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Proceedings of the Budapest Conference, May/June 1976
IAH–IAHS

HYDROGÉOLOGIE DES GRANDS BASSINS SÉDIMENTAIRES

Actes de la Conférence de Budapest, Mai/Juin 1976
AIH–AISH

MŰSZAKI KÖNYVKIADÓ, BUDAPEST

1978

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PREFACE

It was at the Tokyo meeting in 1971 of the *International Association of Hydrogeologists* that the idea of organizing a conference on "*Hydrogeological problems of the great sedimentary basins*" was first aroused. At the same meeting the Hungarian delegate was called upon to make a proposal in this respect. As early as 1972, at the IAH General Assembly held in Montreal, the Hungarian Geological Institute expressed its willingness to organize the conference planned. Consequently, in 1974 a formal invitation was submitted to the General Assembly convened then at Montpellier.

From not later than 1974 the preparatory work for the conference was being shared with the Groundwaters Department of the *International Association of Hydrogeological Sciences,* and the organizers succeeded in taking UNESCO's support for granted. Then the meeting came under the auspices of the Central Office of Geology, the Hungarian National Water Authority and the Hungarian National Commission for UNESCO. In organizational work, the Hungarian Geological Institute was helped by the Hungarian Geological Society and the Hungarian Hydrological Society.

The Organizing Committee included DR. J. KONDA as chairman, DR. A. RÓNAI as secretary general, Mr. L. ÓDOR as secretary and Mrs. Z. LOSONCZY as secretary assistant.

The first circular issued in May 1975 was followed by a second one in February 1976. Discussions were conducted on 1st and 2nd June, and a field trip took place on 3rd and 4th June, 1976.

Organizing Committee

AVANT-PROPOS

C'est lors de la conférence de l'*Association Internationale des Hydrogéologues* tenue en 1971 à Tokyo que l'idée de consacrer une conférence aux *problèmes de l'hydrogéologie des grands bassins sédimentaires* a été évoquée. On a prié le représentant de la Hongrie de soumettre une proposition concernant l'organisation d'une telle conférence. A l'Assemblée Générale de l'AIH à Montréal, l'Institut Géologique de Hongrie a déclaré qu'il était prêt à organiser la Conférence, et il a présenté une invitation officielle à ce propos lors de l'Assemblée Générale tenue en 1974 à Montpellier.

En 1974 la Division des eaux souterraines de l'*Association Internationale des Sciences Hydrologiques* s'est jointe aux efforts de préparations à la Conférence, et les organisateurs y ont acquis du support de la part de l'UNESCO. Le Bureau Central de Géologie de la République Populaire de Hongrie, l'Office National des Eaux et la Commission Nationale Hongroise pour l'UNESCO se sont chargés du patronage de la Conférence. Les travaux d'organisation ont été confiés à l'Institut Géologique de Hongrie auquel la Société Géologique de Hongrie et la Société Hydrologique de Hongrie ont prêté leurs concours.

Le Comité d'Organisation comprenait: le président, DR. J. KONDA; le secrétaire général, DR. A. RÓNAI; le secrétaire, M. L. ÓDOR et son assistante Mme Z. LOSONCZY.

La première circulaire fut distribuée au mois de mai 1975, la deuxième en février 1976. Les réunions de la Conférence ont eu lieu les 1 et 2, l'excursion les 3 et 4 juin 1976.

Comité d'Organisation

PROGRAMME

of the International Hydrogeological Conference (IAH–IAHS)
Budapest

May 31, 1976 (Monday)

Reception of the participants and registration in the Ceremonial Hall of the Hungarian Geological Institute (Budapest XIV, Népstadion út 14).

June 1, 1976 (Tuesday)

The opening of the Conference takes place in the Assembly Hall of the Hungarian Academy of Sciences (I, Országház u. 28).

Introductory addresses

by Dr. J. Konda, Director of the Hungarian Geological Institute
by Dr. J. Fülöp, President of the Central Office of Geology in Hungary
by J. Vincze, Vice-president of the National Water Authority
by Dr. M. Jóború, President of the Hungarian National Commission for UNESCO
by Dr. S. Buchan, President of the IAH
by Dr. Gy. Kovács, Vice-president of the IAHS

Report of the Screening Committee

by Dr. L. Dubertret, Secretary General of the IAH

First session

Chairmen: Mr. Prof. S. Buchan and Mr. Gy. Kovács

Mr. A. Rónai: Hydrogeological Particularities of the Great Hungarian Plain

Mr. Prof. M. Manfredini et al.: Hydrogeological Features of the Po Valley (Read by Mr. Prof. M. Pellegrini)

Second session

Chairmen: Mr. Prof. H. SCHOELLER and Mr. L. SZEBÉNYI

Mr. GY. KOVÁCS and Mr. H. BESENECKER: Keynote addresses on Theme 1 (Subsurface water movements, filtration, hydrodynamics)

Discussion

Reception at the Hungarian Geological Institute (XIV, Népstadion út 14)

June 2, 1976 (Wednesday)

Third session

Chairmen: Mr. L. DUBERTRET and Mr. A. RÓNAI

Mr. Prof. L. STEGENA and Mr. L. SZEBÉNYI: Keynote addresses on Theme 2 (Regional hydrogeological elaborations)

Discussion

Fourth session

Chairmen: Mr. Prof. M. PELLEGRINI and Mr. Prof. W. RICHTER

Mr. Prof. G. GIULIANO and Mr. GY. GRESCHIK: Keynote address on Theme 3 (Hydrogeological mapping, geophysical and geochemical data)

Discussion

Fifth session

Chairmen: Mr. B. REZÁC and Mr. L. ALFÖLDI

Mr. W. A. VISSER and Mr. E. ALMÁSSY: Keynote addresses on Theme 4 (Water exploration, wells, water regime, water pollution)

Discussion

Closing of the Conference

June 3–4, 1976 (Thursday and Friday). Excursion

June 5, 1976 (Saturday). Day of departure

PROGRAMME

détaillé de la Conférence Hydrogéologique Internationale de l'AIH–AISH
à Budapest

31 mai, 1976 (Lundi)

Réception et enregistrement dans la Salle des Conférences de l'Institut Géologique de Hongrie (14, rue Népstadion, Budapest XIV).

1 juin, 1976 (Mardi)

Inauguration de la Conférence dans la Salle des Cérémonies de l'Académie, 28, rue Országház, Budapest I.

Ouverture présidentielle et allocutions par

DR. J. KONDA, directeur de l'Institut Géologique de Hongrie

DR. J. FÜLÖP, président du Bureau Central de Géologie

J. VINCZE, vice-président de l'Office Nationale des Eaux

DR. M. JÓBORÚ, président de la Commission Nationale Hongroise pour l'UNESCO

DR. S. BUCHAN, président de l'AIH

DR. GY. KOVÁCS, vice-président de l'AISH

Compte-rendu du Comité de Sélection des communications reçues par

DR. L. DUBERTRET, secrétaire général de l'AIH

1e Session

Présidée par MM. Prof. S. BUCHAN et GY. KOVÁCS

M. A. RÓNAI: Caractère essentiel hydrogéologiques de la Grande Plaine hongroise

M. Prof. M. MANFREDINI et al.: Particularités hydrogéologiques de la vallée du Pô. (La communication sera présentée par M. Prof. M. PELLEGRINI)

2e Session

Présidée par MM. Prof. H. SCHOELLER et L. SZEBÉNYI

Comptes-rendus des rapporteurs généraux du Thème 1 (Écoulement d'eaux souterraines, hydrodynamique, modèles) par MM. GY. KOVÁCS et H. BESENECKER

Discussion

Cocktail à l'Institut Géologique de Hongrie (14, rue Népstadion, Budapest XIV)

2 juin, 1976 (Mercredi)

3e Session

Présidée par MM. L. DUBERTRET et A. RÓNAI

Comptes-rendus des rapporteurs généraux du Thème 2 (Descriptions hydrogéologiques régionales) par MM. L. STEGENA et L. SZEBÉNYI

Discussion

4e Session

Présidée par MM. Prof. M. PELLEGRINI et W. RICHTER

Comptes-rendus des rapporteurs généraux du Thème 3 (Cartographie hydrogéologique, données géophysiques et géochimiques) par MM. Prof. G. GIULIANO et GY. GRESCHIK

Discussion

5e Session

Présidée par MM. B. REZÁC et L. ALFÖLDI

Comptes-rendus des rapporteurs généraux du Thème 4 (Production d'eau, régime hydraulique, puits, pollution) par MM. W. A. VISSER et E. ALMÁSSY

Discussion

Clôture des sessions de la Conférence

3—4 juin, 1976 (Jeudi et Vendredi). Excursion

5 juin, 1976 (Samedi). Jour de départ des congressistes

INTRODUCTORY ADDRESSES
ALLOCUTIONS PRESIDENTIELLES

Dear Participants,
Ladies and Gentlemen,

It is a real pleasure to me to profite of this opportunity to pronounce, in the name of the Hungarian Geological Institute, a hearty welcome to you all, who have accepted our invitation, who are interested in the subject chosen and who will certainly contribute to our effort towards achieving, during these few days of consultations, valuable results of particular use for both science and practice.

I should like to greet first of all the representatives of the Central Office of Geology and the National Water Authority, our sponsors, who have attributed due importance to the idea of this Conference and who have helped us efficiently in bringing it into completion. Thank is due to the Hungarian National Commission for UNESCO which has paid attention to our meeting and which has followed with attention our organizatory work since the idea of the Conference has been raised.

I feel indebted to the International Association of Hydrogeologists and, personally, to its President, Mr. S. BUCHAN, and its Secretary General, Mr. L. DUBERTRET, who have provided possibility to discuss our special problems together with scientists of international name under the organization and within the frame of this Association. We have also to thank the International Association of Hydrological Sciences and, personally, vice-president Mr. GY. KOVÁCS, for having allowed the Department of Subsurface Waters of the Association to join us in our work and to widen this way the background of our consultations.

From among the organizers I must mention first of all the name of Professor S. VITÁLIS, who was invited to be Chairman of our Conference and who has participated since the very first day, with very great care and attention, in the preparations for the meeting and whose guiding advices and experiences we could not have dispensed with. Unfortunately, his stubborn disease has kept him stuck to bed for many weeks now, so that he has not been able to attend the meeting. It is in his place that I have to expound now also his ideas.

We have been given valuable assistance by both Hungarian scientific associations interested in our subject, by the Hungarian Geological Society and the Hungarian Hydrological Society. We have been similarly backed by the Research Institute for Water Resources, the Roland Eötvös Geophysical Institute and some university institutes which have taken part in the elaboration of the selected topics and in the organization of the excursions.

A positive response has been excited among our foreign colleagues and fellow institutions abroad. Hence the relevant papers received from 20 different countries and the preliminary applications we have received from 37 countries. The number of colleagues virtually present is smaller, but it is not low at all. We have received 244 attending members from 36 different countries and 90 foreign members will participate in our excursion.

I must tell you that, beside the formal, collective support provided by associations, societies, colleges and government organizations, we have enjoyed the contributions of countless private individuals and that members on the staff of the Hungarian Geological Institute, themselves, have excelled in enthusiasm and efficient co-operation.

Our subject is the hydrogeology of large basins. The exploitation of subsurface waters rapidly increases all over the world. Beside the utilization of karstic waters it is the aquifers of the large sediment-filled basins that play an increasing role in this domain. We must be able to distinguish, in the water budget and exploitation of these basins, between fossil waters and waters recharged by surface precipitations. To be able to assess the possibilities for infiltration and accumulation of water in aquifers of different permeability and tectonic setting requires from the geologist to get acquainted, in addition to the traditional research tools of geology, with the regularities of rock physics and hydrodynamics. The assessment of the variation of the quality of the water in the water-bearing rocks and on its path of migration underground requires the use of hydrochemical research methods beside hydrodynamic ones. And this need is becoming more and more urgent because of the many kinds of pollutions, natural and artificial, which are threatening the groundwaters as well. Geologists, engineers and specialists in many other fields are getting involved in the investigations of groundwaters gradually increasing in importance. And they are to play a special role in solving the tasks of the international exchange of experiences and in concentrated international efforts.

In the spirit of these ideas I declare the Conference open and wish all guests present a good and successful work with my heartiest greeting and the desire of real hospitality.

Dr. J. Konda
Director of the Hungarian
Geological Institute
Chairman of the Organizing
Committee

Ladies and Gentlemen, dear Colleagues,

During the history of Hungarian geology, hydrogeology found its own way of development rather early, gradually acquiring rather significant importance. Its development differs to quite a degree from the international pattern. Whereas in most countries hydrogeology represented first of all an aid for mining, seeking to solve the hydrogeological problems arising in the course of mining activities, in Hungary, in addition, it has become the basic discipline of lowland water recovery. This specialization was started as early as the first decades of systematic geological research and, characteristically enough, V. Zsigmondy, a mining engineer enjoying international reputation in the domain of water prospecting, was to play a very active role in this development. Although water control in mines was significant in Hungary, just like in many other countries, and though quite a number of theoretically and practically valuable schemes had been developed and though, in addition, water supplies for cities and industrial plants required, both in the mountainous and hilly regions, plenty of hydrogeological research to be carried out, occasionally combined with large-scale testing and recovery works (e.g. the Municipal Waterworks Project of Budapest, the water supply of the city of Pécs, etc.), the major part of the activities of the Hydrogeological Department established in 1882 in the Hungarian Geological Institute consisted in the drafting of expertises for artesian water well drilling.

By the early 1900's, data on more than 300 wells of this kind had been collected and processed.

Between the two World Wars, the most significant contributions were represented by the publication of the "Hydrogeology of Budapest" by H. Horusitzky and the cadastral registration of the artesian wells as a supplement to Kreybig's map of soils. A work of similar importance, both theoretical and practical, was published in 1929 on the geothermal conditions of the Alföld—the Great Hungarian Plain.

From 1950 on, new data collecting and mapping projects were launched over this lowland area as well as in the Transdanubian Central Mountains, where the karstic water conditions had to be examined in connection with coal- and bauxite-mining.

Beside the registration of the surface water resources, there arose the need for assessing the quantity of subsurface waters, to explore their reserves and the origin of their recharge. This required to apply basically geological methods of research: determine the extension and geological features of the water-bearing subsurface formations and clear up their interconnections and relations to impervious strata.

The very first step toward achieving this goal was to study geologically the phreatic groundwater—the subsurface water most directly communicating with the land surface. After the development of a major cadastral well registration system comprising the measurement and mapping of 1.2 million wells, a monograph on the groundwaters of the Hungarian Basins was compiled by A. Rónai and published by the Hungarian Geological Institute.

At the same time a new survey and registration of the artesian wells was started. The first product of that work was the publication of the "Hydrogeological Atlas of Hungary" compiled by the staff of the Geological Institute's Hydrogeological Department under the direction of R. E. Schmidt.

Later on, the competent Water Authorities took over the charge of data collecting, whose results were published in the National Cadastrial Register of Wells compiled by J. Urbancsek.

At present, hydrogeological research work is being conducted, both in the highland and lowland areas, by the staffs of the regionally organized geological mapping departments of the Geological Institute. The scientific level of these research activities and the achievement of tangible results have been greatly enhanced by the large-scale and complex geological and hydrogeological research projects, based on a proper theoretical and methodological background, which have regularly been launched since 1964 in the Great Hungarian Plain. In executing them, we have relied partly on boreholes of medium to great depths (100 to 1500 m) put down along selected major geological section lines, and partly on closely-spaced shallow boreholes (10 to 20 m) arranged in a dense grid. In addition to yielding new important results, these investigations have produced plenty of evidence, now replacing the earlier hypotheses and worth of consideration even on the international scale.

Reports on these results will be presented at this Conference.

To conclude, on behalf of the Central Office of Geology, I should like to express a most cordial welcome to the organizers of this Conference, who for quite some time have carefully and enthusiastically prepared for this important international scientific consultation. So I am really happy to observe this exemplary co-operation of the representatives of two different disciplines, geology and hydrology. I also welcome the representatives of the International Association of Hydrogeologists, the Hungarian National Commission for UNESCO and the National Water Authority.

Similarly cordial welcome goes to all Hungarian and foreign guests, to the director of the Hungarian Geological Institute, our host, and, in person, to Mr. A. Rónai who has had a lion's share in convening this international meeting of scientists.

Dr. J. Fülöp
President
of the Central Office of Geology

Ladies and Gentlemen,

On behalf of the National Water Authority of the Hungarian People's Republic I cordially welcome our foreign guests, the participants in this Conference who are representatives of a science very important for us. Subsurface waters have for long been of particular importance in this country, because Hungary, as her geological structure shows, is a typically basin territory. Hence the particular attention paid to the topics of this Conference by Hungary's water management officials.

Those of the participants who already visited Hungary previously could have witnessed, and those who have come here for the first time will see, that the Hungarian economy develops at a very fast rate. In this general framework both large-scale-farming agriculture and the industry playing a leading role in the country's economy call for more and more water. On account of Hungary's peculiar climatic conditions, we have to use irrigation over vast areas and there are several, highly developed industrial branches, e.g. the food and pharmaceutical industries, calling for very good quality of water. The multiplication of large cattle-breeding plants; the industrialization of the provinces; the associated housing projects reflecting the rise in living standards and the increasing demand for the amenities of living on a higher cultural level are associated with a simultaneous increase in water demand. So the rational and economical utilization of our resources need a particularly well-founded planning.

What is then the role of the underground waters, what is that of their management in this development? To give an answer to this question, you should know that 95 per cent of Hungary's surface water reserves come from the neighbouring countries situated hypsometrically higher than Hungary and the overwhelming majority of this water travels down in the streambeds of our two major rivers—the Danube and the Tisza. In the Tisza valley we have to build highly expensive barrages and associated water distribution systems in order to meet the demands. And if we wish to supply these waters to farther areas, we must develop sophisticated water management systems.

Favourable hydrogeological conditions, however, allow us to explore the subsurface waters with relatively low inputs even in areas situated farther away from the occurrences of surface waters. The major part of these water reserves in our country is protected by Nature from man's subaerial polluting activities, thus being available in satisfactory quantity and quality in the long run. This will explain that up to 80 per cent of the tapwater supplied by our waterworks are of subsurface origin and this figure is expected to remain practically unchanged in the coming decades.

Peculiar problems are concomitant of these developments. One of these problems is that while previously the demand was smaller and could be satisfied by the use of a great number of single wells, present-day and future needs have to be satisfied by large regional water supply systems capable of servicing two or more settlements. These will require to erect large, concentrated water-producing plants. Since these are built step by step, it is necessary to forecast reliably for a considerable length of time the potential regional range of their influence. In the concerned regions there are other users, whose social and economic importance, though secondary after the water supply of the population, cannot be neglected. Taking into consideration the potential increase of surface water pollution it is quite clear that the subsurface waters of good quality protected from such effects ought to be conserved primarily as the source of communal water supply.

Therefore, first of all, we have to determine the available resources and we have to protect them and keep them in their original state in the vicinity of planned water-producing plants until their utilization is commenced. I think it is clear to all of you, that we have to go on with great care and circumspection, because any underestimation on our part will hinder the development of the planned waterworks in the long run, while exaggerated precaution may force other users to spend unnecessarily large amounts of money on the water consumed.

Economic progress is accompanied, in Hungary too, by increasing environmental pollution hazards. So we have to institute measures and prevent activities letting subaerial pollutions penetrate underground. It is well-known, that measures of this kind presuppose the knowledge of the means of diffusion of different kinds of pollution of surface origin and that many questions still await solution in this field.

The worldwide energy crisis induces us to study how we could better utilize our geothermal energy. In Hungary there are considerable achievements in using thermal water for agriculture, the heating and hot water supply of habitations and balneology. Thermal waters are easy to recover in several regions of this country, the resources being considerable, though not infinite. Therefore, in addition to increasing exploration, we want to improve the means of thermal water exploitation by avoiding excessive use and seeking to enhance versatility.

I think you may know best that we cannot master even these few problems selected as most important from among the topics intriguing us, unless making use of the up-to-date results of hdyrogeology. Therefore we, responsible for water management in Hungary, are eager to hear and look forward to hearing about the results which may help promoting, directly or indirectly, the further development of production and management methods by improving the accuracy of subsurface water reserve calculations; increasing the reliability of forecasts on underground processes and making positive the interaction between man and natural environment.

Ladies and Gentlemen,

I feel that, even though briefly, I could give you an idea of the significance of up-to-date hydrogeological methods and achievements for the development of water management in Hunga ryrI hope, you will be interested in one another's ideas and results. I wish, ou foreign guests may have some opportunity to get familiar with our way of living, our results and worries. I hope, you will have a pleasant stay in Hungary leaving in you good impressions when you will be returning home.

J. VINCZE
Vice-president
of the National Water Authority

Mesdames et Messieurs,

Permettez-moi de vous saluer, tout d'abord, au nom de la Commission Nationale Hongroise pour l'UNESCO et vous souhaiter beaucoup de succès.

C'est depuis de plus que deux décennies que la Hongrie participe aux activités de l'UNESCO, aux actions et projets de cette organisation, entre autres, à ceux consacrés aux sciences naturelles et, précisément, aux projets hydrogéologiques et géologiques. De nombreux spécialistes hongrois ont participé et participent toujours aux différentes conférences, séminaires, symposiums etc. organisés par l'UNESCO et plusieurs experts hongrois travaillent dans différents pays en voie de développement. Autant que je sache, leur travaux sont partout reconnus et appréciés. A ma connaissance, nos scientifiques jouent un rôle important aussi dans vos deux organisations.

La Hongrie est le lieu de beaucoup de réunions scientifiques internationales organisées chaque année par l'UNESCO et par d'autres organisations internationales. Nous sommes très contents de ce fait, car nous sommes persuadés qu'elles contribuent considérablement à une coopération internationale intense dans les domaines respectifs. C'est pourquoi, il a été pour nous une heureuse nouvelle d'apprendre que les dirigeants de l'Association Internationale des Hydrogéologues et de l'Association Internationale de l'Hydrologie Scientifique avaient décidé de tenir cette réunion ici, en Hongrie.

Mesdames et Messieurs,

Ma profession est très-très éloignée de la vôtre, tout de même je sais que l'hydrologie, la géologie et l'hydrogéologie jouent un rôle extrèmement important à notre époque. L'époque dans laquelle nous vivons est caractérisée par le fait qu'alors que la science et la technique ont sauvé l'humanité de beaucoup de dangers et de dévastations par des forces destructives dont on souffrait autant pendant les siècles passés et même pendant les décennies récentes, et alors que la science et la technique ont ainsi rendu la vie des hommes plus humaine de plusieurs points de vue, les possibilités de nouveaux dangers et de nouvelles catastrophes se sont présentés, et la vie d'une grande partie de l'humanité est toujours indigne de l'homme. Je pense que vous êtes d'accord avec mon opinion en disant que, de ce fait, l'importance des sciences et, plus précisément, de vos disciplines est devenue plus accentuée et la responsabilité des savants s'est accrue.

J'ai appris des chercheurs de l'Institut Géologique National de Hongrie que l'Alföld, la Grande Plaine hongroise qui est pour nous, Hongrois et Hongroises, une région bien connue de notre pays, une région charmante et agréable, le pays natal de beaucoup de nous, y compris moi-même et qui, pour beaucoup de touristes étrangers, n'est guère que la « puszta » romantique, où on cherche des « tchardas », des gardiens de chevaux et peut-être même des brigands « betyár », cette même Grande Plaine hongroise représente, du point de vue scientifique, l'un des bassins sédimentaires les mieux étudiés et les mieux explorés de l'Europe. Je suis de l'avis qu'outre la renommée internationale des géologues, hydrologues et hydrogéologues hongrois, ce fait a également contribué à ce que, pour notre grande joie, on a choisi la Hongrie pour le lieu de cette conférence.

Je suis contente de voir dans votre programme qu'en outre des sessions, vous aurez la possibilité aussi de faire des excursions un peu plus loin de la capitale. J'espère que cela vous offrira non seulement des parcours à travers les paysages de notre pays, mais qu'il vous permettra également de connaître quelque-chose de la vie de nos villes et villages et de la manière de vivre de notre peuple.

Finalement, permettez-moi d'exprimer ma conviction que cette conférence, ensemble avec d'autres réunions internationales, contribuera à l'élargissement de la coopération internationale dans le domaine des sciences, qui est un facteur très important de la compréhension internationale et de la coexistance pacifique. Je vous souhaite, d'une manière réitérée, des travaux fructueux, plein de succès.

Dr. M. Jóború
Président
de la Commission Nationale
Hongroise pour l'UNESCO

Mr. Chairman, Presidents, Ladies and Gentlemen,

When the proposal to hold this Conference was first received by the International Association of Hydrogeologists we had no hesitation in agreeing to sponsor it. Hungarian research in the field of hydrogeology is legendary and few hydrogeologists can be unaware of the excellent contributions which Hungarian hydrogeologists have made. I recall with great pleasure receiving a year ago a splendid hydrogeological atlas which could be used as a model anywhere in the world for the presentation of data in concise form. For me the atlas had the added advantage that it included explanations in English.

The proposal to hold this Conference in Budapest gave us an excellent opportunity to invite the International Association of Hydrological Sciences to join us in joint sponsorship and I am very pleased indeed to have Mr. KOVÁCS next to me as evidence of the happy collaboration of both our Associations. About the time of the early arrangements for this Conference the UNESCO programme called the International Hydrological Decade was drawing to a close and plans were being made for a new programme called the International Hydrological Programme. It was appropriate therefore that UNESCO should be interested in this Conference and we welcomed their decision to join as co-sponsors, thus ensuring the support of three bodies with special interest in hydrogeology.

We have had a wonderful welcome from DR. KONDA and we have listened with great pleasure to the encouraging remarks made by Hungarian Presidents of the National Water Authority, the Central Bureau of Geology and UNESCO who have stressed the importance of our meeting.

In the audience I am happy to see a mixture of engineers and geologists. Sometimes engineers and geologists disagree with one another, sometimes there is rivalry between them and sometimes they are impatient with one another. This is understandable, their approach to problems is very different. When the foundations of a bridge collapse the first reaction of the engineer is how to restore them. The first reaction of the geologist is why did it happen and if he takes a long time to find out it is not surprising that the engineer becomes a little impatient. However, when the engineer and the geologist do work together, as they are doing in hydrogeology, they make a formidable team.

Among the aims of the International Association of Hydrogeologists is to encourage studies, to promote collaboration between geologists and specialists of all disciplines who are interested in the study of hydrogeology, to collaborate with other organizations, to promote meetings including those with other organizations and to encourage the publication of the results of scientific studies. These aims are being fulfilled very effectively on this occasion.

One of the activities of IAH is to form commissions of specialists for the intensive study of selected topics. One of its Commissions was set up to make a hydrogeological map of Europe. This work began in the heart of Europe and involved many nations in truly international collaboration which is already bearing fruit by the publication of several sheets of the map of Europe.

Another Commission was set up to study mineral and thermal waters especially in Europe where they are particularly important. Initially the work concentrated in the preparation of a glossary to overcome language difficulties and the preparation of a legend as well as model maps.

The third Commission studies Karst, a subject of special interest in the Eastern Mediterranean. National studies have grown into regional studies and last year culminated in the publication of a Monograph on the hydrogeology of Karst terrains and an international meeting in Alabama to which American and Mediterranean research results were presented and discussed, along with contributions from other countries.

At present, thought is being given to the hydrogeology of volcanic rocks and members of the Italian and Spanish National Committees are making a preliminary study of the state of knowledge and whether or not a study of the subject in depth is likely to yield results of scientific and practical importance. Their recommendations will be considered by IAH along with proposals for the study of other topics.

This brings me back to the present meeting where we are about to discuss the hydrogeology of a great sedimentary basin. Sedimentary basins are among the most important sources of ground water, occurring widely in Africa, North and South America, Australia, India, Europe and other parts of the world. Some of the basins have been intensively studied—the Danubian Basin for example—but others have been neglected.

This Conference focuses attention on the importance of the great sedimentary basins and gives rise to the thought that their study on a worldwide basis would be a suitable subject for the attention of IAH. It may be that our work over the next few days will lay the foundation for a new item in the scientific programme of IAH.

I look forward to a very happy meeting during which we shall have many opportunities to meet scientists with whom we have corresponded or whose published work we have read, and to exchange scientific experiences and ideas for future research. I also look forward to the scientific programme and the presentation of Hungarian research results and their discussion, as well as the privilege of visiting the Danubian Basin.

I understand that Hungary is a wine producing country and that we may hear something about the influence of geology on the wine. Probably before we return to Budapest we shall also have had an opportunity to study the effect of the wine on the geologists.

On behalf of the International Association of Hydrogeologists I thank the Hungarian scientists and their sponsors for the invitation to join this Conference in Budapest and also for the warm welcome we have received.

DR. S. BUCHAN
President of the IAH

Mr. Chairman, Ladies and Gentlemen,

First of all on behalf of the Bureau of the International Association of the Hydrological Sciences I should like to greet all the participants of this symposium. I think it would not be appropriate to take long time for the opening address, the scientific sessions are the places for lecturing. It is the reason why I should like to emphasize only two aspects here.

The first is the co-operation between the two international associations IAH and IAHS as it was already mentioned by Professor Buchan. In our Association we were trying to determine the correct definition of hydrology and we have found that hydrology is not an independent branch of sciences, but it has to include all the other scientific activities which are dealing with the water as it is moving along the large hydrological cycle. If somebody dealing with hydraulics or hydrobiology or hydrochemistry investigates water as a part of the large cycle and in connection with the surrounding branches of the cycle, —in our opinion—this work is a part of hydrology. This is the reason why it was felt by many scientists that the present structure of the scientific international associations separates many-many small parts of this large work, which should be unified. The first attempt to establish a wide form of co-operation was made in 1974, when a Presidential Council of the water-oriented associations was established by four associations: ICID, ICOLD, IAHS and IAHR. One or two years later we were very glad when the International Association of Hydrogeologists has joined also this Presidential Council and recently seven among the most important water-oriented associations are working together in the framework of this new form of co-operation. Naturally all the associations have preserved their original way of work, structure, independence. The task of the Presidential Council is only to harmonize our activities and to promote the organization of such joined meetings symposia as it is done now here, in Budapest. I think, this symposium is one of the first examples of this new type of co-operation. This is the reason why our Association, IAHS highly appreciates the organization of this symposium and emphasizes, that this good example has to be followed in the future.

The other aspect, what I should like to emphasize now, is the importance of the ground-water in the general water resources development. As it is well-known, one of the main objectives of the International Hydrological Decade and now that of the International Hydrological Program (which is the continuation of the Decade) is to determine the regional, continental and world-wide water balances. The practical purpose of the investigations is to promote the best utilization of the very limited water resources. There are many investigations showing that the water resources over our globe form a non-separable unit: We cannot do sharp distinction between ground-water and surface-water resources, because they are so closely interrelated, that if we use ground-water, this amount is missing from the surface-water. At the same time the use of ground-water has many-many advantages. It is—at least in the large basins— evenly distributed, while the surface-water resources are always concentrated in river beds. You will see when visiting some part of Hungary, that the density of the river beds is very small here and therefore between two large rivers there is only one type of resources to be used, and that is ground-water. The other great advantage is that ground-water is better protected against pollution than surface-water. If we want to increase the available surface water resources we have to construct large dams, establish large reservoirs. At the same time below the surface we have natural reservoirs, the ground-water regime is governed by natural storage processes, and if we can govern the storage large quantities of water become available for utilization. For the control of subsurface run-off it is necessary to know all the phenomena influencing and governing the ground-water regime within the large sedimentary basins. The high importance of the ground-water resources in general water resources development and the close contact between surface- and ground-water is the reason why IAHS thought that the investigation of the hydrogeology of the large sedimentary basins is a very important topic also from hydrological point of view. The detailed discussions on this topic provide us with many important aspects for the investigation of the whole large natural hydrological cycle. This is the second reason why we were very glad when we were asked to join the organization of this symposium.

Closing my words I should like also to thank the Organizing Committee and especially the organizers and Dr. Rónai for the enormous amount of job done for the preparation of this meeting and I should like to greet once again the participants of the symposium. Thank you for your attention.

Dr. Gy. Kovács
Vice-president of the IAHS

COMPTE-RENDU DU COMITÉ DE SÉLECTION DES COMMUNICATIONS REÇUES

L. Dubertret
Secrétaire Général d'AIH

La première circulaire de la Conférence de Budapest a été distribuée au printemps de 1976. En réponse à cette invitation, 261 personnes ont annoncé leur intention de participation, et nous avons reçu 62 communications in extenso et les résumés de 84 communications avant l'expiration des délais imposés. En plus, 13 communications complètes ont été reçues après les délais. Ainsi on en a reçu 75 au total. Le Comité d'Organisation a pris la décision de soumettre à la discussion chacune des communications reçues et de différer la décision définitive à propos de leur inclusion au mémoire jusqu'au moment où on aurait les résultats des discussions respectives.

Plusieurs communications ont dépassé les limites prévues; ni les dimensions des illustrations graphiques, ni leur presentation ne correspondaient toujours aux conditions fixées dans la 1e Circulaire. Nous avons sollicité les auteurs de ces communications de réduire leur textes et d'assurer une qualité acceptable des illustrations.

En plusieurs cas, les rapporteurs généraux ont dû recevoir des communications non encore réduites, avec des illustrations pas convenablement remaniées. En effet, à l'égard du temps disponible bien court, on considérait plus sage de ne pas perdre le temps en attendant, car les rapporteurs généraux allaient courir le risque de ne pas pouvoir accomplir leur devoir à temps. Avant de remettre les manuscrits des communications à l'impression, le Comité d'Organisation fera son mieux pour qu'elles satisfassent aux exigences.

En préparant les communications à la discussion, nous les avons groupées par thèmes et nous avons invité des rapporteurs généraux à les étudier et à en donner des commentaires critiques. Nous avons invité deux rapporteurs généraux pour chaque groupe de communications. Les auteurs et tous ceux désirant participer aux discussions sur les comptes-rendus des rapporteurs généraux auront la possibilité de prendre la parole pour une durée limitée.

Rapporteurs généraux des thèmes choisis

1^er^ Thème — Écoulement d'eaux souterraines; hydrodynamique; modèles. Rapporteurs généraux: Dr. György Kovács, vice-président d'AISH, chef de département de l'Office National des Eaux, Budapest. — Dr. Horst Besenecker, Niedersächsisches Landesamt für Bodenforschung, Hannover.

2^e^ Thème — Descriptions hydrogéologiques régionales; tectonique et hydrogéologie. Eaux thermales et minérales.
Rapporteurs généraux: Dr. Lajos Stegena, prof. d'université, Budapest. — Dr. Lajos Szebényi, directeur du Service de Documentation de l'Institut Géologique de Hongrie.

3^e^ Thème — Cartographie hydrogéologique; données géophysiques et géochimiques.
Rapporteurs généraux: Dr. Giuseppe Giuliano, Istituto di Ricerca Sulle Aque, Rome. — Gyula Greschik, ingénieur en chef départemental de l'Entreprise METRO.

4^e^ Thème — Production d'eau; puits; régime hydraulique; bilan d'eaux; énergie géothermique souterraine. Problèmes de pollution.
Rapporteurs généraux: Dr. Willem Visser, Organization for Applied Sciences, Delft, The Netherlands. — Endre Almássy, ingénieur, chef de Section de l'Office National des Eaux, Budapest.

Afin de mettre les participants au courant des problèmes à discuter, nous avons inséré les résumés reçus dans un volume, qui est à distribuer parmi les participants avant l'ouverture de la Conférence. Nous avons inclus dans ce cahier aussi les résumés des communications dont le texte complet ne nous est encore pas parvenu. De cette manière nous avons voulu élargir la gamme des problèmes à discuter et offrir la possibilité d'une brève contribution orale même aux collègues qui n'avaient pas réussi de préparer leur communication jusqu'au délai fixé. — Quelques communications sont arrivées après le délai prolongé. Les auteurs de celles-ci auront la possibilité d'en donner des renseignements brefs. Quant à la question de leur inclusion aux mémoires de la Conférence, elle ne sera décidée qu'après la révision technico-scientifique des textes.

Les auteurs de deux communications spécialement choisies ont été priés de donner un exposé plus détaillé de leurs travaux, mais l'aperçu critique à donner par les rapporteurs généraux embrassera même ces deux communica-
ns.
itoL'une d'elles, écrite par M. A. Rónai, porte le titre: « Caractère essentiel hydrogéologique de la Grande Plaine Hongroise ».

L'autre, écrite par un ensembles de co-auteurs italiens (C. Bortolami — G. Braga — A. Colombetti — A. Dal Pra — V. Francani — F. Francavilla — G. Giuliano — M. Manfredini — M. Pellegrini — F. Petrucci — R. Pozzi — S. Stefanini) (Torino, Pavia, Modena, Padova, Milano, Bologna, Roma, Parma, Trieste — Italie) est consacrée aux particularités hydrogéologiques de la vallée du Pô.

La discussion sur ces deux communications, ainsi que des autres, sera ouverte après les comptes-rendus des rapporteurs généraux du Thème 2, le 2 juin, pendant la matinée.

L'Institut Géologique de Hongrie s'est engagé à publier les communications de la Conférence et à faire parvenir aux membres le volume imprimé avant la fin de 1977.

Le Screening Committee comprenait les membres suivants: DR. J. KONDA, DR. L. DUBERTRET, DR. GY. KOVÁCS et DR. A. RÓNAI.

* * *

REMARQUES DU RÉDACTEUR

Le Comité de Sélection a remis toutes les communications aux Rapporteurs Généraux. Elles ont été discutées lors de la Conférence. Avant d'être préparées à l'impression, elles furent révisées tant au point de vue de leur présentation scientifique qu'à celui de leur langage. Suivant l'opinion des réviseurs, 14 communications n'ont pas pu être acceptées pour publication. Lors de la révision du langage et de la présentation des communications on s'est restreint à des corrections et abrègements inévitables, sans modifier leur contenu essentiel, ni leur style. Par conséquent, il y a des différences sensibles entre les communications en ce qui concerne leurs niveaux scientifique et stylistique. Le Comité n'a pu accepter qu'une seule communication de chaque auteur, 3 communications étaient refusées par cette même raison.

EDITOR'S REMARKS

The papers received by the Screening Committee were, without exception, handed over to General Rapporteurs and later discussed at the Conference. Before being prepared for printing, manuscripts had been scrutinized regarding both their technical content and style. Relying upon the readers' opinion, the Committee had no choice but omitting to publish 14 papers for this very reason or because only one paper could be accepted from the same author. The grammatical checking of manuscripts was limited to the execution of minor corrections and some necessary abridgements. It is hoped, however, that the scientific and practical value of most articles will overshadow any inconsistencies in technical content and style.

Theme 1

SUBSURFACE WATER MOVEMENTS, FILTRATION, HYDRODYNAMICS, MODELS

Thème 1

ÉCOULEMENT D'EAUX SOUTERRAINES, HYDRODINAMIQUE, MODÈLES

GENERAL REPORT 1

Dr. Gy. Kovács

Head, Department of Water Resources Development, National Water Authority, Budapest, Hungary
Vice-president, International Association of Hydrological Sciences

The general objective of the hydrological investigation of ground-water is to determine the available amount of water stored below the surface, or moving along the subsurface section of the hydrological cycle. This type of study, however, is basically different from the determination of surface-water resources. It is necessary, therefore, to develop the special methods of ground-water hydrology.

Dealing with surface-water the series of the collected data of daily flow rates provide us with the inventory of the available natural water resources, the discharge expected with a given probability can be directly calculated from the sufficiently long records by applying the well known statistical methods. The amount of available water can be raised considerably above that ensured by the natural conditions, if reservoirs are constructed and water is stored during the flood-periods. Inventory of water resources cannot be determined directly from the hydrological observations in water systems including reservoirs. Some further investigations are needed considering the way of operation of the reservoirs and their interactions as well. This is the reason, why a new branch of hydrology has recently developed, the so-called system-hydrology or the hydrological application of system-analysis, which try to follow by models the random natural events and the deterministic human activities influencing the development of hydrological processes. These models establish contacts between the inputs and outputs of the system simulating the internal operation of the latter.

Ground-water is one of the natural systems — and perhaps the most important one — where the amount of available water resources is fundamentally influenced by the process of storage. Detailed investigations are necessary, therefore, in all studies dealing with ground-water to prepare the inventory of the ground-water resources. Similarly to any other storage system, the calculation of the water amount available from ground-water depends on the operation of the reservoirs. In this case, however, the operation of the reservoirs is not under human control (or it can be influenced by man's activities only in a very limited extent) the process of storage greatly depends on uncontrolled natural phenomena. Thus ground-water hydrology can be regarded as a special branch of system-hydrology. The basic ideas of system-analysis can be utilized as a guide-line, and methods developed for this concept can be used to create mathematical models and to solve them by applying numerical, analogue or computer techniques.

Considering the general objective explained in the previous paragraphs the direct purposes of the hydrological investigation of the large sedimentary basins can be summarized as follows:

– investigate the subsurface branch of the hydrological cycle and describe the phenomena governing the regime of the ground-water;

– analyse the dynamic processes influencing the storage and the changes of the amount of water stored below the surface;

– explore the interrelations between ground-water and both hydrological and meteorological phenomena occurring on and above the land surface;

– investigate the contact and water exchange between the various aquifers;

– determine quantitative (numerical) parameters characterizing the interactions mentioned in the previous points, the ground-water flow and the dynamic conditions governing the change in storage, giving in this way the complete description of the ground-water regime;

– predict some changes expected as the results of the modified natural conditions or that of planned artificial effects (human activities).

Ground-water hydrology has no well defined place within the structure of sciences, but covers the field of boundaries existing between some of the various disciplines. Only a small part of its domain is real hydrology. It includes a large part of hydrodynamics, because the relationship between hydrological events and the movement of water is closer within a ground-water reservoir than it is in the case of surface-waters. Ground-water hydrology deals with some problems pertaining originally to geology, geochemistry, to soil science and agriculture, because the ground-water phenomena are influenced by the structure and the chemical behaviour of the soil and by vegetation as well. There are, however, some basic principles to be considered in any type of the investigations of the ground-water independently of the discipline, ·om the direction of which the topic is approached.

As a first fundamental aspect the importance of the consideration of the dynamically balanced condition should be mentioned. All ground-water catchments must be regarded as a storage system and, therefore, the equation of continuity is always valid for such catchment. This statement means, that the difference between inflow and outflow summarized for an investigated period should be equal to the change in the amount of water stored in the system. The fluctuation of the ground-water level indicating the storage in wet seasons or periods and the drainage during the dry ones is commonly observed under natural conditions. It can be also stated, however, that the wet and dry periods counterbalance each other, if the investigated interval is long enough and, therefore, the system is in a state of equilibrium, the change of the stored water is equal to zero. The validity of this statement is quite evident considering the fact, that the natural effects influencing the balance of the ground-water space are acting since geological ages without any considerable modification and having only periodical fluctuation. Average water table can only develop, therefore, at a level, where recharge and drainage are in equilibrium.

It was also mentioned, that this balanced condition should be regarded dynamically. The adjective "dynamic" means the consideration of two facts. First of all it has to be taken into account that a considerable part of the ground-water is in continuous movement and, therefore, the equilibrium has to be expressed not only in the form of water amounts (the recharged amount is equal to the drained one if the storage is not changed during the investigated period) but the energy balance (the balanced conditions of the acting forces) has to be also investigated (the difference in the energy contents of the recharged and drained mass—considering also change of the energy of the stored water—has to be equal to the energy consumed by the resistances acting against the flow of water). The other important aspect originates from the first one. The realization of the dynamic concept requires the investigation of the accumulation of the energy content as well apart from that of the changes in the stored amount of water. The condition of equilibrium given in the previous paragraph by equalling the recharged and drained amount of water (supposing that the change of storage is negligible for the investigated period) has to be supplemented. This second condition of the equilibrium states, that the energy accumulated in the system is equal to the amount necessary and sufficient for the maintenance of the flow developed to ensure the mass balance. In such systems, however, where the flow is hindered by impervious layers and movement did not yet started, the energy content has not achieved the upper limit of the latent energy determined by the resistance of the system belonging to the static condition. If the accumulation of energy is a continuous process, it is only a question of time, when the stored energy will surpass the limit and the flow will develop in the system. Thus the imperviousness of a layer is not an absolute term, but it is related to the instantaneous energy content of the whole system.

When a human effect commence to influence the ground-water, the water table will change continuously until the development of a new equilibrium of both natural and artificial phenomena. Naturally this modified balanced condition has to be also interpreted in dynamic sense, its parameters depend on the available energy content increased by the man's influence in question as well. Thus investigating the expected new conditions created by human activities the change in time of both moving and stored amounts of water as well as that in energy content have to be predicted to determine whether there exists a new water regime (described generally by a new position of the water table) at all satisfying the required conditions of the equilibrium, or the continuous change of the situation has to be foreseen for a long period.

The study of natural and artificial balances requires the knowledge of the various processes recharging and draining the ground-water space such as infiltration, evaporation and evapotranspiration, percolation to surface-water bodies (base flow), recharge of ground-water from surface-waters, as well as the water exchange between water-bearing layers and/or ground-water catchments. For this purpose the recharging and draining areas of the ground-water must be distinguished and also the stretches determined, through which the movements create contacts between the two different types of areas ensuring in this way the overall mass-balance within the entire basin. The basic data required for these types of investigations are:

– the geometry of the water transporting layers;

– the physical parameters of the system composed of different layers sufficient to characterize the resistance against the water movement either in static condition or in the presence of a flow with given velocity developed already in the system;

– the dynamic conditions of the system, i.e. the forces and pressure values prevailing in the various layers or along their different stretches resulted by the accumulation of energy content;

– the numerical values describing the flow through the system (i.e. velocity, flow rate, change in energy content, etc.) calculated from the physical and dynamic parameters.

The aspects listed previously raise two further basic principles which have to be considered in the hydrological investigation of the ground-water. The first is the importance of movement and consequently that of the hydrodynamics of the systems. The second originates from the previous one. If the flow is an important factor, the system can be investigated only by taking the whole structure into account at the same time.

In connection with the use of hydrodynamic aspects in ground-water hydrology MAXEY's (1969) lecture delivered at the first International Seminar for Hydrology Professors (Urbana, Illinois) can be quoted. He stated there, that the most important new feature of recent hydrogeological investigations is the combination of hydrological and dynamic methods. He recalled the results presented by TÓTH (1962; 1963) which publications were also quoted in numerous papers submitted to this Symposium. The basic idea of TÓTH's investigation is, that he considers the whole mass of the sediments within a basin as an unified flow-system. The boundary conditions along both the surface and the lower impervious contour of the basin can be determined. The impervious boundary is always a flow line, while the potential conditions at the surface can be characterized by the position of the water-table. The entry- and exit-faces of the flow-space can be also separated if the main recharging and draining effects are known. Geological and geophysical explorations can provide us with informations on the inhomogeneity, the layered character and the anisotropy of the field. Taking into account these flow conditions together with the boundary conditions the probable flow-net characterizing the seepage developing through the entire basin can be constructed by applying e.g. analogue-models.

The considerable change of the local potentials along the surface within relatively short distances, as well as the inhomogeneity and the anisotropy of the field may result the development of separated flow-systems independent of each other within the unified flow space. Local and generally shallow flow-fields develop near the surface directed from the higher terrains towards the valleys. The paths between the recharging and draining areas are relatively short in these systems and the flow rate is, therefore, high in most cases as compared to the deeply penetrating part of the whole water movement. The flow-lines following the contour of the basin may contact regions located far away from each other. These regional flow-systems start generally from the edges of the basins and the water transported along these paths (where the hydraulic gradient and, therefore, both velocity and the flow rate are usually low because of the considerable length) reaches the surface at the main draining points. In most cases there are intermediate systems between the local and regional ones enveloping a few local flow-spaces and penetrating some hundred meters. The development of the local systems is generally governed by the morphology of the surface, while the extension of the intermediate and regional systems are mostly determined by the geological conditions.

This concept which combines the hydrological and hydrodynamic aspects and serves the recent new trend of the hydrogeological methods (as MAXEY has pointed out) is generally accepted now which fact is clearly indicated by many papers presented at the Symposium. The method is very popular also here in Hungary, where a similar way of thinking was developed and published by ALMÁSSY (1962) (just a few months before the publication of TÓTH's investigation) for the characterization of the pressure conditions of the ground-water within the Hungarian basin. Using the data of numerous deep wells he has constructed the equipotential lines in some characteristic sections of the basin. This work was continued by ERDÉLYI (1972), who has supplemented the evaluated data by adding the informations collected from boreholes drilled in the last decades. He was able in this way to extend the investigation to the whole basin instead of the selected sections. All these results are summarized in ERDÉLYI's paper presented at the Symposium.

This dynamic approach of the hydrological investigation of the ground-water is absolutely clear and theoretically well founded. Its practical application still has caused difficulties and sharp debates between the scientists dealing with hydrogeology. This debate is not closed yet, as it is indicated also by the content of the papers presented at the Symposium.

Most of the authors describing the natural regime of a basin or a part of a subsurface system (ERDÉLYI; BÉRCZI and KÓKAI; SEMENOVA et al.) as well as those giving models for the prediction of the human influences on the ground-water reservoirs (CLOUET D'ORVAL and ARCHAMBAULT; LLAMAS and CRUCES DE ABIA) suppose the free flow of water in the direction of the pressure gradients. In two papers (BALLÓ, URBANCSEK) the attention is drawn to the fact, that the pressure difference does not mean the sufficient condition of flow. It indicates only the possibility of movement but there are cases where the higher pressure originates from the weight of the overlying layers, but the accumulated energy is not adequate to overcome the static resistance of the impervious layer. While BALLÓ accepts in some cases, that a part of the energy is consumed by the ground-water flow, URBANCSEK sharply excludes the existance of the vertical communication between the layers. Unfortunately he has failed to explain how his hypothesis might be testified in systems, where the energy content of the subsequent layers within a relatively pervious formation decreases gradually going downwards (as it is shown in Fig. 4. and 5. of URBANCSEK's paper). According to our present knowledge this condition (or using URBANCSEK's terminology a pressure gradient smaller than unity) indicates the vertical recharge of the deeper-lying aquifers.

I should like to ask Mr. URBANCSEK to try and explain this contradiction of his paper.

There is another point not clarified completely yet in connection with the application of the dynamic aspects in the hydrology of the ground-water: i.e. the contact between the shallow ground-water (the first water bearing horizon having direct contact through the soil-moisture with the meteorological phenomena above the surface) and the deep-lying aquifers. Although there was not any direct reference made to this problem in the papers submitted to the Symposium (at least in those the reviewing of which was my task) only DEÁK's paper has indicated slightly its importance, I should like to open a discussion also on this topic. This large international audience composed of well known scientists, is the best forum to determine the way to be followed in the solution of this problem.

The fundamental discrepancy between the two different opinions in connection with the contact of the shallow and deep ground-water lays in the supposed recharge of the subsurface reservoirs. Geologists dealing mostly with deeper aquifers have found smaller energy content in the upper lying layers than in the deeper ones in the central parts of the large sedimentary basins indicating the possibility of vertical flow directed upwards. They have supposed, therefore, that the mass-balance requires a continuous recharge along the edges of the basins (and through their internal pervious terrains having higher elevation than the average level of the basins) and the main draining effect of the flow-system is the evapotranspiration acting through the shallow ground-water. A further assumption (following from the previous one) is that the level of the shallow ground-water is the piezometric indication of the pressure conditions of the deep-lying aquifers. Hydrologists investigating the regime of the shallow ground-water have observed the decrease of the energy content going downwards from the water table. It was also found, that the average vertical positive accretion of the ground-water (infiltration reaching the water table) is higher than the negative value (withdrawn by capillarity from the ground-water to replenish the evaporated soil-moisture) in most cases. The shallow ground-water has to be drained, therefore, either at the bottom of the first aquifer (deep percolation) or by the rivers (base flow).

In my opinion the basic principles listed previously and especially

— the correct and complete determination of the geometry of the flow systems, as well as

— the combined consideration of the mass balanced of both the water and the transported dissolved materials (including that of the environmental isotopes) together with the energy balance of the system, provide us with sufficient theoretical background to characterize the natural conditions of the ground-water space and to forecast the expected anthropological changes. I think one of the main purposes of the Symposium is to promote the development of an internationally accepted concept for the application of hydrodynamic principles. The two topics mentioned as problems to be clarified (i.e. the interaction between the shallow and deep ground-waters as well as the role of the accumulated energy and the ground-water flow in the development of the pressure conditions) are good examples to show the fields to be investigated and the papers submitted to the Symposium contain many important aspects providing guide-lines to find the most realistic way of the solution of the arising problems. It is the reason why the papers listed in the annex of my report (including some papers dealt with by other general reporters but being pertinent also from the aspects discussed here) are reviewed considering their contribution to the solution of the basic problems quoted previously.

Classification of the various types of ground-water

It is a quite natural attempt applied many times in the various branches of sciences to establish a relatively simplified scheme, according to which the versatile phenomena can be classified by considering the most dominant effects in all groups. There are three papers dealing with this special problem (Bogomolov et al.; Kissin; Rogovskaya and Bezrodnov). Only slight differences occur in their classification systems. The most important common feature of these papers is the sharp distinction between the upper layers of the basins (where hydrodynamic processes govern the ground-water regime) and the lower part of the sedimentary formations (influenced by internal forces i.e. consolidation and thermal effects). Considering the two dominating factors Kissin even divides the lower part into two zones: elisian (dominating action is consolidation) and abyssal zones (being thermal dehydration is the most important effect). Bogomolov et al., as well as Rogovskaya and Bezrodnov make distinction in the hydrodynamically influenced zone between unconfined and confined systems divided by the upmost regionally extended impervious formation. The paper prepared by Iasvin et al. deals also separately with the layers having recharge from the neighbouring formations maintained by ground-water flow and with those the resource of which originates from consolidation. They also emphasize the differences occurring from the point of view of water exploitation between the central part of the basins and their edges.

The most important consequence of these classifications is that basically different approaches have to be applied for the investigation of the various regimes:

– shallow ground-water having direct contact with the hydrological and meteorological phenomena occurring on and above the surface;

– deep ground-water within the hydrodynamically influenced zone, where the movement of water influences the regime, but does not create direct contact with the surface only through the shallow ground-water or in the form of drainage directed to the river-beds from the pervious layers;

– closed aquifers through which flow does not develop under natural conditions and the pressures prevailing there are the results of consolidation;

– near the bottom of very deep basins the thermal effects have to be also considered.

The flow net presented by Tóth is applicable only for the investigation of the first two groups, because the interpretation of flow lines is acceptable only if movement really occurs. The construction of the potential lines is more suitable, because the potential difference indicates only the possibility of flow and not the existence of the movement. The closed aquifers have to be excluded, therefore, from the dynamically investigated flow field, and only the remaining layers can be divided into local intermediate and regional systems. A basic problem remains, however, still open, i.e. the determination of the limit between the hydrodynamic zone and that influenced mostly by consolidation. In the papers quoted previously only some slight indications can be found concerning the depth of this border (a depth of 1000 ~ 3000 m is mentioned by Iasvin et al. below which the dominant role of resources originating from consolidation is expected during exploitation, while the upper value of this depth is 300 ~ 500 m, its average is 1000 m but in extreme cases the border can be as deep as 6000 m according to KISSIN). A further difficulty is caused by the fact, that sharp distinction cannot be made between the two different regimes, since gradual transitions are generally characteristic for any changes in the Nature, and only the human brain tries to simplify the processes and to describe them by using generalized and deterministic relationships. The most important consequence of this gradual change is the fact, that the man's activity may basically modify the weight of the acting factors.

Two methods can be proposed to eliminate the obstacles hindering the application of the explained classification, i.e.:

— to determine the physical behaviour of the layers and to give numerical data for the characterization of the developing processes (giving more sensitive quantitative description of the processes instead of the sharp and rigid qualitative distinction);

— to use parameters other than the water-balance (chemical composition, environmental isotopes, heat balance, energy content) to prove the existance of the movement and to determine its numerical parameters mentioned in the previous point.

Physical parameters of the layers

There is only one parameter which can be observed directly in the layers i.e. the pressure, which gives the potential and the potential gradient in the system (after having combined it with the geodesic height of the measured point). The velocity or the flux of flow, the change in storage and the energy consumed for the maintenance of movement can be derived from the measured potential values, if some important soil-physical parameters of the layers are known (hydraulic conductivity, transmissibility, threshold gradient, storage coefficient, specific yield, etc.). Most of these parameters are theoretically well known, and there exist also methods to calculate them from data determined on samples taken out from the layers or to measure them either in laboratories (also on samples) or by applying field methods. Although it is well known, that many uncertainties hinder the determination of these parameters (e.g. the sampling methods disturb the structure of the layer, or an artificial condition is observed with pumping tests and not the natural one, thus the character of flow may be completely different during the measurements from that attempted to be described) the biggest problem is caused by the fact, that the various methods provide only point values, while the regional characterization of the physical behaviour of the layers would be necessary for the investigation of the large basins, as it is emphasized in several papers presented at the Symposium.

Considering the necessity to determine some average parameters of largely extending layers two important ways of further investigations can be indicated, i.e.:

— development of simple and inexpensive methods to measure these parameters; and

— application of statistical methods to calculate the mean and variance of data.

BESENECKER and LILLICH use geophysical data (geoelectric resistance and gamma radiation measured by well logging) to determine the hydraulic conductivity of the semi-permeable layers separating the aquifers. They emphasize, that the characterization of these layers is very important, because the flow develops generally perpendicularly to the interface of the aquifers and the separating layers, thus the cross-section of the flow is extremely large and, therefore, the water conveyance may be high even if the flux is relatively low. Hydraulic conductivity can be calculated from the data of coarse sediments (grain size distribution, porosity) with quite good accuracy, but the same relationships are not applicable for cohesive layers, which fact indicates the importance of the indirect measurements. It is also a great advantage of the method, that it provides informations from measurements generally executed in each bore-hole. The evaluation of the data collected in great number shows also a good example of the application of statistical analysis.

Sarin and Babic use also geophysical methods when introducing the transmissivity index, it being the product of geoelectric resistance and the corresponding thickness of the layer. They have found, that this index is proportional to the transmissibility of the layers, the relative water conveying capacity of the various sections of the basin can be compared on the basis of surface measurement. The application of the method is represented by the example of the Drava valley. One question arises, however, when the published data are analysed, i.e. what is the distorting effect caused by the fact, that the depth of the section is limited at 150 m where the basement lies deeper than this level. It would have been better, perhaps, if the lower limit had been taken into account by considering the actual geological conditions.

Juodkazis and Paltanavicius investigated also the hydraulic conductivity of the semi-pervious layers situated between the aquifers. They have found the depth of the active water exchange to be 100 ~ 300 m in the Baltic artesian basin. Some data are published characterizing the regional averages of the hydraulic conductivity values of the various layers. These parameters are determined by using water and heat balance equations as well as by considering the local head losses, thus the paper provides a good example also for the combined application of the various methods. Laboratory measurements were also executed, the result of which is represented in Fig. 3 of their paper indicating a threshold gradient of about 0.1 or less.

After mentioning the threshold gradient I have felt necessary to analyse the behaviour of this physical parameter not only because in some papers (Balló; Mrs. and Mr. Lorberer; Kesserű) reference is made to my own research, but also because its numerical value compared to the actual potential gradient makes it possible to determine the position of the border between the hydrodynamic zone and that influenced only by consolidation. Recent investigations (Bondarenko, 1973; Kovács, 1976) have shown, that the threshold gradient decreases with increasing temperature and its value becomes negligible at about 60 °C. Although this relationship is valid only under atmospheric pressure, the decreasing character of the parameter indicates, that the influence of the threshold value may be smaller in greater depth then near the surface. This condition may be compensated by the effect of the higher pressure, but the determination of this second influence requires still further investigations. The other important result of the recent research (Kovács, 1976) is, that the pore size is a random variable, and the probability of a pore having twice larger diameter than the average pore is still 1 to 4 percentage. Thus the development of flow is still probable under a gradient being only the half of the theoretical threshold value. The interpretation of the threshold gradient as a random variable is also well demonstrated in KESSERŰ's paper, in which the parameter in question is derived from data measured in mines.

Among the reviewed papers that written by BÉRCZI and KÓKAI gives also numerical informations on the flow developing in a regional system of layers, while ORHAN DUMLU deals with the evaluation of pumping tests creating non steady flow.

Combined application of the investigations of various parameters

It is emphasized in JUODKAZIS and PALTANAVICIUS' paper already referred, that there are many uncertainties influencing the methods used for the evaluation of the various balance equations. More reliable results can be achieved, however, if the different balances are compared to each other. Thus BESBES and DE MARSILY use the piezometric data, the salt content, the water balance and the investigation of the storage process for the characterization of a sedimentary basin having an extension of 3000 sq.km. ERDÉLYI combines also the pressure values and some parameters of the chemical composition of the water to analyse the ground-water regime of the Hungarian Basin. The pressure and chemical data are also suitable, to indicate local interactions between layers as it is done in PALAUSI and POLVECHE's paper. There are examples to demonstrate the use of thermal-balance equations to characterize numerically the ground-water flow in the papers written by ALFÖLDI et al. as well as by WERNER and BALKE.

EINSELE's paper is an important contribution to the evaluation of the hydrochemical data for the characterization of the flow systems. He proves both theoretically and by using observed data, that the water squeezed out of the lower lying layer by consolidation caused by the weight of the upper layers can reach only a limited range, thus the observation of a water of mixed character in a greater distance from the interface indicates a high probability of the development of flow. EINSELE has applied a very simplified model for the derivation of the theoretical results, i.e. the hypothesis of a linear relationship between the porosity of the layer and the depth of the investigated point. Two different assumptions are used concerning the lower limit of the linear relationship, supposing at first that porosity may decrease till zero, while in the second model a finite lower limit was applied. Although it is evident, that the simplified models give higher range of penetration, than the practically expected value (and, therefore, the calculated parameters include some safety as well) the theoretical investigation ought to be repeated by applying more realistic approximations of the porosity v.s. depth relationship.

The investigation of the rate of environmental isotopes provides us with new very effective means for the determination of the flow conditions, which indicates not only the mixing of the various types of water (naturally considering also the penetration of the water squeezed out by consolidation) but it can be used for aging the water and to give numerical values for the determination of the average velocity of propagation. A further advantage of the method is the fact, that the determination of the rate of the various isotopes (deuterium, tritium, O^{18}, C^{13}, C^{14}) gives a possibility of the comparison of the different results and thus the checking is ensured at the same time.

DEÁK's paper is a good example of the application of isotope-investigation in ground-water hydrology. It was found in the investigated area (Nagykunság) that the tritium content of the shallow ground-water is proportional to the effective grain diameter of the covering layer, but the TU parameter was always significantly higher, than the background value, while this difference was not observable in deeper layers. The indicator of the rate of C^{13} was found to be constant (around -20) below a depth of 70 m, while at the surface this parameter was between -8 and -12. The aging (using C^{14} and considering C^{13} as well) has shown a water being about 10.400 years old at a depth of 400 m-s and 14.500 years at 80 m-s. Recent water was indicated at the surface, while the age of waters taken from the depth-interval of $30-50$ m was between 4000 and 7000 years. The index of deuterium concentration was the highest at the surface (-60), had a minimum (-90) at a depth between 50 and 100 m-s and increased gradually with the depth reaching about -70 at 1000 m-s. The data indicate the probability of a flow system through the deep ground-water directed upwards with a flux of 80 mm/year and drained horizontally through a coarse-grained sandy layer at the depth between 50 and 70 m-s, while above this layer the mixture of the recent shallow ground-water and the deep one can be found.

Investigation of the complete flow system

DEÁK's data draw the attention to the importance of the investigation of the compete flow system from the recharging area through the transporting stretch until the draining section, as it was already mentioned in the first part of this report. And really the lack of this complete analysis hinders the evaluation of the collected data in many cases. Thus the very careful collection and representation of the data presented in Mrs. and Mr. LORBERER's paper gives a good example of an up-to-date hydrogeological investigation considering both pressure conditions and chemical parameters. It still does not give a complete evaluation of the situation, because the place and the rate of the draining effects are not known and the development of the local sinks in the pattern of the potential lines cannot be understood without these informations.

The place of the draining point is basically important from the point of view of the investigation of the interaction between deep and shallow groundwaters. The flow system described in connection with DEÁK's investigations can be explained only by supposing the river beds penetrating into the sandy layer extending below the first regional covering layer to be the main draining sections of both the deep and shallow ground-waters. It is the reason why the potential gradient is directed downwards above the sandy layer while in opposite direction in the deep ground-water. The existance of a slight relationship between the character of the autocorrelation of precipitation and that of the regime of the shallow ground-water demonstrated by RÉTHÁTI's paper seems to testify, that the main recharging effect of the upmost aquifer is the precipitation. The potential values represented in the sections constructed by ERDÉLYI justifies also the probability of the explained flow system, demonstrating, that the minimum value of the potential occurs generally in the first confined sandy layer having a direct contact with the rivers.

I have already mentioned, that the generalization of a process observed at a given area may lead to the misinterpretation of the hydrogeological systems developing in other regions. I did not quote the example as a generally acceptable model for the characterization of the interactions between the shallow and the deep ground-waters, but to demonstrate the fact that the correct exploration of the entire flow systems may give the correct interpretation of the observed data and may eliminate the appearant contradictions between the observations. I am quite sure, that apart from the explained type of contact there exists several other ways of interactions as well (the upmost water horizon may recharge the deeper ones, flow can develop from the greater depth directed to the shallow ground-water and drained by evaporation, or there may be impervious boundary between the two systems if the actual gradient is smaller than the threshold value). The character of the interactions has to be determined always by the careful consideration of the local conditions and – because these conditions may change within relatively small distances – only the quantitative description of the processes can be sensitive enough to clarify the actual development of the flow systems and their boundaries.

Without trying to generalize the explained separation of the shallow and deep systems, it is worth-while to note, that the development of similar conditions may be supposed in other regions as well on the basis of some informations presented in papers submitted at the Symposium. Fig. 3. of SEMENOVA et al. paper represents the potential conditions in a section of the Terek-Kuma Basin. The lowest potential is indicated below the first regional semi-pervious layer which is possible only if the sand below this cover is drained by the rivers. PALAUSI and POLVECHE describe a basin, where the shallow ground-water has direct recharge from precipitation of 800 mm/year. The pressure of the second water horizon is higher than that of the first one, but the deeper layer is also drained by the sea. It would be interesting to investigate under natural conditions the development of vertical flow through the semi-pervious layer separating the two water horizons, where such interaction was artificially produced by perforating this layer. The determination of the rate of the two different effects draining the deeper layers (i.e. evaporation through the shallow ground-water and percolation to the sea) and the areal variation of this ratio depending on the local conditions.

Prediction of the effects of human activities

The centre of our interest is always the human being, the final objective of all sciences has to ensure, therefore, the better living conditions of mankind. Thus the hydrological investigation of the ground-water always aims the determination and the best utilization of the available water resources. On the basis of the exploration of the natural phenomena influencing the regime of ground-water the expected environmental changes caused by the exploitation of the water has to be predicted finally. This is the reason, why several papers deal with the models suitable to forecast of the probable influences of ground-water pumping (CLOUET D'ORVAL and ARCHAMBAULT; LLAMAS and CRUCES DE ABIA; SZÉKELY) and evaluate the observed changes caused by human activities (BESBES and DE MARSILY; CLOUET D'ORVAL and ARCHAMBAULT).

The hydrodynamic equations describing the flow initiated by pumping are very complicated if the correct boundary conditions, the layered flow space, the various recharging effects and the role of consolidation have to be considered in a regional investigation, as it is shown in SZÉKELY's paper. Some geometrical simplification has to be applied in any cases to simulate the irregular structure of the water transporting layers (e.g. LLAMAS and CRUCES DE ABIA investigate a system composed of three aquifers, while in the example demonstrated by SZÉKELY two aquifers are situated above one another separated by a semi-pervious layer). The recently developed numerical techniques facilitate the solution of the problems with acceptable accuracy by using computers (e.g. the combination of the forward and backward approximations makes possible to raise the number of the knots within the grid as it is described by SZÉKELY; LLAMAS and CRUCES DE ABIA do not present the applied technique,but it is probable, that the same method was used, considering the 1125 knots). The most difficult problem in connection with the application of the numerical methods is the determination of the physical parameters characterizing the flow and the boundary conditions. The results achieved by using the computers are excellent and very important from hydrogeological point of view as it is demonstrated by the papers applying these mathematical models, but their reliability depends basically on the correctness of the parameters. The further research ought to be concentrated, therefore, to the better determination of such characteristics as hydraulic conductivity, transmissibility, storage coefficient, vertical accretion, etc.

The better understanding of the physical process is the purpose of SCHMIEDER's paper as well. He investigates the extension of the influenced field around a pumped well. The practical observations have shown, that the radius of the influenced zone is not exactly proportional to the square root of time, as it is supposed by the theoretical derivations, but a time-dependent factor of proportionality has to be also considered. He has found, that this factor reaches a maximum after increasing with time, and it starts to decrease later on. It would be interesting to investigate, whether the slowering of the propagation is caused by the non laminar character of flow around the edges of the influenced zone (by the influence of threshold gradient) where the gradient is very small after the considerable extension of the draw-down cone.

KESSERŰ investigates an other important consequence of draw-down, i.e. the land subsidence, and the demages of buildings caused by the movement of the land surface. The analogy between the effect of pumping and that of underground mining is used to determine the numerical values of the expected subsidence.

* * *

As the final consequence drawn up from the reviewed papers the following main principles can be repeated here at the end of the report as the most important aspects of the hydrological and hydrodynamical investigation of the ground-water:

1. The balances of water, dissolved chemical, environmental isotopes and energy should be used always in a combined form.

2. The geological system should be investigated always dynamically considering the influence of the developing flow and the accumulation of energy.

3. The flow systems should be separated from each other as well as from the zones where the accumulated energy does not initiate movement, and always the total characterization of a complete flow system should be the goal of the investigation.

4. The processes governing the regime at a given area and explored by the investigation must not be generalized for the whole basin or for other regions, but the local conditions have to be determined individually within each part of the basin.

5. For the simplified characterization of the various regimes the relative importance of the different acting processes have to be determined and, therefore, the numerical description of the phenomena is inevitable.

6. As a consequence of the previous point the determination of the physical properties of the various layers is one of the most important purposes of the hydrogeological investigations. The knowledge of the numerical values of the soil-physical parameters (together with their probable variances within large regions) is a precondition of the application of the up-to-date numerical methods as well.

GENERAL REPORT 2

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Mister Chairman, Ladies and Gentlemen,

I have the pleasure to introduce to you 21 papers according to our main topic: "Subsurface water movements, filtration, hydrodynamics, models." I read all these papers with great pleasure and can declare to you, that each one is very interesting. It is of course not possible for me, to spend more than a few words on each paper, and neither the number of words—which are sometimes many sometimes few—nor the arrangement of the papers mean any valuation. I have tried to bring the papers into a logical and thematical succession and that means of course sometimes a mere reductional order.

Great sedimentary basins are of primary importance for the satisfaction of the water supply of man. This requires, however, the knowledge of their regional hydrogeological situation, of the subsurface water resources themselves and of their dynamics. To solve the appearing problems it is necessary, besides and together with special regional studies, to find out general characteristics and methods.

Mr. Bogomolov, Mr. Kudelsky and Mr. Lapshin are presenting in their paper a scheme of vertical hydrogeologic zoning of large sedimentary basins, based on long-years exploration work of a great number of scientists. They distinguish:

1. an "upper geohydrodynamic system" which is pressure-free, either under local hydrostatic pressure and has free and impeded subterranean discharge, mainly infiltrative with elements of the so-called elisional discharge, from

2. a middle geohydrodynamic system with water pressured; free, impeded and greatly hampered subterranean discharge; mainly elisional with infiltrative discharge, conditioned by desorbtion of bounded waters and influx of retrograde solutions and

3. a lower geohydrodynamic system, which is pressure-free either under local partial pressure and has greatly hampered subterranean discharge of mainly overheated waters and retrograde solutions with elements of elisional discharge.

Mr. KISSIN describes also three types of hydrodynamic regime in his paper *"The principal distinctive features of the hydrodynamic regime of intensive Earth crust downwarping areas"*:

1. the regime of the infiltration type with the main subtypes *(a)* unconfined ground-water and *(b)* artesian (confined) water

2. the regime of the elisian type with water release on *(a)* gravitational compaction of unconsolidated clays or *(b)* compaction by tectonic stresses and

3. the regime of the abyssal type with *(a)* epigenetic and *(b)* metamorphogenic subtypes.

He pointed out, that the occurrence of regimes of various types depends on a certain zonation that in turn depends on the structure and history of the geological development of the region. This is exemplified by the Eastern Predkavkazye basin, where in some parts for instance very high ground-water heads are observed, which are assumed to have been caused by compaction and dehydration of clays.

Mrs. ROGOVSKAYA and Mr. BEZRODNOV give in their study *"Regional hydrodynamic regularities of platform-type artesian basins"* a hydrophysical model. They found out, that the age of rocks composing the sedimentary formation defines the maturity of the post-sedimentary transformation processes of deposits into older platforms and rather active recent development of these processes in young platforms. Within the Mesocenozoic platforms hydrodynamic pressures reach 2500–3200 m and hydraulic gradients are 10^{-3} to 10^{-2}. Within the older platforms hydrodynamic pressures, as a rule, do not exceed 450–550 m, hydraulic gradients vary from 10^{-5} to 10^{-2}. In young artesian basins the main energetic source is a geostatic pressure, in older platforms a hydrostatic pressure was re-generated by differential tectonic movements.

Hydrogeology of great sedimentary basins means in practice at first *"a study of the relationship between ground-water resources and subsurface geological structures"*, which is carried out for example "in the northwestern part of the Great Hungarian Plain" by Mrs. and Mr. LORBERER. They found out, that within the Pliocene/Upper Pannonian aquifers a deep water flow between the buried Mesozoic basement of Bugyi–Űrbőpuszta and the tectonic graben of Délegyháza–Alsónémedi can be identified. The rate of descending water near the Mesozoic block of Bugyi–Űrbőpuszta can be estimated according to the DUPUIT–THIEM equation to be 1452 litres/minute whereas after the equation of local heat-flow calculation it is of 1209 l/m. The natural flow systems of the subsurface waters and their connection with local flow can be determined by the comprehensive study of the potential distribution of deep waters as well as that of the hydrochemical and geothermal conditions. The flow conditions of deep waters are highly affected by the tectonic setting. A recharge through leakage deriving from the thermal water-bearing karstic reservoirs of Budapest is supposed to be responsible for the increased surplus of heat, gas content, Cl-concentration and hardness of the deep waters in the region of Délegyháza–Alsónémedi.

Also situated within the Great Hungarian Plain is the Szeged Basin in which Mr. BÉRCZI and Mr. KÓKAI carried out *"hydrogeological observations in* *hydrocarbon exploring boreholes"*. The Szeged Basin's basal aquifer shows pressure differences between the southern and eastern parts and other parts of the basin. The highly permeable reservoir rocks (first of all the Middle Triassic fractured, brecciated dolomite, Miocene and Lower Pannonian conglomerate series) are favourable for the flowing system. The rates of flow estimated from tilted hydrocarbon-water contacts and pressure differences are 0.12 and 0.15 m/year in Szeged and in Algyő, respectively. The direction of flows is highly influenced by an area of poor hydrodynamic conductivity in the western flank of the Algyő and in Dorozsma, in which stagnant water conditions were revealed by hydrogeochemical analysis.

The hydrodynamically orientated evaluation of chemical analyses of ground-water samples can be an important qualitative approach in order to identify ground-water-flow systems in large sedimentary basins. This is demonstrated by Mr. ERDÉLYI in his paper: "*Chemical aspects of ground-water flow-regions of the Hungarian basin*". On principle, three water horizons of different chemical character can be distinguished in the Hungarian basin:

1. a bicarbonate type, linked with the surface-recharge areas of the ground-water flow-regions

2. a sodium hydrocarbonate type, in a zone of dynamic chemical equilibrium — the transition zone — which formed between the downward moving colder bicarbonate water and the ascending warmer salt water, and

3. a sodium chloride type at greater depth.

Knowledge of the spatial distribution (depth, shape, thickness) of these types is helpful to verify and explain hydrodynamic conditions especially when investigating irregular flow-patterns. This is shown in several examples.

Spatial variations of geohydrostatic pressure and their reason were studied by Mr. URBANCSEK and described in his paper "*Formation water pressure in the Quaternary sediments of the Great Hungarian Plain*". He pointed out that in the study area the geohydrostatic pressure will increase with depth and the rate of increase is in causal relationship with the grain size composition, and with the consolidation of the grain set concerned. The rate of increase will vary, even in the same profile, according to the proportion of the coarser and finer grain fractions, being higher in the finer grain fractions and lower in the coarser ones. The fact that the different pressures were not balanced in the course of geological time is indicative of the presence of a static strain. The diverse pressure states observed indicate that the water contained in the Quaternary sediments of the Great Hungarian Plain is, under normal conditions, in a static state and that this static state is maintained by the boundary stress corresponding to the permeability of the strata.

Mr. BALLÓ gives a contribution to "*The stress-theory of water-bearing layers based upon the experiences in the Great Plain of the Hungarian share of the Carpathian Basin*". He gives regional examples, discusses the problems concerned with stress and stress-theory and shows a simple model of the geohydrostatical stress space. Finally he points out that human interference by drilling wells into this delicate geological balance of the stressed volumetric continuum means a drastic change in the surroundings of the filter of the well. Concerning underground water use the decrease of stress following the exploitation seems to be more important than water extraction itself. The descent of heads of artesian wells is due to the decrease of geohydrostatic surplus and not to the significant diminution of underground water resources.

In his study "*Rise of primary pore water (connate water) in compacting sediments*" Mr. EINSELE found out by graphical methods, that under normal conditions the vertically ascending pore water flow does not reach the sedimentary surface. Besides the range of upward movement also the velocity of the ascending pore water can be calculated. In the course of time, however, the influence of molecular diffusion and hydrodynamical dispersion may alter the situation.

In his paper "*Formation du facteur de la conductibilité de pression dans le temps dans les aquifères fissurés et stratifiés à structure poreuse*" Mr. SCHMIEDER presents a theoretical and mathematical study based on a wide range of experience gained in mining exploitation. He points out the temporal "factor of pressure-conductibility", which depends on the factors of ground-water inventary.

As a basis for model calculations of available ground-water resources for example it is necessary to know parameters in addition to lithologic logs from boreholes, such as the coefficient of transmissivity and of permeability.

In his paper: "*Determination of uniform and granular aquifer characteristics by linear methods*" Mr. DUMLU shows a simple graphical method to determine the transmissivity coefficient from pumping test data.

Also concerned with transmissivity is the paper of Mr. SARIN, Mr. BABIC and Mr. CAKARUN: "*Transmissivity index along the Drava River valley in Croatia*". Transmissivity index is a new parameter developed to make resistivity surveying better applicable in hydrogeology. As is well known the resistivity sounding results in the determination of the thickness h and specific electrical resistivity of beds laying beneath a surveying station. It is assumed that there is an approximate exponential relation between permeability K and resistivity ϱ for the following sequence of water-saturated grain-size classes: clayey silt – silt – fine sand – medium to coarse sand – pure gravelly sand or gravel. The product $\varrho \cdot h$ receives a property ascribed to the product $k \cdot h$, that is to say to transmissivity and therefore is termed transmissivity index τ and its unit is ohm·m². For the calculation of τ it is necessary that a sufficient number of lithologic logs from deep boreholes is available during the interpretation of measured resistivity surveying data. Results obtained by use of this method are shown on the example of the Drava River valley.

The problem of ground-water movement through slightly permeable, semi-permeable or nearly impermeable layers by seepage is the main subject of the following papers.

Mr. KESSERŰ shows in his paper: "*The communication between unconsolidated subsurface water reservoirs in the light of mining experiences*". In the mining terminology the generally isolating layers between a mine cavity and a water-bearing layer are termed protective layers. Their protective effect is studied by theoretical examination of their capability of loadbearing, of preventing the starting of a seepage and of limiting the rate of the starting seepage and the extent of the limitation. This new approximation of the problem holds out promises of new results.

Common in large sedimentary basins are several superposed aquifers which are separated by semi-permeable layers. As a basis for model calculations, such as for the calculation of the rate of vertical movement of ground water into deeper aquifers, it is necessary to know the order of magnitude of the vertical permeability and thereby of the leakage coefficient of the semi-permeable beds separating the aquifers. Information about these values is of considerable practical relevance in estimating the regeneration of the ground-water in a deeper aquifer from which water is to be produced and on the other hand in estimating the risk of pollution.

Mr. JUODKAZIS and Mr. PALTANAVICIUS studied in their paper the "*Filtration properties of slightly permeable layers of the Baltic artesian basin...*" In the separating layers: morainic loam and sandy loam of the Würmian, Riss and Mindelian glaciation, a complex of Jurassic and Triassic clays and argillaceous Cambrian deposits. On the basis of balance calculations it is possible to determine the filtration coefficient of slightly permeable rocks and the intensity and dynamics of aquifer recharge as well. The results of geothermal measurements give the possibility to determine the rate of vertical leakage of underground waters and filtration properties of slightly permeable layers. On the base of the investigations it is possible to conclude that:

1. the values of the filtration coefficient of coeval semi-permeable layers in the same areas differ slightly,

2. the recharge modulus of head aquifers numerically decrease with depth,

3. the vertical leakage and filtration properties of coeval slightly permeable layers increase from watershed areas to river valleys.

Laboratory tests carried out by the authors using formation samples of undisturbed structure have shown that increasing a load per sample the filtration coefficient decreases according to hyperbolic dependence.

In my own paper, prepared together with Mr. LILLICH, a new method for "*the determination from non-hydrological data of permeability rates of aquifer-separating layers*" is presented. From lithologic and geophysical (resistivity and gamma-ray) logs of boreholes numerical parameters have been determined for the semi-permeable strata separating two superposed aquifers consisting of unconsolidated Tertiary and Quaternary sediments in northwest Germany. To allow comparison, the parameters K_R (from the resistivity log) and K_G (from the gamma-ray) were calculated for each borehole by relating the logged values for the aquifer-separating layer to the logged values for "pure" clay and gravelly coarse-grained sand.

A statistical analysis of the values of K_R and K_G determined for about 300 boreholes showed the anticipated dependence on grain-size trend. Consequently, it appears valid to correlate these parameters with permeability coefficients k_f. Assuming that fine-grained sand has a k_f-value of 10^{-10} m/s (about $9 \cdot 10^{-6}$ m/day), a simple linear correlation of the intervals in between indicates a k_f-value of about $9 \cdot 10^{-10}$ m/s (about $8 \cdot 10^{-5}$ m/day) for silty clay, $1.5 \cdot 10^{-8}$ m/s (about 10^{-3} m/day) for clayey silt, and $1 \cdot 10^{-6}$ m/s (about 10^{-1} m/day) for sandy silt.

Hydrogeological parameters such as coefficients of transmissivity and permeability are needed for model calculations like those of Mr. LLAMAS and Mr. CRUCES DE ABIA called: "*Conceptual and digital models of the ground-water flow in the Tertiary basin of the Tagus river (Spain)*". The authors explain their conceptual models as geometry and hydrogeological parameters of the aquifer system as well as the system of recharge, flow and discharge of the ground-water and the hydrochemistry. The digital model used is very similar to the so-called "three dimensional" one of Prickett and Lonnquist (1971). The prototype is represented in the model by three super-imposed aquifers. The first aquifer is a water table one. The second aquifer is confined and is separated from the first one by an aquitard. The third aquifer is also separated by an aquitard and is limited at the bottom by an aquiclude. The grid that was taken is quadrangular and uniform (twice two km). The calibration process required seven premodels to be carried out. Some principal characteristics of the last one were:

1. Transmissivity of the three aquifers is equal to 50, 100 and 50 m^2/day.

2. Leakage coefficient (calculated as permeability coefficient of a slightly permeable layer divided by its thickness) is the same in the two aquitards and equal to $0.6 \cdot 10^{-5}$ (l/day).

3. Recharge is equal to 50 mm/year in the whole area excluding the urbanized zones, there a higher rate, produced by leaks in the distribution and sewer network, equal to 200 mm/year, is estimated.

A first withdrawal simulation was carried out and has shown, that the project of concentrated pumping should be considered with a good deal of reservation.

In their study: "*L'analyse d'un grand réservoir aquifère en vue de sa modèlisation*" Mr. Besbes and Mr. Marsily give an example how to analyse a large ground-water reservoir with regard to its model calculation. In the study area, the plain of Kairouan in Central Tunesia, there are two reservoirs. Data about the geometry, structure and hydrogeological conditions of the basin were obtained by geophysical surveying, drilling and ground-water heads. Hydrodynamical and hydrochemical investigations together with pressure and recharge calculations were carried out, and hydrogeological parameters such as transmissivity coefficient were evaluated. These results are presented in an organization-programme.

In their study: "*La nappe des sables verts et l'alimentation en eau de la région parisienne*" Mr. Clouet d'Orval and Mr. Archambault present a model calculation of the Cretaceous greensand-aquifer in the region of Paris. After the calibration of the model the possibility of ground-water exploitation in the region of Paris was tested. The simulations have shown that a withdrawal will be possible in some areas.

The last years have seen considerable developments in geostatistics and other applications of the Theory of Regionalized Variables with successfull applications also in hydrogeology. The theory has been developed, its fundamental hypotheses have been widened allowing the tackling of more and more sophisticated problems and original new techniques have been designed and successfully applied, mainly conditional simulations and Universial Kriging. The applications of this technique in hydrogeology are shown by Mr. de Wrachien and Mr. Gruppo in their paper: "*Le Krigeage Universel et ses applications à l'hydrogéologie*".

Mr. Székely in his paper presents a "*mathematical model for the cone of depression of waterworks in loose sedimentary basins*". Within the two layered aquifers of water-table condition as well as within the three-parted confined water-bearing formations of the Quaternary of the Great Hungarian Plain the distribution of the regional drawdown influenced by the water production can be characterized by the heterogeneous "leaky aquifers" model. For the numerical analysis of the nonsteady-state processes a computer program based on the application of differential method was elaborated.

In the surrounding of the town Kalocsa a regional drawdown of the waterworks which have a capacity of $Q = 10{,}000$ m^3/day was determined by computer method where a two-layered aquifer of water-table conditon accompanied by given flow parameters occurs. The vertical leakage of the discharging formation derives here from the drainage network having no complete beds. In the surrounding of the city Debrecen hydrogeological parameters were determined from known water production and drawdown data in the case of a three-parted confined aquifer. The discharging Lower Pleistocene sand formations get here a vertical leakage deriving from the phreatic water bodies. The problem was solved by simulation through the minimization of the difference between the measured and computed drawdowns.

Ground-water exploitation often leads to a usually local but considerable drawdown. A result of this is an alteration of the balance state between water and reservoir rocks which may cause damages at the surface. Mr. KESSERŰ studied this phenomenon in his paper: "*Surface damages due to differences in the pressure of interstice water for deconsolidated sediments*". He carried out a theoretical and practical, quantitative study of displacements and points out the pre-conditions of damages, based on practical experience and measuring data. His results are shown in a table and in several graphs.

This was the last paper dealing with our topic, which last but not least also proves that hydrogeology is calling for far-sighted men.

Thank you very much for your attention!

THE STRESS-THEORY OF WATER-BEARING LAYERS BASED UPON THE EXPERIENCES IN THE GREAT PLAIN OF THE HUNGARIAN SHARE OF THE CARPATHIAN BASIN

LA THÉORIE DE L'ÉTAT DE TENSION DES NAPPES AQUIFÈRES; SELON LES EXPERIENCES REÇUES DANS LA PARTIE HONGROISE DE LA GRANDE PLAINE DU BASSIN CARPATHIQUE

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RÉSUMÉ

L'étude donne la description sommaire des phénomènes du régime hydrostatique particulièrement modifié dans l'espace gravitaire.

La Grande Plaine Hongroise, notamment sa partie hongroise est un modèle naturel parfait — et peut être unique — du terrain de recherche des équilibres fins des nappes aquifères répandues.

L'hypothèse établie en 1960 est vérifiée par les preuves du comportement d'environ 40 milles puits forés dans des profondeurs de 100 à 500 m.

L'étude présente l'hypothèse dans quelques cas démontrés selon le Cadastre des Eaux Souterraines compilé par le Centre de la Gestion des Ressources d'Eau de l'Office National des Eaux hongrois.

Les observations sont déjà suffisantes à ce que l'étude puisse déterminer en 9 thèses la théorie de la géohydrostatique en s'appuyant aux nouveaux points de vues.

On peut supposer, que dans le monde entier il y a beaucoup de bassins de sédimentation pareils, où les expériences acquises à la Grande Plaine Hongroise peuvent être utiles aussi lors de la planification perspectif de l'aménagement d'eau des autres pays.

Nous espérons que les exemples donnés contribueront aux discussions qui suivent, ou encore, qu'ils feront rapprocher les opinions aujourd'hui encore tellement différentes.

Introduction

The ever-increasing importance of requirement for water becomes more and more of common knowledge. The scarcity of pure drinkwater in many corners of the World casts the spotlight to the ground-waters. Ground-waters are scilicet safeguarded against pollution, contamination etc. by Nature itself. With special reference to the serious problems involved, it must be considered as one of the most fundamental basic resources for survival of life in the World.

Nowadays, we are mining these ground-waters without a proper understanding of the phenomena ruling the physical state of the water deposits. Nonetheless, we have not a clear concept of the process incited by drastic human interference due to releasing the stress by drilling wells into the water bearing layers.

The Hungarian share of the Great Plain is approximately 41 thousand km^2. The volume of the continuum of the same in cu.km might contain a porosity-volume of 8.2 thousand cu.km full of water. In these days the number of driven wells in the area treated is about 40 thousand and water exploited cannot be more than $17.7 \cdot 10^9$ cu.m supposing an exponential increase from 1900 as starting point until 1975 (the last statistical data available is $330 \cdot 10^6$ cu.m/year).

The same volume expressed in water stratum and in water-bearing layer might be equal to 43 cm and 215 cm, respectively. Pro rata this is not more than 0.00215 portion of the underground water resources estimated. Metaphorically speaking we exhausted only a drop from the ocean situated underneath.

The pre-estimation of the water volume exploited is valuable only for the verification of the fact; we cannot deduce far-reaching conclusions on the basis of the quasi steady discharges of the wells without a deeper knowledge of physical phenomena taking place in the heterogeneous and anisotrop continuum of water-bearing layers.

According to the common knowledge the artesian phenomenon is a simple case of intercommunicating tubes. If this is true;

a) There must exist a permeable structure of layers reaching from the recharging site to the artesian well.

b) It is necessary the existence of a non-permeable and continuous layer-cup which is able to maintain the stress produced in the natural reservoir of the hill.

c) It is indispensable a head equal to the resistance of the permeable layers and the tube, as well.

The aggregate of the preconditions mentioned afore seems to be an exceptional situation.

Contrarily a new hypothese of stress-theory of underground waters has been set up identified as the theory of geohydrostatics. First publication of the theory expressed also in this paper has been performed in 1960 as a lecture of I. Z. Balló C. E. in Szeged (Hungary) at a session of the Local Group of the Hungarian Hydrological Society. Its first international publication was at the "Conference on Hydraulics", Budapest (Hungary) 1960. Broader explanation of the idea has been published in the "Vízügyi Közlemények", Budapest 1961 N° 4.

Phenomena treated in this paper comprises a description of a specially modified hydrostatical realm which is included into the gravity field of force.

Theses of the theory of geohydrostatics

1. Waters in deeper water-bearing layers excluding the uppermost subsurface waters are of paleogeologic origin.

2. The stress of underground waters shows a special shape which is different from the hydrostatical stress that has been manifested in a general surplus and is increasing vertically with depth.

3. Spatial anomalies in different directions are depending upon the individual structure of water-bearing layers and on the structure of sedimentations.

4. Taking into account the effect of retention of layers working as seasonal reservoirs dynamic subsurface water resources are extremities in the case of karstic water and/or the highest subsoil water and the bank-filtered waters if their annual oscillation follows the normal water budget of the site. In consequence, these water resources might be handled as unseparable shares of surface waters.

5. Deviations—further called as geohydrostatical differences—are depending upon

- the molecular attractive forces on the surface of granuli as a function of size-structure porosity and plasticity,
- the morphology of the surface and of the layers,
- the gravity field of force,
- and the epeirogenic micro-movements of the crest.

Factors enlisted are coming into prominence varying according to the individual features of the position.

6. Gradient vector either positive or negative is only necessary but not adequate precondition of water transfer. The adequacy is connected with the surpass of the geohydrostatical stress limit value.

7. Gradient vector tending toward the surface or the abyss displays always the role of molecular attractive forces the normal consolidation process and/or local epeirogenic rising of the ground in the presence of convenient morphology of layers.

8. Negative gradient vector—tending vertically upstairs—is a normal feature of upper semi-permeable layers.

9. The same negative vector under a non-permeable layer is a proof of epeirogenic rising of layers due to the rearrangement of geohydrostatical stress space inside of the water-bearing layers.

Case of the stress situation in the Hungarian share of the Great Plain

In the course of the Upper Pannonian and the Quaternary the Carpathian Basin underwent a rhythmic deposition of sediments of aeolian and alluvial origin controlled by climatic oscillations. The Hungarian share of the Great Plain has a considerable thickness of fluvian and aeolian deposits too. Many strata of these layers are water-bearing rendering an enormous resource of ground-waters suitable for management purposes. The origin and behaviour of the ground-water were always the topic of hot-tempered polemics among geographers, geologists and civil engineers.

The Hungarian Great Plain i.e. the Hungarian share of the lowland is an excellent—perhaps unique—natural pattern for the investigation of a delicate stress balance inside of the space of vast water-bearing layers.

Evidences for corroboration of the hypothesis set up in 1960 are given by the behaviour of about 40 thousand 100—500 meter deep driven wells.

The purpose of this paper is to illustrate the theses with the aid of some situations drawn up by the Ground-water Thesaurus of the Hungarian National Water Authority—Water Management Centre.

The theses are demonstrated by some examples taken from the case of the Hungarian Great Plain. See sketch map as Fig. 1.

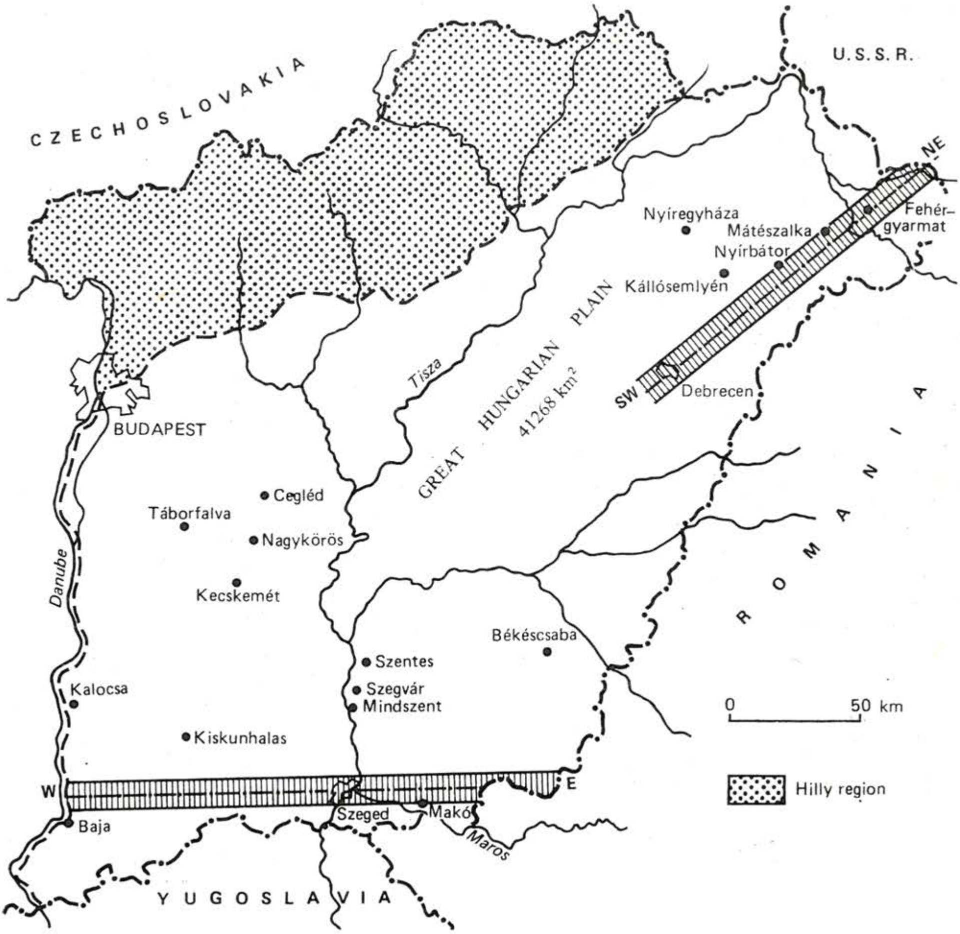


Fig. 1. Sketch map of the Great Plain

The classical artesian theory is disaffirmed by the facts:

a) There are some sites showing up all the necessary and adequate conditions of the set-up without any artesian phenomena, and

b) vice versa, artesian phenomena can be found in a wide range not fitting into the necessary and adequate conditions required by the classical theory.

A profile of the Great Plain is presented in Fig. 2, along the cross-section of the towns Baja—Szeged—Makó of 140 km length and the piezometric heads of a few driven wells. The density of the wells along the straight line is not adequate for drawing up the heads related to the same geodetic superficies of the underside of the wells, yet the overall shape is considerable. The well "α" particularily emphasizes the horizontal stress parameter of the underside of the hole equal to 4.6 kp.cm^{-2} i.e. 4.6 atmosphere at an elevation of 92 meter and 6 meter above sea level and the average level of the Danube, respectively. The distance of the Danube river bed from the well is not more

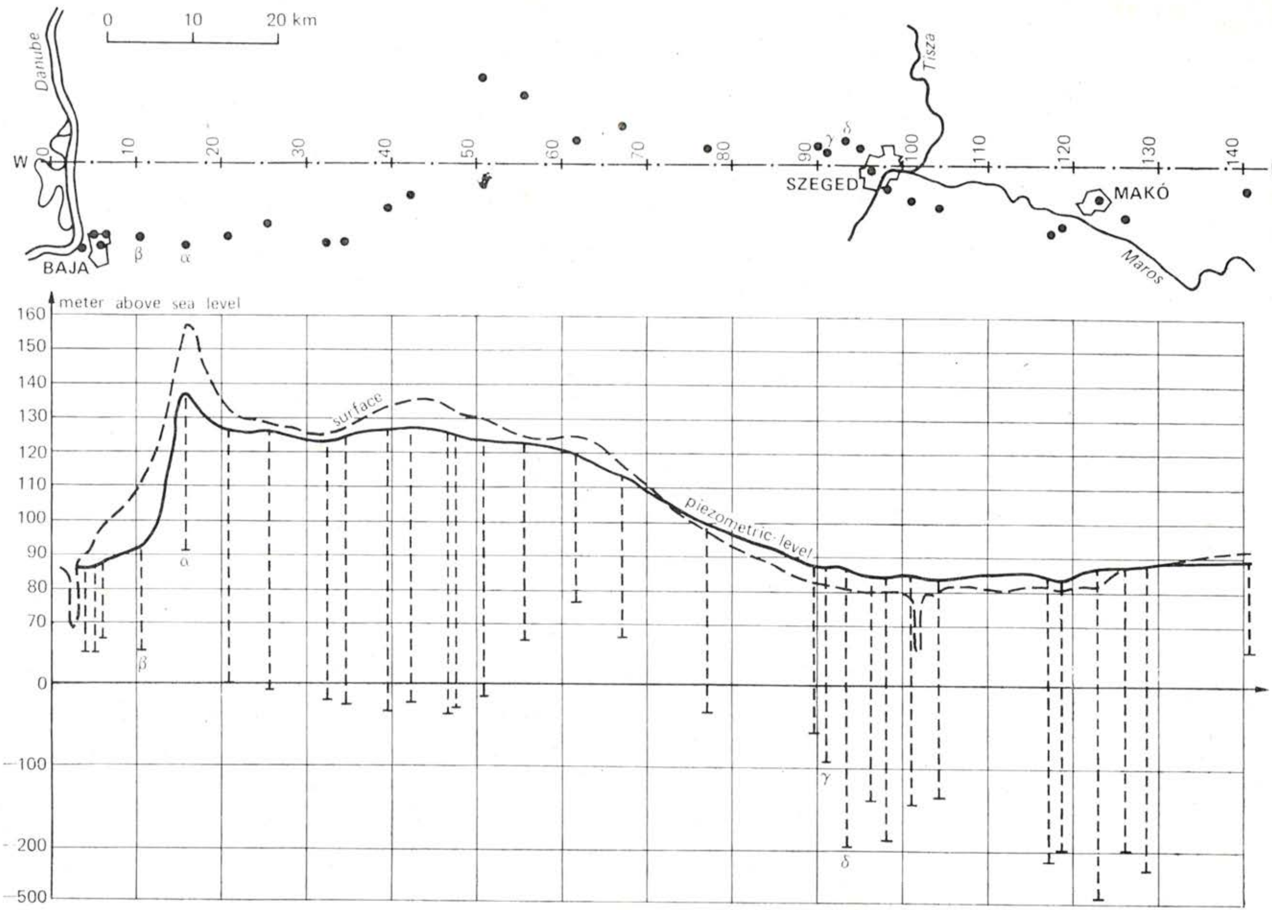


Fig. 2. Cross section of the Great Plain between Baja and Makó

than 13 km. The next well 5.5 km towards the Danube "β" has an underside of 48 m and the static head of the hole is of 92 m above sea level equal to the underside of the well mentioned afore. The static heads of the wells are manifesting the strain situation of the potential space, consequently, the horizontal component of the withstand of water bearing layers is of 44 meter i.e. equal to 8 $m.km^{-1}$ or 0.8 $kp.cm^{-2}$. This is identical to 0.8 atm. The behaviour of the deeper weels "γ" and "δ" with relatively higher surplus born by the resistance of upper layers can be observed on the cross section. The similar shape of the geodetic surface curve line and that of the piezometric line, being latter one somewhat flattened, is rather striking. That cannot be derived from, or interpreted by the classical artesian theory; the shape of the curves does **not** obey the accepted rule.

Building up our theses we have to presuppose the following facts:

- water is not an ideal liquid,
- ground-water flow is hampered by a limit resistance caused by the physical character of the layers,
- under this determined limit of resistance underground water flow is impossible, consequently, the original arrangement of the volumetric stress remains stabile i.e. statical.

Carrying on research work in the potential realm of underground waters different gradients of stress can be observed. We have to take these gradients as statical state because the million years of geological ages has been long enough for any relaxation of discrepancies and underground water flows needed for the evolution of the balance. Water movement has always been hampered by the specific resistance of water bearing layers observable even in the case of open subsoil water table. The only difference is that the appearance of open *subsoil water table is governed only by the gravitational field of force*, while *the other one* is *ruled by a manyfold volumetric stress-space born by the physical endowments of the water bearing layers* in a more complicated realm. For the sake of distinction the first one is determined as "*marginal slope*" and the other as "*stress limit*". The twin-ideas are, certainly, convertible with a simple multiplicator.

The documentation of the marginal slope in laboratories seems hardly available because using rough clastic material the slope would be too gentle as compared to the sensitivity of measuring equipments, while using fine sands when the observation would be possible in anticipation we have no time enough for the developing of the balance.

The existence of stress limit in our geohydrostatical theses is clearly demonstrable with the evidences given in the cross section of the Hungarian Great Plain. The quantification is but difficult because the interferences caused by heterogeneous layers. Preassuming the existence of a homogeneous layer structure large enough to determine the characteristic marginal limits, the correlation between the marginal slope of subsoil water table and the stress limit of deeper ground-water flow could be processed without any laboratory facilities. For time in Nature is practically endless and limit values similar to the surface conditions are yet relaxed in a considerable depth.

Furthermore we have to scrutinize the original features of deeper field of force of underground waters. Starting from our standpoint:

subsoil water is a category of subsurface water with an open surface communicating more or less with the atmosphere having a dynamic oscillation controlled by the climate. On the basis of the theses discussed above the earlier categorization of *ground-waters* as artesian, vadoze, compact, fossil, juvenile, seems to be not suited to the phenomenon, for the characterization among the different sort is not distinguished enough.

Marginal superficies are very worthy of attention showing convenient discrepancies for further analyses.

Upper surface of the ground-water reservoir is the subsoil water table. The subsoil water table known with minuteness of detail is not a flat panel but adopts itself roughly to the morphology of the geodetic surface. The outline of the bottommost edge plane cannot be drawn up so unanimously. We may realize the uppermost surface of the ancient sea deposits because of their different conditions of sedimentation and/or different water quality. The abyssal rock would be seen also similar margin. Preassuming the existence of juvenile or dissociated water recources we may suppose even interconnections with the magma through the fissures of the crest.

Horizontal marking of the borders is also dubious depending upon the structure of layers and the morphology of the surface.

This delineation of the volumetric shape is negligible according to the theses, for water-bearing layers of the Hungarian Great Plain are large enough. Marginal limits are too far removed to be able to alter the volumetric stress picture of the realm.

All underground layers convenient for storage are water-bearing and saturated the static states of which can be disturbed only exceptionally.

This statement deducted from the geologic evolution process is universal. Extremely dry geologic ages or local climate are to be taken as rare exceptions which don't disprove the existence of interconnected large volumes of underground waters at all. The imaginable cause of such a phenomenon might be the quick subsidence of dry sediments into the depth. In this case the dry block might be surrounded by the sea of ground-waters without saturation. Obstacles of saturation can be assigned to the presence of gaseous elements and/or the high stress limit of the layers e.g. clay-membranes.

Classical hydrostatic situation can prevail only in layers of negligible inside stress limits. In a substance of considerable stress limit hydrostatic parameters must be judged in the spotlight of an unusually different contemplation.

Simple hydrostatical situation in the case of driven wells is exceptional in water-bearing layers which are essentially permeable. The continuum of water-bearing layers of total permeability is just as scarce as of total isolation. Practically, we have to count always with some degree of semi-permeability in a wide scope. Stress limits in quasi-permeable layers and quasi-nonpermeable layers are relatively low and high, respectively.

Water-bearing layers are not solid but the more non-permeable, the more plastic. For the stress state of ground-waters the plasticity of the layers is responsible. This is the prevailing cause.

Generally, increasing surplus or other deviation of stress related to the increasing depth of the undersides of wells from the surface is observable.

General conception of the stress-theory

a) *Geological load* is caused by the weight of the sediments in the field of gravitational force. This is, usually, taken into account at engineering design in the field of soil-mechanics.

b) *Hydrostatical pressure* in the water, exclusively, without any interference of other substances (see Fig. 3) tending in the opposite direction of the geological load is calculated, also as the up-lifting Archimedean force in the field of soil-mechanics.

c) The *force elevating water* over the surface *is* neither geological load nor hydrostatic pressure but an *extra deviation*. In the lack of better terminology we shall define this deviation as *geohydrostatic deviation*. The more impervious continuum of sediments the more geohydrostatic deviation is observable. This occures in the tubes of driven wells above the average subsoil water table. Certainly, the physical situation of water in the tube is a clear-cut hydrostatic one.

Horizontal resistance is generally less than vertical one because of the changes of heterogeneous layers are more frequent.

This is the reason of lower heads of wells up the hill of the divide between the Danube and Tisza rivers comparing with the surface. On the contrary piezometric heads of driven wells in the alluvial surfaces are lifting up sometimes considerably in the valleys near to the rivers. (It might be mentioned the 54-meter surplus geohydrostatic head of a one-thousand-meter-deep well near Szeged.)

Geohydrostatic stress effecting the resources of water-bearing layers has been evoluted during the geologic development of the sediments organically and this fact results locally and individually all the components having effect in space and time.

The most simple explanation of the origin of the potential stress situation observed seems to be the long-lasting and simultaneous sedimentation of water and eolic and waterborn clastic materials. Two sorts of sedimentation is to be taken into account. Waterborn sediments consisting of bed-load and suspended soil are commonly known as alluvial deposits. Other sorts of waterborn layers are the relics of lakes and seas. The saturation of these sediments with water is clear.

In the case of eolic sediments we have to compare the relative volumes of sediment budget and water budget of the different geological ages. From the point of view of accumulation of water climatic conditions must be treated relatively. Humidity, independently of its grade might be inadequate for the accumulation if the total of evaporation and run-off exceeds the precipitation in the long run and vice versa small precipitation could fulfil the preconditions of accumulation.

Dealing with the upper level of subsoil water and its water budget it has been already verified that the effect of evaporation decreases along the vertical line towards the depth, atmospherian interference doesn't work any more below a certain level. This underneath limit of subsoil water budget is determined by the physical features of the upper layers and the annual and multiannual climatic oscillations. Phenomena prevailing in the realm of subsoil water budget are governed by thermal budget changing in two-phase and three-phase zones inside of the strata affected, alternatively. Accumulation is still possible if the conditions of entering of vapours and of recondensation are given in the clastic layers. Namely, precipitation is only one though most commonly known form of atmospheric fall-out.

Water budget of clastic layers is characterized by unhindered inflow under the prevailing conditions by the simultaneous accumulation of water during the sedimentation of water-bearing debris. By contrast, the outflow through evaporation cannot be superior to the income still not even during secular periods. The result of secular water budget such in a realm ought to be merely positive.

Supposing that the silting of one meter stratum of sediments lasted three thousand years in the Hungarian Great Plain; the total water turnover of the same period might be assesed as 2400 meters. The ratio shows three order of magnitude. Consequently, the probability of humid years verifies the 200–250 mm surplus during one thousand years. The ratio of fourth order of magnitude seems to be quite adequate for accumulation either in short humid periods or by recondensation in the depth not available by the evaporation effect any more. The precondition of this irreversible phenomenon is given in the Hungarian Great Plain because a period of at least one year during fifteen is needed for the retention of one millimeter water, only. We have to emphasise over (and over) the prevailing significance of the long geologic ages. In consequence, we dare say:

Ground-water originates from the accumulation process of geological ages. Exploiting these ground-water resources by human interference means final consumption of those, thus the final consumption of energies represented by the stress state of the special potential space simultaneously. This statement means the total negation of the existance of dynamic ground-water resources.

A simple model of the geohydrostatical stress space

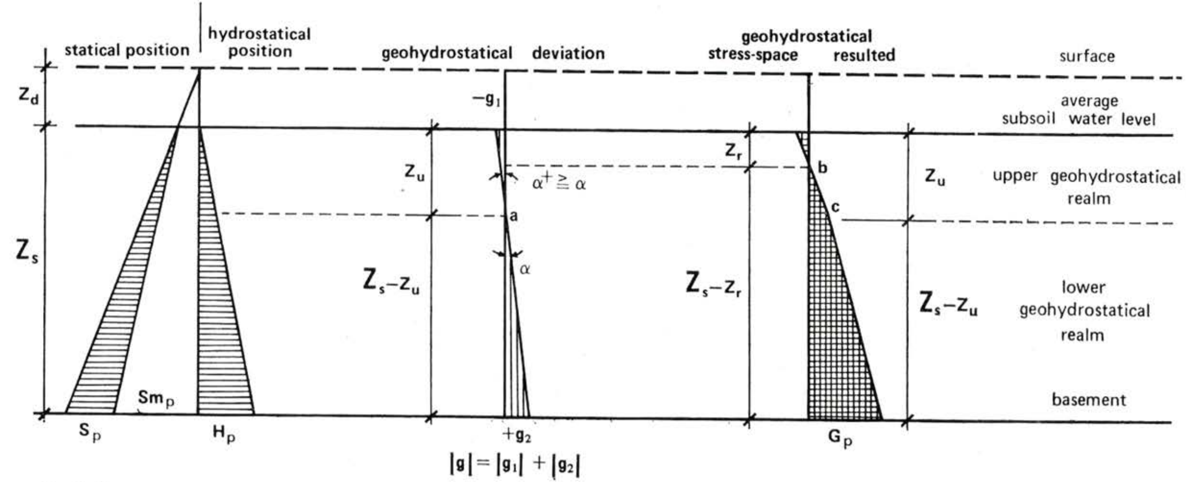
The simplest case of a geohydrostatical continuum is drawn up by symbols used for further mathematical proceedings in Fig. 3. Starting from the left and going to the right the followings are represented in the sketch;

- oblique line leaning to the left shows the abscissas of geological pressure,
- the opposite leaning line of the same set-up represents the abscissas of clear-cut hydrostatical pressure,
- second line to the left means the stress taken into account in soil mechanical engineering,
- central scheme of the figure shows the geohydrostatic deviations as differential abscissas (value "α" is the directional angle of geohydrostatic deviation),
- scheme at the right side is the result of the components i.e. the final picture observed in Nature.

Attention is to be raised to three zones even in that simplest case:

- uppermost zone is that of the dry layers,
- second is defined as upper geohydrostatical zone of saturated layers,
- third, per analogiam, is the lower zone of saturated geohydrostatical layers.

It has to be noted if Z_s is equal to 1.0 all the vectorial values have to be



Symbols:

Linear measures:

Z_d	thickness of dry layers
Z_s	thickness of saturated layers
Z_u	thickness of upper geohydrostatical realm
$Z_s - Z_u$	thickness of lower geohydrostatical realm
Z_r	zone of negative deviation
$Z_s - Z_r$	zone of positive deviation

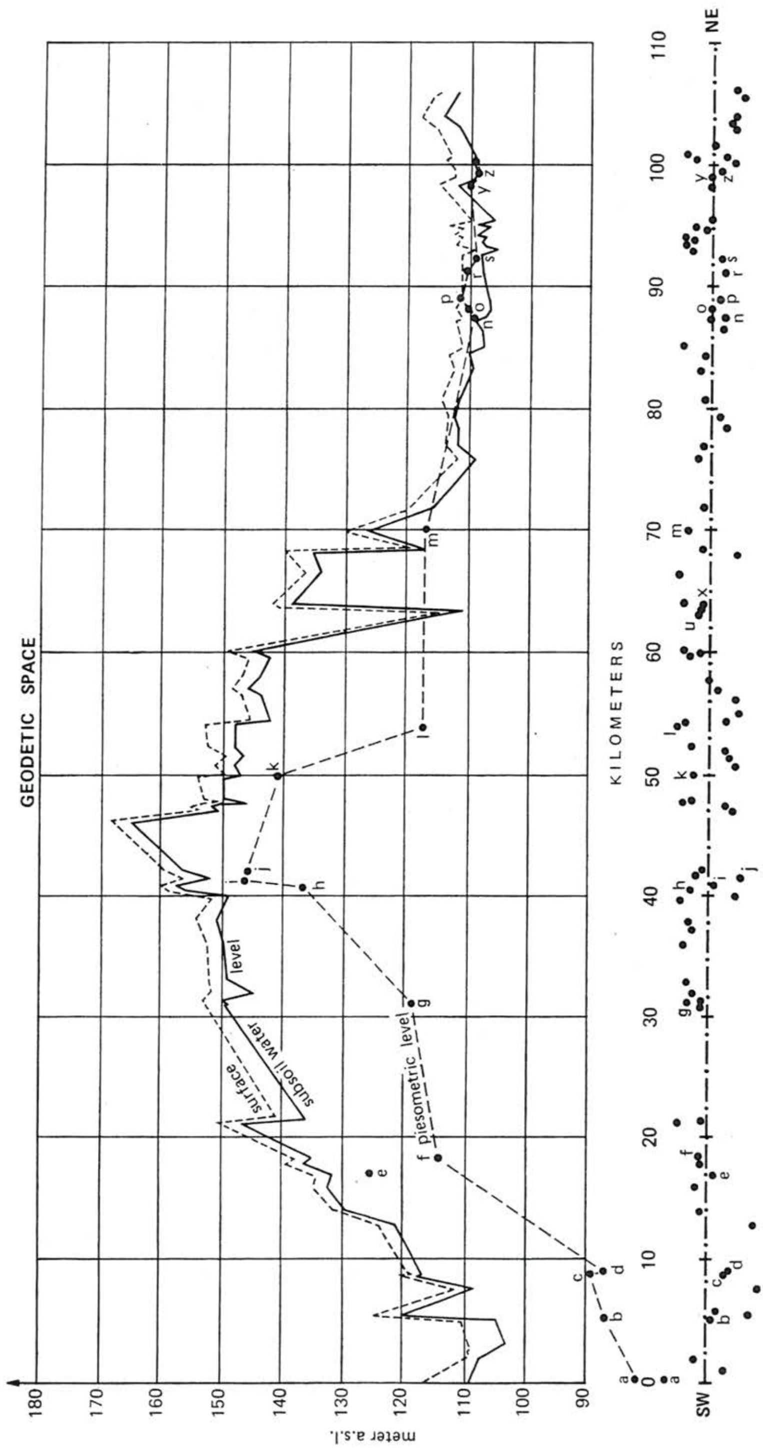
Discrete values of the scheme:

a	point of zero deviation
b	point of zero stress resulted
c	breaking point of stress line
α	directional angle of deviation

Powers and stresses:

S_p	soil pressure at the basement
Sm_p	engineering soil-mechanik pressure
H_p	hydrostatical pressure
$\lvert g \rvert$	geohydrostatical deviation
$-g_1$	negative share
$+g_2$	positive share
G_p	geohydrostatical pressure

Fig. 3. Simplest scheme of the geohydrostatical cross section in semi-permeable quasihomogeneous layers


GEODETIC SPACE
KILOMETERS
meter a.s.l.
surface
subsoil water
level
piesometric level
SW
NE

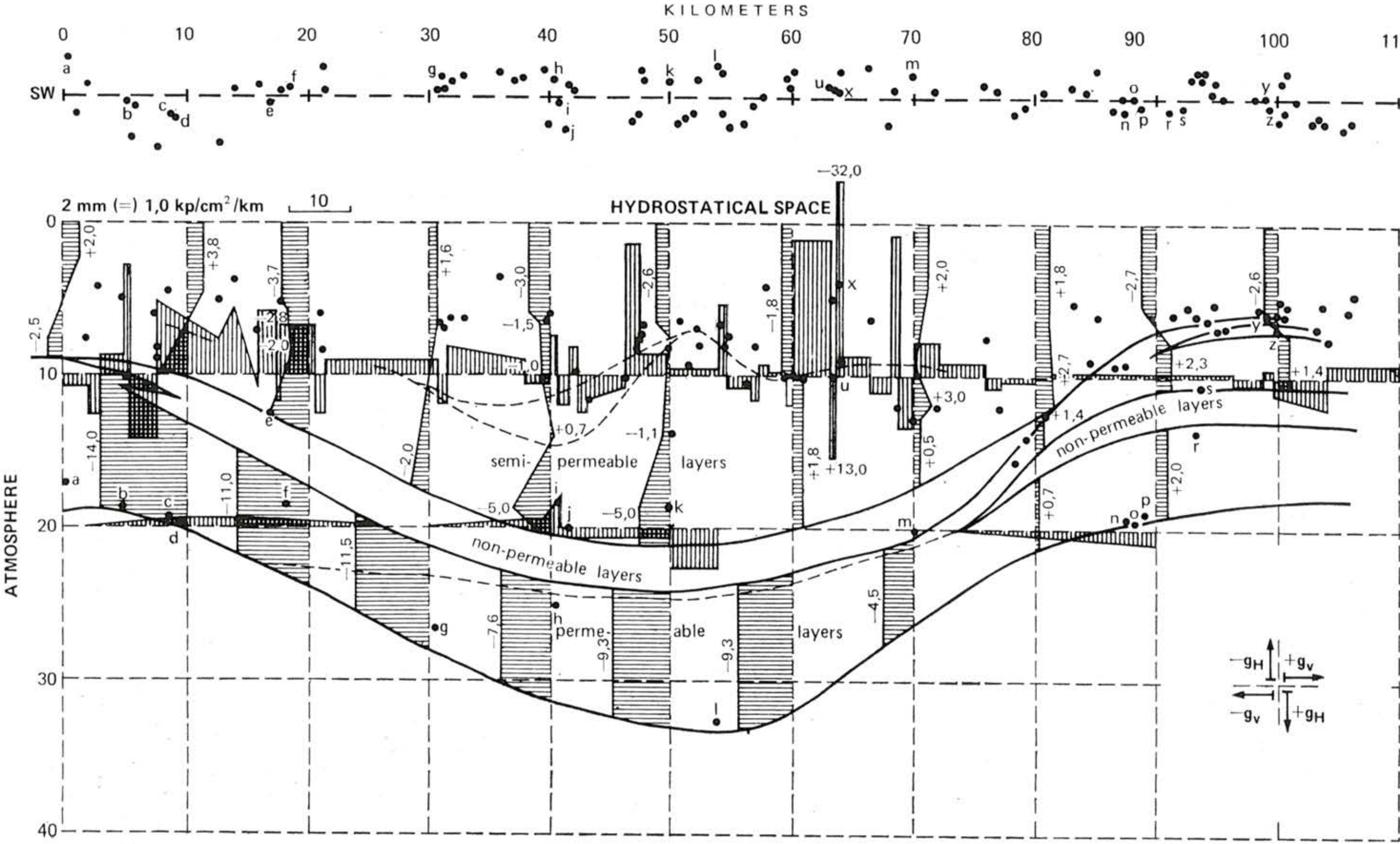


Fig. 4a—4b The cross section of the Nyírség region

taken into account as specific ones. Drawing up Fig. 4 as a practical representation of an individual cross section we used engineering measuring system in kp.cm^{-2}.km (=) Mp.m^{-2}.km units.

Interpretation of the geohydrostatical space by the practical example of the cross section of the Nyírség region

Fig. 4a and 4b represent a study case on the cross section of the Nyírség region. Sampling of relevant statistical data is based upon two thousand driven wells available in the Ground-water Thesaurus of the Hungarian National Water Authority – Water Management Centre. Wells selected for the purpose in a stripe of ten kilometer wide are 214. 94 of these has been used for the calculation of vertical and horizontal parameters of the geohydrostatical stress space. Some of the calculated specific values are represented in the figure in dimensions of kp.cm^{-2}.km. Those figures could be expressed more plastically in Mp.m^{-2}.km specific units.

It has to be noted that the upper representation of the figure is drawn up in the geodetic space while the lower one in a transformed hydrostatical space for the sake of better understanding of geohydrostatical deviations.

Specific vectors of geohydrostatical stress space are represented turned away 90 degrees clockwise i.e.:

- positive vertical deviation is tending from the left to the right,
- positive horizontal deviation showing vectors tending from SW to NE are represented turned down,
- negative vertical deviation tends from the right to the left and,
- negative horizontal deviation means a vector tending from NE to SW and turning up.

It is worthwhile to draw the attention to the rather considerable values of the vertical specific vector at 50 km of the non-permeable layer which is equal to 1410 Mp.m^{-2}.km and the horizontal one of 320 Mp.m^{-2}.km between the 60 – 70 km.

It is also noticeable that the zero deviation curves shown by dotted line on hydrostatical space delineation and the turning points of negative-positive signes are connected with the adverse tendency of vectors.

Even in rough sandy – gravel permeable layers specific horizontal geohydrostatic gradients are showing potential stress differences of 3.1 – 4.9 Mp.m^{-2}.km. These secondary differences of deviations in the lower permeable layers – if decreasing by ten km steps – provide an other necessary but not absolute certain prove of the rearrangement of stresses.

The turning point of Fig. 3 "c" is also easy to find at 198 m at Szeged. From that value "α" might be calculated at individual sites. E.g. at Szeged tg $\alpha = 0.1$ i.e. $\alpha = 5° \, 43'$ and the specific value of geohydrostatic vertical deviation at site is of 112 Mp.m^{-2}.km.

Human interference by drilling wells into this delicate geological balance of the stressed volumetric continuum means a drastic singular change in the surroundings of the filter of the well. The decreasing of stress following the exploitation seems to be more important from the point of view of underground water uses than water abstract in itself. The descent of heads of artesian wells is due to the decreasing of geohydrostatic surplus and not to the significant diminution of ground-water resources.

HYDROGEOLOGICAL FEATURES OF SOME DEEP-BASINS IN SE-HUNGARY AS REVEALED BY HYDROCARBON EXPLORATION

CARACTÈRES HYDROGÉOLOGIQUES DE CERTAINS BASSINS PROFONDS AU SE DE LA HONGRIE SUR LA BASE DES PROSPECTIONS D'HYDROCARBURES

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RÉSUMÉ

Dans la région de sud-est de la Hongrie, dans le bassin de Dorozsma—Szeged—Algyő les forages d'exploration ont développé au-dessous de l'horizon marno-calcaire du Pannonien inférieur un système aquifère complexe se rattachant hydrodynamiquement aux gisements d'hydrocarbures. La roche du gisement est composée de métamorphite fissurée paléozoïque grès quartzeux du Trias inférieur, dolomie du Trias moyen, brèche dolomitique et conglomérat du Miocène et du Pannonien inférieur. Les dolomies et les conglomérats ont les paramètres pétrophysiques les plus favorables.

Sur la base de la distribution de pression régionale, de la surface piézométrique, des niveaux des phases inclinées de même que de la situation régionale de la conductivité hydrodynamique, un écoulement d'eau de vitesse de 0,15 cm par an peut être marqué dans la direction est-ouest et un autre dans la direction sud-est—nord-ouest.

Les deux écoulements sont séparés par une zone avec une conductivité hydrodynamique extrêmement basse, située à l'ouest de la structure d'Algyő. Dans la structure de Dorozsma, les écoulements d'eau sont également faibles à cause des roches ayant une perméabilité basse. En outre des faibles valeurs de conductivité hydrodynamique, ce fait est supporté aussi par la présence de Cl et Mg caractérisant les eaux artésiennes stagnantes.

Introduction

Since the onset of the hydrocarbon-exploring activity in the southeastern part of the Great Hungarian Plain in the early 60's, almost 500 wells have been spudded, tested and completed near the town of Szeged (Fig. 1). About a quarter of these wells have hit the Pre-Tertiary basement complex composed of Paleozoic metamorphic rocks and Mesozoic sedimentary formations of various lithological character. There is a well-defined marker near the bottom of the Lower Pannonian—a 5 to 100-m—thick calcareous marl sequence underlain by a variety of reservoir rocks of different age forming a masstype reservoir complex all over the area involved.

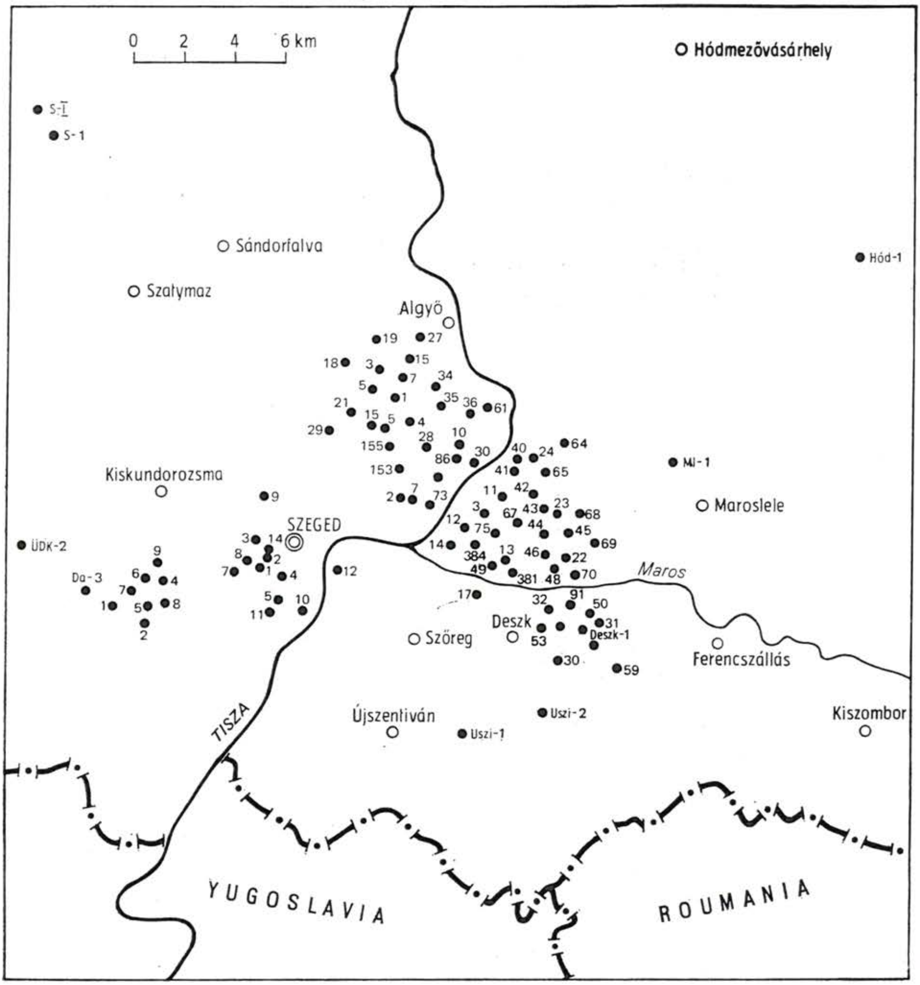


Fig. 1. Location map

General geological setting

The geological boundaries of the Szeged Basin are formed by the Makó Trough in the east and a smaller depression of a more or less local character between Kiskundorozsma and Mórahalom. Similarly, a local depression could be outlined in the north and in the south, as evidenced by data gained from boreholes Sándorfalva-I and Forráskút-1 as well as Újszentiván-1 and 2, respectively (Fig. 1).

The oldest formations of the area are the Paleozoic metamorphic rocks of different lithological character, a detailed discussion of which is beyond the scope of this study. The heavily weathered surface zones of this metamorphic complex act as reservoir rocks of rather poor quality ($\varnothing = 2-8\%$; $k = 1-10$ mD) in many places.

The subsequent lithostratigraphical unit uncovered in this area is a red, poorly-bedded, locally-brecciated, fine-grained Lower Triassic quartzose sandstone series with a thickness varying between 12 and 188 m. This series with variegated shale intercalations can be found on the eastern flank of the so-called Szeged-arch only, but is more frequent in other localities of the Great Hungarian Plain (Ásotthalom-Palic, Gyoma, Dombegyháza etc.). The siliceous cement material and the common quartz overgrowths around well to moderately well-rounded clastic (mostly quartz) grains manifest themselves by rather poor reservoir rock parameters: $\varnothing = 3-6\%$; $k = 5-20$ mD. This series is unconformably underlain by Paleozoic metamorphic rocks and overlain likewise unconformably by Middle Triassic or Miocene dolomite or conglomerate, respectively, depending upon its regional and tectonic position within the Basin (Figs. 2–4).

A massive occurrence of Middle Triassic dolomite, a brecciated dolomite series is characteristic of the western flank of the Szeged-arch only (Fig. 2). This formation underlain by Paleozoic or Lower Triassic and overlain by Miocene or Lower Pannonian rocks, could be detected up to the western flank of the Algyő structure (boreholes Szeged-12, Algyő-18, 26 and 29). The sections of these wells are considered as evidence for a general distribution of this formation of considerable thickness (locally around 200 m) in the area between the arches Szeged and Algyő. The reservoir rock characteristics of this dolomite and brecciated dolomite series are fairly favourable: $\varnothing = 7-13\%$; $k = 50-1000$ mD.

The Paleozoic–Mesozoic basement complex is unconformably overlain by Lithothamnium-and-Foraminifera-bearing Miocene as well as Limnocardium-bearing Lower Pannonian conglomerate series with an average thickness of 16 and 20 m, respectively. Generally speaking, there is no any overlap observable in the regional occurrence of these formations i.e. the Miocene and Lower Pannonian conglomerate beds do not occur together in the same borehole.

Detailed petrological investigation of this coarse-grained sequence (Bérczi, I. 1970, 1972, 1973) has revealed that its mineralogical composition dominated by rock fragments corresponds to the rejuvenation litharenite's group (Fig. 5). As for the clay minerals of the formations involved (Bérczi, I.–Viczián, I. 1975), a well-defined transformation zone could be observed at a depth of 2500–3000 m. This transformation accounts for the disappearance of the expanding clay minerals and for a sudden increase in the illite content below this zone (Fig. 6). In addition, the kaolinite could also be transformed into illite in this depth interval. The phenomenon of kaolinite transformation, however, is not general. The Lower Pannonian conglomerate series of Algyő, for instance, became saturated with hydrocarbons between the kaolinite enrichment in the early stages of diagenesis and the kaolinite degradation in the late diagenetic stages. The presence of hydrocarbons has prevented the inflow of aggressive interstitial waters necessary to form physico-chemical conditions favourable to kaolinite degradation. Thus high kaolinite content of the Lower Pannonian conglomerate has been preserved (Fig. 6).

As evidenced by a detailed textural investigation of the coarse-grained sequences (CM-plots, Sahu's discrimination equations, interpretation of

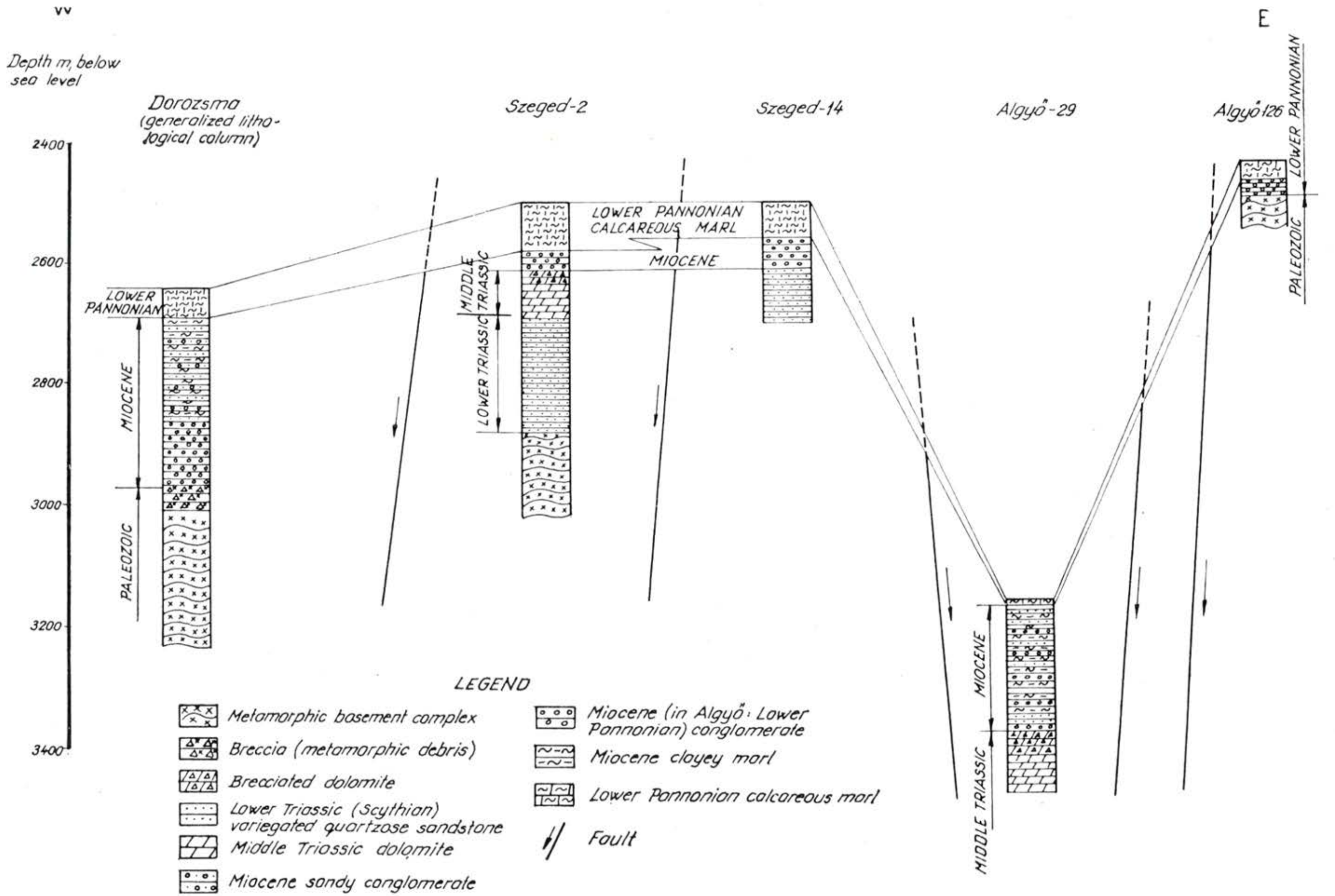


Fig. 2. Generalized lithostratigraphical section across the basal aquifer in the Szeged Basin

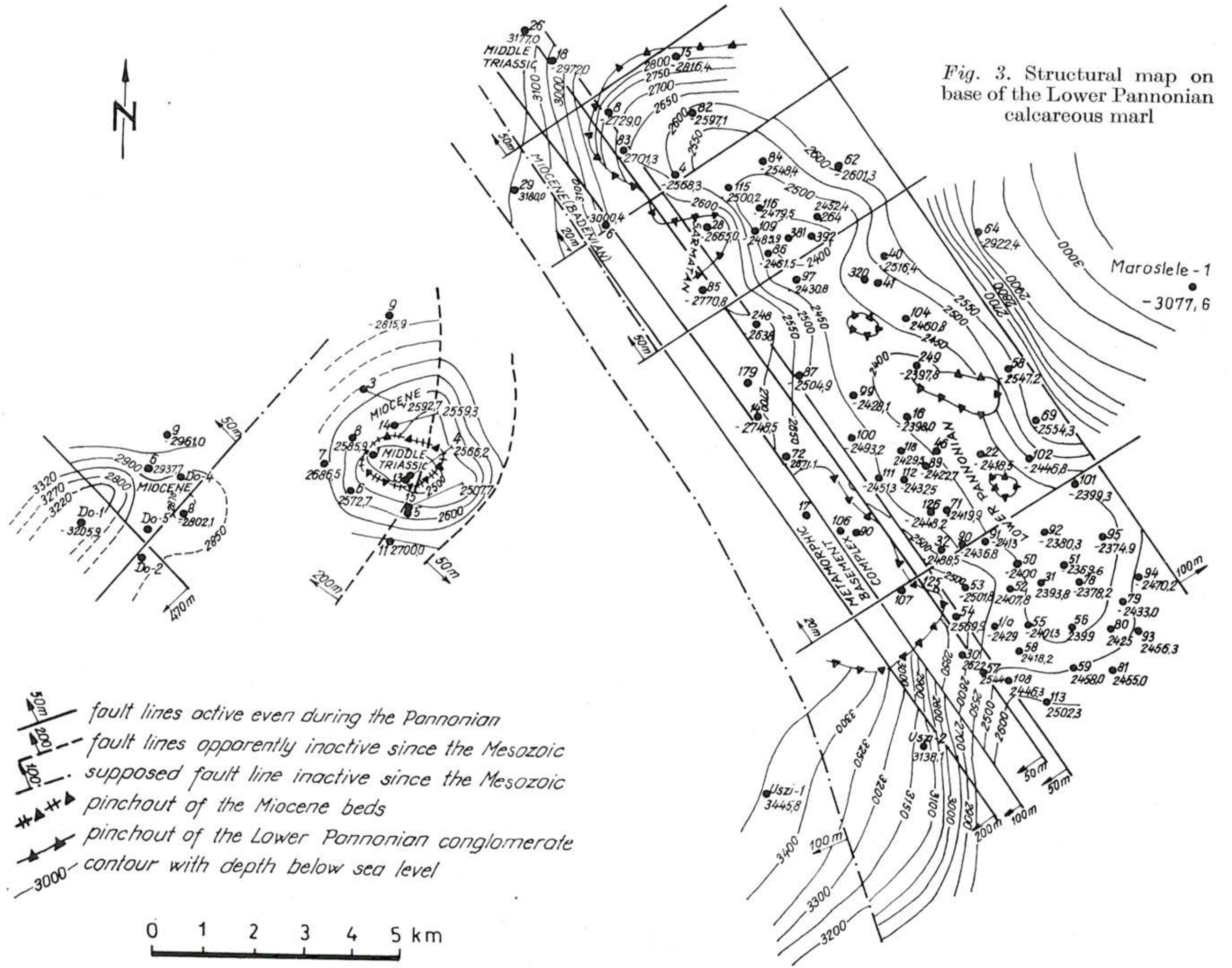


Fig. 3. Structural map on base of the Lower Pannonian calcareous marl

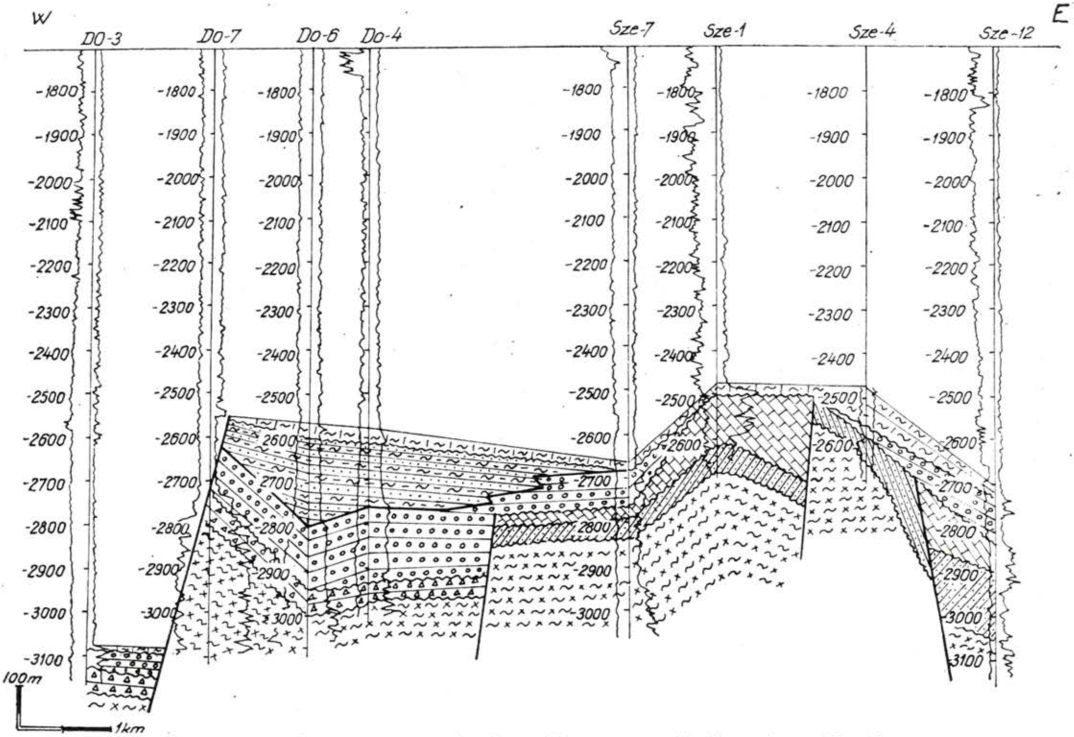


Fig. 4a Structural profile from Dorozsma via Szeged to Algyő
For legend see Fig. 4b

bedding characteristics), the turbidity currents dominated the depositional environments during the Miocene and the earliest Pannonian (Bérczi, I. 1970, 1972, 1973).

The Lower Pannonian calcareous marl is widely distributed in the area involved, and is considered to have as the principal role in trap-forming.

Tectonic setting

Three generations of faults can be separated within the Szeged Basin. The earliest one is of Mesozoic age supposedly without any renewal during the Tertiary and Quaternary. These slightly curved faults with an NNE–SW strike are characteristic of the Szeged-arch (Fig. 3). Their position and throw (200 and 50 m, respectively) are evidenced by the absence of the Middle Triassic dolomite series in the central part of the Szeged-arch (boreholes Szeged-4, 5, 14 and 15). The fairly smooth "blanket type" setting of the Miocene beds overlying the Mesozoic formations in this arch serves as an evidence against any rejuvenation of these tectonic lines since the Mesozoic era.

A more complicated tectonical evolution is characteristic of both Algyő and Dorozsma. The western flank of the Algyő-arch had been displaced along

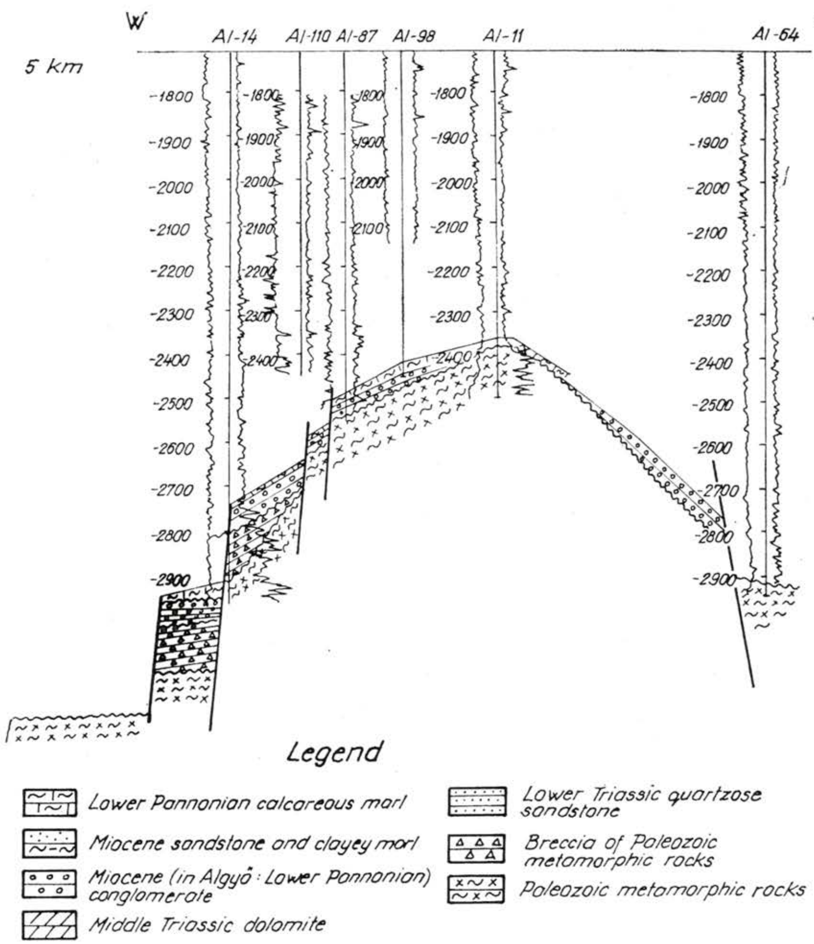


Fig. 4b

NW–SE faults between the Early and Middle Tortonian (Badenian). As a result of this, the Paleozoic and Mesozoic formations here were overlain by Middle Tortonian (Badenian) conglomerate beds, while in other areas a continental period was continued. The transition from the Badenian to Sarmatian was marked by more widespread tectonic movements leading to the emergence of all the Szeged Basin, as evidenced by the general absence of Sarmatian beds. (The only exception is made for the conglomerate bed in borehole Algyő-28, displaying brackish-water microfaunal elements.) At the same period the Dorozsma-arch has also been split into two parts by an NW–SE fault with a throw as high as 470 m.

At the time of passing of the Miocene into the Pannonian, the central and eastern part of the Algyő structure began to sink along longitudinal faults. This event means the onset of the Lower Pannonian conglomerate sedimentation. Simultaneously, the western flank of Algyő as well as the Dorozsma and Szeged arches, all covered by Miocene conglomerate beds got to and remained in an emerged position. This can explain why the Lower Pannonian conglomerate beds are missing over the Miocene formations of similar lithological character.

The latest tectonical lines are cross-faults with ENE—WSW and/or NE—SW strike in Algyő and Dorozsma, respectively. These are of post-depositional character in relation to the Lower Pannonian conglomerate series. Their throws are generally less considerable (20—50 m) as compared to those of the longitudinal faults (50—470 m). Their significance is in a supposed sealing effect as evidenced by some irregularities in the hydrocarbon-water contact in Algyő and by a small, tectonically screened, Upper Pannonian pool in Dorozsma. Contrarily, the older tectonical lines do not show hydrodynamic sealing effect, owing to the fractured character of the Paleozoic and Mesozoic reservoir rocks.

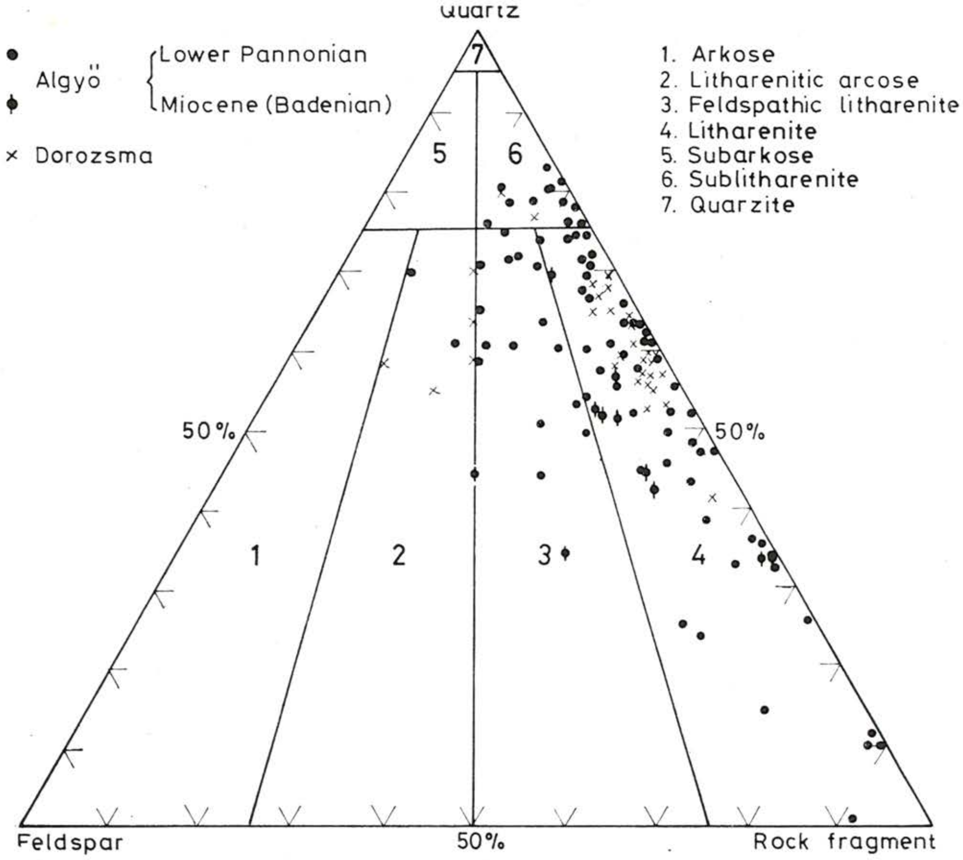


Fig. 5. Mineralogical composition of conglomerate beds of different age in the Szeged Basin

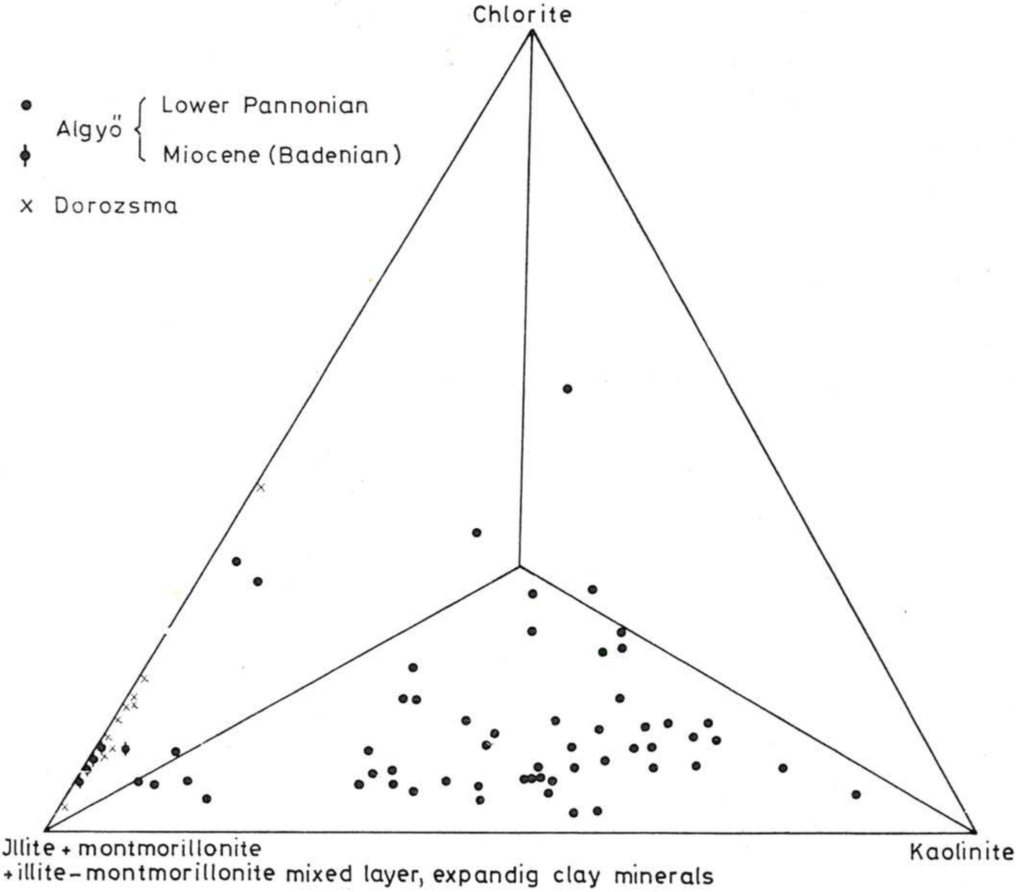


Fig. 6. Clay mineralogy of conglomerate beds of different age in the Szeged Basin

Reservoir geology

A number of production tests were carried out in the basal aquifer of the Szeged Basin. The measured production rates have shown that the Middle Triassic fractured and brecciated dolomite and the Miocene conglomerate series (especially in Szeged) as well as the Lower Pannonian coarse conglomerate beds are characterized by the most favourable flowing conditions, as evidenced by the production test data listed below:

Middle Triassic dolomite and breccia

Szeged-3 2995—2999 (2660—2664 m b.s.l.): 113 m^3/d brine flowing in 4 mm choke;
Szeged-9 3417—3428 (3331.0—3342.0 m b.s.l.): 146.4 m^3/d brine flowing in 6 mm choke;
Algyő-26 3255—3292 (3169.9—3206.9 m b.s.l.): 360 m^3/d brine flowing in 10 mm choke;
Algyő-81 2613—2640 (2529.5—2556.5 m b.s.l.): 411 m^3/d brine flowing in 24 mm choke;

Miocene conglomerate

Algyő-21 3209–3300 (3126.2–3216.2 m b.s.l.): 90 m^3/d brine flowing in 6 mm choke;
Szeged-2 2770–2774 (2685.3–2689.3 m b.s.l.): 43 m^3/d brine flowing in 10 mm choke;
Szeged-11 2804–2808 (2719–2723 m b.s.l.): 50 m^3/d brine flowing in 6 mm choke;

Lower Pannonian conglomerate

Deszk-1/a 2571–2573 (2489.3–2491.3 m b.s.l.): 145 m^3/d brine flowing in 6 mm choke;
Algyő-87 2590–2596 (2508.9–2514.9 m b.s.l.): 316 m^3/d brine flowing in 8 mm choke.

The rates of production from the Lower Triassic quartzose sandstone, brecciated locally and from the breccias composed of metamorphic rock fragments, as well as from the weathered zone of the metamorphic basement complex are highly varied, apparently depending upon the pattern of the

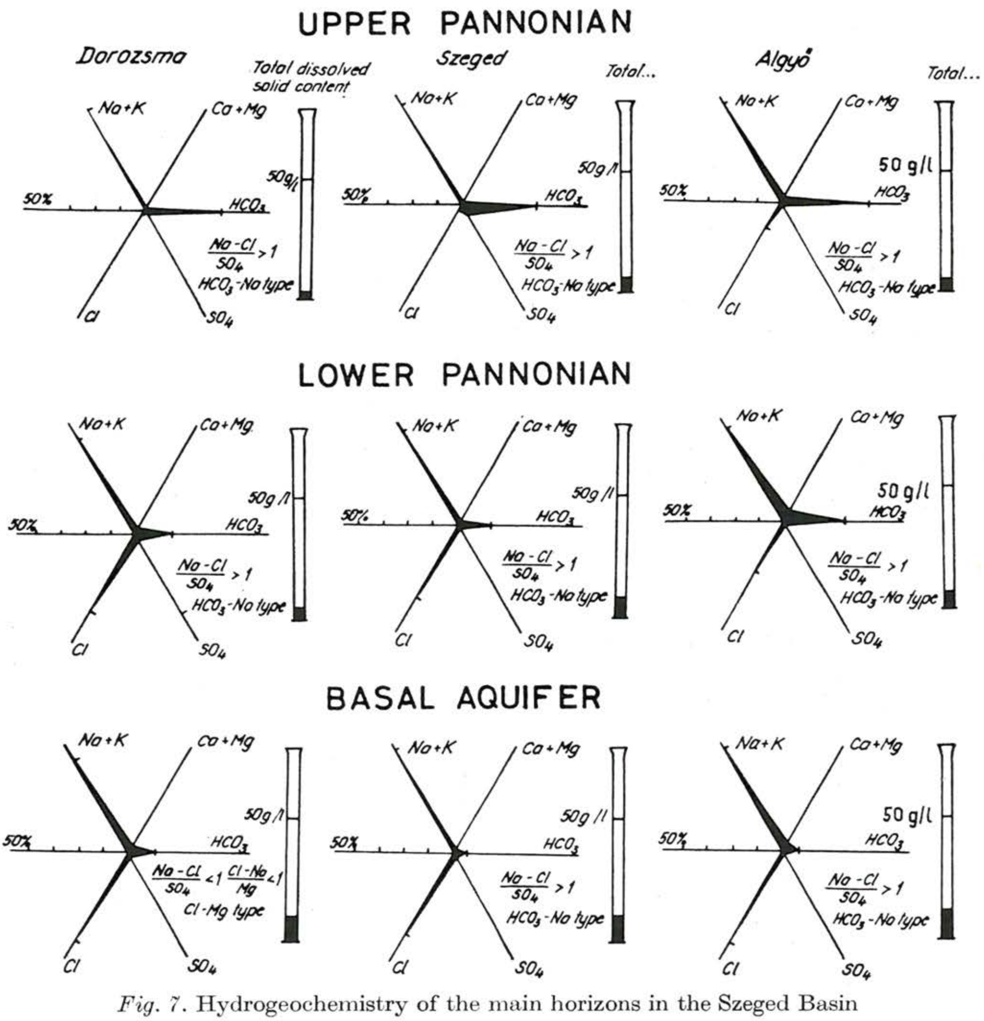


Fig. 7. Hydrogeochemistry of the main horizons in the Szeged Basin

natural fracturing. Apart from some tests that resulted in recording considerable rates of production from these reservoir rocks, there has been a number of tests without any fluid inflow observable:

Algyő-22 2523.5 – 2525.0 (2440.0 – 2441.5 m b.s.l.) metamorphic basement complex: no inflow;
Algyő-57 2653 – 2663 (2567.4 – 2577.4 m b.s.l.) metamorphic basement complex + Lower Pannonian conglomerate: no inflow;
Algyő-72 2777 – 2785 (2695.1 – 2703.1 m b.s.l.) metamorphic basement complex: no inflow;
Algyő-85 2924 – 2932 (2840.8 – 2848.8 m b.s.l.) metamorphic breccia: no inflow;
Algyő-248 2719.5 – 2780.0 (2638.1 – 2698.1 m b.s.l.) total interval of the basal aquifer: no inflow.

These dry tests listed above have the utmost significance, by testifying to the existence of an impermeable barrier in the western flank of the Algyő structure. This sealing zone indicated also by the regional pattern of contours on the map of hydrodynamic conductivity (Fig. 16), influences heavily the flow system of the basal aquifer.

Similarly, the reservoir rocks in Dorozsma (Miocene conglomerate with sandstone and siltstone intercalations + breccia from Paleozoic metamorphic rocks + Paleozoic fractured metamorphites) are characterized by extremely poor hydrodynamic conductivity as testified by production tests and by permeability values measured on core samples.

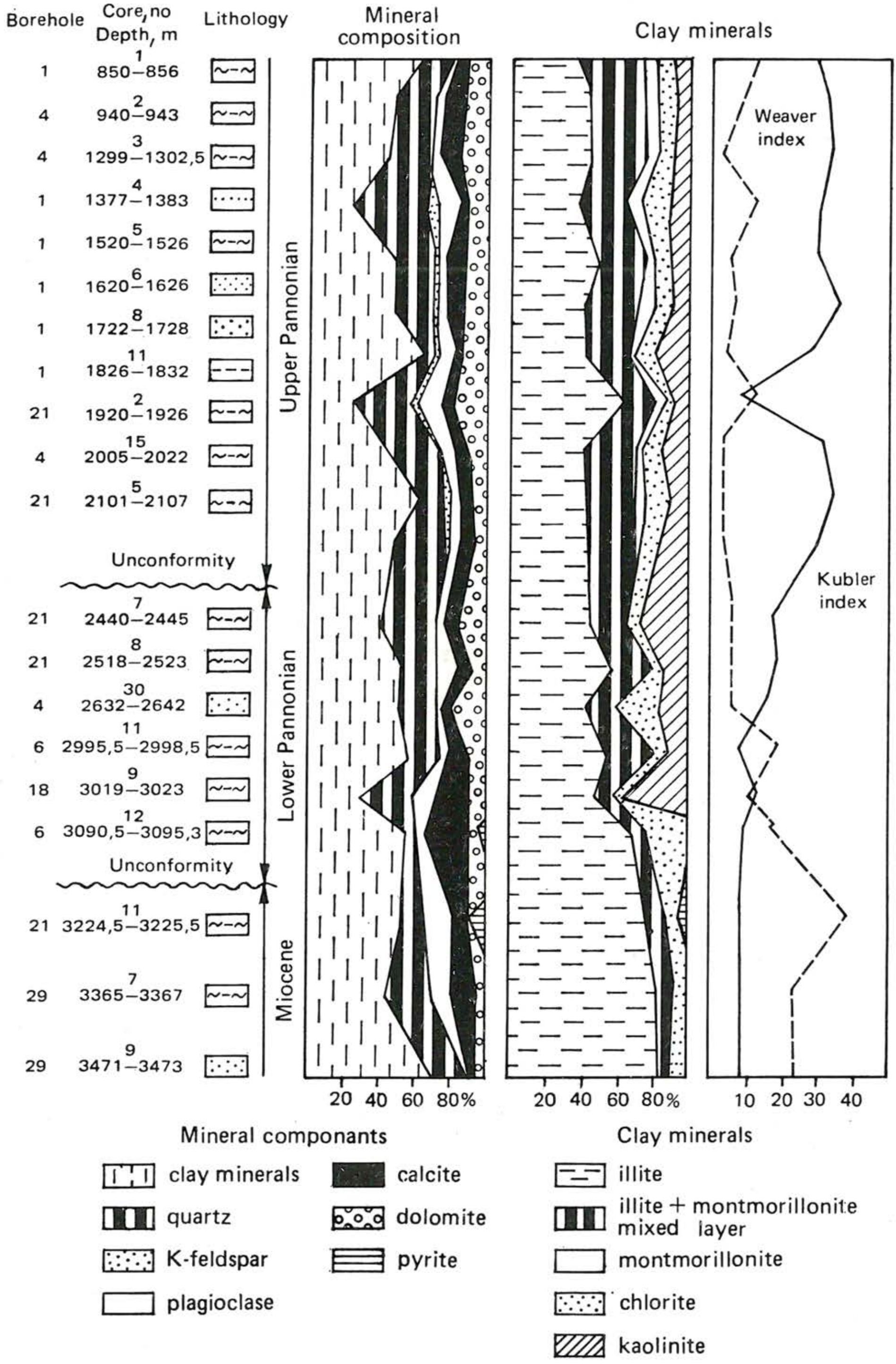
Hydrogeochemistry

The chemical characteristics of water gained from the basal aquifer have been thoroughly investigated and compared to those of waters from aquifers in higher stratigraphical position (Figs 7 to 12).

The principal hydrogeochemical characteristics of the different horizons in the Szeged Basin are shown in Fig. 7. A continuous increase in the total dissolved solids content can be observed from the Upper Pannonian (4 – 6 gr/l) on to the basal aquifer (10 – 15 gr/l) all over the area investigated.

The Upper Pannonian waters are dominated by HCO_3^- and $Na^+ + K^+$ ions; the Lower Pannonian waters are in half-way between the $HCO_3^- + Na^+$-dominated Upper Pannonian and $Cl^- + Na^+$-dominated basal waters (Fig. 7).

The gradual increase with depth in the Cl^- content at the expense of HCO_3^- is considered as diagenetic phenomenon. This assumption is corroborated by the changes in the mineralogical and clay-mineralogical features shown by two vertical profiles studied thoroughly (Figs 8 – 9). Drastic diagenetic alterations characterized by the disappearance of K-feldspars, expanding clay minerals as well as of kaolinite, and by an increase in crystallinity of illite started as deep as 2500 m and ended in 3000 – 3500 m. The alteration of formation waters due to the interaction of water and reservoir rock can be indicated in form of sudden decrease in Na/Cl and $SO_4 \cdot 100/Cl$ ratios around 2100 m i.e. at about the Upper and Lower Pannonian boundary. It is worth noting here that some years ago the authors were engaged in analysing the trace elements of pelitic rocks from the same formations of this area, in order to fathom out statistically significant differences for separating facies i.e. different depositional environments. In connection with this study, it has been a surprising result that a gap in trace element distribution between the Upper



Borehole
Core, no
Depth, m
Lithology
Mineral composition
Clay minerals
1 850–856 (1)
4 940–943 (2)
4 1299–1302,5 (3)
1 1377–1383 (4)
1 1520–1526 (5)
1 1620–1626 (6)
1 1722–1728 (8)
1 1826–1832 (11)
21 1920–1926 (2)
4 2005–2022 (15)
21 2101–2107 (5)
Unconformity
21 2440–2445 (7)
21 2518–2523 (8)
4 2632–2642 (30)
6 2995,5–2998,5 (11)
18 3019–3023 (9)
6 3090,5–3095,3 (12)
Unconformity
21 3224,5–3225,5 (11)
29 3365–3367 (7)
29 3471–3473 (9)
Upper Pannonian
Lower Pannonian
Miocene
Weaver index
Kubler index
20 40 60 80%
20 40 60 80%
10 20 30 40
Mineral componants
clay minerals
quartz
K-feldspar
plagioclase
calcite
dolomite
pyrite
Clay minerals
illite
illite + montmorillonite mixed layer
montmorillonite
chlorite
kaolinite

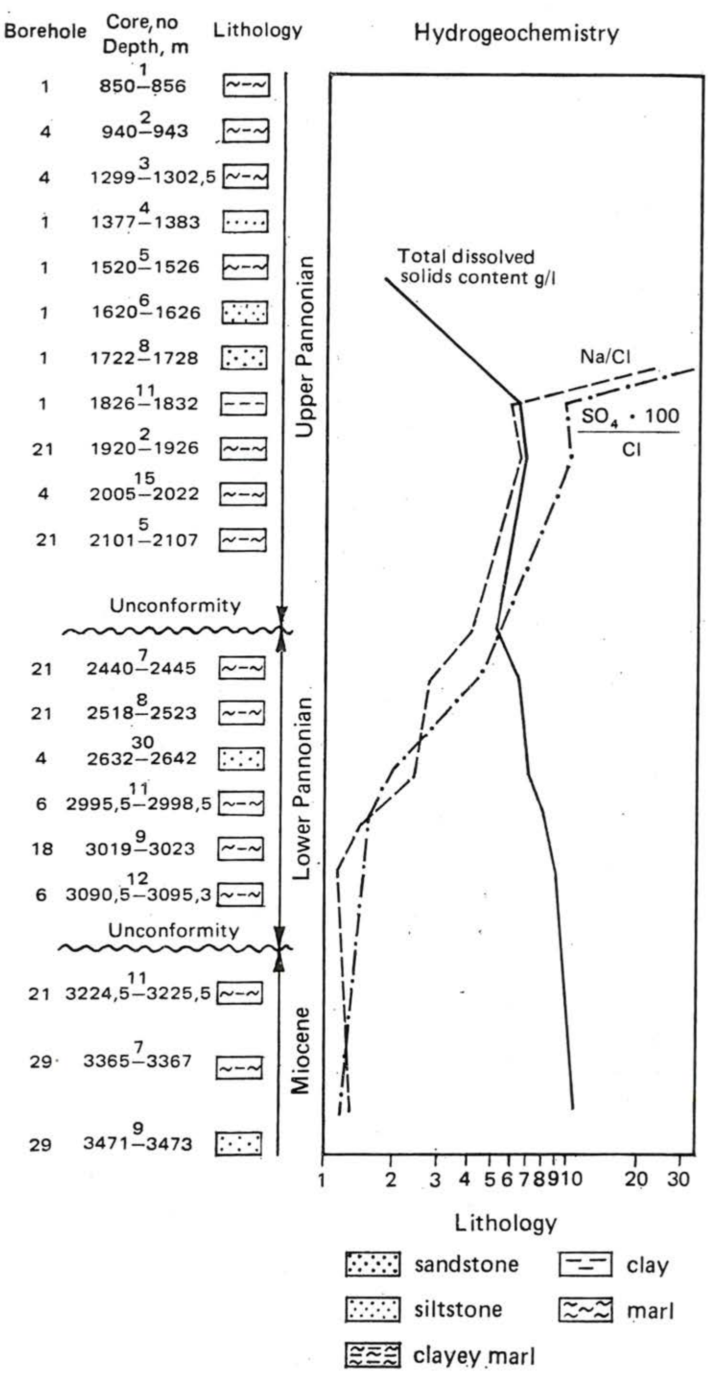


Fig. 8a—b Generalized lithological and hydrogeochemical columnar section of the Algyő area

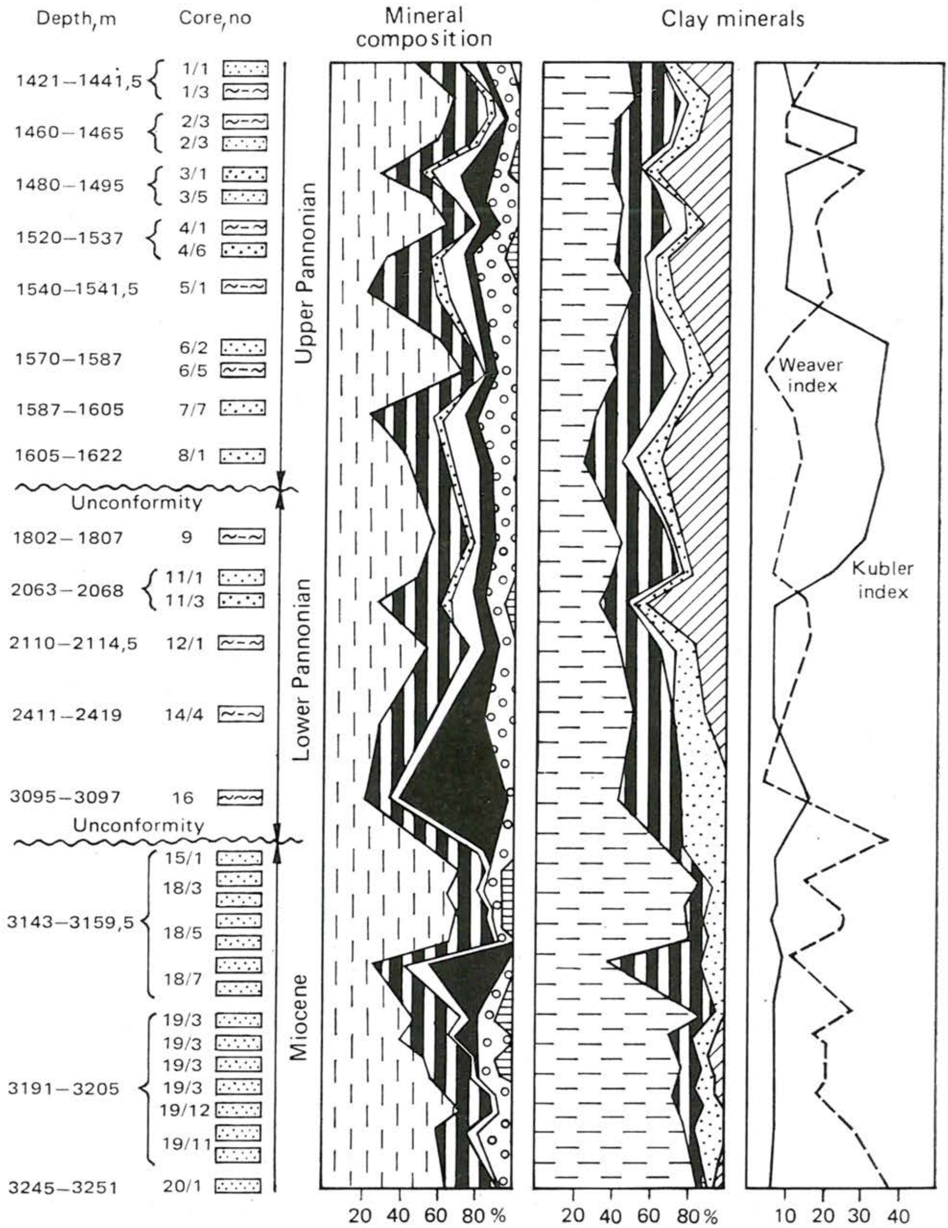


Fig. 9a–b Generalized lithological and hydrogeochemical columnar section of the Dorozsma area. (For the legend see Fig. 8a–b)

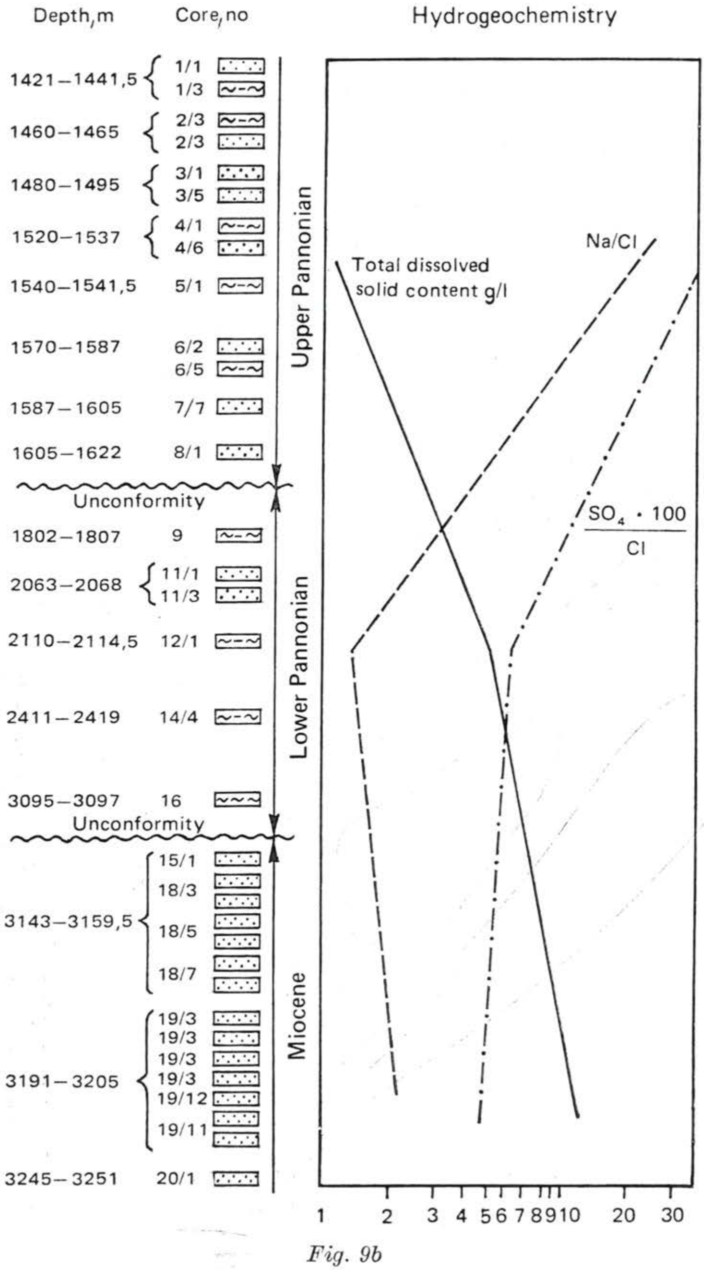

Depth, m
Core, no
Hydrogeochemistry
1421–1441,5
1/1
1/3
1460–1465
2/3
2/3
1480–1495
3/1
3/5
1520–1537
4/1
4/6
1540–1541,5
5/1
1570–1587
6/2
6/5
1587–1605
7/7
1605–1622
8/1
Unconformity
1802–1807
9
2063–2068
11/1
11/3
2110–2114,5
12/1
2411–2419
14/4
3095–3097
16
Unconformity
15/1
18/3
3143–3159,5
18/5
18/7
19/3
19/3
19/3
3191–3205
19/3
19/12
19/11
3245–3251
20/1
Upper Pannonian
Lower Pannonian
Miocene
Total dissolved solid content g/l
Na/Cl
SO4 · 100 / Cl
1
2
3
4
5
6
7
8
9
10
20
30

Fig. 9b

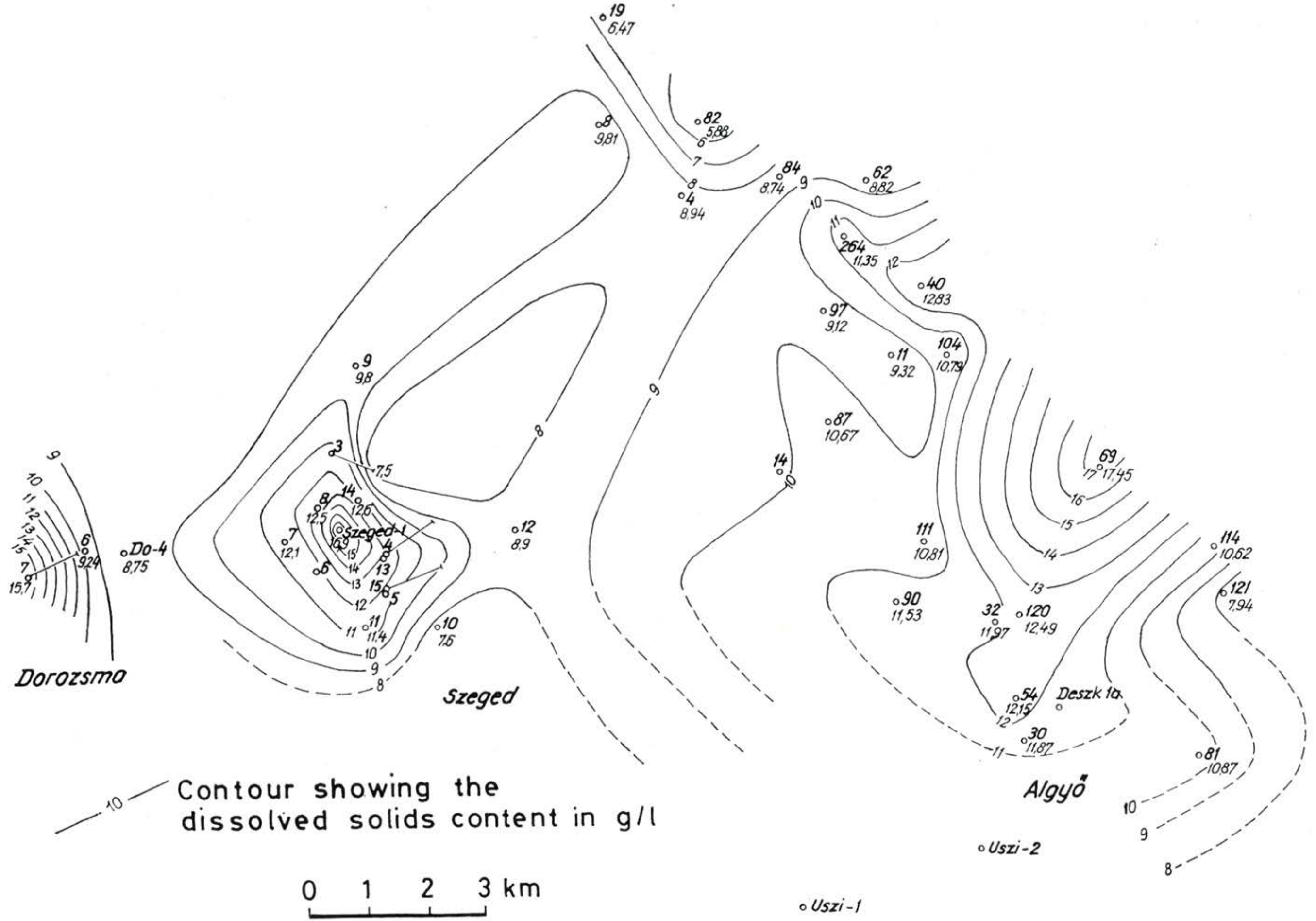


Fig. 10. Map of the total dissolved solids content in waters of the basal aquifer of the Szeged Basin

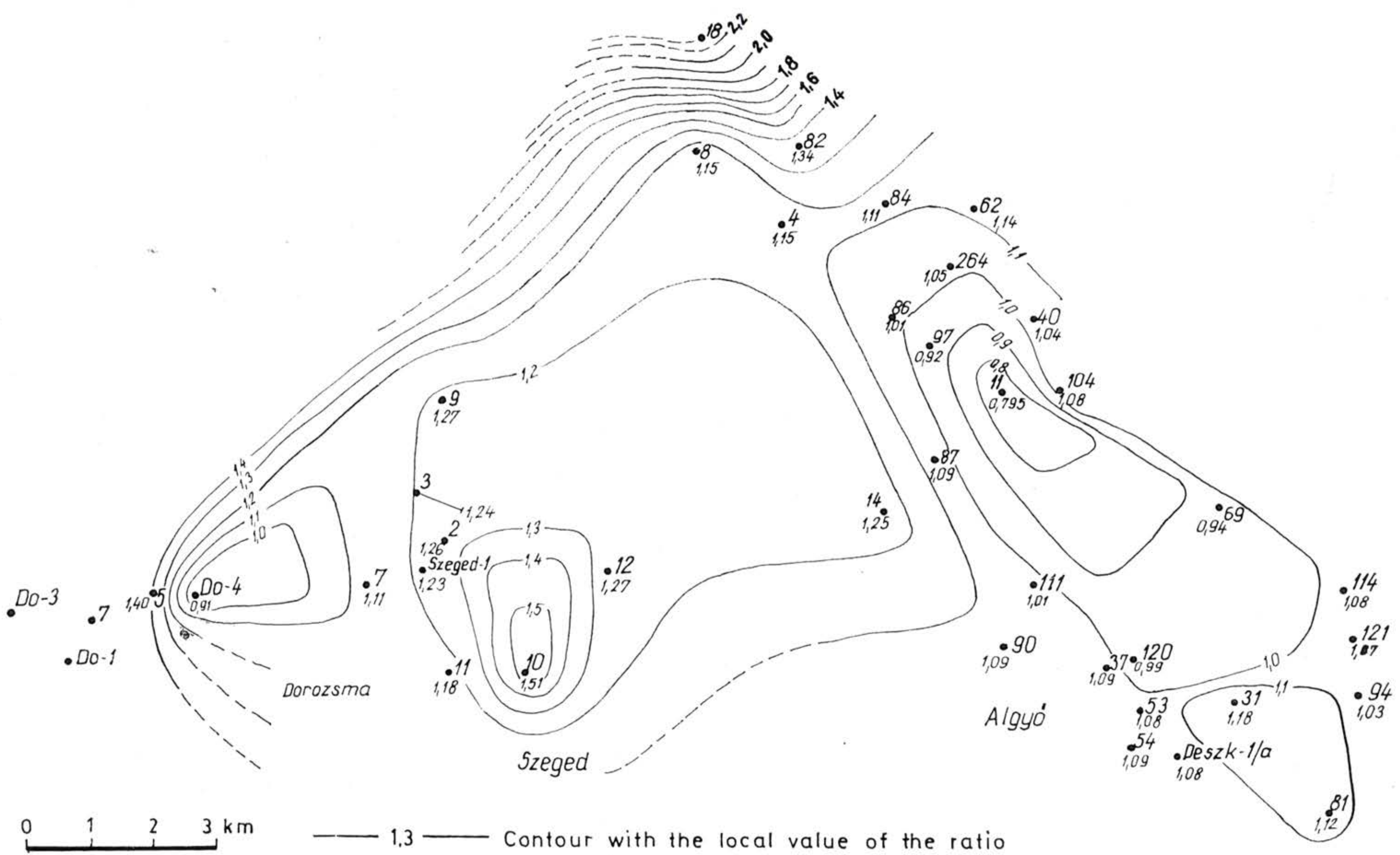


Fig. 11. Map of the Na/Cl ratio in waters of the basal aquifer in the Szeged Basin

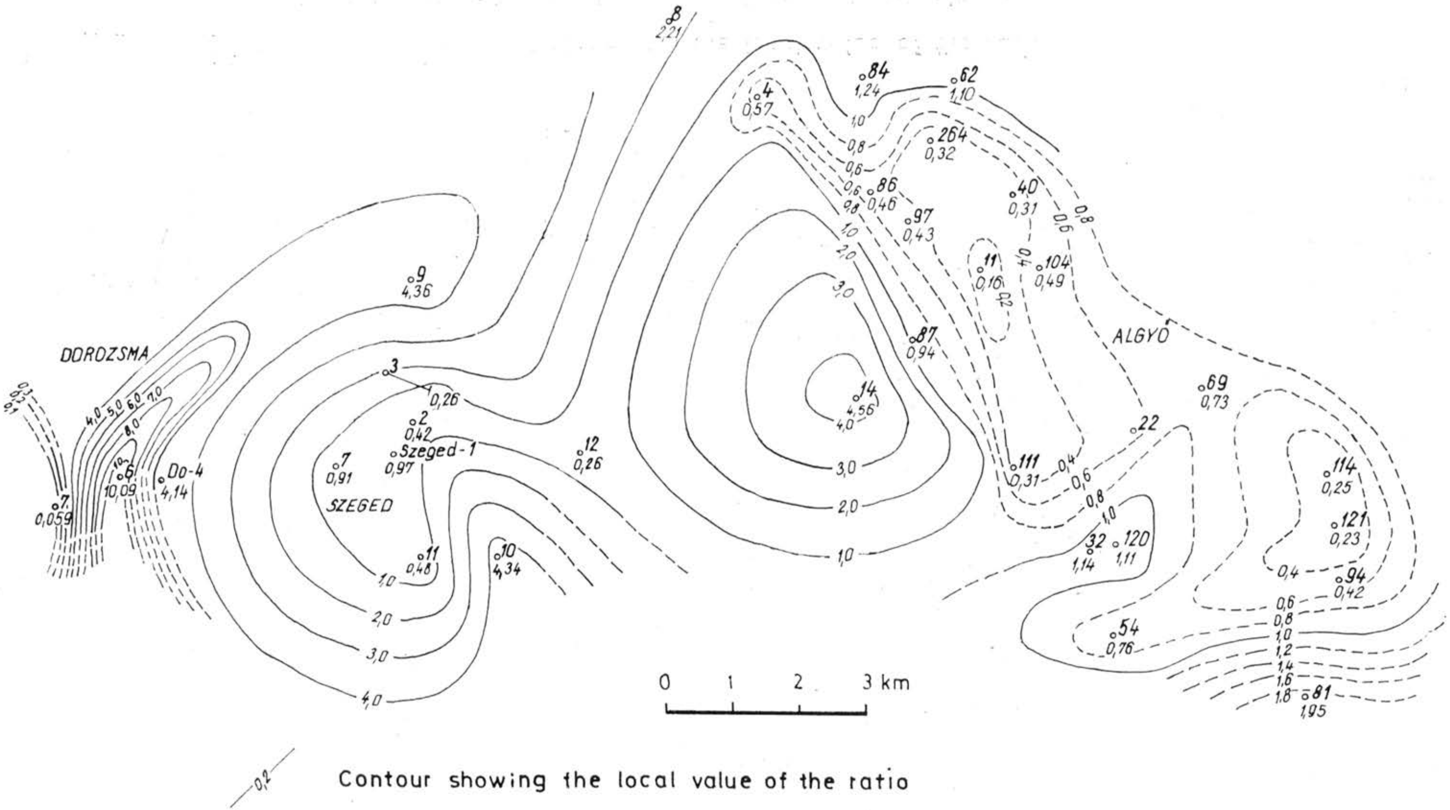


Fig. 12. Map of the $\frac{SO_4 \cdot 100}{Cl}$ ratio in waters of the basal aquifer in the Szeged Basin

and Lower Pannonian, characterized likewise by "brackish" fossils, could be found ampler than that between the marine Miocene and the brackish Lower Pannonian.

It is interesting, but not surprising anymore, that the situation is similar in the case of the hydrogeochemistry of waters raised from the formations listed above—there is a lesser difference in the content of total dissolved solids and in Na/Cl as well as in $SO_4 \cdot 100/Cl$ ratios between the Miocene and Lower Pannonian than between the Lower and Upper Pannonian (Fig. 7).

Regional variance in the three hydrogeochemical parameters of the basal aquifer, listed above, are shown in Figs 10—12. The total dissolved solids content shows positive anomalies over the hydrocarbon-bearing areas, supposedly as a result of interactions between the hydrocarbons and brine (Fig. 10).

Inversely, negative anomalies in Na/Cl and $SO_4 \cdot 100/Cl$ ratios pinpoint to the productive areas. In any case, the high Na/Cl ratio in the environs of borehole Algyő-18 may refer to recharge along faults (Fig. 11), while the relative high SO_4 contents in the margins of the productive areas might refer to more vigorous flow of water preventing the completion of process of SO_4 reduction according to some classical theories on hydrogeochemisty (Collins, A. G. 1975; Karcev, A. A. 1972; Krotova, A. V. 1969).

Finally, we would return to the diagrams in Fig. 7 to which the main Sulin-parameters of each horizon have been computed. Surprisingly enough, all but one formation belong to the group of HCO_3^-—Na^+-type waters characteristic of fairly dynamic systems, only the water of the basal aquifer in Dorozsma belongs to the Cl^-—Mg^{++} type characteristic of rather stagnant waters. This fact is in accordance with the observations concerning the extremely poor permeability conditions in Dorozsma, which prevent any significant water flow, while in other areas of the basin the prevailing hydrodynamic conditions might not allow the subsurface waters to be kept stagnant i.e. with high contents of Cl^- and Ca^{++}.

Reservoir pressure and temperature

More than 100 pressure values have been determined from pressure build-up curves registered during the production tests. The formation pressure-depth plot for the Szeged Basin is given in Fig. 13. For distinguishing the reservoirs characterized by anomalous pressures (i.e. overpressures), the hydrostatic pressure gradient is also shown. It can be stated that hydrostatic conditions are characteristic of the Pannonian reservoirs up to the lower part of the Lower Pannonian. The lowermost sandstone layers of the Lower Pannonian are characterized by an overpressure of about 15%, while an overpressure of 23—25% could be registered in the basal aquifer, with the exception of the deepest parts of the basin (Újszentiván, Maroslele), where the overpressure is as high as 50%. The pressure gradient is of about 0.15 at/m, but the authors are of the opinion that the gradient changes (presumably increases) with depth, consequently the linear approximation given by the dotted line could be considered as the simplest correlation only.

The temperature gradient estimated from the formation temperatures recorded in different boreholes of the area (Fig. 14) is 67 °C/km (or 15 m/°C in

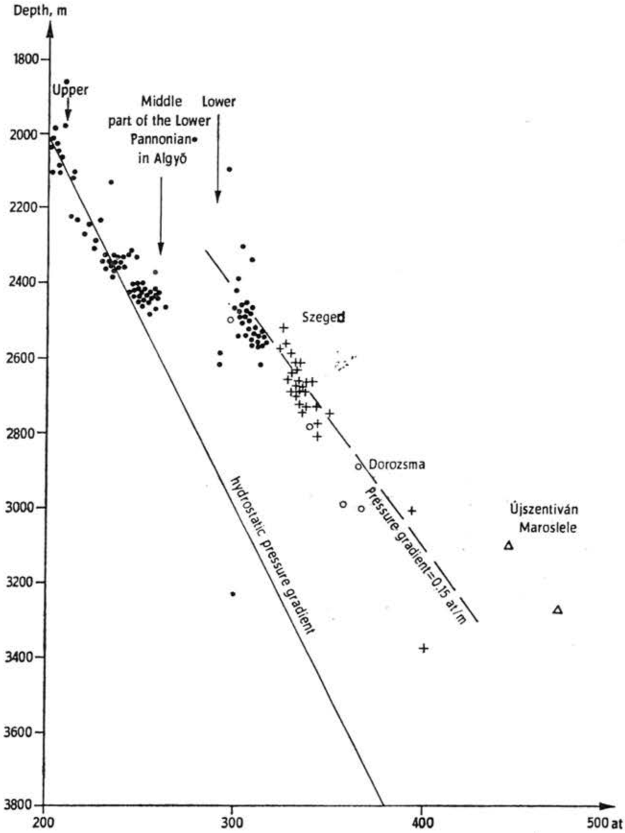


Fig. 13. Pressure depth plot of the Szeged Basin

inverse gradient) in the higher horizons of the basin, while it decreases to the normal value (30 °C/km or 33 m/°C) in the deeper parts of the basin. This vertical change in the temperature gradient is an interesting phenomenon, which is worth of analysing thoroughly in other areas intersected also by deep and ultradeep wells.

Regional pressure distibution and water flow

The hydrogeological studies of oil fields are always aimed at the determination of the direction and rate of water movement, if any, in aquifers attaching to the pools. Both exploration and production could benefit from the conclusions of these investigations.

In the case of the basal aquifer in the Szeged Basin, three questions have arisen to be replied.

1) Is there any sign of dynamic conditions in the sequence of layers involved?

2) Could be pressure differences indicated as significant enough to generate and maintain any water movement?

3) Is the hydrodynamic conductivity high enough to allow the water to flow and in which direction?

As for the first question, tilted hydrocarbon-water contacts could equally be observed in Algyő, Szeged and Dorozsma. In the eastern flank of the Algyő area, the hydrocarbon-water contact is in a position higher than in the western flank. The difference is as large as 25 m. Similarly, there is a difference of 20 m in the position of the hydrocarbon-water contact between the northern and southern flank of the Szeged-arch. According to the classical theories concerning hydrodynamics of the oil fields, these differences could be explained by water flows directed from E to W and/or from S to N, respectively. (The similar phenomenon in Dorozsma could be elucidated by special characteristics of the reservoir rocks, discussion of which is beyond the scope of our present study.)

For determining the regional pressure distribution of water flows and outlining their possible direction, piezometric surface values of each well have

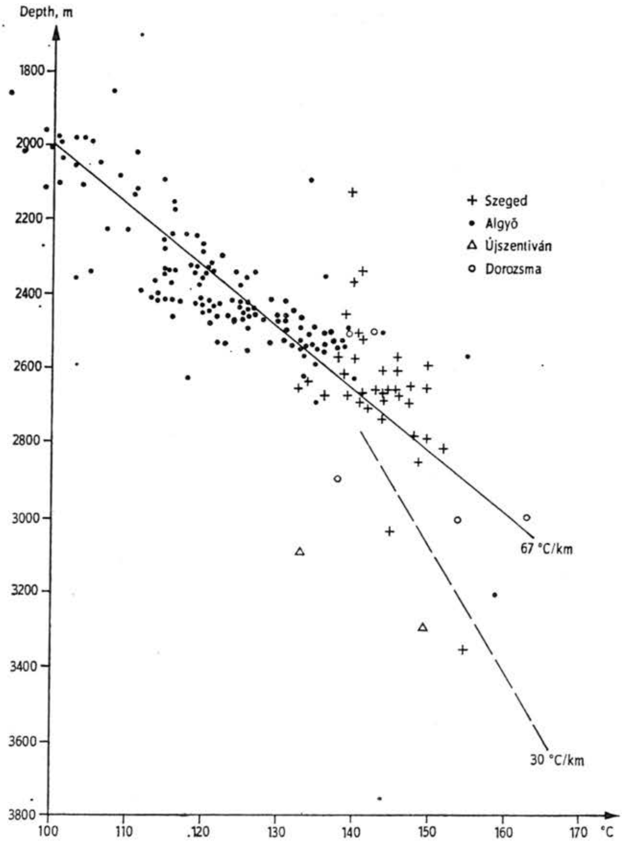
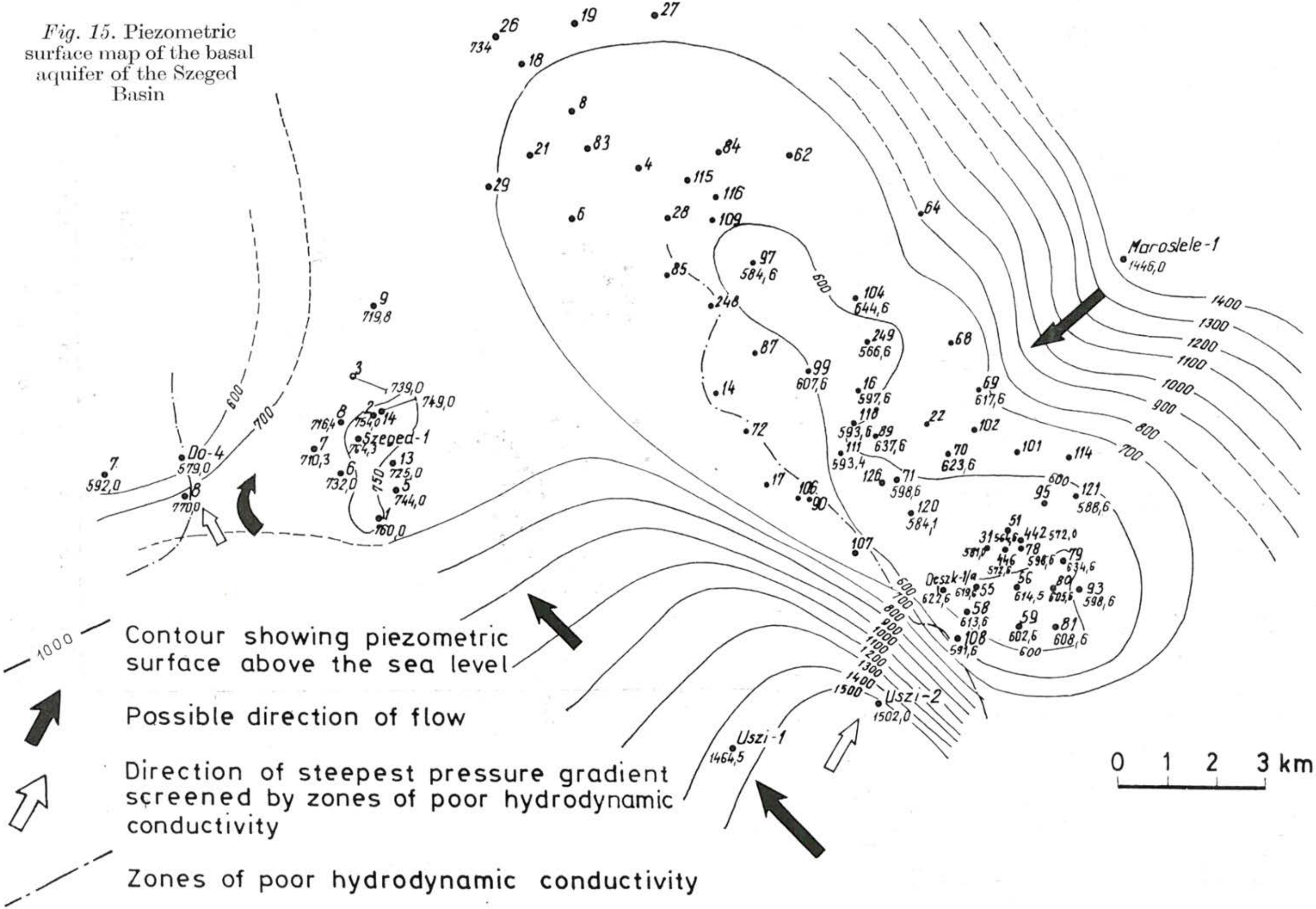


Fig. 14. Temperature depth plot of the Szeged Basin

Fig. 15. Piezometric surface map of the basal aquifer of the Szeged Basin



been determined taking the hydrocarbon-water contact of the lowest position (2965 m b.s.l., Dorozsma) for a datum level.

The piezometric surface map (Fig. 15) shows considerable highs of 1400–1500 m a.s.l. in the environs of boreholes Újszentiván and Maroslele. These areas could be considered as local sources of water flows.

As for the possible direction of these flows, we have to consider the data of the regional map showing hydrodynamic conductivity as presented in Fig. 16. It can be stated that areas of high hydrodynamic conductivity exist in the inner parts of both the Szeged and the Algyő-arch, due to the considerable thickness of highly permeable fractured dolomite and conglomerate series, respectively. At the same time extremely poor hydrodynamic conductivity values are characteristic of the western flank of Algyő and the whole Dorozsma structure as well. These practically impermeable barriers heavily influence the flow system of the aquifer by preventing any fluid movement in the direction of the steepest pressure gradients; i.e. from Újszentiván to Algyő-South and from Dorozsma towards the west. Consequently, the flow from the Újszentiván area would be directed towards NW, and a component of this could account for the northward tilting of the hydrocarbon-water contact in Szeged. The other part of this subsurface stream might be diverted towards N and NE by the impermeable barrier existing in the environs of the borehole Dorozsma-8 (Fig. 15).

Finally, on the basis of average reservoir rock parameters and pressure differences recorded as measured and evidenced by tilted hydrocarbon-water contacts, the velocities of the water flows have been estimated by the Darcy equation;

$$Q=\frac{k\cdot h\cdot w\cdot P}{\mu\cdot L}, \tag{1}$$

	Algyő	*Szeged*
where Q = flow velocity, in cm^3/sec		
k = permeability, in D	0.15	0.15
w = width of the flowing path, cm	$7\cdot10^5$	$3\cdot10^5$
h = high (= effective thickness) of the flowing path, cm	$5\cdot10^3$	$15\cdot10^3$
P = pressure differences between two selected points of the path, at	2.15	2.3
μ = viscosity of water under reservoir conditions, cP	0.3	0.3
L = length of the flow path between the two selected points mentioned above, cm	$5\cdot10^5$	$3\cdot10^5$

The results are 0.12 m/year and 0.15 m/year for Szeged and Algyő, respectively, which are highly probable in the light of the reservoir rock characteristics of the area involved.

Conclusions

The basal aquifer of the Szeged Basin is characterized by fairly well defined hydrodynamic conditions formed by pressure differences between the southern/eastern parts and other parts of the basin. The highly permeable reservoir rocks (first of all the Middle Triassic fractured, brecciated dolomite,

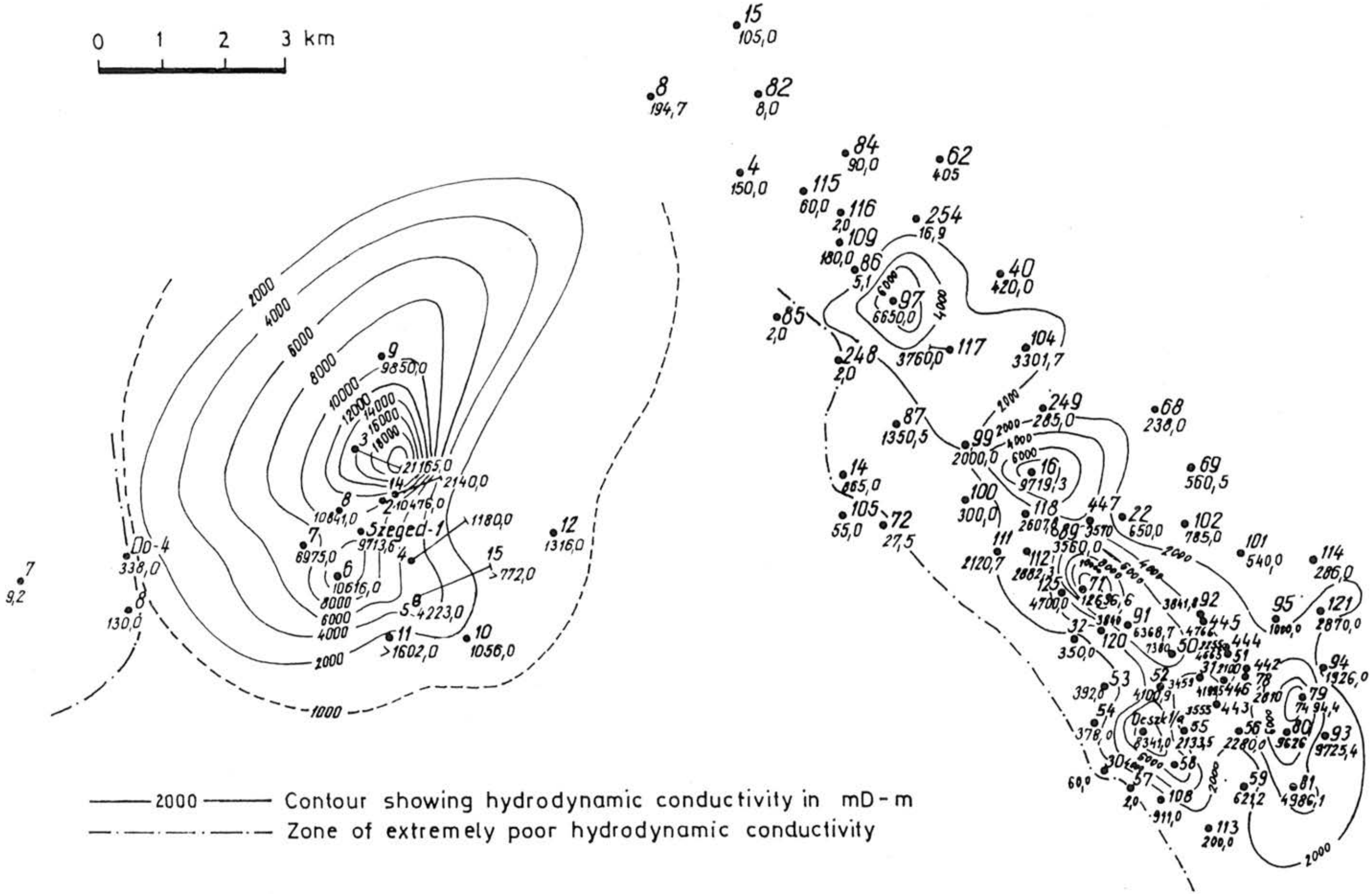


Fig. 16. Regional variation of the hydrodynamic conductivity of the basal aquifer in the Szeged Basin

and the Miocene and Lower Pannonian conglomerate series) are favourable media for water flow. The direction of flows are highly influenced by areas of poor hydrodynamic conductivity in the western flank of Algyő and in Dorozsma. The existence of this last-mentioned poor conductivity could also be evidenced by the stagnant water conditions revealed by hydrogeochemical analyses.

The rate of flow estimated from tilted hydrocarbon-water contacts and pressure differences are of 0.12 and 0.15 m/year in Szeged and in Algyő, respectively. These values are sufficient to preserve hydrodynamic conditions without affecting the hydrocarbon accumulations of the area.

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L'ANALYSE D'UN GRAND RÉSERVOIR AQUIFÈRE EN VUE DE SA MODÉLISATION

THE ANALYSIS OF A LARGE AQUIFER FOR THE PURPOSE OF BUILDING A MODEL

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ABSTRACT

In order to determine the groundwater resources of a basin, if its structure is at all complex, a digital model must be built. With the plain of Kairouan in Tunisia as an example, it is shown how a hydrogeological approach, considering the aquifer as an underground reservoir, simplifies the building of the model and its adjustment.

This aquifer is enclosed in a sink-hole of nearly 3000 km² filled with continental detrital deposits from the Plioquaternary age, of a thickness often exceeding 500 m.

The analysis contains a schematic definition of the different aquifer layers: nature, geometry and structure of the reservoir and conditions at the boundaries of the reservoir. Then, a study of the pattern of piezometric heads and of the distribution of salinity gives a picture of all the hydrological mechanisms which play a part in the underground water system.

Finally, the preliminary estimation of the water balance of the basin gives comparative values, which are useful for the building of the model.

The digital model on this basin uses the hydraulic parameters obtained from the pumping tests on the wells and the boreholes. If no tests are made, values are used which are obtained by less precise methods, but which contribute to shorten the time of fitting of the model.

Introduction

La plaine de Kairouan, limitée au Nord par l'exutoire d'El Haria, constitue par ses dimensions le plus important réservoir d'eau souterraine de Tunisie Centrale. Les nappes contenues dans ce grand bassin endoréïque, alimentées principalement par les crues des oueds Zeroud et Merguellil, circulent au sein d'une alternance de terrains sableux et argileux pour aboutir à l'aval à un système de sebkha (lacs salés).

Jusqu'au début des années 60, seule la partie septentrionale de la plaine était exploitée par des forages. L'étude hydrogéologique de l'ensemble de la plaine débuta en 1967. Depuis, 40 forages, sur les 80 existant actuellement, ont été réalisés, permettant de reconnaître un aquifère profond à bonne productivité dans la zone méridionale de la plaine.

En outre, un réseau de surveillance des nappes phréatiques et profondes a été mis en place. Ce qui a permis d'enregistrer un historique de la piézométrie sur plusieurs années, et particulièrement la perturbation causée par les crues exceptionnelles de l'automne 69.

Les données archivées à la Division des Ressources en Eau de Tunisie constituent une masse d'informations volumineuse dont l'exploitation permet la construction d'un modèle mathématique de simulation qui devra répondre aux objectifs de gestion des eaux. Mais préalablement à l'élaboration du modèle mathématique, une analyse fine des mécanismes hydrologiques souterrains est rendue nécessaire par la complexité du remplissage sédimentaire.

Dans la présente note, nous présentons cette phase préliminaire de l'étude qui doit déboucher sur l'établissement d'un véritable « modèle hydrogéologique ». L'adaptation d'un programme de calcul adéquat permet alors de passer au modèle mathématique, dont la mise au point (ou calage) se trouve d'autant plus facilitée que l'analyse préliminaire aura été menée à bien.

Nature, geométrie et structure du réservoir

Étendue sur près de 3000 km², la plaine de Kairouan est une vaste cuvette limitée à l'Ouest par une série de reliefs à formations secondaires et tertiaires se relayant du Nord au Sud.

Ainsi délimitée, la plaine alluviale est essentiellement constituée par une cuvette d'effondrement centrale comblée par des dépôts détritiques continentaux d'âge Plio-quaternaire. La sédimentation y est lenticulaire et formée d'alternances de sables plus ou moins grossiers et de marne, avec toute la gamme des terrains intermédiaires, sur une épaisseur supérieure à 500 m.

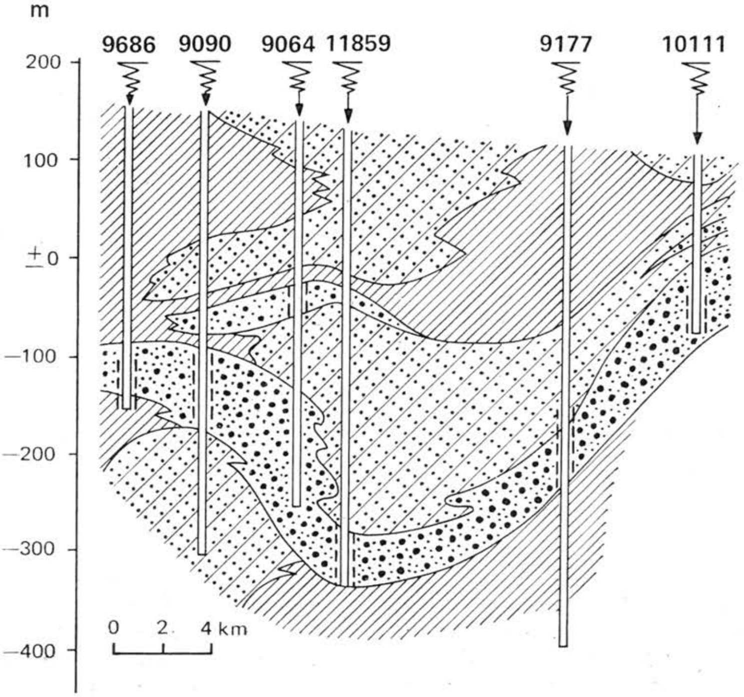

Fig. 1

Alors que la nappe phréatique, premier niveau productif et nappe des puits, présente une continuité relativement bonne sur l'ensemble du domaine, les forages réalisés, au nombre de 80, sur des profondeurs allant de 100 à 700 m, confirment le caractère lenticulaire de la sédimentation, rendant la définition d'un réservoir profond assez délicate. En effet, au niveau de Kairouan, sur un forage de 260 m, on ne dénombre pas moins de 7 horizons productifs séparés par des niveaux semi-perméables! L'absence de carottage mécanique, liée au fait que les essais de production ne sont pas réalisés d'une manière systématique, rend les corrélations entre les niveaux aquifères d'un forage à l'autre quasiment impossibles. A cet effet, le seul recours reste donc le carottage électrique, lequel est réalisé sur tous les puits. Par l'analyse simultanée des coupes du sondeur et des logs électriques, et au prix d'une simplification parfois très poussée, on arrive à individualiser en profondeur un niveau perméable unique, de 50 à 100 m d'épaisseur, inclus dans un ensemble semi-perméable, et que l'on retrouve dans toute la plaine sur les coupes litho-électriques des forages. La profondeur de ce niveau est variable, elle se situe généralement entre 100 et 400 m, ce qui le distingue nettement de l'aquifère phréatique, sauf au Nord-Ouest, au débouché du Merguellil, où il y a une absence de séparation entre les deux niveaux. Ailleurs, l'analyse litho-électrique ne suffit pas elle seule à apprécier la qualité des communications verticales entre les deux aquifères (Fig. 1).

Fonctionnement hydraulique du réservoir

Alimentation des nappes

Les parties amont des oueds Zeroud et Merguellil apparaissent comme des zones d'alimentation essentielles de la nappe phréatique. En effet, les débits d'étiage de ces deux oueds s'infiltrent entièrement à l'entrée de la plaine. De plus, la mesure des hauteurs piézométriques, réalisée sur plusieurs années, a mis en évidence une participation importante des crues à l'alimentation des nappes.

Si l'on peut estimer faible l'infiltration efficace de la pluie à la surface du sol de la plaine, une multitude de ravins, parfois importants, concentrent un ruissellement non négligeable aux piedmonts des reliefs limitrophes, contribuant pour une part relativement importante à l'alimentation des nappes.

Enfin, si l'on peut considérer comme étanche l'ensemble des limites géologiques amont de la plaine, un apport souterrain en provenance du bassin hydrogéologique amont du Merguellil (aquifères de Haffouz et Aïn Beïda) devrait transiter par le seuil d'El Haouareb qui en constitue l'unique exutoire.

Exutoires naturels

La reprise par évaporation constitue l'unique exutoire naturel des eaux souterraines du bassin. Cette évaporation s'effectue soit à l'aval de Kairouan où la nappe phréatique est subaffleurante, soit dans les sebkha de bordure drainant également le ruissellement de crue des oueds les plus importants. L'écoulement des eaux vers l'aval est sollicité par deux systèmes de drainage

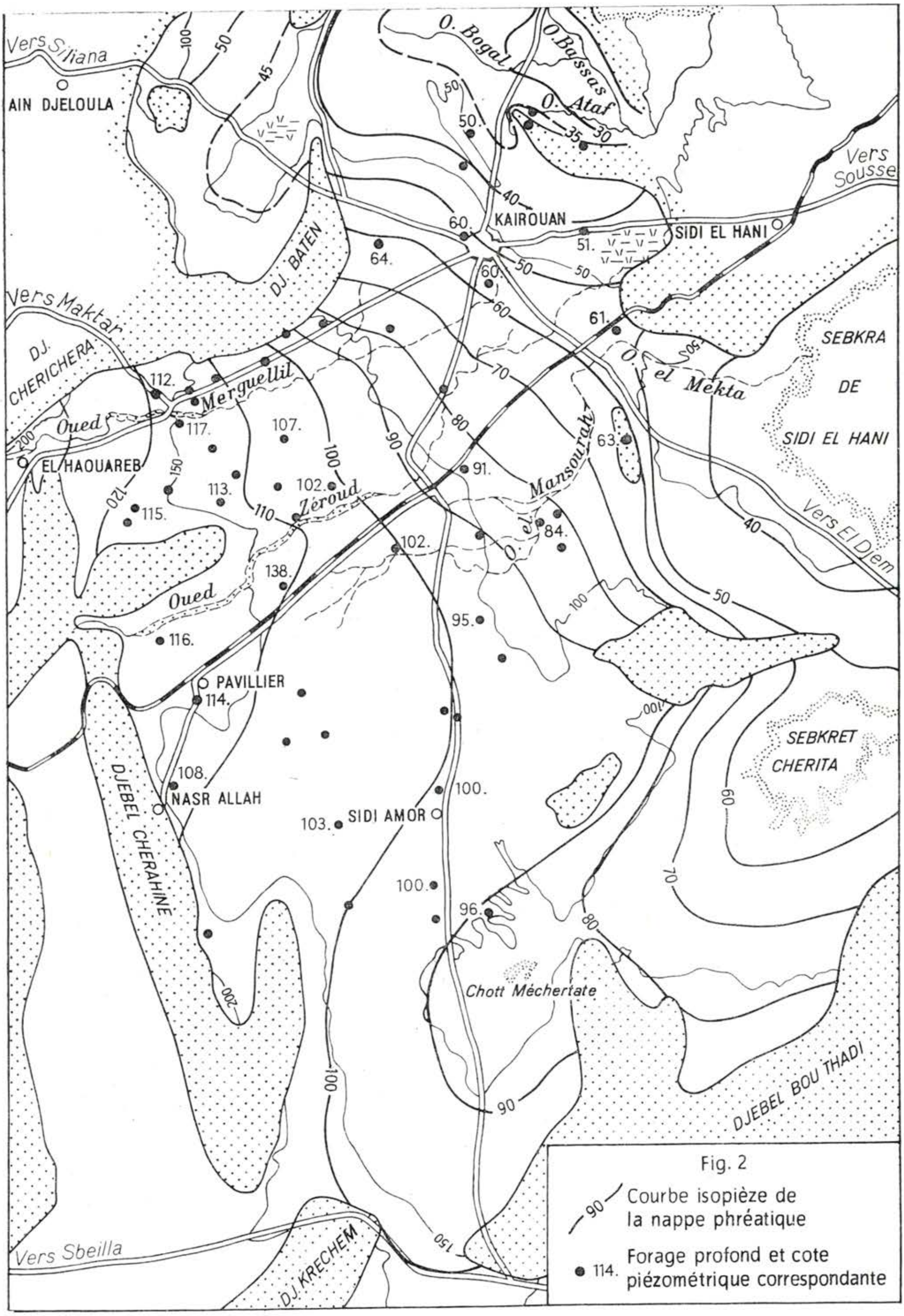

Fig. 2

Courbe isopièze de la nappe phréatique

Forage profond et cote piézométrique correspondante indépendants, organisés sous forme de sebkha en relais et séparés par les collines de Sidi El Hani:

— Exutoire Nord: El Haria → Metbasseta → Kelbia.

— Exutoire Est: Mechertate → Cherita → Sidi El Hani.

Les sebkha aval du bassin constituent donc les deux niveaux de base des écoulements, à la côte +30 m pour la sebkha Sidi El Hani, et +20 m pour la sebkha Kelbia. C'est cette dernière qui constitue sans doute l'exutoire le plus important des nappes de la plaine de Kairouan. En effet, en plus de la différence des niveaux de base, si l'écoulement souterrain vers le Nord s'effectue « librement », la circulation des nappes vers l'Est est frainée par la présence d'un chapelet de « verrous » litho-morphologiques que sont les collines Plio-quaternaires orientales.

Percolations verticales

La piézométrie rend également compte des relations hydrauliques existant entre les aquifères phréatiques et profonds. Il semble clair que dans la zone amont, la nappe profonde soit alimentée par l'horizon supérieur, mais la communication semble bonne puisque la perte de charge entre les deux niveaux atteint rarement 5 m. Par contre, à l'Est et au Nord, après un passage par une zone de transition et d'équilibre des niveaux, l'aquifère profond se met en charge et percole dans la nappe phréatique. La perte de charge entre les deux horizons dépasse 10 m au Nord de la ville de Kairouan, zone d'artésianisme des niveaux profonds, lesquels seraient donc, là, mieux isolés de la nappe phréatique.

La salinité: un traceur hydrologique naturel

A) L'étude de la qualité des eaux confirme les mécanismes hydrogéologiques que nous avons décrits :

— Eau douce (salinité inférieure à 1.5 g/l) dans la zone d'alimentation par l'oued Merguellil en relation avec la salinité des eaux de ce dernier.

— Eau à salinité moyenne (1.5 à 3 g/l) dans la zone d'alimentation par les crues du Zeroud également en relation avec la salinité de cet oued, tandis qu'au débouché de ce cours d'eau dans la plaine, une poche de salinité plus élevée rend compte de l'infiltration des débits d'étiage à forte concentration.

— Gradient de salinité amont-aval beaucoup plus élevé dans la nappe phréatique (où l'on passe de 1 à plus de 7 g/l) que dans la nappe profonde (gradient très faible : 1.5 à 3 g/l), conforme aux gradients piézométriques en rendant bien compte du sens et de la qualité des échanges verticaux entre les deux nappes : alors qu'à l'amont, la bonne liaison permet une quasi-conservation de la qualité des eaux dans leur percolation de haut en bas, à l'aval, la séparation par des couches très peu perméables provoque un gradient vertical de salinité élevé.

— Enfin, une zone à forte salinité (3 à 5 g/l) au Nord du Drâa Affane est peut-être due à l'infiltration des eaux de ruissellement au piedmont de cette colline formée de sédiments gypseux (Fig. 3).

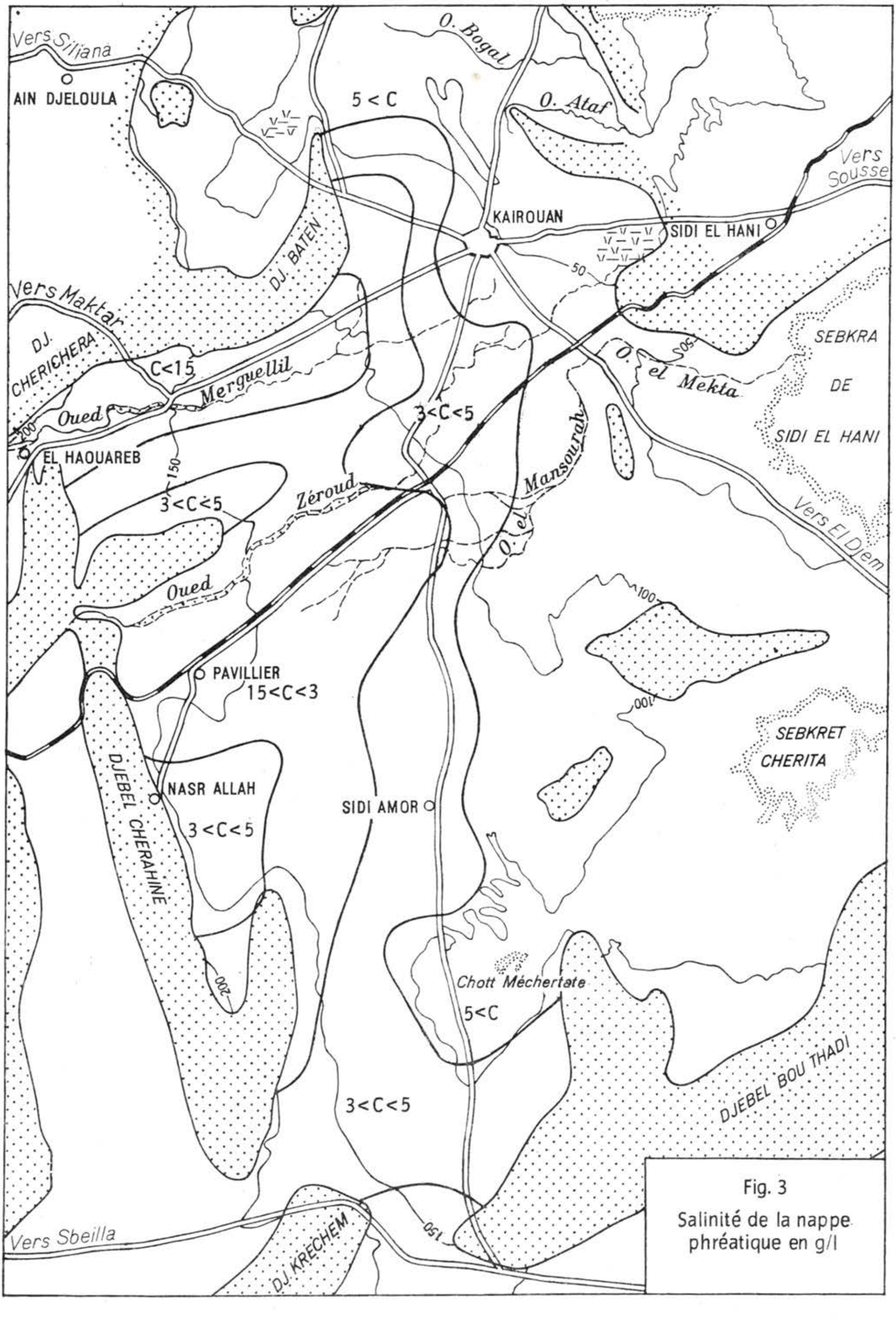

Fig. 3
Salinité de la nappe phréatique en g/l

B) La concentration des eaux de la nappe phréatique vers l'aval est due à plusieurs facteurs :

— Au Sud du Zeroud, le réservoir superficiel devient très peu perméable à l'aval du méridien de Kairouan, entraînant un contact prolongé avec la rochemagasin argileuse. De plus, les zones à plus fortes salinités correspondent à celles où la profondeur de la surface de la nappe devient très faible, favorisant l'évaporation.

— Au Nord du Zeroud, le critère de perméabilité intervient moins, et c'est l'évaporation qui semble être la cause principale de la salinité des eaux. Cependant, dans les secteurs où la profondeur de la surface de la nappe interdit toute reprise par évaporation, l'explication réside peut-être dans le fait suivant, sous réserve d'une analyse plus approfondie du phénomène : la seule anomalie dans l'évolution d'amont en aval des salinités coïncide précisément non seulement avec une zone de forte concentration des prélèvements dans la nappe phréatique (bled Abida), mais aussi avec la moitié aval du lit du Zeroud où continuent de s'infiltrer les crues importantes et moyennes. Cette remontée vers l'amont de la salinité des eaux phréatiques ne saurait pourtant être imputée à ces apports de crue puisque sous la moitié Ouest du lit, les eaux souterraines sont plus douces. L'origine de la contamination proviendrait donc de l'irrigation avec l'eau relativement salée de la nappe phréatique : il y aurait dépôt de sel par évapotranspiration des plantes au niveau du sol, puis par infiltration des eaux de crues, retour de ce sel à la nappe et aggravation de la concentration. Mais une autre cause de cette anomalie n'est pas à exclure : cette zone correspond également à la confluence du Zeroud avec l'oued Melaïeh, un « oued de plaine » qui draîne un drâa Affane gypseux. Une étude géochimique précise de ce phénomène devrait pouvoir résoudre ce dilemme, car s'il était établi que l'oued Melaïeh n'était pour rien dans cette anomalie et que cette dernière avait son origine dans une contamination à caractère évolutif, les conséquences sur la gestion de la nappe phréatique pour l'irrigation en seraient d'une grande importance.

Essais hydrodynamiques

Essais de longue durée aux forages

Sur les 60 forages productifs de la plaine, 16 ont fait l'objet d'un essai de pompage de longue durée (3 à 6 jours) à débit constant, avec observation d'un ou plusieurs puits témoins.

Dans la majorité des cas, il a pu être observé l'amorce plus ou moins nette d'un palier de stabilisation dû au débit retardé en provenance de la nappe phréatique.

Du fait de la structure lenticulaire du réservoir, la transmissivité obtenue sur un forage pompé est en général inférieure à celle calculée au même point lorsqu'il joue le rôle de puits-témoin au cours d'un essai sur un forage voisin. Il suffit en effet d'une différence de complétion entre les deux forages ou de la discontinuité du niveau capté d'un puits à l'autre pour que le débit dont participie l'horizon capté par le puits-témoin soit inférieur au débit pompé. Il en résulte une surestimation systématique de la valeur de T calculée aux piézométres. La pénétration partielle peut aussi expliquer ce genre d'anomalie :

les forages considérés étant souvent éloignés les uns des autres, l'effet de pénétration partielle se trouverait éliminé au puits-témoin qui offrirait alors une transmissivité plus représentative, la valeur de T au forage pompé ne correspondant plus qu'au niveau crépiné.

Les réactions aux puits-témoins n'étant pas interprétables dans tous les cas, nous avons choisi de ne retenir que les valeurs de T au forage pompé, solution pessimiste certes, mais offrant une plus grande homogénéïté quant au critère de ce choix. En certains points, les paramètres de drainance ont pu être calculés et serviront à l'ajustement des valeurs de transmissivité de passage entre nappe profonde et nappe phréatique.

Essais de courte durée à plusieurs paliers de débit

Étant donné la difficulté d'interprétation des essais de longue durée, liée à la grande anisotropie du réservoir, et l'étendue de l'aquifère non couverte par ces essais, les valeurs de T sur 45 forages ont dû être déduites des essais de réception des puits ou essais du sondeur.

Une analyse de ces essais, réalisés à plusieurs paliers de débits, permet en effet une évaluation de la transmissivité aux forages [3]. La répartition des valeurs à introduire dans le modèle s'en trouve renforcée et l'ajustement de ce dernier facilité, même si au cours du calage certaines anomalies devaient être rectifiées.

Essais sur la nappe phréatique

A proximité de 7 piézomètres répartis sur la zone d'étude, des essais de pompage de 3 à 4 jours ont été réalisés dans des forages peu profonds captant rigoureusement le même niveau que le piézomètre correspondant. Les résultats obtenus montrent des caractéristiques de nappe libre sur 3 des piézomètres. Sur les autres, une amorce de stabilisation et une reprise du rabattement en cours d'essais traduit une mise en charge purement locale.

La couverture par ces rares essais hydrodynamiques reste cependant très lâche, et ne suffit pas à représenter l'ensemble de l'aquifère. Pour pallier cet inconvénient, nous avons entrepris d'évaluer les transmissivités à l'aide d'une approche différente, moins élaborée, mais permettant une estimation de T en chacun des 900 puits en production. Nous n'aurons besoin, pour l'étude sur modèle, que d'une valeur unique par maille et nous nous intéresserons à la valeur moyenne pour chacune des mailles de la nappe phréatique.

La démarche entreprise repose sur deux hypothèses : on suppose que chacun des exploitants agricoles s'efforce d'utiliser toute la quantité d'eau que son puits est en mesure de fournir, et que par ailleurs, le diamètre et la tranche d'eau disponible dans tous les puits sont à peu près équivalents. Si la première hypothèse peut être aisément admise, la seconde par contre n'est certainement pas vérifiée si l'on considère les puits individuellement, essentiellement pour ce qui est de la tranche d'eau, le diamètre intervenant d'ailleurs beaucoup moins dans le débit.

En revanche, si l'on considère des groupes de puits, comme c'est le cas puisque nous nous intéressons à des moyennes par maille et que les mailles du modèle comprennent le plus souvent plusieurs puits chacune, les valeurs erratiques joueront beaucoup moins. De plus, lorsque l'on étudie l'ensemble du domaine, les discontinuités locales dues à ces valeurs extrêmes si elles persistent (puits unique dans une maille) seraient pondérées sur le modèle par la méthode de composition des transmissivités utilisée dans les calculs : dans le programme utilisé, on calcule une moyenne arithmétique pour composer les transmissivités de deux mailles voisines, ce qui a pour effet d'atténuer des contrastes de ce genre ne présentant pas un caractère régional.

On obtiendra donc, pour chacune des mailles du modèle à construire, la capacité moyenne de production de l'aquifère, égale au débit annuel prélevé dans la maille, rapporté au nombre de puits dans la maille produisant ce débit. La capacité des mailles où il n'existe pas de puits sera obtenue par interpolation entre les points connus, ou par extrapolation vers les limites.

Apports aux nappes et estimation du bilan en eau

Une évaluation précise des apports et des pertes aux exutoires du bassin devrait résulter de l'ajustement du modèle mathématique. Il est bon toutefois, étant donné le nombre de paramètres que l'on sera amené à ajuster par approximations successives, de se donner au préalable un ordre de grandeur de ces débits d'échange du système avec l'extérieur.

Infiltration des crues d'oueds

Dans le lit du Zeroud, l'infiltration des crues a été évaluée [5] par l'analyse graphique sommaire des hydrogrammes aux piézomètres et puits d'observation de la nappe phréatique. Cette analyse porte sur deux années dont les caractéristiques hydrologiques sont les suivantes :

	1967–68	1968–69
ruissellement total du Zeroud :	$135 \cdot 10^6$ m^3	$68 \cdot 10^6$ m^3
infiltration évaluée des crues :	$40 \cdot 10^6$ m^3	$2 \cdot 10^6$ m^3
pluie à Kairouan :	438 mm	180 mm

– moyenne des pluies annuelles à Kairouan ≈ 300 mm,
– moyenne des apports en ruissellement du Zeroud $\approx 94 \cdot 10^6$ m^3/an.

Malgré la grande variabilité d'une année à la suivante, la moyenne des deux années étudiées serait donc sensiblement proche de l'année hydrologique moyenne, ce qui porterait l'apport moyen des crues du Zeroud à environ $21 \cdot 10^6$ m^3/an infiltrés en première analyse.

La même méthode d'approche sommaire a permis d'évaluer l'infiltration dans le lit du Zeroud, consécutive aux crues exceptionnelles de l'automne 1969, à $100 \cdot 10^6$ m^3, tandis que dans le lit du Merguellil, le volume infiltré correspondant au même épisode de crue s'élèverait à $60 \cdot 10^6$ m^3, soit un rapport de 0.6 entre les deux oueds, alors que les superficies des bassins-versants à l'entrée de la plaine sont dans un rapport bien plus faible : 100 km^2 à El Haouareb et 8950 km^2 à Sidi Sâad, soit un rapport de 0.125.

Mais ce paradoxe pourrait n'être qu'apparent car, bien que l'on ne possède aucune donnée précise sur le ruissellement hors étiage à El Haouareb (Merguellil) alors que plus de 20 années de mesures sont enregistrées à Sidi Sâad

(Zeroud), la morphologie et la structure lithologique des deux bassins-versants militent en faveur d'un coefficient de ruissellement plus élevé sur le Merguellil. D'ailleurs, l'ensemble des estimations d'apport effectuées donnent un module compris entre 18 et $32 \cdot 10^6$ m^3 à El Haouareb, soit un rapport au Zeroud compris entre 0.2 et 0.33. De plus, la morphologie des lits dans la plaine et la fréquence des épisodes de crue sont comparables pour les deux oueds. Si l'on considère que dans l'infiltration d'une lame d'eau libre interviennent surtout la surface et le temps de contact avec le sol (sensiblement comparables pour les deux oueds), la charge intervenant beaucoup moins, le débit de percolation après saturation étant régi par la perméabilité du terrain que l'on n'a pas de raison de ne pas considérer équivalente pour les deux oueds, on voit que le rapport de 0.6 entre les volumes infiltrés dans les deux oueds ne constitue pas une aberration.

En se basant sur cette approximation et en appliquant ce coefficient aux apports aux nappes des crues du Merguellil en année moyenne, on peut donc estimer ces apports à $21.0 \cdot 0.6$, soit 12.6 millions de m^3/an pour l'ensemble du bassin-versant.

Infiltration du débit de base

L'apport en débit de base moyen du Zeroud à Sidi Sâad est de 12 millions de m^3/an. En aval de la station hydrométrique, une partie de ce débit était prélevé par deux sequia importantes servant à l'irrigation de deux unités agricoles. La valeur moyenne de ce prélèvement, résultat de huit campagnes de jaugeages différentiels effectuées durant les années 50, est estimée à près de 6 millions de m^3/an. Mais, en raison du creusement important du lit de l'oued, ces sequia n'étaient plus en service après les crues de l'automne 69.

A l'aval des anciens points de prélèvement, et après un parcours de plus de 10 km dans la plaine de Sidi Sâad où le drainage de la nappe phréatique locale semble compenser la reprise sur l'oued par évaporation, le débit de base va s'infiltrer à la surface du sol à l'entrée de la plaine de Kairouan. Les apports moyens de base du Zeroud, infiltrés en surface, peuvent donc être évalués à 12 millions de m^3/an pour la période actuelle et à 6 millions de m^3/an pour la période antérieure aux crues de 69.

Sur le Merguellil, la station hydrométrique de Haffouz ne contrôle que la moitié du bassin-versant. Une série de 44 jaugeages effectués pendant les années 1970 à 72 indiquent un rapport moyen égal à 0.6 entre le débit de base à Haffouz et celui mesuré à l'aval (Sidi Bou Jdaria). Or, sur les quatre dernières années (1970 à 74), on a enregistré à Haffouz un apport de base moyen égal à 5 millions de m^3/an. Ce qui donnerait près de 8 millions à l'aval en appliquant le coefficient de 0.6. Si l'on tient compte d'un prélèvement de 1 million de m^3/an pour l'irrigation, il resterait donc près de 7 millions de m^3/an qui iraient s'infiltrer en surface à l'entrée de la plaine de Kairouan.

Contrairement aux débits de crue où l'on a évalué les quantités arrivant à la nappe, les estimations que nous venons d'effectuer concernent la disparition en surface du débit de base: il s'agit donc d'une infiltration apparente. En dehors des nappes de la plaine de Kairouan, il n'existe pas d'autre structure hydrogéologique d'accueil pour cette infiltration, mais l'évaporation dans les premiers mètres du sol peut en retenir une partie que nous ne sommes pas en mesure d'évaluer actuellement.

Infiltration directe de la pluie

Le bilan hydrique global de la plaine indique un déficit important à l'échelle annuelle et même mensuelle. En effet, alors que la pluviométrie moyenne annuelle à Kairouan est de 300 mm, l'évapotranspiration potentielle moyenne calculée par différentes formules varie de 1000 à 1800 mm/an. Seules les averses de forte intensité arrivent à combler le déficit de saturation et parviennent à ruisseler ou s'infiltrer sur la plaine. L'étude comparative du bilan en eau d'un certain nombre d'aquifères de Tunisie Centrale indique des coefficients d'infiltration efficace de la pluie compris en 2 et 4% sur les plaines alluviales.

Les terrains quaternaires de la plaine de Kairouan offrent des conditions de perméabilité de surface peu favorables. Nous admettons, en première hypothèse, un coefficient d'infiltration égal à 2% en dehors des zones à salinité élevée, soit une superficie utile d'un millier de km² et un apport d'environ 6 millions de m³/an. Mais il est bien évident que, dans ce genre de calcul, l'incertitude qui demeure est du même ordre de grandeur que la valeur à estimer.

Autres éléments naturels du bilan

Ce sont les termes dont on ne peut avancer une estimation chiffrée aussi imprécise soit-elle. Il s'agit des infiltrations aux piedmonts des reliefs de bordure et des apports souterrains provenant essentiellement par le seuil d'El Haouareb, mais surtout des exutoires naturels par les différentes sebkha limitrophes et par évaporation sur la nappe phréatique, là où elle se rapproche de la surface du sol.

Prélèvements dans les puits et forages

L'inventaire des points d'eau et l'enquête sur les débits extraits qui se sont déroulés sur une période d'environ cinq années (1968 à 73) ont permis de cataloguer 1300 puits dans la nappe phréatique. 920 puits, dont 500 équipés de pompes, sont en production. L'ensemble de ces puits prélève 22.2 millions de m³/an en moyenne. Bien que cette valeur dépende de plusieurs facteurs d'imprécision, tels que la pluviosité de l'année en cours, mais aussi de la bonne volonté de l'utilisateur interrogé à l'enquête, nous nous efforcerons de la considérer représentative d'une réalité moyenne dans la suite de l'étude. Il semble en effet difficile d'en estimer la variabilité dans le temps.

Par ailleurs, il existe actuellement 60 forages d'eau exploitables dans la plaine. La première enquête systématique sur la production de ces forages indique un prélèvement global de 10 millions de m³ en 1969 et 11.5 millions de m³ en 1970.

Bilan en eau du bassin

Le tableau récapitule l'ensemble des estimations précédentes et résume l'état de nos connaissances actuelles:

Entrées (millions de m³/an):

infiltration des crues Zeroud	=21
infiltration des crues Merguellil	=12.6
infiltration débit de base Zeroud	=12
infiltration débit de base Merguellil	=7
infiltration directe pluie	=6?
— id — aux piedmonts de bordure	=?
apports souterrains	=?

Sorties (millions de m³/an):

prélèvements nappe phréatique	=22.2
prélèvements nappe profonde	=11.5
exutoires naturels	=?

Les cases vides et les valeurs incertaines devront être précisées par le modèle mathématique, mais les estimations effectuées permettront d'entamer le calage avec des ordres de grandeur réalistes.

Conclusion

Dans un bassin sédimentaire complexe, la construction d'un modèle mathématique passe nécessairement par une analyse aussi fine que possible des réservoirs à simuler. Pour cela, on s'efforce généralement d'individualiser les différents réservoirs du bassin en définissant leur disposition les uns par rapport aux autres, leurs conditions aux limites et leurs caractéristiques hydrauliques, ainsi qu'une estimation des débits d'alimentation ou d'exhaure.

Il n'existe pas cependant de méthodologie précise en la matière et une certaine forme de pragmatisme peut parfois être souhaitable. Dans le cas de la plaine de Kairouan, résumé par l'organigramme précédent, nous avons ainsi établi une première carte de transmissivités de la nappe profonde à partir des résultats d'essais de courte durée, et celle de la nappe phréatique au vu des simples prélèvements annuels. Dans d'autres bassins, d'autres possibilités peuvent se présenter, telle la classification lithologique des coupes de forages [1] par exemple. Dans l'exemple étudié, nous avons pu différencier deux réservoirs au sein des épaisses formations sablo-argileuses de la plaine et l'architecture de l'aquifère profond a été mise en évidence grâce à la géophysique. Les conditions aux limites et les relations entre les deux réservoirs ont été définies par l'examen de la répartition des pressions, la géochimie et les essais hydrodynamiques. Enfin, une évaluation des flux d'alimentation a pu être effectuée par des procédés analytiques de calcul [5].

Cette phase préliminaire d'analyse des réservoirs aboutit à la conception d'un modèle hydrogéologique dont l'adaptation sur un programme numérique à structure multicouche permet un ajustement rapide. Dans le cas de la plaine de Kairouan, malgré l'ètendue et la complexité du bassin et la finesse du maillage adopté (maillage variable avec 1 km de côté pour les petites mailles, soit un total de 2000 mailles), il est ainsi suffit de 15 passages sur ordinateur pour que le modèle soit entièrement ajusté en régime permanent avec une précision suffisante.

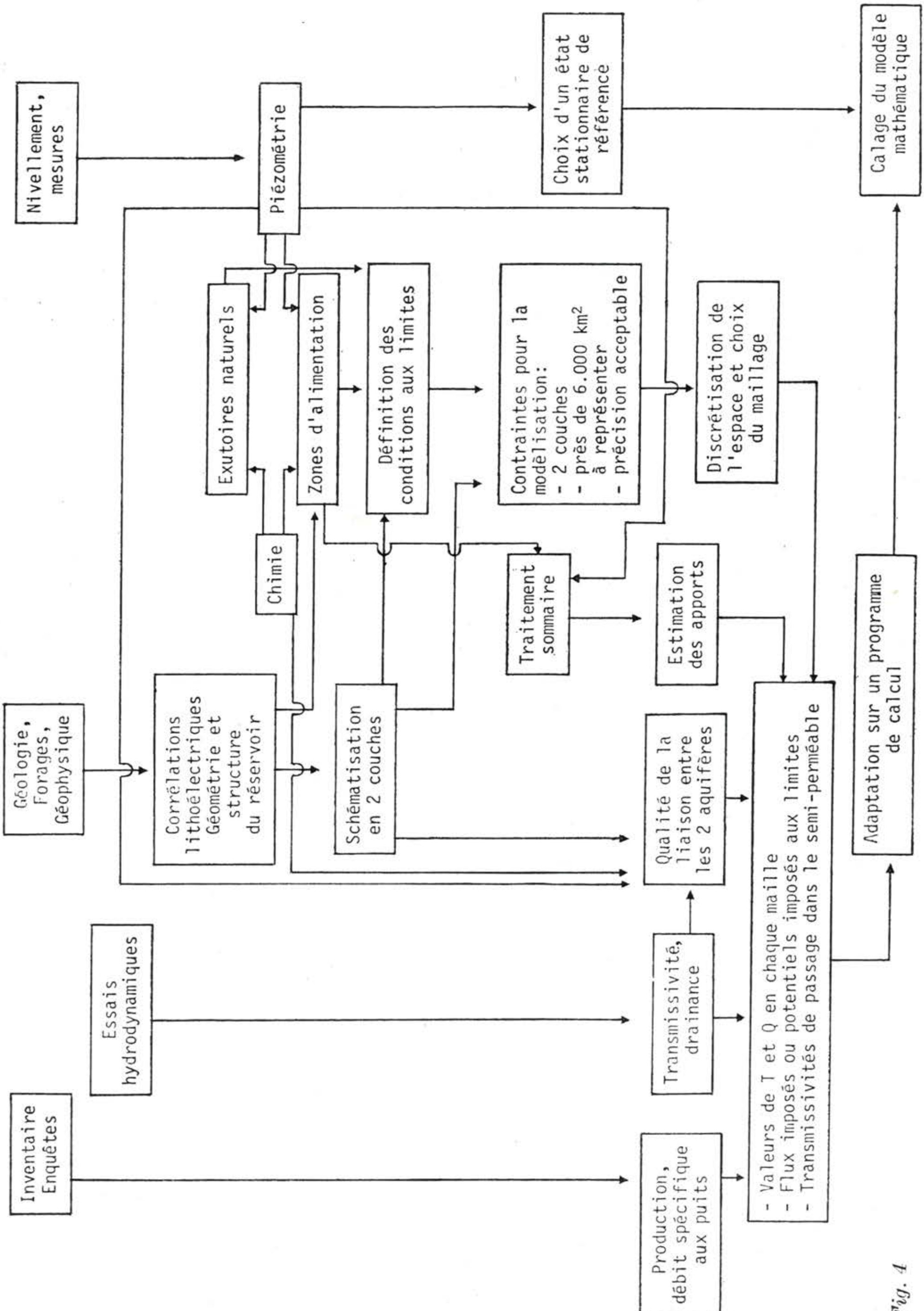
Nivellement, mesures
Piézométrie
Choix d'un état stationnaire de référence
Calage du modèle mathématique
Exutoires naturels
Zones d'alimentation
Définition des conditions aux limites
Contraintes pour la modélisation:
- 2 couches
- près de 6.000 km²
à représenter
- précision acceptable
Discrétisation de l'espace et choix du maillage
Chimie
Traitement sommaire
Estimation des apports
Géologie, Forages, Géophysique
Corrélations lithoélectriques Géométrie et structure du réservoir
Schématisation en 2 couches
Qualité de la liaison entre les 2 aquifères
- Valeurs de T et Q en chaque maille
- Flux imposés ou potentiels imposés aux limites
- Transmissivités de passage dans le semi-perméable
Adaptation sur un programme de calcul
Essais hydrodynamiques
Transmissivité, drainance
Inventaire Enquêtes
Production, débit spécifique aux puits

Fig. 4

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THE DETERMINATION OF PERMEABILITY RATES OF AQUIFER-SEPARATING LAYERS FROM NON-HYDROLOGICAL DATA

LA DÉTERMINATION DE L'ORDRE DE GRANDEUR DE LA PERMÉABILITÉ DE COUCHES SEMI-PERMÉABLES INTERCALÉES ENTRE LES COUCHES AQUIFÈRES A BASE DE DONNÉES NON-HYDROLOGIQUES

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RÉSUMÉ

Dans de grands bassins sédimentaires, il se trouve fréquemment plusieurs couches aquifères situées les unes au-dessus des autres tout en étant séparées par des couches semi-perméables. Afin de disposer de bases du calcul de modèle, par exemple, sous le rapport de la recharge verticale d'eaux souterraines vers des couches aquifères plus profondes, il faut avoir recours à des connaissances de l'ordre de grandeur relatif à la perméabilité verticale et, de cette façon, aussi de la drainance s'appliquant à de telles couches semi-perméables intercalées. Des valeurs obtenues sous cet aspect revêtent une importance pratique considérable, s'il s'agit d'évaluer la régénération des eaux souterraines dans une couche aquifère plus profonde utilisée pour le captage d'eaux souterraines.

Dans une aire d'environ 1000 km² du pays plat de l'Allemagne du NW dont la structure accuse des sédiments non consolidés du Tertiaire et du Quaternaire, et où, en règle générale, se trouvent formées deux couches aquifères situées les unes au-dessus des au tres, des valeurs caractéristiques relatives aux couches semi-perméables ont été dérivées en se fondant sur les coupes de sondage et sur les carottages (diagraphies de résistivité et des rayons gamma) de forages d'exploration. Afin de permettre des comparaisons mutuelles, ces valeurs caractéristiques sont définies comme le rapport de deux distances de colonne de coupe qui contiennent de l'argile « pure » ou, autrement, du sable grossier graveleux et qui servent de grandeurs de référence. Pour chaque sondage, ces valeurs caractéristiques sont déterminées sur la base des diagraphies de résistivité (K_R) et des rayons gamma (K_G).

Une investigation statistique, qui fut réalisée sur les valeurs caractéristiques K_R et K_G sur la base des déterminations d'environ 300 sondages, accuse les différences directionelles, comme supposé, existant en fonction de la granulométrie. De cette façon, il paraît justifié de réaliser un classement de ces valeurs caractéristiques en relation avec les valeurs du coefficient de perméabilité (k_f). En sélectionnant comme des points fixes un sable fin caractérisé par un coefficient de perméabilité autour de 10^{-4} m/s aussi bien qu'une argile caractérisée par un coefficient de perméabilité autour de 10^{-10} m/s, on obtient pour résultat, sur la base d'une subdivision uniforme, des coefficients moyens de perméabilité (k_f) suivants:

pour une argile limoneuse environ	$9 \cdot 10^{-10}$ m/s
pour des limons argileux environ	$1.5 \cdot 10^{-8}$ m/s
et pour des limons sableux fins environ	$1 \cdot 10^{-6}$ m/s

Introduction

The Northwest German Lowland is part of a large sedimentary basin filled with unconsolidated Tertiary and Quaternary deposits. In this area there are mostly two but, locally, up to four superimposed aquifers. The aquifers are separated by slightly permeable layers of clay, silt, and glacial till of varying thickness. For model calculations, such as the calculation of the rate of vertical movement of groundwater into deeper aquifers, it is necessary to know the order of magnitude of the vertical permeability and thereby of the leakage coefficient (or hydraulic resistivity) of the semipermeable beds separating the aquifers. Information about these values is of considerable practical relevance in estimating the regeneration of the groundwater in a deeper aquifer from which water is to be produced.

In principle it is possible to calculate the vertical permeability of the separating layers from data obtained by pumping tests. For a model of a certain area, however, the number and density of pumping tests are usually not sufficient because of the varying nature of the separating layers. For this reason an attempt has been made to determine the permeability from borehole data consisting of lithologic, resistivity and gamma logs.

With special consideration of the regional geology, an area of about 1000 km^2 approximately 30 km south of Hamburg was selected for study. The methods, developed and the interpretation of the data obtained are presented and discussed below.

Methods to determine parameters of aquifer-separating layers from well logs

Until now, in groundwater exploration, resistivity and gamma-ray logs of boreholes in freshwater-bearing unconsolidated sediments have only been used for qualitative correction of the lithologic logs based on ditch samples and not cores. Slightly permeable rocks like clay, silt, and glacial till are known to have low resistivity and rather high gamma radiation whereas sand and gravel have a comparatively high resistivity and low gamma radiation. Consequently, both the resistivity and the gamma radiation of a rock often indicate the type and permeability of the rock.

The geophysical well-log data for specific strata can be meaningfully compared only if the conditions in the boreholes are very similar during logging. Only resistivity and gamma-ray logs measured in 120-mm-diameter holes drilled, with clear water as drilling fluid, into freshwater-bearing unconsolidated sediments were used for the present study. Distorted results due to a varying hole diameter were generally minor because the fine-grained sediments are cohesive and there was little caving of the drillhole walls.

Resistivity and the parameter K_R

Rewiew of numerous resistivity logs of boreholes in the study area showed

a) as expected, fine-grained sediments, such as clayey silt and silty clay, can easily be numerically distinguished on the basis of their resistivities;

b) the resistivities of very similar sediments commonly differ considerably from one borehole to the other.

It is apparent that for *(a)* the conditons for logging of a single borehole were almost constant and that for *(b)* the differences were due to either well-to-well variations, varying water conditions or instrumental effects. A direct comparison of resistivities measured in neighbouring boreholes is often not possible. Therefore, the dimensionless parameter K_R (defined below) is calculated for every layer of interest so that direct comparisons are possible.

Fig. 1 shows the lithologic section of a borehole together with the resistivity log (RES, short normal) and the log of natural gamma radiation (GRL). To save space, an uninteresting section between 68 and 160 m below sea level has been left out. The lithologic section and the water levels show the presence of an upper, unconfined aquifer consisting of Quaternary glaciofluvial sand and of a lower, confined aquifer formed of the "lignite sand" of Miocene age. The aquifer-separating layers are glacial till.

The resistivity log (Fig. 1) shows an average resistivity of $x = 60\ \Omega\text{m}$ for the aquifer-separating glacial till. The resistivity ($a = 32\ \Omega\text{m}$) is lower only in the clay underlying the lower aquifer. The highest average resistivity ($b = 244\ \Omega\text{m}$) was encountered in the gravelly coarse-grained sand of the lower aquifer.

The dimensionless parameter K_R was defined as the difference between the resistivities of the aquifer-separating layer (x) and "pure" clay (a) related to the difference between the resistivities of gravelly coarse-grained sand (b) and pure clay (a) :

$$K_R = \frac{x-a}{b-a}\ .$$

Using the above-mentioned values, K_R equals 0.13.

All boreholes studied are more than 150 m deep and all penetrated similar coarse-grained sand containing fine gravel in the lower aquifer and almost-pure clay in the underlying rock. Assuming that these rocks have almost uniform natures in the study area, the above-mentioned equation will yield parameters K_R suitable to characterize the aquifer-separating layer encountered in the various boreholes. These K_R-values can then be used to compare the separating layer in one well with that in another. In order to determine K_R, the values for a and b in addition to the value for x have to be individually calculated from the resistivity curve for each well.

Gamma-ray-log and parameter K_G

As is well-known, gamma-ray-logging means measuring the natural γ-radiation of the rock. The number of counts per second is registered in a measuring curve. In sedimentary, fine-grained clastic rock sequences, the γ-radiation generally increases with increasing clay content.

Because the measured gamma-ray intensities are proportional to various factors, a numerical comparison is only possible if the values are taken relative to a defined standard. Consequently, for the gamma-ray-logs, the dimensionless

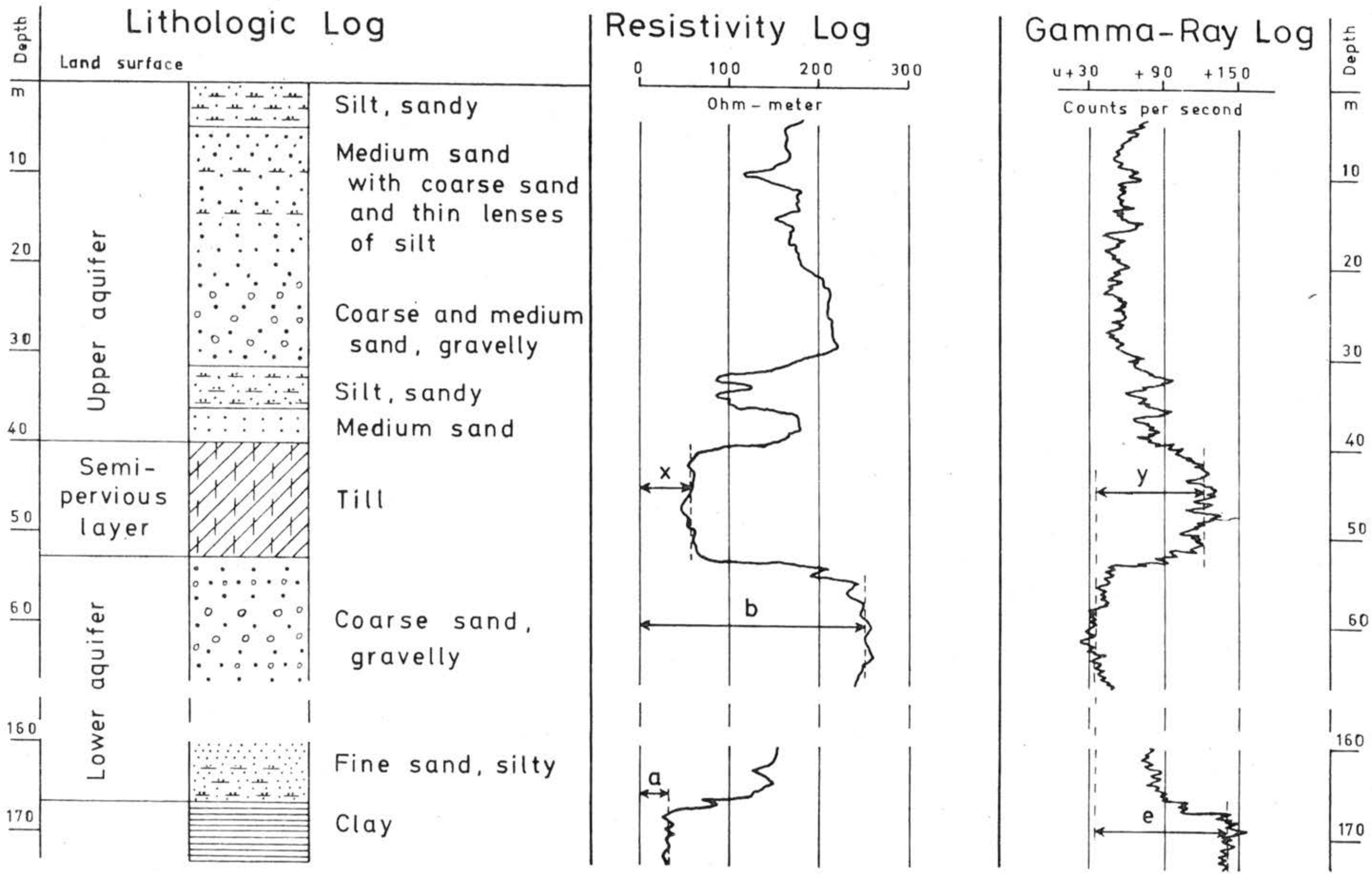

Fig. 1. Resistivities and gamma-ray activities used to calculate the parameters K_R and K_G of aquifer-separating layers

parameter K_G (defined similarly to parameter K_R) has been introduced (see also Fig. 1):

$$K_G = \frac{e-y}{e}.$$

Here, e equals counts/sec. for clay and y = counts/sec. for the separating layer.

In the example shown in Fig. 1 $e = 108$ counts/sec. for the clay and $y = 90$ counts/sec. for the glacial till forming the separating layer which gives a parameter K_G of approximately 0.17. In order to determine the parameter K_G, not only y but also e must be calculated for every borehole in order to circumvent technical variations.

Statistical analysis and the relationship between K_R and K_G and permeability

Statistical analysis

Figs. 2 and 3 show a statistical evaluation of the parameters K_R and K_G calculated for the aquifer-separating layers encountered in approximately 300 boreholes. Fig. 2 shows histograms of the parameters K_R and K_G for silty clay, clayey silt and sandy silt. At the interval (0.1) chosen, the distributions are unimodal and asymmetrical to slightly asymmetrical. The interval containing the largest percentage of K_R values varies from $0 < K_R < 0.1$ for silty clay, to $0.1 < K_R < 0.2$ for clayey silt, to $0.3 < K_R < 0.4$ for sandy silt. The distribution of K_G is almost the same except that the interval with the highest percentage of K_G values is $0.1 < K_G < 0.2$ for silty clay, $0.2 < K_G < 0.3$ for clayey silt, and $0.4 < K_G < 0.5$ for sandy silt.

The values of K_R and K_G for each individual layer are plotted with K_R as abscissa and K_G as ordinate in graphs *a*, *b*, and *c* of Fig. 3. For the resultant point array a computer program* has been used to calculate the middle points (centroids) and regression lines of the point arrays and the coordinates of the corners of rectangles which enclose 60%, 75% and 90% of all points. As it was expected, the positions of the centroids of the three types of sediments differ significantly. The middle point for silty clay has the coordinates ($K_R = 0.09$, $K_G = 0.13$), for clayey silt ($K_R = 0.18$, $K_G = 0.24$), and for sandy silt ($K_R = 0.33$, $K_G = 0.42$). The rectangles emphasize the differences in the three arrays. The 90%-rectangles overlap each other considerably.

As there is no systematic relationship between the magnitude of the resistivity and the intensity of the natural gamma radiation (not even in terms of grain size), the individual regression lines for the point arrays only served for determining the position of the rectangles represented. By means of the rectangles (Fig. 3), a probable sediment type can be assigned to slightly permeable strata (such as till) for which there are geophysical logs but no reliable grain-size data.

A more precise division into very clayey and slightly clayey silt, into clay-free, slightly sandy and very sandy silt is, in principle, possible. However the available analyses of ditch samples are not adequate for such a divison.

* Written by Dr. H. Koschel (Hannover).

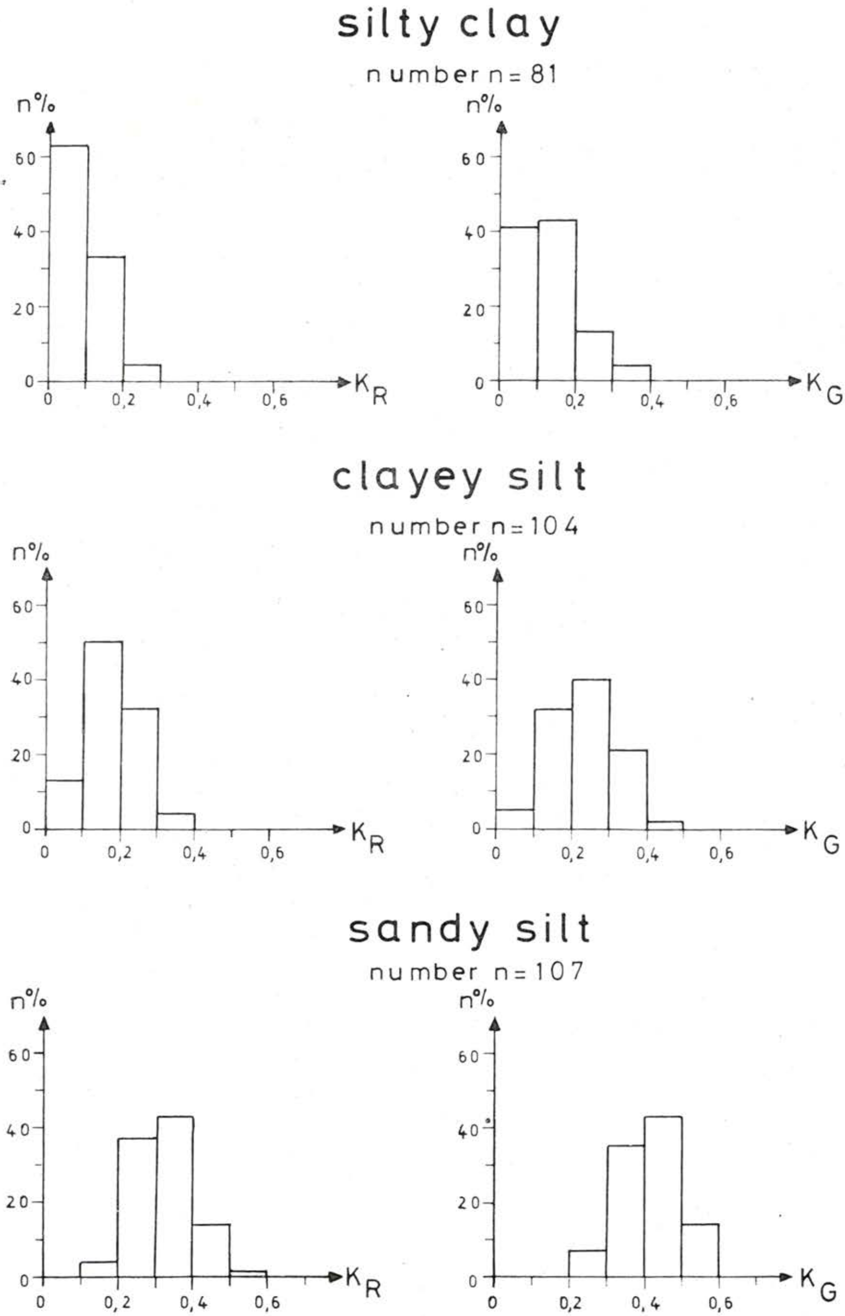

Fig. 2. Histograms of the values of K_R and K_G for silty clay, clayey silt and sandy silt

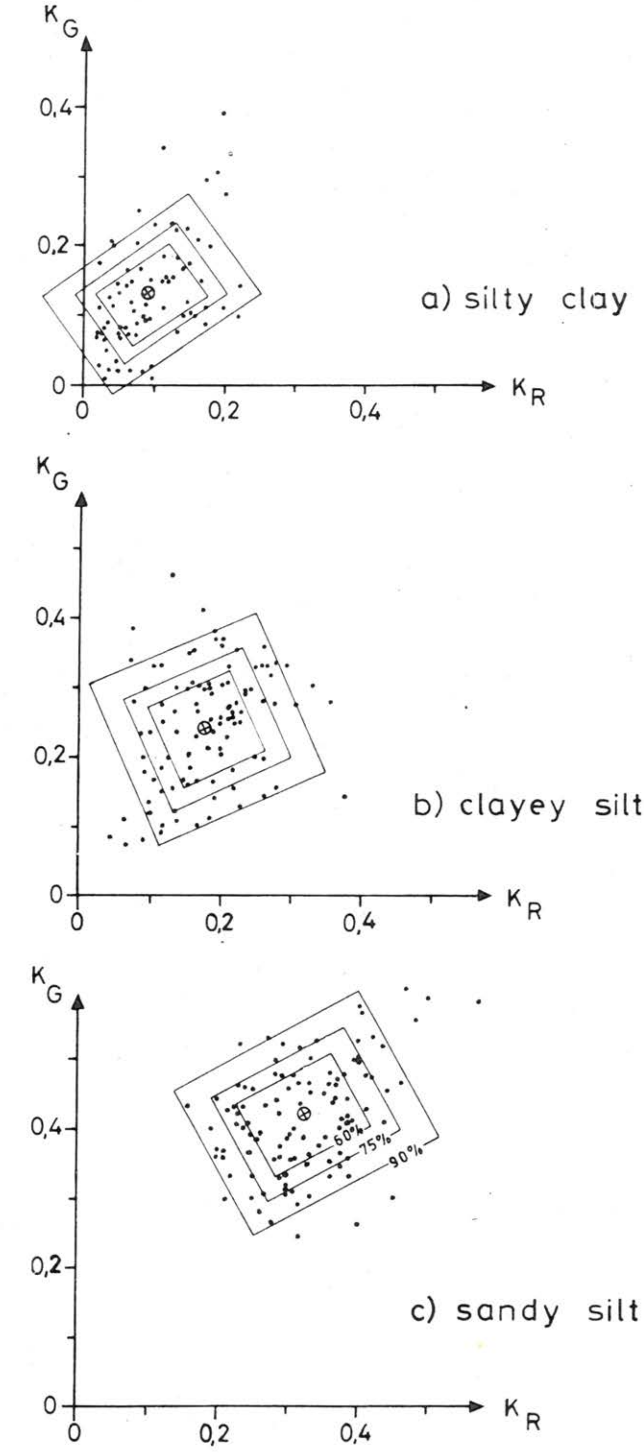

Fig. 3. Relationship between K_R and K_G for 3 types of sediment

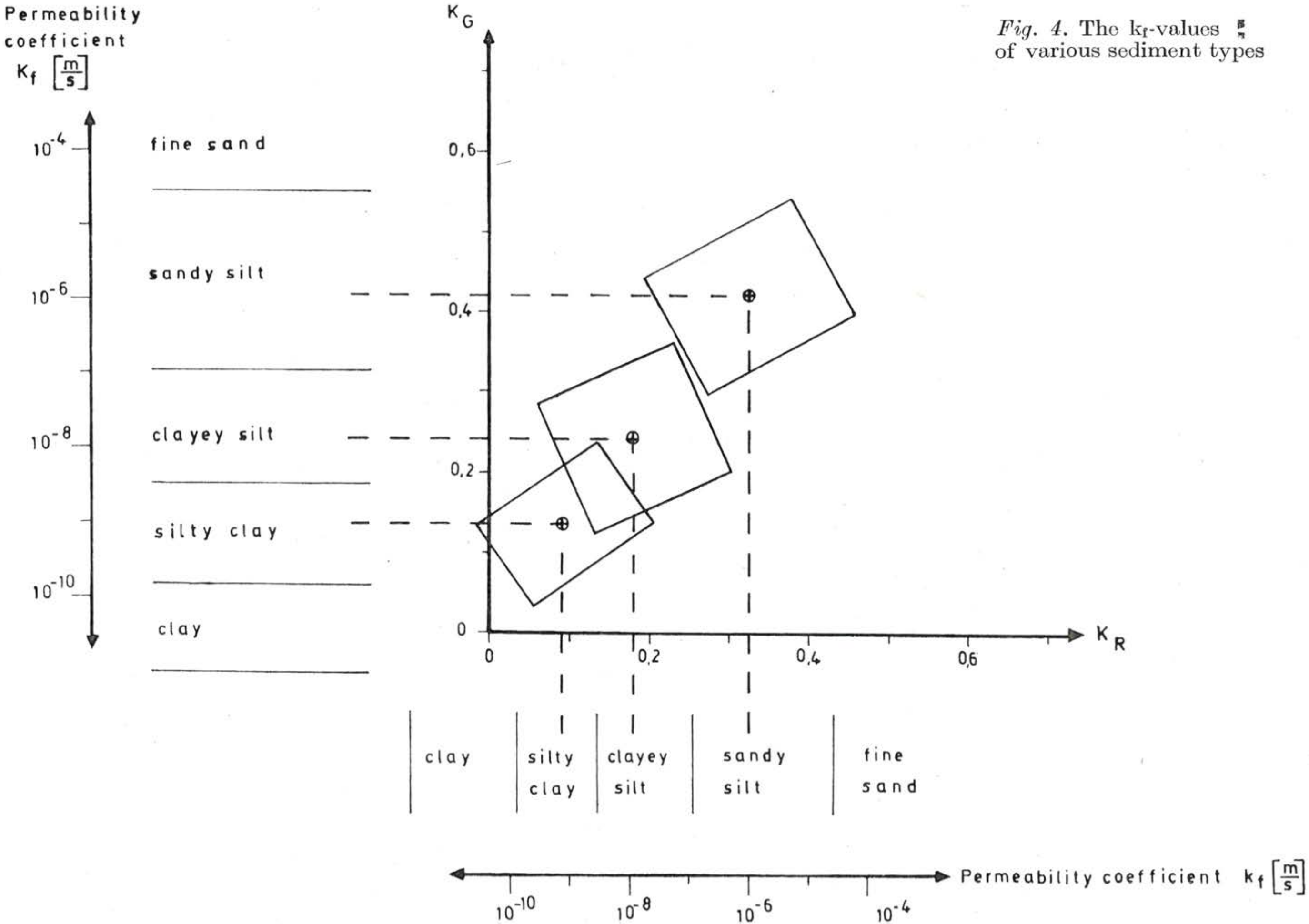

Fig. 4. The k_f-values $\frac{m}{s}$ of various sediment types

Relationship between the parameters K_R and K_G and the permeability coefficient k_f

In order to be able to calculate the quantity of groundwater flowing vertically (under known pressure conditions) through the separating layers, it is necessary to know at least the order of magnitude of the permeability coefficient k_f of these layers.

The k_f values of various grain-size distributions present in aquifers of unconsolidated sediment are well known. Numerous studies using a number of different methods (such as pumping tests, grain-size analyses and distance-velocity measurements) have yielded k_f-values for size fractions from fine sand to fine gravel; the k_f-values range between 10^{-4} m/s and 10^{-2} m/s.

However, slightly permeable layers with $k_f < 10^{-4}$ m/s have received little attention because they are not of interest for groundwater production. Only general data are available and k_f-values ranging between 10^{-5} and 10^{-9} m/s for silt and between $< 10^{-8}$ and 10^{-11} m/s for clay are assumed here (S. N. Davis et R. J. M. de Wiest, 1966; W. Richter et W. Lillich, 1975; H. Schoeller, 1962). Since the parameters K_R and K_G make it possible to determine the grain-sizes of the sediments forming slightly permeable layers, it seems to be justified to relate K_R and K_G to values of k_f.

Fig. 4 shows the plots of K_R against K_G together with the 75% rectangles (and their centroids) for the three grain-size distributions recognized in the layers separating the aquifers in the study area. At the margin of the graphs there are appropriate scales with the k_f-values for the relevant sediment sizes. Fine-grained sand with k_f-values around 10^{-4} m/s and clay with values of $> 10^{-10}$ m/s were selected as fixed points and the interval between was simply divided equally. Reading from these scales silty clay has a k_f of about $9 \cdot 10^{-10}$ m/s, clayey silt of $1.5 \cdot 10^{-8}$ m/s, and sandy silt of $1 \cdot 10^{-6}$ m/s.

The scales allow a k_f-value to be determined for every individual value of K_R or K_G. Sometimes, however, there are considerable differences between the k_f-values determined for the K_R and K_G of a particular $K_R - K_G$ pair. In these cases the k_f-value for K_R should be given preference, since the resistivity data generally correlate better with the grain-size.

It should be pointed out that the method introduced here can only give an indication of the probable order of magnitude of the k_f-values of semi-permeable layers. Further laboratory studies, such as the determination of k_f-values and of the grain-size distributions for a number of drill-cores, are required in order to circumvent or eliminate errors in the method.

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HYDROGEOLOGY OF LARGE SEDIMENTARY BASINS

HYDROGÉOLOGIE DE GRANDS BASSINS SEDIMENTAIRES

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RÉSUMÉ

Les auteurs donnent les caractéristiques des systèmes géohydrodynamiques de sous-sol, artésien, d'élision et thermobarométrique, mis en italique dans « le schéma de la zonalité verticale hydrogéologique ». L'article donne un brève historique de l'évolution des vues sur la hydrogéologie de grands bassins sédimentaires.

The early notions of "artesian waters" and "artesian basins" formed in France where at the beginning of the XIIth century (1126) self-effluent waters were for the first time exposed in the Artesia province (its present name is Artois). In those days French scientists maintained that the movement of subsurface waters is similar to the liquid movements in communicating vessels. In other words, the sites of outflow from the aquifers to the surface were related to the feeding area of the Artesian basin, while the lowered sites linked to the transitional area or that of the basin's water discharge. The gradual accumulation of facts in hydrogeology has led to a change of views regarding the depth of the subterranean hydrosphere. The recently developed thesis (N. K. Ignatovich, 1945) dealing with the existence of three geohydrodynamic zones as well as the zone of intense water exchange, the zone of impeded water exchange and that of the stagnant water regime, is shared, to a certain extent, by many specialists.

According to latest views (I. K. Zaitsev, V. M. Tarasov, 1972), the hydrodynamic zonal distribution of the large confined (artesian) aquifer systems with subaerial and subaqueous boundaries within the entire subterranean hydrospheric stratification (the line of the water saturation limit ought to be adopted as its lower boundary) may be tentatively subdivided into two geohydrodynamic stages different in the dynamics and origin of their water stored.

The upper stage may incorporate geohydrodynamic zones of free and impeded water discharge and the lower stage may include the zone of a highly impeded water discharge. Accordingly, the outcrop of crystalline rocks at the Earth's surface (crystalline rock mass) i.e. its direct emergence onto the surface (crystalline shields) ought to be regarded not merely as areas of confined (artesian) aquifer systems but as independent subterranean water basins (I. K. Zaitsev, 1974), contrary to the view shared earlier and supported erroneously by some specialists even up to now.

The hydrogeological section of the Earth's crust exposes pressure-free or hydrostatically pressured ground-water basins and confined aquifer systems of artesian, elisian and thermobaric basins (see table). The principal difference between them is given by the nature of the subterranean water pressure system, the position in space of the feeding areas and of the discharge of the water-bearing complexes. However, the hydrogeological basins directly corresponding to those mentioned, have nothing common with the actual picture. Each hydrogeological basin is featured by a multiform: geohydrodynamics, and the basin's vertical section reveals a numerous series of regular qualitative transformations, which fit in the following pattern (in a top-down direction): geohydrodynamic ground systems, artesian, elisian and thermobaric systems.

The geohydrodynamics ground systems are spread in the hydrogeological basin section's upper parts (without exception). The feeding of the water-bearing horizons is secured at the cost of the infiltration from atmospheric precipitation and, in a number of cases, by waters ascending from deeper confined aquifer system levels.

The movement is directed towards the lower ground-water levels, and in first approximation they follow DARCY's law. In case of a high position the water temperature usually corresponds to the annual mean of air temperature. The geohydrodynamic ground systems, in their vertical section, may transform into artesian ones and, in the absence of the latters, they are gradually replaced by elisian systems. Geohydrodynamic ground systems within the hydrogeological basins are characterized by a considerable thickness and are

Table 1

A scheme of the vertical hydrogeologic zoning*

Geohydrodynamic system	Features of the geohydrodynamic system (hydrogeological basin)	Rock formation stage within the geohydrodynamic system $\frac{\text{Temperature °C}}{\text{Pressure kbar}}$
Ground system	Pressure-free or under local hydrostatic pressure. Free and impeded subsurface discharge, mainly filtrating, with elements of the elisian discharge	Diagenesis, initial catagenesis $\frac{<350}{<25}$
Artesian system		
Elisian system	Water under pressure. Free, impeded and greatly hampered subterranean discharge, elisian filtrating and conditioned by desorption of bounded waters and influx of retrograde solutions.	Late catagenesis, early metagenesis $\frac{\simeq 350}{\simeq 25}$ ($\pm$ 15%)
Thermobaric system	Pressure-free or under local i.e. partial pressure. Greatly hampered subterranean discharge of mainly overheated waters and retrograde solutions (close to fluids) with elements of elisian discharge.	Late metagenesis, regional metamorphism $\frac{>350}{>25}$

* The scheme of the vertical hydrogeologic zoning is based on the explorations of the following authors: A. M. BLOKH, G. V. BOGOMOLOV, Y. G. BOGOMOLOV, N. B. VASSOEVICH, V. I. DUKHANINA, I. K. ZAITSEV, N. K. IGNATOVICH, A. A. KARTSEV, V. V. KOLODIY, V. I. KONONOV, B. I. KUDELIN, A. V. KUDELSKY, A. F. LEBEDEV, F. A. MAKARENKO, J. V. MUKHIN, A. Mo. OVCHINNIKOV, A. I. SILIN-BEKCHURIN, N. I. TOLSTIKHIN, E. U. FRANK, A. E. KHODKOV, A. V. SHCHERBAKOV, V. ENGELGART etc.

closely connected, and they gradually transit into thermobaric systems (crystalline rock masses or shields, for example, the Byelorussian, Ukranian, Voronezh, African, South African, Australian etc.).

The transition of the geohydrodynamic ground systems into artesian systems is observed in many regions (Paris, North African basin, some intermountain Kopet Dag synclines, the Caucasus etc.). Notions dealing with the nature and position of artesian geohydrodynamic systems in the Earth's crust section originated and developed owing to the study of shallow-seated low pressure aquifer systems to which DARCY's law is applicable. The latter is based on the assumption of the physical indifference of the rocks of water intake. It regards the rocks as a matrix, providing that the shape and size of the hole (pores) may change merely under the dissolving effect of moving waters. In the nature hydrogeological basins are rarely observable as insertable in an artesian geohydrodynamic pattern. In a proper sense they are detectable in areas of extremely low influences made by geostatic load and temperature affecting the rock matrix. In contrast with this, it stands the striking action of subsurface waters driven under the forces of hydrostatic pressure (V. V. KOLODIY, A. V. KUDELSKY, 1972).

The arenaceous to argillaceous formations, a relatively homogeneous rock matrix system, is subjected to irreversible physico-chemical changes under the effects of increased geostatic loads and temperatures (reduction in pore storage capacity and removal of solutions from the pores, piezoelectric effects on crystal contacts with an outflow of element-admixtures, thermobaric destruction of the scattered organic matter etc.), and is converted into an active geochemical system. The origin and dynamics of the underground waters, vapour-water mixtures and gases of the deeper areas of the sedimentary rock sequences are greatly indebted to the initiating action of this very system. Such geohydrodynamic systems, in which the dynamics of the subsurface waters is mainly determined by the reduction of the pore space of clays and by the removal of pore water into bedded reservoirs under the effect of geostatic loads, are named elisian systems (N. B. VASSOEVICH, 1958; A. A. KARTSEV, S. B. VAGIN, 1966; G. V. BOGOMOLOV etc. 1972, 1973). The difference between the elisian hydrodynamic system and the artesian and thermobaric systems is expressed by the fact that both the distribution of pressures and the replenishment of the underground waters are here connected with the geostatic densification of clayey formations and with the removal of pore waters into bedded reservoirs. At the same time, in artesian and thermobaric systems the formation of reservoirs, the pressure and shifting of the subsurface waters, respectively, are under the effect of the filtrating recharge of the water-bearing complexes, moreover under thermic dehydration and phase transitions taking place within the rock-water-gas-scattered organic matter (SOM) system.

The elisian geohydrodynamic system might lack in film-water. Experimental studies of the dependence of thermophysical properties of the moisture-saturated rocks upon temperature (G. V. BOGOMOLOV, etc., 1975; P. P. ATROSCHHENKO etc., 1976) as well as data on the change of electric potential in the course of moisture migration in arenaceous to argillaceous rocks (V. P. BOROVITSKY, 1968) confirm this assumption.

The availability of mineralized water in crystalline rock masses beneath the thick sedimentary deposits is a widely-known fact and the nature of this phenomenon is quite clear. The presence of mineralized water and brine in crystalline shields and rock masses is rather unexpected and hardly explainable by the classical theory of vertical hydrochemical zoning. Most investigations interprete this presuming that these waters are of a relict nature. It appears, however, that their formation is imposed by the warping of the Earth's crust, and, consequently, with the squeezing out of saline waters from the dominating clayey beds into the body of crystalline rock mass and shields. Hence, it is quite explainable that in a number of points of the Baltic, Ukranian and Canadian shields and in crystalline rock masses occurring in the Voronezh and Byelorussian areas one encounters saline waters and brine (Y. G. Bogomolov, 1971).

At depths exceeding the first thousand metres, the compaction under geostatic loads of the clayey rocks is practically finished and the squeezing off of the pored solutions is gradually replaced by other types of pressure. It is believed that the processes of thermobaric metamorphism and rock degradation are most significant. It is a convenience to have such geohydrodynamic systems to be classified as thermobaric systems. The process of mineral dehydration within the geohydrodynamic system develops most intensely at temperatures exceeding 200–300 °C. Overheated waters and supercritical homogeneous fluids of very high fluidity (E. U. Franck, 1968; A. M. Blokh, 1969) and rock wedging capacity, penetrate into the microfractures, the fine capillaries and intercrystalline cavities, stimulating thereby the discharge and redistribution of the products of mineral dehydration, the sublimation of many metal compounds (hydroxids etc.) and SOM destruction. These processes result in the liberation of huge amounts of water, hydrocarbon, nitrogen, carbon dioxide, hydrogen etc. The dissolving capacity of supercritical fluids is reduced due to the decrease of the rate of the water molecule association. However, owing to matrix injections, the fluid's substance concentration constantly increases. Moreover, it appeared that the co-existence of dense homogeneous mixtures of water and hydrocarbons is possible even at temperatures exceeding 400 °C when the pressure reached a few kilobars (E. U. Franck, 1968). On the other hand, substances dissolved in water, particularly hydrocarbon gases, stimulate sufficiently the processes of water decomposition (L. K. Gutsalo, 1974).

The correlation of the sedimentary mantle-base folding boundary with the geochemical boundary of fresh and mineralized waters within the structure-geological system is established in a formal way, in accordance with the chlorine content of the subterranean waters (40 mg/l), and the total dissolved solids content (Ig/l) is most important as a governing low of the hydrogeological zoning. The mirror reflection of the boundary surface separating the fresh and mineralized waters by the relief of a base folding permits to forecast, to a certain extent of authenticity, the depth of its occurrence. This regularity appears to be a general nature within the range of the epi-Karelian—epi-Kaledonian areas of a stabilized Earth's crust, providing the latter's development is connected with the deepest level of tectonospheric instability. The amplitude and the fluctuation period of such a hydrogeological boundary depends upon special factors, however, it ought to be maintained for coeval structures.

As it rises from the concept of unity of the natural waters and lithosphere, the development of the hydrosphere is determined by the evolution of the Earth's crust. The latter is formed by its partial fusion from the Earth's upper mantle ("zoned fusion", V. V. BELOUSOV, 1966), whereas the gradually revealed periodicity in the vertical shifting of the Earth's crust material and mantle determines tectonic periodicity (the cyclic recurrence), which in its turn controls hydrogeologic cycles exposed in all areas. Thus, it may be assumed that planetary pulsations of the Earth's substance determine the development of geohydrodynamic systems. In the final analysis this development is directed to a predominant formation of infiltrated confined aquifer systems (artesian basins, A. A. KARTSEV et al., 1969). And, hence, the elaboration of the theoretical foundation of hydrogeological zoning, forming the very basis of regional hydrogeology, may be achieved only from the viewpoint of recognizing the mentioned leading tendency in the development of the Earth's hydrosphere as a whole and the pressure aquifer confined system (artesian) basins in particular (G. V. BOGOMOLOV, Y. G. BOGOMOLOV, 1972).

Once the regional subdivision of the subsurface waters is made in accordance with the actual physicogeographical zones, the regional subdivision of deep waters is involved with the outlines and size of the abyssal water bodies. The lower limit of such deep reservoirs is determined by regional confined aquifer systems with the boundary of the water saturation limit, bearing in mind that the latter's vertical position depends on the rate of the regional heat flow and, respectively, on the age of the tectonic structures (Heat regime of the USSR interiors, 1970; Y. G. BOGOMOLOV, 1971). As a rule, the upper limit of such geological forms does not reveal any structural/lithologic limits and is determined directly by the erosion basis i.e. by the provisional boundary of the deeper and the surface discharge. Thus, the principal geostructural subdivisions in hydrogeology are presently regarded as hydrogeological structures characterized by peculiar conditions of feeding, transit, calming down and formation of subterranean waters mainly related to the geometry of geological formations. Hence, the typification of similar structures calls primarily for a special attention to be paid to the shape and dimension of geological bodies, rather than to the genetic relationships of the hydrogeological and structure-geological conditions emerging from the theory of evolution of the Earth's crust and hydrosphere. A series of paleo-relationships such as paleohydrogeologic, paleogeothermic, paleogeologic, paleotectonic and other paleo-correlations ought to be regarded merely in their first and very rough approximation, owing to the fact that "age" is a most complex parameter. And, nevertheless, the genetic envolvement of hydrogeologic and geologic factors is an indispensable element, which in a short while will undoubtedly constitute a methodological basis for the hydrogeologic zoning.

This approach raises an interesting analogy borrowed from geotectonics concerning the evolution of concepts in determining platform age. Some scientists (N. A. KRYLOV etc., 1974; G. I. AMURSKY, 1972) believe that the establishment of the platform regime is intimately connected merely with the period of the early shaping of the sedimentary mantle. Others (N. S. SHATSKY, 1945; V. S. ZHURAVLEV etc., 1975) maintain that the early shaping of the platform regime is connected with the age of basement's folding i.e. with end of the geosynclinal structures' development passing over to the emergence of the platform structures. Having shared the second viewpoint, the development of the platform regime, with an early period of taking shape of the platform mantle, leads to the exclusion of the long-lasting shaping of the basement's internal elevation from the class of "platform structures", providing that the former frequently reaches huge dimensions. The same is observed in the course of hydrogeological zoning. The schematization of the hydrogeological structures according to the feeding areas (the basement's elevation), the transit areas (slopes) etc. appear to be an outdated approach; the simple reason for this is that the formation of the subterranean hydrosphere is an integrate process of interactions of the fluid and solid phases influenced by intra-Earth and extra-Earth sources of energy, whereas the geohydrodynamic dismemberment of the subterranean hydrosphere in the lateral and vertical directions represents a particular phenomenon with its own features.

It is believed that the recent stage in the development of hydrogeological knowledge urgently requires a statistical analysis of hydrogeologic and structure-geologic parameters within the limits of several large-scale confined aquifer systems, with the aim of revealing the actual genetic relationships between the hydrogeologic and structure-geologic contents (zoning).

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LA NAPPE DES SABLES VERTS ET L'ALIMENTATION EN EAU DE LA RÉGION PARISIENNE

THE GREEN SAND AQUIFER AND WATER RECHARGE IN THE PARIS AREA

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ABSTRACT

The "green sand" aquifer—or Albian Aquifer—reached at a depth of more than 600 meters underground level in Paris has been put into production since 1841.

Formerly, this aquifer was artesian; nowadays pumping off is necessary, following a considerable drawdown of water table. In 1935 a strict regulation of the aquifer exploitation was undertaken resulting in a stabilisation of the water table at about 25 meters underground level for a mean rate of discharge of about 0.8 m^3/s.

Recent investigations, using mathematical modelling and simulation, were carried out by BURGEAP for the Paris Suburbean Water Corporation: they provided evidence on the balance of the aquifer, and on its boundary or confinement parameters (transmissivity, storage capacity, parameters of leakage with over- or underlying strata).

In order to improve the distribution of drinking water resources in the Paris area, and to reduce the length of water conveying pipes—which are expensive in urban areas—a new method of ground water exploitation could consist in recharging the aquifer by injection of surface water into boreholes, and pumping from other wells. Comprehensive testing of such artificial recharge was conducted in a borehole reaching the "green sand" aquifer. The complete success of the operation prooved that this ground water resource, omnipresent in the Paris area, was likely to be used within the imperative conditions of its balance preservation.

A l'intérieur du grand bassin sédimentaire qui correspond à la partie centrale et septentrionale de la France, et au centre duquel se trouve Paris, la nappe de l'Albien (Crétacé inférieur), dite encore « nappe des sables verts », couvre une superficie de l'ordre de 70 000 km². C'est aux environs mêmes de Paris que la profondeur de la nappe (plus de 500 mètres) est maxima; elle y est recouverte en effet par plus de 400 mètres d'un calcaire tendre et plus ou moins marneux dénommé « Craie ».

Il y a beaucoup plus d'un siècle que l'on a commencé à mettre cette nappe en exploitation, et son utilisation s'est poursuivie au cours des temps, à un rythme variable, jusqu'à ce que, en 1935, des dispositions règlementaires y interdisent tout nouveau prélèvement. C'est cette histoire de la nappe de l'Albien que nous voudrions résumer ici, avant de montrer que des techniques particulières paraissent permettre, sous certaines conditions et sans modifier son équilibre général, de l'utiliser à nouveau.

Dès le début du dix neuvième siècle, certains observateurs, sur la base d'études géologiques assez approfondies, soupçonnaient la possibilité de recouper au-dessous de la capitale les niveaux sableux qui avaient été reconnus à la périphérie du bassin sédimentaire, et escomptaient le jaillissement à la surface du sol de la nappe aquifère correspondante: un jaillissement* dont on augurait beaucoup, à une époque où l'état de la technique ne permettait guère d'exploiter, par rabattement à l'intérieur d'un forage ou d'un puits, des débits importants.

Les crédits, pour l'exécution à Paris même d'un forage destiné à recouper les formations sableuses de l'Albien et prévu pour une profondeur de 400 m, furent votés par le Conseil Municipal de Paris en 1833.

Il s'agissait, en fait, d'un pari, car c'était la première fois en France et sans doute dans le monde qu'un forage de recherche d'eau aussi profond était envisagé; de plus, les connaissances géologiques du Bassin Parisien étaient telles que la présence de sables aquifères artésiens ne constituait qu'une présomption.

La réalisation du forage fut confiée à un ingénieur mécanicien du nom de Mulot, qui commença les travaux le 24 décembre 1833 dans la cour de l'Abattoir de Grenelle. Le matériel dont il disposait était évidemment fort précaire: une chèvre, des palans et des tiges de bois de 8 mètres de long, travaillant au battage, des treuils entraînés par des roues horizontales actionnés d'abord par des hommes, puis par des chevaux . . .

Les incidents de forage furent très nombreux, et ce n'est qu'au bout de sept ans, en 1841, que Mulot atteignait la nappe artésienne de l'Albien à 548 mètres. Ce fut un succès, qui se traduisit par un enthousiasme général: 160 m^3 d'eau à l'heure jaillissants, avec un niveau d'équilibre égal à 33 mètres au-dessus du sol. Mulot fut récompensé de ses efforts par l'attribution d'une rente viagère et de la Légion d'Honneur.

Les travaux de Mulot eurent pour résultat d'encourager la réalisation d'assez nombreux forages à la périphérie du Bassin Parisien, où les sables albiens sont beaucoup moins profonds, et notamment à sa bordure occidentale. Ils devaient confirmer la régularité de la formation aquifère sous une grande partie du Bassin de Paris.

La productivité du forage de Grenelle ne devait, cependant, durer que quelques mois: par suite d'une détérioration du tubage, le débit chuta à environ 40 m^3/h, ce qui amena à critiquer le bien fondé de l'ouvrage réalisé.

En dépit de ces déboires, la ville de Paris décida quinze ans plus tard, en 1855, l'exécution d'un nouveau forage à Passy, non loin du précédent. Les travaux furent cette fois-ci confiés à un ingénieur allemand du nom de Kind, mieux équipé que Mulot. Les travaux durèrent tout de même plus de cinq ans, et ce n'est qu'en 1861 que la nappe jaillissait, avec le débit très important de 700 m^3/h, qu'on utilisa à alimenter les rivières du Bois de Boulogne.

Dès la mise en production de ce nouveau puits, on s'aperçut que le débit baissait dans le puits de Grenelle et c'est approximativement à cette date que des nouvelles idées sur le fonctionnement des nappes artésiennes furent émises par Degoussee et Laurent. Ces auteurs expliquaient en effet les interactions entre les deux puits par des remarques tendant à modifier notablement la théorie antérieure des « vases communicants » : le débit recueilli dans un trou de sonde dépend non seulement de la différence de niveau entre les affleure-

* Le nom de « nappe artésienne » vient de l'Artois, province du Nord de la France où des puits jaillissants furent réalisés dès le Haut Moyen-Age (en 1120 à Lillers) pour exploiter une nappe de sables tertiaires maintenue captive sous un recouvrement marneux.

ments et le captage, mais aussi des fuites possibles à travers différentes issues possibles et du « frottement » entre les affleurements et le trou de sonde. Nous pensons que ces remarques sont les premières à faire appel à la notion de transmissivité.

Entre 1863 et 1904, d'autres forages furent réalisés à Paris et sa proche banlieue:

– les forages de la place Hébert et de la Butte-aux-Cailles, destinés chacun à alimenter une piscine pour profiter de la température de l'eau, qui est voisine de 30°. Les travaux, d'une extrême lenteur, durèrent respectivement 28 et 31 ans.

– le forage de la Raffinerie SAY, terminé en 1869 sans incident notoire.

– le forage du Bois de Vincennes, réalisé en 1900 à titre de démonstration, dans le cadre de l'Exposition Universelle.

C'est seulement en 1904 que la décision fut prise d'utiliser la nappe albienne pour l'alimentation en eau potable de la région parisienne, à partir de deux forages:

– le forage de Carrières-sous-Poissy en 1904,

– le forage de Maisons-Laffitte en 1909.

Par la suite, avec l'apparition du système de forage Rotary, à partir de 1930, les puits à l'Albien se multiplient:

– en 1931, la Société Lyonnaise des Eaux fait exécuter avec succès et rapidité par la Société LAYNE-FRANCE, trois forages à Orsay, Poissy et Viry-Chatillon.

– de 1932 à 1937, on assiste à une véritable prolifération des forages (21 forages en 5 ans), la plupart réalisés au Rotary.

Cette multiplication des forages se traduit alors par une diminution très nette des débits des forages et de leur artésianisme. Le puits de Grenelle avait déjà cessé de jaillir en 1910, celui de Passy en 1930. A partir de 1930, la baisse de l'artésianisme se fit sentir de façon encore beaucoup plus nette. Les puits les plus anciens s'arrêtèrent les premiers, parce qu'équipés de manière artisanale, mais les puits modernes eux-mêmes accusèrent une baisse de leur niveau piézométrique.

En 1934, au moment même où certaines personnalités voudraient lancer une politique de « l'eau blanche » à Paris et proposent 100 nouveaux puits à l'Albien, certains auteurs dont Lemoine, Humery et Soyer lancent un véritable cri d'alarme en démontrant que depuis la réalisation du puits de Grenelle (c'est-à-dire en 92 ans) la nappe a baissé de 74 mètres.

Le Conseil Municipal de Paris est saisi du problème et son rapporteur l'expose en ces termes :

« Il faut surtout incriminer la concentration des forages au voisinage de la capitale, qui en créant un appel d'eau considérable vers le même point, a eu pour conséquence des pertes de charges internes qui ont abaissé la pression de l'eau à son point d'utilisation et fait disparaître le jaillissement. »

On montrait ainsi que ce n'est pas l'insuffisance d'alimentation en eau par les bordures qui était responsable de la baisse piézométrique, ni l'épuisement de la nappe, mais les pertes de charge à l'intérieur du réservoir.

Non seulement le projet des « 100 nouveaux puits » fut définitivement abandonné, mais on rechercha le moyen de stabiliser la situation, par l'interdiction de nouveaux puits à proximité de Paris. Un décret-loi, promulgué le 8 août 1935, précise à cet effet:

Article 1er : « Aucun puits ou sondage de plus de 80 mètres de profondeur ne pourra être entrepris dans les départements de la Seine, Seine-et-Oise et Seine-et-Marne, sans autorisation préalable. »

Il était suivi du décret d'application du 4 mai 1937 dans lequel on relève :

« Les exploitants des puits et sondages pour le captage d'eaux souterraines doivent conserver trace à leur date de toutes les mesures de débit, de température et analyses auxquelles il sera procédé, ainsi que des incidents d'exploitation survenus et des changements constatés dans le régime des eaux. La mesure des débits dans les conditions normales d'exploitation devra être faite une fois par an au minimum. »

Les problèmes posés par la nappe de l'Albien continuent toutefois à susciter diverses recherches et études. Telle celle de A. Vibert qui indique et montre que la baisse de l'artésianisme n'est pas une affaire de réalimentation :

« Ce n'est pas la nappe qui s'épuise, c'est la conduite servant au transport souterrain des eaux, qui devient insuffisante, eu égard au débit, toujours plus grand qui lui est demandé. »

L'année 1939 voit une étude descriptive très détaillée sur la nappe par Lemoine, Humery et Soyer avec une synthèse sur la structure et la stratigraphie des terrains crétacés, la coupe de tous les forages, le bilan de la nappe, ses qualités chimiques. Cet ouvrage reste aujourd'hui un document fondamental.

En 1953 commence la prospection systématique du pétrole dans le Bassin de Paris, avec les moyens considérables dont disposent plusieurs grandes compagnies. Plusieurs centaines de forages pétroliers traversent les terrains crétacés, et en explorent la structure. Les gisements d'hydrocarbure étant localisés dans le Jurassique, les sondages se sont surtout intéressés à cet horizon. Mais les terrains crétacés et l'Albien font l'objet de diagraphies qui permettent de déceler avec précision l'épaisseur et la superposition des diverses intercalations perméables de l'Albien.

Un organisme officiel, la Direction des Carburants, réalise en 1965 une étude de synthèse et une première ébauche de l'écoulement de la nappe en utilisant un modèle mathématique de simulation.

Un peu plus tard (1967), Lauverjat publie une thèse initulé « Contribution à l'étude géologique et hydrogéologique de l'Albien dans le centre du Bassin de Paris » qui réunit l'ensemble des données acquises à cette date.

A partir de 1935, la nappe de l'Albien (qui ne fournit d'ailleurs que quelque 3% de la consommation en eau de la région parisienne) n'a donné lieu qu'à peu de nouveaux forages. Pourtant, sa supériorité sur d'autres ressources en eau est considérable :

— c'est une ressource très régulièrement répartie, pouvant donc être captée en n'importe quel point de la région parisienne,

— les eaux de l'Albien sont de très bonne qualité (leur température, voisine de 30°, ne permet toutefois pas de les distribuer immédiatement après la sortie du forage),

— la nappe est parfaitement protégée des pollutions, et les captages ne nécessitent pas de périmètres de protection,

— étant captive, elle transmet rapidement les pressions,

— elle peut fournir des débits ponctuels substantiels (de l'ordre de 150 à 200 m^3/h par forage),

— enfin son exploitation ne nécessite pas de frais d'exhaure élevés, compte tenu de la faible profondeur de son niveau piézométrique.

On a de ce fait, été conduit à se demander si la technique de recharge des nappes souterraines (peu répandue encore en France, qui dispose d'assez larges ressources) ne pourrait pas permettre sous certaines conditions, à débit limité et sans modifier le niveau d'équilibre général, d'utiliser à nouveau la nappe de l'Albien. A cet effet, le Syndicat des Communes de la Banlieue de Paris pour les Eaux confia en 1974 au bureau d'études BURGEAP la mise au point d'un dispositif expérimental d'alimentation artificielle.

Un essai de recharge de longue durée (6 mois) fut réalisé en 1973, dans un forage de la banlieue Est (Noisy-le-Grand). L'expérience a permis d'examiner en détail des problèmes très divers que pose ce genre d'opération: colmatage du forage d'injection, compatibilité des eaux d'injection et des eaux de la nappe, débits injectables etc. Les résultats ont conduit à conclure de façon très positive sur la possibilité de recharger la nappe.

Il était toutefois nécessaire, compte tenu de leur prix de revient, avant de réaliser dans la pratique des opérations cumulées d'injection et de reprise d'eau, de déterminer avec précision, compte tenu de leur prix de revient les dispositifs à mettre en oeuvre et leur influence sur la nappe.

Dans ce but, un modèle mathématique de la nappe dans la région parisienne a été réalisé, après qu' aient été réunies toutes les données disponibles sur la nappe et sur son exploitation antérieure : données que nous résumerons brièvement ci-dessous.

Données prises en compte pour le modèle mathématique

Le réservoir albien est constitué d'horizons argilo-gréseux et sableux (Sables de Frécambault, Sables des Drillons, Sables Verts) plus ou moins isolés les uns des autres par des horizons marno-argileux (Marnes de Brienne, Argile du Gault, Argiles Tégulines, Argiles de l'Armance). L'ensemble des horizons gréseaux et sableux représente une épaisseur cumulée de 30 à 40 m, et peut être considéré comme un aquifère unique, dont la transmissivité est comprise entre 10^{-3} et 10^{-2} m^2/s dans la région parisienne. L'horizon aquifère se situe vers 600 mètres de profondeur et son niveau piézométrique est voisin de la surface du sol. La nappe est donc fortement captive, avec un coefficient d'emmagasinement adimensionnel faible, de l'ordre de 10^{-4}.

Les horizons albiens sont relativement bien isolés des niveaux qui les encaissent, mais l'étanchéité n'est pas parfaite et il se produit des échanges par percolation verticale :

— d'une part avec les horizons sablo-argileux sous-jacents du Barrémien et du Néocomien,

— d'autre part avec les horizons crayeux superposés du Cénomanien, Turonien, Sénonien (seulement dans la partie Ouest du Bassin de Paris).

Les circulations verticales se font, selon le lieu, de l'Albien vers les terrains encaissants, ou inversement, des terrains encaissants vers l'Albien. Lorsque la nappe est soumise à un pompage, elle est notablement déprimée, ce qui entraîne soit une augmentation des percolations des épontes vers l'Albien, soit une diminution des percolations de l'Albien vers les épontes.

La structure de l'horizon albien est assez monotone, en forme de cuvette centrée sur Meaux (Est de Paris), comme l'ensemble des terrains secondaires du

Bassin Parisien. Il faut toutefois signaler (Fig. 1) quelques particularités tectoniques :

— tectonique souple, marquée par l'anticlinal de Beynes, l'anticlinal de Saint-Illiers, l'anticlinal de Vernon Rouen, l'anticlinal du Pays de Bray,

— tectonique cassante, mise en évidence par les failles de Rouen-Vernon, le décrochement qui prolonge vers l'Est la structure du Pays de Bray, la faille qui longe la vallée de la Seine au Nord de Mantes. Certains de ces accidents jouent un rôle hydrodynamique, particulièrement la faille de la Seine, qui semble favoriser les échanges verticaux entre l'Albien et les terrains superposés, (ils sont en ce qui les concerne drainés par la Seine).

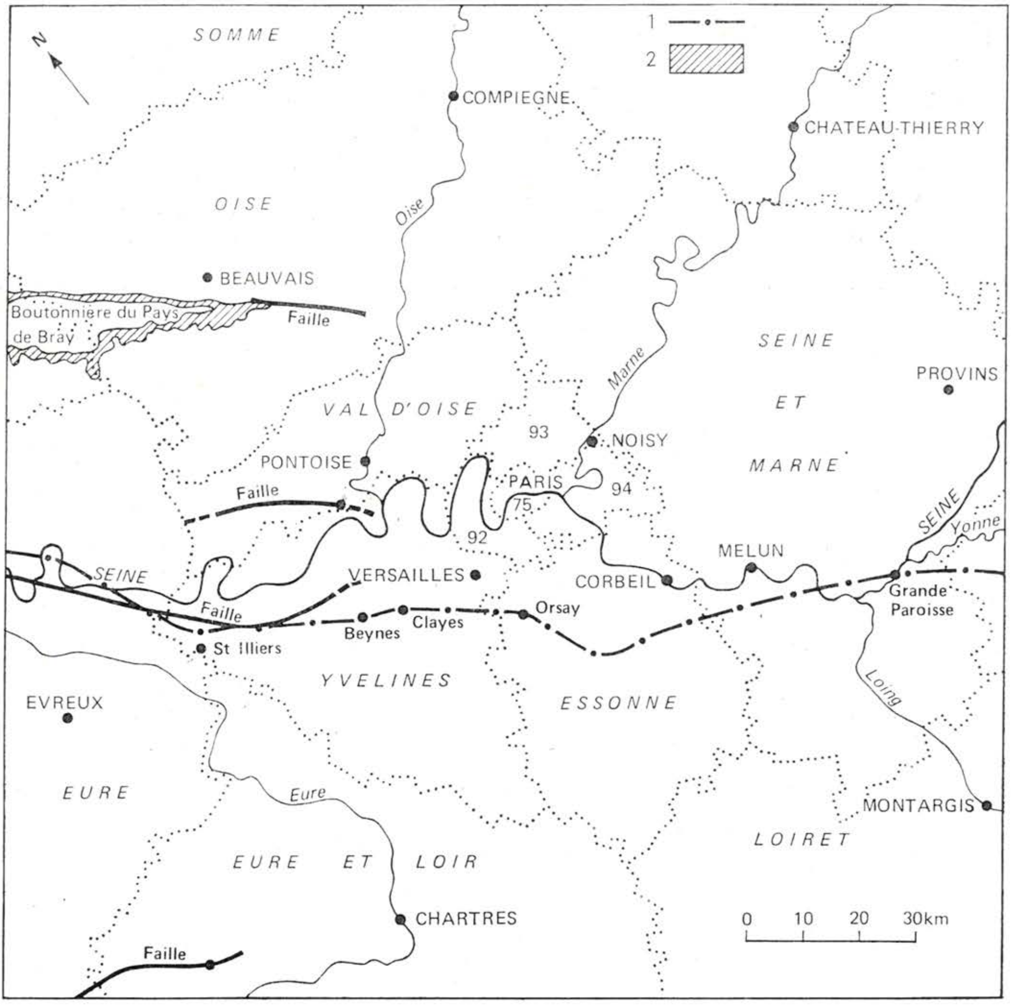

Fig. 1. Carte des affleurements et des principales failles de l'Albien

1. Tracé de la coupe Vernon-Sens, *2.* affleurement de l'Albien

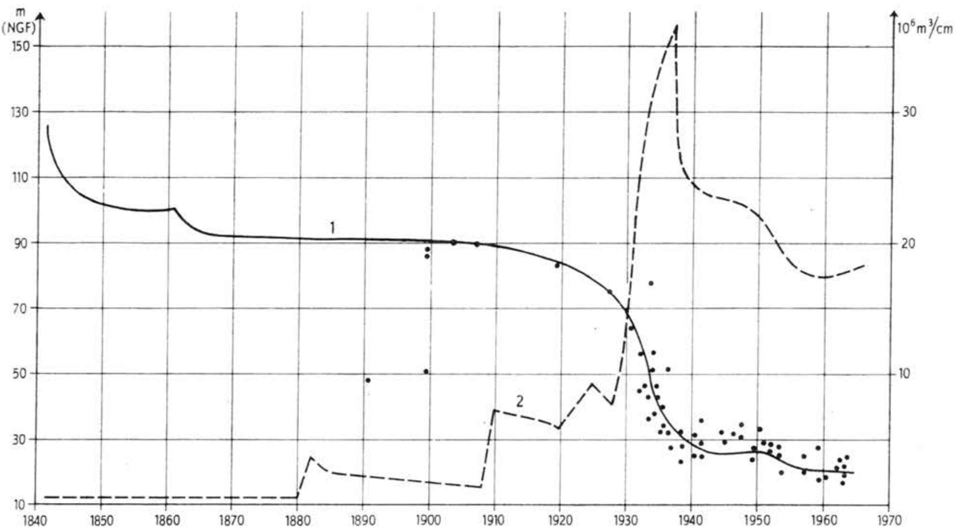

Fig. 2. Historique des niveaux piézométriques et des débits de prélèvement dans la région parisienne. (Document établi d'après la Direction des Carburants)

1. Historique des niveaux piézométriques dans la région parisienne, *2.* historique des débits pompés dans la région parisienne

La piézométrie de la nappe a considérablement évolué depuis sa mise en exploitation en 1841. Un graphique de l'évalution d'ensemble des niveaux piézométriques dans la région parisienne, établi par la Direction des Carburants d'après les mesures faites dans une vingtaine de puits disséminés (Fig. 2) met en évidence une baisse importante du niveau piézométrique. On constate cependant que cette baisse, qui s'effectue par paliers, semble se stabiliser vers les années 1960 après avoir connue une chute importante vers 1935.

Toutefois, cette stabilisation est discutée par les auteurs: pour certains elle est réelle, pour d'autres, dont J. LAUVERJAT, l'évolution n'est past terminée (certains auteurs parlent même d'une baisse de niveau de 1 mètre par an). Nous estimons, en ce qui nous concerne que des baisses importantes peuvent se produire localement (mise en route ou réfection d'un forage), sans que le niveau général moyen s'en ressente: c'est ce qui parait ressortir de la courbe de la Fig. 2 qui retrace l'évolution des niveaux dans plusieurs forages.

Avant 1920, il n'existait que peu de forages et cela interdit pratiquement de tracer une carte piézométrique antérieure à cette époque. Sur les figures 3, 4, 5 nous avons fait figurer les cartes piézométriques de 1920 et 1935 établies par la Direction des Carburants, celle de 1963—66 établie par J. LAUVERJAT. Ces trois cartes soulignent combien s'est accrue la dépression de la nappe dans la région parisienne. On peut constater notamment qu'entre 1920 et 1966 le niveau a baissé d'environ 70 mètres au droit de la ville de Paris.

Une des principales difficultés de l'étude a été d'établir un historique, relativement précis des débits pompés depuis 1920. En fait, on a été amené pour prendre ces données en compte sur le modèle, à scinder cet historique en 4 périodes :

1920—1930 débit moyen d'exploitation de 0.25 m^3/s,
1930—1950 débit moyen d'exploitation de 0.77 m^3/s,

1950—1964 débit moyen d'exploitation de 0.72 m³/s,
1964—1974 débit moyen d'exploitation de 0.8 m³/s.

Ces coupures ne font pas d'ailleurs apparaître la pointe d'utilisation, qui s'est produite vers 1935 avec un débit de 1.2 m³/s.

Réalisation du modèle

Des données qui ont été rassemblées il ressort que, sous l'effet du rabattement provoqué par l'exploitation, l'eau fournie par la nappe est susceptible de provenir de la décompression du réservoir d'une part, de l'apport de systè-

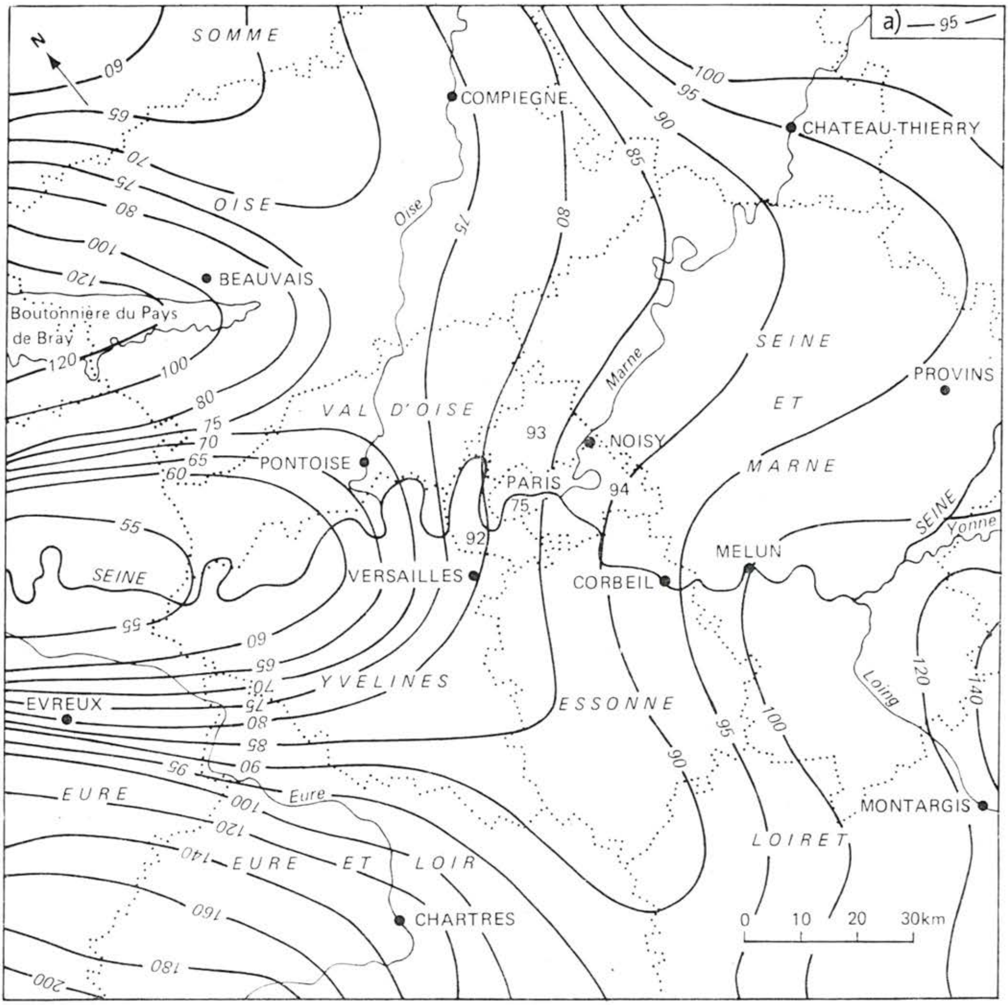

Fig. 3. Carte piézométrique de la nappe albienne en 1920. (D'après la Direction des Carburants)

a) Courbe piézométrique en mètre NGF

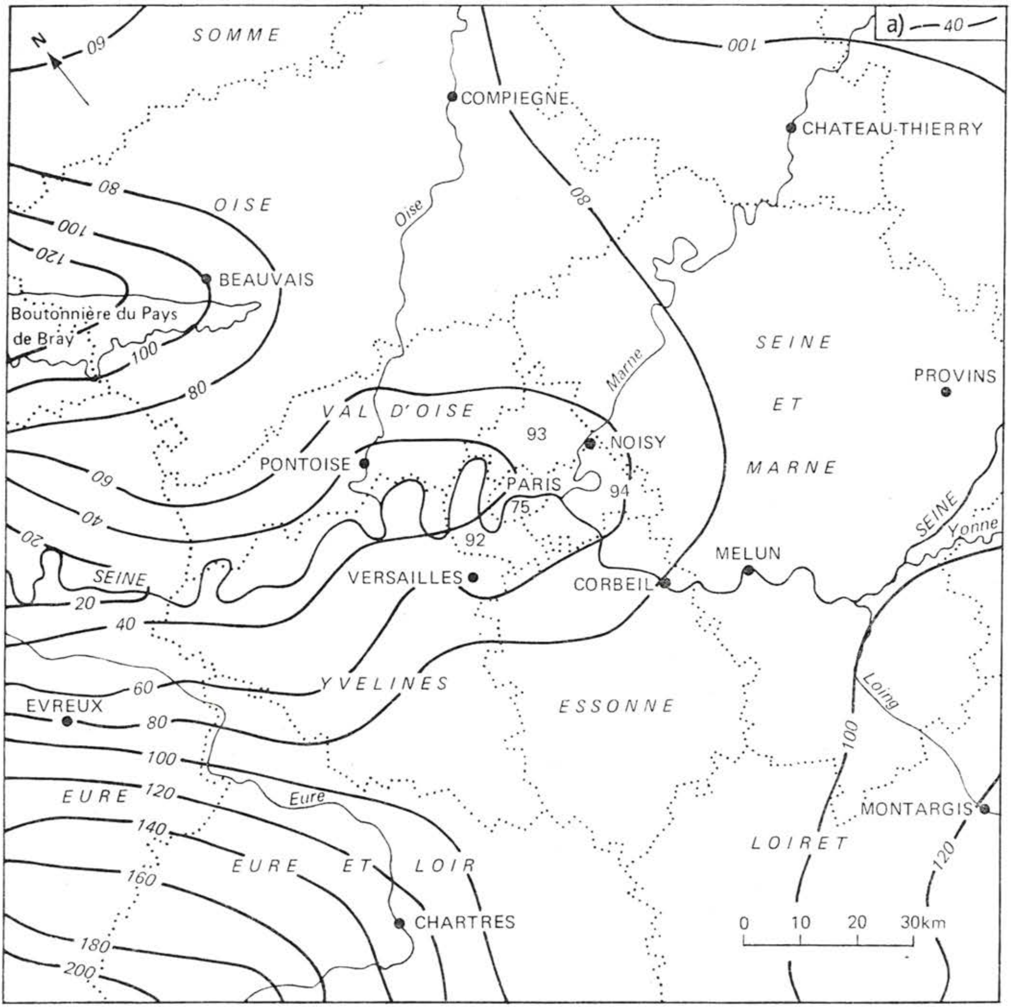

Fig. 4. Carte piézométrique de la nappe albienne en 1935. (D'après la Direction des Carburants)

a) Courbe piézométrique en mètre NGF

mes aquifères adjacents ou environnants, d'autre part. Parmi ces derniers apports :

— la nappe de la Craie,
— la nappe du Barrémien-Néocomien,
— les affleurements de bordure, eux-mêmes alimentés par les pluies.

Dans la série stratigraphique générale, l'aquifère albien se comporte comme une unité de circulation d'eau privilégiée, liée à d'autres unités aquifères dont on ne peut pas ne pas tenir compte du point de vue hydrodynamique. Il est manifesté, toutefois, que l'Albien est assez bien isolé des unités de réalimentation possibles, car dans le cas contraire, la baisse du niveau piézométrique n'aurait pas été aussi importante depuis la mise en exploitation de la nappe.

Ce sont précisément les mécanismes de circulation et de décompression au sein de la nappe albienne et de réalimentation par les épontes qui ont pu être précisés et quantifiés au moyen de la simulation des historiques sur modèle mathématique.

Alors que le modèle qui avait été réalisé en 1965 par la Direction des Carburants couvrait l'ensemble de la nappe albienne dans la Région Parisienne, celui dont il est question ici ne couvre qu'un carré de 180 km de côté, centré sur Paris. En effet, le but du modèle étant d'étudier un schéma d'ex-

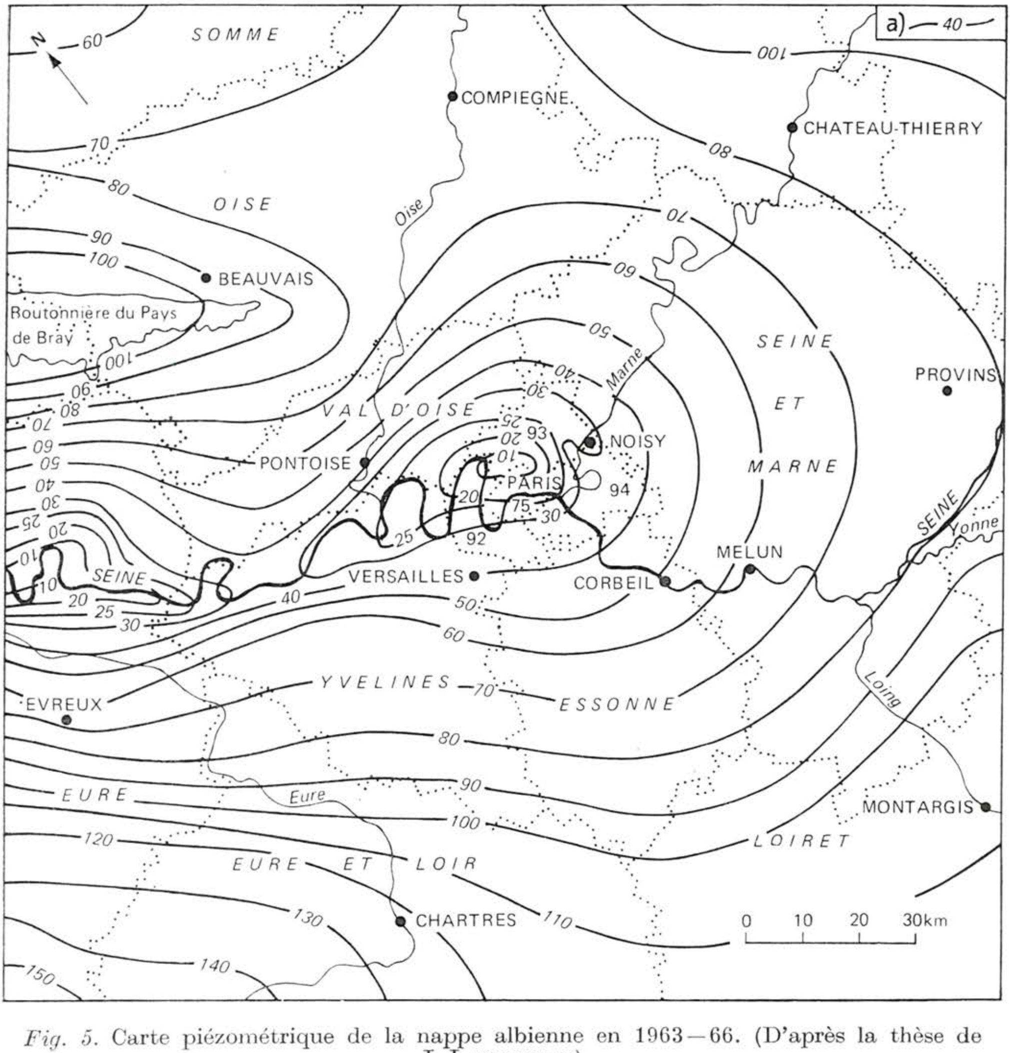

Fig. 5. Carte piézométrique de la nappe albienne en 1963—66. (D'après la thèse de J. LAUVERJAT)

a) Courbe piézométrique en mètre NGF

Tableau 1

Année	1930	1950	1964	1974
Débit aux limites du modèle	331	498	511	510
Débit d'échange avec la Craie	−29	220	158	185
Débit d'échange avec le Barrémien-Néocomien	−52	77	43	65
Débit d'exploitation	−255	−793	−717	−800
Débit dû à la décompression de la nappe	5	2	5	40

ploitation ne modifiant pas le résultat global du débit d'exploitation à long terme, on s'est volontairement limité à la zone représentée sur les planches. On a pu ainsi préciser beaucoup des phénomènes en jeu à l'intérieur de la région parisienne.

Le modèle a d'abord fait l'objet d'un calage de façon à restituer les évolutions piézométriques les plus voisines possibles de la réalité.

A partir de ce calage, on a pu déterminer le bilan de la nappe, comme le montre le tableau 1.

Le calage a mis en évidence et a permis de chiffrer certains phénomènes particuliers :

— la drainance de la Craie est surtout importante dans la Basse Seine, au niveau de la faille de Vernon-Rouen. Au contraire, à l'Est du Bassin Parisien les communications avec la Craie sont pratiquement nulles, ce qui s'explique par le fait que la base de la série crayeuse est constituée dans cette région par des niveaux marneux très épais et très peu perméables.

— la drainance du Néocomien-Barrémien est assez homogène et correspond à des perméabilités verticales de l'ordre de 10^{-9} m/s.

— la perméabilité moyenne de l'Albien est plus forte au centre du Bassin Parisien que sur les bordures.

— le coefficient d'emmagasinement par décompression est de l'ordre de 10^{-4} pour une épaisseur moyenne de la couche d'environ 30 mètres.

A la fin des opérations de calage, le modèle a été utilisé pour déterminer la possibilité d'utiliser la nappe dans le cadre d'un dispositif associant injection et reprise comme on l'a indiqué ci-dessus.

Le résultat de ces dernières simulations a été positif. Il devrait ouvrir des perspectives pour l'utilisation de la nappe de l'Albien en tel ou tel point de la région de Paris, en respectant l'impératif qui consiste à ne pas modifier son niveau général d'équilibre. Il sera bon toutefois de procéder par étapes, de manière à tirer le maximum d'enseignements des dispositifs qui pourraient être successivement réalisés. A plus forte raison, un projet qui viserait à une mise en exploitation avec injections intensives, intéressant un grand nombre de partenaires, nécessiterait-il un examen beaucoup plus poussé, et éventuellement de nouvelles simulations mathématiques.

RISE OF PRIMARY PORE WATER (CONNATE WATER) IN COMPACTING SEDIMENTS

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Introduction

Transport of pore water and its dissolved constituents is generated or influenced by different physical and chemical processes. In hydrogeology, generally two mechanisms are considered: *1)* flow induced by pressure gradients due e.g. to differences in groundwater levels, and *2)* hydrodynamical dispersion of solutes.

If, however, deep aquifers in large sedimentary basins are explored, the additional effects of compaction or advective flow as well as ionic diffusion should be taken into account. These processes enable the migration of primary pore water or formation water resp. the transport of its solutes into higher strata independently from near-surface meteoric water circulation. This mechanisms may be especially interesting in basins which have been exposed to changing environmental conditions during deposition, e.g. transition from marine to fresh water sediments or evaporites. Such changes occurred in many large sedimentary basins on the continents.

In this study only the effect and range of compaction flow is dealt with. Compaction takes place during sedimentation, but it may be maintained also after the filling up of a basin as long as compaction equilibrium is not yet established. In thick sedimentary sequences formation pressures considerably higher than hydrostatic pressures have been reported (e.g. Marsal and Philipp 1970, Engelhardt 1973, Rieke and Chilingarian 1974). These pressures are explained by low-permeability layers strongly delaying the escape of pore water under the overburden load. Here, some compaction flow may still be in progress. In the following study, however, chiefly conditions of compaction equilibrium are considered.

Deductions from models

To determine the range of compaction flow for varying porosity/depth relationships a graphical method is used. The base layer of the sedimentary column is assumed to be solid and impermeable. Compaction flow moves

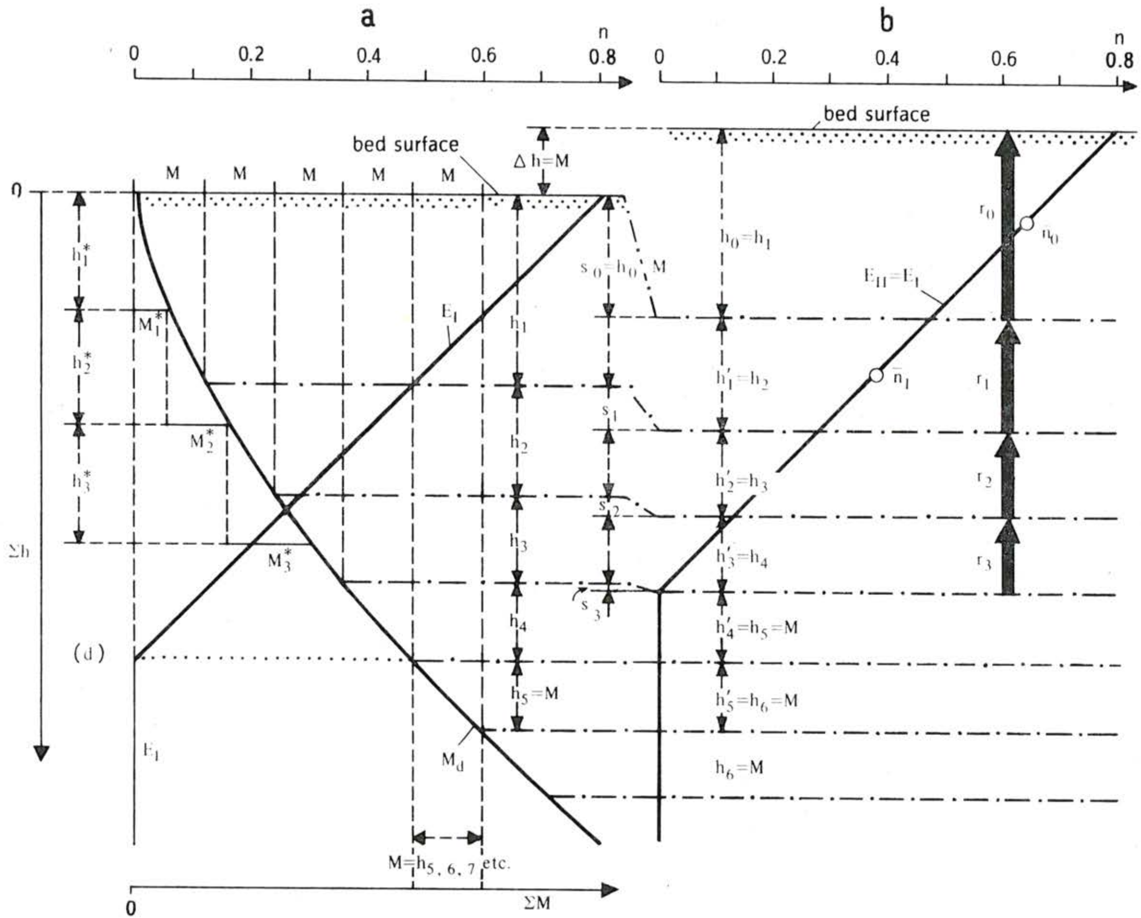

Fig. 1. Upstreaming pore water r in a model with straight-line porosity/depth relationship E and porosity n dropping to 0

upwards vertically. Therefore, refracted flow lines occurring in tilted beds of different permeabilities are not discussed.

Model A (Fig. 1): In a first approach a simple model is introduced. It is assumed that the porosity/depth relationship E_I is described by a straight line. This starts at the sedimentary surface with a porosity $n=0.8$ and leads to $n=0$ at some depth Σh. According to ENGELHARDT (1973) the sedimentary column is then subdivided into thin layers of equal height $h_1^*, h_2^*, \dots h_i^*$. These layers contain different volumes of pore space and compact mineral substance. The mineral substance of the individual layers is here expressed by the heights $M_1^*, M_2^*, \dots M_i^*$. If now the total heights h^* and the heights M^* of mineral substance of the single layers are summed up and their relationship is plotted on the $\Sigma M/\Sigma h$ diagram (Fig. 1), curve M_d is found.

After that, on the ΣM-axis discretionary equal heights M of compact mineral substance are plotted. Then, by means of the $\Sigma M/\Sigma h$-relationship (curve M_d), the heights of single layers h_1, h_2 are determined, which contain the same height of mineral substance. In other words, the layers h_2, h_3 etc. have been formed by compaction of layer h_1. During this process, only pore water has been squeezed out, but no mineral substance has been lost to overlying layers. In layer 5 and all deeper layers of Fig. 1, the porosity is $n_5=0$ and hence $h_5=M$.

Note: In this study heights of layers or mineral substance of layers etc. instead of volumes per unit area of a horizontal plane are used. For this reason, e.g. also certain "heights" of pore water are squeezed out which, however, can be easily turned into units of volume.

In the second step a new increment of sediment with the mineral substance M is deposited on top of layer h_1 (Fig. 1b), and the same equilibrium of compaction with a straight line porosity/depth relationship E_{II} as before E_I is established. When the growth of sediment has reached the height $\Delta h=M$ over the initial sedimentary surface, the same 4 layers on top of the completely solid sediment can be distinguished as before: h_0 corresponds to h_1, h_1' to h_2, etc., and h_4' is compacted to $h_5=M$.

By the new increment of sediment the initial top of layer h_1 has been lowered by the difference h_0-M and deeper border lines of layers by $h_1'-M$, $h_2'-M$, etc. To enable this lowering, the same heights of pore water (without mineral substance) have to be squeezed out from the underlying sediment, e.g.

$$\begin{aligned}
&\text{through the base of } h_0: s_0=h_0-M=h_1-M && (1)\\
&\text{through the base of } h_1': s_1=h_1'-M=h_2-M,\\
&\text{through the base of } h_2': s_2=h_2'-M=h_3-M, \text{ etc.}
\end{aligned}$$

In a medium of 100% porosity ($n=1$), the heights s signify the pore water movement relative to the top of individual layers the water comes from. In order to calculate the range r of pore water flow in the interstices of the sediment, the mean porosity $\bar{n}$ within the zones of upstreaming pore water has to be considered:

$$r_0=\frac{s_0}{\bar{n}_0}=\frac{h_1-M}{\bar{n}_0}\,; \qquad r_1=\frac{s_1}{\bar{n}_1}=\frac{h_2-M}{\bar{n}_1}\,; \text{ etc.} \qquad (2)$$

In general, but not in the case of model A, the zones of ascending pore water according to equation *(2)* are thinner than the corresponding layers analysed. For this reason, $\bar{n}_0$, $\bar{n}_1$, $\bar{n}_i$ often become smaller than the mean porosities of the layers h_0, h_1, h_i.

By this calculation it is possible to get quantitative data, how far the pore water, squeezed out from different layers by an additional weight (layer with the height h_0), has migrated upwards relative to boundaries of settling layers (Fig. 1). The pore water movement r_0 is generated by water from the underlying layer h_1'. Therefore, water included in the new increment of sediment h_0 is replaced by older pore water and forced to flow back into the overlying water body. Since sedimentation as well as compaction often proceed continuously, the new pore water of very small new increments can also perpetually be substituted by upstreaming pore water. Hence, in model A no new water from above can enter the sediment, as soon as the porosity curve E has dropped to zero at some depth.

This result can be demonstrated also by Fig. 3a, where the squeezed-out pore water is represented by the area on the left from the E_{I} and E_{II} lines. This area remains constant for different heights of the total sedimentary column.

The total or partial replacement of new pore water at the bed water interface is neglected in the further discussion. The following statements refer to pore water from layers below the uppermost increment.

Model B (Fig. 2). The porosity/depth relationship E is now represented by a curve which does not fall back to zero as in model A, but ends at some depth with a certain value $n > 0$, and then remains constant. After following the same procedure as in model A and replacing M in the equations *(1)* and *(2)* by the height h_f of a firm layer not any more affected by compaction, it is evident that the pore water movement r starting at the base of different layers does not reach the sedimentary surface or the top of these layers any more. Under these conditions the upstreaming pore water of all the earlier deposited layers is "buried" by the new layer. The same can be expected for all porosity curves in natural sediments, which do not go back to zero. From considerations similar to those in this study, FÜCHTBAUER (1974, p. 5–6) got the same result.

A similar situation as in model B is also shown by Fig. 3b. The porosity curve now, however, does not drop to a constant value at some depth. In Fig. 3b the area representing pore water squeezed out between the porosity curves E_{I} and E_{II} is smaller than the area of pore water necessary for the new increment Δh of sediment. Furthermore, one can see that the range r of compaction flow depends on the depth of a solid base (hardrock, e.g. basalt) below the sediment-water interface or on the height of the total sedimentary column. In general, thick deposits enable a higher range of upstreaming pore water than thin ones, because the porosity at their base becomes small and a high column of sediment is affected by pore space reduction.

If in model B further increments Δh_2, Δh_3 both equal to Δh_1, are added (Fig. 2), a water particle A having moved by the increment Δh_1 from the initial interface to B, now ascends by the increment Δh_2 from B to C, and by Δh_3 from C to D. Similarly, the path of deeper situated water particles can be traced. These movements are relative to a fixed datum line or to the solid base of the sequence. The arrows r, on the other hand, always show the range of pore water flow relative to a certain marker bed within the sequence. A third possibility is to describe the position of a water particle relative to

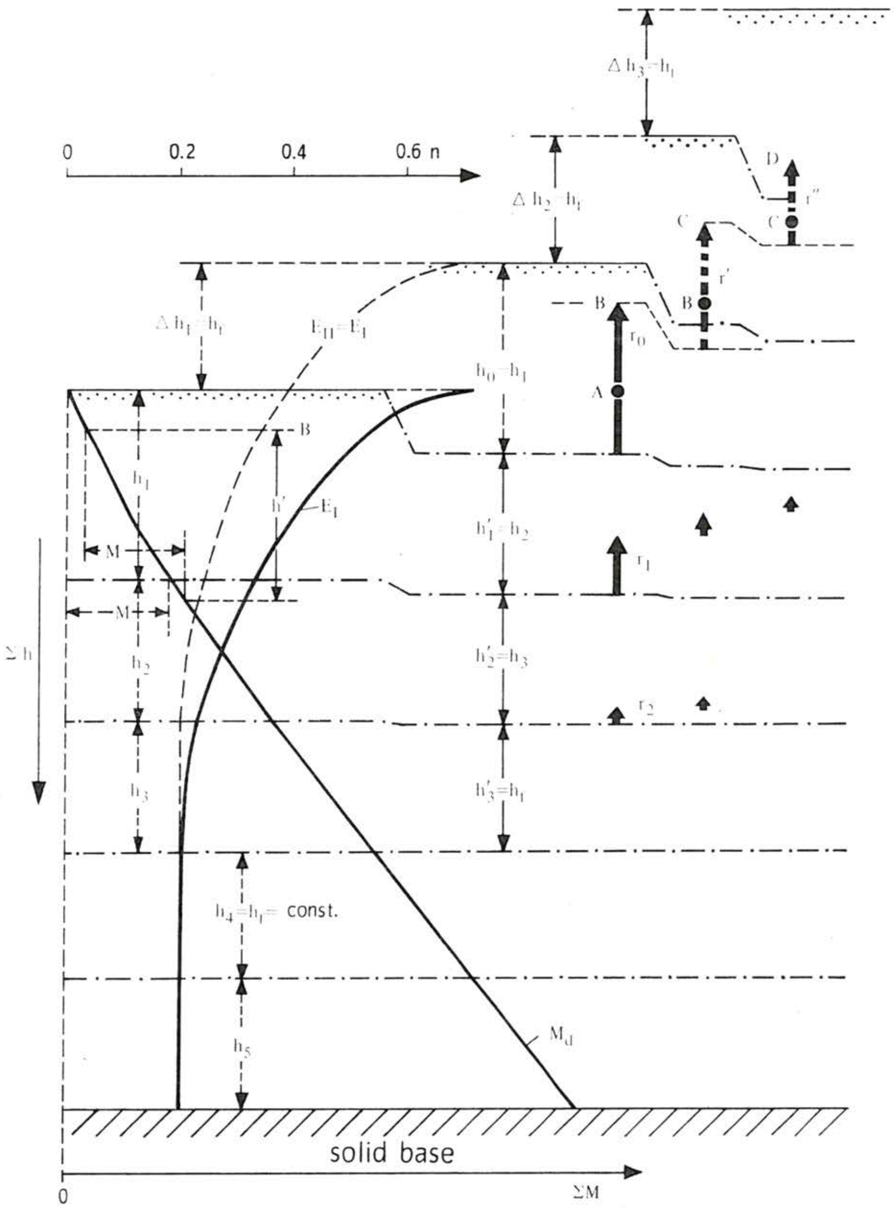

Fig. 2. As Fig. 1, but porosity/depth relationship represented by curve E, which does not fall back to zero

Equilibrium of compaction is maintained during deposition of new increments (layers h_0 or growth of the sequence by Δh_1, Δh_2, Δh_3); for further explanations see text

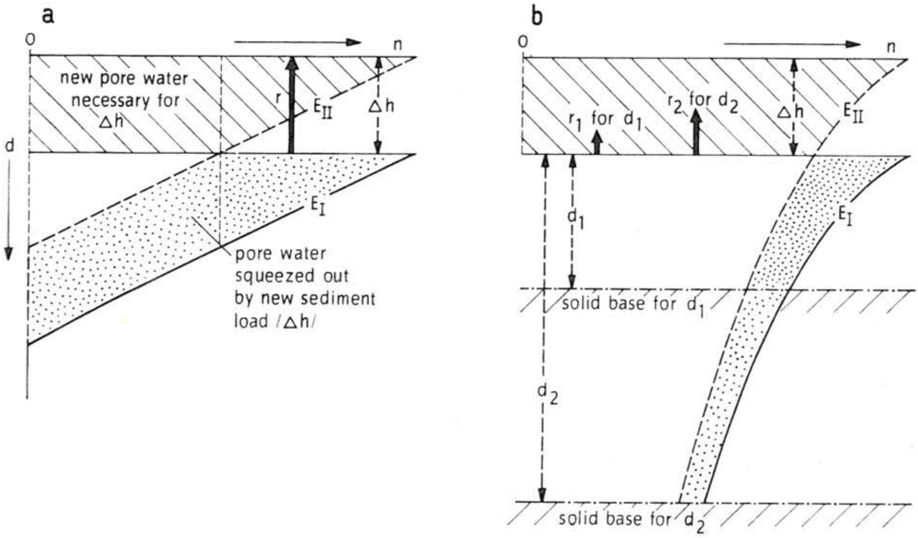

Fig. 3. Scheme of porosity/depth relations with loss and need of pore water below resp. above a former sedimentary surface, *a)* with straight-line porosity after deposition of Δh, *b)* with porosity curve and different thickness of total sediment d_1 and d_2 after deposition of Δh

the upward-moving sediment-water interface (definition of advection by Lerman 1975). Then the water particle in Fig. 2 moves downward from the changing interface.

Further models dealing with the compaction flow of under-consolidated sediments or with the affects of occasional erosion on top of a growing sequence are described in Einsele (1977). Besides under-consolidation also diagenetic processes such as dissolution or recrystallisation of mineral substances can initiate compaction flow. If the reduction of porosity generated by these processes can be estimated, the range of vertical ascending pore water is found similarly as described above.

Pore water movement in sedimentary sequences

The porosity curves of the following examples are taken from the literature. They represent normal cases of different sediment types. In the calculation of compaction flow the influences of under-consolidation, recrystallisation or dissolution as mentioned above are neglected, though e.g. Morgenstern (1967) has shown that some under-consolidation may occur in clayey muds even under relatively low sedimentation rates. The permeability necessary for maintaining normal consolidation during continuous sedimentation is discussed in detail by Bredehoeft and Hanshaw (1968, cited in Rieke and Chilingarian 1974).

For the assumption that normal consolidation or compaction equilibrium has been realized all the time, the analysis of porosity/depth relationships by the graphical method renders minimum ranges of ascending pore water. This is the case for the following examples. Furthermore it must be emphasized that the porosity curves of Figs. 4 and 5 are strongly generalized neglecting the influence of layering and of many irregularities common in nature.

Fig. 4 shows the porosity curve from site 178 of the Deep Sea Drilling Project, which was performed in the Alaskan Abyssal Plain in 4218 m water depth (Huene et al. 1973). In this graph the laboratory porosity measurements on samples from several sites are combined. Therefore, according to the authors, the curve represents the best approximation to the actual in situ curve for North Pacific distal deep-sea turbidite sequences. The first thin silt and fine sand turbidites occur in the Pliocene on top of a Miocene deep-ocean pelagic sequence of comparatively reduced thickness (Kulm, Huene et al. 1973).

In a second example (Fig. 5) a generalized porosity curve E combining data from McCulloh (1967) and Ham (1966, cited and compared with other data in Rieke and Chilingarian 1974, Fig. 17) is analysed. This curve represents the maximum probable porosity for many clastic sedimentary rocks (mainly argillaceous), but not the upper limit of porosities found at certain depths. Furthermore it is assumed that the sequence consists of three principal layers A, B, and C, all containing the same height (or volume per unit area) of solid mineral substance, i.e. $M_A = M_B = M_C$. Consequently, the initial heights h_A, h_B, h_C of those layers are taken to have been equal, as long as they were not buried by younger deposits. In the final stage of deposition, the deeper layers B and A are compacted to the heights h'_B respectively h''_A, ($h_C > h'_B > h''_A$), and corresponding heights of pore water are squeezed out.

If during deposition the environmental conditions have changed from fresh water sediments *(A)* to marine deposits *(B)* and back to fresh water deposits *(C)*, the ascending pore water or formation water passes the environmental boundaries of the solid part of the sediments. After deposition of layer B (surface 2) the original sedimentary surface 1 is lowered to 1′ and the fresh pore water from layer A replaces part of the marine pore water of layer B. By deposition of layer C, the former surfaces 1 and 2 are depressed to 1″ and 2′. Simultaneously, the upper limit of fresh pore water from layer A is further raised to such an extent that, in this example, marine formation water of layer B is entirely expelled from its original layer *(B)* and pushed into the uppermost layer C.

Note 1: The further rise of the upper limit of pore water from layer A (in depth d below surface 2) is found by a similar graphical procedure as explained for model B (Fig. 2). By means of the curve M_d a layer with the mineral substance M_A is plotted in depth d below the final surface 3. The height of this layer becomes h', and the further rise of the fresh pore water limit results from calculation and plotting of the range r'_C according to equation *(2)*.

Note 2: Due to higher density or, in special cases, caused by abnormal fluid pressure gradients, the marine pore water of layer B may locally or over a larger area also move downward and force the underlying fresh pore water to escape laterally or upward at other places. E.g. in the northern Alpine Molasse Basin such a case has been reported for tertiary marine deposits on top of fresh water molasse and karstified Jurassic limestones (Lemcke and Tunn 1956). Such mechanisms are not further discussed here.

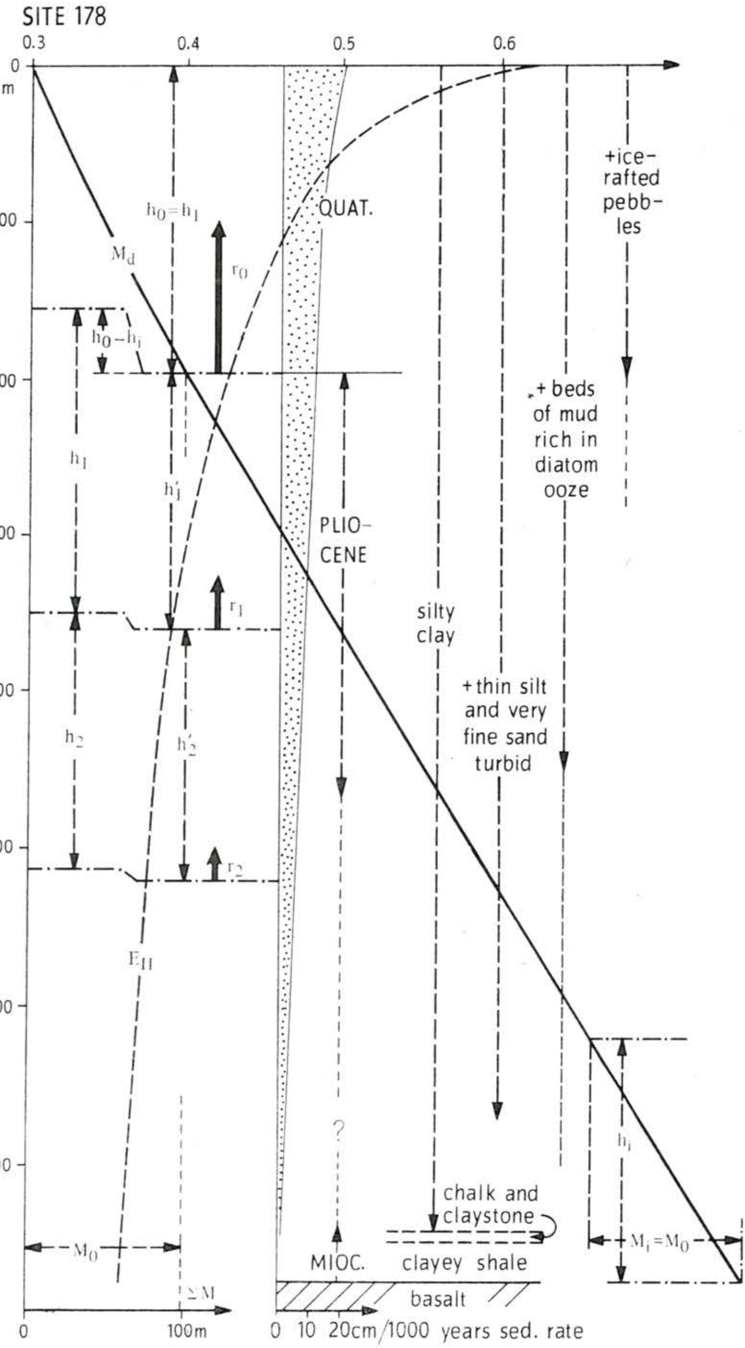

Fig. 4. Pore water flow r induced by deposition of the Quaternary on top of a distal turbidite sequence, Site 178, Alaskan Abyssal Plain

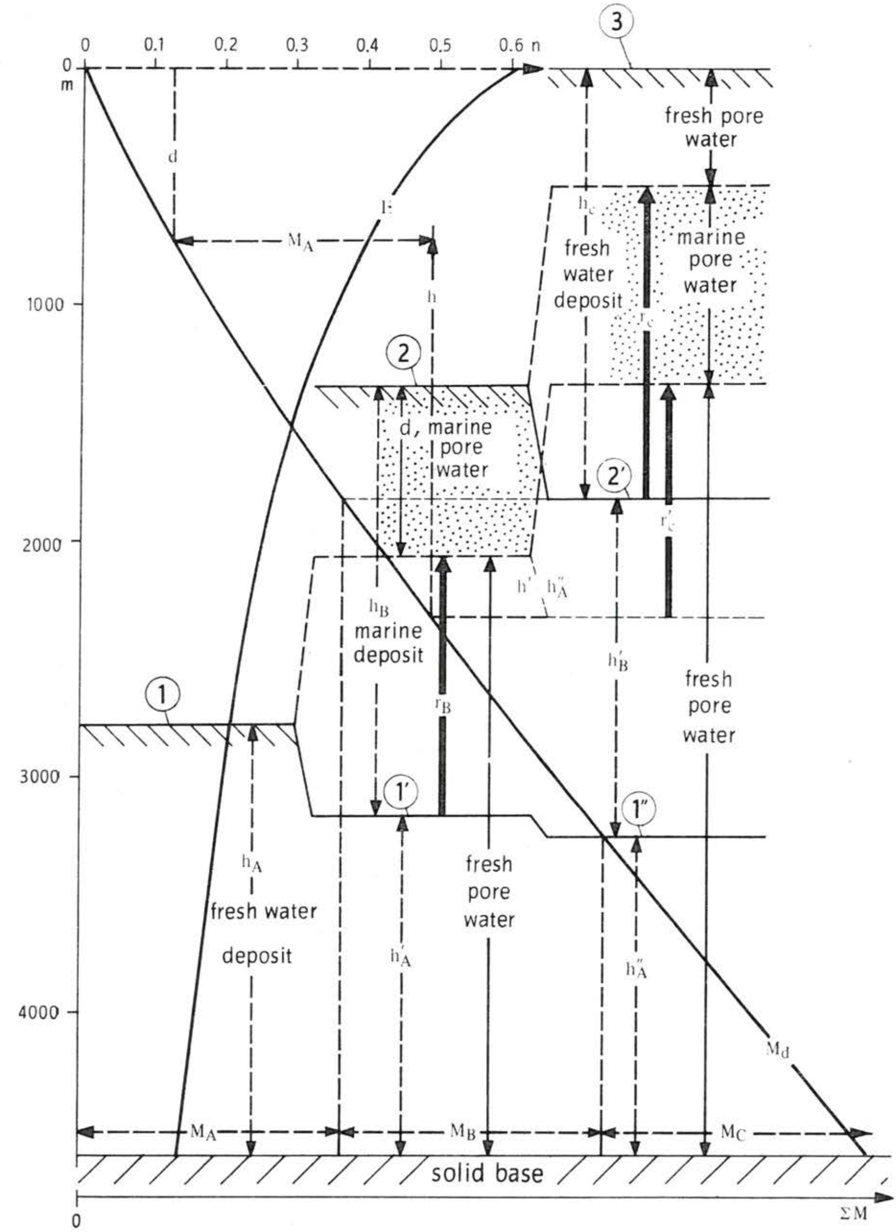

Fig. 5. Hypothetical three-layer sequence with alternating fresh water and marine sediments in 3 stages of deposition, settling of sedimentary and rise of pore water boundaries by compaction. The porosity curve corresponds to argillaceous sediments; for further explanations see text

In c o n c l u s i o n, the study of the three-layer example demonstrates the following trends:

1) The pore water of the lowermost compressible layer of a sequence, exerts a strong influence on the total sequence (no loss, high rise of pore water).

2) In the medium layer, the uptake of pore water from its depositional environment is reduced by ascending pore water from the underlying layer. This effect is still stronger-developed in the uppermost layer.

3) The original pore water of the medium and uppermost layer can partly or entirely be replaced by pore water from the underlying layer *(s)*. Therefore, e.g. fresh water deposits are contaminated by marine formation water due to compaction flow.

Results and conclusions

1) The range of compaction flow in growing sedimentary sequences depends on the difference between the initial porosity at the bed-water interface and the porosity at the base of the sequence and, hence, in general on the total thickness of the sedimentary column (Fig. 3b). Only from the uppermost increment of sediment, new pore water being continuously replaced by upstreaming compaction flow can partly be driven back into the overlying water body.

2) The range and velocity of upward moving pore water caused by a certain increment of newly added sediment decreases from top to bottom of a sequence. The velocity relative to the boundaries of initial layers is proportional to the sedimentation rate, but in general is smaller than this value (for further treatment of this problem see EINSELE 1977).

3) By compaction the solvent itself is moved together with its solutes. From velocity and concentration of certain chemical species mass transport of solutes can be calculated (EINSELE 1977).

4) Because during deposition no pore water from deeper layers is lost to the overlying water body, the pore water of deeply buried sediments exerts a strong influence on the total sequence. This is important for alternating fresh water and marine deposits. The uptake of new pore water in younger layers is successively reduced by compaction flow, and the initial pore water of higher layers is partly or entirely replaced by ascending pore water. The shifting of pore water boundaries in relation to the originally alternating environmental boundaries of sediments can be determined quantitatively.

5) It is to be expected that such pore water boundaries still exist in parts of sedimentary basins which are not or only slightly affected by diffusion (see point *7*) and/or meteoric water circulation. The chemistry of such "formation waters" or connate waters, however, is often considerably changed by reactions between the original pore water and the solid phases of the sediment (see e.g. HÖLTING 1970, HAHN 1972, ENGELHARDT 1973, CARLÈ 1975). Especially in old deposits often influenced by a varying geologic history, the situation may become too complicate by different paleohydrogeological events (see e.g. MICHEL et al. 1974) to enable the recognition of former pore water boundaries.

6) In contrast to mass transport by diffusion, compaction flow or advective flux is always directed upwards. It is most important during high rates of deposition and generally ceases with the end of depositon. Compaction flow, however, can also be maintained at a small rate after deposition as long as compaction equilibrium is not yet reached or when the porosity is secondarily reduced by diagenetic processes. In order to determine quantitative data, the initial and secondary porosity-depth relationship must be known.

7) Mass transport by diffusion is not treated in this study. It may be only mentioned that diffusional flux depends on concentration gradients and is, therefore, independent from certain directions or the sedimentation and its end. A comparison of the quantitative influence of advective flux (compaction flow) and diffusional flux on mass transport of solutes within sediments is not possible in a general way. Many authors, most of them being geochemists, hold the opinion that material transport by diffusion is the principal mechanism (e.g. ANIKOUCHINE 1967). BERNER (1975), GIESKES (1975), or LERMAN (1975), however, they emphasize that advective flux becomes increasingly important with growing sedimentation rate.

Whereas according to this study and EINSELE (1977) advective flux can be determined rather precisely, diffusional flux often appears to be less known because of highly variable diffusion coefficients and other uncertainties (see e.g. TZUR 1971, BERNER 1971, GIESKES 1975). This especially applies to low-porosity sediments in greater depth, which often show alternating layers of well and less permeable material. Here, the calculation of diffusional flux becomes problematic. Furthermore, filtering of ions by semi-permeable layers may become an important factor (RIEKE and CHILINGARIAN 1974). The occurrence of connate water in many ancient sedimentary basins proves that diffusional flux often was not able to remove strong concentration gradients. The same is true for the fresh water/salt water boundary in many areas which are exploited for groundwater.

Therefore, a general comparison between mass transport by advective of diffusional flux hardly makes sense. This question has to be answered separately for the special conditions of each sedimentary basin and its various dissolved chemical constituents.

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HYDRODYNAMICS OF THE HUNGARIAN BASIN

HYDRODYNAMIQUE DU BASSIN HONGROIS

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RÉSUMÉ

On peut distinguer deux étages structuraux dans le grand bassin hydrogéologique. L'étage topographique supérieur est constitué par des cuvettes sédimentaires pliocènes et l'étage topographique inférieur se trouve dans les fossés d'effondrement quaternaires de la Plaine.

La région d'écoulement régionale du bassin est alimenté par les eaux souterraines provenant des montagnes se situant autour et à l'intérieur du bassin. L'alimentation des nappes d'écoulement intermédiaire des sous-bassins a lieu d'une part du côté des montagnes, d'autre part sur les terrains élevés dans le bassin. Ceux-ci sont constitués par les dépôts quaternaires à forte perméabilité. Les nappes d'écoulement de la Grande Plaine Hongroise communiquent entre elles, malgré que les dépôts du Pliocène supérieur renferment des nappes presque captives en continuité.

La répartition des niveaux potentiométriques représentent les caractères du dynamique des eaux souterraines du bassin.

Les terrains caractérisés par les hautes gradients verticaux positifs coïncident avec les points des coupes, où les courbes équipotentielles sont parallèles aux caractères géologiques horizontaux (par ex. la base des couches quaternaires) indiquant les eaux souterraines profondes stagnantes. L'autre preuve des eaux souterraines stagnantes est la courbe isopièze fermée de 85 mètres et le très bas gradient horizontal dans les dépressions profondes.

Les dépôts à forte perméabilité constituent des zones profondes de recharge à côté des bordures des régions d'émergences ou de décharges indiqués par les courbes équipotentielles élargissantes ou par des courbes isopièzes verticales représentant des conditions hydrostatiques.

Les basses plaines sans drainage sont des sites importantes de décharge. Les sous-bassins sont très rarement drainés par les fleuves où les lits des fleuves s'enfuissent dans les dépôts superficiels imperméables. La part de l'alimentation de la région d'écoulement intermédiaire « flotte » au-dessus d'une région d'écoulement thermale.

Une zone peu profonde des gradients verticaux négatifs dans la région des gradients positifs coïncide toujours avec l'endroit où *(a)* les couches sableuses aquifères peu profondes sont en affleurement entouré par des terraines imperméables et *(b)* le fleuve s'enfouit dans les terrains imperméables. Dans tous les deux cas un équilibre dynamique existe entre les eaux d'infiltration ou le drainage par un fleuve et entre les eaux profondes salées ascendantes. S'il n'y a aucun drainage par un fleuve ou par des sables en affleurement il peut résulter une accumulation de sel dans la partie supérieure des cuvettes sédimentaires des sous-bassins, dans des conditions arides.

The major advance in hydrogeology in the past decade has been a consequence of the application of physical principles in hydrology. The basic work on the theoretical principles of a groundwater flow-system was written by HUBBERT in 1940. The next major advance started when TÓTH presented his computer-developed models in 1962 and 1963 defining the local, intermediate and regional flow-regions.

Investigations on groundwater-flow carried out in the last decade have cleared the essentials of the hydrodynamics of regional flow-systems (MEYBOOM, WINOGRAD, FARVOLDEN, MAXEY, etc.).

Characteristics of a multi-layer flow-system

The character of a flow-region is shown by:

a) the pressure conditions, which are different in discharge areas and recharge ones;

b) the chemical composition of the groundwater;

c) the geothermic anomalies: which may indicate the upward or downward movement of water;

d) natural isotopes.

The various aquifers building up an artesian system are dynamically connected because the confining beds are practically more or less pervious. Water may move from an aquifer into another due to different heads prevailing in them.

In a multi-layer flow-system there are three kinds of vertical pressure patterns:

1. Pressure within the same interconnected formation shows a more or less uniform increase with depth, *higher* than that of the hydrostatic pressure. Where the energy potential increases with depth, there is an upward movement of water through the confining strata.

2. There are areas where head in the same hole is *less* in a deeper than in a shallow aquifer. The downward increase of pressure is less than hydrostatic.

The decline in pressures with depth shows that a confined aquifer is recharged by the downward movement of water through the overlying confining bed.

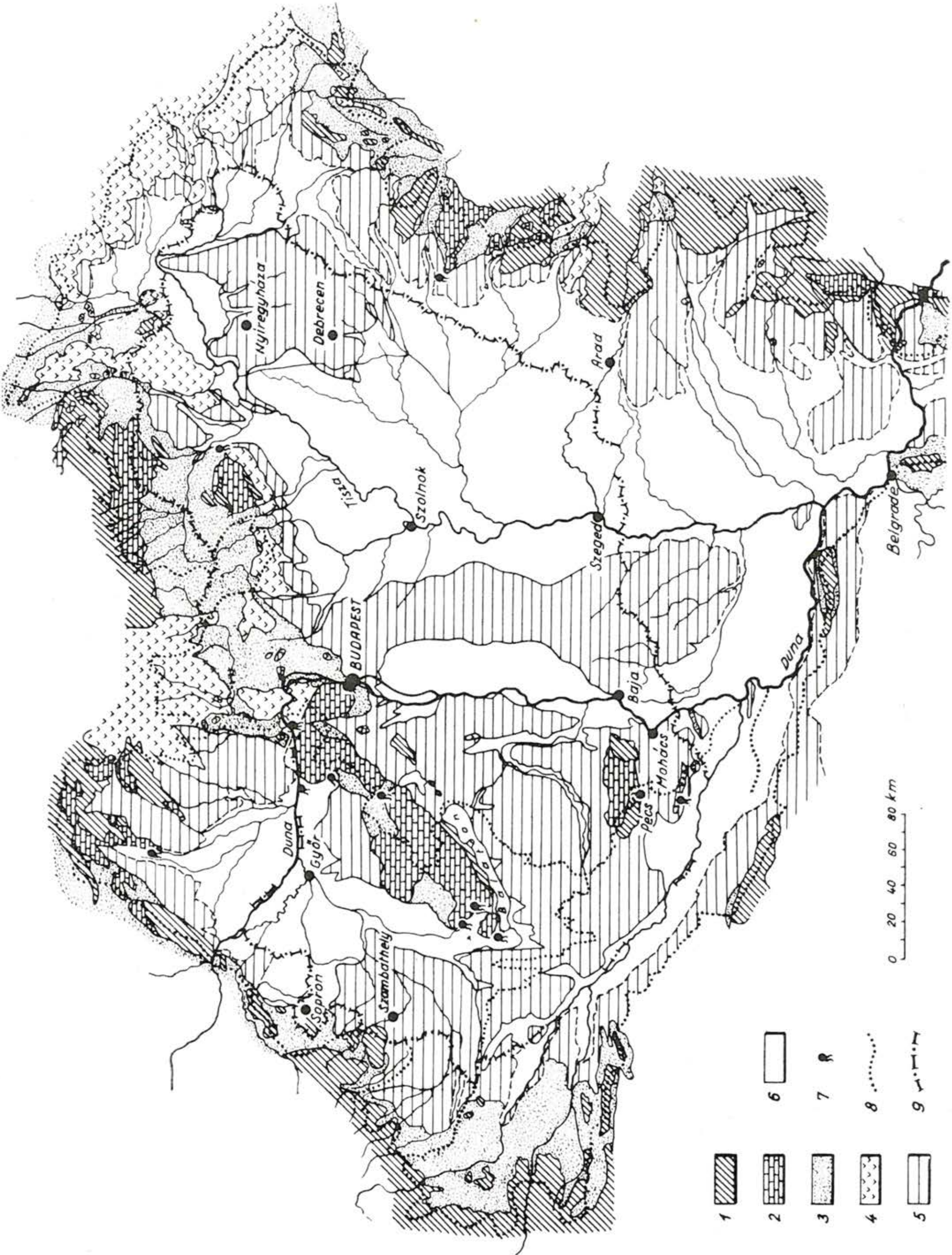
It may also be an explanation that the deeper aquifer is much more permeable than the shallow one (TÓTH 1970). In this case each aquifer has its own piezometric surface independent of the other. A further explanation may be that a lower aquifer is sealed off by an unconformable overlapping impermeable formation. It is also a possibility that the elevation of the recharge area of the deeper aquifer is less than that of the shallow one.

3. Neutral pressure occurs in the zones between areas of vertically increasing and decreasing heads. Here the downward increase of the pressure is hydrostatic.

Outlines of the Hungarian Basin's hydrogeology

The Hungarian Basin is a large and non-uniform flow-system and structurally a deep depression formed mainly in the late Tertiary and Quaternary.

A short description of Hungary's hydrogeology is given here for the better understanding of the details.



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Tisza
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The geology of the bordering mountainous areas (Alps, Carpathians and Dinarides) and the mountains inside the great central basin between the Little and Great Hungarian Plains can be described in brief as follows (Fig. 1).

a) *The Paleozoic basement* consists of crystalline rocks exposed not only in the mountains around but also within the central basin.

b) *Karstic Mesozoic limestones and dolomites* occupy large areas around and inside the basin, the majority of them being of Triassic age.

c) These basement complexes are overlain by diverse *cover formations*, which are briefly described as follows:

1. Deposits of excellent permeability: post-Liassic hard fractured Mesozoic limestones, Nummulitique limestones of the Middle and Upper Eocene and limestones of the Middle and Upper Miocene.

2. Medium to poorly permeable deposits: thick sandy marine sequences of the Upper Oligocene and Lower and Middle Miocene (schlier); Lower Eocene and Middle Miocene continental and littoral deposits and coal measures. Sandstones and flysch deposits of great variety; continental, deltaic and littoral beds of the Pliocene outside the central depression in smaller intermontane basins.

3. Impermeable formations: a thick sequence of marine Mesozoic formations younger than the Middle Liassic, Upper Cretaceous continental clays, thick Upper Eocene and Middle Oligocene marine clays.

4. Volcanics.

The Hungarian Basin is a hydrological unit. The hydrogeological boundary runs along the surface divide connecting river gaps cut into impermeable bedrock (Fig. 1) where there is no subsurface flow. As a consequence the total amount of surface water flowing into the central basin can be assessed by river discharge measurements in the river gaps.

The boundaries of the subbasins are running generally along the surface contact between the Pliocene deposits and older formations.

The central basin can be divided into two structural storeys. The upper one consists of surface Pliocene deposits with or without a thin Quaternary cover. The lower storey is the area of the Quaternary subsidence (Fig. 2).

The hydrogeological structure of the two Hungarian lowlands is very similar. The Pliocene "thermal water aquifer" is capped by a practically impermeable formation almost all over the sub-basins thus separating the regional Pliocene flow-region from the overlying intermediate flow-region of the Quaternary (Fig. 3).

Fig. 1. Hydrogeologic map of the Hungarian Basin (ERDÉLYI 1973)

A. Recharge area of the intermediate and regional flow-regions: *1.* crystalline rocks (Paleozoic), *2.* karstic rocks (Mesozoic limestones and dolomites), *3.* sedimentary deposits of the Eocene, Oligocene and Miocene, *4.* volcanics (Neogene), *5.* Pliocene and Quaternary. *B.* Discharge area of the intermediate and regional flow-regions: *6.* Pliocene and Quaternary, *7.* high discharge artesian karstic spring, *8.* inner divides of the Hungarian Basin, *9.* international boundary

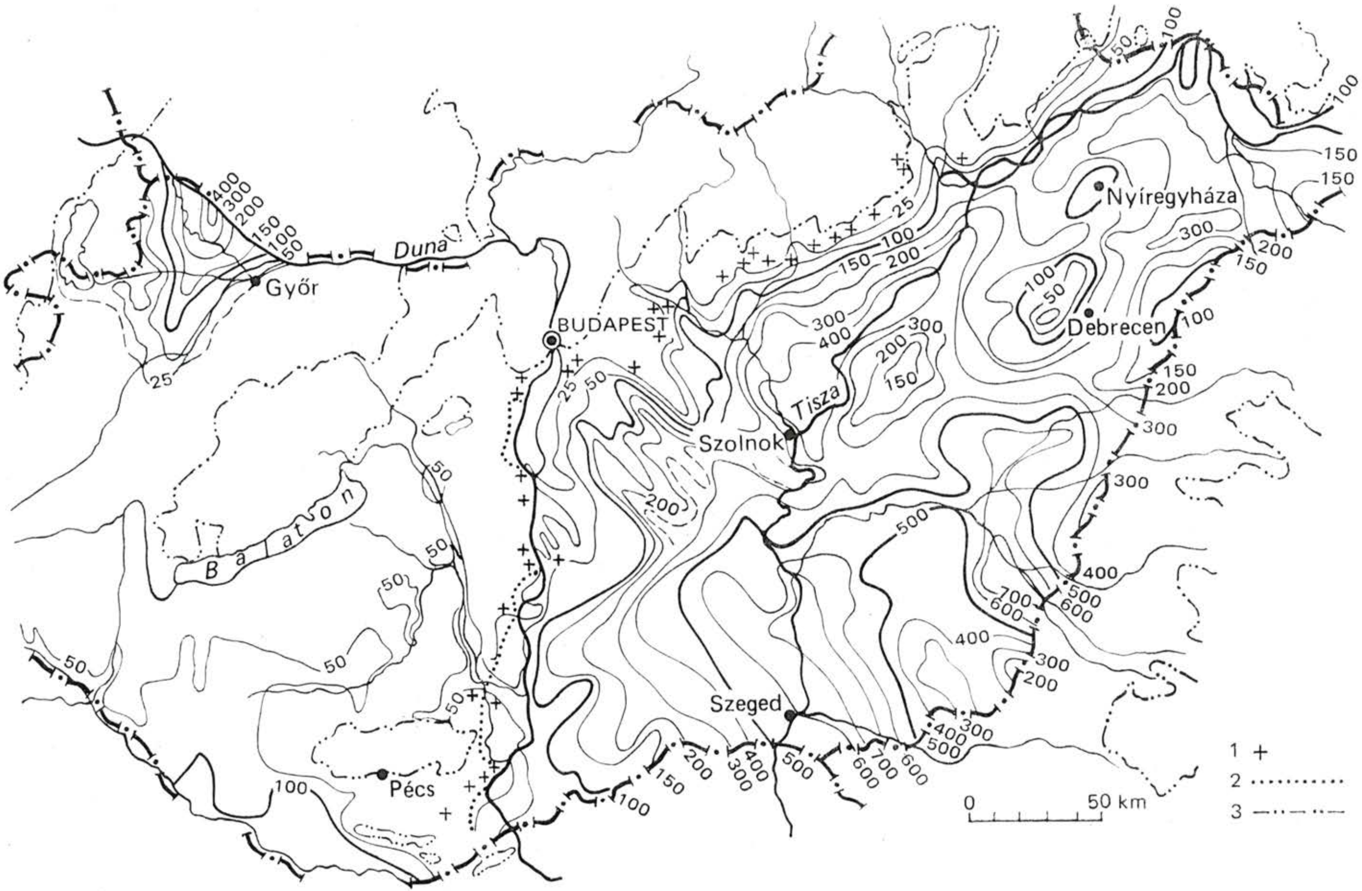

Fig. 2. Thickness of the Quaternary-type deposits in Hungary (Erdélyi 1973)
1. Outcrop of the Pliocene farthest from the boundary of the Great Plain, *2.* western boundary of the Great Plain, *3.* boundary of the mountainous area

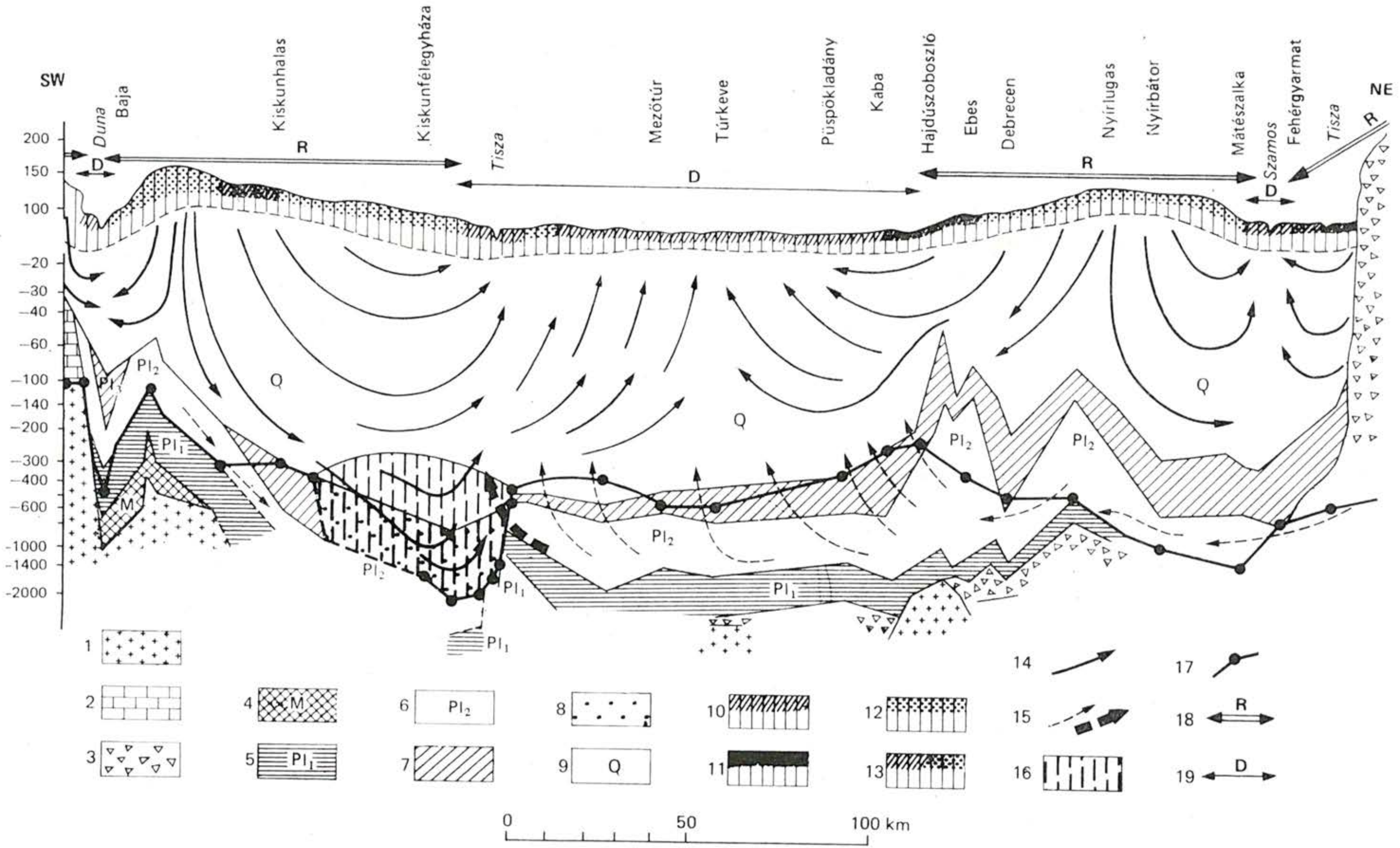

Fig. 3. Diagrammatic flow-pattern of the Great Hungarian Plain (Erdélyi 1973)

***Lithology and stratigraphy** (1–9): 1.* Crystalline basement, *2.* Mesozoic limestone, *3.* Miocene volcanics, *4.* Miocene sedimentary deposits, *5.* Lower Pannonian (Piliocene) impermeable or poorly permeable, *6.* Upper Pannonian (Pliocene) permeable, *7.* impermeable or poorly permeable top formation of the Pliocene, *8.* highly permeable Upper Pliocene, *9.* Quaternary. ***Surface deposits** (10–12): 10.* impermeable soils, *11.* moderately permeable soils, *12.* dune sand and loessy sand. ***Flow-regions** (13 –16): 13.* local, *14.* intermediate, *15.* regional, *16.* intermediate and regional. *17.* Fresh- and salt-water interface, *18.* recharge area, *19.* discharge area

Hydrodynamics of the Hungarian Basin

The results of the early research into the hydrodynamic conditions of the Hungarian Basin carried out in the 1950's were published in 1962 (Almássy in Schmidt's Hydrogeological Atlas of Hungary).

Piezometric conditions of the Hungarian Basin

Comparing the heads in the wells drilled a few decades earlier or before the intensive pumping had begun, the original water levels of wells not influenced by withdrawal were established and used for the plotting of piezometric profiles (Fig. 4). This was possible by favourable circumstances:

1. The number of drilled wells of known aquifer depth and with elevations established by geodetic survey was about 8200 at the end of 1970.

2. Packer-tests have been made in hundreds of boreholes traversing two or more aquifers.

3. A large number of drilled wells is situated close to each other in certain areas tapping different aquifers.

4. Electric logs of 5500 drilled wells were available at the end of 1971.

5. A detailed inventory of drilled wells (50,000 in 1971) is being continuously completed.

The vertical hydrodynamic gradient was obtained by dividing the difference in heads of two confined aquifers by the vertical distance between them in an exploratory borehole or in two closely situated drilled wells with exact elevations measured by geodetic survey.

Parts of the Hungarian artesian flow-system

Fig. 1 shows the recharge and discharge areas of the regional and intermediate flow-systems. The flow-system diagram of the Great Hungarian Plain is also given (Fig. 3).

The piezometric conditions shown in Fig. 4 refer mostly to depths not more than 400 m below surface and only exceptionally to greater depths.

1. *Unconfined water of the mountains and hills*

The mountains and hills around and inside the basin are characterized by an unconfined water table. The water table of the crystalline basement (Fig. 1) follows the topography. The karstic water table of the carbonate basement (Triassic to Lower Jurassic) is a deep-lying regional water table which becomes confined below impermeable or slightly permeable cover formations.

The volcanic hills and mountains are the recharge areas of the confined aquifers of adjacent valleys, of intermontane basins and of the margin of the main basin (Fig. 3).

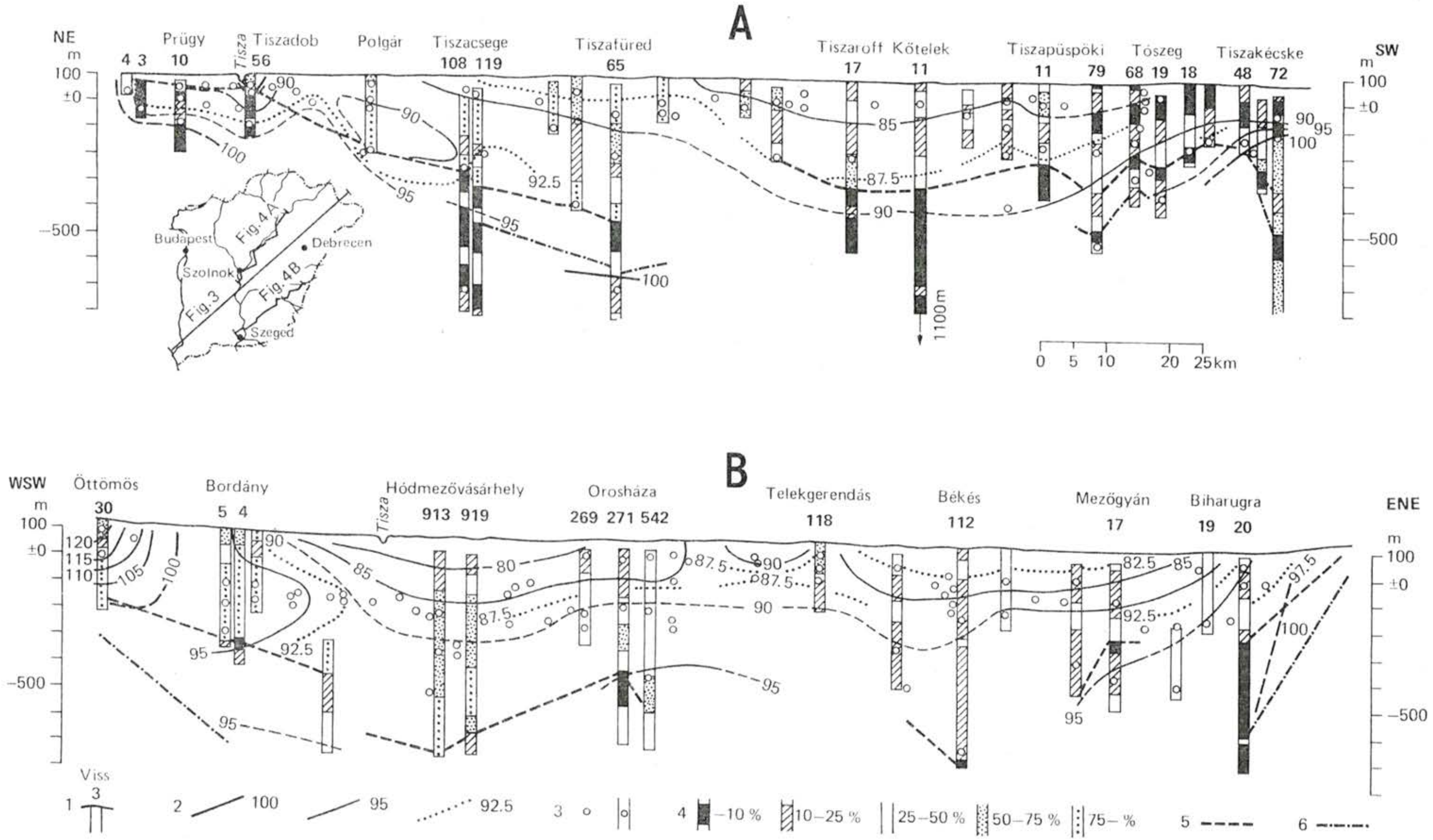

Fig. 4. Hydrodynamic cross-sections and lithologic logs (Erdélyi 1973)

1. Municipality and serial number of the drilled well, *2.* equipotential contours, *3.* middle of the screened portion of a well, *4.* sand percentage, *5.* base of the Quaternary-type deposit, *6.* base of the impermeable top-formation of the Pliocene

The Tertiary volcanics must be separated on the map from the outcropping crystalline basement because they are different from hydraulic aspect. The Tertiary volcanics are inhomogeneous because of the vertically alternating lava flows and tuffaceous beds.

In the mountains around and inside the central lowlands there are smaller and individually confined flow-systems.

2. *Areas of energy potential decreasing with depth*

Areas with decreasing energy potential can be found inside the basin in topographically distinct places:

a) A comparatively narrow belt more or less parallel to the foot of the mountains where the shallow artesian aquifers are exposed and are recharged by the unconfined groundwater.

The deeper confined beds are recharged partly through the fractured volcanic rocks of the mountains bordering the basin. The water of the shallow confined aquifers is chemically very close to the water of the Quaternary aquifers in the valleys. The deeper artesian aquifers contain water chemically very similar to the deep water encountered inside the volcanic mountains in deep mine shafts. The same chemical character can be found in the water of the springs of the adjacent volcanic areas. This is an evidence that the shallower artesian aquifers are mostly recharged directly from the surface or through surficial deposits whereas the deeper confined water-bearing beds are recharged at an increasing rate through the water-table flow-system of the mountains (Karácsonyi—Scheuer). Similar conditions have been found in and near the carbonate rock areas (Karácsonyi—Scheuer).

These examples and many more bear witness to the fact that the contact between the late Tertiary basin deposits and the carbonate and volcanic rocks of the mountains is mostly of tectonic nature along the zones of deep marginal faults.

b) The low undulating terrains and higher hills are built up of Pliocene deposits of continental, deltaic, littoral and lagoon origin in Transdanubia and along the boundary of the Great Plain. The Pliocene deposits are exposed on large areas and in small outcrops or capped by Quaternary sands, sand-dunes, loess and sandy loess of various thickness (5—55 m). The thick loess sequence of the southeastern half of Transdanubia is divided by 2—5 fossil soils of poor permeability which retard infiltration through the loess. There is no general downward increase in the permeability of the Pliocene beds contrary to the Quaternary alluvial fans on the topographic highs inside the basin between the rivers Danube and Tisza and the northeastern part of the Great Plain.

c) Areas of low undulating topography in the southern part of Transdanubia covered by dune sand, sandy loess and occasionally by impervious swamp deposits underlain by a Quaternary alluvial sequence of a thickness between 20 and 100 metres. This sequence consists mostly of thin beds of medium- and fine-grained sands separated by clay and silt deposits.

d) The most important areas where the energy potential is declining with increasing depth are the topographic highs inside the central basin. These areas are mostly underlain by thick Pleistocene alluvial deposits and covered in more than 80 per cent by dune sand, thin loessy sand and sandy loess, all highly permeable. These areas are bordered wholly or mostly by Upper Quaternary subsidence areas in which energy potential increases with depth.

Packer-tests carried out in pilot boreholes on areas with declining potential showed that a deeper lying coarse-grained fluviatile stratum or cavernous carbonate aquifer has a much lower head than expected. Such conditions have been found in areas adjacent to the karstic mountains and on the topographically high parts of the Quaternary alluvial fans. This extremely great decline in pressure can probably be explained by CASAGRANDE's reflection law. In both examples mentioned previously the permeability of the lower-lying aquifers is many times higher than that of the overlying water-bearing horizons, consequently their piezometric surfaces dips in very low angles.

The piezometric surface of the deep Upper-Pannonian thermal water aquifer (regional flow-region) is not a subdued replica of the surface contrary to those of the overlying two Pleistocene aquifers (intermediate flow-region) but follows the general inward slope of the surface of the Great Hungarian Plain.

This can be explained simply because the Quaternary and Upper Pannonian aquifers are separated by an extended Upper Pliocene formation ranging from impervious to poorly permeable deposits, and the deepest aquifer becomes unconfined at the edge of the basin.

The unfavourable drainage conditions may partly be due to a tectonic process (subsidence) also affecting other areas with high sand percentage. Possibly the major factor accounting for the unfavourable drainage conditions has been the highly different rate of compaction between the areas of low and high sand percentage with the same amount of subsidence.

The sand percentage was calculated from the total thickness of the clean sand and gravel beds of an individual thickness over 1.5 m.

The recharge areas of the artesian flow-system inside and around the Hungarian Basin are bordered by more or less narrow belts of lateral flow in the Quaternary formation which shows that there is no difference in head between the Lower or Upper Quaternary aquifers, the energy potential does not vary with depth. There are smaller recharge areas fairly inside the Great Plain. These are narrow strips of sands and sand-dunes isolated or merging into sandy patches as islands in impermeable or poorly permeable surroundings. The sands and dunes are actually outcrops and thereby recharge areas of permeable alluvial deposits underlying the surface deposits.

3. *Areas of artesian energy potential increasing with depth*

The lowest areas of the Hungarian Basin are the areas of typical artesian flow. The surface deposits are mostly loess-like silts and the topographically deeper areas are covered by completely impervious alkali soils (solonetz and solontchak). A small percentage of the area is sandy. The sands are outcrops of water-bearing alluvial deposits.

There are occurrences of gaseous artesian water over large areas of the Great Plain (Fig. 5).

The discharge areas of the Great Hungarian Plain occupy about half of the area of the basin inward the inner recharge belt adjacent the bordering

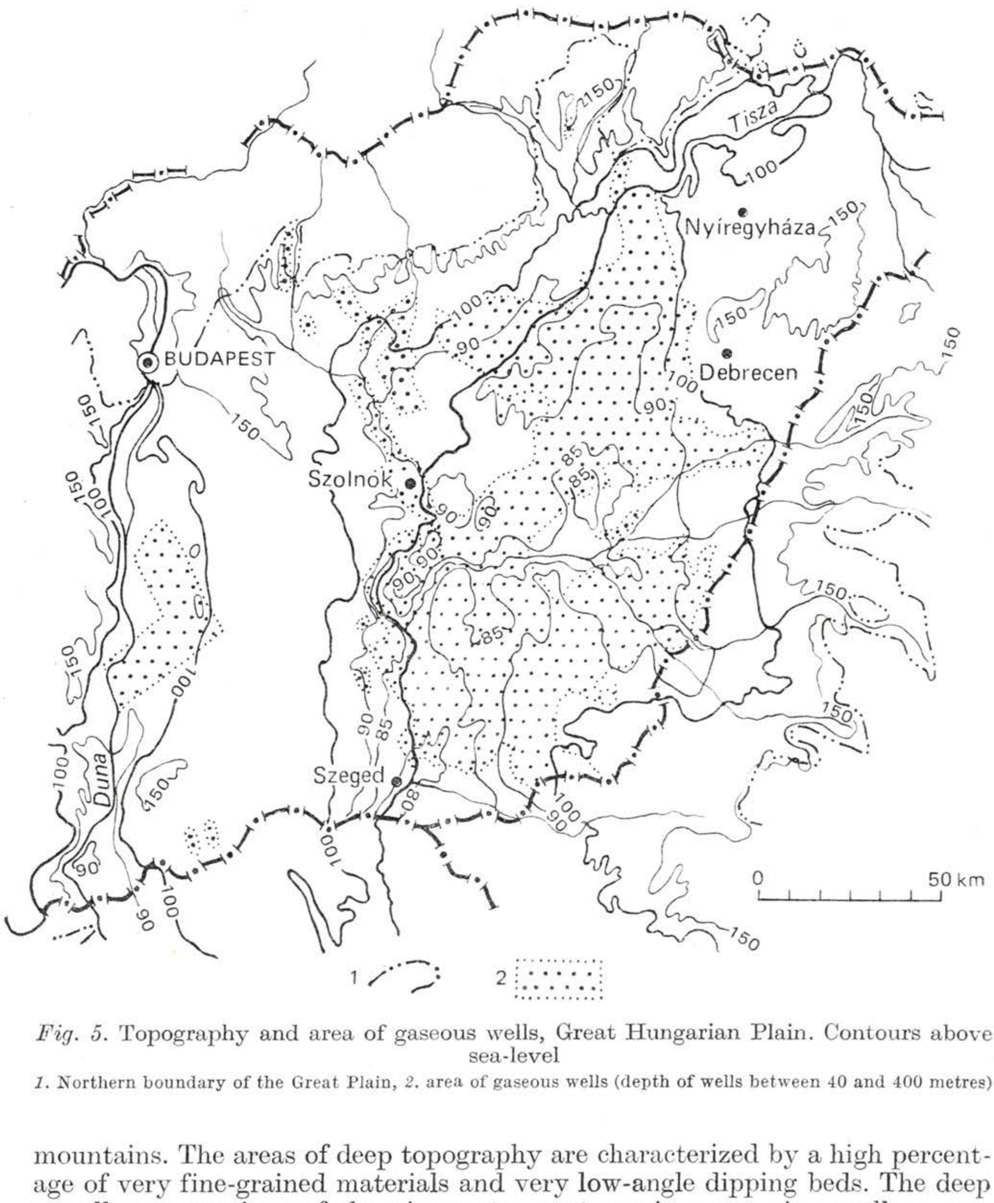

Fig. 5. Topography and area of gaseous wells, Great Hungarian Plain. Contours above sea-level

1. Northern boundary of the Great Plain, *2.* area of gaseous wells (depth of wells between 40 and 400 metres)

mountains. The areas of deep topography are characterized by a high percentage of very fine-grained materials and very low-angle dipping beds. The deep upvalley protrusions of the piezometer contours in many river valleys, e.g. those of the Marcal and Galga rivers are indications of higher-velocity deep underground flows probably along tectonic lines. The Marcal river follows a deep-reaching tectonic zone whereas this peculiar pattern of the piezometric contours is absent in the case of the much greater Rába river because this river does not follow a structural line.

Vertical distribution of the potential

The map of the vertical hydrodynamic gradients of the Great Plain (Fig. 6) shows the approximate vertical change of pressure between 100 and 400 m depth below the ground level.

The vertical gradients were calculated from heads measured before intensive withdrawal had begun.

The 400 m depth is a practical limit, because:

1. Up to about 400 m depth the change in the density of water is negligible.

2. Depth of the intermediate flow-region exceeds only exceptionally the 400 m below ground level. This is partly due to the large areal distribution of the impervious to poorly permeable Upper Pliocene formation which underlies and separates the Quaternary sequence from the thermal water aquifers of the Pannonian (Pliocene) formation. The depth of the top of the Upper Pliocene exceeds only exceptionally 400 m.

3. Scarcity and areally uneven distribution of artesian head measurements of wells deeper than 400 m.

The hydrogeologic-hydrodynamic cross-sections show the flow-regions of the Hungarian Basin as well as the irregularities in the potential distribution.

The porosity of the formations is indicated in the sand percentage of the logs.

The equipotential contour lines show the natural potential distribution undisturbed by withdrawal. Potential contours were constructed from early heads or those measured later but corrected by initial potentials.

Conclusions drawn from potential distribution

From the potential distribution (Fig. 4) many conclusions may be drawn. A few will be sufficient here to indicate the general characteristics of a major sedimentary basin.

1. The areal distribution of strong positive hydrodynamic gradients coincides broadly with those portions of the cross-sections where the equipotential lines are parallel with major geologic features. The most striking feature of this kind is the basement of the Quaternary formation. The parallel arrangement of contour lines in itself is an indication of stagnant deep groundwater. There are long and broad stripes in the deepest sub-basins of the Great Plain where the pressure is the same in every well with the same depth of the screen. This is another indication of the lack of horizontal seepage. Over these sections the upward moving amount of water is equal to the amount of water pressed out due to the compaction of the sedimentary sequence loaded with the material deposited on the ground of the subsiding basin. Water pressed out is moving upward either locally along tectonic disturbances or regionally by very low-velocity seepage through the confining strata (Scherf 1948 and 1967, Erdélyi 1964 and 1972). It may be objected that the seepage velocity parallel to the bedding is many times higher than that perpendicular to the layer. It is true but large amounts of water have moved upward considering


NYÍREGYHÁZA
DEBRECEN
MISKOLC
SZOLNOK
SZEGED
BUDAPEST
Galga river
A
A
A'
A'
1
2
3
4
5
> 0,15
0,15-0,05
0,05 >
> 0,05
0,05-0,02
0,02 >

Fig. 6.

the long time passed since the surface and subsurface geometry and lithology of the Hungarian Basin had become very similar to that of today.

2. Potential distribution shows that there are deep lateral recharge belts of highly permeable formations along borders of discharge areas indicated by bulging equipotential lines (Fig. 4).

3. The low vertical gradients are indications of high seepage velocity. An extreme case is the practically hydrostatic pressure distribution in the thick Quaternary gravel of the Danube below the Little Plain (Fig. 7).

4. Equipotential lines indicate the small role of the rivers in the natural discharge of the basin. The only exceptions are a few portions of the major river courses (Danube and Tisza) where the river beds are penetrating the impervious surface cover and are draining the underlying regional aquifer formation.

The main natural-discharge places of the plains are not the rivers but the large deep-lying undrained or poorly drained flats covered by thick fine-grained deposits with alkaline soils.

5. A few cross-sections show how the recharge part of an intermediate flow-region is "floating" on the thermal water-bearing regional flow-region. The dense equipotential lines in the upper few hundred meters of the cross-sections will be widely spaced toward the depth and in the neutral zone where the trend of the potential will be reversed. Below the neutral pressure zone the downward increase in potential is an indication of the regional flow-region.

6. A belt of near-surface negative vertical pressure gradients inside a region of positive gradients coincides always with an area where:

a) The shallow groundwater-bearing sand is cropping out in long strips and in patches from under the poorly permeable or impermeable soils. The phreatic water infiltrates through the outcrops into the aquifer and dilutes the upward moving deep groundwater. The consequence is a dynamic equilibrium between the upward and downward moving parts of the flow-region causing a weakening of the artesian tendency eventually changing the positive gradient into a weak negative one in the upper portion of the section.

b) The river bed had cut through the impermeable surface layer and is draining and recharging the subsurface water-bearing beds underlying the impermeable cover. The river drain intercepts the upward moving highly saline water.

Chemical analysis of groundwater samples is the best way for localizing the depth of neutral pressure which is an important factor in groundwater management of the area concerned. Overdraft of the shallow groundwater may dangerously disturb the natural equilibrium causing the upward shift of the neutral pressure zone.

Fig. 6. Map of the vertical hydrodynamic gradients of the intermediate flow-region (depth of aquifers between 100 and 400 metres below ground), Great Hungarian Plain (Erdélyi 1973)

1. Northern and *2.* western boundary of the Great Plain, *3.* negative vertical gradients, *4.* area of weak near-surface negative vertical gradients, *5.* positive vertical gradients

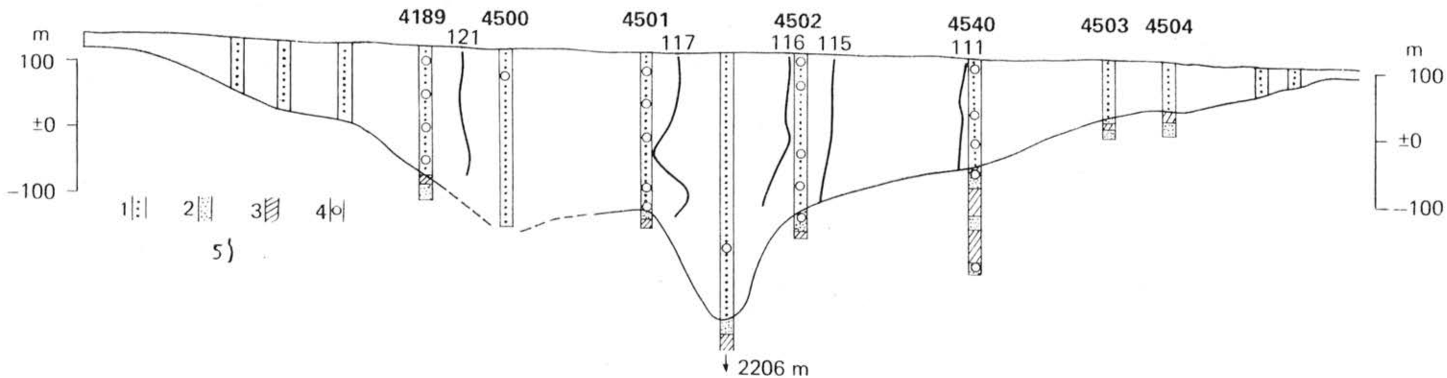

Fig. 7. Lithologic logs and vertical gradients in the Quaternary gravel of the Little Plain

1. Quaternary gravel and sand, *2.* sand (Pliocene), *3.* clay (Pliocene), *4.* depth of the filter, *5.* equipotential lines

We have seen that the upward moving water of the discharge areas is of dual origin. A part thereof moves up in the deep and intermediate flow-regions. Another part is the water being released from the compressed deposits. Both parts increase the salt content of the near-surface water in an unidirectional traffic of salt transport.

The high number of reliable data has made possible to show not only the general hydrodynamic pattern of the Hungarian Basin but also the actual deep flow-regions in 44 long equipotential profiles. Here are only published two long profiles (Fig. 4).

The interdisciplinary approaches in the evaluation of lithologic, structural, pressure, geothermal and hydrochemical data are indispensable for the understanding of the groundwater flow under large sedimentary basins. The author hopes to give evidence of this concept.

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FILTRATION PROPERTIES OF SLIGHTLY PERMEABLE LAYERS OF THE BALTIC ARTESIAN BASIN AND METHODS OF THEIR INVESTIGATION

LES PROPRIÉTÉS DE FILTRATION DES COUCHES PEU PERMÉABLES DU BASSIN ARTÉSIEN BALTIQUE

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RÉSUMÉ

Le bassin artésien baltique, à l'aide duquel sont examinées les propriétés de filtration des couches peu perméables, est situé dans la partie occidentale de l'Union Soviétique. La série sédimentaire du bassin est formée des dépôts paléozoïques, mésozoïques, cénozoïques et quaternaires, d'une épaisseur totale de 100—300 m à 2000—2800 m. Une zone de l'échange active des eaux souterraines s'étend de 100 m jusqu'à 300 m. Dans cette zone se trouvent les couches séparatives suivantes, formées des dépôts argileux: *1°* terre argileuse morainique et sable argileux des glaciations wurmien, rissien et mindelien qui séparent les horizons aquifères intermoraines, *2°* le complexe des argiles jurassiques et triassique, et *3°* les dépôts argileux cambrien et vendien. Les propriétés de la filtration des couches peu perméables étaient étudiées en employant la méthode du bilan hydrodynamique et géothermique.

Les recherches accomplies permettent de tirer la conclusion que les valeurs du coefficient de filtration des couches argileuses peu perméables diffèrent assez peu de celles de la zone de l'échange active. Cependant les valeurs régionales des coefficients de filtration des dépôts de la même âge, obtenues par la même méthode à plusieurs endroits, ne sont pas les mêmes, mais changent dans certaines limites (Table 1). Les résultats des recherches du bilan hydrodynamique permettaient à même temps, de déterminer la diminution quantitative du module d'alimentation des horizons aquifères. Une régularité analogue de diminution de la rapidité de filtration de l'eau à travers les couches peu perméables par rapport de la profondeur est déterminée à base des résultats des recherches géothermiques (Fig. 1). Outre cela, l'analyse des résultats des recherches permet de tirer la conclusion que l'infiltration verticale et les propriétés de filtration des couches peu perméables s'accroissent à partir des bordes du réseau fluvial vers les vallées des rivières (Table 2). Ce phénomène peut être expliqué par le processus de disconsolidation des dépôts argileux.

La dépendance entre les valeurs du coefficient de filtration et la profondeur d'incision des vallées des rivières est indirectement affirmée et pas précisément exprimée (Fig. 2).

Les recherches ont montré qu'une comparaison des propriétés de filtration des couches peu perméables ne doit être pratiquée que sur la base de l'analogie de la structure géologique-hydrogéologique des régions étudiées.

Information on filtration properties of slightly permeable layers and on degree of vertical interdependence of aquifers is necessary when solving many hydrogeological problems. It acquires a paramount importance, in regional estimation of operation reserves of ground-waters and in their protection from pollution. Therefore recently a great attention is payed on study of vertical water exchange through and filtration properties of slightly permeable layers lying in the active water exchange zone of artesian basins.

Methodological problems of the filtration properties of slightly permeable deposits and those of the interdependence of aquifers are analysed in numerous papers. In spite of this fact, the use of the present methods in regional studies needs additional investigations. Thus a principal task consists in working out a complex of rational methods giving the possibility to determine the filtration properties of slightly permeable deposits and regularities of their changes within the limits of their distribution area with an accuracy requested by the practical application.

The Baltic artesian basin taken as an example to study the filtration properties of slightly permeable layers is located in the western part of U.S.S.R. and includes the Baltic synclinal, the slopes of Baltic shield and of Byelorussian—Mozurian anticlinal. The sedimentary strata of the basin is formed of Paleozoic, Mesozoic, Cenozoic and Quaternary deposits having a total thickness from 100—300 to 2000—2800 m. The zone of active water exchange of ground-waters averages from 100 to 300 m.

Alternation of water-bearing and interbedded slightly permeable layers characterizes the sedimentary strata of the artesian basin. The slightly permeable layers in the zone of active water exchange do not ensure a complete isolation of aquifers thus the conditions of vertical water exchange between aquifers are favourable. Within the limits of the zone of active water exchange of the investigated artesian basin the following basic dividing layers can be found: *1.* morainic loam and sandy loam of Würmian, Riss and Mindelian Glaciation dividing intermorainic aquifers; *2.* a complex of Jurassic and Triassic clays; *3.* argillaceous Cambrian or Vendian deposits. Regional study of filtration properties of slightly permeable layers is inseparably linked with the problems of vertical filtration of ground-waters through clayey deposits. As it is generally known, the possibility of filtration of water through slightly permeable (relatively water-resisting) clayey layers in natural conditions was a subject to critical discussions for a long time. Theoretical investigations in the sphere of vertical filtration carried out during the last decades as well as the regional investigation of recharge and discharge conditions of artesian waters either under natural conditions or disturbed by operation indicate, however, that a vertical water exchange through slightly permeable layers is one of the main forms of interdependence of ground-waters in the zone of active water exchange. There can be no doubt that phenomenon of local hydraulic interaction between aquifers through old valleys and erosional cuttings, lithological "windows" and tectonic disturbances is superimposed on a general background of vertical water exchange through slightly permeable layers. This circumstance should be taken into account using regional investigations.

When studying vertical interdependence of aquifers and filtration of ground-waters through slightly permeable layers, hydrodynamic, hydrochemical and geophysical methods are used in the practice of hydrogeological investigations based on determination of peculiarities of corresponding fields. Existing methods can be devided into qualitative and quantitative groups. The first testify only the fact that the process of vertical water exchange has developed in the system while the quantitative methods give the possibility to estimate the intensity of this process and the parameters of geological-hydrogeological media, where it proceeds. The reliability of the investigations depends on both the method applied in the study and the quantity and quality of the initial geological information, the collection of which requires considerable investments in many cases (Fig. 1).

The most trustworthy information concerning the interdependence of aquifers and the filtration properties of slightly permeable deposits can be obtained within those limited areas where the representative investigations were carried out. The calculation of hydrodynamic balance based on data of observations of wells with filters arranged by stages in bunches located in various aquifers is the most important part of such representative investigations. On the basis of balance calculations it is possible to determine the filtration coefficient of slightly permeable rocks, as well as the intensity and dynamics of aquifer recharge. The results of geothermal measurements giving also the possibility to determine the rate of vertical leakage of ground-

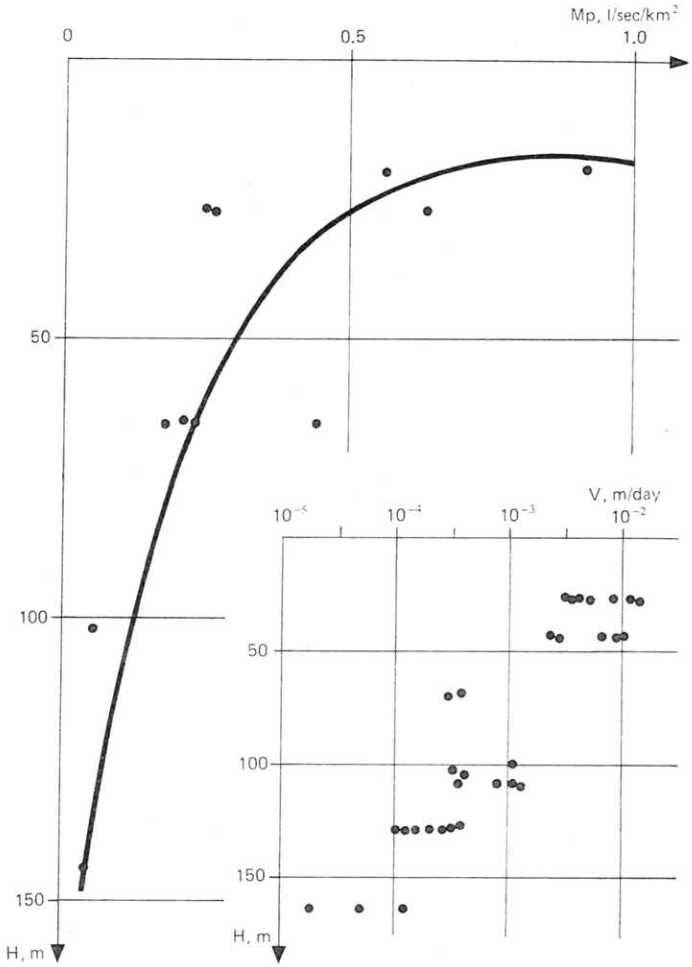

Fig. 1. Modulus of aquifer recharge (*Mp*) and water filtration rate through slightly permeable layers (*V*) as the function of depth (*H*)

waters and the filtration properties of slightly permeable layers may be a source of additional information when studying the interdependence of aquifers. Possible errors using geothermal methods are connected, on the whole, with absence of exact data on coefficient of thermoconductivity of rocks, from which the slightly permeable horizon is formed. It is advisable, in regional hydrogeological investigations that the balance and geothermal methods should be supplemented with methods of solution of reversed dimensional hydrogeological problems. These investigations are based on geological, hydrogeological works of different purposes. The degree of reliability of initial data and also that of the final results is, therefore unequal. In spite of this fact, the solution of reversed problems in regional expression is of great practical importance since it gives the possibility to determine filtration parameters of slightly permeable horizons and modulus of vertical leakage on large territories taking into account the heterogeneity of the area. The analysis of head relations of special regional flows (from recharge area to drainage area) is of a definite interest as well.

On the basis of investigations carried out it is possible to make a conclusion that the values of the filtration coefficient of coeval slightly permeable argillaceous layers in the zone of active water exchange of the Baltic artesian basin determined in the same areas by a complex of the above mentioned methods differ slightly.

Table 1

Filtration properties of basic slightly permeable layers of the Baltic artesian basin

Slightly permeable layers	Geological index	Filtration coefficient [m/day] and method of determination			
		Hydrogeological balance	Geothermal	Solution of reversed dimensional problems	Analysis of relation of heads of regional flows
Würmian*	Q_3	$5\cdot10^{-5}-6\cdot10^{-3}$	–	$10^{-5}-10^{-2}$	$10^{-5}-10^{-4}$
Riss*	Q_2	$2\cdot10^{-5}-7\cdot10^{-3}$	$10^{-3}-2\cdot10^{-2}$	$10^{-5}-5\cdot10^{-3}$	$9\cdot10^{-4}-10^{-1}$
Mindelian*	Q_1	$3\cdot10^{-5}-10^{-3}$	$2\cdot10^{-4}-6\cdot10^{-4}$	$5\cdot10^{-5}-10^{-3}$	$2\cdot10^{-5}-3\cdot10^{-3}$
Jurassic**	J_3	–	10^{-3}	–	–
Triassic**	T_1	$4\cdot10^{-6}-4\cdot10^{-4}$	$3\cdot10^{-5}-6\cdot10^{-4}$	–	–
Jurassic – Triassic**	J_3+T_1	–	–	$2\cdot10^{-6}-5\cdot10^{-3}$	$3\cdot10^{-6}-2\cdot10^{-5}$
Lontovian**	C_1	–	–	$10^{-6}-3\cdot10^{-6}$	–
Kotlinskian**	PR_3^{kt}	–	–	$10^{-5}-2\cdot10^{-6}$	–

* = morainic loam
** = clays

Regional values of the filtration coefficient of coeval deposits obtained by the same method in different areas are different, however, and vary within definite limits (Table 1). This phenomenon is influenced by both lithology and the conditions of occurrence of slightly permeable layers. It was also possibly based on the results of hydrodynamic balance studies to determine decrease of the recharge modulus of confined aquifers with depth numerically. Analogous regularity of decrease with depth of water filtration rate through slightly permeable layers is established on the basis of the results of geothermal

THE COMMUNICATION BETWEEN SUBSURFACE WATER RESERVOIRS IN THE LIGHT OF MINING EXPERIENCES

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The study of water flow processes taking place in sedimentary rocks built up of different layers of water-bearing and impermeable rocks often requires the determination of the rate of the water percolation (seepage) between these layers. In cases when a seepage can take place between two water-bearing layers through a separating layer, the quantitative study of this seepage will require the knowledge of the conditions under which a so-called separating rock will remain impermeable to water and also of the grade of its impermeability. As these processes cannot be studied directly, our conclusions as to the possibility of seepage through confining rocks must be drawn from data and observations, while in certain cases even the rate of the same can be estimated.

A necessary though not the only condition for generating a flow is the existence of differential pore-water pressure sufficiently high to maintain the flow between the reservoirs separated by impermeable layer. The existence of differential pressure, however, does not necessarily involves also the existence of seepage, as in open layers this differential pressure can develop also due to external and internal limiting conditions (Schmieder, 1965), while in closed layers it is derivable also from the genetics of the sediment formation.

In certain cases, direct proofs can be obtained from the chemical composition of underground waters as to the communication of water reservoirs separated by confining beds.

In the course of mining, extensive cavity-systems are being formed and maintained for some decades in the surroundings of water reservoirs separated more or less from these cavities by isolating layers. Thus, the conditions under which a confining bed is actually proved to be impermeable to water, and the extent of its impermeability could be examined on the basis of a multitude of empirical data. This multitude of experiences then served as a basis for laying the foundation of the "theory of protective layers", the development of its quantitative analytical methods and for its application in the everyday practice of the fight against water inrushes in mines.

In the mining terminology, generally speaking, the isolating layers between a mine cavity and a water-bearing layer, which are capable of limiting both the number of the points and the rate of the inflow into this cavity, are

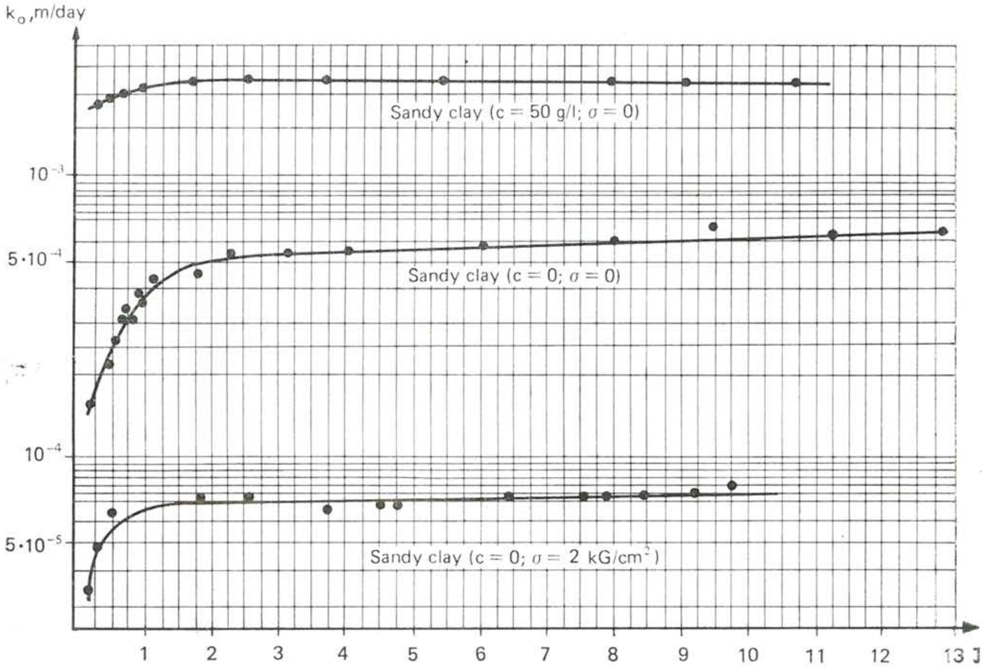

Fig. 3. Relationship between filtration coefficient of slightly permeable rocks (k_0) and gradient (I) obtained using different regimes of laboratory investigations (c = concentration of filtrating liquid, σ = load)

Changes of the filtration properties of slightly permeable layers may occur also due to the differences in vertical gradients of head. As is generally known, the filtration coefficient of slightly permeable rocks in the range of low gradients (when it does not reach a given value) is not constant, but increases with increasing gradient (2, 8, 20 and others). This regularity is well determined experimentally for undisturbed samples taken from the zone of active water exchange of the discussed artesian basin (Fig. 2, 3). It would be necessary, however, to determine the influence of gradient on the filtration properties of slightly permeable layers more precisely by performing field investigations in watershed areas and within the regions of river valleys.

The investigations carried out and the analysis of the obtained results gave the possibility to reveal basic regularities of interdependence of aquifers and changes of the filtration properties of slightly permeable layers within a zone of active water exchange. Investigations showed that filtration properties of slightly permeable layers depend on a number of factors and, therefore, the analogy of geological-hydrogeological structure of the studied areas has to be considered when analysing the obtained results. Informations achieved as the results of the investigations are widely used for regional estimation of operation reserves of fresh ground-waters of the Baltic artesian basin.

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In the mining terminology, generally speaking, the isolating layers between a mine cavity and a water-bearing layer, which are capable of limiting both the number of the points and the rate of the inflow into this cavity, are termed protective layers. Their protective effect is examined in the following respects:

1. Capability of bearing, as a load-bearing structure, the loading caused by water.

2. Capability of preventing starting of a seepage.

3. Capability of limiting the rate of the starting seepage and the extent of the limination.

The possibility of seepage

Let us take the theorem known and applied in groundwater hydraulics as a starting point according to which the water surrounding the rock particles and filling up the pores in rocks has, in its state of repose, a certain resistance to shearing due to the action of the Van der Waals (electrostatic) bonding forces (Kovács, 1972).

The value of the so-called hydraulic threshold gradient (I_0), i.e. the hydraulic gradient required for the starting of a flow depends on these bonding forces. The threshold gradient I_0 thus depends on the measurements of the water channels which can be characterized also by their hydraulic radius r_0. In channels smaller in size, the water molecules lie closer to the rock molecules in the side walls of the channel, consequently the molecular bonding forces are higher in such channels than in those larger in size.

From experiments it was found that

$$I_0 = \frac{C}{\gamma_0 \, r_0} \tag{1}$$

where γ_0 = the specific gravity of the fluid;

C = constant at a given physical state, the value of which depends, according to Bondarenko and Sárosi, solely on the absolute temperature. From experiences, conclusions can be drawn also as to the effect of other factors.

Since several decades, the effectiveness of a protective layer has been characterized by its so-called specific thickness, i.e. the reciprocal of the hydraulic gradient in the mining. From Eq. *(1)* it follows that the threshold gradient of the channel largest in size is the determinative one. Such channels are given by tectonical fractures in the impermeable rocks. Experiences gained in the Hungarian mining show that 95 per cent of the water inrushes occurring in underground mines take place in the surroundings of identified faults.

If the micro-channels along tectonic fractures were developed according to a similar size in the rock of quality given, no seepage (i.e. no water inrushes in a mine) would be experienced below a threshold gradient of a certain value, while at a value higher than that, water seepage would take place at every possible point. Relating the actual number of the seepage points to the possible maximum number obtainable in the absence of a protective layer, and plotting the rations thus obtained as a function of the hydraulical gradient I as shown in Fig. 1, we obtain the discontinuous straight line a parallel to the axis I, in which straight the point of discontinuity lies in I_0. Due to the known inhomogeneity of both the rock material and the tectonical stresses, the channels differ in size and, consequently, the value of the threshold gradient I_0 varies from point to point within an interval. Numerically, this variation can be characterized by an empirical probability distribution (curve *b* in Fig. 1). The relative frequency of the number of seepage points is the reciprocal of the integral curve of this empirical probability distribution as a function of the hydraulical gradient (curve *c* in Fig. 1).

The empirical curves obtained with the data of about 800 water inrushes occurred in Hungarian mines are fairly similar to those theoretically derived, as shown in Fig. 2.

In Fig. 2, curve *a* relates to the Middle Miocene clays at Várpalota, while curve *b* and *c* belong to the Lower Eocene clay-marls of Tatabánya and Dorog, respectively.

The hydraulic threshold gradient has its highest value in the plastic clays where the faults close best. In the more brittle clay-marls it is of a lower value. The relatively larger average size of the channels in the confining beds in the Dorog Coal Basin can be explained by the greater tectonic stresses in this Basin (WILLEMS 1973).

Curves *a'*, *b'* and *c'*, obtained by the derivation of the respective empirical curves, represent the probability distribution of the hydraulical threshold gradient I_0.

Curve *d'* in Fig. 2 shows furthermore the interval of the hydraulic threshold gradients determined from sound (i.e. tectonically undisturbed) rock specimens (V. NAGY—KARÁDI 1960). This is the largest one and fairly fits in with the

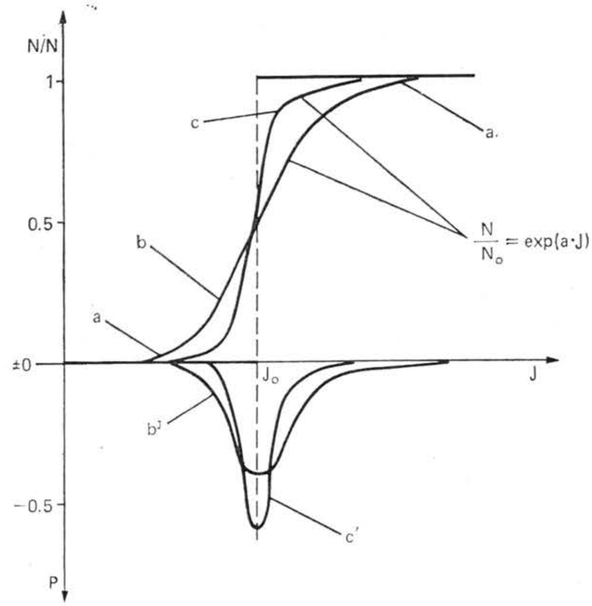

Fig. 1. The variation of the relative frequency of the number of seepage points N/N_0 as a function of the hydraulic threshold gradient I_0 for the following cases: *Curve a)* homogeneous and isotropic impervious layer penetrated by channels of similar measurements. *Curve b)* genuine rock. *Curve c)* as under *b)*, but the thickness of the layer is greater. *Curves a'*, *b'* and *c'* represent the empirical probability distribution (p) of the hydraulic threshold gradient I_0 for the cases under *a)*, *b)* and *c)*, respectively

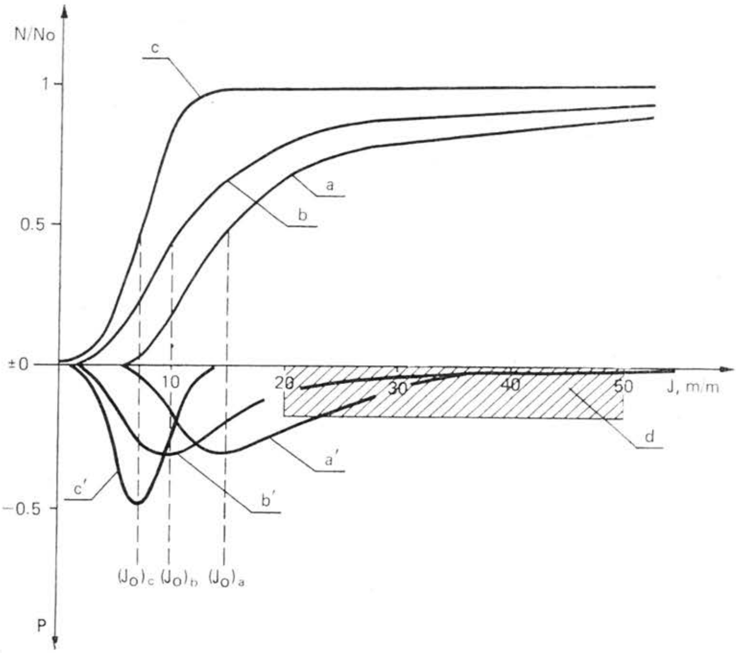

Fig. 2. The variation of the relative frequency of the number of seepage points N/N_0 as a function of the hydraulic threshold gradient I_0, based on experiences gained by the Hungarian mining activity for tectonically stressed and other kinds of separating impermeable beds: *Curve a)* Miocene plastic clay at Várpalota. *Curve b)* Lower Eocene clayey marl at Tatabánya (according to WILLEMS, T.). *Curve c)* tectonically highly stressed Lower Eocene clayey marl at Dorog (according to WILLEMS, T.). *Curves a′, b′* and *c′* represent the empirical probability distribution of the hydraulic threshold gradient I_0 for the cases under *a), b)* and *c)*, respectively. *Curve d)* shows the probability distribution of the same gradient for sound clay as determined on specimens by laboratory tests

experiences. The fair fitting of the relations obtained by deduction with the correlated ones obtained from a high number of empirical data shows that both the hydraulic threshold gradient interpreted as the quotient of the thickness of the confining layer and the differential pressure head as well as its empirical probability distribution characterize even numerically the possibility of a seepage taking place through this isolating layer.

Nevertheless, it should be noted that the average hydraulic gradient calculated with the thickness of the confining layer does surely not equal that actually existing in the determinative (critical) narrowest cross-section of the series-connected narrowing-and-widening channel sections. This actual hydraulic gradient is always unknown, solely just the average hydraulic gradient as calculated between the two endpoints of the channel is known. If there were certain differences they would be due to the differences either in the hydraulic radius r_0 of the channel or in the constant C, as follows from Eq. *(1)*.

The formation of underground cavities results in the modification of the state of stress of the surrounding rocks. In both the roof and the floor of extensive level face excavations "zones of tension stresses" exist up to a height and down to a depth, respectively, of 5 to 20 m. The original fissure size of the faults is here an increased one. Accordingly, the expected hydraulic threshold gradient will be, due to the fissure measurements, higher than that experienced in mining.

Constant C represents the molecular forces' effect that depends also on the properties of the material as well as on the values of the absolute pressure and temperature. Respecting the seepages between water reservoirs, these effects are similar to those of seepages between water reservoirs and mine cavities, at least down to a depth of about 500 m. Furthermore, it is not wholly improbable that also the time factor plays here a role not to be neglected. The "sustained" shearing strength of the "boundary surface" of the water/rock system can be lower than that formed under load but acting for a short time. In rocks, such phenomena are long since known to occur.

These qualitative conclusions, however, do not serve as a substitute for the examination of the phenomena connected with the shearing strength of the water, which examination is already in progress with the subject to solve the practical problems presented by the fight against water inrushes in mines. These qualitative conclusions have rendered probable the assumption that —according to our present knowledge—the determinative hydraulic threshold gradient may be rightly assumed to have a value lying in the vicinity of the domain of those determined from empirical data obtained in mining.

The rate of seepage

In channel systems, the relationship between the effective hydraulic gradient I_h and the velocity of flow W, as based upon the general law of filtration, can be described by the formula (Schmieder et al. 1975)

$$I_h = \frac{W}{k} + \frac{W^2}{k_T^2} \tag{2}$$

where k = the coefficient of laminar filtration in the channel;
k_T = the coefficient of turbulent filtration in the channel.

The coefficient of laminar filtration k in one single channel with a diameter d, is (Schmieder et al. 1975)

$$k = \frac{g}{2\nu\ (\mathrm{Re}\lambda)} d^2 \tag{3}$$

where Re = the Reynolds number;
ν = the kinematic viscosity of the water;
g = the gravitational speed;
λ = the coefficient of resistance to laminar flow.

The coefficient of turbulent filtration k_T is (Schmieder et al. 1975)

$$k_T = \sqrt{\frac{gd}{\lambda T}} \tag{4}$$

where λT = the coefficient of resistance to turbulent flow.

Depending upon the magnitude of the principal forces determining the hydraulic head losses, one or the other term on the right side of the Eq. *(2)* becomes determinative.

The rate of seepage in a channel of a length L and a diameter d being

$$Q = Ldw, \tag{5}$$

the rate of seepage is in any state of flow a function of the known hydraulic gradient I and the unknown channel diameter d.

The rate of seepage cannot be calculated without the knowledge of the determinative channel measurement. However, if the rate of seepage were known, the channel measurement could be calculated.

Based upon the knowledge of the rates of flow, the hydraulic heads and the concerned confining layers of some 800 water inrushes that have occurred in Hungarian mines. Fig. 3 shows the variation of the ratio q/q_0 expressing the relative decrease of the rate of flow of the water inrushes as a function of the effective hydraulic gradient I_h. Here q_0 denotes the rate of flow which would develop in the absence of a protective layer, while I_h is the effective hydraulic gradient maintaining the flow in the series-connected water reservoir and confining layer. In the mining, regarding the best part of all the data obtained in the past, only this latter was known.

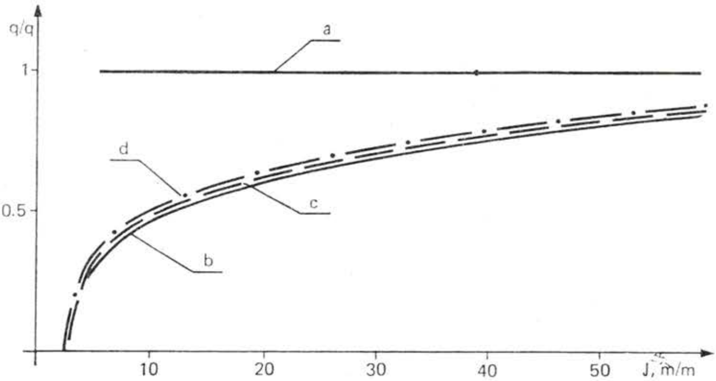

Fig. 3. The variation of the relative rate of seepage q/q_0 as a function of the hydraulic gradient I for the following cases: *Curve a)* Miocene plastic clay at Várpalota. *Curve b)* Lower Eocene clayey marl at Tatabánya. *Curve c)* Lower Eocene clayey marl at Dorog. *Curve d)* theoretical curve

The empirical curve a in Fig. 3 relating to the plastic clays in the floor of the coal seams in the Várpalota Coal Basin is a straight lying parallel to axis I. Should seepage start here, the isolating layer would not be able to limit the rate of flow in the channels, as they would soon be widened out to an extent at which no notable resistance to flow could be offered any more.

The curves b and c in Fig. 3 represent the coal basins Tatabánya and Dorog, respectively, and show a definite throttling effect and fairly fit in with the curve d derived for a constant channel diameter, the equation of which latter curve is

$$\frac{q}{q_0} = \frac{1}{\sqrt{1 + \frac{A}{I - I_0}}} \tag{6}$$

where A is a constant, the value of which depends upon both the characteristics of the layer and the water carrying characteristics of the water reservoir (KESSERŰ 1973).

The equivalent channel diameter calculated for the area Tatabánya and Dorog from the respective equation *(6)* closely fitting the factual data of these areas, is $5-6$ cm, indicating thereby that also the channels in these areas are widened out, though to a smaller extent.

This phenomenon can be interpreted also by an analytical relationship. Introducing the sliding stress τ between water and channel walls, Eq. *(6)* can be brought to the following form (SCHMIEDER et al. 1975)

$$\frac{q}{q_0}=\sqrt{1-A'\frac{\tau}{I-I_0}} \tag{7}$$

In rocks with shearing strength the channel dimensions are steady, however, if the rock gets soaked—as it is the case with the plastic clays of Várpalota—the channels will widen out to a great extent and will then offer no hydraulic resistance to flow.

In Eq. *(7)*, the state of stress of the rocks surrounding the channel and also the static water pressure, as a "form re-establishing" load were left out of consideration.

Opening a horizontal slit in a semi-infinite body of an own weight, the principal stress acting in the direction perpendicular to this slit at a depth z will be

$$\sigma_z = z\gamma_k$$

where γ_k = the specific gravity of the rock.

In a direction opposite to this will act the static water pressure

$$p = h\gamma_v$$

where γ_v = the specific gravity of water and
h = the hydraulic head.

The resultant i.e. the effective stress σ'_z will now be

$$\sigma'_z = z\gamma_k - h\gamma_v > 0$$

as the hydraulical head h reaches the proximity of the surface.

In general, it can be said that $z \approx h$ and $\gamma_k \approx 2\gamma_v$. In a vertical channel of infinite length the primary effective normal stress σ'_h acting on the channel walls will be

$$\sigma'_h = \frac{z\gamma_k}{m-1} - h\gamma_v$$

where m is the Poisson ratio. Under the conditions of $z=h$; $\gamma_k = 2\gamma_v$ and $\sigma'_h = 0$, the value of m will equal 3.

The limiting state of equilibrium is thus characterized by $m=3$ (SCHMIEDER et al. 1975) as $m=3$ being characteristic for sands and clays, in which measures even fault planes somewhat deflecting from the vertical will close.

The equilibrium of steep channels can be maintained by a water pressure of $p=z\gamma_v$ as derivable from rock mechanical relationships and also proved by experiences gained in the mining. Although in case of communicating unconsolidated reservoirs of water, in absence of large cavities capable of receiving the transported rock material, no widening-out of the channels to a degree experienced in mining can be expected. In case of a seepage between the water bearing layers the channel in the separating layer will nevertheless widen out as a result of the transport of rock materials sufficiently fine-grained to be received by the pores of the receiving water reservoir layer. Such internal suffusion phenomena are known from the practice of civil engineering and hydraulic construction.

From the examination of the conditions of equilibrium of rocks the conclusion can be drawn that in water-yielding rocks the developing courses of communication waters are of a "channel character". Crevices of extensive surface areas can remain open in solid rocks or when propped by coarse-grained material.

Unfortunately, from the experiences gained in underground mines no conclusions can be drawn as far as the channel measurements determinative for the communication between unconsolidated reservoirs of water separated by an impervious layer, are concerned.

Conclusion

Though the experiences hitherto gained and evaluated in the Hungarian mining and the theoretical examinations based upon these experiences could quantitatively not be directly applied for the examination of the communications between underground water reservoirs (aquifers) separated by water-impermeable layers, nevertheless, they supplied methods seemingly suitable for a better approximation of the phenomenon as well as some informative data, proving thereby that the new approximation of the problem holds out promises of new results.

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THE PRINCIPAL DISTINCTIVE FEATURES OF THE HYDRODYNAMIC REGIME OF INTENSIVE EARTH CRUST DOWNWARPING AREAS

CARACTÈRES DISTINCTIFS PRINCIPAUX DU RÉGIME HYDRODYNAMIQUE DES RÉGIONS DE LA TERRE SUBISSANT DES SUBSIDENCES INTENSES DE L'ÉCORCE TERRESTRE

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RÉSUMÉ

Plusieurs grands bassins artésiens coïncident avec les régions du fléchissement de l'écorce terrestre. Dans ces bassins, avec la grande puissance des dépôts, la zonalité verticale des différents régimes hydrodynamiques des eaux souterraines est la plus évidente. On a démontré 3 types de ces régimes: d'infiltration, d'élision et profond. Les causes de l'origine des gradients de la pression dans les conditions des ces régimes sont la différence des hauteurs des régions d'alimentation et des régions de recharge; la condensation et la déhydratation des argiles, la réduction de la porosité et la déhydratation thermale des roches sous l'épigénèse et le métamorphisme. On a examiné les particularités et la zonalité des régimes indiqués dans les différentes structures artesiennes. Les régimes d'élision ou profond dans les bassins artésiens se manifestent comme des maxima piézométriques, des anomalies hydrodynamiques. On donne la caracteristique de ces anomalies sur l'exemple du bassin artésien de la Précaucasie orientale.

The areas of intensive earth's crust downwarping may be considered as very complicated water-pressure systems with the distinctive feature of a large thickness of sedimentary and igneous-and-sedimentary strata attaining to 5–10 km and even more. In these areas, large quantities of ground-water penetrate deeply into the earth's crust and are subjected there to the strong effect of high temperatures, pressures and various geological processes. Such areas are expressed tectonically as troughs and depressions with high amplitude (syneclises, foredeeps, etc.). Many large artesian basins (submontane, intermontane and intraplatform; for example, Azov-Kuban, Kura and Dnieper-Donets basins) belong to them.

The differences in ground-water hydrodynamic regimes and their vertical zonation most fully manifest themselves under the conditions of artesian basins with a large sedimentary cover.

Three types of the hydrodynamic regime in the underground hydrosphere–infiltration, elisian and abyssal–were earlier distinguished by the author (Kissin, 1967a). It is possibile to present the principal distinctive features of such regimes as applied to intensive downwarping areas, taking into account new data.

Water movement within the domain of the regime of the infiltration type is governed by a head difference in the zone of recent infiltration and water head formation, and in the discharge zone. Ground-water is recharged through infiltration of atmospheric and surface water, but water of different origin (for example, of sea origin) may be present under conditions of the given regime. The maximum pressure of ground-water within an aquifer does not surpass the hydrostatic pressure of a water column confined between the aquifer and its outcrop in the area of recharge. The necessary condition of existence of the infiltration regime is the hydraulic connection of the aquifer with the ground surface. The main subtypes of the infiltration regime, the regime of unconfined ground-water and the regime of artesian (confined) water, have been studied sufficiently well.

The distinctive features of the infiltration type in artesian structures, being deep-seated depressions, depend first of all on the lithologic composition of rocks filling the depression. The geological section of such depressions is commonly a complicated alternation of rocks with different permeabilities, therefore the seepage field is characterized by great heterogeneity here. Certain differences exist between submontane artesian basins, confined to foredeeps, and intraplatform basins, situated in syneclises.

Submontane basins have an asymmetric structure. A steep limb ajoining a mountain structure is composed of coarse-grained rocks away from mountains. Lateral water movement is retarded due to the poor transmissivity of strata and an appreciable part of the flow is discharged through upward seepage. High temperatures in the abyssal zones of foredeeps contribute to the permeability of clay and therefore facilitate such a seepage. The recharge areas highly elevated in submontane regions provide an appreciable ground-water head.

In intraplatform basins, variations of the infiltration field are expressed not so sharply, and recharge areas are situated at relatively small altitudes. Therefore, here, the role of upward movement of water is not so great, and pressure gradients are smaller than in submontane basins.

Under conditions of the elisian* regime, water flows as a result of its release from rocks compacted under the action of geostatic pressure or tectonic stresses. The development of such a regime is more often connected with the processes of clay compaction. Depending on the factors causing compaction, two subtypes of the regime, namely sedimentation and tectonic, are distinguished.

The water pressure under the conditions of the sedimentation regime is generated by water release on gravitational compaction of the unconsolidated clays. In this case, the head value may appreciably exceed the head value in surface areas of infiltration and head formation. For example, in the axial part of the Terek-Caspian trough the head of Lower Cretaceous strata rises to an absolute altitude of 3300 m, while in the vicinity of the outcrops of the strata, on the nothern slopes of the Caucasian Ridge (the area of recent infiltration and head formation), heads do not exceed 1250 m. Heads in the aquifers occurring under sedimentation regime conditions may approach the value of the geostatic pressure of the covering strata, but they are more often within 0.6 – 0.9 of the geostatic pressure value.

Two stages of the sedimentation regime related to different stages of the trough development should be distinguished. The first stage corresponds to the sedimentation period when buried sediments subside and their lithification

* The term "elisian" is derived from Latin, "elisio" meaning "squeeze out". It is suggested to use by a Soviet scientist, N. B. Vassoevich.

takes place. In this case, rocks are subjected to increasing geostatic load. After sedimentation, the second stage begins when the total geostatic load does not increase, but the consolidation of clay sediments is not yet completed and their compaction continues as a result of redistributed pressure on the rock skeleton and liquid pore pressure. Both the first and second stages are accompanied by water release from clays. The first stage of the formation of the sedimentation system is mainly due to developing troughs, occupied by sea basins, and the second stage may appear in young continental troughs.

The tectonic subtype of the elisian regime is observed in artesian systems recharged through water release from rocks (mainly clayey) being compacted by tectonic stresses. Liquid release from clays subjected to tectonic stresses may also occur when clays were already consolidated under the geostatic load of the overlying layers.

The principal difference between the tectonic regime and the sedimentation regime lies in the close relation with young tectonic movement and very high aquifer (reservoir) pressures, which may not only reach the values of the geostatic load of overlying strata, but also exceed them. For example, in the Sunzhensky anticlinorium complicated by the Terek-Caspian trough, the aquifer pressure/geostatic pressure ratio attains 0.95 (Kissin, 1967b), and at the foot of the Great Himalayas Ridge this ratio has been estimated, using the data of Thomeer and Bottema (1961), at 1.21.

A distinctive feature of the elisian-type regime is the non-equilibrium state of artesian systems, the recharge of which varies depending on the degree of the consolidation of clay sediments subjected to compaction. The rate of the consolidation of clay strata and their impact on the dynamics of the artesian system depend not only on the rate of downwarping or acting tectonic stresses, but also on the conditions of released water outflow. These conditions are governed by the thickness of the clay layer and the presence of permeable (sandy) layers in it. If the clay strata include a sufficient amount of sand layers, water outflow from them is facilitated; a fact that contributes to a rapid compaction and consolidation of clays. In this case, even under intensive subsidence, the effect of clays being compacted on the dynamics of the water-pressure system, happens to be relatively short-term, and large aquifer pressures are not observed in such sections. Such a role of the sand layers is observed in the Maikop strata of the Terek-Caspian and Western Kuban troughs (Kissin, 1967b; Teslenko and Korotkov, 1966)

The recent manifestations of the elisian regime in artesian structures of deep-seated depressions depend on the age of depressions, lithologic facies and tectonic features. These manifestations are most often observed in troughs of the Alpine geosyncline area and young platforms. Piezometric maxima generated by the elisian regime are often encountered in deep-seated aquifers of the Predkarpatsky, Azov-Kuban, Eastern Predkavkazye and Kura artesian basins. They are also observed in many piedmont basins of the Soviet Middle Asia. Recent manifestations of this regime are, as a rule, absent in artesian basins occupying troughs of the ancient platform.

The elisian regime has the strongest effect on the hydrodynamic situation of the basins containing thick clay strata. If clays play a minor role and permeable rocks prevail, manifestations of the elisian regime are not observed despite an appreciable amplitude of downwarping (for example in the Rioni artesian basin).

The regime of the abyssal type is characteristic of artesian systems, which are not influenced by surface areas of head formation and discharge, and are not subjected to the action of abyssal geological processes. The dynamics of such systems is governed by two principal factors: the decreasing porosity and permeability of rocks as they sink deeper and the release of additional water due to thermal dehydration of rocks.

Both the decrease in porosity and the release of additional water in appreciably isolated water-pressure systems lead to an increase in water-pressure that may approach the geostatic load value and even exceed it. Retarded upward water migration occurs under abyssal regime conditions.

Varieties or subtypes of the abyssal regime may be distinguished according to geological processes acting in a certain zone. The epigenetic regime is connected with the processes of abyssal epigenesis (solution of mineral grains under pressure and cementation) that results in porosity decrease and water release. The metamorphogenic regime is formed due to the dehydration and recrystallization of sedimentary rocks in the course of metamorphism.

The elisian regime acts mainly in the zone of initial epigenesis, and the abyssal regime in the zones of abyssal epigenesis and metagenesis. Therefore the abyssal regime manifests itself in deeper parts of troughs, in the domain of high temperatures and pressures. This regime may be formed not only in young, but also ancient troughs.

The effect of geostatic and tectonic pressures on liquid creates a certain common character of abyssal and elisian types of the regime, and predetermines the fundamental difference of these regimes from the regime of infiltration type. Common features of the abyssal and elisian regimes are also the very important role of elastic forces, as a factor of water migration, and the absence of a direct hydraulic connection with the ground surface of the water-pressure systems, being under conditions determined by these regimes.

The occurrence of regimes of various types depends on certain zonation that in turn depends on the structure and history of the geological development of the region. This zonation is different in platform geosyncline areas. The infiltration-type regime prevails in the upper parts of the underground hydrosphere. The lower boundary of its distribution is marked by the contact with the zone of the elisian or abyssal regimes. Infiltration regime conditions occur most deeply in major troughs within ancient platforms. Aquifers with an infiltration-type regime may be encountered at a depth of 6–8 km. For example, in well W. C. Tyrren No. 1. in Texas, after layers with anomalous high pressures had been drilled and cased from a depth of 4964 m down to the planned depth of 6595 m, clay mud with a specific gravity of 1.08 g/cm^3 (Murphy, 1965) was used. This is indicative of the occurrence of layers with normal pressure characteristics of the infiltration regime within this interval of strata.

Elisian regime conditions are typical of structures formed under intensive subsidence terminated only recently. Manifestations of this regime are encountered at depths of 300 to 500 m and even more often at depths of over 1 km. The elisian regime is transformed into abyssal regime in the abyssal epigenesis zone. However, there are cases when water-pressure systems with an elisian regime are replaced in depth by infiltration water-pressure systems. Such a regime zonation is observed in Eastern Predkavkazye (Kissin, 1967b), moreover in the area of the Texas well, mentioned above, and some other regions.

The abyssal-type regime seems to be detectable at minimum depths in regions of recent volcanic activities, where powerful thermic fields occur and thermometamorphic processes take place near the land surface. Their greatest depth corresponds to the lower boundary of the liquid-phase water distribution in the earth's crust.

The zone of reservoir pressures transient from hydrostatic to geostatic (MUKHIN, 1965) is of particular interest. In this zone, being at the lower boundary of artesian systems with an infiltration regime, appreciable pressure gradients occur. YEZHOV and VDOVIN (1970) considered the depth of transition of aquifer pressures from hydrostatic to geostatic for geological structures of younger and old areas of folding. According to average data published by these authors, the depth where such a transition takes place is 2—3 km in the Alpine folded areas, 3—4 km in the area of the Hercynian folding, and 7 km or deeper in the Baikal and Pre-Riphean areas of folding.

In artesian structures, manifestation of the elisian or abyssal regimes should be considered as hydrodynamic anomalies. Deviations from the infiltration regime are characteristic of these structures (KISSIN, 1967b). Such anomalies are commonly localized as zones i.e. separate areas appreciably isolated from the adjacent parts of the water-pressure system. Extremely high water heads, which cannot be due to the elevation of the recharge areas of aquifers with infiltration replenishment, are characteristic of them.

Piezometric maxima caused by hydrodynamic anomalies form a barrier to water flow directed from the surface areas of recharge to areas of discharge. The effect of such a barrier on the regional dynamics of the artesian basin depends on the orientation of the anomaly zones in respect to water flow and is most pronounced where anomaly zones intersect the direction of the flow as in Predkavkazye. In the boundary belt between the anomaly zone and the area with common infiltration regime, retarded water makes a move from the anomaly zone through semipermeable beds due to pressure difference.

Zones of hydrodynamic anomalies display unfavourable conditions for the ground-water development. The isolation of aquifers of these zones from adjacent parts of the artesian system and recharge areas reduces to a minimum the role of the natural ground-water resources in the development of such aquifers. In anomaly zones, due to very high aquifer pressures, appreciable capacities of elastic water storage are available. In the case of large aquifer dimensions their (fairly long) development may only be ensured by elastic water storage. Thermal and industrial waters are more often connected with the zones of hydrodynamic anomaly, fresh water commonly does not occur under such conditions. Evaluations made by the author for one of thermal water reservoirs in the area of the town of Georgievsk (Eastern Predkavkazye artesian basin), have showed that an average of about 6 million cu.m per 30 atm of water is released as result of a head drop in an aquifer having a radius of 10 km. Such an elastic water storage provides water development with a yield of 2000 m^3/day for 8 years.

Very high pressures, characteristic of the elisian and abyssal regimes, appreciably complicate the drilling of deep wells. The excessive pressure at the mouth of wells tapping infiltration regime aquifers does not exceed, as a rule, several units or tens of atmospheres, while in a 6-km-deep well tapping an aquifer under geostatic pressure the mouth pressure rises to 800 atm, and in petroleum-or-gas-bearing beds it is much higher. In planning and drilling deep wells, the forecast of the expectable pressures based on the study of hydrodynamic regimes allows to avoid grave complications.

The Eastern Predkavkazye basin is exemplary for a large artesian basin where the hydrodynamic regimes discussed above are encountered (KISSIN, 1964, 1967b). This basin covers a vast area (about 200,000 km^2), including the northern slopes of the Caucasus, the Terek-Caspian trough and the eastern half of the epi-Hercynian platform of Predkavkazye. The Paleozoic basement of the basin dips to a depth of 6 – 8 km or more in the area of the foredeep and flatly thrusts up to the north. The sedimentary cover is composed of Mesozoic to Cenozoic terrigenous and carbonate deposits.

The thick (up to 1.5 km) clay strata of the Maikop Series dissect the basin into two hydrogeological stages. The whole upper Neogene-Quaternary stage shows conditions of the infiltration regime. The aquifer systems of Pliocene to Quaternary sediments display an appreciable storage of fresh water. They are recharged on the sloping Kabardinskaya plain composed of gravel and pebbles.

The aquifer systems of the Podmaikopsky stage are replenished through infiltration in the northern slopes of the Caucasus Ridge. The zone of partial discharge of these systems is confined to the Karpinsky rampart. These aquifer systems contain mineralized thermal water.

Over the largest part of the basin, in the Podmaikopsky stage, a gradual decline of water heads towards the north-east, in correspondence with the position of recharge and discharge areas, is observed. However, extensive zones of hydrodynamic anomalies are found in the southern and central parts of the basin (Fig. 1). Very high water heads show that there are no infiltration regime conditions in these zones. In addition to the three zones of anomalies confined to the middle reaches of the Kuma River, Terek and Sunzhensky anticlinorium, a number of piezometric maxima are found in separate wells or groups of wells.

In the Kuma zone, hydrodynamic anomalies are observed in sand bands of the Maikop strata, moreover in the Paleocene-Eocene and Upper Cretaceous aquifer systems, but in the Lower Cretaceous system they are absent. Here, the largest absolute water heads attain 1245 m. In the Terek and Sunzhenskaya zones, anomalies are observed in Maikop; Upper Cretaceous, Lower Cretaceous and Jurassic aquifer systems and heads come to 4000 – 4500 m. For these zones, the water heads and the aquifer pressure/geostatic pressure ratio have maximum values in the Maikop strata and they decrease, as a rule, in deep-seated aquifer systems.

The detailed analysis of hydrogeological and structure-geological situation of these zones allows to conclude that the main cause of the very high groundwater heads lies in the compaction and dehydration of the Maicop clays. Water, released from these clays, penetrated into the underlying aquifers that resulted in an increase of pressure in them. In the Kuma zone, a relationship between the anomaly intensity and the rate of the Maikop clay strata dip is observed. Therefore, conditions of the sedimentation subtype of the elisian regime exist here. In the Terek and Sunzhenskaya zones, the clay compaction is mainly provoked by tectonic stresses that had folded the Maikop strata (diapiric folds). In these zones, hydrodynamic anomalies have mainly been formed under conditions developed by the elisian regime subtype.

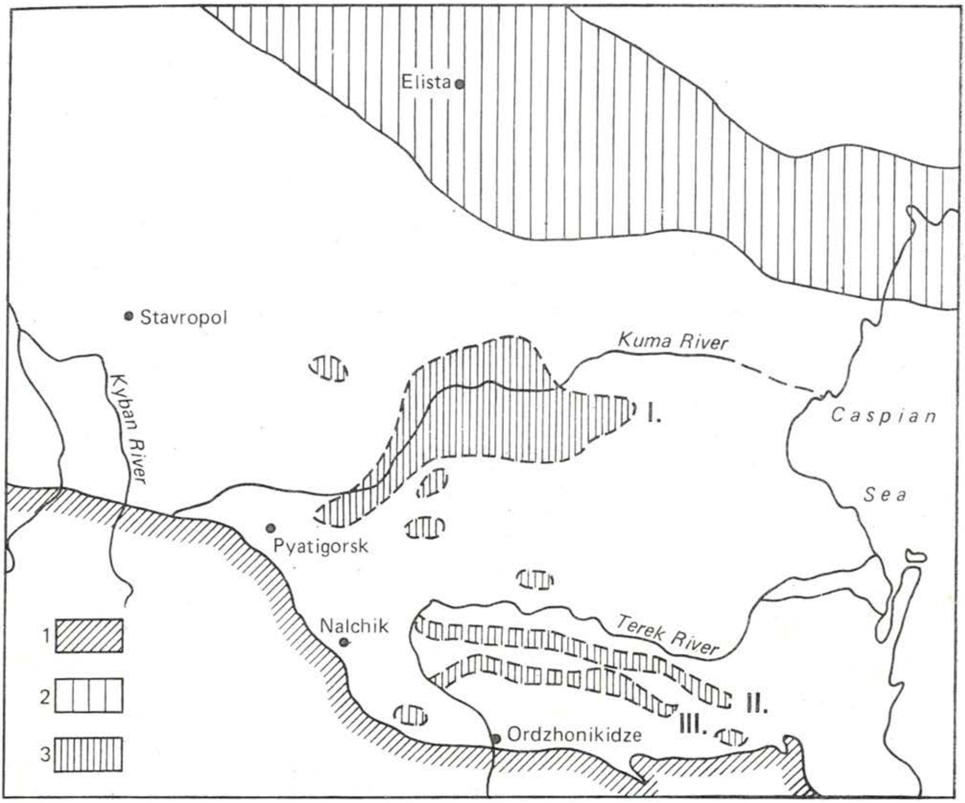

Fig. 1. Scheme of the Eastern Predkavkazye artesian basin

1. The northern boundary of the infiltration recharge area and head formation, *2.* the area of partial groundwater discharge, *3.* the zones and areas of hydrodynamic anomalies. *I.* The Kuma zone, *II.* the Terek zone, *III.* the Sunzhenskaya zone

In one area situated to the north-east of the Kuma zone, hydrodynamic anomalies in the Lower Cretaceous aquifer system are evidently connected with epigenetic processes. Microscopic studies of rocks have showed that here the siltstones show solution structures typical of these processes. Thus it may be assumed that manifestations of the abyssal hydrodynamic regime exist in deep parts of the Eastern Predkavkazye basin.

Known zones and areas of hydrodynamic anomalies in the basin mentioned cover an area of over 9000 km^2. The total area of anomalies, together with regions where they possibly exist, may take up 2000 km^2, that is, 10% of all the artesian basin area.

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CONCEPTUAL AND DIGITAL MODELS OF THE GROUND WATER FLOW IN THE TERTIARY BASIN OF THE TAGUS RIVER (SPAIN)

MODÈLE CONCEPTUEL ET DIGITAL DE L'ÉCOULEMENT D'EAUX SOUTERRAINES DANS LE BASSIN TERTIAIRE DE LA RIVIÈRE TAGE (ESPAGNE)

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RÉSUMÉ

La zone centrale du bassin du fleuve Tage est constituée par une fosse tectonique comblée de matériaux tertiaires et quaternaires d'origine continentale. Elle occupe en surface un peu moins de 15 000 km^2 et son épaisseur semble dépasser, dans certaines zones, les 2000 m. L'exploitation des eaux souterraines de ces matériaux a connu un accroissement rapide ces dernières années en raison de l'accroissement démographique et industriel rapide de l'Aire Métropolitaine de Madrid.

Dans cet article, les modèles conceptuels utilisés aujour-d'hui pour l'interprétation des données disponibles sont exposés. Dans l'essentiel, ces modèles rejoignent ceux des systèmes de flux régional, intermédiaire et local proposés par Tóth. Sont exposés aussi les premiers résultats quantitatifs obtenus par application d'un modèle digital simplifié ou pré-modèle à l'étude d'une zone de 1500 km^2. Le modèle digital utilisé suppose trois aquifères superposés et séparés entre eux par des nappes semiperméables. Le premier aquifère est libre, reçoit une recharge constante en provenance de la surface et est rélié aux rivières. Dans différents pré-modèles, le régime permanent naturel et le régime variable introduit par les pompages sont analysés.

Introduction

The central zone of the Tagus River basin is a large graben filled with continental Tertiary materials. It is bordered on its northern and southern sides by impervious rocks. Its eastern border is formed by the calcareous Altomira Mountains. The total surface area of the graben is something less than 15,000 km^2 (Fig. 1).

The sediments lie almost horizontally. The relief is generally smooth. It is composed of extensive valleys whose bottoms are usually covered by Quaternary alluviums. Hills and often horizontal tablelands form the watersheds between the valleys. There is rarely more than 200 m-s difference between the level of the bottom of the valleys and that of the nearby crest lines. The climate is of continental type with an average rainfall of 400 to 600 mm/year. The average yearly temperature is about 15 °C.

The Metropolitan Area of Madrid is in the central zone of the area and covers some 1000 km^2-s. It contains a population of more than four million inhabitants and the population is predicted to increase to nine million by the end of this century. The water demand for urban and industrial supply is

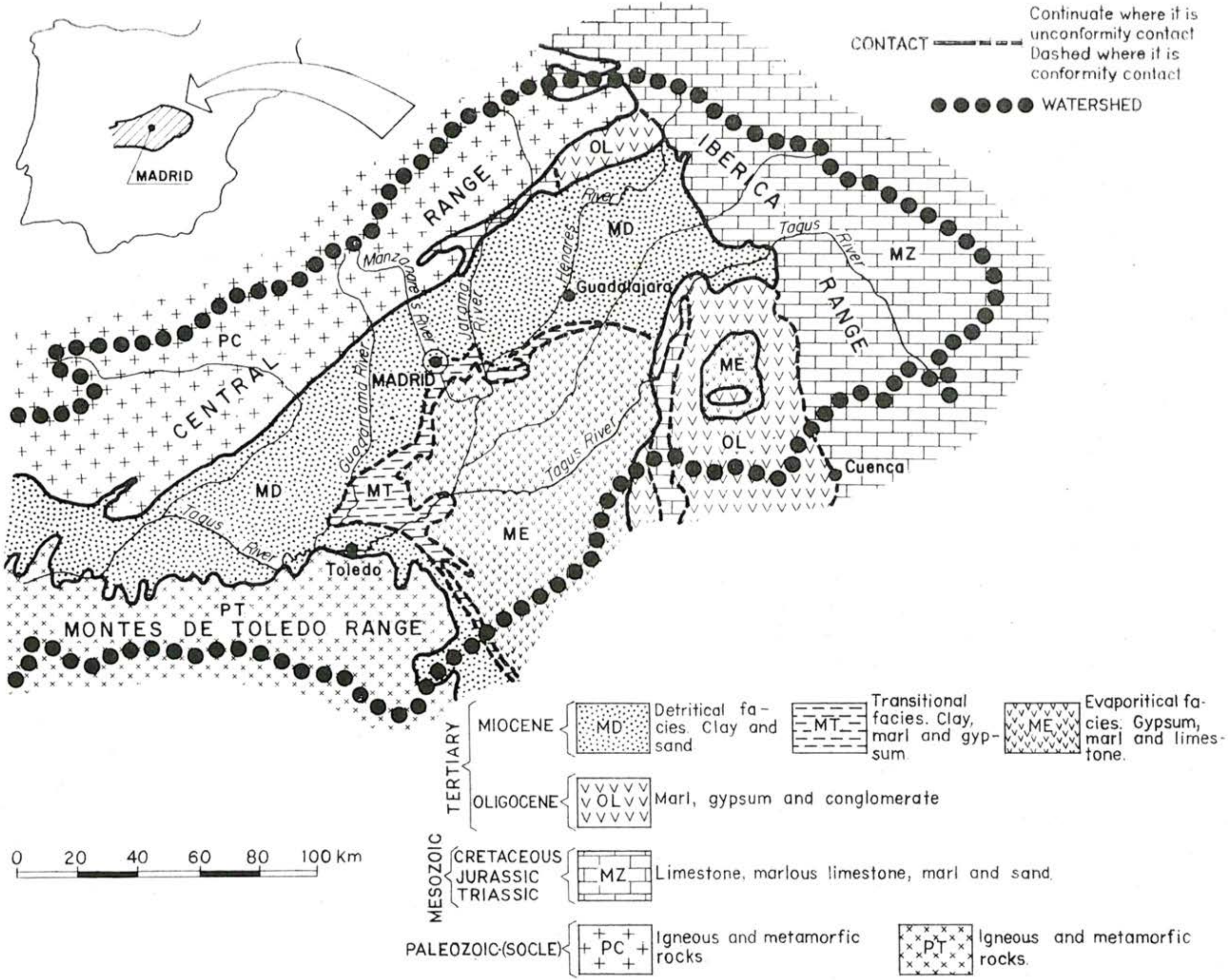


Fig. 1 SCHEMATIC GEOLOGICAL MAP OF THE TAGUS RIVER TERTIARY BASIN

currently about 600 hm^3/year and it is predicted to increase to the amount of 2000 hm^3/year by the year 2000. The current water demand for irrigation is small (200 hm^3/year) and is not expected to increase considerably in the future.

The water supply of the city of Madrid was based exclusively on ground water until 1858. From that time until very recently, practically the entire water supply of Madrid has been ensured from several surface water reservoirs located in the northern mountain ranges. During the last few years the technical feasibility and the economic advantages of the utilisation of ground water has become more apparent (LOPEZ CAMACHO, 1974). This fact has given rise to a good deal of activity in the construction of wells.

With the goal of contributing to the necessary knowledge of ground water in the Metropolitan Area of Madrid, the Water Resources Research Section of the Higher Council for Scientific Research (C.S.I.C.) has begun an extensive research programme which includes different subprogrammes (LLAMAS, 1975). This paper will only take up the subprogramme dealing with the study of flow systems by means of digital models.

Conceptual models of the aquifer system

The geometry and hydrogeological parameters of the aquifer system

The geological and hydrogeological characteristics most relevant to this paper will be explained briefly (a more detailed description can be seen in LLAMAS and LOPEZ VERA, 1975).

The main geological characteristics (Figs. 1 and 2) can be summarized as follow:

a) The graben of the Tagus was formed throughout the Tertiary and filled with sediments of continental origin that came from the bordering mountains.

b) The maximum thickness of the Tertiary seems to be 2000 to 3000 metres.

c) The lithology of the Tertiary sediments is the usual one of an endorheic continental basin in an arid climate. Fundamentally, three large lithological facies can be distinguished: one *marginal or detrital facies*, a *central or chemical* one, which usually contains gypsum and marls or lacustrine limestones, and an *intermediate facies* which has the characteristics of both of the first two to a greater or lesser extent.

d) Lastly, with the advent of the Pliocene the basin became exorheic and an erosive period began which continues at the present time. At the same time, alluvial type sediments were deposited in the bottom of the valleys. These sediments are sometimes very extensive, but generally have thicknesses of less than 10 to 20 metres.

In the marginal facies the aquiferous formations are mainly composed of sand and gravel lenses, which are surrounded by a mass of finer materials, including clay and silt. The frequency and permeability of these lenses depend to a large extent on the nature of the rock in the source area. So it is practical

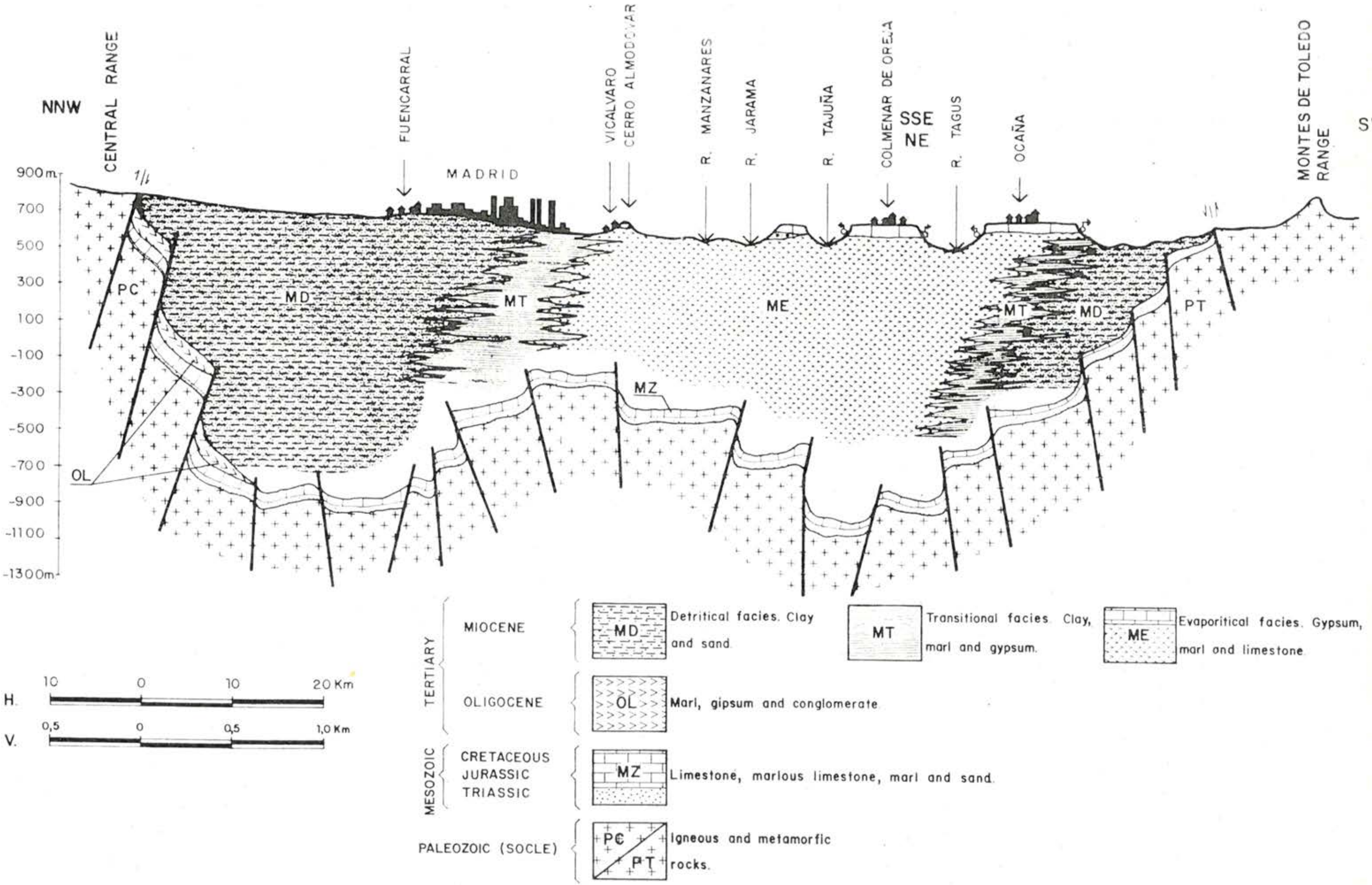


Fig. 2. SCHEMATIC GEOLOGICAL CROSS-SECTION OF THE TAGUS RIVER TERTIARY BASIN

to distinguish a fairly sandy "Madrid facies" originated by the erosion of the granites and gneiss of the Guadarrama Mountains, and a "Guadalajara facies", more clay-like, having originated from the shales of the Somosierra Mountains.

Because of their minor thickness, the Plio-Quaternary alluvial deposits are not usually important aquifers, except when they are hydraulically connected to river beds.

The Miocene limestones form extensive layers that lie almost horizontally. They can be interesting aquifers but their flow system is almost independent of the Tertiary and Quaternary detrital flow systems.

The specific capacity of the "Madrid facies" wells usually fluctuates between 10 and 300 m^2/day. The specific capacity of the "Guadalajara facies" wells is usually smaller and rarely reaches 50 m^2/day. In the Quaternary alluvium the specific capacity can reach sometimes 3000 m^2/day, but the dynamic drawdown which can be produced is generally very limited.

The system of recharge, flow and discharge of the ground water

Considering the great thickness and the large surface area of the graben, the most suitable conceptual model for interpreting the currently available data is considered to be that of the regional, intermediate and local flow as proposed by Tóth (1963).

Fig. 3 shows a hypothetical diagram of the regional, intermediate and local flows in a geological profile parallel to the Central Range.

Fig. 4 shows a diagram of a local flow system (Hubbert type), considering also the inhomogeneous character of the medium. If the lenses were frequent and distributed with a certain statistical uniformity, this aquiferous system would be practically equivalent to a homogeneous anisotropic aquifer with a vertical permeability smaller than the horizontal permeability. In a more simple way, it may also be considered to be equivalent to a "multilayer aquifer", i.e. a system composed of a series of aquifers and aquitards. Fig. 5 shows one of these multilayer systems which is made up of three aquifers, the first of which is unconfined and the last one is limited at the bottom by an impervious layer.

The flow system is fairly similar in the three conceptual models described previously (Figs. 3, 4 and 5). Among their distinctive characteristics the followings are worth-while to mention: *a)* recharge originates mainly from infiltration in the interfluvial zones and the system is drained at the bottom of the valleys; *b)* the piezometric levels fall as a well or drilling goes deeper in a recharge zone, the opposite character can be observed in discharge zones; *c)* in the Tagus graben there are probably not only local flow systems, but also intermediate and regional ones such as have been indicated in Fig. 4.

Hydrogeochemistry

As a general rule, the available data indicates that the mineralisation of the ground water coming from the marginal or detrital facies is usually limited and almost always has a TDS of less than 500 ppm. On the other hand, in the

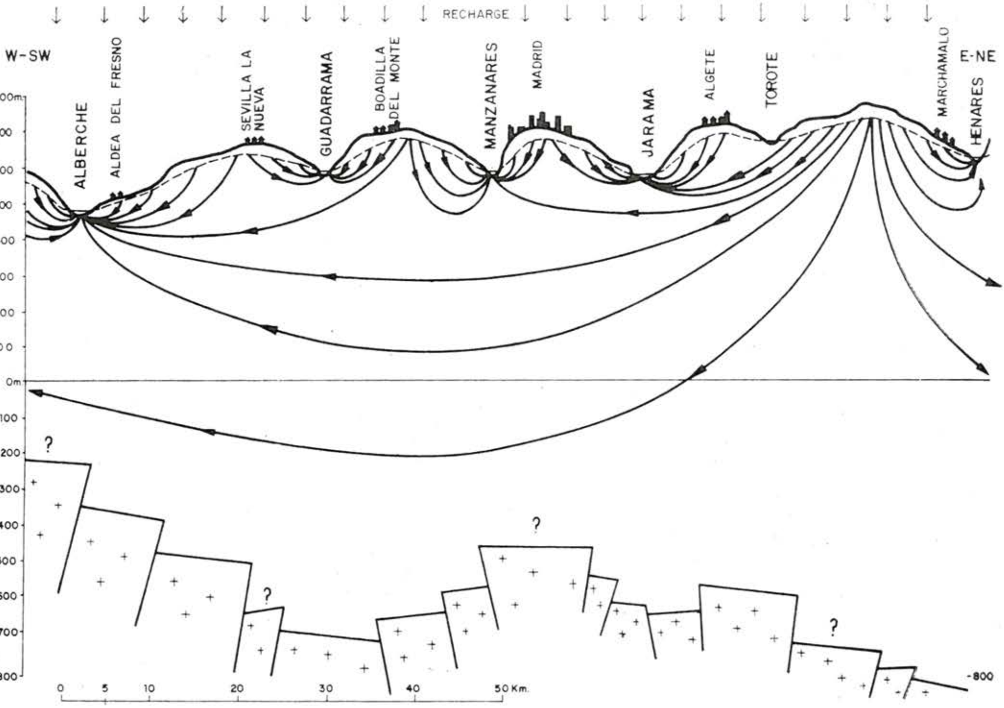


Fig. 3. Hypothetical local, intermediate, and regional ground water flow systems in the Tagus River Tertiary basin

central or chemical facies the TDS is usually greater than 2000 ppm. Fig. 6 shows a map of the variation of the electrical conductivity of the ground water of the local flow system in the middle basin of the Jarama River. There exists a close correlation between the lithology and the quality of the ground water.

Nevertheless, there may be zones in which this correlation is not so evident. These zones appear to belong to areas in which the discharge of the regional or intermediate flow is important in comparison to the discharge from the local flows. This seems to be the case of the western zone of the graben, near Talavera de la Reina, where the TDS of the ground water is greater than 8000 ppm., in spite of the absence of evaporitic rocks in the vicinity. This fact seems to be consistent with the flow diagram in Fig. 3.

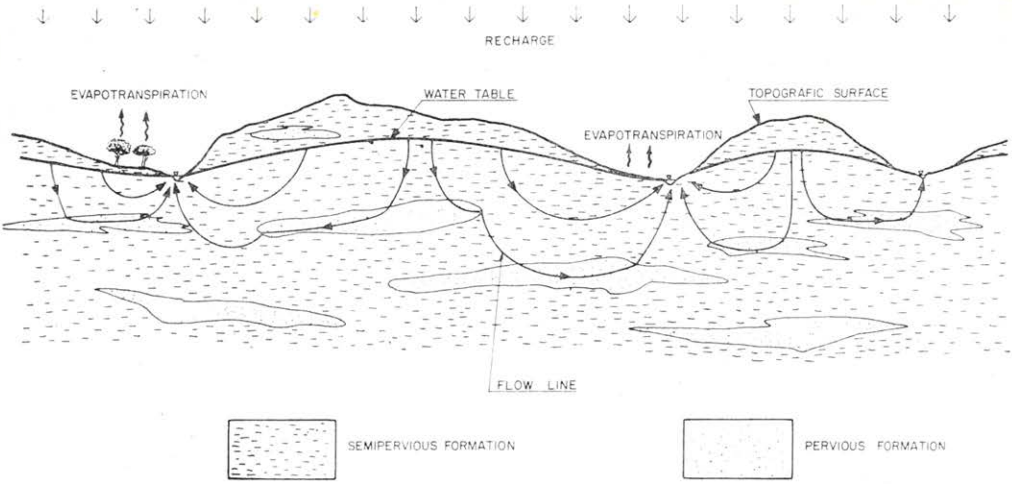


Fig. 4. Local ground water flow system in heterogeneous media

Digital models

Basic hypotheses and method of calculation

The large demand for water in the Metropolitan Area of Madrid and the growing number of wells drilled there have made it advisable to prepare a mathematical or digital model in a relatively short time to simulate the effects of the different exploitation hypotheses. Research was begun using a digital model of which a certain experience had already been attained in other cases in Spain (Aragonés et al., 1972, López Camacho and Llamas, 1975). Nevertheless, the possibility is currently being studied to apply more sophisticated models, more or less similar to those proposed by Freeze and Witherspoon (1967).

The digital model that has been used is very similar to the so-called "three dimensional" one of Prickett and Lonnquist (1971). Their main characteristics are described below. The prototype is represented in the model by three superimposed aquifers (see Fig. 5). The first aquifer is a water table one, but it is supposed that its saturated thickness is great enough to consider its T constant throughout the entire time; it is connected to the rivers and adjoining alluvial aquifers through layers of low permeability. It obtains a recharge from the infiltration of the rain which is assumed to be a constant through time and space; in the urbanized zones the recharge by rain is replaced by the recharge originating from the leaking distribution and sewer networks. The recharge from these leaks is considered to be greater than that produced by the rain. The second aquifer is confined and is separated from the first one by an aquitard. The third aquifer is separated from the second one by an aquitard in the same way and limited at the bottom by an impermeable layer or aquiclude.

The method of calculation used to determine the hydraulic parameters of the flow system is based on the application of finite differences and it is the

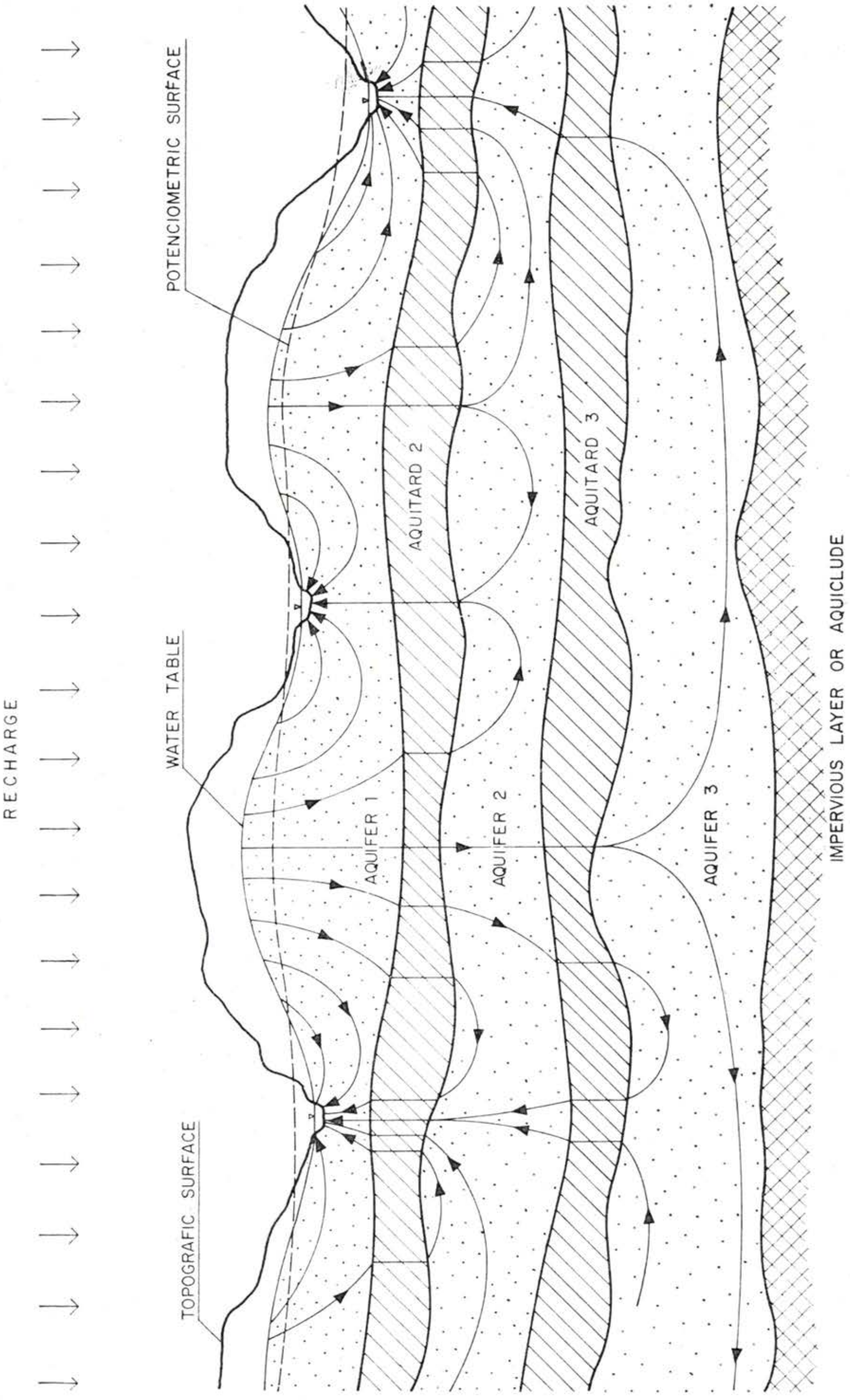


Fig. 5. Scheme of multilayer-aquifer

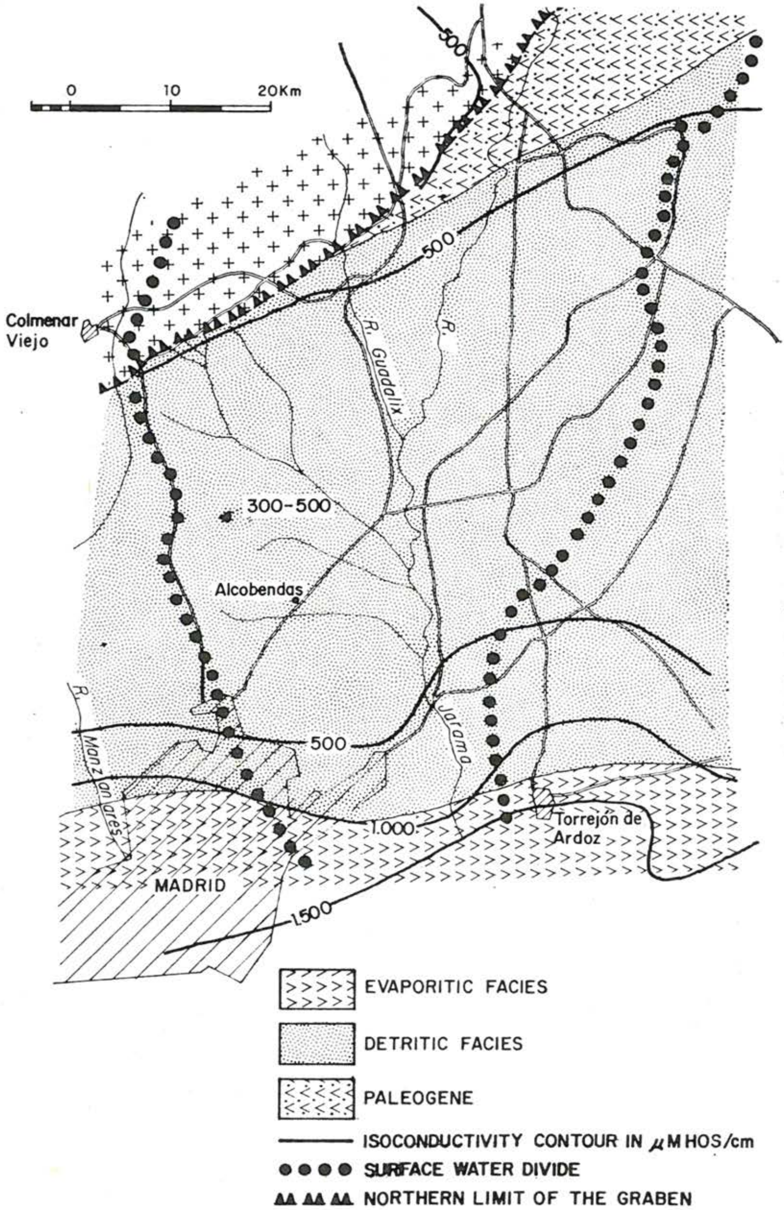


Fig. 6. Electric conductivity of ground water in the Jarama River basin (after LLAMAS and LÓPEZ VERA, 1975)

so-called Implicit Iterative Method of Alternating Direction. The calculation programme has been processed by a UNIVAC computer from the Ministry of Science and Education.

The characteristics and calibration of the premodels I, II and III

During this first stage it was assumed that the transmissivities and the storage coefficients were constant through time and space in each of the three aquifers; in the same way, a single value is considered in each aquitard for its vertical permeability (k') and its thickness (b'). The variation in the storage of water in the aquitard due to the change in the pore pressure is not considered. Because of these simplifications of the prototype, this type of models was called "premodel" and it was used to select the parameters for the final model.

Two simple two-dimensional cases were tested at first i.e. two vertical hydrogeological sections with two and three superimposed aquifers, respectively. Premodels I and II served:

- to conclude the final preparation of the programme;
- to test the coherence of the average values for T, S and (k'/b');
- to obtain an order of the magnitude of time that was necessary to reach a steady regime, starting from an initial situation in which the piezometric level was the same in all three aquifers (for more details, see LLAMAS, 1975).

The next step was the preparation and calibration of a three-dimensional model which corresponded to a 50-by-30-kilometre zone (Fig. 7). Although its surface area represents only a little more than 10% of the surface of the graben, it is of great interest since a large part of the Metropolitan Area of Madrid is located there. The grid used to make the area discrete is quadrangular and uniform (2×2 km) and gives $25\times15\times3=1.125$ nodes.

The calibration of the premodel was performed by comparing the calculated and measured piezometric surface corresponding to the steady regime in the confined aquifers, while in the first aquifer the regional water table deduced from the measurements of the water level in wells penetrating less than 20 metres into the saturated zone was used for the comparison. The goodness of fit between the premodel and the prototype looks acceptable – except in the lower left-hand corner – where the differences between the computed and the measured values are greater than 10 metres. The calibration process required seven premodels III to be carried out. The principal characteristics of the last one, premodel III-7, were as follows (Fig. 8):

a) Transmissivity of the three aquifers were 50, 100 and 50 m^2/day respectively. The values for S are hardly of interest because the purpose is to analyse the steady natural regime without spending an excessive amount of time on the computer.

b) The value of the leakage coefficient (k'/b') is the same in the two aquitards and equal to $0.625\cdot10^{-5}$ ($days^{-1}$).

c) Between the first aquifer and the rivers or the alluvial aquifers connected to them, a layer of low permeability is supposed to exist (0.01 m/day) with a thickness of one metre. Both the level of the rivers and the level of

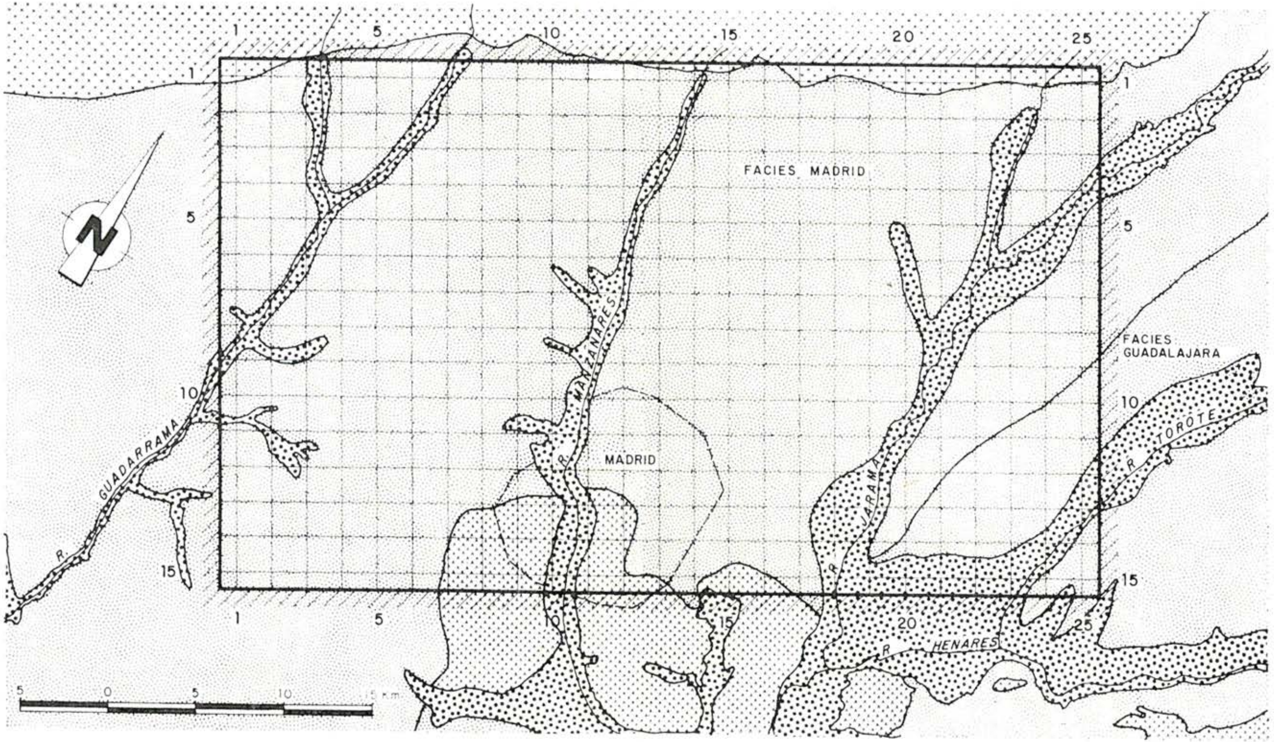


Fig. 7. Grid for premodels III and IV

the Quaternary aquifers connected to them are taken as constants along the time.

d) The recharge of the rain is equal to 0.05 m/year in all of the nodes except in those of the urbanized zones of Madrid, where the recharge is taken as being equal to 0.20 m/year.

This premodel (Figs. 8a–8b) shows that the ground water flow divides in the second and third aquifers also coincide, approximately, with the divides of the superficial watersheds. This would seem to indicate that there is no regional flow. This is probably due to the fact that the model zone is too small (10% of the total extension), or because it might be necessary to use a fourth aquifer in order to take the deepest flows into account. Both possibilities shall be analysed in future studies.

Results of the first withdrawal simulation

In spite of the limitations of premodel III-7, its calibration was considered to be sufficient to begin a trial of the effects that would be produced by the next ground water withdrawals in the Metropolitan Area of Madrid. The data and results of one of the simulations that have been carried out are summarized as follows:

All of the data are the same as that obtained from premodel III-7 except for the following: *a)* the value of S in the first aquifer is equal to 0.10; in the other two aquifers $S = 0.0001$; *b)* the time of the simulated period is equal to 128 years; *c)* the pumping is taken as a constant throughout time and is concentrated in six nodes (three of them very close together). A rate of 10 hm^3/year is pumped from each of them; the screens of the wells are always in the second aquifer.

In Figs. 9a–9b the piezometric surfaces of the three aquifers are shown at the end of 128 years, as well as the hydrographs of nodes (13, 8, 2), (13, 8, 1) and (1, 15, 1) which correspond to the node of the greatest drawdown, the node that is immediately above it and to the point in the upper aquifer which lies at the greatest distance from the rivers and pumping centres.

Among the conclusions deduced from this simulation, the following are worthy of note: *a)* the drawdowns created by concentrated pumping were very great and were attained very rapidly in the two lower aquifers (if the screens were placed in the first aquifer they would be slower, but larger); *b)* the influence on the upper aquifer caused by the pumping in the second aquifer was relatively slow and it would be even slower if the water released from storage in the aquitards were taken into account; *c)* the stabilising effect of the rivers on the progressive drawdown of the piezometric levels is small, if the wells are not located in the first aquifer and in a zone that lies close to streams.

The foregoing conclusions would seem to indicate that the project of carrying out concentrated pumping in the Tertiary of Madrid should be considered with a good deal of reservation, at least until more detailed studies have been carried out.

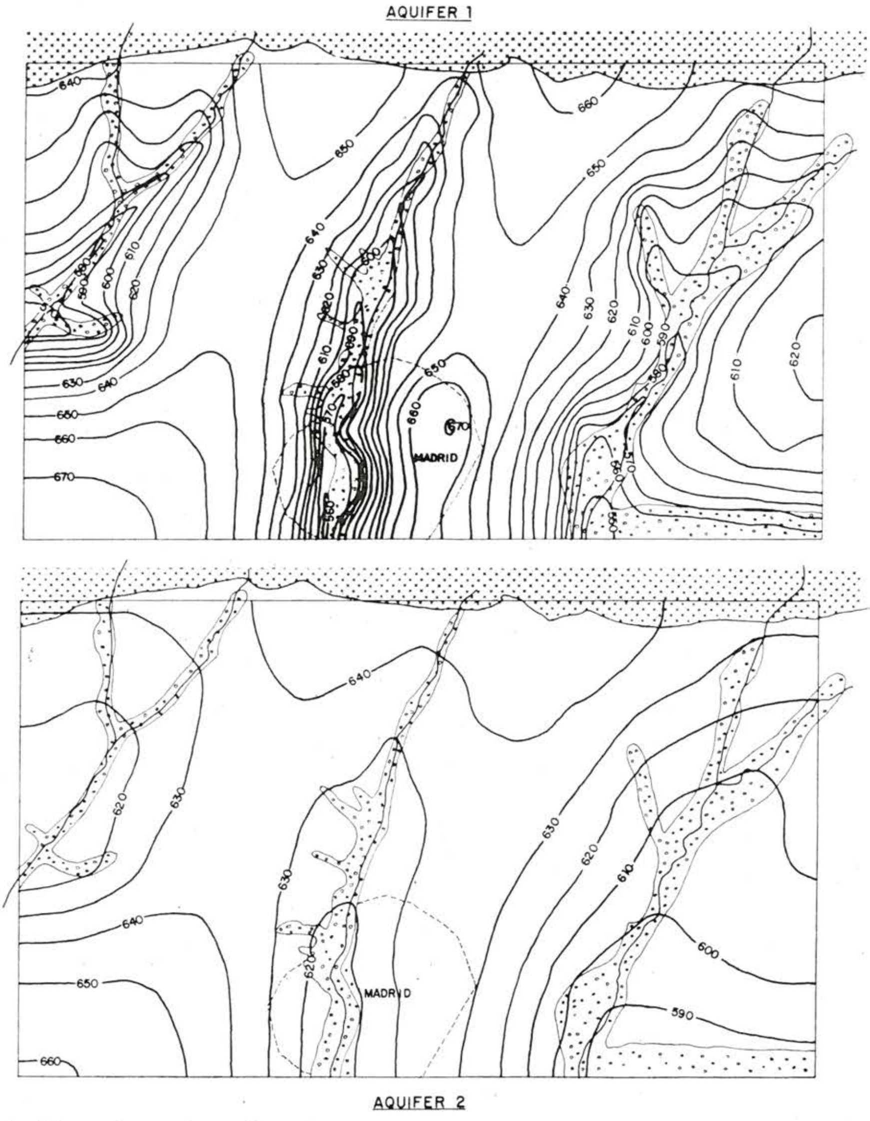


Fig. 8a Potentiometric surfaces in steady flow (premodel III-7). (Ground water levels in meters a.s.l.)

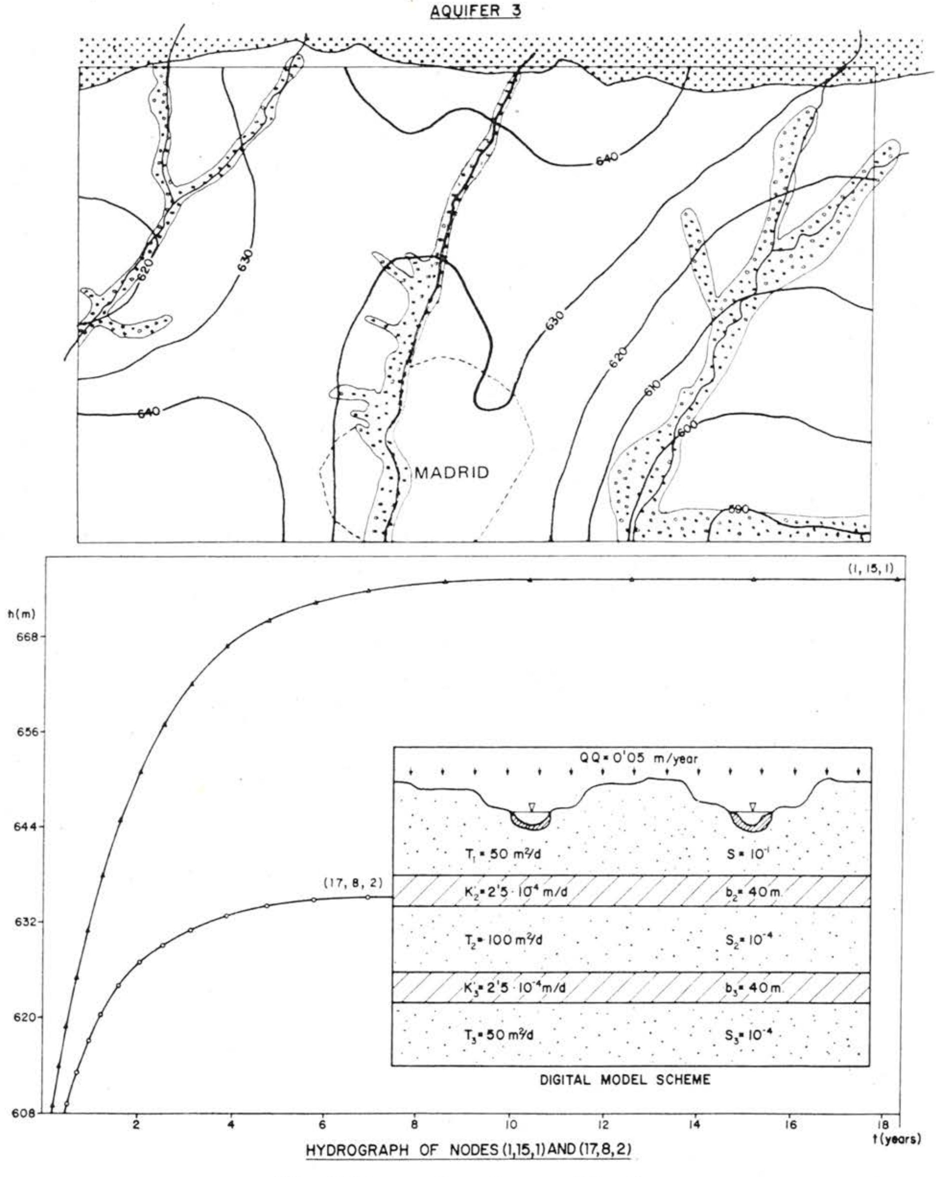


Fig. 8b For explanation see Fig. 8a

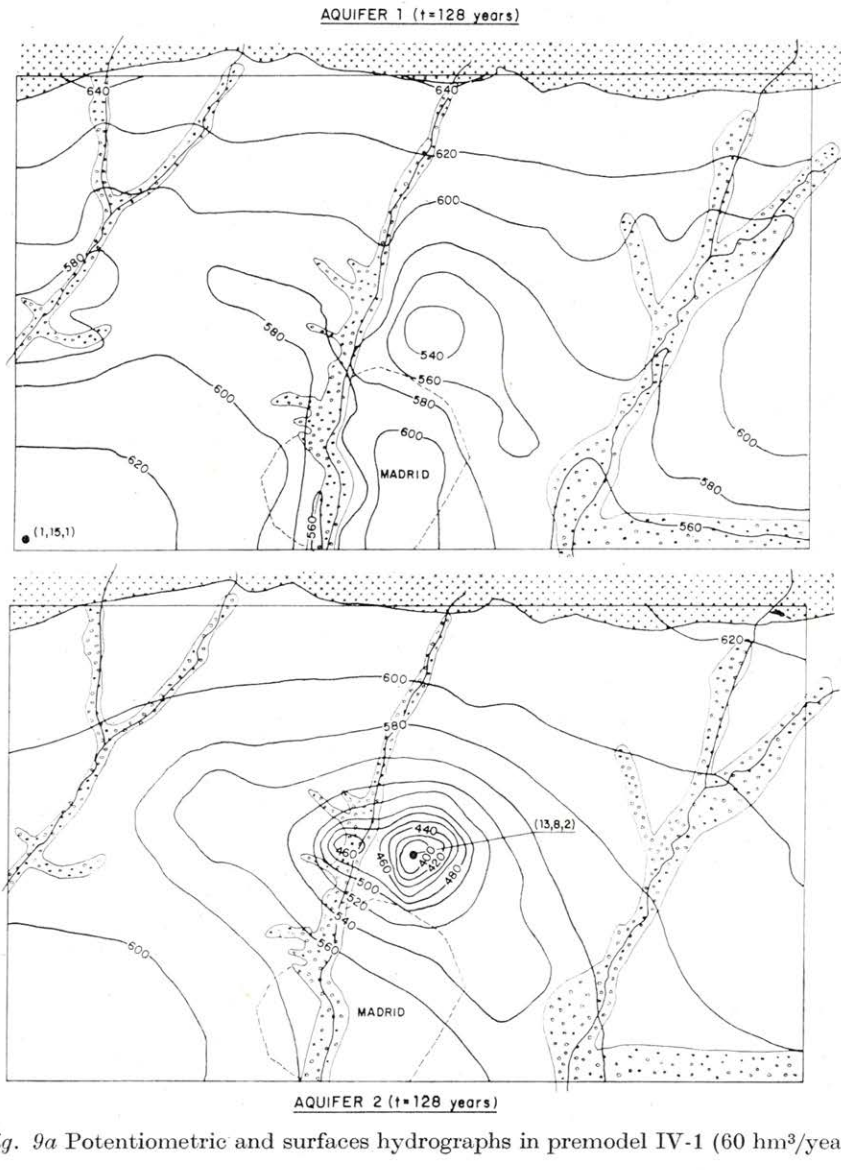
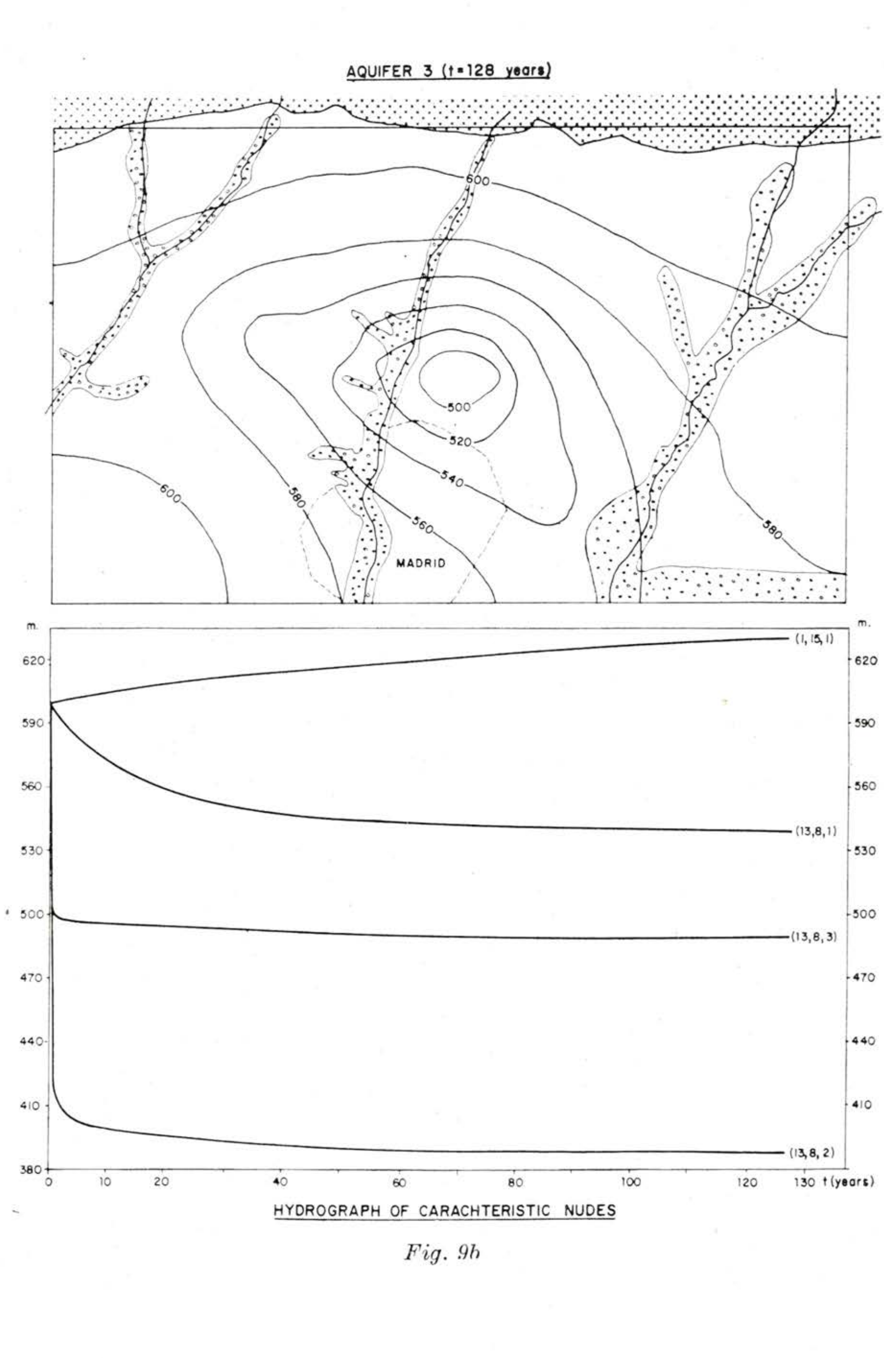


Fig. 9a Potentiometric and surfaces hydrographs in premodel IV-1 (60 hm³/year)


AQUIFER 3 (t=128 years)
600
500
520
540
560
580
600
580
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m.
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380
(1, 15, 1)
(13,8,1)
(13,8,3)
(13,8,2)
0
10
20
40
60
80
100
120
130 t (years)
HYDROGRAPH OF CARACHTERISTIC NUDES

Fig. 9b

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A STUDY OF THE RELATIONSHIP BETWEEN GROUND-WATER RESOURCES AND SUBSURFACE GEOLOGICAL STRUCTURE IN THE NORTHWESTERN PART OF THE GREAT HUNGARIAN PLAIN

ÉTUDE DE RELATION ENTRE LES RESSOURCES EN EAUX SOUTERRAINES ET LES STRUCTURES DE SUBSURFACE DANS LA PARTIE NW DE LA GRANDE PLAINE HONGROISE

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RÉSUMÉ

Les auteurs ont étudié le régime des eaux souterraines sur un territoire de 2500 km^2 au sud du territoire métropolitain de Budapest. Au cours de dresser la carte hydrogéologique à l'échelle de 1 : 100 000^e plusieurs méthodes de calcul et d'établissement ont été mises au point.

A partir de la relation et de la répartition entre les niveaux d'eaux statiques des aquifères du Pannonien supérieure et les structures de subsurface et les gradients géothermiques on peut déduire un écoulement local des eaux souterraines de la direction SE—NW, perpendiculaire à l'écoulement régionale. Cet écoulement local des eaux souterraines se déroule entre le bloc (« horst ») mésozoïque de Bugyi-Ürbőpuszta possédant une zone de descente et le « graben » de Délegyháza-Alsónémedi avec une zone d'ascension. Dans la région des eaux de descente la dépression locale indiqué comme une anomalie residuelle peut-être décrite d'après la formule Dupuit-Thiem et le débit des eaux de descente peut-être estimé à 1452 l/min. Au cours de l'étude de la dépression les fractures construites d'après les caractéristiques conçues par Schmieder s'accordaient avec les failles bien connues dans la région.

Dans la région des eaux de descente les composants verticaux des gradients piézométriques ont une valeur négative et ils ont la valeur positive dans la région des eaux d'ascension. Les anomalies géothermiques locales vérifient les conditions d'écoulement analoguement aux conditions de la pression. Le débit des eaux de descente est de 1209 l/min calculé d'après l'équation *(4)* de la variation du flux de chaleur local.

Dans la région de Délegyháza—Alsónémedi caractérisée par des eaux d'ascension on peut constater un surplus de chaleur et les valeurs positives locales des composants verticaux des gradients piézométriques augmentent vers la profondeur. La répartition des concentrations de l'ion de chlorure, de la teneur en gaz dissous et de la dureté des eaux souterraines et leurs conditions géologiques locales indiquent que le surplus local de chaleur peut être la conséquence de l'alimentation indirecte provenant des réservoirs d'eaux karstiques. Ce phénomène est conforme à la théorie de l'écoulement profonde d'après Vendel—Kisházi.

On peut constater que les eaux souterraines de différents types (eaux karstiques, eaux de formation et eaux phréatiques) forment un système d'écoulement unique et continu. L'ascension des eaux karstique thermales vers les gisements du bassin se déroule au bord du S et SE des roches carbonatées de la géosynclinale de la Montagne Centrale Hongroise près de la ligne du lac Balaton.

Theoretical hydrogeologic and hydrologic studies made during the last 15 years could give better ideas about the general rules of the underground water movements within the Hungarian Basin. Relationships between the ground-water reserves in aquifers of different geologic age, moreover some

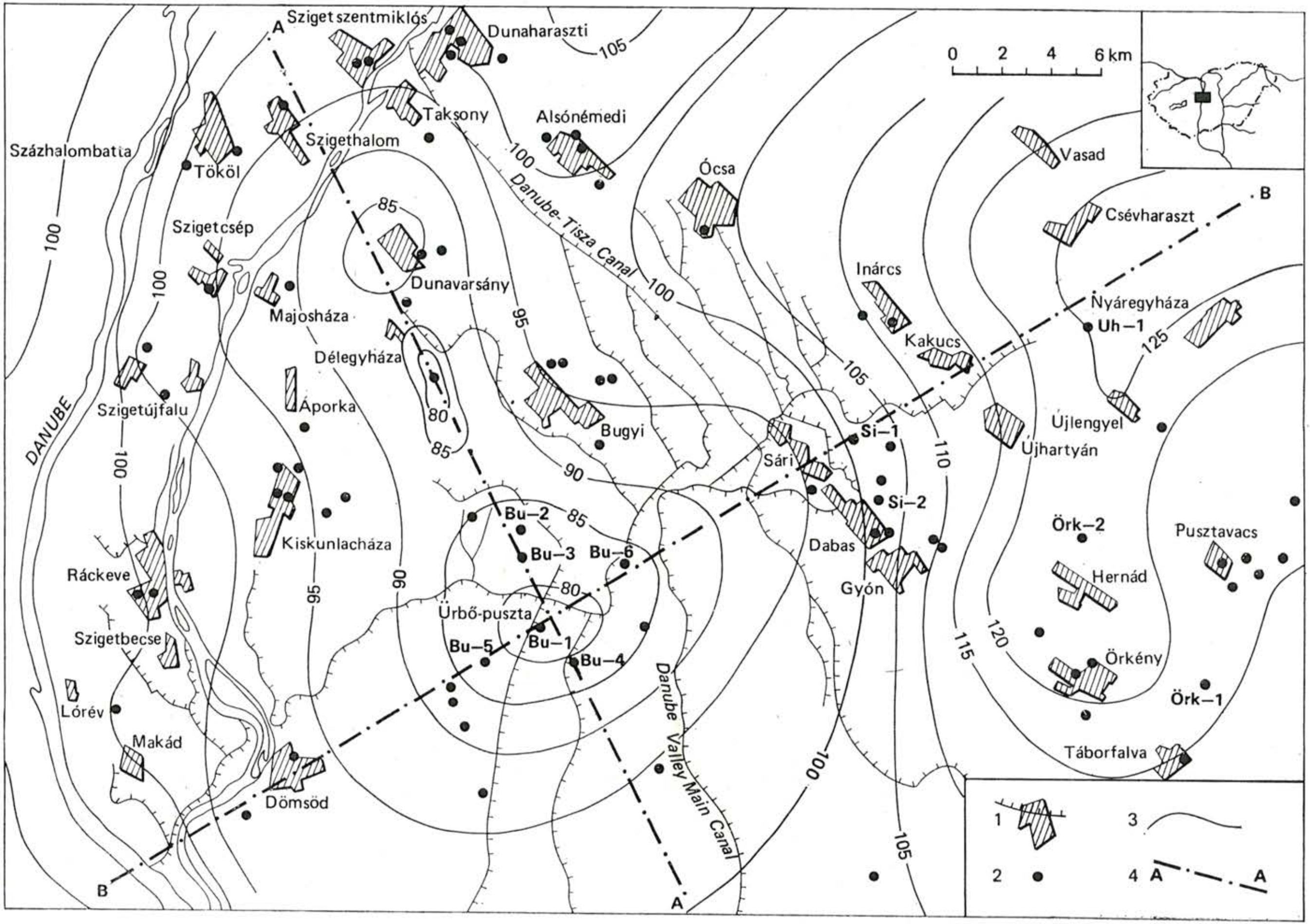


Fig. 1. The study area and the water level contour map of the Upper Pannonian aquifers

1. The inner areas of communities, drainage network, *2.* deep drillings and wells, *3.* average piezometric surfaces of the Upper Pannonian " Aquifer system for Water Work " m a.s.l., *4.* direction of the hydrogeological cross-sections relationships of the chemical composition and pressure conditions of the formation waters, the natural water flows developing within the basin and their origin (threshold gradient of seepage by GY. KOVÁCS, 1972) were revealed and described. All the statements referred to could be made more specific in this study based upon a detailed survey carried out in an area of 2500 sq.km along the NW border of the Great Hungarian Plain (Fig. 1).

Geologic setting and structure of the study area

The Upper Permian and Lower Triassic shale-marl formations, penetrated by borehole Bu-5, may be considered as the oldest sedimentary rock known in the study area. The bulk of the buried hill is consisting mainly of Lower and Middle Triassic limestone and dolomite covered by Upper Cretaceous to Lower Eocene terrestrial sequence on the flanks and along the northern side. Borehole Bu-4, located on the southern side, penetrated pyroclastic rocks in a thickness of 600 m without reaching older formations. Basement rocks formed by Middle Triassic dolomite were drilled by drills Si-2 and Uh-1. These *carbonate rocks* could be assigned to the *Dinarid* or *Bükk Facies* and the area of this high, uplifted block ("horst") has been related to the *Igal—Bükk eugeosyncline zone* by GY. WEIN (1969) (Fig. 2). The southeastern part of the area concerned belongs to the Precambrian-Paleozoic crystalline zone, the so-called *Lóczy-ridge*, and the two structural units are separated by the *Zagreb—Kulcs Main Structure Line*. The *Alpine Paleozoic—Mesozoic series of the Central Mountains Trough* is to be found north from the tectonic graben of Délegyháza—Alsónémedi.

In geologic sections on exaggerated vertical scale (Figs. 3a, b) a strikingly marked SSE trend of the Bugyi—Ürbőpuszta asymmetrical *thrust-sheet* of Triassic age can be seen. It might be regarded as a *symmetrical horst* in a WSW-ENE direction. The thrust-sheet is not of a local character and it can be traced along the so-called Zagreb—Kulcs Main Structure Line towards Újhartyán—Farmos. Its development can be related to the intense compressional movements of the Middle Cretaceous (Austrian phase). Owing to contraction and partly to dilatation, *secondary tectonic zones* were formed (L. KŐRÖSSY, 1963) resulting in a *fault-block system built up of deep-seated and also elevated blocks* in the Paleogene. *Tertiary tectonic zones* were developed as separating these blocks into several block units. The above-mentioned horst is also related to this Neogene block-faulting process (Fig. 3b). The rocks of the Bugyi structure were exposed to karstification for a relatively short period, since they turned to be covered by detritus proceeding from the southern crystalline basement, as early as the Late Cretaceous. The crystalline basement remained in uplifted position until the Middle Miocene and, therefore, the Mesozoic block could not be subjected to a continuous karstification in the Neogene. After the subsidence of the Lóczy-ridge, only the impervious covering was eroded in some places. The horst was buried again by Lower Pannonian sediments.

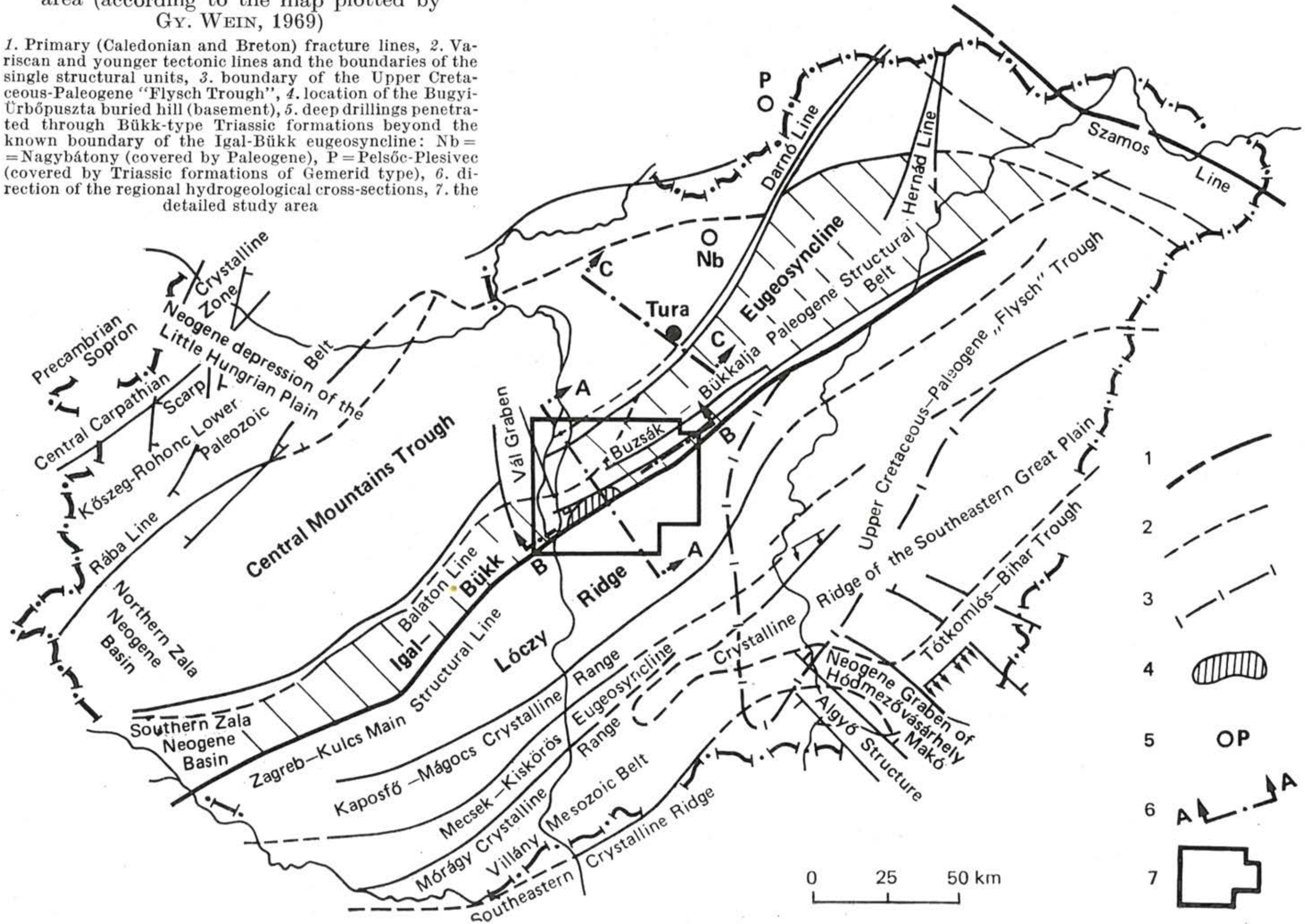


Fig. 2. Regional geologic setting of the study area (according to the map plotted by Gy. Wein, 1969)

1. Primary (Caledonian and Breton) fracture lines, *2.* Variscan and younger tectonic lines and the boundaries of the single structural units, *3.* boundary of the Upper Cretaceous-Paleogene "Flysch Trough", *4.* location of the Bugyi-Ürbőpuszta buried hill (basement), *5.* deep drillings penetrated through Bükk-type Triassic formations beyond the known boundary of the Igal-Bükk eugeosyncline: Nb = =Nagybátony (covered by Paleogene), P = Pelsőc-Plesivec (covered by Triassic formations of Gemerid type), *6.* direction of the regional hydrogeological cross-sections, *7.* the detailed study area

Hydrogeological conditions

None of the oil and gas exploratory boreholes was converted into water well. From borehole Bu-1 a flowing fresh water production of only 6 litres/minute and of 20 °C temperature was obtained from an uncased hole section ranging from 57 to 279 m in Lower Pannonian and Triassic limestones. In borehole Bu-5 an interval from 611 to 615 m was perforated in Middle Triassic limestone; a cold water discharge of 37 cu.m/day (i.e. 22 l/min) was obtained by swabbing along with an operational water level 560 m deep.

According to these data, the *Bükk-type Middle Triassic limestones show only a low-degree karstification if any.*

Karst-water occurrences, nearest the study area, are in Budapest: a new well at Budafok (1973) and that of the hospital of Tétényi street are tapping water in *Triassic limestone of Hungarian Central Mountains type,* whereas bored wells to the open-air bath at Csepel were located on *Upper Eocene limestone formation,* which form a reservoir system common with the underlying Triassic basement. These thermal waters occurring in Budapest are classified into different water types of lower temperature, but their chemical character indicates an origin similar to that of the thermal water of higher temperatures (all water types are cold karst-water). This seemingly controversial fact — which cannot be explained by the mixing of waters—has been presumed by L. ALFÖLDY (1968) on the ground of the theory of M. VENDEL—P. KISHÁZI (1963—1964), *referring to a water type that would have a shorter path of underflow.* Thus he concluded that *the direction of flow should turn back along the line of the Délegyháza—Alsónémedi graben,* which may coincide equally with the crystalline threshold separating the two geosynclines or with a tectonic contact zone. He considers this overthrust, apart from its geomechanical aspects, to be very significant in outlining *all flow-paths of the overall underflow system.*

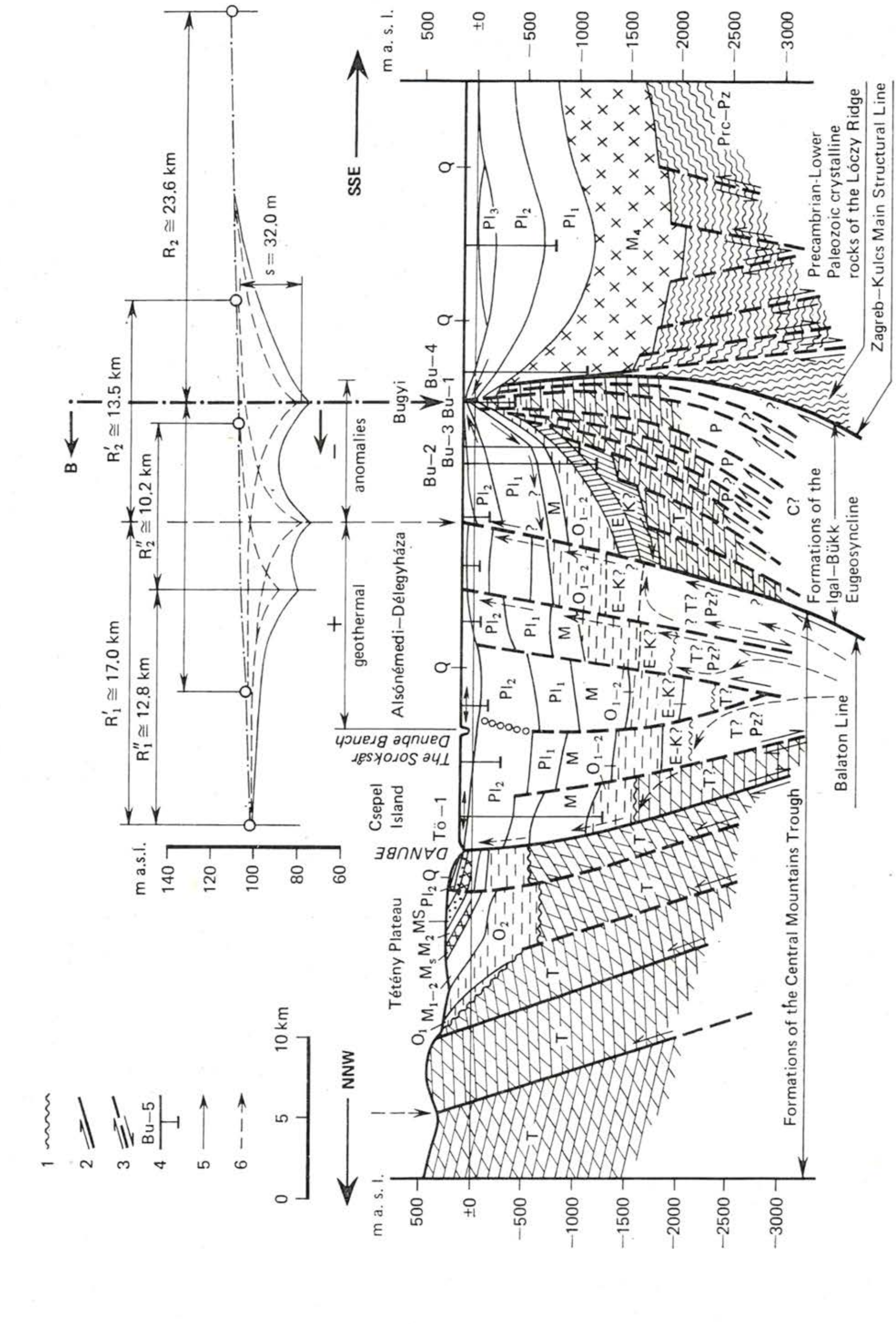
Pre-Upper Pannonian formations that overlie the basement—not to speak of the Upper Eocene limestone—are not active hydrogeologically.

The lower part of the Upper Pannonian sequence—similarly to other basin areas of Hungary—is thermal water-bearing, especially in deep-seated tectonic graben (e.g. thermal well in Ráckeve town).

The thermal water horizons are overlain by a clayey-sandy series of a maximum 600 m thickness, which ensure a gradual transition from the thermal waters (as reservoir fluid) to shallow formation waters.

The majority of drinking water wells are located on the uppermost Pannonian sand deposits (see Fig. 1). Our study is referring mainly to this aquifer system yielding potable water, which can be considered as a unit. This unit has a direct connection with the Lower Pleistocene aquifers consisting of coarse sands (it is called also "Aquifer system for Water Work").

The total thickness of this sequence varies from 50 to 200 m. Sand beds to be screened can well be correlated by means of electrical well logs. The specific yield of the individual wells is about 80 l/min/m, their maximum discharge ranges in average from 250 to 300 l/min. The water is slightly of sodium bicarbonate type, its hardness is less than 20 (German degree) and contains less than 1000 mg/l of total dissolved solids. The sodium ion concentration is increasing versus depth (E. R. SCHMIDT et al. 1962; A. RÓNAI, 1965, 1967).



NNW
SSE
m a. s. l.
Tétény Plateau
Csepel Island
DANUBE
Tö–1
The Soroksár Danube Branch
Alsónémedi–Délegyháza
geothermal
anomalies
Bugyi
Bu–1
Bu–2
Bu–3
Bu–4
B
R1′ ≅ 17.0 km
R1″ ≅ 12.8 km
R2″ ≅ 10.2 km
R2′ ≅ 13.5 km
R2 ≅ 23.6 km
s = 32.0 m
Formations of the Central Mountains Trough
Balaton Line
Formations of the Igal–Bükk Eugeosyncline
Precambrian-Lower Paleozoic crystalline rocks of the Lóczy Ridge
Zagreb–Kulcs Main Structural Line
Prc–Pz

Fig. 3a Regional hydrogeological cross-section A-A after L. Kőrössy (1953) and G. Csíky (1968). (Exaggerated by 12.5× in geological formations and by 250× in the Upper Pannonian water-levels)

Q = Quaternary, Pl_3 = Upper Pliocene (Levantine), Pl_2 = Upper Pannonian substage, Pl_1 = Lower Pannonian substage, MS = Sarmatian, M_4 = Badenian (Tortonian), M_3 = Karpatian (Upper Helvetian), M_{1-2} = Eggenburgian-Ottnangian (Burdigalian-Lower Helvetian), M = Miocene (in general), O_2 = Middle Oligocene (Kiscellian, Rupelian), O_1 = Lower Oligocene (Lattorfian), E = Eocene, K = Cretaceous, T = Triassic, P = Permian, C = Carboniferous, Prz-Pz = Precambrian-Lower Paleozoic metamorphic rocks. *1.* Unconformity, *2.* overthrust, *3.* fault, *4.* deep drilling, *5.* flow directions of ground-waters, *6.* assumed flow direction of karst thermal waters

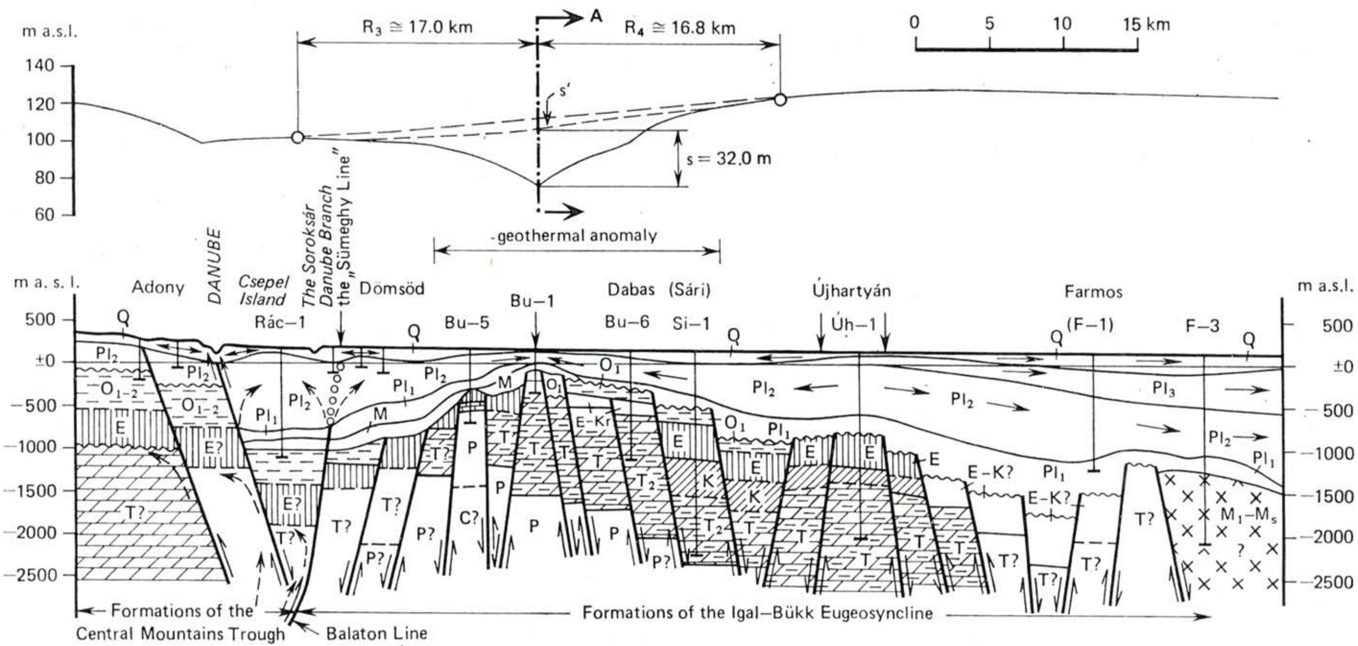


Fig. 3b Regional hydrogeologic cross-section B-B (Scale and symbols see at Fig. 3a)

The Pleistocene-Holocene gravel bed, 30 to 70 m thick, of the Danube Valley can yield 500 to 1000 l/min of water at a drawdown less than 1.0 m. The dissolved solids content is high in the phreatic water near the surface, whereas in deeper horizons it comes to less than 1000 mg/l. The difference may be contributed to the fast repetition of oxidizing-reducing processes taking place within the capillary fringe.

In the southern areas of the Danube – Tisza Interfluve the thickness of this sequence even may attain to 200 – 300 m. A specific yield more than 100 l/min/m can be expected, however, only from the coarse-grained deposits of the old riverbeds of Danube. The ground-water table can be deeply under the ground surface and it is not separable, even by quality, from the formation waters. Their total solids content is about 500 mg/l within the interval ranging from 100 to 200 m. The hardness is also low, due to the direct infiltration from precipitations.

Water wells characterized by dissolved gas content were first reported by J. SÜMEGHY and E. R. SCHMIDT from the northern and northeastern part of the study area, from the zone of the Szigetszentmiklós – Délegyháza – Alsónémedi line. Later investigations carried out by various companies and institutes indicated here measurable contents of methane in water of wells. The gas-bearing Upper Pannonian formation water is more concentrated, it can be characterized by more than 1000 mg/l of salt content and by a high Cl-ion content which *indicates a recharge from deeper formations* (e.g. at Szigethalom where water from a well of 250 m total depth has a salt content of 1270 mg/l and its Cl-ion concentration turned to be increasing parallel with the extent of drawdown).

Hydrodynamic survey

General relationships

One of the main recharge areas of the Quaternary and Upper Pannonian formation waters is the Danube – Tisza Interfluve as well as the Gödöllő – Albertirsa Pannonian Sand Ridge which are characterized by negative formation pressures. On the contrary, *areas along the Danube Valley and the Tisza Valley represent a discharge area of positive formation pressure.* Between these areas situated along the eastern and western border of the Sand Ridge area, there is a neutral zone where approximately *hydrostatic conditions are prevailing. The pressure anomalies are in accordance with the geothermal anomalies and with the assortment of the sediments* (L. SIMON, 1964).

Precipitation water infiltrates within the Sand Ridge area to the draining fracture lines of the Danube and – presumably – through *local* (shallow), *intermediate* and *regional* (widely extended and deep) flow-paths of the Tisza.

One of the principal proofs of the natural ground-water circulation within great sedimentary basins is that the formation pressure, in contrast with the hydrostatic one can be characterized by *excessive* or *deficient formation pressure*. Each *piezometric gradients* shows the difference between measured head and hydrostatic pressure,

Since under hydrostatic conditions the piezometric surface is nearly constant, th pressure anomalies may be characterized by $\Delta h/\Delta z$ which can be calculated from th values of static water level. In this case the pressure is measured as water head.

The gradient itself is a vector which has a definite direction at every point of the flow field. Therefore we cannot speak on vertical, horizontal, dip-oriented gradients but only on the *vertical component of the piezometric gradient.* In the practice, the values of $\Delta h/\Delta z$ referring to a given interval will be defined instead of the differently oriented components related to certain points.

Most of the authors agree upon the existence of a continuous subsurface flow system extending to the whole Hungarian Basin, with water derived ultimately from the precipitation. The relationship between the pressure conditions and their corresponding ground-water regimes have been pointed out by the water level observations made by the Hungarian Geological Institute and Research Institute for Water Resources.

Hydrodynamic survey of the "Aquifer system for Water Work"

The static piezometric surfaces of the Upper Pannonian freshwater-bearing formation is shown in Fig. 1. The water level contours refer to the *sand formation lying at a depth of about 80 to 120 m.* Reference wells were checked up according to their electrical logs and screened intervals. The water level movements follow the aforementioned rules according to their trends only — a definite water level subsidence could be observed in the Danube—Tisza Interfluve, along the Délegyháza—Alsónémedi tectonic graben and the Ürbőpuszta Triassic horst. The lowest points of water level are by 20 to 30 m deeper than they would be with regard to the drainage effect of the Danube (see the level contours in Figs. 3a, b).

It can be supposed, for the northern part of the area of depression that an ascending ground-water flow is directed from the Upper Pannonian aquifers into the Pleistocene-Holocene beds, water budget of which is negative due to a water withdrawal by gravel pits and unautorized drilled wells.

The Bugyi—Ürbőpuszta sub-depression has an asymmetrical form similarly to the basement: the *ground-water will be accumulated southwards accompanied by the lowest point of the cone of depression in the northern side. Both areas correspond to the compressional and dilatational zone of the thrust-sheet.* Regarding that in some places the basement is in direct contact with the Pannonian formations (Figs. 3a, b), *it can be concluded that in the vicinity of the horst a ground-water withdrawal from the Upper Pannonian formations occurs, directing into the fractured-shattered zones of the older rocks.* This weir-like or sinkhole-like behaviour of the basement gives an indirect proof for the thrust-sheet structure.

The observed cone of depression has no contact with the drainage network of the study area, operational level of which ranges from 95.0 to 98.0 m a.s.l. in the surroundings of the Bugyi—Ürbőpuszta horst, thus being higher than the Pannonian aquifers. The descending waters cannot recharge the karst thermal water system in S Budapest, since its pressure and temperature is higher than the same parameters observed in the thermal well of Ráckeve ($p_0 = 100$ m a.s.l.; $T = 71$ °C at a depth of 1059 m), not to speak on its improbable connection with the Ürbőpuszta horst consisting of Bükk-type sediments.

Since the fact of the ground-water ascension in the vicinity of Délegyháza—Alsónémedi and of the ground-water descent around the uplifted basement (horst) is supported by hydrochemical data too, it is likely that the two areas have a hydrodynamic communication.

It follows from the spatial nature of the assumed flow (which is mainly vertical in the surrounding of the horst) that the pressure anomalies observed cannot be restricted to the Upper Pannonian "Aquifer system for Water Work" but—in lack of a regional impervious top formation (cap)—they may occur also in the Quaternary formations. The local depression shown in Fig. 1 can be seen on the map plotted by L. SIMON (1964) indicating the average static water levels of the *Lower Pleistocene* formation. On the comparison of the two maps it can be stated that the *Lower Pleistocene water levels are higher along the inferred paths of descent, and they are lower-situated along the paths of ascension, as being compared with the static water levels of the Upper Pannonian aquifers.* According to data obtained from the Bugyi-Ürbőpuszta region *the static water levels in boreholes tapping the Triassic limestone formation were lower than those of the Pannonian aquifers.* The vertical components of the piezometric gradients show negative values within the assumed area of descent and, in turn, positive values appear in the ascension area.

The *local water level subsidence* in the vicinity of the horst *is similar to that of the asymmetrical drawdown around a well which was located on a confined artesian aquifer.*

In estimating the rate of the water withdrawal, *a linear superposition is to be taken meaning that the drawdown resulted by the common effects of natural discharges* (Fig. 1) can be formed by *the sum of the individual drawdowns, due to the influence of the various discrete discharges.* In this case the single local depressions can also be reconstructed from the actual water levels by means of the *method of residual anomaly* and *digital filtering process* applied widely in geophysics (K. KIS, 1974).

The "Map of drawdown", shown in Fig. 4a, was plotted by using a map by L. SZEBÉNYI (1971), which refers also to the aquifers lying at a maximum depth of 200 m below the ground surface. The depression of residual anomaly was checked up by the method of A. SCHMIEDER (1971).

If the depression shown in Fig. 4a can really be described by the DUPUIT—THIEM equation, then the values of water level subsidence, s_i, corresponding to the arbitrary points along the profiles, have to form straight lines in the function of the adequate logarithmic distances, ϱ_i, measured from the centre of the cone of depression.

In this case, on the grounds of tangents of the obtained straight lines the average transmissivity of the Pannonian formations can approximately be determined by the following equation

$$T_{av} = (k_h \cdot m)_{av} = -0.366 \cdot \frac{Q}{\operatorname{tg} \psi}, \tag{1}$$

where ψ is an angle between the straight lines and the horizontal (logarithmic) axis (A. SCHMIEDER, 1971).

It is shown in Fig. 4b that the profiles display really straight lines in accordance with the assumption, and the so-called *flow characteristics* or *curves* show typical breaks. According to equation *(1)* the transmissivity is suddenly changed at these break-points, which is resulted from the change either in the thickness "*m*" of the layer or of the seepage coefficient, k_h. In a geological sense both possibility can be inferred to structure lines, i.e. faults which may have been formed before or even after sedimentation.

The tangent of the individual curves is conspicuously low near the lowest points of the drawdown, which testifies the high transmissivity of the *Pannonian basal detritus* proved also by drilling in the surroundings of the uplifted blocks. It means that uplifted blocks (i.e. buried hills) should be taken into account, not only *in areas of descending waters but within areas of ascending waters near Délegyháza, as presumed.*

The break-points on the curves are shown along the corresponding profiles also in Fig. 4a, and it was found that many of them are lining up into straights. In addition, even the direction of breaks for these sections are identical here. *After linking these points several break-systems can be distinguished.* The most conspicuous of them are the *Late Neogene tectonic lines* which fully correspond to the fault lines indicated by K. MIKE (1971). The Pannonian block cropping out near Dunavarsány outlines itself when plotting the profile.

In the quantitative estimation of descending ground-waters the DUPUIT—THIEM equation can be applied but the calculation be carried out separately in the structurally privileged directions. Since it is not allowed by the poor accuracy of the basic data, thus only *an approximation is available* on the ground of regional averages. Average horizontal seepage coefficient of the Upper Pannonian "Aquifer system for Water Work" is of $k_{h_{av}} = 10^{-5}$ m/s, the thickness is $m = 50$ m; the drawdown, $s = 32.0$ m; the average effective well radius, $R_{av} = 19.6$ km $= 1.96 \cdot 10^4$ m (Figs. 3a, b). The effective well radius, r_0 was determined to be 307 m on the basis of the extent of the basal detritus. The value of $\ln \frac{R_{av}}{r_0}$ was defined thus 4.15 which can be considered as a real value. After substituting these date, the DUPUIT—THIEM equation can be written up for a well located on a confined aquifer as

$$Q = \frac{2 \cdot \pi \cdot m \cdot k_h \cdot s}{\ln \frac{R}{r_0}} = \frac{2.0 \cdot 3.14 \cdot 10^{-5} \cdot 50.0 \cdot 32.0}{4.15} = \tag{2}$$

$$= 0.0242 \ \text{m}^3/\text{s} = 1452 \ \text{litres/minute.}$$

That is, the water withdrawal is equivalent to the discharge of two tube wells (for irrigation) located on the gravel bed in the Danube Valley or to the discharge of 6 water wells (for drinking water supply) located on the Upper Pannonian sand formation. Regarding its entire water budget, this water withdrawal is insignificant, since one portion of this discharge is recharged from the top-situated Pleistocene formation waters.

The hydrodynamic unity of the subsurface waters

The pressure anomalies of the "Aquifer system for Water Work" and the quantity of descending waters in the surroundings of the horst can supply only indirect informations about the natural underground water circulation of the sub-basin. In studying the relationship between the character of the water movement and the geologic-tectonic setting the *pressure distribution of the overall flow-field* has to be known.

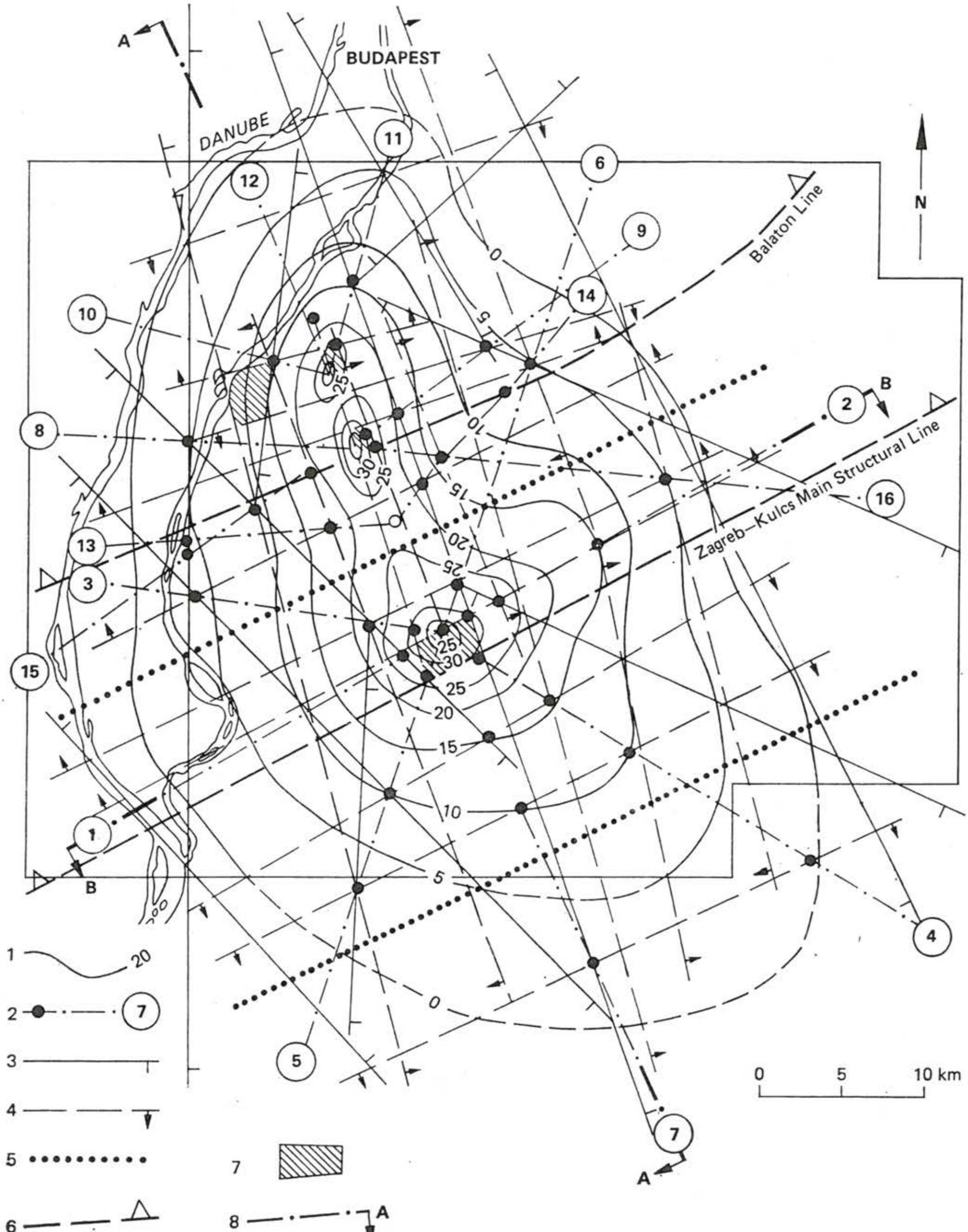


Fig. 4a Fault-tracing by analysis of the depression surface

1. Water level difference between the water level contours of the map after SZEBÉNYI (1971) and **Fig. 1**, *2.* location of drawdown profiles and break-points, *3.* most recent (Post-Pannonian to Quaternary) fracture lines, *4.* older faults inferred when plotting, *5.* older faults of no hydrogeological significance, *6.* primary (main) tectonic lines, *7.* uplifted (elevated) fault-blocks, *8.* location of hydrogeologic cross-sections

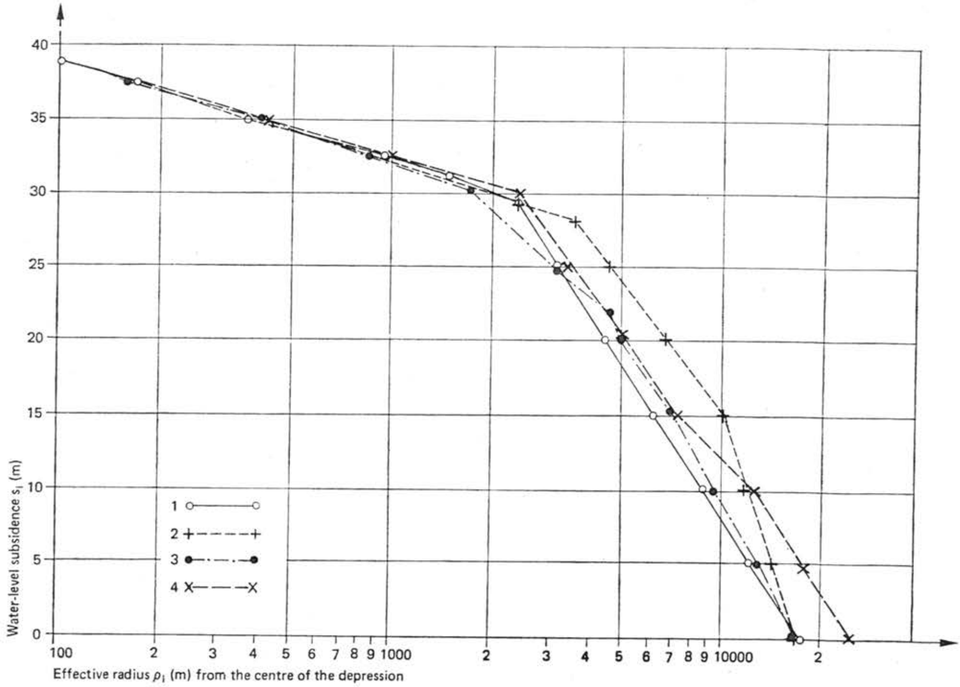


Fig. 4b Analysis of flow characteristics by drawdown profiles

The piezometric surfaces were analyzed by the following discharge levels: +100, +50 and ±0 m a.s.l. It has been stated that the *near-surface water levels were adjusted to the ground surface conditions*, to minor water courses and drainage channels whereas *versus depth they would reflect gradually the subsurface geologic conditions* (Fig. 5, for example). Fig. 6 shows the differential equations of the water level contours, that is, the *vertical components of the piezometric gradients* within the discharge interval of ±0 and +50 m a.s.l. The values $\Delta h/\Delta z$ differ greatly by various depth-intervals and *their absolute value is usually decreasing near the impervious bottom formation*, in accordance with the drop in discharge, together with horizontal discharges and growing compaction. *In the surroundings of the Délegyháza—Alsónémedi tectonic graben* or *trough*, however, *higher and higher positive values were obtained versus depth.*

Vertical distribution of the static water levels for different water-bearing horizons is illustrated by the *near-surface profiles of piezometric surface* (Fig. 7). For their plotting, water level contour maps as well as well test and formation test data of the deep water wells and of petroleum exploratory wells were applied.

The assumption of local ground-water flow between the Bugyi-structure and the Délegyháza area is directly verified by figures 5 to 7. The pressure minimum at Délegyháza, developed under the effect of water withdrawal, represents the lowest value of that area, meaning that descending waters can be moved in this direction from the vicinity of the Mesozoic block, supposedly by the

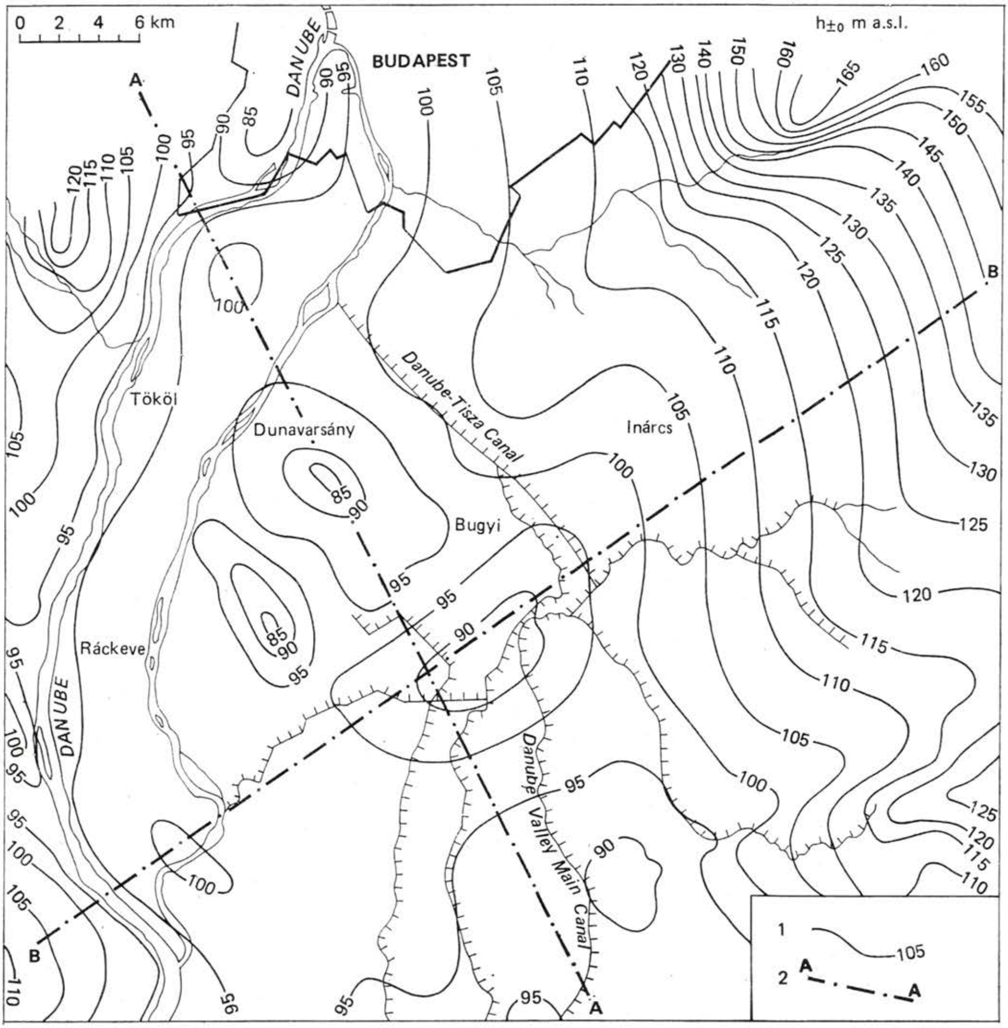


Fig. 5. Static piezometric surfaces of the subsurface waters at the water-yielding horizon of ± 0 m a.s.l.

1. Static piezometric surface (m a.s.l.), *2.* location of near-surface piezometric surface profiles

intermediary highly permeable coarse detritus lying within the interval of -200 m and ± 0 m levels (Fig. 7).

The near-surface drain effect of the Danube is acting only beyond the local flow system while the *regional ground-water flow, by-passing the local circulation, would be increasingly important versus depth* as presumed by M. Erdélyi (1973–1975).

The vertical changes in the values of the water levels and gradients might indicate that the *natural ground-water flow cannot be in correlation with geologic ages:* the impervious bottom (boundary) of the water body being in steady-state motion may be represented by rock formations of strikingly different geologic ages (Figs. 3a, b and Fig. 7). The rate of flowing water within different depth intervals may be changed not merely continuously but also abruptly due to the tectonic setting, local transmissivity characteristics, hydraulic gradients as well as to the concentrated (e.g. fault-controlled) discharge-recharge effects.

The horizontal and vertical ground-water movement (regimen) may only be defined along certain well-developed "sections" and the result might be, in every case, *of a local value* only.

The illustration of piezometric surfaces in horizontal and vertical profiles give us good ideas about the *subsurface structural conditions*. Tectonic (fault)

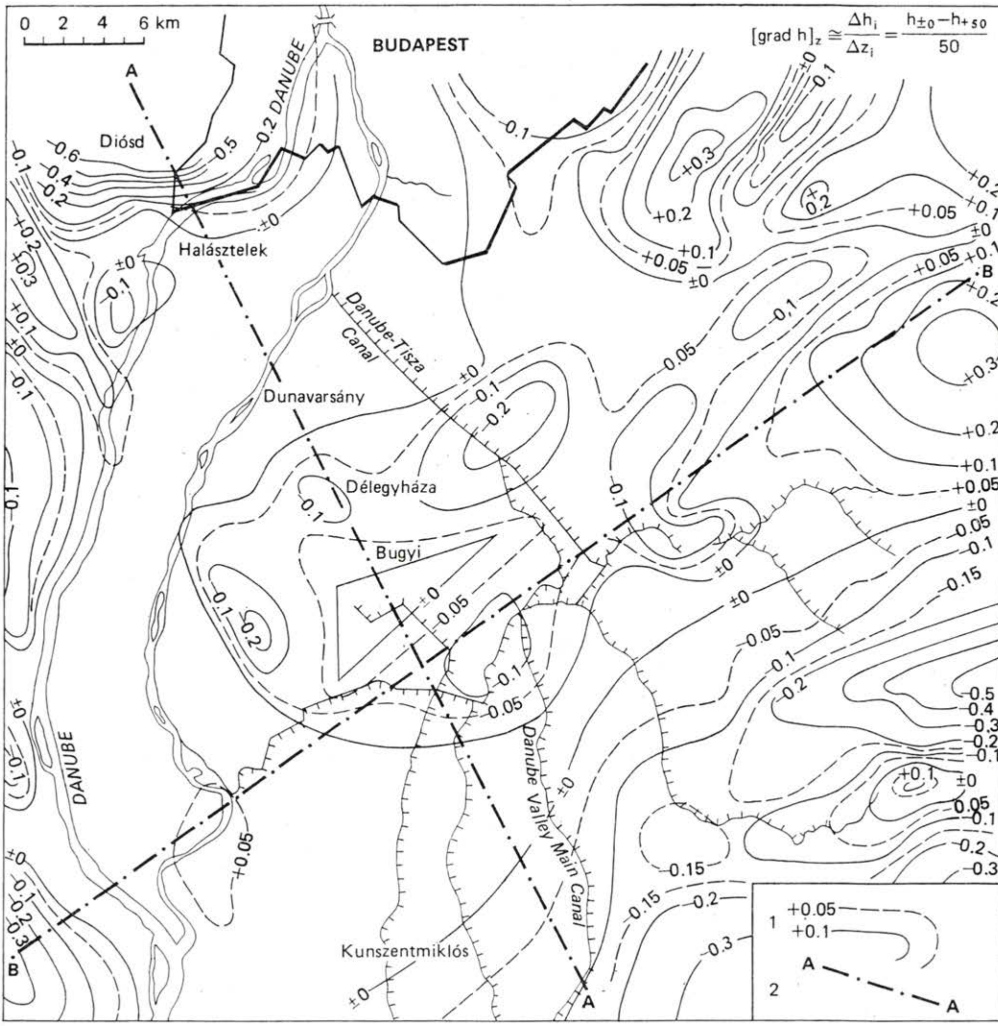


Fig. 6. Average values for the vertical components of the piezometric gradients within the water-yielding interval of $+50$ m and ± 0 m a.s.l.

1. Isogradient-line (metre/metre), *2.* location of the near-surface piezometric surface profiles

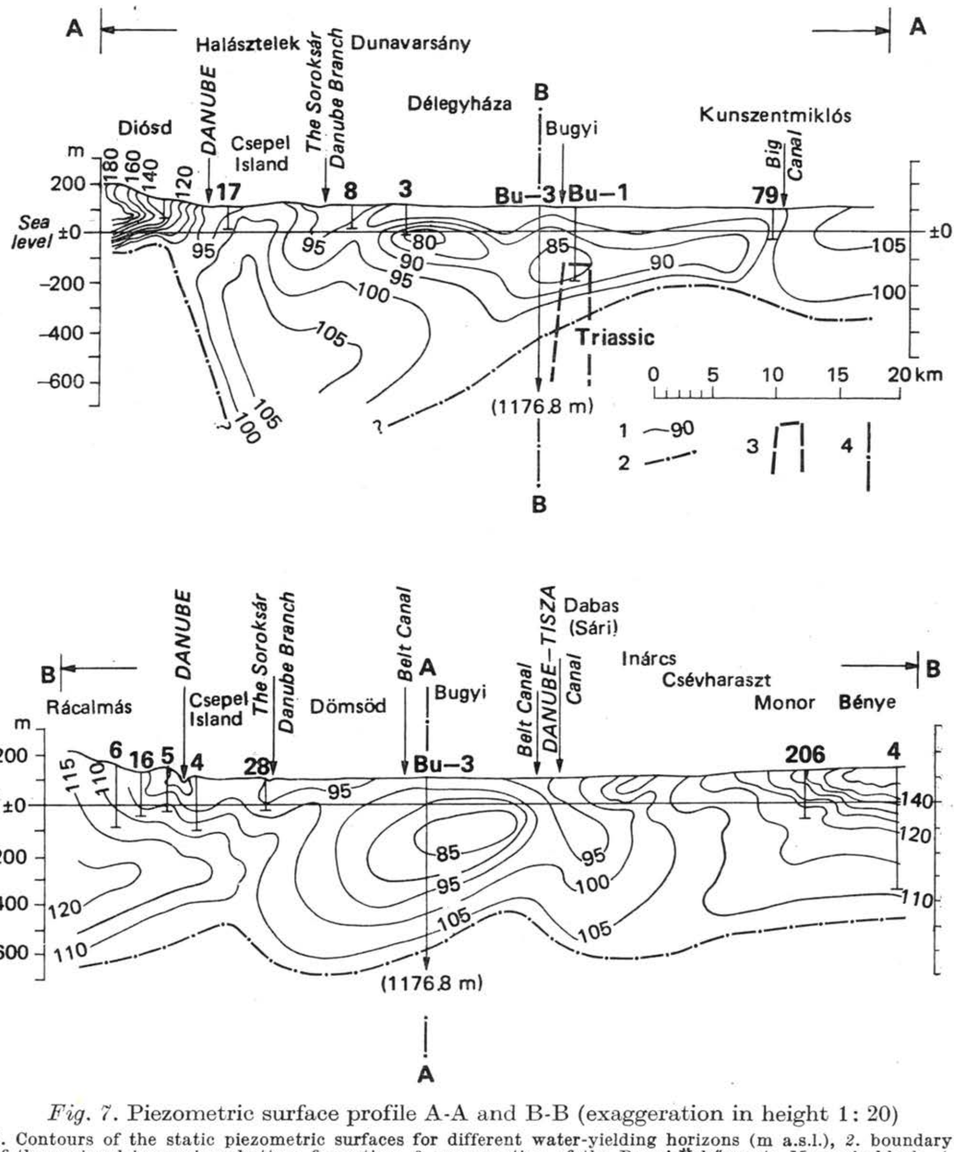


Fig. 7. Piezometric surface profile A-A and B-B (exaggeration in height 1: 20)

1. Contours of the static piezometric surfaces for different water-yielding horizons (m a.s.l.), *2.* boundary of the regional impervious bottom formation, *3.* cross-section of the Bugyi-Ürbőpuszta Mesozoic block, *4.* location of the intersecting cross-sections

lines can be here determined by *autocorrelation of the orientation of water levél contours*. The plotting and statistical evaluation of the individual maps can be carried out by computer programs widely used in geophysics. Tectonic lines recognizable by such means in the study area are identical with the fractures shown in Fig. 4a. On the basis of pressure profiles, the fault can be traced i.e. their direction determined.

Hydrodynamic studies could verify the existence of a *continuous ground-water flow system* in the study area, which also communicates with the shallow karst waters of the SE marginal Buda Mountains and possibly with the thermal water-bearing buried karst (Fig. 7).

Geothermal studies

The geothermal conditions of the study area were first described by J. SÜMEGHY (1929). On the basis of flowing water temperatures measured in shallow wells he determined the *contours of thermal anomalies* and, consequently, *Quaternary tectonic lines* were indicated in virtual coincidence with the fracture lines of greatest throw, reported by K. MIKE (1971), as well as with some structure lines indicated by hydraulic methods (Figs. 3a, b and 4a).

Only a few data of bottom-hole temperature measurements are available for the time being, beside the data supplied by the thermal well at Ráckeve. Among them a value of 7.0 m/°C (as reciprocal geothermal gradient) was reported from a 120-m-deep well at Alsónémedi, and others have reciprocal geothermal gradients ranging from 18.0 to 26.2 m/°C.

The flowing water temperature tested in wells in function of the depth of the tapped section is shown in Fig. 8a.

As a first approximation, it has been assumed that the ground-water of the Upper Pannonian "Aquifer system for Water Work" would be transferred to the Lower Pannonian basal detritus and to the Upper Cretaceous to Lower Eocene terrestrial formations, respectively.

The increase of the ground-water temperature down to a depth of 100 m is

$$\Delta t = 3.5\ °\mathrm{C}$$

and the increment of the geothermal gradient (Fig. 8a)

$$\Delta G_{t_{av}} = 0.047\ °\mathrm{C/m} = 4.7 \cdot 10^{-4}\ °\mathrm{C/cm}$$

The average heat conductivity of the basal conglomerate or that of the terrestrial formation is equally $K = 10^{-2}$ cal/cm.s.°C.

Accordingly, the increment of heat-flux variation can be expressed as

$$\Delta \Phi = \Delta G_t \cdot K = 4.7 \cdot 10^{-4} \cdot 10^{-2} = 4.7 \cdot 10^{-6}\ \mathrm{cal/sq.cm.s} \qquad (3)$$

This heat will be reduced by the water flow which is warming up during the downward movement:

$$\Delta \Phi = \mathrm{c} \cdot v_v \cdot \Delta t, \qquad (4)$$

where $c = 1.0$ cal/g·°C as specific heat of water.

The average velocity of the descending water at a reference datum level of 100 m below the ground surface,

$$v_v = \frac{\Delta \Phi}{\mathrm{c} \cdot \Delta t} = \frac{4.7 \cdot 10^{-6}}{3.5} = 1.342 \cdot 10^{-6}\ \mathrm{cm/s} = 1.342 \cdot 10^{-8}\ \mathrm{m/s}.$$

Referring to a surface of contact extending to about 7.5 sq.km within the range of the horst, and taking as minimum 20 per cent for the gravitational

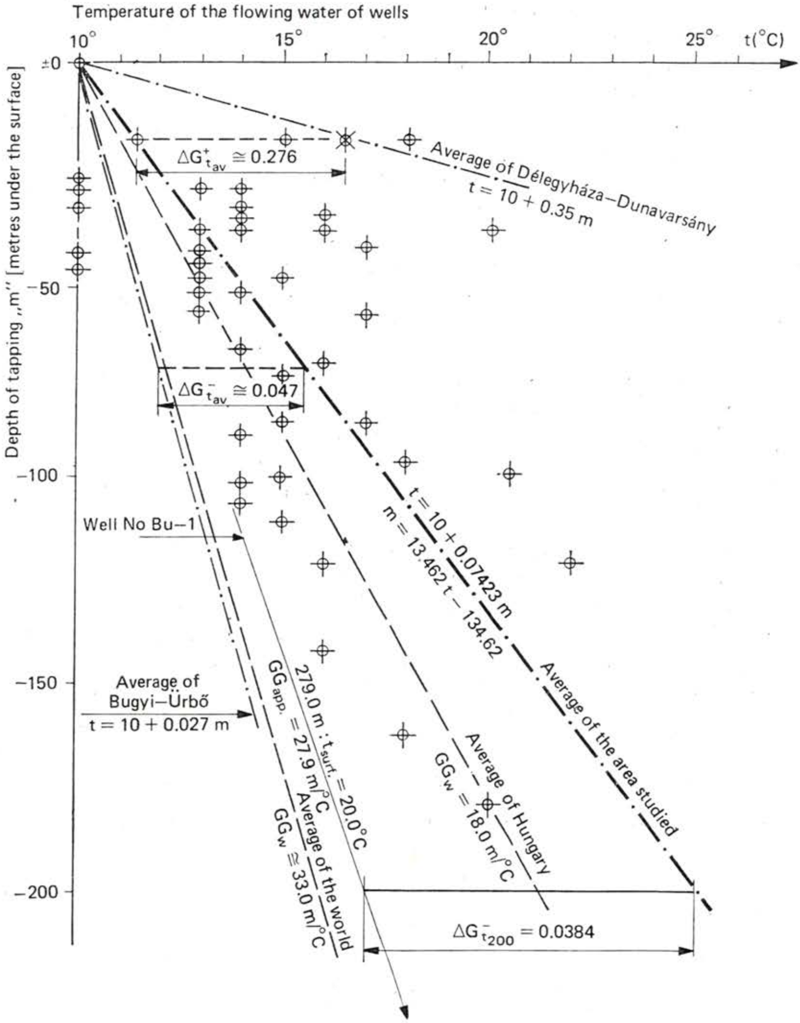


Fig. 8a Geothermal analysis of the study area

pore volume, then the cross-section of the flow system is about $1.5 \cdot 10^6$ sq.km, whereas the discharge is

$$Q = S_{flow} \cdot v_v = 1.5 \cdot 10^6 \cdot 1.342 \cdot 10^{-8} = 2.015 \cdot 10^{-2} \text{ cu.m/s} = 1209 \text{ litres/minute}$$

The geothermal computation will practically supply the same results as the hydraulic calculation. Since both methods of estimation can be considered as approximation, therefore the calculation was based on the assumption that the downward movement of the ground-water would be taking place within the basement rock itself. From the data obtained from borehole Bu-1, the increase of temperature at -200 m is $\Delta t = 7$ °C, the variation of gradient $\Delta G^{-}_{t} = 3.84 \cdot 10^{-4}$ °C/cm. The average heat conductivity of the limestone and dolomite was $K=7.5\cdot10^{-3}$ cal/cm.s.°C, the pore volume was estimated to be 4 per cent.

From these data the velocity is $3.49\cdot10^{-9}$ m/s with a discharge of 63 l/min showing the same order of magnitude as in case of the discharge from borehole Bu-5. At the same time this would suggest that *most descending waters will flow laterally before reaching the basement rock*, as predetermined by transmissivity of the carbonate basement rocks and of the overlying clastic sediments.

For the area of ascending ground-waters, velocities some orders higher in magnitude were obtained both from the average and extreme values, taking a

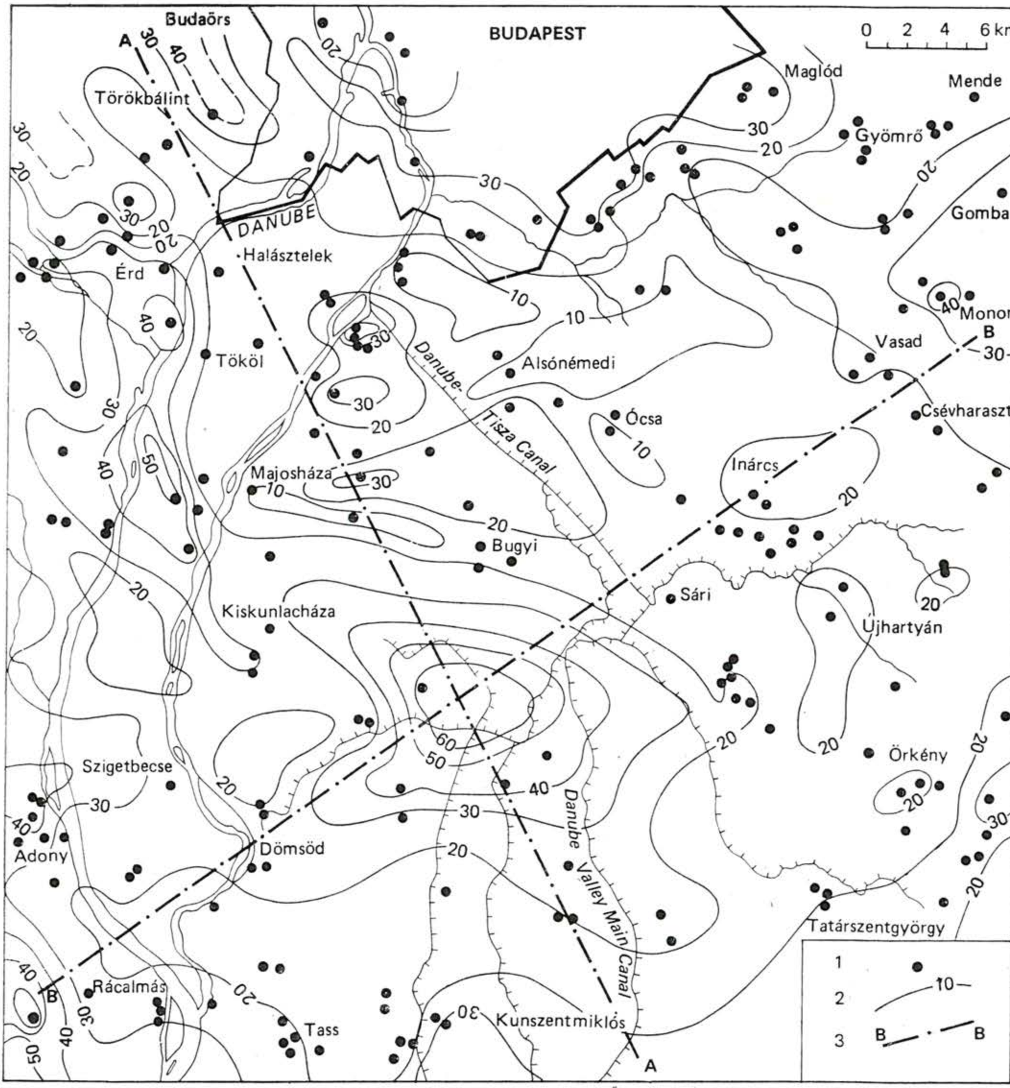


Fig. 8b Areal distribution of the geothermal gradient $GG_{app.}$ [m/°C]

1. Location of wells used for plotting of profiles, *2.* isogeothermal gradients [m/°C], *3.* trace of hydrogeological cross-sections reference depth of 100 to 200 metres and dry sand as seepage media ($K_{av} = 6.0 \cdot 10^{-3}$ cal/cm.s. °C). In determining the discharges an uncertainty regarding the flow cross-section was similarly felt.

Areal distribution of the geothermal gradient is shown in Fig. 8b. The representation required data of all wells deeper than 50 metres and, in order to estimate single values, an annual mean temperature of 10.5 °C was taken into account.

It can be seen that *within the assumed area of descending waters a high negative anomalous value has been found:* the value of G_m is higher than 60 m/°C and the isogeothermal lines will follow the WSW–ENE strike of the Zagreb–Kulcs Main Tectonic Line. This "cooling effect" of the Bugyi–Ürbőpuszta horst was firstly indicated by means of shallow soil temperature measurements made by L. STEGENA (1957). Due to the negative temperature anomaly *no thermal water development of economic significance within the area of the horst can be expected.*

On the ground of the study of local flow system and tectonic setting *positive anomalies* those observed within the Szigethalom–Délegyháza–Alsónémedi area (some spots having reciprocal geothermal gradients less than 10 m/°C) are of great importance. *Their orientation is also more definite here than in the surroundings of the horst* indicating the upward flow of the warmed formation waters along the "Balaton-Line". Similarly, areas characterized by regional ground-water flow anomalies of NE–SW trend can be revealed within the Danube-Tisza Interfluve Area and also along the Danube.

Relationship between underground waters of different types

According to concordant hydraulic and geothermal data the existence of a local flow system can be verified.

In the surroundings of Délegyháza–Alsónémedi, formation waters occur which had been descended and warmed up near the Bugyi–Ürbőpuszta horst area. Besides this, their discharge might be increased, by a lateral and even (to a lesser extent) downward-directed *cold water inflow* arriving from the top (Figs. 4a and 7). *Thus the positive anomalies of temperature would have been even greater than those shown in Figs. 8a, b.*

From this point of view the bottom-hole temperature of the well at Alsónémedi as well as the dissolved gas content, Cl-ion and total dissolved solids content of the wells located on the Alsónémedi tectonic trough (graben) are very important factors. The temperature of 126 °C, measured at depth of 1500 m of the borehole Tököl–1 proves the convective heatflow of the ascending karstic water, located in a fractured overthrust zone dipping 50° to the WNW.

It should be deduced from the tectonic setting, the existence of relatively uplifted blocks occurring along the "Balaton Line" and also from the pressure distribution that *the surplus of heat responsible for the Délegyháza positive anomaly can indirectly be derived from the ascending karst thermal waters.*

Regarding the regional tectonic setting (Fig. 2) and the so-called "underflow-theory" suggested by M. VENDEL–P. KISHÁZI (1963–1964) *such a conclusion can be drawn from the results of our study that the ascension of karst thermal waters are bound to the southern and southeastern boundaries of the Hungarian Central Mountains Range, and the flow-paths of the upward-moving* *waters are determined by the structural (tectonic) characteristics of the "Balaton Line"*. The circulation of formation waters are closely related to the ascending zone and *all the underground waters form a common, continuous groundwater flow system.*

Conclusion

1. Within the Pliocene (Upper Pannonian) aquifers of the study area a deep water flow of SE–NW direction can be indicated between the buried Mesozoic basement of Bugyi-Ürbőpuszta, representing an area of descending waters and the tectonic graben of Délegyháza–Alsónémedi as a zone of ascending waters. The current drainage of the ascending deep waters is man-made.

2. The cone of depression affected by the deep water flow shows a close relationship with the occurrence of the water-bearing, coarse-grained clastic sedimentary sequence and with the major tectonic lines developed in the study area (Figs. 3–4).

3. Due to the geological conditions, the rate of descent of water can unambiguously be determined only near the Mesozoic block of Bugyi-Ürbőpuszta. It can be estimated, according to the DUPUIT–THIEM equation *(1)*, to be 1452 l/min whereas upon the equation of the local heat-flow calculation it comes to 1209 l/min.

4. The natural flow systems of the subsurface waters in the study area and their relationship with the local flow verified within the aquifers of "Waterworks" can be determined by the comprehensive study of the potential distribution of deep waters as well as that of the hydrochemical and geothermal conditions.

According to the piezometric surface maps (e.g. Fig. 5) and profiles (Fig. 7), moreover the maps of the vertical components of piezometric gradients (e.g. Fig. 6) the areal distribution of the geothermal gradients (Fig. 8b) is directional due to the tectonics and the flow conditions of deep waters highly influenced by the tectonic setting. Earlier statements of various authors can be confirmed upon detailed studies proving that the Danube is the main factor in draining descending waters within the sand ridge of the Danube-Tisza Interfluve. Due to the structural properties experienced near the uplifted block of Bugyi-Ürbőpuszta, a minor local circulation has been developed under natural conditions but distorted by artificial influences in the surroundings of Délegyháza.

5. The surplus in heat, gas-content, Cl-concentration and hardness of the deep waters in the region of Délegyháza–Alsónémedi cannot be explained by the water movements occurring within the Pliocene formations or by the artificial water withdrawal. As a consequence, a recharge through leakage deriving from the thermal water-bearing karst reservoirs of Budapest should be taken into account.

6. It can be stated that the subsurface waters of different type (karst water, deep and shallow ground-water) form a closely related flow system. An upward movement of karst thermal waters directed towards the porous basin sediments takes place at the S–SE boundary of the carbonate rock formations of the Hungarian Central Mountains Trough, along the so-called "Balaton Line".

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REGIONAL HYDRODYNAMIC REGULARITIES OF PLATFORM-TYPE ARTESIAN BASINS

RÉGULARITÉS HYDRODYNAMIQUES DES BASSINS ARTÉSIENS DE TYPE DE PLATE-FORME

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RÉSUMÉ

L'élargissement de la sphère de l'utilisation des eaux souterraines dans l'Économie Nationale exige le modèle unique hydrophysique de la Terre basé sur la combinaison des structures hydrodynamiques, hydrochimiques et hydrogéologiques des bassins des eaux souterraines. L'analyse de comparaison de la structure hydrodynamique des bassins artésiens paléozoïques et mésozoïques de l'U.R.S.S. a permis de mettre au point les distinctions principales dans les conditions de formation du champ de la perméabilité de filtration et du champ des pressions hydrodynamiques et, par conséquent, les distinctions principales dans la formation des eaux souterraines et des minéraux utiles solubles dans l'eau.

Recent problems of regional hydrogeology are closely associated with the expansion of a sphere and extent of ground-water utilization in the national economy. This primarily concerns the problems of industrial waste water disposal, the artificial ground-water recharge, the ground-water protection and the interaction between man and the earth. In this connection, there arise new problems before the satisfactory solution that is needed to develop a hydrophysical model of the earth, based on hydrodynamic, hydrogeochemical and hydrogeothermic structures of the artesian basins. Studies of this problem may determine the development of the theoretical principles of hydrogeology and thus are of great practical importance.

In the forties one of the fundamental empiric principles of the geospheric zonation of ground-waters was formulated by I. K. Ignatovich. He proposed to regard a regional hydrochemical zonation of ground-waters as being adequate to reflect the hydrodynamic structure of any artesian basin. As a result of studying the distribution of hydrodynamic pressures in the U.S.S.R. territory, a more complex nature of the hydrodynamic conditions has been evidenced.

A model of the hydrophysical zonation of the lithosphere is based on a spatial combination of physical and geological fields which constitute its hydrodynamic, hydrogeothermic and hydrogeochemical structures.

The hydrodynamic structure is defined by variations in filtration permeability and hydrodynamic pressures. A major component of the hydrodynamic structure is transmissivity as a generalized index of the storage capacity and the seepage properties of rocks. The analysis of spatial changes in transmissivity shows that in the marginal parts of artesian basins it varies from 100 to 10,000 m^2/day and depends on the age and petrographical composition of rocks. In deep parts of the piedmont troughs the value of transmissivity varies from 0.001 to 1 m^2/day and is determined only by thermodynamic conditions of the lithosphere. Azonal variations in transmissivity, exceeding, as a rule, a typical value by one-two orders, are observed within the local tectonic structures as well as in the zones of tectonic disturbances.

Another major element of the hydrodynamic structure stipulating the stages of the artesian basins of platforms are aquicludes among which global, regional and local aquicludes (according to thickness and area) may be distinguished.

Above the first regional aquiclude the upper hydrodynamic stage occurs. It includes strata of the zone of aeration, unconfined water and the upper part of a confined aquifer. This nearly corresponds to the zones of fresh water, heliothermozone and the upper part of the geothermozone. The processes of the postsedimentary transformation of rocks are developing in this stage under diagenesis and hypergenesis. A field of hydrodynamic pressures is formed mainly under the influence of infiltration from precipitations and the pattern of a river system.

A regional ground-water flow in the upper hydrodynamic stage depends on the relation between the elevations of recharge areas and bases of drainage. Ground-water movement occurs over regional aquicludes and in the marginal parts of artesian basins (beyond the development of aquicludes) over the formation of sedimentary rocks to the basement.

The lower hydrodynamic stage occurs under the regional aquiclude. It includes the zone of confined mineralized waters and brines, nearly corresponds to the zone of catagenesis (partially of early metagenesis) and the main part of the geothermozone. The field of hydrodynamic pressures herein is genetically connected with the effects of the internal energy of the earth (temperature, geodynamic stress). The mechanism of the ground-water movement in the lower hydrodynamic stage is a combination of vertical and horizontal discharge from areally bounded blocks of the earth's crust subjected to geodynamic stresses. Vertical gradients of the formation water head are by some orders greater than lateral ones. Therefore a tendency towards vertical ground-water movement along the zones of tectonic disturbances is better defined than that towards ground-water movement along the beds the transmissivity of which is negligable at great depths.

A regional study of the hydrogeological conditions of the U. S. S. R. territory shows that the classic areas of recharge (of mountain systems) do not make influence in the lower hydrodynamic stage.

Here the ground-water flow is usually missing and has a local nature in the zones of fracture formed by the internal energy of the earth in tectonically active periods. There is no water exchange here. The main process determining the replenishment of ground-water resources is dehydration of rocks.

The lack of theories regarding ground-water movement in deep parts of artesian basins provokes serious difficulties in solving a lot of problems in claryfying the conditions of the formation of ground-water, and also raises discussion when defining a concept or phenomenon.

For instance, the classification of the hydrodynamic regime types in artesian basins is of great interest for the development of fundamental trends of hydrogeology. This relates to the fact that each type of the regime of the lithosphere may define a different trend of the evolution of thermal, mineral and industrial waters, moreover oil and gas deposits.

Yet, it is a field of knowledge characterized by a complete uncertainty. Some researchers, when dealing with general regularities of the distribution and redistribution by tectonic movements of water heads, distinguish infiltration, reinfiltration, quasi-infiltration regimes. Others count with a complex of factors to distinguish infiltration, infiltration-sedimentation, sedimentation, magmagenetic regimes and many other types of regimes.

In our opinion, at a modern level of studying the point, three hydrodynamic regimes of the lithosphere may be distinguished on the basis of the recharge source of ground-water resources, genesis of seepage potential and a general level of rock dehydration in the system of the regional development of the processes of metamorphism as leading factors.

In this connection, infiltration (upper hydrodynamic stage) epigenetic and metamorphic types of the hydrodynamic regime (lower hydrodynamic stage) are distinguished.

Comparative analysis of the hydrogeological conditions of platforms based on the known distinction in the structure of their basement and sedimentary mantle leads to the following conclusions.

Age of the rocks composing the sedimentary formation defines the maturity of the postsedimentary processes of transformation into older platforms of deposits and the rather active recent development of these processes in young platforms.

Owing to this, a comparatively slight differentiation of the field of filtration permeability (except piedmont troughs) and, as a whole, high values of permeability in mesocenozoic artesian basins are observed. On the contrary, in paleozoic artesian basins the field of filtration permeability is differentiated and shows a very clear connection with geological structures and depth of rock occurrence, variations in seepage properties being of a progressive nature.

A very high tectonic activity of young platforms, as compared with older ones, is expressed by a sharp distinction in hydrodynamic situation of the mesocenozoic and paleozoic artesian basins.

Within the mesocenozoic platforms hydrodynamic pressures reach $2500-3200$ m and hydraulic gradients 10^{-3} to 10^{-2}. Within the older platforms hydrodynamic pressures, as a rule, do not exceed $450-550$ m, and hydraulic gradients vary from 10^{-5} to 10^{-2}.

In young artesian basins the main energetic source is a geostatic pressure, and in older platforms it is a hydrostatic pressure regenerated by differential tectonic movements.

Energetic differences between the platforms come to a role, only in the areas of tangential tectonic movements (stress).

TRANSMISSIVITY INDEX ALONG THE DRAVA RIVER VALLEY IN CROATIA

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Introduction

The demand for water has become rapidly increasing in municipal, industrial and agricultural areas all over the world. This growing demand has resulted in problems concerning both the quantity and the quality of water available. Surface waters e.g., if available, are getting more polluted every year. These problems have stimulated the allocation of considerable funds and manpower to the development of this industry and related sciences. Many new methods for exploring ground-water were developed or old methods improved during the last two decades.

This paper deals with the continuation of efforts to make the resistivity surveying more applicable in hydrogeology [12, 13, 14] that has recently resulted in the development of a new parameter called the "transmissivity index". Although it has been explained in detail in another paper [14], which is being published almost simultaneously with the presentation of this one, the definition, determination, and application of this parameter are outlined subsequently because of the objectives of this Conference.

The Drava River valley was selected as a case study area for the application of the transmissivity index because of the availability of numerous geophysical and other data regarding this deep Quaternary sedimentary basin, and because of the size and significance of this highly promising source of ground-water. Also, the authors have been engaged in long-term and wide-range hydrogeologic explorations of the Drava River watershed within the Croatian territory, and they just entered the last phase. These explorations comprised the inventory of ground-water phenomena and structures, the setting up of observation and test-pumping wells, chemical analyses of ground-waters, long-term monitoring of ground-water level fluctuations, resistivity survey, hydrogeologic mapping, and ground-water budgeting. They are part of a program to explore the large drainage basins in Yugoslavia in order to estimate regional hydrogeologic conditions [2].

Furthermore, the knowledge of hydrogeology of the area concerned constitutes a common interest of Hungary and Croatia expressed by a long-term intergovernmental program for the evaluation of water resources of the Drava River valley.

Transmissivity index

The resistivity sounding results in determination of the thickness, h, and specific electrical resistivity, or simply resistivity, ϱ, of beds lying beneath a survey station. Under unfavourable conditions—given e.g. by layers with frequent vertical and sharp lateral changes in lithology showing slight differences in resistivity (thick alluvial deposits) and by lacking in a sufficient number of control test holes—accurate single values of h and ϱ are difficult to obtain. Under the same conditions, the product ϱh is subject to a slightly smaller error.

The product ϱh gains particular significance when assuming an approximately exponential relation between hydraulic conductivity (coefficient of permeability), K, and resistivity for the following sequence of grain-size classes of rocks each saturated with common ground-water, as they occur in alluvial deposits: clayey silt—silt—fine sand—medium to coarse sand—pure gravelly sand or gravel [1, 9, 11, 12]. Here, the resistivity receives a property ascribed to the hydraulic conductivity and, hence, the product ϱh receives that of the Kh, i.e. of the transmissivity (coefficient of transmissibility), T.

Due to its approximate resemblance to transmissivity, the product ϱh has been termed as "transmissivity index", τ, and its unit is ohm-m² [14].

In multilayered media, $\tau = \tau_1 + \tau_2 + \ldots + \tau_i$, where τ is the total transmissivity index of the entire group of layers surveyed and $\tau_1, \tau_2, \ldots, \tau_i$ are transmissivity indices of single layers. The same relationship exists also between T and $T_1, T_2, \ldots, T_i$.

A favourable feature of the transmissivity index is given by the fact that it reflects geohydraulic properties not only of a narrow vertical "columnar section" of rock materials under a resistivity sounding station but also of a considerably wider zone around the station.

Moreover, the possibility of representing by a single number all the data obtained from a resistivity sounding station (resistivities and depths to their changes) makes τ extremely suitable for the preparation of maps showing the areal distribution of this hydrogeologic parameter. Such maps can serve particularly well in determining proper sites for drilling production wells, and even better in the construction of aquifer models for which they can be used instead of expensive transmissivity maps [14].

Physiographic setting of the Drava River valley

The Drava River is one of the longest in Yugoslavia, forming also one of the broadest valleys. After taking its source in Italy it soon crosses the border, flows a long way through Austria and even farther through Yugoslavia, to a most part in Croatia, where it runs into the Danube River (Fig. 1). Out of its total length of 749 km, it flows through the northern part of Croatia by 275 km, including 95 km along the border to Hungary [10]. The river's mean rate of flow is 340 m³/sec where it enters Croatia and this rate is nearly redoubled at its mouth [7]. The Drava is unique in Yugoslavia for its abundance of water during summertime, due to the thawing of snow in the Alps.

Within the explored area, the river valley gets steadily wider beginning with a width of 11 km and increasing gradually to 25 km on the right-hand

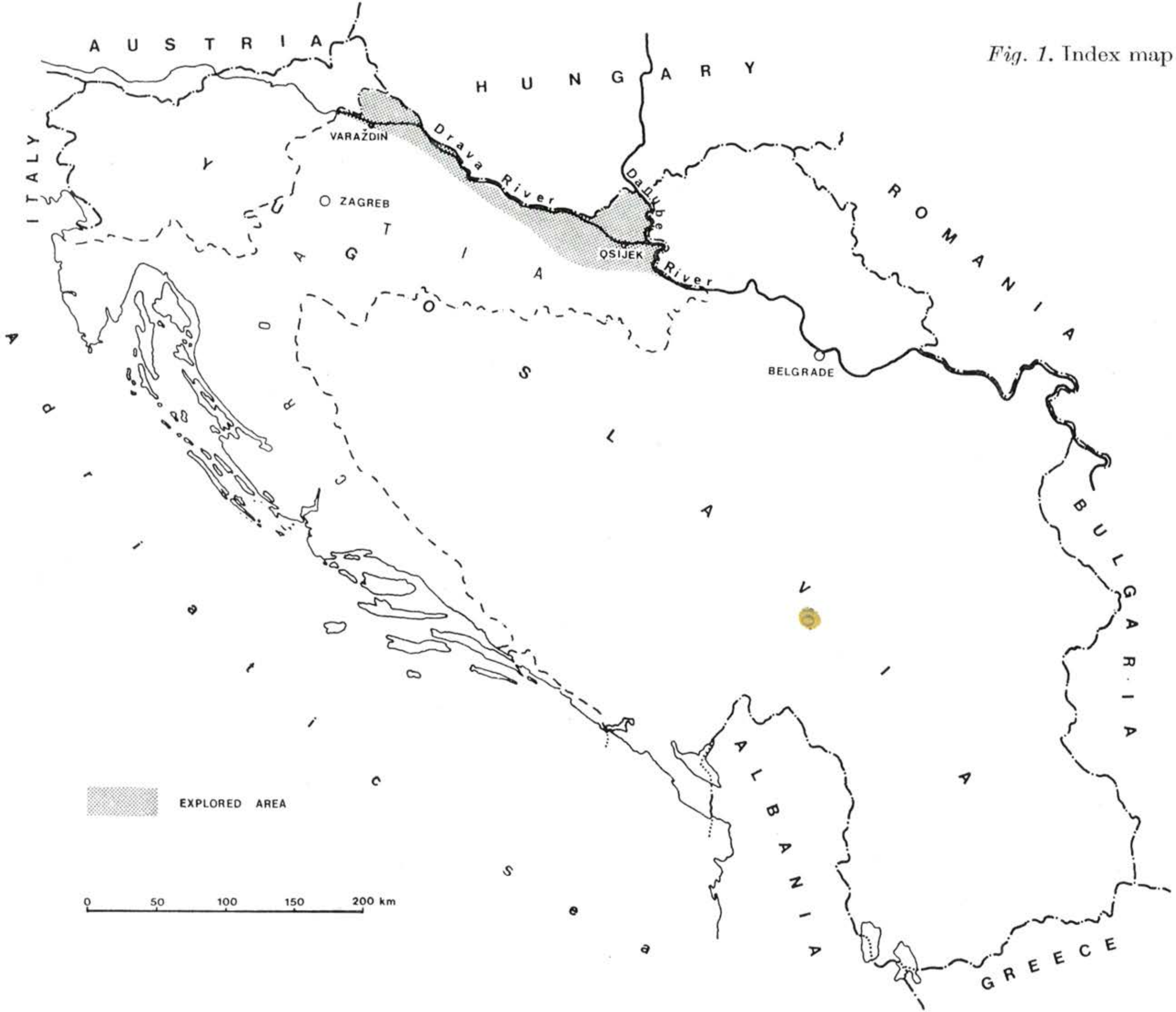

Fig. 1. Index map

riverside only, where reaching the Danube. The valley gently rises southward intersecting the slopes of a series of low mountains.

The valley, within the area studied, represents a deep depositional basin formed in a period of longitudinal faulting and sedimentation, and is filled with Pleistocene loess deposited in dry and aqueous environment and with alluvial deposits of Pleistocene and Recent ages. The alluvium mostly consists of gravel and sand, and the loess is composed of silt and silty clay. Thick layers of gravel prevail in the western part of the basin and appear as far downstream as the town of Podravska Slatina [15, 16].

The bedrock predominantly consists of Tertiary marls, conglomerates, sands and sandstones.

The uniform river fill over large span of time indicates a continuous subsidence of the basin bottom that has reached the following depths, gradually increasing downstream: from a few meters at the westernmost edge of the studied area, through 80 m at Varaždin, 150 m near Koprivnica, to probably more than 2500 m at Osijek. The loess is, in general, several meters thick, with the exception of a large area near Vukovar where it has a thickness up to 25 m and forms a plateau.

The Drava River valley is almost entirely cultivated, with numerous small towns, villages, and farms spread all over it. The water demand for domestic, rural, and industrial purposes is great but so far as the resources be available.

Transmissivity indices of alluvial deposits

For the calculation of transmissivity indices, τ, 14 traverses with 423 resistivity soundings were selected from comprehensive resistivity surveys carried out during hydrogeologic explorations of the Drava River watershed and from other minor ground-water explorations. Within each traverse, resistivity soundings ranging in distance from 0.5 to 2.0 km for each interval were set. The Schlumberger electrode arrangement, with maximum electrode separations, $L/2$, ranging from 500 to 1000 m, was applied [3, 4, 5, 6, 9]. A sufficient number of lithologic logs of deep boreholes was available to geophysicists during the interpretation of measured resistivity data.

In order to avoid the aeration zone, the shallow part of aquifers with no yield during prolonged pumping and other surface irregularities, the upper boundary chosen for the vertical interval used in the calculation of τ was 10 m below the land surface. In most cases the interval ended at the aquifer bottom. In the case of a water-bearing horizon ranging deeper than 150 m under the ground surface or when the base level of alluvial deposits was uncertain to determine, the lower level for calculation was arbitrarily established in 150 m.

In Fig. 2 are shown the locations of the resistivity survey traverses and tabulated average values of τ calculated from the single values for each sounding. The average depth of the applied resistivity soundings, the length of resistivity traverses, the number of soundings used for calculations, and the maximum single value of τ are tabulated for each traverse.

The average τ displays the ability of the unit columnar sections of the alluvial aquifer to transmit ground-water either into natural discharge areas or wells.

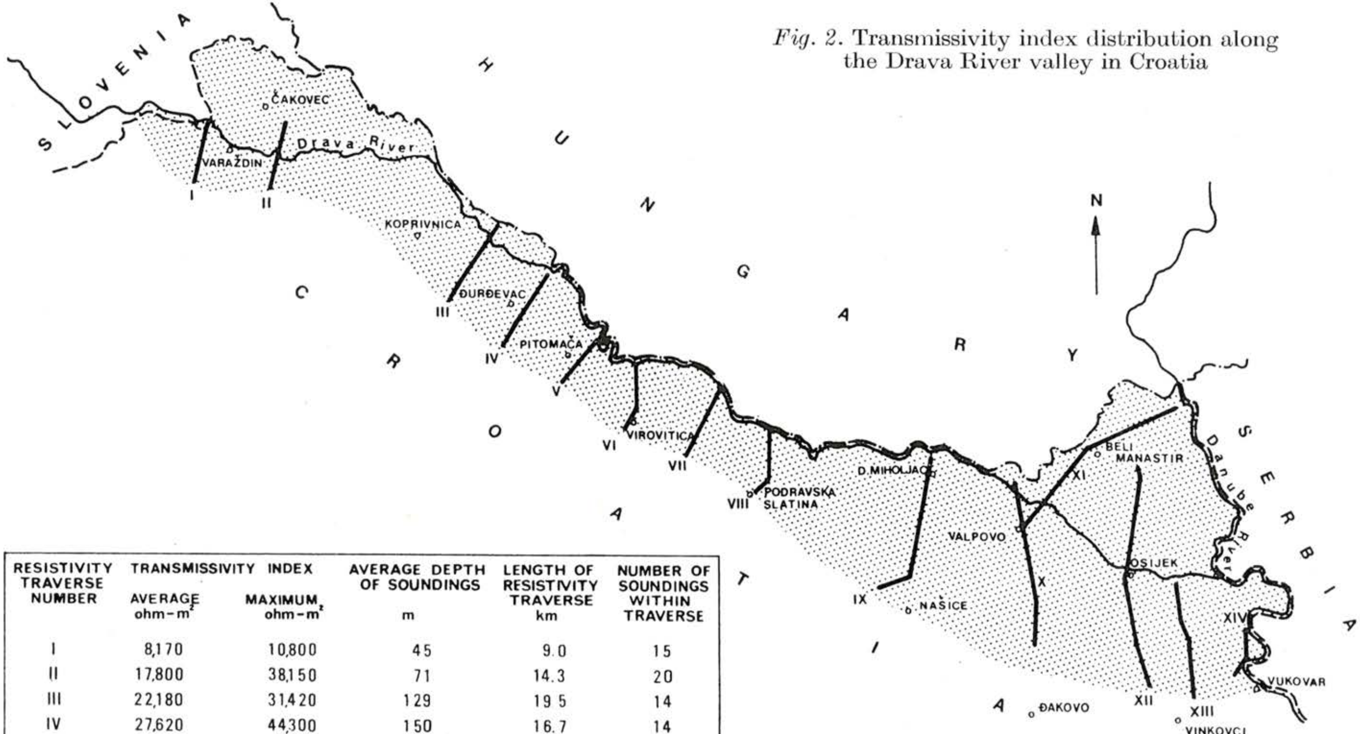

Fig. 2. Transmissivity index distribution along the Drava River valley in Croatia

RESISTIVITY TRAVERSE NUMBER	TRANSMISSIVITY INDEX AVERAGE ohm-m²	TRANSMISSIVITY INDEX MAXIMUM ohm-m²	AVERAGE DEPTH OF SOUNDINGS m	LENGTH OF RESISTIVITY TRAVERSE km	NUMBER OF SOUNDINGS WITHIN TRAVERSE
I	8,170	10,800	45	9.0	15
II	17,800	38,150	71	14.3	20
III	22,180	31,420	129	19.5	14
IV	27,620	44,300	150	16.7	14
V	18,400	32,950	143	11.5	7
VI	16,580	26,650	150	12.4	10
VII	11,690	19,110	141	14.9	32
VIII	10,030	22,750	131	12.0	26
IX	7,490	10,900	136	32.0	70
X	5,560	9,140	104	31.7	62
XI	2,700	4,750	76	37.9	78
XII	2,890	5,150	75	45.0	28
XIII	4,910	7,750	86	15.3	29
XIV	2,310	4,670	82	10.5	18

Two areas substantially differ in τ: the western part extending from Varaždin to Virovitica and the eastern part, between Donji Miholjac and the Danube River. There is also a central transitional zone situated approximately between Virovitica and Podravska Slatina.

The western part was allotted with resistivity traverses I to VI and is characterized by high transmissivity indices ranging from 17,000 to 28,000 ohm-m^2, a result of increased gravel content. Traverse I would display considerably higher τ if it were deeper, because the average depth of its resistivity soundings is four times smaller than in traverses III to VI. The western part might be subdivided into two zones, the first one including traverses I and II and the other, traverses III to VI. In traverses I and II, the alluvial aquifer is represented by one resistivity layer* only and by two layers in traverses III to VI. In this zone, although the lower layer extends as deep as 220 to 300 m below the land surface, the arbitrarily chosen common depth of 150 m has been properly applied here as the lower boundary of the interval used for calculations of transmissivity indices, from the following reasons: *(1)* it is necessary to keep equal base for comparison of τ in all traverses, and *(2)* the boundary between the alluvial deposits and the underlying poorly cemented Upper Pliocene gravels is uncertain. Thus, the assumed depth, 150 m, probably does not differ considerably from the actual one.

The eastern part includes traverses IX and XIV and displays τ smaller than 7500 ohm-m^2. In fact, if there was no difference in average depth of soundings among the traverses, all the values τ would vary only within limits of 5000 and 7500 ohm-m^2. The alluvium aquifer consists of two to three laterally and vertically interconnected resistivity layers corresponding to higher or lower sand contents in a fine-sand-to-silt-constituted medium.

The transitional zone encompasses traverses VII and VIII and is represented by transmissivity indices ranging from 10,000 to 12,000 ohm-m^2. These elevated values derive from two interconnected and very thick resistivity layers displaying rather high resistivities and representing the last but still considerably thick beds of gravel and sandy gravel in deposits where sand predominates.

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* The "resistivity layer" is a subsurface body outlined in a resistivity cross section and characterized by the lateral extent manifoldly bigger than vertical (thickness) and displaying equal or similar resistivities attributed to one layer or a group of layers with similar lithology [14].

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FORMATION DU FACTEUR DE LA CONDUCTIBILITÉ DE PRESSION DANS LE TEMPS DANS LES AQUIFÈRES FISSURÉS ET STRATIFIÉS A STRUCTURE POREUSE

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Introduction

En formant un système de cavités (p.ex. des puits) pour la production de l'eau ou d'une autre matière, p.ex. d'une matière minérale solide utilisable, dans l'écorce terrestre, l'état d'équilibre naturel existant jusqu'ici cesse et une série de mouvements de matière et d'énergie compliqués se déclenche dans le voisinage proche et plus éloigné du système de cavités. En résultat, on peut observer un mouvement des roches et de l'eau ainsi que le changement de l'énergie du système évacué, dans la série de couches au voisinage du système de cavités.

Le champ auquel les mouvements s'étendent peut être nommé champ de mouvement, et la distance entre le système de cavités et le champ de mouvement est la portée. Habituellement, on exprime la portée des mouvements par le rayon d'action $R(r, t)$ dépendant du lieu et du temps, et dans le cas de l'eau, par l'action de la dépression à distance.

Dans les roches vraies, le champ d'effet des mouvements d'eau et de ceux de roche sont unifiés et on ne peut pas les séparer, puisque le mouvement d'eau induit le mouvement de roche et la déformation de roche induit le mouvement d'eau. Par suite, l'action de la dépression à distance détermine non seulement le champ de mouvement de l'eau, mais aussi le champ d'effet du réservoir d'eau.

La connaissance sûre de la distance de l'influence de la dépression est nécessaire à la solution de plusieurs tâches, p.ex. à la détermination d'un profil de protection géologique des actions des eaux, à la présignalisation des effets nuisibles de l'écoulement forcé des eaux p.ex. des drainages miniers exercés sur l'environnement, à la détermination du champ d'influence des mouvements superficiels par suite de l'élévation de l'eau etc., ce qui rend nécessaire la détermination détaillée des relations quantitatives concernant l'étendue de la distance d'influence de la dépression. L'auteur s'occupe de ce thème à l'aide des méthodes inductives et déductives.

Caractérisation quantitative de l'action de la dépression à distance

L'action de la dépression à distance peut être caractérisée par le rayon départant du centre du champ de dépression et s'étendant jusqu'à la limite du champ de mouvement, ou bien par la valeur absolue d'un tel vecteur de lieu dans le point final, duquel la vitesse de déformation de roche ou bien la vitesse de changement du niveau d'eau ou de son niveau piézométrique est égale à zéro, c'est-à-dire $dh/dt=0$.

On peut déterminer sa valeur numérique par mesures, par méthodes analogues et par calcul. Nous employons les mesures à la détermination du champ d'influence des mouvements terminés, les autres méthodes à la présignalisation du champ de mouvement probable.

On peut déduire les relations causales nécessaires au calcul à partir de l'équation de continuité exprimant la conservation de la matière, généralement supposant que le facteur de conductibilité de pression du système étudié est constant dans le temps. En départant de cette supposition, on peut écrire aussi la relation suivante bien connue dans la littérature:

$$R=c\sqrt{at} \tag{1}$$

ce qui dépend, auprès de la constante de multiplication c, du facteur de conductibilité de pression a et du temps t.

Aux calculs pratiques, la supposition de la constance de la conductibilité de pression dans le temps veut dire qu'on néglige le changement des réserves d'eau par suite de l'écoulement transversal et de l'infiltration. Par conséquent, une déviation beaucoup plus grand que tolérable peut se présenter entre le résultat des calculs et des données de l'observation. Pour le prouver, nous présentons quelques résultats de mesure.

Détermination empirique du facteur de conductibilité de pression

Pour pouvoir suivre le changement du champ de dépression, la méthode la plus sûre est l'observation directe, la mesure directe. Nous employons à cette fin un réseau des puits pour l'observation du niveau d'eau, bien établi. En mesurant dans les puits le changement du niveau ou dans le cas de plusieurs couches, des niveaux d'eau dans l'espace et dans le temps, on peut construire le champ de dépression et en le connaissant, on peut déterminer le champ d'influence appartenant aux temps différents.

La forme du champ d'influence est en général irrégulier conformément au schéma de principe à la fig. 1, à cause de l'inhomogénéité des caractéristiques de roche et des conditions de limite différentes. Conformément aux changements, il faut étudier la formation de l'action de la dépression à distance, dans plusieurs directions.

Prenons p.ex. la situation hydrogéologique selon la fig. 2. La mine de Halimba extrait l'eau du karst stratifié pour une part couvert, ayant un ravitaillement des eaux des condensations atmosphériques. Sur l'effet du débit d'eau évacué par les cavités minières, le niveau d'eau karstique originaire s'est abaissé et un champ de dépression variant dans le temps s'est formé autour du champ d'abattage. En mettant les données du niveau d'eau mesurées à temps et distances différents, nous obtenons des droites qui donnent au

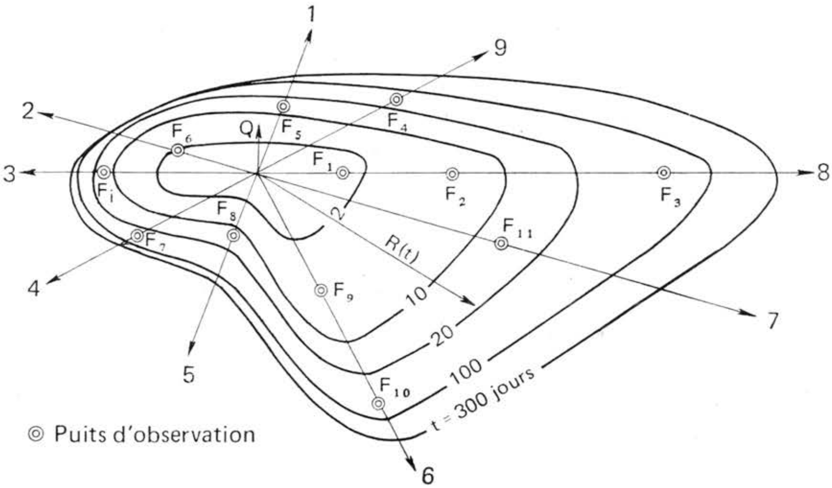

Fig. 1. Changement de la limite du champ de mouvement dans le temps

point $s=0$ l'action de la dépression à distance R_i, appartenant au temps et à la direction donnés. En représentant les valeurs $R(t)$ ainsi obtenues en fonction du t, nous obtenons pour chaque direction une courbe qui montre la formation de l'action réelle de la dépression à distance. Les courbes obtenues à l'aide de cette construction sont présentées à la fig. 3a. L'allure des courbes est caractérisée par une expression parabolique semblable à l'expression *(1)* :

$$R(t)=c\sqrt[n]{t} \tag{2}$$

qui montre que la relation déduite théoriquement poursuit bien la tendance réelle. On peut apprécier le degré de l'adaptation à partir de la formation du facteur de conductibilité de pression dans le temps qu'on peut calculer dans le cas d'une image des courants obtenue à partir de la relation *(1)*:

$$a^*=\frac{R^2}{2.25 \cdot t} \tag{3}$$

Le changement de la caractéristique déduite des valeurs R selon t est présenté à la fig. 3b.

La courbe *1* présente dans l'intervalle de temps $0-t_{01}$ une valeur augmentante, dans le domaine $t_{01}-t_{11}$ une constance de temps, et dans l'intervalle $t_{11}-t$ une diminution monotone. Les équations de corrélation de ces parties de courbe sont les suivantes :

dans l'intervalle $0-t_{01}$:

$$a^*=\frac{(kM)_1}{S_0\left[1+c_1\left(\ln\frac{t_{01}}{t}\right)^m\right]} \tag{4}$$

dans l'intervalle $t_{01}-t_{11}$:

$$a^*=\frac{(kM)_1}{S_{01}} \tag{5}$$

dans l'intervalle $t_{11}-t$:

$$a^* = \frac{(kM)_1}{c(t-t_{11})+S_{01}} \tag{6}$$

Dans ces relations

$(kM)_1$ = facteur de transport d'eau moyen appartenant à la direction *1*,
S_{01} = facteur de réservoir :

$$S_{01} = \frac{(kM)_1}{a^*_{1\,\max}} \tag{7}$$

appartenant au maximum de la courbe,

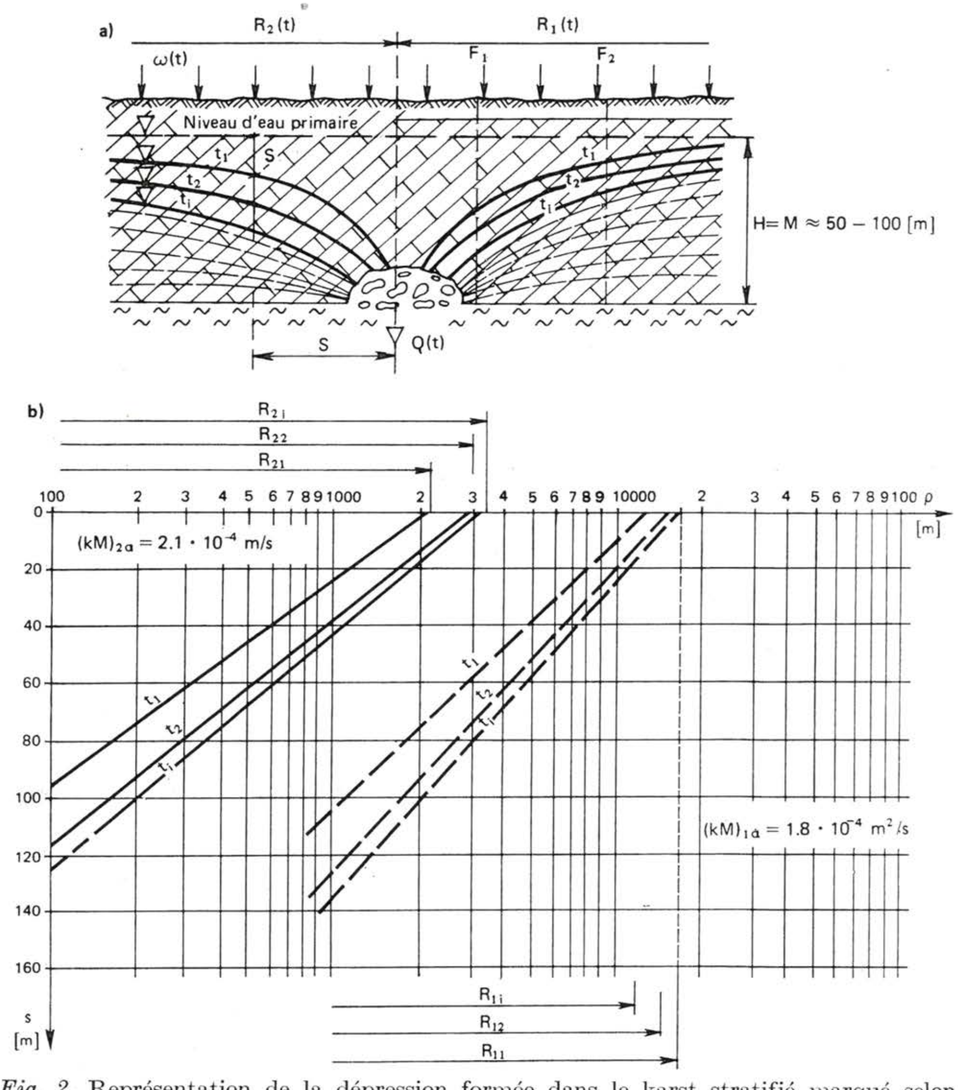

Fig. 2. Représentation de la dépression formée dans le karst stratifié marqué selon THIEM

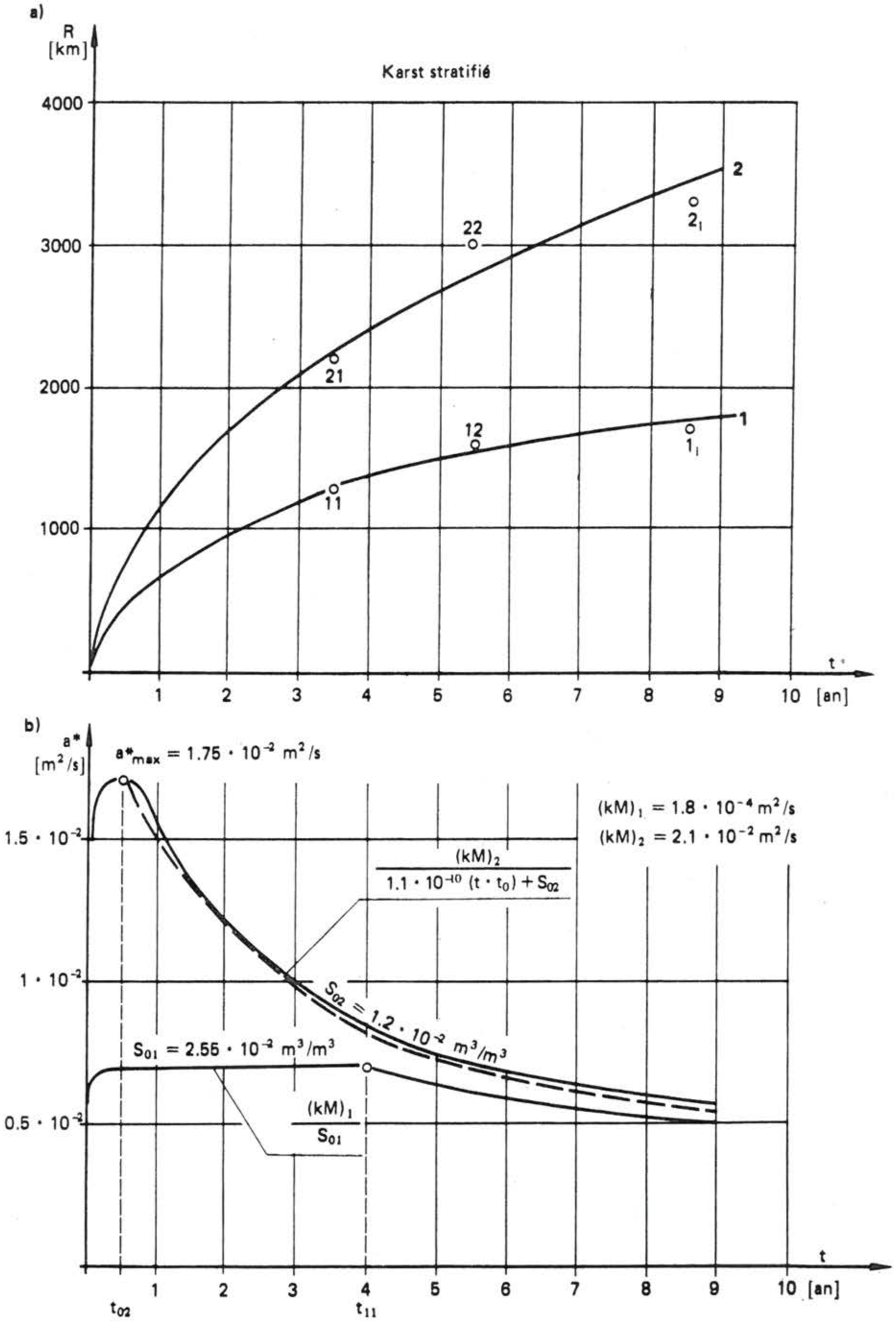

Fig. 3. Changement de l'action de dépression à distance et du facteur de conductibilité de pression en fonction du temps

m = l'exposant de la fonction exponentielle à déterminer empiriquement,
c_1 et c_2 = constantes de multiplication empiriques.

Aussi l'analyse des caractéristiques de mouvement d'eau des nappes souterraines conduit à un résultat semblable. Je présente dans les fig. 4 et 5 un exemple y relatif pour étudier les sédiments proches de la surface en utilisant les données de mesure du VIZITERV de Leninváros.

La série des couches est composée des couches successives de sable et d'argile. La série de couches et le module de compression moyen des roches sont présentés dans la fig. 4a. Sur l'effet de l'évacuation, un champ de dépression étendu se forme dans le nappes souterraines d'une puissance de 53 m environ, marqué par II. L'action de la dépression à distance présentait des différences importantes dans les directions différentes (1, 2, 3 . . . etc.) à cause des données géologiques différentes et des conditions aux limites. En résultat de ces différences, on peut observer une différence importante aussi dans la valeur absolue des facteurs de conductibilité de pression, pendant que l'allure des courbes ne présente aucune anomalie.

La connaissance des relations fonctionnelles à déduire théoriquement peut contribuer à la reconnaissance plus sûre des formes de mouvement.

Détermination théorique du facteur de conductibilités de pression

Sous l'effet de l'exhaure produite dans le puits, une circulation d'eau se met en mouvement dans les roches, dont le débit est égal au changement des réserves d'eau dans le système de roche réservoir, dans chaque moment. Cette condition d'équilibre peut être exprimée — à partir la loi de conservation de matière — par le bilan d'eau, dans le cas d'un mouvement laminaire, par l'équation différentielle suivante :

$$\frac{\partial^2 h}{\partial^2 r} + \frac{1\,\partial h}{\frac{1}{\mathrm{r}}\partial r} = \frac{S(t)}{(kM)} \cdot \frac{\mathrm{d}h}{\mathrm{d}t} \tag{8}$$

où :

h = la hauteur de la pression,
$S(t)$ = le facteur de réservoir du système,
(kM) = le facteur de transport d'eau du réservoir,
r = le rayon de la coordonnée polaire.

Dans le cas d'une valeur constante de S, la solution connue de l'équation *(8)* donne la fonction EVERDINGEN-HURST et dans le cas d'un grand t, la relation MUSKAT. Dans la plupart des tâches pratiques, le facteur de réservoir est une quantité variable dans le temps, comme le prouvent les exemples présentés, qui contient auprès du changement statique des réserves d'eau, aussi les valeurs momentanées du changement des réserves d'eau par suite de l'infiltration et de l'écoulement transversal. Puisque les réserves d'eau de gravitation n_0, ainsi que les changements élastique et de consolidation des réserves d'eau $\beta^* \gamma M$ et $\sigma^* \gamma M$ par suite de l'expression sur l'effet du changement de l'hauteur de pression par unité sont quassi-constants dans le temps à

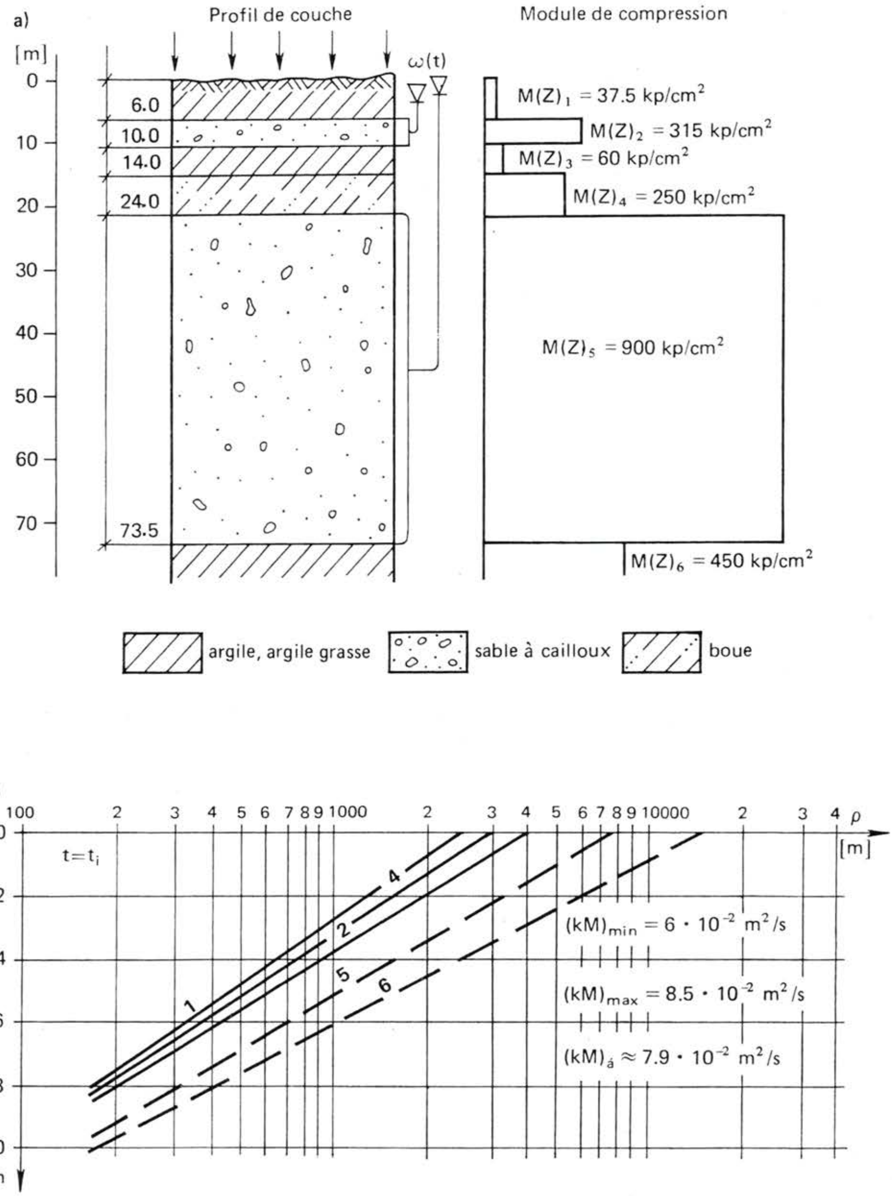

Fig. 4. Représentation de la dépression formée dans la 2[e] couche du réservoir d'eaux souterraines selon THIEM

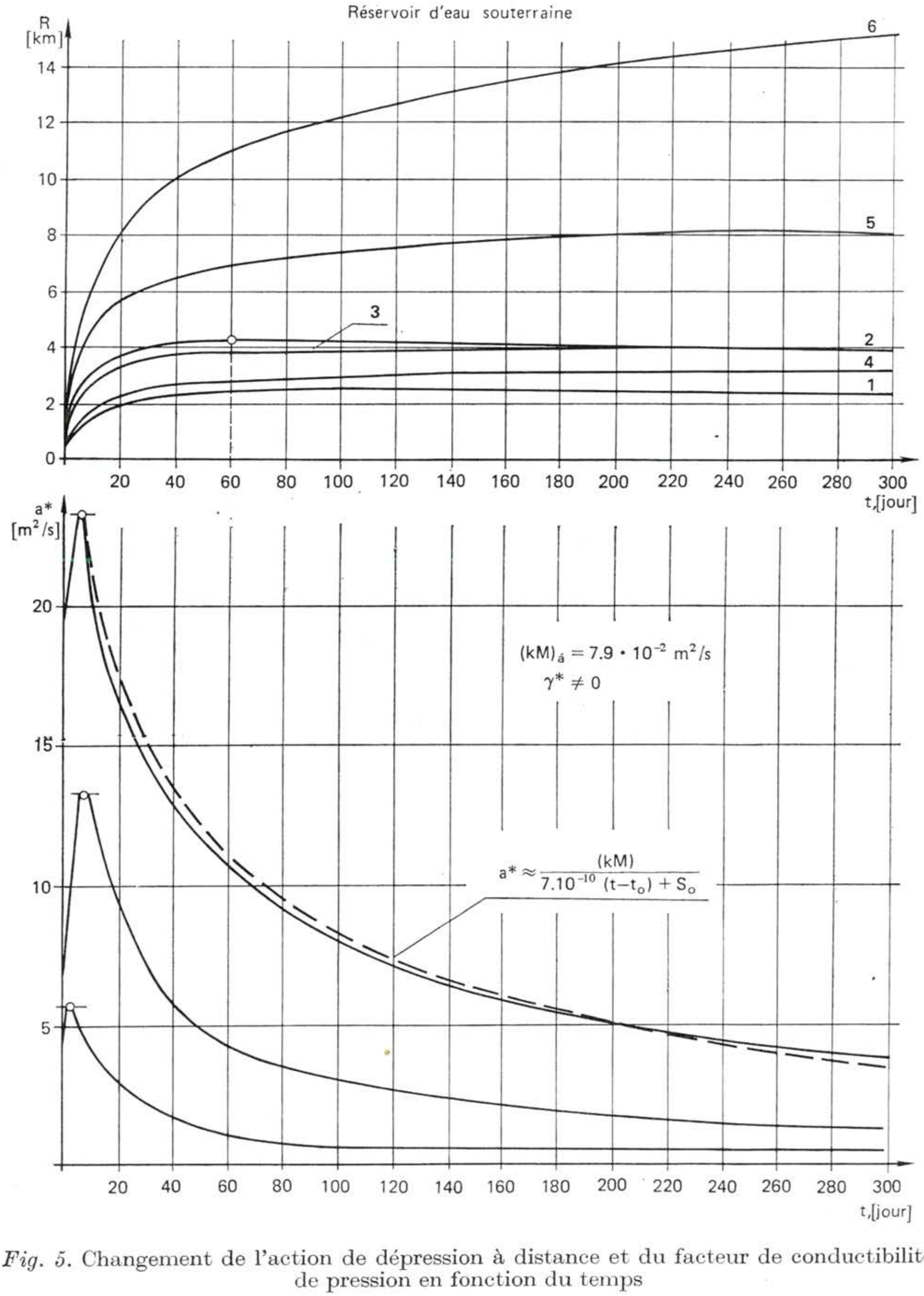

Fig. 5. Changement de l'action de dépression à distance et du facteur de conductibilité de pression en fonction du temps

partir du temps $t=t_0$, le facteur de réservoir $S(t)$ constant peut être écrit en somme de l'S_0 constant et des composants de S variables. C'est à dire :

$$S(t)=S+n_0+(\beta_e^*+\sigma_e^*)\gamma M=S+S_0 \tag{9}$$

où :

β_e^* = le facteur d'élasticité de volume résultant,
σ_e^* = la valeur réciproque du module de compression résultant,
γ = le poids spécifique de l'eau,
M = la puissance du réservoir d'eau évacué.

Pour pouvoir déterminer la composante variable dans le temps du facteur de réservoir, il faut écrire le côté droit de la relation *(9)* en prenant en considération tous les facteurs de bilan.

On obtient en résultat de la déduction la relation générale, ensuite il faut résoudre l'équation ainsi obtenue par itération. En résultat, nous obtenons pour le facteur de conductibilité de pression variable dans le temps la relation suivante :

$$a^*\approx\frac{(kM)}{\left[\frac{2\omega}{s_0}+\left(\frac{k}{m}\right)_{0e}\right]t+S_0} \tag{10}$$

Le facteur de réservoir résultant est, exprimé à partir de la formule *(10)* :

$$S(t)=\left[\frac{2\omega}{s_0}+\left(\frac{k}{m}\right)_{0e}\right]t+n_0+(\beta^*+\sigma^*)\gamma M \tag{11}$$

Son allure dans le temps est présentée à la fig. 6 en fonction du temps sans dimension. L'action de la dépression à distance est :

$$R(t)=c\sqrt{a^*t} \tag{12}$$

En comparant la relation *(10)* obtenue en résultat de la déduction à la formule *(6)* obtenue par voie empirique, on peut constater à l'aide des corrélations écrites pour les gisements étudiés que la relation causale *(12)* exprimant la conductibilité de pression variable dans le temps, détermine qualitativement bien conformément à l'expérience les processus temporels se déroulant dans les systèmes de roche réservoir.

Conclusions

A partir de l'analyse des processus de mouvement réels, on peut constater que la conductibilité de pression a^* variable dans le temps règle les processus temporels des mouvements.

La variation du facteur de conductibilité de temps dépend surtout du changement des facteurs de bilan d'eau, au premier lieu de celui des réserves d'eau de consolidation et du ravitaillement d'eau provenant de l'infiltration et de l'écoulement transversal.

C'est la formation du facteur de réservoir dans le temps qui reflète le changement des facteurs de bilan, où on peut distinguer deux sections d'un caractère différent.

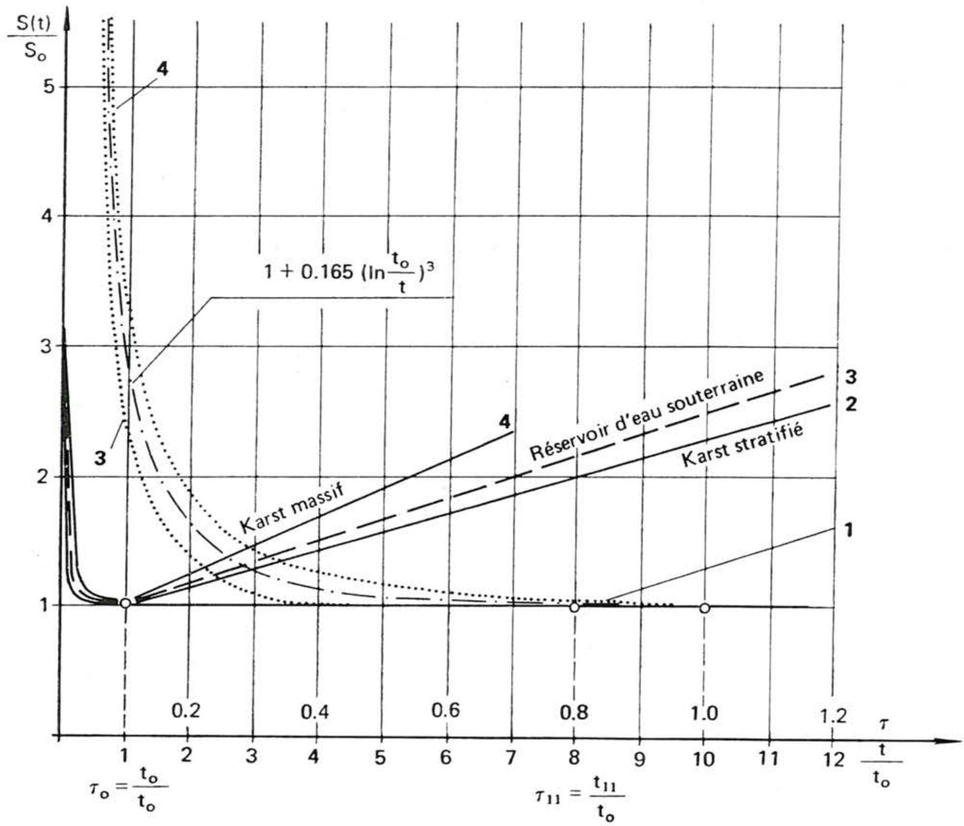

Fig. 6. Changement du facteur de réservoir relatif en fonction du temps τ sans dimension

— Pour *la première phase,* le changement intense des réserves d'eau statiques — surtout de celles élastique et de consolidation — est caractéristique. Dans ce domaine — l'intervalle $0-\tau_0$ dans la fig. 6 —, l'abaissement du niveau d'eau est très important, et par conséquent aussi la vitesse de déformation élastique et permanente, ensemble avec une expression des réserves d'eau statiques à grande intensité au commencement, ensuite à moindre intensité en fonction du temps.

— Dans *la deuxième phase* du processus temporel, le changement des réserves d'eau élastique et de consolidation devient proportionnel à la vitesse de l'abaissement du niveau d'eau; l'écoulement transversal se met en marche ou bien l'infiltration se fait valoir ce qui augmente linéairement la valeur du facteur de réservoir. Dans les systèmes sans ravitaillement d'eau, la tangente de direction de la droite est égale à zéro, comme dans l'intervalle $\tau_0-\tau_{11}$ de la droite *1*. Dans le cas d'un ravitaillement d'eau, la tangente de direction de la droite n'est plus zéro, elle prend une valeur positive. Et dans la première et dans la deuxième phase, on peut déterminer le facteur de réservoir par essais ou par calcul.

Dans la deuxième phase du processus temporel, on peut employer pour le calcul la relation théorique *(12)* qui est en accord avec les expériences.

En connexion avec les exemples présentés, il faut remarquer que dans la roche réservoir karstique fissurée à surface nue, un écoulement général à niveau en dépression se forme en sens cinétique à cause du degré différent de la karstification et de la fermeture locale des zones de failles [2].

Les données du tableau 1 montrent que dans le cas des données étudiées, surtout dans un système de sédiments aquifères granuleux une circulation intense, un écoulement transversal se forme entre les réservoirs évacués et non évacués dans la proximité dépendant de la dépression produite, de la conductibilité d'eau des épontes verticale à la stratification et de leur puissance. En résultat, il est nécessaire que dans les réservoirs d'eau d'une position plus haute ou plus profonde, une égalisation de la pression plus lente ou plus rapide se produit s'étendant comme une réaction en chaîne qui peut étendre le champ de mouvement à tout le milieu rocheux. Il s'ensuit qu'il faut étendre nos études aussi aux couches de couverture et sous-jacentes du reservoir d'eau évacué pour pouvoir reconnaître les caractéristiques du champ de mouvement se produisant dans un milieu rocheux à structure stratifiée.

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MATHEMATICAL MODEL FOR THE CONE OF DEPRESSION OF WATERWORKS IN LOOSE SEDIMENTARY BASINS

MODÈLE MATHÉMATIQUE DE DÉPRESSION PAR SERVICE DE DISTRIBUTION D'EAU DANS LES BASSINS HYDROGÉOLOGIQUES CONSTITUES PAR DES CUVETTES SÉDIMENTAIRES MEUBLES

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RÉSUMÉ

Les sédiments quaternaires aquifères de la Hongrie appartient à trois types hydrodynamiques principaux. Les cuvettes quaternaires à mince épaisseur avec une nappe libre contiennent des aquifères bicouches. A la base se situe une couche aquifère graveleuse et grossière et au sommet il y a un toit imperméable. Vers l'intérieur du bassin il y a une cuvette tripartite avec des nappes captives. Il se trouve ici une couche aquifère de sables grossiers à la base, une couche intermédiaire surtout argileuse au milieu et une couche d'alimentation des sables fins au sommet. A l'intérieur du bassin le type multicouche ayant plusieurs cents mètres d'épaisseur avec la série de sables et d'argiles est caractéristique, mais ce n'est pas le sujet de cette étude. Dans le cas du premier type pendant l'opération des puits de Service de distribution d'eau la nappe inférieure peut augmenter par alimentation direct ou indirecte (« vertical leakage ») provenant des canaux, des lacs, des cours d'eau, etc. . . . tandis qu'au cas du deuxième type par drainance de la nappe phréatique. Dans l'étude mathématique de répartition de dépression les puits du Service de distribution d'eau étaient simulés ensemble et non pas séparément comme le drainage bidimensionnel distribué sur une certaine surface, F. Pour l'intégration numérique d'une équation differentielle de dépression régionale produite par le drainage bidimensionnel c'est la méthode de différence qui est appliquée, et la réalisation des algorithmes obtenus s'est fait par la programmation de FORTRAN. L'emploi du programme est présenté par deux exemples. Dans le système de la nappe phréatique aux environs de Kalocsa appartenant au type bicouche, une dépression du Service de distribution d'eau avec une capacité de 10 000 mètres cubes par jour était déterminé à partir des nappes caractéristiques bien connues eu égard aux influences du Danube à lit complet et aux canaux incomplets. Le modèle mathématique du Service de distribution d'eau de la ville Debrecen représente l'étude d'un système tripartite d'une nappe captive. Le but de cette étude était l'optimalisation des caractères des terrains par la simulation de l'histoire de la production d'eau des Services de distribution d'eau. Le problème était résolu par la minimisation des déviations entre les dépressions mesurées et calculées.

Main types of the Pleistocene water-bearing sedimentary sequence

One of the most important basis of the municipal, agricultural and industrial water supply within the lowland areas in Hungary is the alluvial Pleistocene sequence consisting of sandy-clayey layers and lenses in alternating succession. This series, locally as thick as many hundreds of meters has a highly varied sedimentary character both vertically and horizontally. On the basis of the vertical distribution and occurrence of the coarse-grained sediments (gravels and sands) traversed and tapped by the water producing wells within the

Pleistocene aquifer system the following three main types can be distinguished (Fig. 1).

a) Coarse-grained formations of relatively shallow occurrence covered by thin clayey-sandy-silty layers of a few ten meters. In this case the groundwater production from the lower series will result a water level decline also within the cover sediments which means the so called *water-table condition.* Due to the drawdown and discharge of the leaky aquifers a vertical leakage can be taken place from the river beds, lakes and marshes not completely impervious as well as from the atmosphere since evaporation will be usually decreased owing to the decline of the phreatic water level. The Little Hungarian Plain and the border areas of the Great Hungarian Plain are characterized by this type of sediment sequence (The floodplain of the Danube, the alluvial fan of the Sajó river, the Bodrog interfluve area, and so forth). This is the so called *two-layered type.*

b) Gravelly and sandy water-bearing deposits are occurring at larger depths from 100 to 200 m overlain by thick silty-clayey semi-permeable series of regional extension. The uppermost part of this series is often sandy again, however, these sands are finer-grained compared to the sandy layers of the bottom part. The discharge of the wells located in the coarse-grained lower deposits is assured by the water released by the compression of the aquifer and by the expansion of the water, by lateral flow and by vertical leakage due to the drawdown. The vertical leakage deriving from the Upper Pleistocene aquifers of water-table conditions is proportional to the drawdown, s, developing within the discharging aquifer system since the decline of the phreatic water level may be neglected. Its reason is the low specific value of the vertical leakage from the phreatic waters as well as the sources of vertical leakage for the phreatic water as mentioned before. Therefore within the *three-parted* Pleistocene *sediment sequence* a lower water-yielding horizon, a middle transmitting horizon and an upper recharging horizon can be distinguished. The most characteristic area for this three-parted Pleistocene series is the surrounding of Debrecen where one of the most concentrated groundwater withdrawal

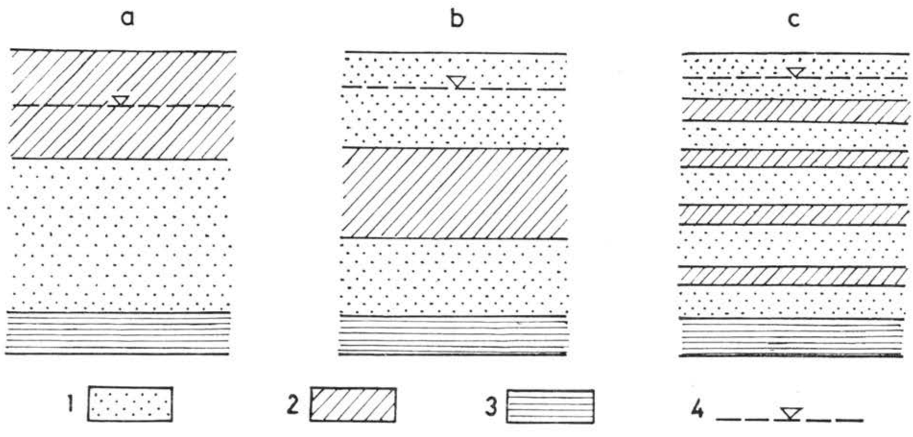

Fig. 1. Hydrodynamical types of the Quaternary water-bearing sediment sequence
a) Two-layered, *b)* three-parted, *c)* multi-layered. *1.* Water-bearing formation, *2.* semi-permeable formation, *3.* impervious formation, *4.* shallow groundwater level

is taking place in Hungary. The Quaternary sediment sequence around Kecskemét has also similar features.

c) Pleistocene sedimentary sequence consists of monotonously-alternating sandy aquifers and semi-permeable clayey layers and lenses. Within this *multi-layered type,* however, does not occur any single, distinct aquifer of good water-yielding properties since water-bearing layers more or less in even distribution can be found within the series as thick as many hundred meters. Consequently, the depth intervals of the screened sections of water producing wells are extremely varied and the water production and drawndown show a three-dimensional distribution. Such types of sediment sequences occur in the central part of the Great Hungarian Plain, along the lower reach of the Tisza-valley. In Szeged, for example, water producing wells completed within the depth interval from 200 to 550 m are supplying the town's water demand.

This study deals only with the mathematical modelling of the wells of waterworks which discharge from aquifers of the two-layered or three-parted Pleistocene deposits. The elaboration of the computer program describing three-dimensional nonsteady-state flow developing within the multi-layered aquifer systems will be carried out in the next future.

Regional depression (drawdown surface) of the groundwater production wells of waterworks

Regional depression, s, means a drawdown surface of the two-dimensional discharge of q [m/day] intensity distributed within the assumed area F in the surroundings of the waterworks. The intensity, q, of the two-dimensional plane discharge can be estimated by the formula

$$q = Q_{\Sigma} / F \tag{1}$$

where Q_{Σ} is the total discharge of the wells of the waterworks [cu.m/day].

Due to the local interferences of the wells, partial screening of the water-bearing series and to some other losses the value of the drawdown within the individual wells will differ from that of the regional depression. These differences, however, in distance L from the center of the waterworks may be neglected and the regional depression equals to the real drawdown. After theoretical studies it can be stated that the value L corresponds approximately to the maximum diameter of the group of wells. Previous studies [8, 9, 10, 11] dealt with the estimation of the local leakage losses and therefore its review will be here omitted.

Determination of the regional depression by electric computer

The regional drawdown related to the two-dimensional plane discharge is independent from the number of wells of waterworks, their location, discharge, depth of screened intervals, etc., and it will depend only on the regional location of the waterworks, on its total groundwater production, on the regional variation of flow parameters as well as on the boundary conditions of the sedimentary sequence.

Within the systems described in Chapter 1a and 1b the spatial and time-dependent variation of the regional drawdown is given by the partial differential equation:

$$\mu(x,\ y)\frac{\partial s}{\partial t}=\operatorname{div}[T(x,y)\ \operatorname{grad} s]-b(x,\ y)s+q(x,\ y,\ t) \qquad (2)$$

where

- $\mu(x,\ y)$ is the storage coefficient by gravity drainage (Type 1a) or by elasticity (Type 1b) of the sediment sequence (dimensionless).
- $T(x,\ y)$ transmissivity of the aquifer system [sq.m/day]
- $b(x,\ y)$ leakage coefficient [day^{-1}]
- $q(x,\ y,\ t)$ intensity of the two-dimensional plane discharge [m/day]
- $x,\ y$ coordinates [m]
- t time [day]

The initial condition is zero for the differential equation *(2)*, $s(x,\ y,\ 0)=0$, the boundary condition can be given by equation *(3)*

$$s(\Gamma,t)\,\varrho_\Gamma+T_\Gamma\frac{\partial s(\Gamma,t)}{\partial \bar{n}_\Gamma}=0 \qquad (3)$$

where

- $s(\Gamma,\ t)$ the value of regional drawdown along the boundary Γ of the flow field, [m]
- ϱ_Γ function of the hydraulic conductivity along the boundary [m/day]
- T_Γ transmissivity of the aquifer along the boundary [sq.m/day]
- $\bar{n}_\Gamma$ the united exterior normal vector of the boundary Γ

The value of the hydraulic conductivity, ϱ_Γ along the boundary will define the features and rate of recharge from the boundary. If $\varrho_\Gamma=0$, an impermeable barrier is present, if $\varrho_\Gamma^{-1}=0$, then the groundwater is unconfined in the recharge area (recharged directly from stream, lake etc.) whereas the relation $0<\varrho_\Gamma<\infty$ will characterize border water courses of differently plugged beds.

For the numerical integration of the differential equation *(2)* the method of differences is applied. The difference equation proposed by E. C. Du Fort and S. P. Frankel [2] was used for the estimation of the approximating value of the regional depression, s, by the following difference equation [9, 11]:

$$\begin{aligned} s_{i,j,p+1}&=\omega_{i,j}[\omega_{i,j}+\Delta t(\beta_{i,j}+\gamma_{i,j}+\beta_{i,j-1}+\gamma_{i-1,j}+2\lambda_{i,j})]^{-1}\cdot\\ &\cdot\{s_{i,j,p-1}+2\Delta t\omega_{i,j}^{-1}[s_{i,j+1,p}\beta_{i,j}+s_{i+1,j,p}\gamma_{i,j}+s_{i,j-1,p}\beta_{i,j-1}+\\ &+s_{i-1,j,p}\gamma_{i-1,j}-0.5(\beta_{i,j}+\gamma_{i,j}+\beta_{i,j-1}+\gamma_{i-1,j})s_{i,j,p-1}+\alpha_{i,j,p+1}]\} \end{aligned} \qquad (4)$$

- $s_{i,j,p+1}$ An approximative value of the regional drawdown at the node of $i,\ j$ coordinates within the flow-net superimposed by the grid at time, t_{p+1} [m]
- $\alpha_{i,j,p+1}$ The discharge of the difference element related to the node i, j [cu.m/day] at time, t_{p+1}
- $\beta_{i,j}$ Hydraulic conductivity of the discrete grid element between nodes of i, j and $i, j+1$ coordinates [sq.m/day]

$\gamma_{i,j}$ Hydraulic conductivity of the grid element between nodes i, j and $i+1, j$ [sq.m/day]

$\lambda_{i,j}$ Areal integral of the function $b(x, y)$ within the difference element related to the node i, j [sq.m/day]

$\omega_{i,j}$ Areal integral of the storage coefficient (x, y) within the difference element of coordinate i, j [sq.m]

$\Delta t = t_{p+1} - t_p = t_p - t_{p-1}$ – time interval [day]

On the basis of the difference equation *(3)* a computer program was elaborated in the Research Institute for Water Resources Development, when a so-called KONV program was also prepared in FORTRAN language. Using this program, two examples are shown here.

Determination of the regional drawdown of wells for waterworks in the surroundings of Kalocsa

The town of Kalocsa is situated on the left-side alluvial floodplain of the Danube. In this area Quaternary deposits represent one of the most important basis of water supply. The alluvium is two-layered (Type 1a). The lower part includes a 30 to 60 m thick gravelous-sandy aquifer overlied by a 8 to 10 m thick cover. The transmissivity, T, of the lower coarse-grained sequence was determined from pumping-tests and from granulometric data [9, 11]. The areal variation of function $T(x, y)$ is shown in Fig. 2 as a contour map.

The most important river of this area is the Danube which may be considered as having a bed hydraulically perfect, that is, $\varrho_r^{-1} = 0$. Within this area a rather dense irrigation and drainage network is available (Fig. 3). The bed of the drainages cutting the cover sediment does not reach the top layer of the lower aquifer, but these drainage channels are hydraulically interconnected with the gravelous-sandy water-yielding formation through the cover sediment. The values of function $b(x, y)$ characterizing the vertical leakage deriving from the channels were determined from the water level data of the phreatic water observation well arrays located perpendicular to the channels as well as from the water-stage data of the channels [9, 11]. On the ground of measurements for the function $b(x, y)$ related to the unit surface area of channels, the following formula can be obtained.

$$b(x, y) = T(x, y)/500B \qquad (5)$$

where B is the width of the channel [m].

The vertical leakage deriving from the atmosphere can not be taken into account due to the lack of measurements and thus beyond the channel beds, $b(x, y) = 0$.

In the planning of waterworks a total discharge of 10,000 cu.m per day was taken into account divided equally between the two water-yielding blocks shown in Fig. 3. The nodes applied in the computations are spaced 1 km in E–W direction and 5 kms in N–S direction (the spacing of 5 kms was justified by other regional estimations made in the study area). Fig. 3 shows the static water level contour map of the regional drawdown. The slight maximum value of 0.85 for the regional drawdown supports the availability of groundwater for favourable water supply in this area.

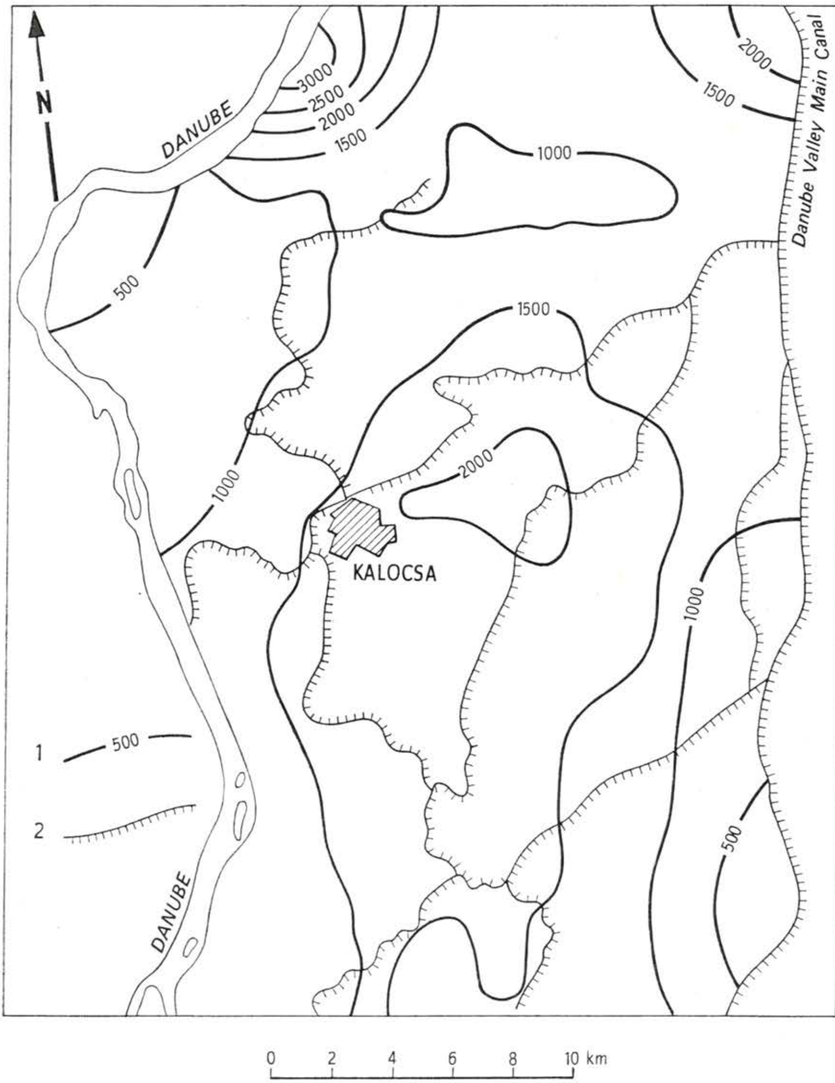

Fig. 2. Transmissivities of the Pleistocene water-yielding formation in the surroundings of Kalocsa

1. Isotransmissivity contour sq.m/day, *2.* drainage network with incomplete beds

Mathematical model for the cone of depression of the Waterworks in Debrecen

The water demand of Debrecen is supplied predominantly by the wells located on the 30 to 50 m thick Lower Pleistocene gravelous and sandy aquifer (so-called aquifer system of the Waterworks). The depth of this aquifer system varies from 110 to 180 m overlied by a semi-permeable clayey deposit of regional extent (transmitting layer) and its uppermost fine-grained sand bed is a water-bearing formation of water-table condition (recharging layer). The typical three-parted Pleistocene sediment sequence of an average thickness of 150 to 200 m (Type b) is fastly wedging out westwards and it is only 30 to 50 m thick within the elevated structure of the Macs-area [3]. Within this area the lower coarse-grained aquifer is already wedged out and thus the isopach line of 50 m interval, as hydrogeological barrier ($\varrho_\Gamma = 0$) can be regarded (Fig. 4).

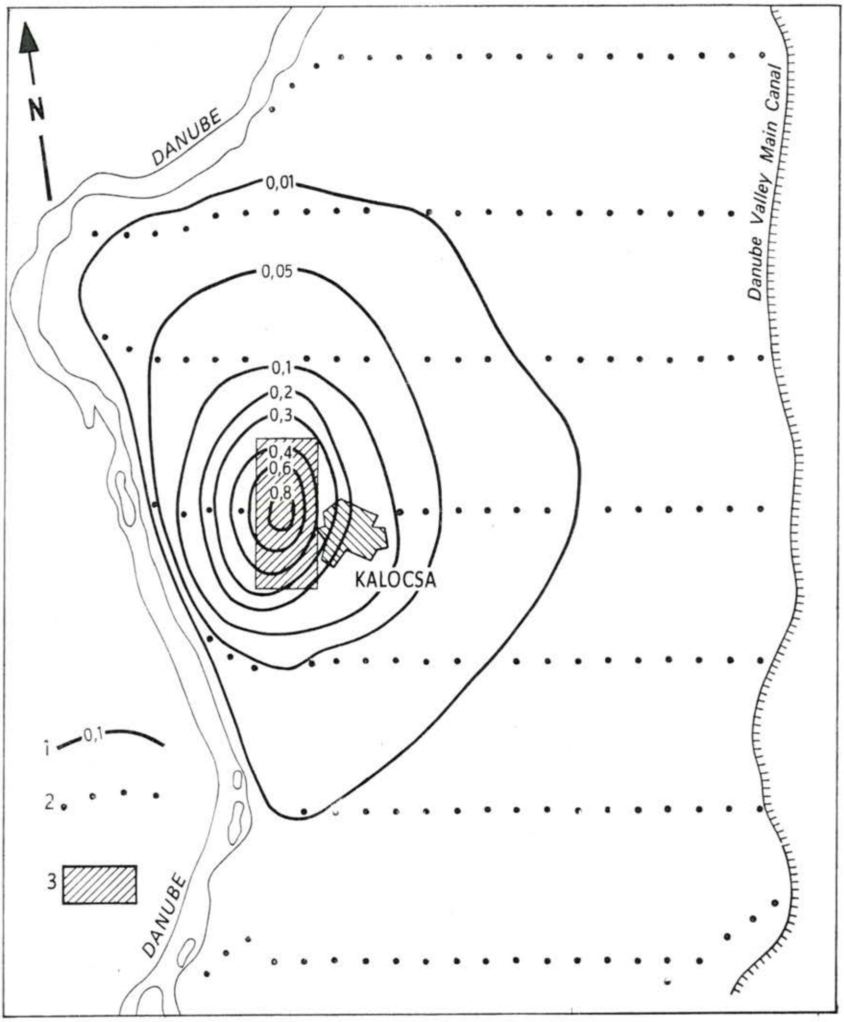

Fig. 3. Steady-state drawdown surface of the waterworks in town Kalocsa
1. Contours of the regional drawdown surface, *2.* nodes of the mathematical model, *3.* area of waterworks

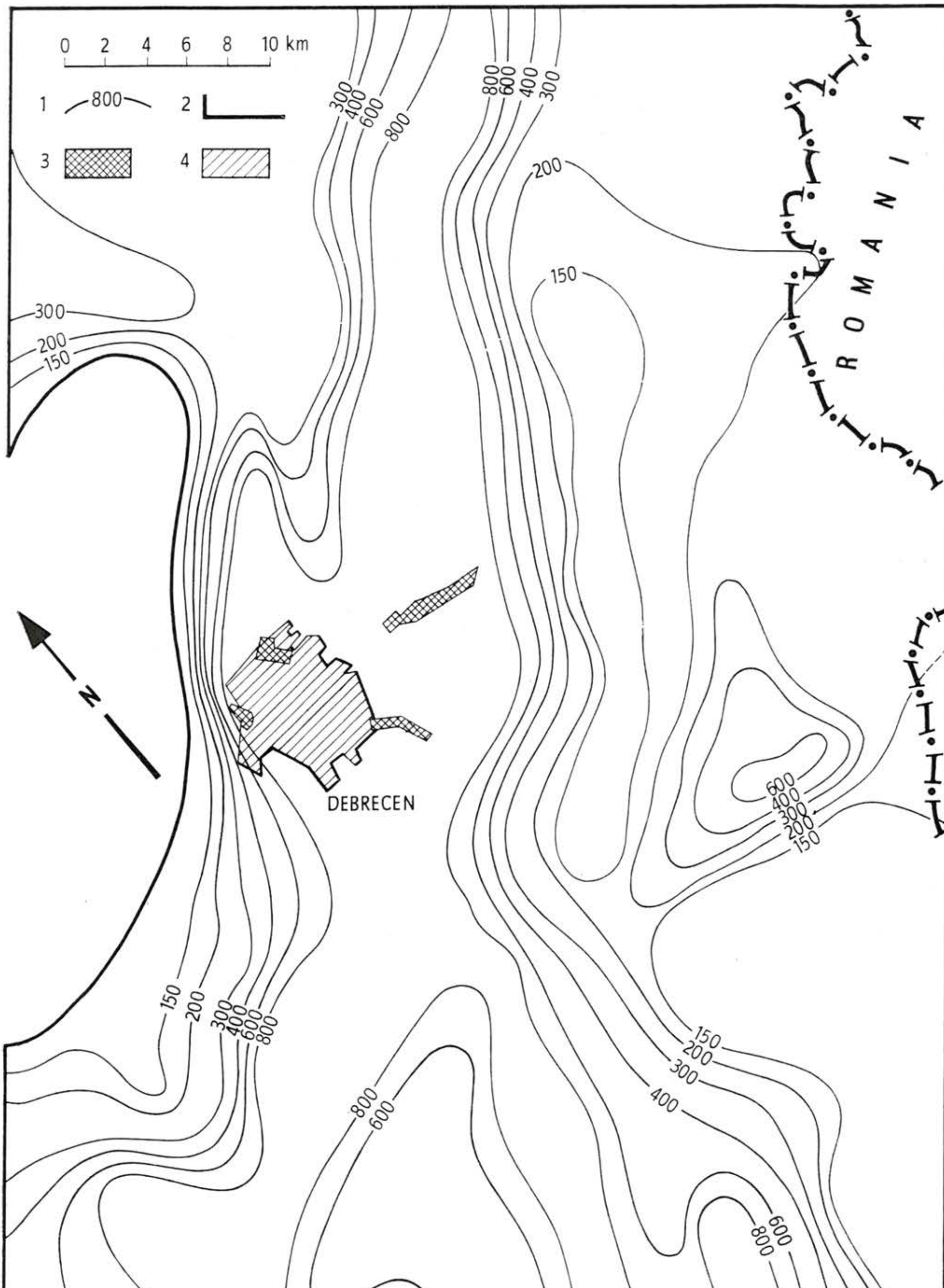

Fig. 4. Transmissivities of the Pleistocene water-yielding formations in the surroundings of Debrecen

1. Isotransmissivity contours sq.m/day, *2.* barrier, *3.* waterworks with groups of well, *4.* urban area where other wells occur

The aim of the mathematical model was to promote the determination of the optimum flow (seepage) parameters of the deposits by simulation of the drawdown surface measured in the field. In the course of the simulation depression was defined from the parameters of the deposits. These parameters were characterized by approximative values of a given distribution and it was compared with the real values of the drawdown surface. The deviation of both depressions was reduced to a minimum value by linear transformation of the parameters representing a given areal distribution.

The first alternative of the function of transmissivity, $T(x,y)$ was plotted using the geologic records and specific yield data of water exploratory drillings located in the study area as well as values, T computed from the recovery curves. This version showed a T value of 1200 sq.m/day within a trough of NE–SW trend crossing Debrecen along with high transmissivities.

As to the vertical leakage coefficient $b(x, y)$, it was assumed that it is a value constant, $b=10^{-5}\,\text{day}^{-1}$.

The storage coefficient, μ, of the elastic confined aquifer was estimated from the data of land subsidence due to the groundwater withdrawal. Comparing land subsidence determined by geodetic surveys [1, 6, 7] with the drawdown surfaces simultaneously observed, a value of $\mu \approx 2\cdot 10^{-3}$ was obtained.

In the model-study the drawdown surface referring to the end of 1974 was simulated. A highly concentrated groundwater production is taking place in the groups of wells Nos. I, II, III and IV of the waterworks, however, considerable quantity of groundwater is produced by other single wells scattered within the entire urban area (Fig. 5). The total discharge of the waterworks was amounted in 1973 as 14,300, 15,400, 5900 and 14,800 cu.m/day, respectively whereas other wells yielded a total of 9500 cu.m/day. In 1974 the total groundwater production amounted 16,300, 16,500, 5700, 15,300 and 9600 cu.m/day, respectively.

The spacing of the model-grid was 1 km at the central part of the depression cone which was further increased to 2 and 4 kms, respectively (Fig. 5). Computing with the afore-mentioned parameters, the calculated drawdown surface was less by 40 per cent on the average than the real values determined by P. Liebe [5]. By the analytical solution of Hantush–Jacob [4] referring to the homogeneous leaky layers such a rate of decrease for the parameters T and b was defined by which the best agreement was reached between the computed and measured values of the drawdown surface. Consequently, functions of $T(x, y)$ and $b(x, y)$ were multiplied by 0.7 and 0.6, respectively and the drawdown surface was defined by computer which showed now a good agreement with the field data. Fig. 4 shows the contours of the final function $T(x, y)$ while in Fig. 5 the calculated drawdown surface can be seen. The model indicates that within this area the drawdown surface referring to a given discharge of the waterworks may be stabilized in 2 years. Therefore, in determining of the drawdown surface related to the end of 1974 only the groundwater production data for the years 1973 and 1974 are required and the modelling of earlier data of water withdrawal might be neglected.

The model of optimum parameters can be applied for the prediction of the drawdown surface within the existing waterworks as a function of the increased discharge and in order to design new waterworks which will be later realized.

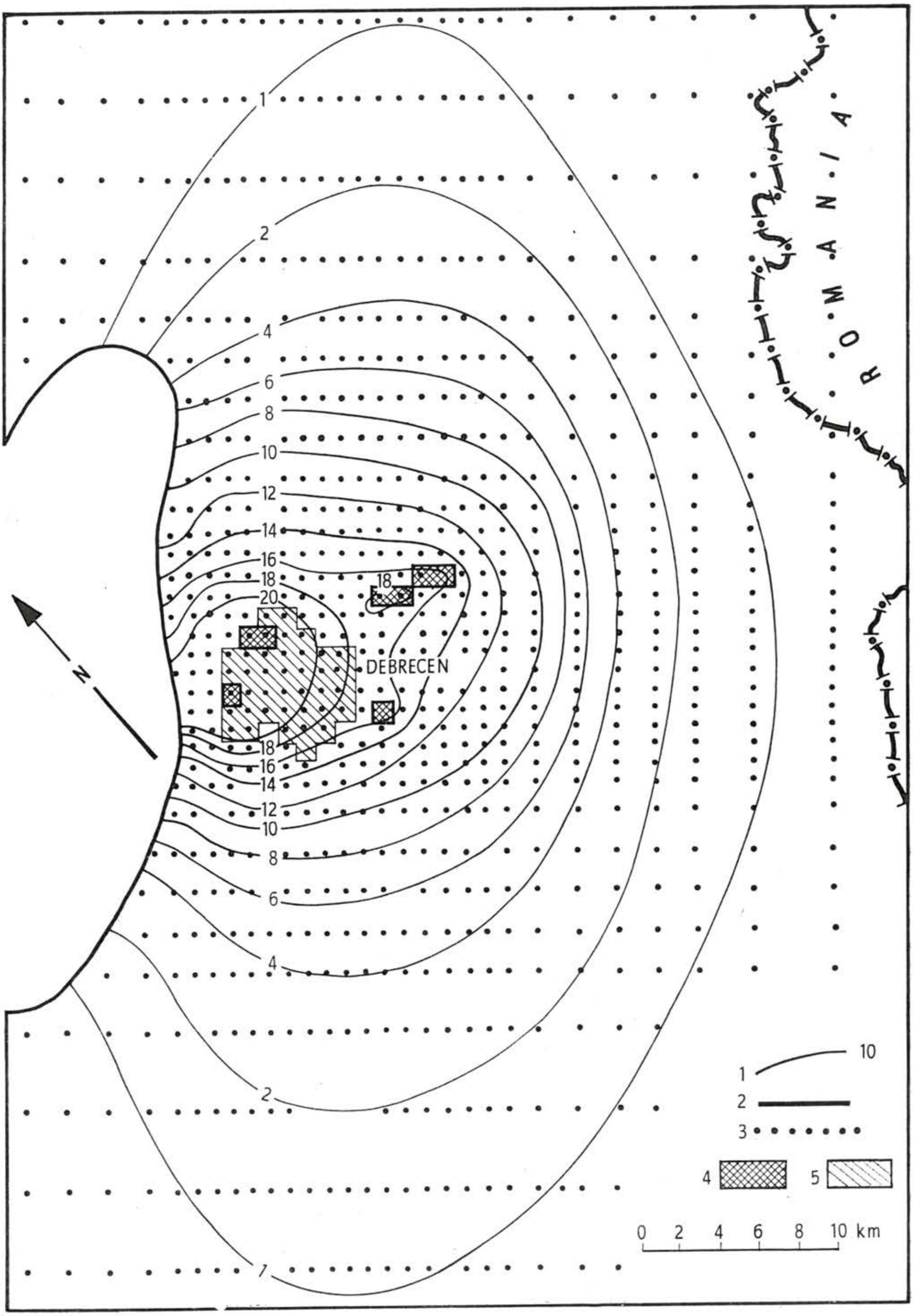

Fig. 5. Computed regional drawdown surface of the waterworks in Debrecen at the end of the year 1974

1. Contours of the regional drawdown surface m, *2.* barrier, *3.* nodes of the mathematical model, *4.* waterworks with groups of wells, *5.* urban area where other wells occur

Conclusion

1. The Quaternary water-bearing sediments of the Great Hungarian Plain can be divided into two-layered, three-parted and multilayered types. This paper deals only with the first and second type.

2. Within the two-layered aquifers of water-table condition as well as within the three-parted confined water-bearing formations the distribution of the regional drawdown influenced by the water production can be characterized by the heterogeneous "leaky aquifer" model.

3. For the numerical analysis of the nonsteady-state processes occurring within the heterogeneous "leaky aquifer" a computer program based on the application of difference method was elaborated.

4. In the surroundings of the town Kalocsa a regional drawdown of the Waterworks having a capacity of $Q = 10{,}000$ cu.m/day was determined by computer method where a two-layered aquifer of water-table condition accompanied by given flow parameters is occurring. The vertical leakage of the discharging formation is deriving here from the drainage network having no complete beds.

5. In the surroundings of the city of Debrecen hydrogeological parameters were determined from known water production and drawdown data in the case of three-parted confined aquifer. The discharging Lower Pleistocene sand formations get here a vertical leakage deriving from the phreatic water bodies. The problem was solved by simulation through the minimization of the difference between the measured and computed drawdowns.

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FORMATION WATER PRESSURE IN THE QUATERNARY SEDIMENTS OF THE GREAT HUNGARIAN PLAIN

PRESSION DES NAPPES AQUIFÈRES ARTÉSIENNES DANS LES SÉDIMENTS QUATERNAIRES DE LA GRANDE PLAINE HONGROISE

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RÉSUMÉ

Ce qu'on entend par pression géohydrostatique c'est la tension due au poids de l'eau artésienne et la pression des couches du toit à laquelle l'eau artésienne est soumise. C'est la pression qui est active dans une nappe aquifère artésienne, mais qui est dans les cas les plus rares égale à la pression hydrostatique.

Le gradient géohydrostatique est la valeur de l'augmentation de la pression et s'il vaut de 0,99 à 1,01 att., on parle d'un état de pression « normal », dans le cas, où il est supérieur à cette valeur-là, il s'agit d'un état « positif », s'il y est inférieur, on parle d'un état de pression « négatif ».

Dans le cas d'un ensemble de grains varié, la pression n'augmente pas d'une manière uniforme en fonction de la profondeur, mais les sédiments d'une granulométrie plus fine présentent un gradient de pression plus élevé. Cela est témoigné par la pression des nappes aquifères artésiennes enregistrée dans la série sédimentaire grossière de la fosse tectonique du Danube d'une part et dans la série à grain fin de la dépression des rivières Körös d'autre part.

L'établissement d'un état de pression « négatif » particulièrement anormal a pu être observé au territoire entre la dorsale de Nyírség et la Plaine de Szatmár. Dans la série sédimentaire à grain fin du Pleistocène supérieur le niveau statique est proche du niveau de la nappe phréatique; le gradient de pression augmente avec la profondeur; par conséquent, un écoulement d'eau descendant est impossible, de sorte que les précipitatison ne peuvent pas s'infiltrer dans les couches plus profondes. Une différence de pression considérable peut être enregistrée entre les deux régions, mais cette différence ne s'équilibre pas à cause de la tension limite existant dans l'ensemble des grains. Par conséquent, c'est une pression statique qui existe dans les sédiments du Pleistocène supérieur. Le Pleistocène inférieur est représenté par des graviers et le niveau statique des eaux qu'ils renferment dans le Nyírség se trouve, d'une façon « anomale », à une profondeur de 30 à 32 m. Cela est dû à ce qu'étant donné la tension limite très basse caractéristique des sédiments grossiers, la différence de pression entre la dorsale et la dépression a été compensée déjà au cours des temps géologiques, de sorte qu'à présent un état statique peut être observé ici aussi.

La pression géohydrostatique n'augmente pas uniformément avec la profondeur, mais elle varie en dépendance de la granulométrie du sédiment. La pression n'a pas été équilibrée même à l'inférieur d'une aire restreinte, ce qui indique également la présence d'une tension statique.

Due to the need of increasing water supplies, communal, industrial and agricultural reasonable management of water resources is becoming a more and more urgent task to solve. This requires, however, the knowledge of the subsurface water resources and the dynamics of subsurface movements.

Before discussing the results of measurements and summing up the conclusions deduceable therefrom, let us give now the interpretations of some notions that will be referred to hereinafter.

The weight of a water head without the rock pressure is called the *hydrostatic pressure.*

Geostatic pressure is due to the own weight of the rock. In a coarse-grained sequence the weight of the overlying beds is carried by the grain set itself, in a fine-grained stratum the weight of the rock is partly taken over by the water contained in it.

Geohydrostatic pressure is interpreted as a stress generated by the weight of the subsurface water plus the pressure of the fine-grained rock upon the formation water. This is the pressure that prevails in any deep-situated aquifer, agreeing in rarest cases with the hydrostatic pressure. The geohydrostatic pressure is indicated by the piezometric water level observed in a bore-well (the well acts as a manometer) and its value is defined by the difference in altitude between the level of water entry into the well on one hand and the piezometric level on the other.

Since only the pressure conditions of Quaternary sediments are to be examined, the temperature and gas regimes, also influencing the pressure prevailing in the strata, have been ignored, as the role of these, compared to that of the geohydrostatic pressure, is unessential.

The *geohydrostatic pressure gradient* represents the size of pressure increase which is the ratio of the height of the water head measured in the well to the column of water flowing into the well. The terms *"normal"*, *"positive" and "negative" pressure state*, which have already found general use in the literature and which refer essentially to the pressure gradient, are products of an artificial categorization and their usage may be omitted. Accordingly, a gradient in the range of 0.99 to 1.01 would be normal; one lower than that would be negative and one higher than that would be positive.

The occurrence of a *hydrostatic pressure state* in the formation-water-bearing space is exceptional, being possible only in a set of unstratified, homogeneous, coarse grains, where the weight of the rock is carried only by sediment grains. This state is approximated by the gravel and coarse-grained sandstone sequence, almost 300 m thick, deposited at Ásványráró in the Little Hungarian Plain, where during the testing of one well at five different depths the static water head settled at almost one and the same altitude, though not hydrostatically even there, but at 0.1 to 0.4 m above the topographic level (Fig. 1). (The well-logging and geological profiles of the nearby water bore-well of Lipót presented herewith are to confirm the admitted presence of a coarse-grained sequence, by instrumental measurements.)

Differing from the afore-mentioned "normal" pressure conditions, the geohydrostatic pressure is brought about in a stratified sequence of different grain size, where, independently of the degree of consolidation, both the grain set and the water will take over the weight of the overburden. And, as a natural consequence of this, the less consolidated rock (clay, loamy clay, loamy fine sand) is, the higher the load applied to the interstitial water and the higher the pressure accumulated in the rock. In the opposite case, in turn, it is the pressure that is lower. In the Danube's graben (e.g. at Szeged, South Hungary), where the share of sands is 40 to 80%, the geohydrostatic pressure is lower than in the depression of the Körös river system (e.g. at Békéscsaba), where

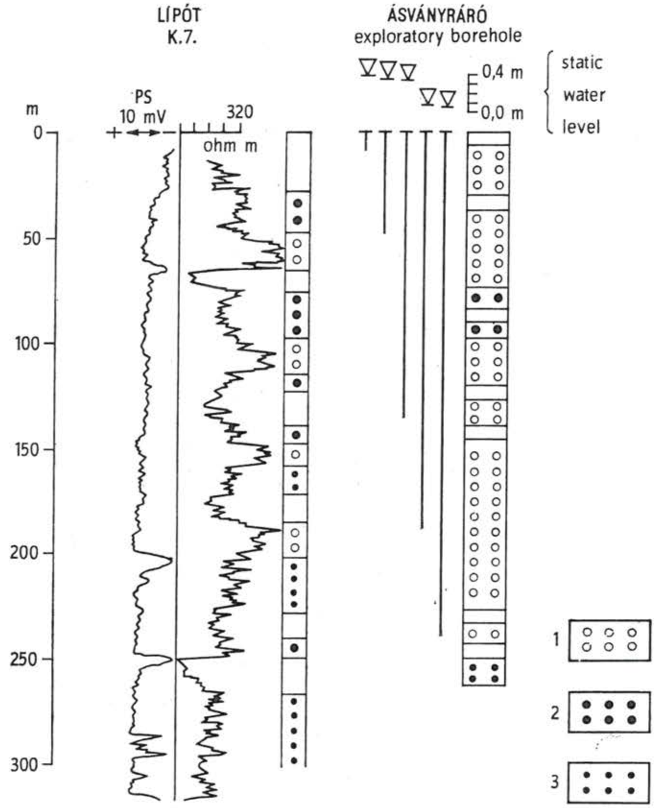

Fig. 1. Geohydrostatic pressure state of the formation water of a gravel sequence

1. Gravel, gravelly sand, *2.* coarse- and medium-grained sands, *3.* middle-, small- and fine-grained sands (loamy)

the share of porous sediments is as low as 10–15–20% (Fig. 2). The pressure in both areas increases as a function of depth, but the rate of this increase is different. Let us call attention, however, to the fact that the depth-dependent increase of pressure is never uniform. Notably, the rate of increase of the pressure will vary, even in one and the same profile, according to the proportion of the coarser and finer grain fractions, being higher in the finer grain fraction and lower in the coarser one. This can be read off the behaviour of the pressure graph of Fig. 2. Notably, at Békéscsaba, in the more sandy sequence of −150 to −250 m altitude, the pressure increase is lower; at Szeged, in turn, in the −75 to −150 m interval of diversified lithology not affected by any tectonic stress, the pressure increase varies in accordance with the variation of the grain set. As evident from the information thus far available and from their representation, the geohydrostatic pressure will increase with depth and the rate of increase is in a causal relationship with the grain size composition, respectively with the consolidation of the grain set concerned. The fact that the different pressures were not balanced in the course of geological time is indicative of the presence of a static strain.

Let us examine now the problem of the so-called "negative" formation-water pressure which is differently interpreted by various research workers. The divergency of opinions indicates that the problem has not yet been settled.

To account for it, let us start with an analysis of the pressure conditions of the subsurface water-bearing spaces beneath the Nyírség Ridge and the Szatmár Plain and their possible interaction. The geological section of Fig. 3 shows the structural pattern and lithology beneath the Szatmár Plain's ridge and marginal depressions. In addition, the piezometric (isoatt) surfaces have also been indicated. Generally speaking, the geological make-up of the territory is characterized by the fact that the Upper and Middle Pleistocene is made up of medium-, small- and fine-grained sands of loamy character and that the porous strata are interlain by impervious layers which are, again, of different

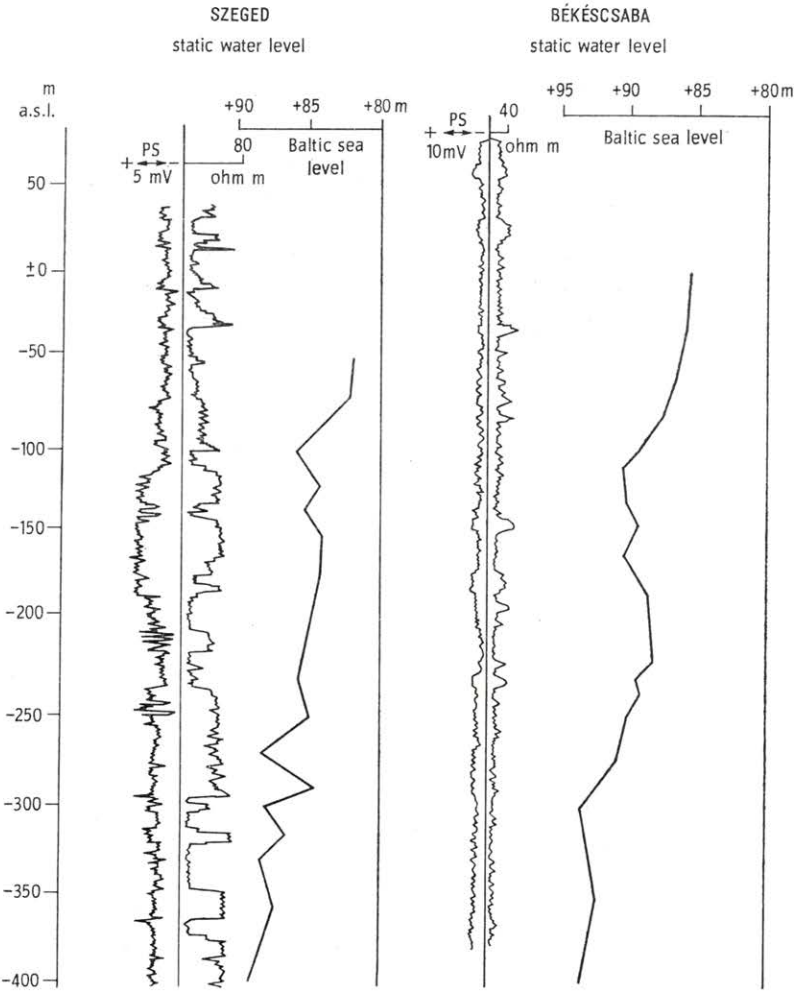

Fig. 2. Formation water pressure of basins of different geological structure

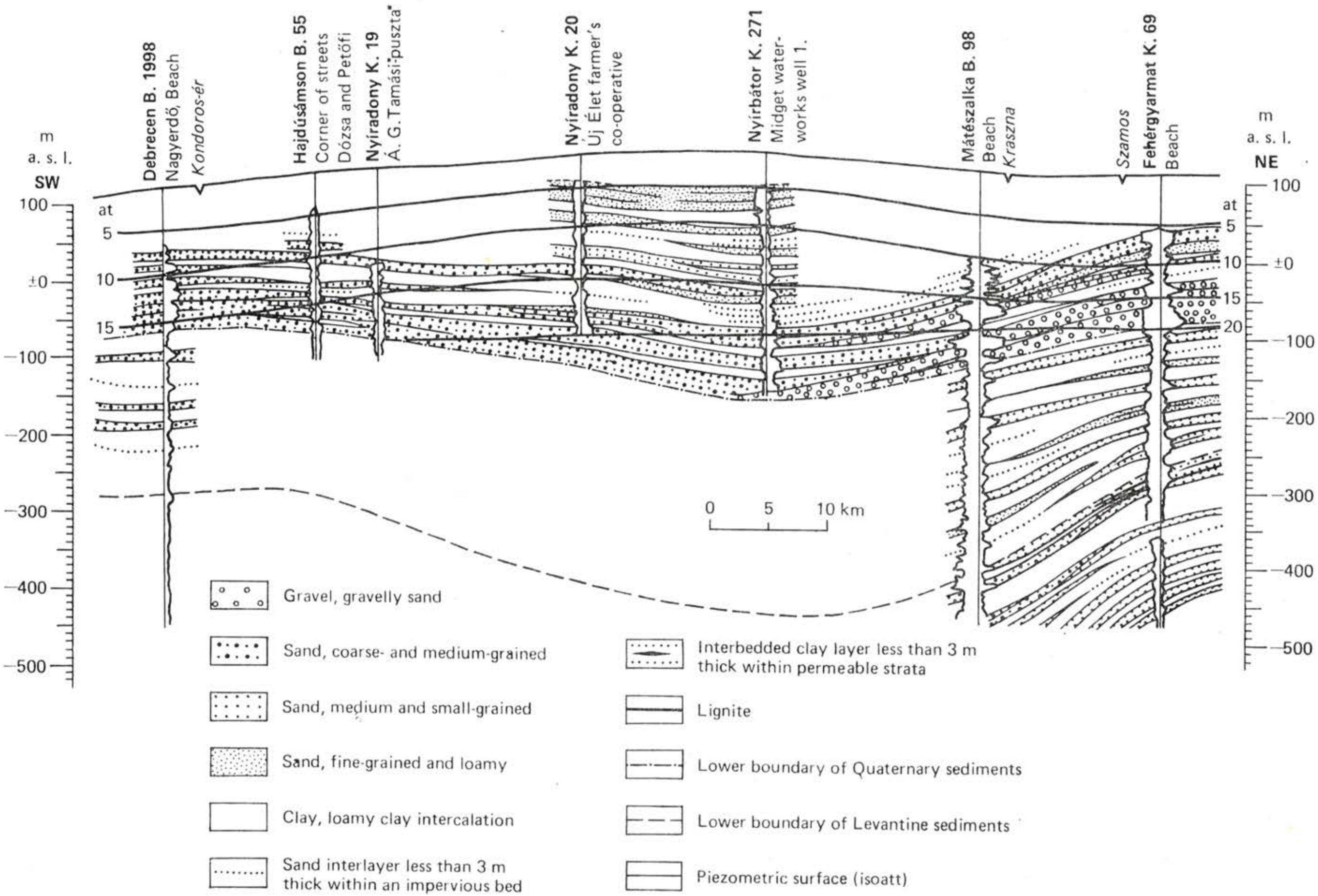

Fig 3. Piezometric surfaces between the Nyírség ridge and the Szatmár plain

facies (clay, loamy clay, fine-sandy loam). Consequently, these grain sets are very diversified in composition, though consisting as a rule of fine-grained sediment from 100 to 150 m thick. The Lower Pleistocene sequence, however, is everywhere of very coarse grain composition (gravel, sandy gravel, gravelly sand, coarse sand), being known throughout the Nyírség and representing the typical sediment of the Szatmár Plain. The base of the sequence at Nyírbátor is at −320 m depth, at Mátészalka −250 m, at Fehérgyarmat −200 m, being followed by hundreds of metres of impermeable Levantine sediment underneath. The 5- to 10-isoatts and, in part, the 15-isoatts within the fine-grained Upper and Middle Pleistocene sequence form a surface reflecting, like in the Great Plain, the relief of the ground surface while the pressure gradient increases with depth, like in any other region. No depthward water flow is thus possible. This provides a remarkable evidence of the fact that the geohydrostatic pressure gradient increases in the Upper Pleistocene sediments throughout the Great Plain and thus no distinction in terms of formation water pressure types ("positive" or "negative") can be made between ridges on one hand and depressions on the other (Fig. 4).

As shown by the geological section, between Nyírbátor and Mátészalka, a decrease in pressure in the Upper and Middle Pleistocene formation waters can be observed: a disequilibrium which could not be compensated even during nearly half a million years. This is quite natural, because the boundary stress characteristic of the fine grain fraction makes such a compensation impossible, as expounded eloquently in I. Balló's paper (1966). Consequently, a static pressure state prevails even laterally, in the Upper and Middle Pleistocene sequence. L. Simon (1968) showed the same static pressure state to occur in the young sediments of the Nyírség. He called it "local pressure" and he even considered "the assumption of the existence of static reserves" to be justified. It would be unproper, however, to imagine the Nyírség playing the role of a recipient of precipitations in which the infiltrated meteoric water would reach great depths underground and then enter an abyssal circulation system with waters flowing towards the interior of the Great Plain and the Szatmár Plain and then returning to the surface there.

A tension apparently different from the former pressure state has developed in the Lower Pleistocene formation waters of the Ridge. At Nyírbátor the static level in the wells drilled to tap the gravelly coarse-sand aquifer of the 160−240 m interval was established at 30 to 32 m. This is an anomalous pressure state which we earlier called the "negative" pressure of formation waters. If the 20-isoatt coinciding with this layer between Nyírbátor and Mátészalka would be scrutinized, it will be found that this surface traverses the Szatmár Plain, the Nyírség Ridge and the Körös Depression almost horizontally, in compliance with the altitude of the footwall of the Pleistocene at Mátészalka. Consequently, this surface, which would connect equal pressure values of the Middle to Upper Pleistocene formation waters, does not reflect the relief of the ground surface in the Nyírség either. An explanation for this is provided by the fact that, because of the low boundary stress, the pressure in the gravelly sediments totally intercommunicating in the space between Mátészalka and Nyírbátor was balanced already during sedimentation or, in the case of the post-sedimentary uplift of the Nyírség, after it. This process took place already in pre-Holocene time, so that at present a hydrostatic state can be observed here, too.

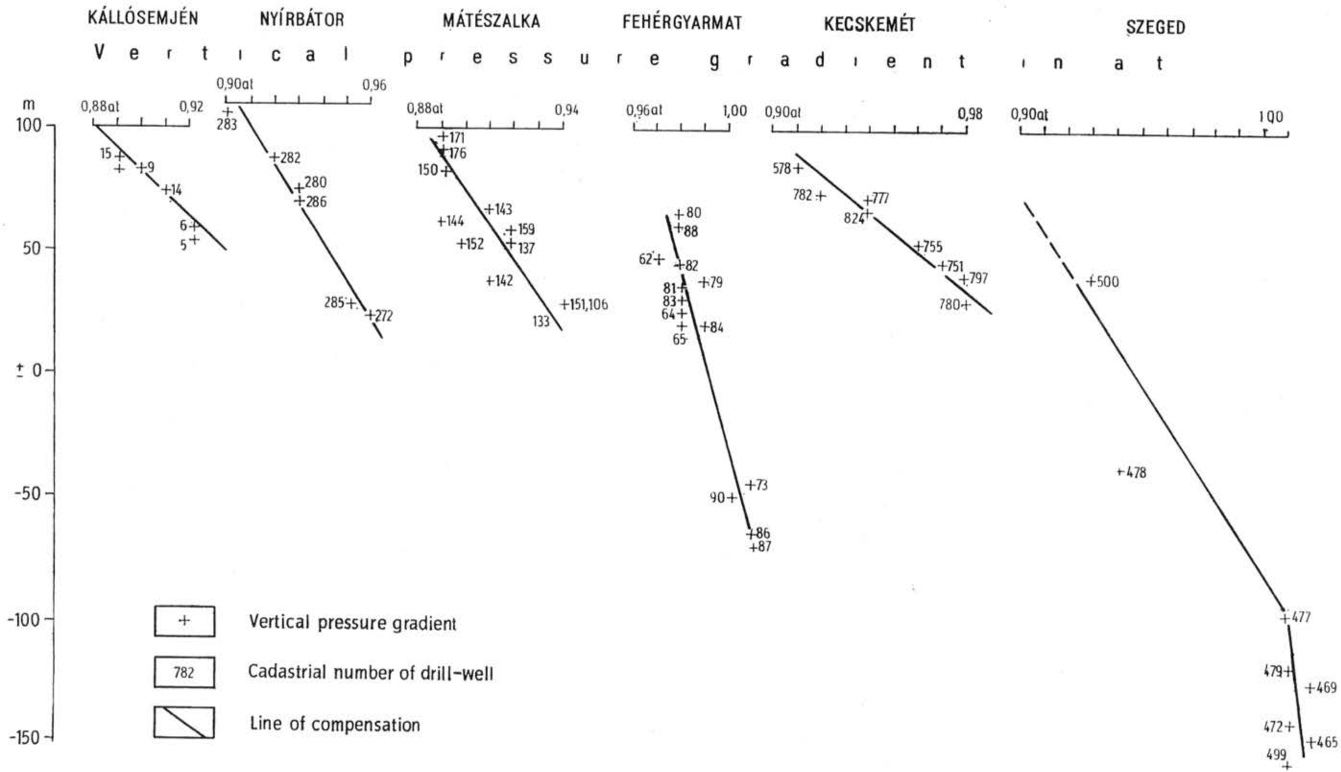

Fig. 4. Vertical geohydrostatic gradient in the Upper Pleistocene beds

Thus the mechanism of the pressures prevailing in the Nyírség's Upper and Lower Pleistocene sediments are controlled by a boundary stress corresponding to the grain structure which, naturally, plays one and the same role both in the Nyírség and throughout the basin interior.

According to L. Simon, "since the coarsening of the sediment with depth is a common rule, it is possible that the water in the deeper-situated (coarser) layers is under lower stress" than it would be the case with a finer grain composition. In his opinion, "This results in a negative pressure state". This would mean that the pressure in the Lower Pleistocene sediments of gravel facies is — even in depression areas — lower than in a finer set of grains. This opinion is contradicted by the pressure conditions existing in the sandy gravel aquifer occurring in the 150 to 350 m interval at Polgár and Tiszacsege on the alluvial fan of the river Sajó and between 350 and 500 m at Berettyóújfalu in the Körös Depression. Notably, at this third locality, a water surging above the surface is being tapped, while at the first two localities the static level is close to the groundwater table. The same situation can be observed at Nyírbátor, in the gravel layer of the 240 to 320 m interval, where, given the presence of a closed subbasin of lower altitude compared to the Szatmár Plain, the available stresses could no longer be balanced. Here the pressure gradient rises again or the pressure conditions are normal, compared to the 20-isoatt surface.

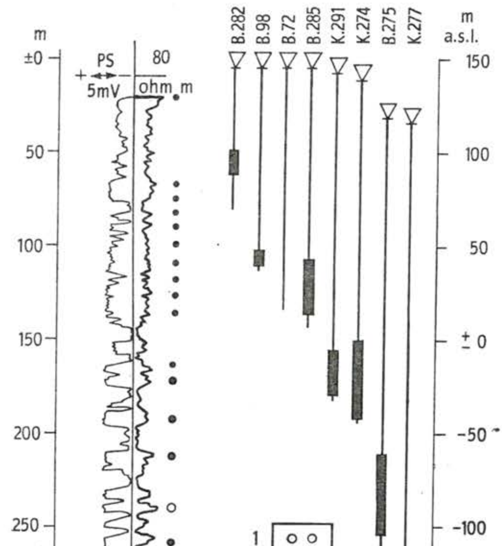

Fig. 5. Formation water pressure of a ridge area: Nyírbátor

1. Gravel, gravelly sand, *2.* coarse- and medium-grained sands, *3.* medium-, small- and fine-grained sands (with loam), *4.* perforation points, *5.* cadastral number of drill-well in the municipal area of a settlement, *6.* cadastral number of a drill-well outside the municipal area of a settlement

The fact that the Lower Pleistocene formation water pressure is lower than the stress corresponding to its geodesic position is due to causes other than the loss of energy by the infiltrating water, as believed by some scientists. The real cause is that the afore-mentioned pressure has been compensated so as to correspond to the pressure level prevailing in the Lower Pleistocene horizon of the adjacent depression.

The establishment of a formation water pressure substantially lower than the geohydrostatic one, i.e. a "negative" one, depends not only on the permeability of the strata, but an initial overpressure, implying a pressure gradient, is also necessary. This, in turn, is possible to occur on a ridge formed either similarly to the case of the Nyírség, as a result of a post-sedimentary upheaval, or like the case of the North

Bácska- and Kiskunság Ridge, where the surroundings have subsided and the low boundary stress of the Lower Pleistocene coarse-grained sequence has enabled the compensation of the pressure gradient provoked by changes in altitude.

The static level of the porous strata tapped at Nyírbátor is shown in Fig. 5.

The pressure conditions of the water-bearing space (aquifer) of a diversified grain set are shown in Fig. 6. The static water level (pressure) increases with depth, but this increase is nowhere steady, which indicates that local disequilibria of pressure have not been compensated and that, consequently, the existing pressure state is a static one. The reason for which we have chosen the vicinity of Mindszent and Szegvár as the site of investigation, is that a considerable number of wells of different depth have been drilled in this area, but any sizeable water recovery that might substantially upset the initial equilibrium has so far been avoided. The two well-logging graphs attached to the figure represent type sections which are to provide an instrumental evidence of the diversified stratification and grain size distribution characteristic of the territory. Down to the altitude of -170 m, the pressure increases at an almost steady rate and, as indicated, among other things, by the graph of resistivity logging, the porous strata have almost identical grain size composition. In the $170-275$ m interval, the rate of pressure increase, the geohydrostatic gradient, diminishes strikingly, to show again an increase at greater depths. In the same interval and underneath, a striking change is recorded by instrumental measurements as well and it is this change that accounts for the change in gradient, i.e. in pressure state. In the given interval the share of the porous sediments as well as the size of the grains is larger and, consequently, the lateral resistance of the aquifer is smaller; in the higher and deeper horizons, in turn, some excess of pressure will appear owing to the consolidation of the finer-grained sediments. Because of the threshold value of the boundary stress, however, the difference in pressure in the sedimentary sequence of diversified lithology cannot be compensated.

The diversified pressure states recorded in the above tectonic and stratigraphic setting indicate, all, that the water contained in the Quaternary sediments of the Great Hungarian Plain is, under natural circumstances, in a static state and that this static state is maintained by the boundary stress corresponding to the permeability of the strata.

After the above considerations there may well be no question of whether water movement in the Quaternary beds of the Great Hungarian Plain is possible at all.

Yes, it is, but only in the course of human intervention and this is exactly, where the extraordinary difficulty of assessing the reserves and forecasting them, is rooted.

In case of water recovery the initial equilibrium would be upset and a new field of stresses will develop in the neighbourhood of places characterized by an anomalous pressure decrease and then increasingly farther away of them. In these places, the change (increase) in pressure will initiate a water flow as soon as the threshold value of the boundary stress is exceeded. The highest rate of flow will take place there, where the boundary stress is the lowest, consequently, on alluvial fans, precisely in their coarse-grained Lower Pleistocene sequence. Knowledge on the spatial position of the alluvial fans may

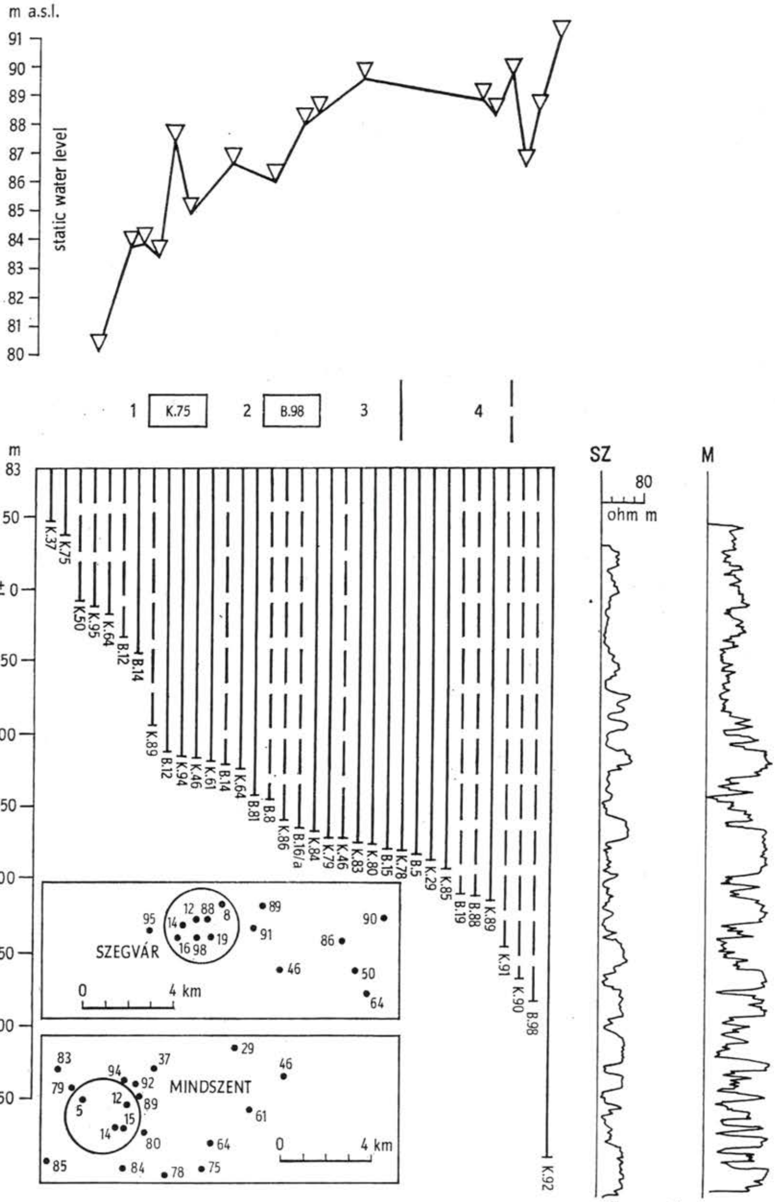

Fig. 6. Geohydrostatic pressure state of a grain set of diversified lithology

1. Number of well outside the municipal area of a settlement, *2.* number of well within the municipal area of a settlement, *3.* Mindszent, *4.* Szegvár lead to important paleogeographic conclusions which, in turn, may give information not only on the paths of sediment transportation, but on the channelways of water flow and recharge as well.

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Discussion

M. Besbes, comme premier, illustre quelques documents de leurs rapport par la projection de diapositives.

J. Urbancsek answers to Mr. Gy. Kovács: "Some scientists believe, erroneously, that the subsurface water reserves in the Nyírség are recharged by atmospheric precipitations. Notably, in absence of a horizontal water flow, a vertical one cannot take place either. Another unacceptable opinion, which would advocate the presence of a dynamic state, suggests that the static pressure here is just an apparent one, maintained exactly by a continuous recharge from meteoric waters. This is contradicted by the fact that the vertical geohydrostatic pressure gradient in the Nyírség increases up to 160 m in the same way as it does in the other large sub-basins, whereas, in case of a dynamic state, it ought to decrease. For, the negative pressure condition known in the Lower Pleistocene is not the consequence of a pressure decrease due to a loss on percolation of the infiltrating water, but it stems from an overall compensation of pressure in the Lower Pleistocene sequence throughout the Great Hungarian Plain.

Consequently, that large-scale circulation of subsurface waters which was supposed to result from infiltration into the sand ridges (Nyírség, Danube-Tisza Interfluve) and from the water flow ascending from the deep basins, is impossible. The subsurface waters can be recharged primarly through the surfacial, gravelly and coarse-sandy sediments of the marginal detrital zones, where a water flow may be generated owing to the low boundary stress.

Considering that the formation waters in the Upper Pleistocene sandy (and everywhere silty) sediments are in a static condition, a mobilization of the waters confined by clays and characterized by a higher boundary stress, is inconceivable. Consequently, these waters cannot be regarded as possible sources of recharge."

Gy. Kovács: "A few words about the role of the threshold gradient in the development of pressure difference. It is well known that in very fine-grained sediments there exists a limit value of the pressure gradient, below which there is no water movement through the strata. Recent investigations by Mr. Bondarenko and others in Leningrad have shown that the threshold gradient largely decreases if the temperature increases. Somewhere between 50° and 60° centigrade the value of the gradient becomes practically 0."

J. Deák insists: "At 300 m depth we have got a layer filled with water 10,000 years old and nearer to the surface, at 80–100 m depth, the age of the water varied between 8000–20,000 years. The source of supply cannot be very far away. The upper strata are replenished from the deeper ones.

At the foot of the Mátra Mountains it was possible to demonstrate that the Upper Pannonian layers receive a continuous supply and the water flowing at about 15–20 km from the recharge area is at least 15,000 years old. It means that, in my opinion, the circulation system which Mr. Urbancsek has mentioned is inconceivable. The waters are too young to receive a recharge from such a great distance."

J. Tóth: "I should like to ask just one question of Dr. Urbancsek. Has any fluctuation in water levels been observed in these deep wells or not. And if so, how can he explain this by the static theory? Another comment. I guess from a former paper by Dr. Urbancsek, that the areas of negative pressure condition coincide consistently and regularly with the topographical highs while positive pressure heads correspond to topographically low-situated areas. These conditions fit very well the general dynamic picture of ground-water flow. I did not find in his papers any proof for his attempt at correlating the positive and negative pressure conditions with grain size."

M. Mucsi gives detailed informations on the paleogeographical and sedimentological studies made on core samples by a team at Szeged.

"The area investigated by us lies in the southern part of Hungary. In this part of the Neogene geosyncline of the Great Plain varied rock formations and facies were observed. Together with the sequence of 100 to 250 metres thickness which can be assigned to the Quaternary, the total thickness of the Neogene varies between 1000 and 6000 metres; over a smaller part of the area values greater than 6000 metres can also be assumed."

I. Balló explains the model dealt with in his paper, which is a conservative and empirical one. It deals with the electric analogy of the stress pattern of subsurface waters. He analyzes the short comings and confusions of the terminologies used by geologists and engineers, respectively. He expresses his doubts as to the reliability of the radioactive watches, because the background radioactivity is not well known.

J. Deák answering to Mr. I. Balló defends his radioactive results. "We may face problems in evaluation, indeed. There are several methods. But the possible deviation is 2000–3000 years according to the method."

L. Szebényi emphasizes that the piezometric level of the subsurface waters changes fairly in time. "In a 30-year period the amplitude of these changes can attain 4 m. So, the present levels of the Pannonian Basin cannot show retrospectively the situation of a million years ago, not even that of a few centuries ago, unlike Mr. Balló suggested it."

F. Székely: "In his contribution Mr. Urbancsek has pointed out that according to his point of view there is no water movement in the fine-grained Quaternary deposits of the Great Hungarian Plain and the water withdrawn from the coarse-grained, sandy-gravelly layers is recharged solely on the margins. At the Research Institute for Water Resources Development we have investigated the recharge of some additional well-fields of important municipal waterworks. We have set down that the depression cone does not extend to the edge of the Great Hungarian Plain but it can be usually traced for a distance of 10–20 km.

This points to the fact that recharge of ground-water exploited from deeper aquifers takes place locally from shallow ground-water by vertical seepage through the overlying fine-grained, loamy clayey, lenticular semipermeable layers. In the influence area of the Debrecen Waterworks we have determined, by mathematical modelling, the average areal vertical seepage factor of the Middle Pleistocene semipermeable layer as corresponding to $4 \cdot 10^{-4}$ m/day. Although this value is very low but it supports the vertical recharge of the exploited confined water, owing to the great area of the depression cone."

J. EINSELE: "We have got two processes which may influence the transport of the dissolved constituents. One process is pore-water flow by compaction or consolidation, the second is molecular diffusion. There is a controversy about the influence of these two mechanisms. Most people think nowadays that as long as the sedimentation rate is very low, molecular diffusion is the more important factor. But as soon as sedimentation rate becomes higher, the compaction flow may become more important. We have a very simple model for studying the problem, I will try to demonstrate things better with a few slides."

Now the slides are projected.

M. R. LLAMAS demonstrates some slides on the Tertiary continental basin of Spain. The maximum thickness of the Tertiary sediments is 3000 m. They represent a multi-aquifer system. The speaker deals with the different interpretations of the transmissibility and exposes the underground flow-lines in the concerned territory. The slides gave informations on the digital model used for the investigation of three aquifers.

A. SARIN explains the meaning of the "transmissivity index", as it is used in Croatia. He speaks of resistivity surveying and how one can substitute the resistivity curves by numerical values to obtain a "mappable" material. He deals with the difficulties of interpreting the resistivity data and the thickness of the layers.

GY. KOVÁCS: "I had the possibility to study MR. SARIN's and BABIC's paper in detail. I have found it very impressive and I am quite sure that the geoelectric method to estimate the transmissivity is very useful for comparative studies. I should like to raise only one problem. Where the basement was near the surface there the transmissivity index was calculated from the surface down to the basement. If the depth of the basin was greater than 150 m, there an artificial limit was applied, which—in my opinion—practically disturbs the calculated value. I should like to ask, whether any investigation was done or not to check the error caused by neglecting greater depths."

A. SARIN: "Deep resistivity explorations in Croatia, ranging between 200 and 500 m, showed the accuracy of interpretation of the thickness of the strata within $\pm 10\%$, but only under favourable conditions: thick layers of homogeneous resistivity."

J. TÓTH: "I should like to point out two questions concerning MR. SARIN's paper. It seems to be a bit dangerous to apply a hydrogeological term, 'transmissivity', to an electrical parameter. Resistivity multiplied by the thickness of a stratum is an electric parameter. If you fill that stratum with sodium chlorite, it will have a very low resistivity, and still has nothing to do with actual physical transmissivity. My second point is that the transmissivity is directly comparable with conductivity times the thickness of the stratum, rather than with resistivity times the thickness of the stratum."

A. SARIN: "The geoelectric term, ϱh, under special conditions is in an approximate exponential relationship with aquifer transmissivity. I think it may be called the 'transmissivity index'. The filling of a stratum with sodium-chlorite will change the basic conditions and the discussed transmissivity index concept cannot be applied."

G. DASSONVILLE: « Mon intervention a pour but de susciter une discussion avec M. SZÉKELY, à propos de l'optimisation des valeurs de l'emmagasinement et des transferts dans une multicouche, dans la restitution de l'historique d'une dépression causée par des pompages.

Le « calage » d'un modèle mathématique de nappe s'effectue en tenant compte des « entrées » et des « sorties » d'eau dans le système aquifère étudié et consiste à ajuster, par tâtonnements, les transmissivités et, le cas échéant, les flux aux limites, de manière à retrouver par le calcul, les niveaux piézométriques (ou potentiels) mesurés sur le terrain. Dès lors, tous les scénarios que l'on veut simuler, en régime permanent et en régime transitoire, quantitativement, mais aussi du point de vue de la qualité des eaux, pourront l'être au moyen de l'ordinateur.

Il n'y a donc aucun démon à exorciser!

Un modèle de nappe a pour seul objet de permettre un test de cohérence entre les valeurs des différents termes du bilan et sur les hypothèses faites à propos des inter-actions entre cours d'eau et nappes ou nappes entre elles dans le cas de systèmes multicouches. »

Theme 2

REGIONAL HYDROGEOLOGICAL ELABORATIONS, GENERAL DESCRIPTION, TECTONICS AND HYDROGEOLOGY. THERMAL WATERS. MINERAL WATERS

Thème 2

DESCRIPTIONS HYDROGÉOLOGIQUES RÉGIONALES, TECTONIQUE ET HYDROGÉOLOGIE. EAUX THERMALS. EAUX MINÉRALES

GENERAL REPORT 1

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This article is a review of twenty papers to the Theme 2 of the IAH – IAHS International Hydrogeological Conference, held in Budapest, 31 May – 5 June 1976.

The Pannonian Basin and its environment

A comprehensive study has been presented by A. Rónai, overlooking on the long-time fundamental work of the Hungarian Geological Institute. Details of geology, hydrogeology, water chemistry, temperature and pressure conditions of the subsurface waters have been presented especially for the upper part (0 – 1 km) of the Basin's sediments. The Pannonian Basin is one of the best-explored sedimentary basin of the world. The Institute's network of boreholes for continuous scientific observations is of special interest. This network put in action several years ago, has revealed clear interactions between ground-water level fluctuations on the one hand and terrestrial and atmospheric tides on the other.

Based on the data of Rónai's paper and other studies (e.g. Stegena, 1970), the present reviewer makes an attempt to draw up a generalized scheme for the subsurface water movements of the Pannonian Basin (Fig. 1).

1. *The zone of infiltration* includes the uppermost 300 – 700 m (approx. down to the bottom of the Upper Pannonian sediments). In this zone, there is a free communication between meteoric and subsurface waters. As a consequence, fresh waters and approximately hydrostatic pressures predominate in this zone; slight hydráulic pressures are controlled by topography.

2. *The zone of expellation* ranges from below the zone of infiltration to the depth of 1 – 3 km (approx. down to the bottom of the Neogene). In this zone the water migration is controlled by compaction and consolidation of sediments; the water of the layers (connate old sea-water) expelled from the sediments by compaction (alteration by pressure) and by consolidation (alteration in time) moves up-and-sidewards. Mineralized waters and local or regional overpressures characterize this zone.

3. *The zone of re-infiltration.* At greater depths the process of compaction of the rocks comes to an end, and the consolidated rocks may give room again for a second filtration process. Beneath the Pannonian Basin, this zone, if

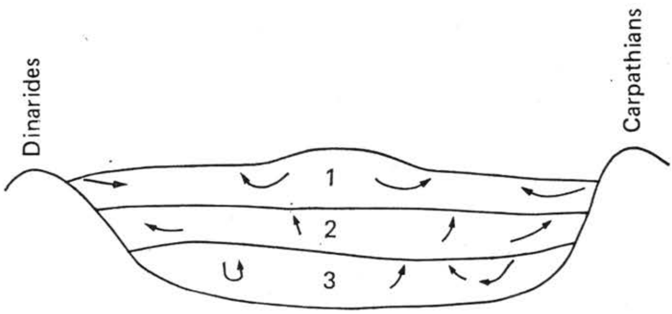

Fig. 1. The author's generalizing scheme for the subsurface hydraulic water provinces in the Pannonian Basin
1. Zone of infiltration, *2.* zone of expellation, *3.* zone of re-infiltration

any, may exist in the Mesozoic and older basement of the basin. No sufficient experiences allow to characterize this zone; hydrostatic pressures and mixed waters are presumably present.

If this model were adopted for the Pannonian Basin, it would be an open question what to say about its general validity for large, closed sedimentary basins? Evidences of some other papers, however, support or do not contradict to these ideas.

In a smaller area of study, S. Karácsonyi and Gy. Scheuer carried out a complex investigation of a part of the Danube riverside near Dunaújváros, a new industrial centre in Hungary. The riverside of the Danube displays here high banks, the failures of which have been proved damaging. In observation boreholes, an 8-year-observation of water level fluctuations has verified to exist a marked relation of the latter with level fluctuations of the Danube river.

The *eastern margin of the Pannonian Basin* exhibits hydrogeological conditions similar to those of the Great Hungarian Plain, as showed by A. Mehmed, T. Bandrabur, P. Craciun, C. Ghenea, P. Polonic and M. Visarion. Here, in the Romanian part of the basin the Pannonian sediments (<2500 m) represent the main aquifer for thermal waters as well. B and R (Basin and Range) structures ("tectonique cassant") in the basement are better developed in this marginal area of the Pannonian Basin than in its central parts. E.g. the Cadea-Gálospetreu-Mecentin graben, discovered seismically, exhibits thermal waters and geotemperatures of 140 °C and even higher. A geothermal map (for the depth of 1 km) has been presented as showing a good accordance with the embracing an adjacent zone on the Hungarian side. Local geothermal highs are controlled by fractures (like in the Hungarian part of the basin; see the case history of Tiszakécske by L. Alföldi, M. Erdélyi, J. Gálfi, K. Korim and P. Liebe, among themes of this conference).

B and R structures in the basement occur also in other marginal parts of the Pannonian Basin, as contributed by E. Kullmann, regarding the NE part of the Vienna Basin. *The Zohor graben* ($100 \times 5-10$ km, $50-150$ m proluvial sediments) is very favourable to water production.

M. Stojsic and A. Kukin report on the little (5.6 km^2) Lake Palić (southern Pannonian Basin, Novi Sad, Yugoslavia) formed by aeolian action, a victim of water pollution (by the town of Subotica).

Other European basins

Besides the Pannonian study by A. Rónai, another comprehensive study was presented at the conference on the hydrogeology of the *Po Valley*, by C. Bortolami, G. Braga, A. Colombetti, A. Dal Pra, V. Francani, F. Francavilla, G. Giuliano, M. Manfredini, M. Pellegrini, F. Petrucci, R. Pozzi and S. Stefanini. Their work represents the first general synthesis made by nine Italian scientific institutions, after a 4-year common work. The Po Valley constitutes a definite geographic and geological entity, a large subsiding basin filled by Pliocene and Quaternary sediments and bordered by the Alps in the north and by the Apennines in the south. In comparison with the Pannonian Basin the main difference consists in the fact that the Po Valley features a basin *open* to the Adriatic Sea. The thickness of the Quaternary sediments increases from west to east (Adriatic Sea) from some hundred metres to 2000 metres and more. – As it is known, these Adriatic parts of the Valley have been subjected to strong recent crustal subsidences. Whether this sinking of the shores has mainly been provoked gas and water exploitation (as suggested by Caloi) and/or by isostatic sinking, this remains an open question, at least for the reviewer.

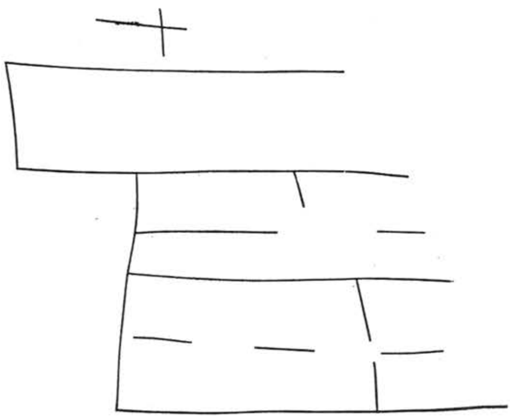

Fig. 2. Coulisse-like structure in Southern Ukraine (overview, after Komarova and Shtengelov)

In Southern Ukraine (*Crimea* and the *NW Pritsernomorsky region*), there are very interesting coulisse-like structures, as presented by M. V. Komarova and E. S. Shtengelov (Fig. 2). These structures have been defined as consequence of recent tectonics: a regional compression had caused joint fractures, in the sense of Becker's strain ellipsoid for principal tangential stresses. The fractures are zones of high permeability, filled with water, providing an opportunity for water exploitation. – As known from deep seismic measurements (Sollogub), in Ukraine there are lots of very deep fracture zones penetrating down to the Moho discontinuity. It would be of interest to know whether a connection between these seismically detected fractures did exist.

The mineralization of deep waters in *NE Poland* does not contradict to the generalized model presented above for the Pannonian Basin. According to C. Kolago and Z. Plochniewski, a regional SE deepening from 0 to 1200 m of the fresh-water bearing layers has been verified. Beneath this zone, low-to-medium mineralized NaCl waters occur. – The former can be identified with the zone of infiltration, the latter with the zone of expellation.

In the *Moscow Artesian Basin*, as showed by N. A. Lebedeva, the fresh-water ("infiltration") zone is 200–250 m thick and is subdivided into 2 stages: an upper one (150–200 m) with higher (~10 cm/year) and a lower-situated one with lower (2–3 cm/year) speed of water filtration and storing slightly salty waters ($\sim$10 g/l). Topography plays an important role in the control of groundwater movements (similarly to the Pannonian Basin; see the paper of M. Erdélyi, this book). — No data deal (in the paper) with the question whether expellation or static circumstances prevail beneath the fresh-water zone.

At the conference only one paper was presented showing a quantitative model study of a big ($\sim$100,000 km^2) basin: the study of M. Besbes, G. de Marsily and M. Plaud on the *Aquitanian Basin* (see Fig. 2 of Besbes and others in this book). On the building up of a unique digital model, it was possible to establish the water balance of the whole basin—an excellent work from every point of view. — It emerges however the question: what is the reality of the four-digit numbers (as shown in Fig. 2 of Besbes)? The introduction of error analysis (calculation of the partial derivatives according to Holzman or Abramovici, etc.) could be of value, like it is in other branches of science, too.

F. Medina Salcedo, R. Fernández-Rubio and A. Gordillo Martin presented a detailed study of a little intermountain basin: the *Marquesado basin, Granada, Spain.* The influence of a permanent pumping is analysed preciously.

E. Sobotha emphasizes the importance of the topography in connection with the water balance of a basin, at first on the example of the area near *Frankenberg*, B.R.D. This statement agrees with the result of Lebedeva (Moscow Basin) and of Erdélyi (Pannonian Basin).

P. Udluft gave a synthesis of water-chemism of the Alpine *marginal areas* in *Southern Bavaria.* A mingling of northern "German-type" $K—Ca—Cl—SO_4$ water and of southern (Alpine-type) water ($K—HCO_3—Cl$) takes place in the layers of this area. Ion-exchange reactions between water and bedrocks are supposed to exist. This may be an important process, often controlling the chemical composition of waters in layers at many places; the natural ion-exchange processes have not yet been studied satisfactorily.

Africa

From Africa, only one case history was presented in regard with the *Douala basin, Cameroun,* from A. Chiarelli. This Atlantic coastal basin is characterized by high overpressures (like the Gulf Coast and the Nigerian coastal areas). The probable common cause of these overpressures may be given by a very rapid sedimentation and a consequent compaction of the argillites, creating near geostatic pressures.

This is in agreement with H. M. van Montfrans's *(Netherland)* idea: the building-up of the lithostratigraphic (=hydrogeological) units is controlled by subsidence velocity.

South America: the Maranhão Basin

A. C. Reboucas presented a comprehensive study of the *Maranhão Basin, Brasil*, of an area of 700,000 sq.km (!). Reboucas' study does not enter details, notwithstanding, that it gives an estimation of the subsurface water supply of this enormous basin. There are permeable sediments, aged from Devonian to Cretaceous with thicknesses ranging to 3000 m. Altogether, about 14,000 km^3 of exploitable water is presumed to be available.

Australia

In the *Gippsland Basin*, according to B. R. Thompson, high geothermal gradients are related to the presence of brown coal deposits. It has been suggested, that this geothermal high might have developed by means of the isolating effect of the coal (this being a bad thermal conductor). Among the factors conceivable (coal oxidation, chemical reactions), the author's idea seems to be most realistic, although quantitative considerations exhibit the existence of some theoretical problems.

Asian basins

A case study of the *Elazig–Uluova Basin, Anatolia*, by M. Sakarya outlines a reservoir system of detrital aquifers; the basin constitutes part of the Euphrates basins. Forecasts the possibility of water exploitation are given till as late as 1990–1994.

The *Mongolian–Baikal orogenic belt*, is described by I. S. Lomonosov and B. I. Pissarsky. In this area, practically, two kinds of water are to be distinguished: *I*. Waters of an area with young volcanism (Baikal rift): nitric thermae $\rightarrow CO_2$ thermae $\rightarrow$ cold CO_2 waters (in the order of age). *II*. Intermountain subsidence's waters (Zabaikalia, North Mongolia): nitric thermae $\rightarrow CH_4$ thermae $\rightarrow CO_2$ thermae $\rightarrow$ cold CO_2 waters.

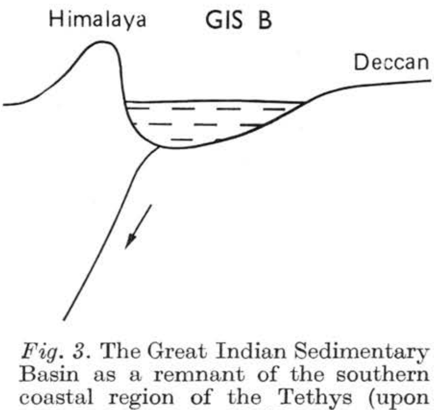

Fig. 3. The Great Indian Sedimentary Basin as a remnant of the southern coastal region of the Tethys (upon Bardhan's idea)

Relationships between seismic activity and variations in thermal water temperatures and chemism have been discussed. This relation may be realistic; earthquake, however, represents a deeper crust—or mantle—process, and some manifestations in near-surface waters are consequences of the secondary deformations that had taken place in the near-surface rocks only.

The study of the *Great Indian Sedimentary Basin*, by M. Bardhan, has suggested a new hypothesis for the development of this giant basin lying in the gap between the Himalaya and the peninsular shield. According to the author, this basin may be a *remnant* of the continental shelf of the *southern coast* of the old Tethys (Fig. 3). This idea is, indeed, in favour of plate tectonics as first put into relation with the region of India by WEGENER.

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GENERAL REPORT 2

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The term "hydrogeology" has been taken in a wider sense. So the papers reviewed hereinafter include both descriptive works and studies discussing genetic and dynamic relationships on the basis of the observed geological make-up. The interpretations have raised issues concerning the way how to recover and appraise new subsurface water resources.

A total of 21 papers, amounting to 376 pages in manuscript, have been received on this subject. Therefore, with no discussion of each paper, we have reviewed the topics that had been treated more exhaustively, however, with making reference to the papers that give ampler details for any theme concerned.

The papers received have concerned the following countries (listed in an alphabetical order):

Australia	1
Brazil	1
Cameroon	1
Czechoslovakia	1
Federal Republic of Germany	2
France	2
Hungary	2
India	1
Italy	1
Poland	1
Romania	1
Spain	1
Turkey	1
U.S.S.R.	3
Yugoslavia	2

It is worth of mention that five of the above papers have discussed the Pannonian Basin that takes up the greater part of Hungary and even extends well beyond her national border.

Before any discussion, let us see how the term "great sedimentary basin" has been defined by us, i.e. by the authors.

One of the authors shows to have been investigated a "large" area of 5 km × 3 km, lying, indeed, within one of the great basins. Others reported from India and Brazil attain half a million square kilometres in size. On the other hand, one paper has discussed such a hydrogeologically independent basin, data of which, in spite of its size as small as a few hundred square kilometres, are very instructive even as projected into the frames of the largest basins.

Basin areas in full could be tackled in a few papers only, as the work of the respective authors was usually hindered by confines such as national boundary or coast-line. Therefore the reviewed papers have been classified according to their complex or special character reflected by the work methods used.

According to the logical succession of description and evaluation, the concerned topics can be detailed as follows:

General hydrogeological description,
Hydrodynamic evaluation,
Water reserve calculation,
Special problems.

There are, however, papers dealing with all the above subjects. Here they are:

M. Besbes, G. de Marsily and M. Plaud make a description of the 100,000-km^2-large Aquitanian Basin in France, open towards the sea, giving an account of the results of appraisal of the renovable water reserves therein. An interesting feature is that numerous semi-permeable layers have had to be taken into consideration. In addition to the infiltrating precipitations, recharge from surface waters and from the sea, moreover by waters from adjacent communicating aquiferous units have had to be taken into account. A balance of subsurface waters for the major stratigraphic units is also given.

E. Kullman deals with a sub-basin as being the Czechoslovakian part of the Vienna Basin, with the calculation of the static i.e. stored and dynamic (renovable) water reserves. The area under discussion, though some hundred square kilometres large, exemplifies the important role that an adjacent mountainous region may play in the water recharge of basins.

N. A. Lebedeva reports on the geological conditions of a basin of open hydraulic system, relying on the example of the Moscow Basin, where the aquifers are predominantly carbonate rocks. Because of the intricate geological and geomorphological conditions, the meteoric and surface waters communicate even with the deepest aquifers. The most essential quantitative data on the calculated subsurface water budget are also given.

A. Cunha Reboucas reports on a typical artesian basin of classic make-up, the Maranhão Basin of about 700,000 km^2 in Brazil. A continuous sedimentation can be shown to have taken place from the Silurian till the Cretaceous. The piezometric levels show a close relationship with topography. A characteristic hydrochemical zonality can be shown to exist both horizontally and vertically, a fact that may refer to the vertical intercommunication of various water horizons. In spite of the comparatively few data available, both the static and dynamic water reserves were calculated.

While the papers reviewed above have been distinguished here because of their complex character, then in the papers to be quoted hereafter some sub-topics have been enfocussed.

Hydrogeological descriptions allow us to draw up on a hydrogeological basis some basin types generalizable in worldwide dimensions. From this point of view, the tectonic setting appears to be crucial.

The sedimentary basins situated between young orogenic belts are built up largely of clastic sediments. Basins of this kind are the Vienna Basin already referred to, moreover the Pannonian Basin in Hungary and around, described by A. Rónai, and the Po Basin reported by C. Bortolami et al., Italy. The detrital cones exposed or buried at the basin margins are of particular importance.

A special kind of the basins in question is demonstrated by F. Medina Salcedo et al., reporting on the Guadix-Baza Basin from Southern Spain, a basin filled up almost exclusively with coarse detritus.

The other main type of sedimentation is represented by the basins developed on platforms or their margins. In these the most important aquifers are often built up of carbonate rocks, while the role of detrital cones is secondary due to the fact that mainly finer-grained sediments are transported from the platforms. Basins of this type are the Moscow Basin already mentioned and the Maranhão Basin, Brazil, described by de Cunha Reboucas.

A transition between the two types is represented by the basins flanked by an orogenic belt on the one side and by a platform or a massif on the other. Such a basin is the Aquitanian Basin already referred to and the large Indian sedimentary basin to the south of the Himalaya, discussed by M. Bardhan.

Some authors lay particular stress on the role of the basin substratum, whence a basin can also be fed with water. Naturally, this depends also on where the boundary of the basin substratum is taken vertically.

The connections between geology and hydrodynamics are tackled by most authors and even a surprisingly uniform approach is adopted by them. These problems are discussed in a most comprehensive way by A. Rónai reporting on the Hungarian part of the Pannonian Basin and by C. Bortolami et al. discussing the Po Basin, Italy. So the following most essential relationships concerning the dynamic connections in sedimentary basins have been pointed out.

It is most important for the clarification of the hydrodynamic conditions to determine the piezometric characteristics of the basin.

A close relationship can be shown to exist, as a rule, between surface morphology and the phreatic ground-water level.

The existence of a connection between the piezometric level of confined waters and the morphology can also often be revealed, though this may be rigorously influenced by stratification.

Piezometric levels showing a tendency to decrease with depth are indicative of the zones where infiltration prevails. On the other hand, an increase towards the depth of the same levels would imply the existence of water discharge. (There may be, of course, exeptions too.)

The concentration of dissolved salts is characteristic of the velocity of underground flow i.e. of the rate of recharge.

Important connections can be shown to exist between temperature and hydrodynamic conditions of waters, not only in the case of thermal waters.

The role of tectonics i.e. faults is pointed out by several authors.

It is important to separate open, closed and semi-closed portions, if exist, in a basin, moreover to estimate the role of semi-permeable strata and to recognize the presence of stratigraphic-hydrogeological windows. A closed system is discussed in only one paper by. A. CHIARELLI taking for example the Douala Basin, Cameroon.

Data useful for the solution of the above problems are presented by C. GRIOLET from the Rhone Valley, France and by P. UDLUFT from the northern foothills of the Bavarian Alps, FRG. These areas are not regarded as large sedimentary basins. It is the paper of E. SOBOTHA to be mentioned here as dealing with the significance of the climatic and hydrogeological conditions of the basins' edges, concerning the Rhine Valley region, FR of Germany.

As already referred to in the introduction, the statements made in the papers contribute to the solution of the problems of subsurface water recovery or they are directly aimed at the calculation of the water reserves as being exemplified by the first-mentioned four papers (M. BESBES et al. from France, E. KULLMAN from Czechoslovakia, N. A. LEBEDEVA from the U.S.S.R. and DE CUNHA REBOUCAS from Brazil). It has been evidenced by them that the feasibility of water reserves appraisal primarily depends both theoretically and technically on the number of data and on the degree of homogeneity of the geological makeup. The respective extremities are marked, on the one hand, by the water reserve calculation in the Aquitanian Basin, France, where the data from 10,000 grid points of observation over a total area of 100,000 km^2 could be processed by computer, and, on the other, by M. SAKARAYA who could assess the reserves and the prognosis or production of Elazig Uluova, Turkey, in a well-outlined basin of 370 km^2, by simple mathematical tool using just a couple of formulae.

The most essential types of water reserves are treated by the authors according to uniform principles, described though with unlike denomination. By this we allude to the static and dynamic or renovable water reserves. Because of the changing nomenclature, it is often the only basis the dimension of the unit of measurement used that indicates the type of reserves under consideration.

As can be seen from the above, the works devoted to the great basins discuss mainly the problems of data evaluation. The methods of research have scarcely been dealt with. The only work devoted to this subject is the paper of Z. I. KRULC and S. KOVACEVIC demonstrating the significance of geophysical research in the hydrogeological study of large sedimentary basins. He describes, as an example, the Zagreb Basin, one of the Yugoslavian sub-basins to the Pannonian Basin.

S. KARÁCSONYI and GY. SCHEUER discuss technical problems implied by observation wells tapping confined aquifers in the Hungarian part of the Pannonian Basin. They also consider the influence of the water level of the river Danube upon the variation of the piezometric level of nearby confined waters.

Of the special problems, let us quote those of the thermal waters dealt with in three papers. These works are of particular actuality now, at the time of an energy crisis. The problematics is profoundly analysed by B. R. THOMPSON taking for example the Gippsland Basin, Australia. He points out particularly the role of the strata of low heat conductivity, underlain by confining beds. In such a structure water can flow beneath a heat-insulating layer (e.g. coal bed) and accumulate heat supplied by the heat flux and, after getting closer to the surface if enabled by the geological structure, it would provoke positive thermal anomaly. It is quite possible that the heat flux at the site of the anomalies is greater in the basin substratum as well. B. R. THOMPSON verifies his conclusions by thermal balance calculations. He describes a method how to take into account the variation in volume and viscosity for the evaluation of thermal waters.

The thermal waters of Romania's share of the Pannonian Basin are dealt with by E. ALI-MEHMED—T. BANDRABUR and P. CRĂCIUN et al. They show a close connection to exist between geothermal anomalies and geology.

I. S. LOMONOSOV—B. I. PISSARSKY and S. D. KHILKO deal with a mountainous region within the Mongolian-Baikal orogenic belt. They indicate the relationship between the hot springs and tectonics, observing the predominance of carbon dioxide hydrothermae near the volcanoes and that of the nitric hydrothermae farther away from the volcanoes.

M. V. KOMAROVA and E. S. SHTENGELOV call the attention to the sudden increase of permeability appearing mainly in carbonate rocks along fault systems in young orogenic belts in the Ukraine (U.S.S.R.). Their work was supported by an investigation regarding geology, seismic measurement, radiometry, well-logging, radio waves.

C. KOLAGO and Z. PLOCHNIEWSKI have found remarkable relationships between the chemical composition and the geological setting, in aquifers with waters of 1 gram per litre concentration in the plains of N Poland.

The problem of environmental control is dealt with in only one paper. M. STOJSIC and A. KUKIN are reporting on the change of the chemical composition of Lake Palić in the Yugoslavian part of the Pannonian Basin and on the upset of its biological equilibrium as a result of pollution by sewage. At the same time, they give some information on the measures undertaken for the regeneration of the lake. This is alimented partly by phreatic groundwaters, so the experiences are also important for the knowledge of the subsurface waters. As far as damaging effects by man are concerned, this problem is treated by C. BORTOLAMI et al., taking for example the Po Basin, Italy. They call attention to the fact that infiltration, and so the recharge of groundwaters from the surface, is considerably reduced by the growing urbanization at the expenses of the extension of irrigated lands. In addition, very harmful is the effect of the chemical substances discharged into rivers, as these decrease the leakage through the riverbed which also leads, in turn, to a decrease in the recharge of the subsurface waters. These environmental problems are manifested on a worldwide scale and, unfortunately enough, a further worsening of the situation is still to be expected. Great efforts should be made in finding out the causes of the harmful effects, in order to that protective measures be taken before it would be too late.

As evident from the above, the papers reviewed here have raised a good many actual problems. Although not covering entirely the pertinent topics, the authors have obviously intended to resume their latest and orientative results.

The importance of the problems raised might be changing in time and from place to place, according to the conditions given by the respective geological setting and water production. Nevertheless, it can be concluded that the geological research tends to become more quantitative, above being based on descriptions and theoretical considerations.

EVOLUTION OF THE REGIONAL HYDROGEOLOGIC UNITS OF THE GREAT INDIAN SEDIMENTARY BASIN IN RELATION TO PREVAILING TECTONIC MOVEMENTS

ÉVOLUTION DES SECTIONS HYDROGÉOLOGIQUES RÉGIONALES DU GRAND BASSIN SÉDIMENTAIRE INDIEN PAR RAPPORT AUX MOUVEMENTS TECTONIQUES

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RÉSUMÉ

L'auteur a étudié le rôle des mouvements tectoniques dans la formation des principales sections hydrogéologiques régionales du Grand Bassin Sédimentaire Indien, situées dans l'ouverture entre la chaîne montagneuse de l'Himalaya au Nord et le rempart péninsulaire au Sud. En s'appuyant sur les données géohistoriques et les résultats des récentes explorations géologiques, il propose une nouvelle hypothèse expliquant l'origine du bassin. Il démontre que les mouvements tectoniques ont exercé une influence décisive sur la détermination de la configuration, de la structure, de la stratigraphie, de la lithologie et même des caractères physico-chimiques des quatre différentes sections géologiques du bassin, à savoir, les zones de Siwalik, Bhabar, Terai et Alluviale. L'auteur étudie également les principales caractéristiques hydrogéologiques de ces sections en vue de souligner leurs traits distinctifs.

Introduction

The Great Indian Sedimentary Basin, drained by the Indus – Ganges – Brahmaputra river system, is one of the largest and most productive groundwater provinces of the world. It is flanked by the Great Himalaya in the north and the Deccan Shield in the south (Fig. 1). The history of development of this basin and its sedimentary units is closely related to the orogenic and neotectonic events of the late and post Himalayan times. These events are traced back to assess their impact on the formation of the basin and its filling materials. For this purpose, reliance is placed on the geohistorical and recent geotechnical informations. It is shown that there had been both direct and indirect influences of these movements in shaping the basin as well as in deciding the distribution and nature of the sediments. The existing ideas pertaining to the origin of the basin are discussed and a new hypothesis is proposed, which appears to better explain the present structural and hydrogeologic features of the basin as well as the history of sedimentation leading to the development of the regional hydrogeologic units.

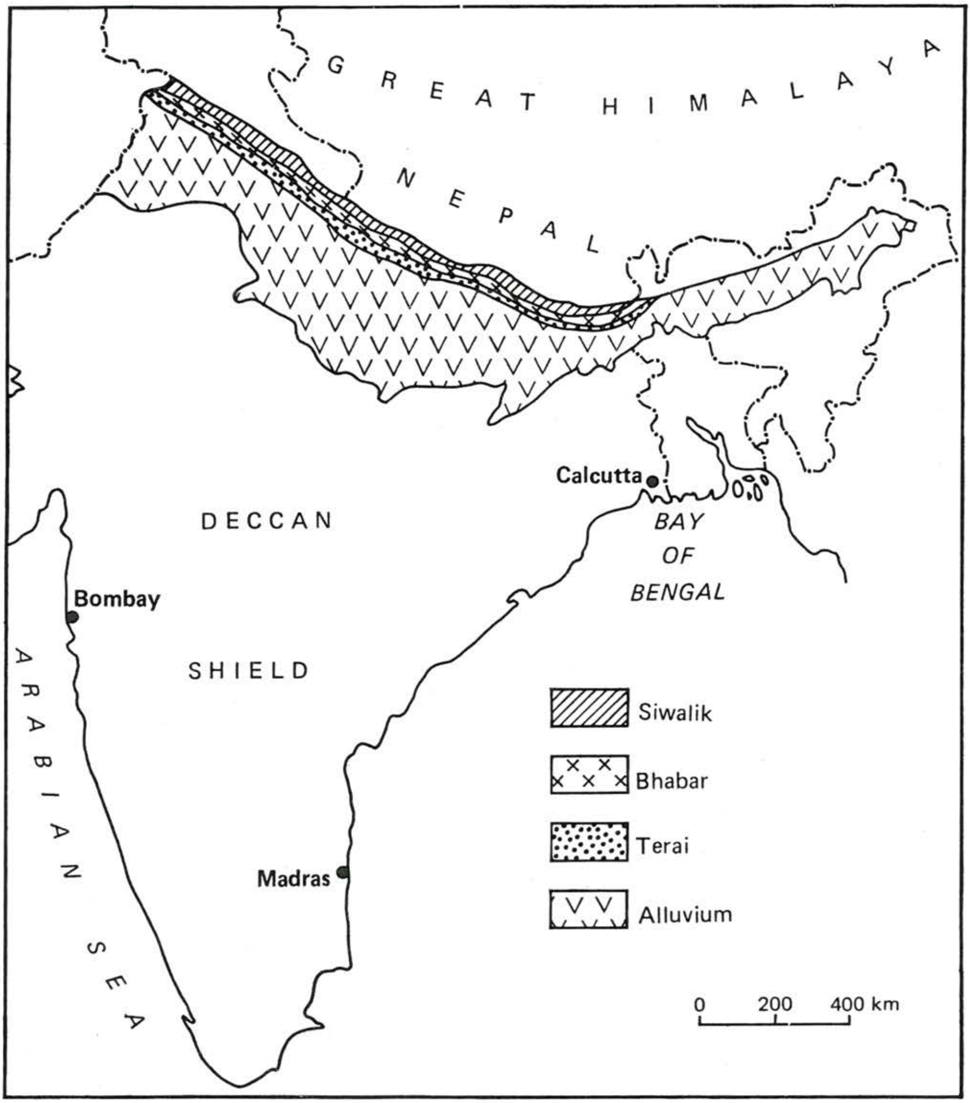

Fig. 1. Distribution of regional hydrogeologic units in the Great Indian Sedimentary Basin

Description of the basin and its hydrogeologic units

The basin runs a length of over 2400 km from Punjab in the west to Assam in the east. Its width is variable, the maximum being over 400 km and the minimum as small as 25 km (Fig. 1). The floor of the basin appears to be highly disturbed by NE–SW trending cross faults, as have been located at Patna, Lucknow, Moradabad and other places in the basin, giving rise to a good number of horsts and grabens (MITHAL and SRIVASTAVA, 1959). While the configuration of the basin floor has not been worked out in detail, it can be reasonably inferred from the available information that the presence of these horsts and grabens together with the existence of a large number of apophyses of Peninsular shelf margins, is liable to render the topography of the basin bottom highly rugged and irregular. This situation should lead to the uneven spatial distribution of thickness of the sediments in the basin, so also the groundwater regime, particularly in the deeper zones. It is further postulated that there exist at least five to six deeper sub-basins along the northern margin of the basin with different groundwater regimes (MITHAL, 1966).

The basin is filled with four distinct sedimentary units designated as the Siwalik, Bhabar, Terai and Alluvial formations (MEDLICOTT, 1873), which are disposed in a nearly parallel fashion between themselves as well as with the Himalayan range. Though the major part of the Siwalik formation is now present as the Siwalik hills along the northern border of the basin and considered as a separate geologic system, from a tectonic angle it can be considered as an integral part of the Great Indian Sedimentary Basin, as will be clear from the discussions to be taken up later. In the following the relevant salient features of the four units of the basin are presented to facilitate further treatment of the subject matter at hand.

Siwalik system

The Siwalik system derived its name from the Siwalik hills lying in the Hardwar region of Uttar Pradesh. It continuously extends along the southern foot of Himalaya from the Brahmaputra valley in the east, to the Potwar Plateau and Bannu plaines in the west. Its equivalents are seen in Sindh, Baluchistan, Assam and Burma. The Siwaliks comprise consolidated and semi-consolidated rocks namely sandstones, grits, conglomerates, pseudo-conglomerates, clays and silts. They have the character of deposits formed by torrential streams and floods in shallow and fresh-water basins. The fossils found in them indicate that the earlier beds were deposited in a somewhat brackish environment. The Siwalik system is divided into three major divisions namely the Lower, Middle and Upper Siwaliks, ranging in age from Middle Miocene to Lower Pleistocene. While there are no marked unconformities within a system, there are indications that the Upper Siwaliks were deposited on the Middle Siwaliks after severe tectonic disturbances resulting in folding and uplift.

So far the Siwaliks are very little exploited for groundwater. However, due to high elevation, in general the groundwater conditions in them appear to be difficult except in rare, wider longitudinal valleys.

Bhabar formation

The Bhabar formation comprising boulders cobbles and gravels as piedmont deposits occurs all along the southern slope of Siwaliks as a distinct belt, varying in width between 3 and 24 km. The formation occurs as an accumulation of talus materials and coalescent alluvial cones built by the hill streams. Groundwater is known to stay in this formation as unconfined. The water table is generally deep but varies between 5 and 90 m below ground level. The ground slope is high and towards the south in the range it descends 8 to 17 m per km. The individual alluvial cones in the Bhabar belt have their own set of aquifers and the adjacent cones appear to have poor interconnection (SAH, 1966). The water table contours generally reflect upon the surface topography. Yearly fluctuation of water table is rather high and a figure of 8 m is quite common. In general the ground water conditions are considerably better along the southern portion of the belt, both in respect of depth of occurrence and quantum of availability. Most of the surface streams in the Bhabar zone remain dry except during monsoon (July to September), though they may be perennial both upstream in the Siwalik hills and downstream in the Terai zone, respectively. This is due to the absorption of the flow of these streams in the highly porous materials constituting the Bhabar.

Terai formation

Immediately following the Bhabar belt on its south is the Terai belt, composed of alternate layers of clay and sand-pebble beds. A spring line is usually seen to separate the Bhabar from the Terai. These sand beds, except the topmost one, usually form artesian aquifers, in which the piezometric level lies at 0.3 to 1.5 m above ground surface. The pressure head shows a tendency to decrease from the north to the south. The confined aquifers occur at varied depths, but more often below 60 m. The general slope of ground is towards south averaging about 0.4 m per km. The sand-clay ratio in the Terai formation is of the order of 25:75. The granular beds, mostly confined to the stream channels, appear as tongue-like projections into thick clays, often maintaining a lithological continuity with Bhabar in the north. The permeability of the granular zone in the Terai formation show large spatial variation due to the presence of highly irregular textural gradation. In this zone the streams are perennial, and many of them receive discharge from the spring-line intervening the Bhabar and Terai.

Alluvial Plain deposits

On its south, the Terai belt is followed by the vast alluvial plain comprising of sand and clay with kankar. The sand beds constitute highly rich aquifers. It is interesting to note that in the northern half of the Plain the aquifers maintain a continuity in the N–S direction, whereas in the southern half an E–W continuity is exhibited. On a regional scale the aquifers are unconfined but subartesian conditions have developed locally (SINGHAL and GUPTA, 1966). However, flowing wells are practically absent in the alluvial Plain. The depth of water table or piezometric surface lies within about 4 to 12 m from the terrain level. The aquifers commonly show a lenticular character indicating that the sand and gravel layers were deposited in the channel beds whereas the silt and clay beds were formed in the flood plains (MITHAL et al., 1973). It is estimated that the alluvial plain together with the Terai and Bhabar zones, covering an area of about 1,048,500 sq.km holds a groundwater reserve of approximately $9.08 \cdot 10^{13}$ cu.m within a depth of 300 m from the terrain level (MITHAL, 1966).

Evolution of the basin

The history of development of the basin is interlinked with the formation of Himalaya and hence with the past existence of the great mediterranean sea or Tethys geosyncline. The basin emerged as a by-product of the tectonic process that moulded the sediments of the Indian portion of the Tethys geosyncline into the Himalaya mountains. The movements associated with this tectonic process acted intermittently throughout the Tertiary period in 5 to 6 major stages, beginning at the end of Cretaceous and culminating in the Early Pleistocene. The basin appears to have begun to form in the Late Eocene and attained the fullest development in the Middle Miocene, during the third and most violent Himalayan upheavel, at the end of which the Tethys basin disappeared. While the major tectonic movements built the initial basic framework of the basin, finishing touches towards its final shaping were given by the neotectonic movements. There are at least three different opinions regarding the type of tectonic movements responsible for the emergence of the Great Indian Sedimentary Basin (KRISHNAN, 1960), as discussed below.

According the EDUARD SUESS, the basin is a "fore–deep" formed due to compression exerted on the Tethys sediments by the moving Central Asian mass in the north, against the stable Indian Peninsular mass (Deccan Shield). But SIR SYDNEY BURRARD considered the basin as a rift valley formed by parallel faulting on both sides, that is, along its northern and southern borders. The third view is that the basin is a sag in the crust in front of the rising Himalaya, caused by the northward drift of the Indian sub-continent against the Tethys sediments. However, taking indications from the recent geotechnical, especially geophysical data, a new hypothesis for the origin of the basin can be proposed. In the opinion of this author the basin is the remanent southern continental shelf zone of the old Tethys sea which refused uplift during Himalayan orogenesis with the support of the resistant Deccan Shield. The following observations tend to back up this hypothesis.

Firstly, recent gravity anomaly data (QUERSHY, 1964) give evidence that in the north the Deccan Shield forms a wide continental shelf margin under the Indo-Gangetic alluvium of which the apophyses appear to extend even below the Siwaliks. Secondly, the Siwalik sediments indicate that water of the basin of deposition was first brackish and then it became increasingly fresh with time. Thirdly, recent deep geophysical survey and drilling activities have proved that contrary to the earlier belief, the thickness of the sediments in the basin is not of the high order of 12,000 to 15,000 m but much lower. As for example, in Ujhani, Kasganj and Tilhar in Uttar Pradesh the thickness of alluvium averages only 400 m. Forthly, it appears from the study of the general shape of the depression of the basin that it is deepest within a few kilometres of the Himalayan foothills, and progressively shelves up towards the Deccan Shield in the south.

Development of the sedimentary units

The tectonic movements were not only responsible for shaping the basin, as already discussed, but dictated the terms of sedimentation and the nature of the materials deposited therein. The tectonic forces acted in three principal directions namely upward, downward, and lateral as will be further discussed later. At the outset of this section, it should be mentioned that while the long-term sedimentation sequences were principally controlled by the paleotectonic and neotectonic movements. The short sequences were mainly determined by the seasonal climatic cycles. This is true of all the four sedimentary units of the basin, but conspicuously traceable in the case of Siwalik system. Hence this aspect will be discussed in greater detail with reference to the Siwalik formations.

The basin took its root in the Middle Miocene, that is after the third episode in mountain building, and since then the sedimentation history became prominent facilitating the deduction of the tectonic and climatic environments then prevailing. The sandstones of the Siwalik system are coarse-grained and mostly ungraded, indicating that they were borne by torrential streams. The great lateral extension of the sedimentary beds suggests that the basin of deposition was practically continuous from Assam in the east to Punjab in the west. Further, the sediments are extraordinarily similar over long stretches along the strike of the basin signifying that they were derived from the same or similar source rocks. It appears that the prevailing tectonic conditions helped in deriving most of the sediments from the north due to the denudation of the newly risen Himalaya, offering terrain with steep slopes to the south. The coarse sediments may have been derived in the wet flood season. The presence of the ancient stable Deccan Shield in the south offered low relief to the streams originating from it, which appears to have contributed mainly the finer ferruginous materials.

The character of the sedimentary sequences and the high rate of deposition of coarse materials reflects upon the presence of monsoon climate during the greater part of the Siwalik time. This climate might have favoured high rainfall, being induced by the orographic effect caused by the establishment of Himalaya. The abundance of fossils of plants, molluscs, fishes, reptiles and mammals in some areas in Siwalik also testifies to the presence of favourable climate and hydrologic conditions during that time.

The thickness of the Siwalik sediments is 5000 – 6000 m, while they are mostly coarse. Such large thickness course of shallow water deposits gives evidence of the interplay of neotectonic movements during the period of sedimentation. It leads one to think that with an increasing accumulation of the sediments there had been gradual downward movement or sinking of the basin. Further, the intermittent activity of the lateral tectonic forces may be inferred from the fact that the basin of deposition has shifted in steps towards the south, particularly at the end of Middle Siwalik time. Probably these shifts were associated with the pulses of uplifts. The evidence of the occurrence of the upward movement, even late after the creation of the basin is given by the fact that while the sediments are thrown up as the Siwalik hills along the northern border of the basin, to the further south the Siwalik formation continues below the younger sediments of Bhabar, Terai and even Alluvial Plain. This phenomenon of asymmetrical uplift of the basin made the earlier-deposited sediments, along the northern part of the basin, the source rocks for the later sediments. Hence, it is seen that the tectonic movements have greatly influenced the order and pace of sedimentation as well as the location of the source rocks. Thus the stratigraphy and the texture of the formations in the basin were influenced by the tectonic movements.

This is not all. In the following it will be discussed how the structural, hydrophysical and even the hydrochemical characteristics of the sediments were influenced by the tectonic movements.

The intermittent southward shift of the basin coupled with the progressive rise of Himalaya helped the Bhabar to overlap the southern portion of the Terai and further the Terai, in turn, to come over the alluvium of the Plain. This process of overlapping, engineered by the tectonic movements, made the boundaries between these formations irregular and confusing. Further, this process is perhaps responsible for the creation of some perched aquifers in the Bhabar and Terai belts. A significant impact brought forth by the tectonic movements is the frequent change in the drainage pattern of the basin which induced far-reaching geohydrological changes, such as rapid spatial variation in thickness, particularly of the Plain. As a result of the frequent shift of the drainage channels the disposition of the individual aquifers is rendered highly complex, so much so that on a regional scale the various aquifers in the basin are interconnected, although locally they may exhibit semi-confined and leaky conditions (SINGHAL and GUPTA, 1966). The frequent dislocation of the stream courses and change in their gradient, caused by the tectonic movements, also influenced the stratigraphy, lithology and configuration of the alluvial cones in the Bhabar zone, such as the rivers Gola, Nihal, Nandhaur, Bhakra, Kosi etc.

Regarding the influence of the tectonic movements on the quality of water in the basin the following observations can be made. Due to the uplift of the northern portion of the basin into the Siwalik hills a high ground slope is established along its northern border, which is gradually moderated in the direction of south, towards the Plain. This created the condition for maintaining a favourable hydrochemical regime in the northern three hydrogeologic units of the basin, namely the Siwalik, Bhabar and Terai. The high ground slope induced steep hydraulic gradient in the aquifers of these units, thereby facilitating good flushing and quick subsurface drainage. As a result, the groundwaters in these units are generally fresh with low content of dissolved solids. However, the alluvial deposits in the Plain, occupying low elevation, hold groundwater of variable chemical quality. While in the well-drained areas of the Plain the groundwater is fresh, in clayey and stagnated pockets brackish and even saline water is seen to occur. It may be noted, however, that in the Terai belt while the confined aquifers are generally fresh, the top water table aquifer is often brackish. This may be due to the accumulation of salt in this aquifer owing to evaporation from the groundwater table, which is quite shallow in this belt. Further, it would appear resonable to say that the hydrochemical regime of the basin, particularly along its northern part, has undergone intermittent changes due to the redisfriction of the drainage network from time to time. This situation leaves the possibility of existence of some subterranean fossil hydrochemical regimes in some areas of the basin.

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BILAN DES EAUX SOUTERRAINES DANS LE BASSIN AQUITAIN

BALANCE OF THE UNDERGROUND WATER RESOURCES IN THE AQUITAIN BASIN

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ABSTRACT

A model of the entire Aquitain Basin was built. It covers a surface of more than 100,000 km^2 and is intended to simulate the hydraulic behaviour of all the sedimentary formations of the basin. To begin with, the analysis of the hydraulic system has made it possible to distinguish eight main aquifers, which communicate through aquitards whose ages run from the Jurassic to the Quaternary.

The building of a unique digital model has made it possible to establish the water balance of the whole basin and to estimate the magnitude of the flow exchange; internal exchanges by leakage between the different aquifer layers and flow at the boundaries of the basin.

Introduction

Le Bassin Aquitain occupe une superficie de plus de 100,000 km^2 dans le Sud-Ouest de la France. En vue de planifier l'utilisation des eaux souterraines de la région, un modèle mathématique à structure multicouche a été construit. Ce modèle doit simuler le comportement actuel et futur de l'ensemble des formations sédimentaires du bassin.

La grande extension des formations aquifères continues sur tout partie du bassin n'a évidemment pas permis de disposer d'une information dont la répartition dans l'espace soit homogène. On a alors eu recours à différentes techniques de la géologie [2] pour combler certaines lacunes. Ainsi, l'examen lithologique des coupes de forage a permis de dresser une carte approximative des transmissivités sur une partie du bassin. La géochimie met en évidence des échanges de flux, la tectonique explique certaines circulations préférentielles et la géologie des affleurements permet d'y évaluer l'alimentation des nappes.

Inventaire des réservoirs

L'analyse du système hydraulique a permis de différencier huit réservoirs aquifères principaux allant du Quaternaire au Jurassique, séparés par des couches semi-perméables, dont la disposition est schématisée sur la coupe Fig. 1.

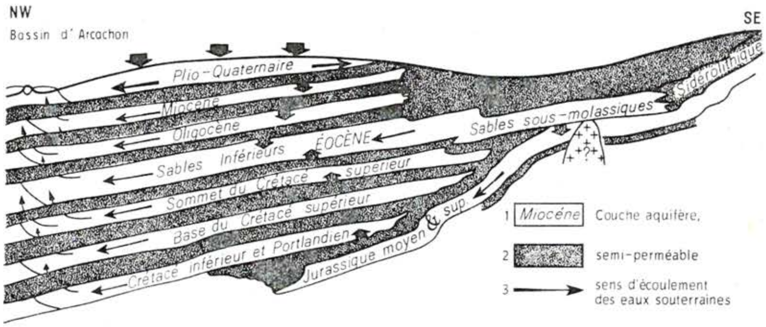

Fig. 1. Schéma de la structure multicouche représentée sur le modèle

Le Plio-Quaternaire: Constitué par le sable des Landes et les graviers Pliocènes, il renferme la nappe phréatique. Celle-ci reçoit l'infiltration de la pluie sur toute son étendue.

De nombreuses études de bilan sur la nappe phréatique, localisées dans l'espace, ont été réalisées par l'Université de Bordeaux et la plupart des auteurs aboutissent à des coefficients d'infiltration compris entre 15 % et 25 % de la pluie, qui est de 900 mm/an en moyenne dans la région. Une partie de cette eau va rejoindre les aquifères profonds par percolation, mais cette fuite peut difficilement être évaluée par les moyens du calcul hydrologique. Cependant, une grande partie des quantités infiltrées s'écoule, soit directement, soit par l'intermédiaire d'un système de canaux de drainage, vers l'Océan et les lacs ou les cours d'eau limitrophes.

Le réservoir Miocène: Il constitue un aquifère continu limité à l'Est par la Gironde et la Garonne, et au Sud par l'Adour. L'alimentation se fait au toit à partir du Plio-Quaternaire et les exutoires se trouvent en mer et dans certaines vallées où existent des affleurements perméables, et un débit non négligeable devrait percoler en profondeur vers l'Oligocène sous-jacent.

Le réservoir Oligocène au faciès carbonaté détritique, est bien développé sur la moitié Ouest du Bassin. A l'Est, il passe aux formations de la molasse aux caractéristiques hydrauliques médiocres. L'alimentation se fait essentiellement par le toit, en provenance du Miocène, mais un certain débit transite à travers le mur, en provenance de l'Éocène dans la partie Sud du Bassin. L'exutoire principal est constitué par le système hydrographique de la Garonne, mais la nappe s'écoule également en mer et dans le Miocène par percolation verticale à proximité du littoral.

Le réservoir Éocène: Il s'étend sur l'ensemble du Bassin. De par ses dimensions et ses bonnes caractéristiques hydrauliques, il constitue l'aquifère le plus important de la région. Il se présente sous différents faciès, mais l'aquifère principal est sableux: « sables inférieurs » au Nord et « sables sous-molassiques » dans le Sud du Bassin. Cette nappe, qui s'écoule dans l'estuaire de la Gironde, est alimentée essentiellement par les affleurements de bordure. Elle est intensément exploitée depuis près d'un siècle dans la région de Bordeaux où la baisse des niveaux piézométriques atteint actuellement 30 m, posant un grave problème, dans un avenir proche, de pollution par les eaux salées de la Gironde.

Signalons l'existence des « sables du Libournais » au Nord-Est, lentille sableuse d'extension limitée, située au-dessus des sables Éocènes dont elle est séparée par un semi-perméable à travers lequelle elle s'alimente.

Les réservoirs du Secondaire: Situés parfois à grande profondeur, ils sont surtout reconnus par l'exploitation des résultats de forages pétroliers. Ils présentent un faciès essentiellement carbonaté. Alimentés sur les reliefs de bordure, ils s'écoulent dans l'Océan et surtout à travers leur toit par percolation verticale en direction de l'Éocène à l'aval. On a pu y distinguer quatre couches aquifères:

- le sommet du Crétacé supérieur,
- la base du Crétacé supérieur,
- le Crétacé inférieur et le Portlandien,
- le Jurassique moyen et supérieur.

Le modèle de simulation

L'ensemble des réservoirs mis en évidence représente une superficie cumulée de près de 250,000 km². Or, si on voulait que le modèle soit significatif des écoulements réels, il ne fallait pas dépasser une dimension de maille égale à 5 km dans la région de Bordeaux, où les prélèvements sont importants et les gradients hydrauliques élevés. Dans ces conditions, il aurait fallu construire un modèle de 10,000 mailles, dont l'exploitation aurait été ruineuse, alors que dans certains secteurs, la qualité médiocre des données ne justifiait aucunement une dimension des mailles aussi réduite.

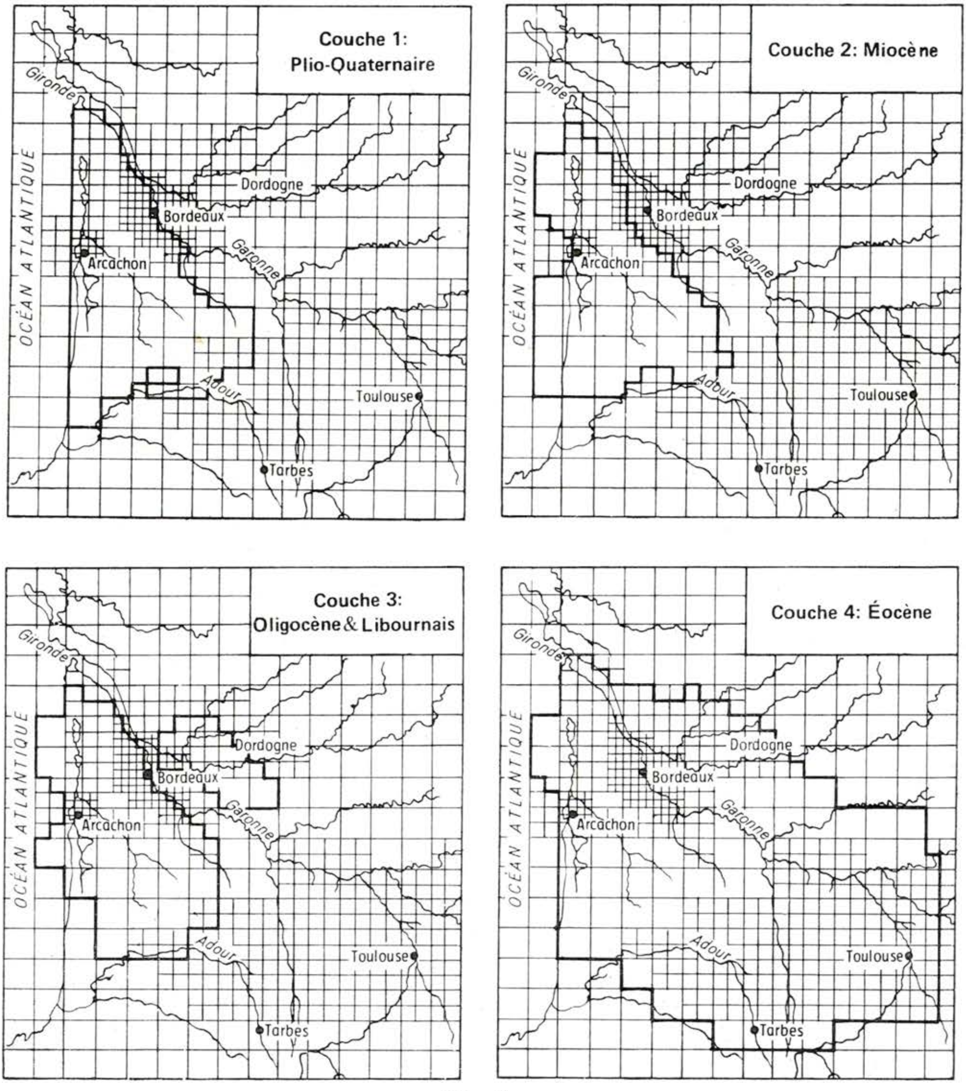

Fig. 2a

Nous avons donc eu recours à un maillage variable dans l'espace en adaptant le modèle du Bassin Aquitain sur le programme SAMMIR (Simulation des Aquifères multicouches à Mailles Irrégulières) mis au point au Centre d'Informatique Géologique [3].

Le maillage à été focalisé sur la région bordelaise et le Bassin d'Arcachon où les mailles ont 5 km de côté. Une dimension des mailles égale à 10 km de côté a été adoptée dans le domaine intéressant la nappe infra-molassique du Bassin Sud-Aquitain. Enfin, des mailles de 20 km ont été disposées dans les zones intermédiaires ou marginales, là où l'information disponible était la moins élaborée (Fig. 2a – 2b).

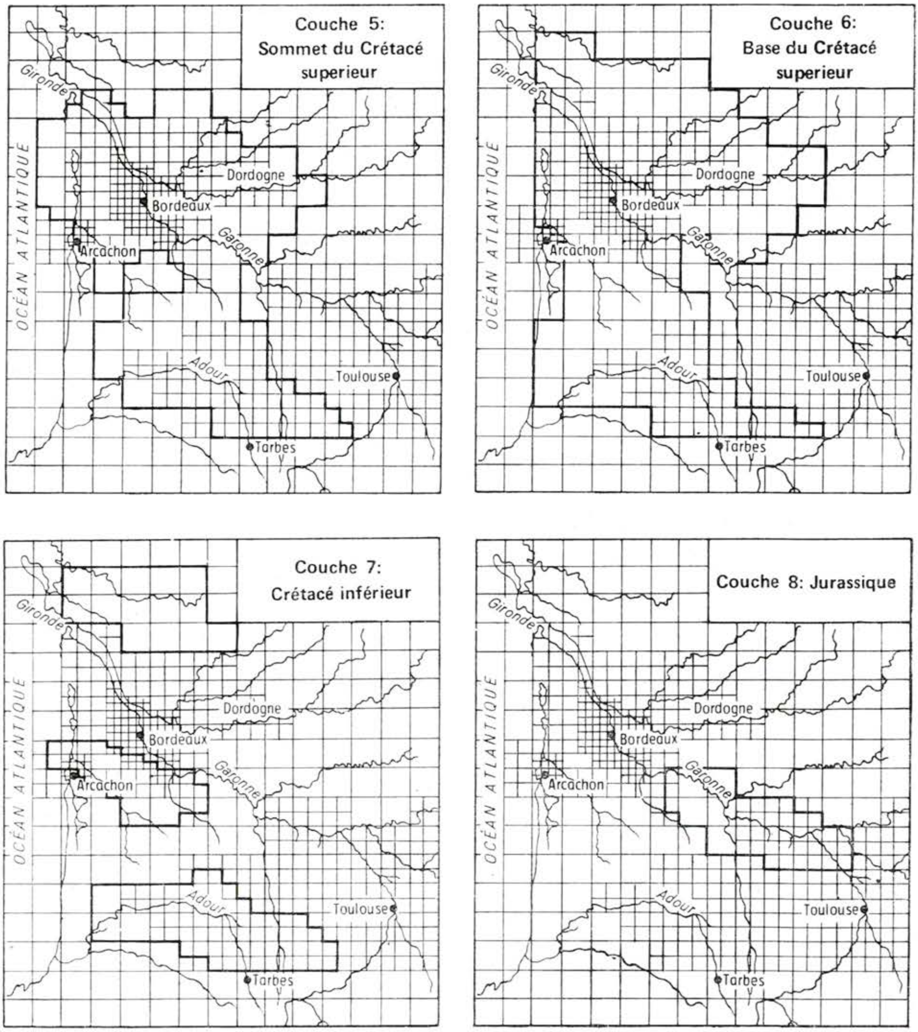

Fig. 2b

Le modèle ainsi conçu comprend 2186 mailles utiles (mailles de travail et mailles de frontières: à flux imposé ou potentiel imposé) dont 1612 mailles de travail, au centre desquelles est calculée la hauteur piézométrique à chaque simulation. Ces mailles sont réparties sur huit couches de travail représentant les huit entités perméables, séparées de couches semi-perméables.

Dans le modèle, on fait l'hypothèse que l'écoulement est bidimensionnel plan-horizontal dans les couches perméables et monodimensionnel vertical dans les couches semi-perméables. L'ensemble du Bassin est alors représenté par une succession de nappes communicant entre elles par drainance verticale.

L'ajustement du modèle est réalisé en régime stationnaire et l'état des nappes observé en 1965 a été choisi pour servir de référence au calage. Ce choix constitue une approximation puisque les prélèvements dans l'Éocène, représentant la majorité des exhaures dans le Bassin Aquitain, sont importants depuis près d'un siècle. A la limite, vouloir choisir un régime rigoureusement permanent reviendrait à simuler l'état initial du système, avant toute mise en production. Mais cette simulation ne saurait aboutir à un ajustement satisfaisant des caractéristiques des réservoirs, les données piézométriques relatives à cette période (année 1880) étant rares et imprécises.

Cependant, le choix de 1965 se justifie par une relative stabilisation de la piézométrie entre 1965 et 1969. *A posteriori*, ce choix s'est de plus trouvé renforcé par l'étude du système en régime transitoire. On s'aperçoit alors que la disposition des captages actuels par rapport aux limites du Bassin et la diffusivité de l'aquifère sont telles que 75 % des rabattements permanents sont obtenus sur une grande partie du réservoir, et cela dans un délai de deux années après une modification du régime de production. Sur l'historique connu des niveaux piézométriques, les variations sur un tel intervalle de temps sont faibles et justifient donc l'approximation d'une étude en régime stationnaire.

Au Sud du Bassin, où la profondeur des couches aquifères dépasse souvent 1000 m, l'information est beaucoup plus rare et est fournie principalement par les sondages pétroliers et la géophysique, cartes structurales, cartes de faciès, quelques prises de pression relevées au droit des formations aquifères en cours de forage, et quelques tests de production. Les cartes piézométriques tracées sont très approximatives.

L'essentiel du calage du modèle a consisté à restituer une répartition précise de la piézométrie dans le bassin septentrional cohérente avec la répartition des transmissivités et les débits d'alimentation, en assurant au Sud :

— la reproduction des cartes piézométriques vraisemblables;
— des taux d'infiltration aux affleurements compatibles avec les observations;
— une schématisation des échanges entre nappes en accord avec les hypothèses structurales faites;
— enfin une répartition des paramètres hydrauliques issue des descriptions de faciès rencontrés (logs des sondeurs et logs électriques).

Bilan en eau du système

Une fois le calage terminé, le modèle permet d'établir le bilan des flux de chacun des aquifères, en prenant en compte les échanges de flux avec l'extérieur du Bassin et les échanges entre les différentes couches par drainance.

L'organigramme de la Fig. 3, sur lequel les entités perméables sont représentées par de grands rectangles et les unités extérieures par des cercles, présente la situation en régime d'équilibre correspondant à l'état des nappes en 1965. La valeur des flux d'échange y est exprimée en litres/sec.

Le tableau 1 fait le bilan pour chacune des couches.

Sur la graphique de la Fig. 4, où l'on sépare la nappe phréatique des nappes profondes, se trouve résumé le bilan pour l'ensemble du bassin (valeurs en l/sec).

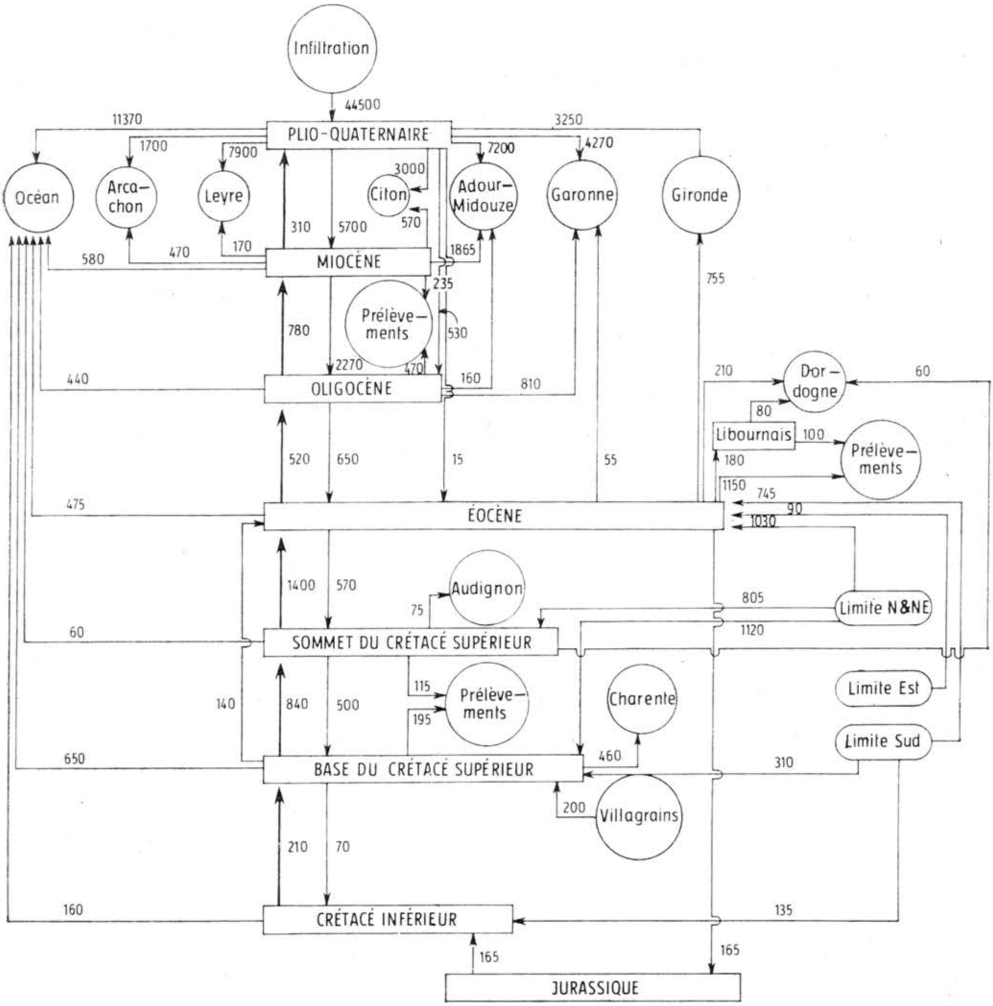

Fig. 3

Considérations sur la signification et l'utilisation du bilan

Le bilan des flux calculé sur le modèle est cohérent avec les transmissivités introduites et la piézométrie de référence. Sa représentativité reste donc étroitement liée à la qualité de ces paramètres, et nous avons vu que si dans

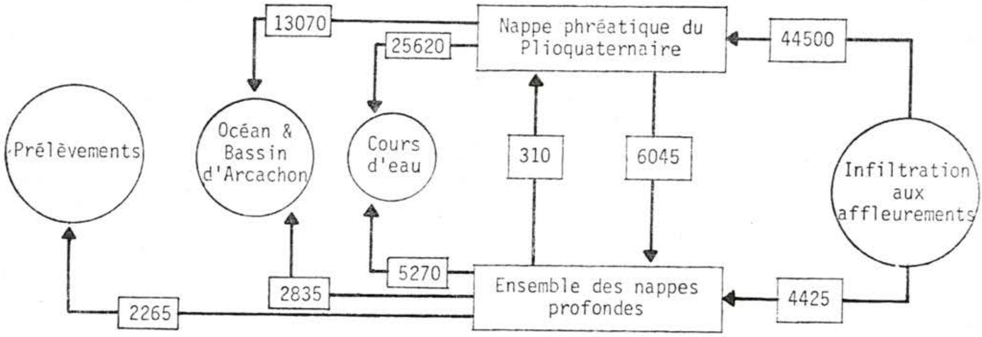

Fig. 1

le Nord du Bassin, elle était jugée satisfaisante pour les aquifères du Tertiaire, l'information sur le Bassin Sud-Aquitain restait à compléter et la construction du modèle peut guider la collecte future des données nécessaires.

Notre objectif était une modélisation en vue de permettre une gestion de la nappe des Sables Éocènes, principal aquifère du Bassin, et le maillage de cette couche permet une représentation aussi fidèle que possible du milieu physique simulé. Pour la représentation d'aquifères ayant moins d'incidence sur l'objectif du modèle, la nécessaire schématisation, imposée par l'échelle de l'étude, nous a parfois amenés à des simplifications que l'on justifie.

Dans le Plio-Quaternaire, les dimensions des mailles adoptées ne sont généralement pas à l'échelle de la représentation des nombreux ruisseaux qui drainent cette nappe. Pour en tenir compte, on peut soit diminuer le taux d'infiltration efficace affiché, soit augmenter les transmissivités, celles-ci représentant alors une moyenne sur la maille intégrant à la fois les plans d'eau et l'aquifère proprement dit. Il s'agit alors de « transmissivités apparentes » non transposables au terrain, et cet artifice, inhérent à une question d'échelle (il aurait fallu des mailles de 100 m de côté pour s'en dispenser entièrement !), diminue l'importance de la notion d'infiltration efficace sur la nappe libre. Cette schématisation a, par ailleurs, motivé le fait de négliger les prélèvements dans la nappe phréatique. Disséminés en une multitude de puits, ces débits n'avaient pu être recensés. Ils sont néanmoins implicitement simulés dans le modèle par ajustement de l'alimentation dont la valeur par maille représente la somme algébrique des entrées et des sorties.

Dans ce bilan du Plio-Quaternaire, un poste important est à retenir. Il s'agit du débit de percolation en profondeur, égal à 6 m^3/s. Une grande partie de ce débit sera drainée par les cours d'eau après avoir transité par le Miocène et l'Oligocène, et seulement 11 % de cette quantité (soit 665 l/s) parviendront jusqu'à l'Éocène, aquifère le plus exploité du Bassin.

Les prélèvements à l'Éocène sont concentrés dans la région de Bordeaux et des tests réalisés sur le modèle ont montré la faible importance de la drainance au toit de cet aquifère. Lors de simulations d'une surexploitation de la nappe, on s'aperçoit que la participation de la drainance par le haut est négligeable. Le débit prélevé provient essentiellement des limites d'alimentation Nord-Est, soit directement par infiltration sur les affleurements sableux, soit indirectement par drainance au mur de l'Éocène, mettant à contribution le Crétacé. C'est dans ce sens que la représentation de tous les aquifères dans un modèle unique prend son importance. En effet, le Crétacé ne saurait, comme producteur, concurrencer les excellentes perméabilités des sables Éocènes et son intégration permet surtout d'évaluer sa contribution dans la production Éocène.

Les affleurements du Crétacé sont extrêmement étendus et la faiblesse des débits d'alimentation de cette couche (= débit d'apport total diminué du drainage par les cours d'eau), calculés sur le modèle, amène à se poser la question de la signification de ces débits en comparaison des grandes surfaces d'affleurements perméables. Cela s'explique parce que la mise en production n'a pas encore affecté d'une manière sensible les limites du Bassin. De ce fait, les parties libres des couches crétacées se trouvent en situation

Tableau 1

Couches	Entrées (l/s)		Sorties (l/s)	
Plio-Quaternaire	Infiltration	44 500	Miocène	5 700
	Miocène	310	Océan Nord-Arcachon	2 780
	Total	44 810	Océan Sud-Arcachon	8 590
			Bassin d'Arcachon	1 700
			La Leyre	7 900
			Le Ciron	3 000
			La Garonne	4 270
			Gironde	3 250
			Midouze-Adour	7 200
			Éocène	15
			Oligocène	530
			Total	44 935
Miocène	Plio-Quaternaire	5 700	Prélèvements	235
	Oligocène	780	Bassin d'Arcachon	470
	Total	6 480	Plio-Quaternaire	310
			La Leyre	170
			Océan Nord-Arcachon	180
			Océan Sud-Arcachon	400
			Oligocène	2 270
			Ciron	570
			Midouze	1 865
			Total	6 470
Oligocène	Plio-Quaternaire	530	Prélèvements	470
	Miocène	2 270	Miocène	780
	Éocène	520	Éocène	650
	Total	3 320	Océan Nord-Arcachon	270
			Océan Sud-Arcachon	170
			Garonne	810
			Midouze	160
			Total	3 310
Libournais	Éocène	180	Prélèvements	100
			Dordogne	80
Éocène	Limite Nord	930	Prélèvements	1 150
	Limite Nord-Est	150	Libournais	180
	Limite Sud (Bigorre)	700	Oligocène	520
	Petites Pyrénées	45	Crétacé supérieur	570
	Albigeois-Castrais	20	Jurassique (Auvillar)	165
	Quercy	20	Océan	475
	Crétacé supérieur	1 540	Garonne	55
	Oligocène	650	Gironde	755
	Plio-Quaternaire	15	Dordogne et Isle	210
	Total	4 070	Total	4 080
Sommet du Crétacé supérieur	Affleurements NE	805	Prélèvements	115
	Base du Crétacé sup.	840	Base du Crétacé sup.	500
	Éocène	570	Éocène	1 400*
	Total	2 215	Océan	60
			Exutoire d'Audignon	75
			Dordogne et Isle	60
			Total	2 210

Tableau 1 (cont.)

Couches	Entrées (l/s)		Sorties (l/s)	
Base du Crétacé supérieur	Affleurements			
	Nord et Est	1 120	Prélèvements	195
	Villagrains-Landiras	200	Océan	650
	Limite Sud	310	Sommet Crétacé sup.	840
	Sommet Crétacé sup.	500	Éocène	140
	Crétacé inférieur	210	Charente, Seudre et Isle	460
	Total	2 340	Crétacé inférieur	70
			Total	2 355
Crétacé inférieur	Limite Sud	135	Bassin d'Arcachon	
	Jurassique	165	(percolation)	150**
	Crétacé supérieur	70	Océan (Nord)	10
	Total	370	Crétacé sup. (base)	210
			Total	370

* dont 500 sous la Gironde, 360 sous la Bassin d'Arcachon, et 200 dans le Double et le Landais.
** Sommet Crétacé supérieur + Éocène.

de trop-plein, et le débit circulant dans les nappes profondes se trouve limité à la « capacité d'ingestion » des affleurements, notion définie par le Professeur SCHOELLER [4]. Dans cette situation, une partie importante de la quantité infiltrée se trouve aussitôt reprise par le drainage des cours d'eau traversant ces affleurements.

Cependant, il en serait tout autrement si l'on simulait des prélèvements beaucoup plus importants, par exemple tels que les limites du Crétacé soient sollicitées d'une manière sensible. Dans ces conditions, il y aurait augmentation des gradients et des flux en provenance de ces limites. Actuellement, les lits des cours d'eau sont figurés par des mailles à potentiel constant pouvant drainer ou fournir un débit non limité. Au cours des simulations de surexploitation, on veille à ce que le débit circulant en profondeur ne dépasse pas la valeur de la « capacité d'infiltration » [4] des limites, soit le débit d'alimentation actuel auquel s'ajoute le débit réellement drainé par les cours d'eau. Au cas où cette valeur viendrait à être atteinte, on remplace simplement dans le modèle les mailles à potentiel constant considérées par des mailles à potentiel variable dans le temps, à l'intérieur desquelles on impose un flux égal à la capacité d'infiltration.

Conclusion

L'Aquitaine est sans nul doute le plus grand bassin sédimentaire français, et le plus riche en eau souterraine. Il y a aussi plus d'un siècle qu'est exploitée la nappe principale de cet empilement aquifère, les sables Éocènes.

Depuis une dizaine d'années, et sous l'impulsion initiale du Professeur SCHOELLER, on a tenté de maîtriser la prévision des écoulements souterrains dans le Bassin, aux fins de gestion de la ressource, grâce à l'utilisation des modèles mathématiques.

La présente synthèse en est l'aboutissement actuel. Si les problèmes de gestion se posent essentiellement pour l'Éocène, on s'est aperçu que l'intercommunication qui existe entre l'ensemble des nappes d'un bassin, vu à cette échelle, nécessite de représenter tel qu'il est cet ensemble, des nappes les plus superficielles aux nappes les plus profondes, même si elles sont inexploitées aujourd'hui.

Du point de vue de l'outil, le modèle mathématique a permis de se plier à cette exigence, principalement grâce à l'emploi de la technique des mailles variables.

Du point de vue des informations nécessaires pour construire un tel modèle, l'outil mathématique a, en fait, montré qu'il est en même temps un moyen d'approfondissement des connaissances: on y introduit ce que l'on sait, fut-ce même qualitatif, et on en vérifie la cohérence interne. L'outil permet ainsi de constamment valoriser les nouvelles informations recueillies.

Ainsi, le bilan des flux échangés entre les différents horizons perméables du bassin, présenté en Fig. 3, présente-t-il en quelque sorte la synthèse des connaissances acquises à ce jour sur le Bassin.

C'est semble-t-il, le premier essai quantitatif de schématisation des échanges dans un aussi important ensemble aquifère.

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HYDROGEOLOGICAL FEATURES OF THE PO VALLEY (NORTHERN ITALY)

PARTICULARITÉS HYDROGÉOLOGIQUES DE LA VALLÉE DE LA RIVIÈRE PO (ITALIE DU NORD)

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RÉSUMÉ

Cette note donne pour la première fois une synthèse des conditions hydrogéologiques de la Plaine du Pô (Italie du Nord): en effet les études entreprises jusqu'à présent ne concernaient que des secteurs très limités.

Un programme de recherche, organisé et financé par l'Institut de Recherche sur les Eaux du Conseil National des Recherches et realisé avec la collaboration des Instituts de Géologie de la région du Pô, a permis d'étudier les aspects hydrogéologiques de vastes unités territoriales de la Plaine.

Sur la base d'environ 10 000 coupes stratigraphiques de puits d'eau ainsi que des données hydrauliques et chimiques de 21 000 autres puits il a été possible d'arriver à une evaluation approximative satisfaisante des conditions structurelles des aquifères, qui se trouvent presque exclusivement dans les dépôts continentaux du Quaternaire (dépôts morainiques, fluviatiles et fluvio-glaciaires). La Plaine du Pô est en effet un grand bassin subsidant et qui a été caractérisé au Quaternaire par une grande vitesse de dépôt (épaisseur moyenne de 800 m); il est limité au Nord et à l'Ouest par les Alpes, au Sud par les Apennins et à l'Est par la Mer Adriatique.

Les conditions hydrogéologiques sont très variables en fonction de celles géologiques et morphologiques locales. Une première évaluation approchée permet de distinguer la Plaine au Sud des Alpes, formée par les dépôts morainiques d'âge différent des grands glaciers alpins, par les alluvions des affluents de rive gauche du Pô et par les fleuves de la Plaine de Venise; la zone axiale dans laquelle on retrouve surtout les sables et les argiles déposés par le Pô; et enfin la Plaine au Nord des Apennins formée par les conoïdes des affluents de rive droite du Pô et par les fleuves de la Plaine à l'Est de Bologne. Dans la Plaine Apenninique les mouvements néo-tectoniques ont eu une grande importance dans le contrôle du dépôt des alluvions et ont parfois déplacé les aquifères les plus superficiels (holocènes).

A l'heure actuelle on constate dans plusieures zones que l'équilibre hydrique a été gravement perturbé par les prélèvements excessifs effectués (même 10 puits par km^2 ont été relevés) ainsi que par d'autres facteurs concomitants tels que l'imperméabilisation des lits fluviaux due à la pollution chimique et organique, l'excavation du gravier des lits qui, baissant même jusqu'à plus de 10 m, drainent à présent la nappe phréatique, et enfin la couverture de grandes étendues en correspondance des zones d'alimentation des nappes, par suite du grand développement des centres urbains. Dans plusieures zones les prélèvements sont souvent aussi conditionnés par les pollutions industrielle et urbaine.

La présente note donne, entre autres pour certaines localités, des indications concernant la perméabilité des aquifères, l'alimentation des nappes, la vitesse de l'écoulement souterrain, le chimisme et la base des eaux douces. En effet, dans certaines parties de la Plaine et surtout au Nord des Apennins, l'approche à la surface des eaux saumâtres profondes constitue une limite pour l'approvisionnement en eau.

Introduction

The water supply for urban and industrial areas in the Po Valley comes almost exclusively from underground sources representing both unconfined and confined aquifers in the thick Quaternary cover of the Valley itself.

In recent years the creation and development of industrial centres, together with the intense urbanization of the Po area, have excessively increased the water demand. This has led to a striking withdrawal of groundwater level.

The industrialization and urbanization have given origin to the worrying phenomenon of the increasing underground water pollution, too.

The Water Research Institute of the National Research Council (C.N.R.), on the proposal of prof. M. MANFREDINI, member of its Scientific Council, put forward in 1971 a multi-annual programme of hydrogeological research in the area including the Po river Valley (s.s.), its north-eastern extension, from the river Adige to the river Isonzo, and its south-eastern extension, that is the Romagna Plain crossed by the river Reno, dealing with the problem in a co-ordinated manner*.

The Institutes of Geology of the Universities of Turin, Pavia, Milan, Modena, Parma, Bologna, Ferrara, Padua and Trieste, with the collaboration of AGIP and the Large Masses Dynamics Laboratory of the C.N.R., all take part in the carrying out of the programme.

The principal aims of the research programme are:

a) The determination of the principal hydrogeological characteristics of the area, and the evaluation of the deep-aquifer water resources.
b) The allotment of new technical elements, useful in programming water supply systems for the Po Region, to the Bodies responsible for water management.
c) The setting up of research techniques and work plans of a scientific and technical nature to be used in similar investigations.

The tasks of the initial phase of this programme are to make, first of all, a systematic collection of all litho-stratigraphical and hydrogeological data recorded from water wells, and to correlate these data on a local scale in order to obtain a first synthesis for the entire Po Valley.

In this report the results of a first general synthesis carried out after the first four years' work** are dealt with. The continuation of the research will make possible to attain a greater precision and refinement of the knowledge of the hydrogeology of the Po Valley.

Geological and hydrogeological background

Geologically speaking the term "Po Valley" applies to the region covered by post-Calabrian deposits, represented mainly by alluvial sediments of the Po and other rivers. Altogether, the area concerned covers about 46,000 km^2, with a length (E—W) of 400 km and a maximum breadth of 120 km (Fig. 1).

* All throughout the paper the term of Po Valley is used in a wider sense of meaning, including also the Venetian and Romagna Plains.

** Detailed information on the results attained to in this phase of the research programme on the hydrogeology of the Po Valley are presented in reference [3].

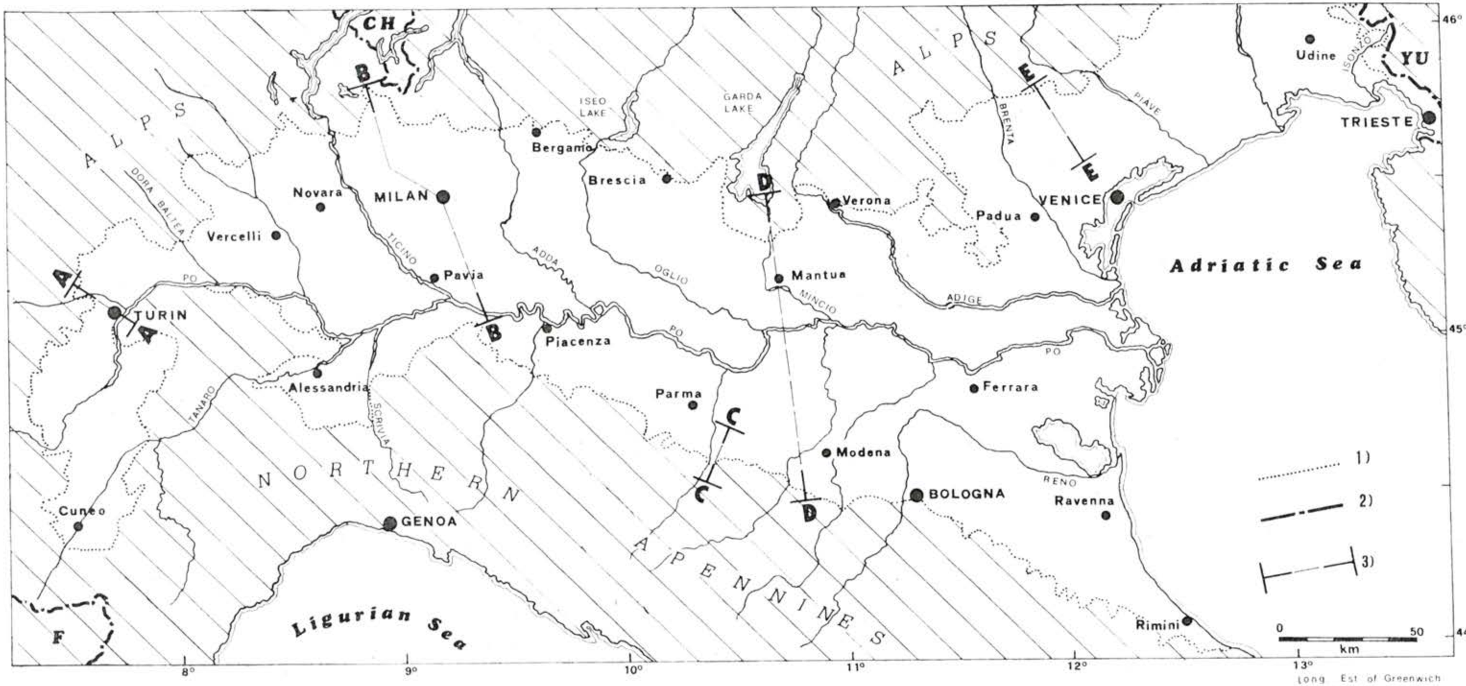

Fig. 1. Geographic outline of the Po River Valley: *1)* boundaries of the Po Valley; *2)* State frontiers; *3)* traces of the geological sections

The Po Valley constitutes 71% of the level ground in Italy; the major Italian industrial and agricultural activities and the most urbanized areas are concentrated here.

From a geological point of view, the Po Valley has long since represented a subsident region, at least for all the Neogene, and an accumulation site of thick sedimentary series; also today, at least in the central and eastern part, the region is still subject to notable subsidence. The available information on the Po Valley enables to estimate the volume of the Quaternary sediments accumulated in the Po Valley to 40,000 km³, corresponding to an average thickness of about 850 m. The subsidence, already very noticeable in the Pliocene (average thickness 900 m) was notably increased in the Quaternary. Thus the Po Valley, constitutes a definite geographical and geological entity, as a great subsident basin, in which the Quaternary sediments reach the maximum thickness in correspondence with the axis of the Valley (over 1000 m), increasing from the west (western Alps) towards the east (Adriatic Sea), where the top of the Pliocene was found at depths of over 2000 m.

The stratigraphy and the structural behaviour of the pre-Quaternary substratum and, more generally, the geological setting of the Po Valley were the subject of detailed studies for petroleum search by E.N.I. [1, 2].

From the hydrogeological point of view, the main interest is in the Quaternary deposits and to a lesser extent, especially along the edge of the Apennines, in the Pliocene sequence.

The post-Pliocene sediments are continental, with gravels and sands in the upper part, and marine, with clays and sands, in the lower part. The marine Quaternary (Calabrian) constitutes the continuation of the sedimentary cycle that began in the Pliocene; therefore usually there is no discordance between Pliocene and Calabrian.

The lacustrine facies (Villafranchian) stratigraphically includes the Upper Pliocene and the Lower Pleistocene. It is cropping out in some areas (e.g. along the pre-Alpine margin of Piedmont), while in other areas it is buried by more recent fluvio-glacial and glacial sediments (e.g. Milan); in other areas it was not yet recognized.

It is possible to single out three large units, with rather different hydrogeological characteristics, arranged according to the principal axis of the Valley:

1. The piedmont band and the high Alpine Plain, where the Mesozoic and Cenozoic substratum outcrops from the Quaternary continental deposits of the great glacial and inter-glacial cycles.

2. The central area along the river Po, made up of alluvial sediments carried down by the main water stream and its tributaries, especially the Alpine ones. In this region, as in the Venetian Plain and that at the foot of the Apennines to the east of Modena, there are local intercalations of marine sediments, also inside the Holocene alluvial cover.

3. The band at the foot of the Appennines where, mainly at the eastern and western limits, together with the alluvial deposits of the Calabrian and the Upper Pliocene are also of practical importance.

In this sector the areas of a greater interest are the large fans situated near the outlet into the Plain of the Apenninic rivers and torrents. Hydrochemical problems connected with the presence of salt waters and, more generally, with mineralized strata become prevailing here.

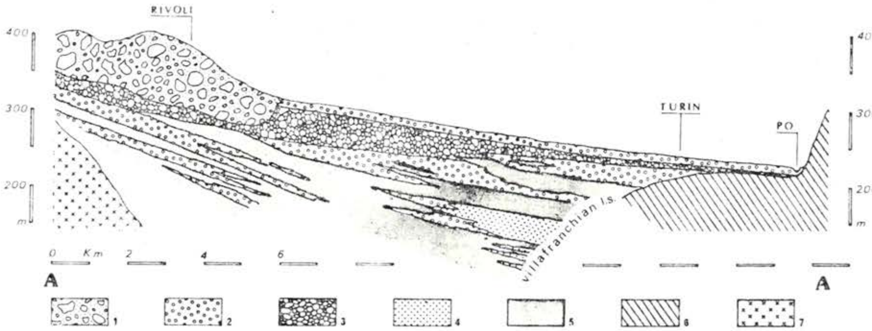

Fig. 2. Geological section A—A: *1)* moraine deposits; *2)* fluvial and fluvio-glacial gravels; *3)* cemented gravels of the inter-glacial Günz-Mindel ("Ceppo" *Auct.*); *4)* fluvial and fluvio-glacial sands; *5)* fluvial and lacustrine clays and silts; *6)* Tertiary sediments; *7)* crystalline base

From the hydrogeological point of view the fluvioglacial and glacial deposits, made up of loose gravelly-sandy sediments with small to fine-grained, less permeable fractions, are of considerable practical importance. Thus, for example, the inter-glacial Gunz-Mindel ("Ceppo Lombardo") deposits, made up of polygenic fluvial conglomerates, more or less cemented and somewhere turned to loose gravel, with Villafranchian fluvio-lacustrine deposits at the bottom (moraine amphitheatre of Rivoli in Piedmont) or other bottom morainic and glacial-lacustrine deposits in a muddy-clayey facies (North and East Lombardy), contain notable water reserves.

The study commenced some years ago with the collaboration of a large group of researchers being carried out on an experimental basis, regarding mainly the hydrogeological characteristics of the aquifers, and by means of the collection and analysis of existing data related to stratigraphic and hydraulic information obtained when drilling water wells. Up to now, this last phase of the investigations has allowed to plot over 10,000 stratigraphic columns and data from other 21,000 water wells have been collected.

The research programme is devoted to determine the stratigraphic and granulometric characteristics of the Quaternary beds of the Plain and, on the other hand, the hydrogeological situation is also to be monitored.

The western sector of the Po Valley: the Turin Plain

1. *Geological and hydrogeological sketch*

The Turin Plain that makes up the highest extreme western part of the Po Valley forms a basin stretching from the south towards the north (from Cuneo to Turin) at right angles to the general trend of the Po Valley line. In the west it is bordered by the crystalline rocks of the Alpine chain and, to the east by the Tertiary rocks of the "Langhe", and of the Monferrato-Hills of Turin. To the north the Turin basin merges with the Po Valley through a "pass" where the Tertiary Hill of Turin approaches the Alpine chain. Despite its relatively small dimension and its marginal position with respect to the Po Valley, the Turin Plain was affected by a notable subsidence. The Quaternary alluvial cover reaches a thickness of about 600 metres in the middle of the area, while the Tertiary deposits and the crystalline basement, respectively, become abruptly deeper-situated.

Also from a hydrogeological point of view, the Turin Plain represents a well-defined basin, a tributary to the actual Po basin itself, with particular characteristics for the natural recharge and main direction of the underground flow, which passes from the south northwardly. All the surface and underground waters of this basin are forced to cross through the Turin "pass", which therefore represents an ideal section for the water balance. The river Tanaro, however, is an exception, for in recent times it has changed its course towards the east cutting through the Tertiary hills and flowing into the river Po near Alessandria.

With regard to the composition of the alluvial deposits, it may be outlined as follows: coarse materials with good permeability (gravel and/or sand) for the deposits coming from the Alpine catchment areas; fine materials (fine sands, silts and clays) for those coming from the breakdown of the Tertiary formations that form the hills to the east of the basin. For this reason, the Plain near the Alpine margin is made up mostly of coarse deposits forming a single, well-developed, free-surface aquifer, which comes under pressure only locally due to limited confining beds. In the axial and eastern area this near-homogeneous alluvial cover tends to be sub-divided into a multi-aquifer complex with the presence of clayey intercalations.

2. *Characteristics of the aquifers*

The aquifers in the Turin basin store both free-surface water and water under pressure. The production is mainly connected with the Quaternary deposits. The specific discharge reaches $20-30$ l/s$\times$m near the Alpine margin, upon better conditions of permeability.

Generally the wells have been completed at a depth smaller than 100 m below the ground level. Near the Alpine margin as well as near Turin the wells easily reach the Villafranchian deposits or even those of the Tertiary, due to the modest thickness of the alluvial cover (a few dozen metres). On the eastern hilly margin, towards Cuneo, the wells go down to a depth of $150-200$ m in order to tap the sandy-gravelly and sandy intercalations of the marine clays of the Pliocene: the specific discharges are of about $3-5$ l/s$\times$m.

The aquifers are recharged from the Alpine catchment areas and, for smaller areas and for the lower depths, from those of the Tertiary hills.

The streamflow of the Alpine rivers reaches the Valley without any appreciable loss in the mountainous part of the catchment areas, since these are constituted by crystalline rocks with no high permeability.

In the central part of the Valley, which is constituted by a multi-aquifer complex, it proves more difficult to establish the recharge areas because of the difficulty in following the behaviour of the underground basins and because of the convergence and mixing of the water inflows coming from different catchment areas.

The contribution of recharge from the small catchment areas of the hilly region situated easterly appears to be rather scarce.

The aquiferous beds are sometimes characterized by not common mineralizations, with the presence of evaporites in the Tertiary series (Messinian).

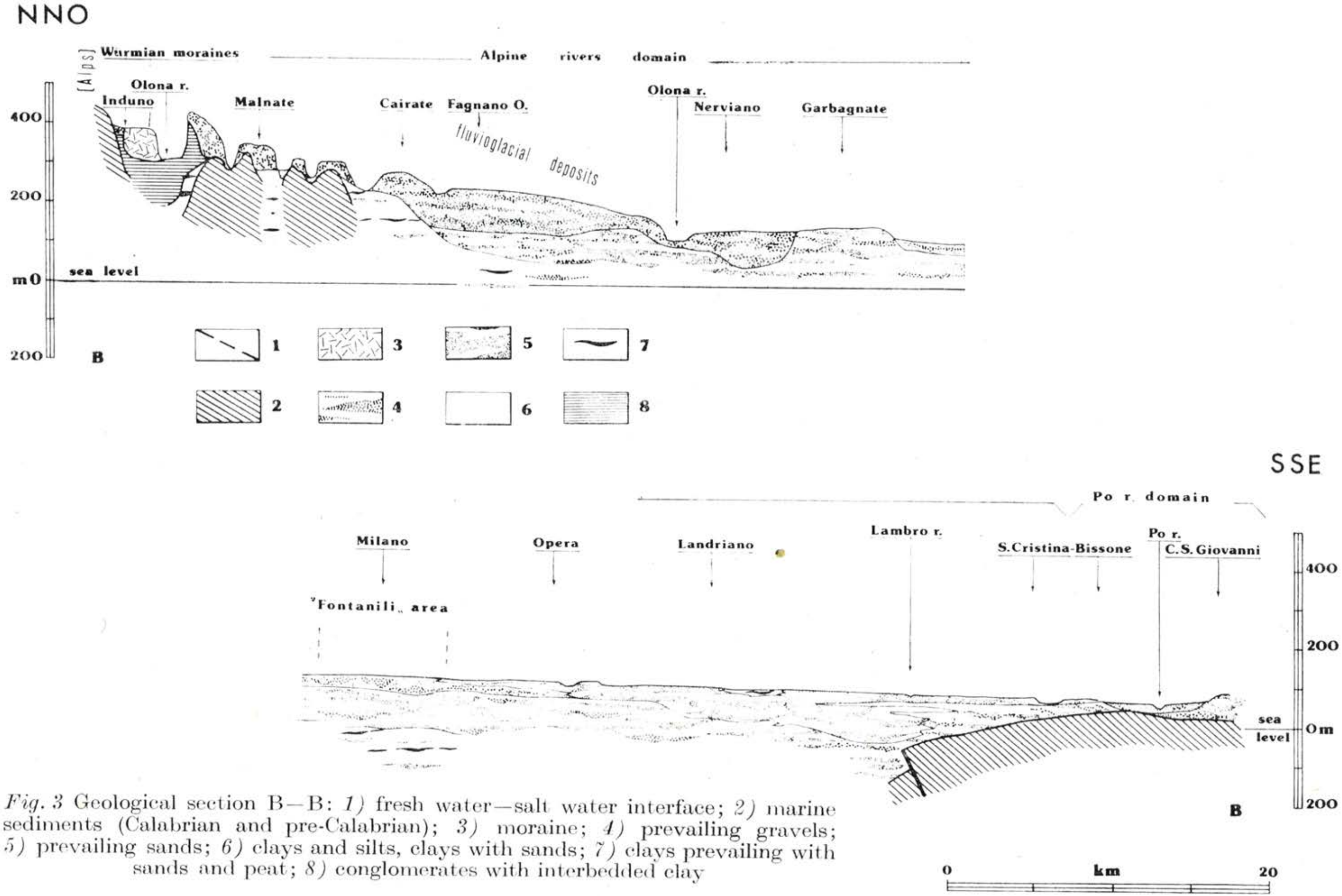

Fig. 3 Geological section B—B: *1)* fresh water—salt water interface; *2)* marine sediments (Calabrian and pre-Calabrian); *3)* moraine; *4)* prevailing gravels; *5)* prevailing sands; *6)* clays and silts, clays with sands; *7)* clays prevailing with sands and peat; *8)* conglomerates with interbedded clay

The northern sector of the Po Valley: from the River Dora Riparia to the River Mincio

1. *Geological and hydrogeological sketch*

The Plain of the Alpine rivers, tributaries on the left of the river Po (Dora Riparia, Dora Baltea, Sesia, Ticino, Lambro, Adda, Oglio and Mincio), corresponds to the northern margin of the Po Valley.

In the western part between the rivers Dora Riparia and Sesia, along the river Po, coarse fluviatile deposits prevail (sands and gravels), while to the east of the river Sesia finer sediments are predominant.

The total thickness of the concerned alluvial deposits is notable (about 600 m near Vercelli) and, as a general rule, there is a change from coarse sediments forming a single aquifer in the upper Plain, to more marked alternations of permeable coarse deposits and of impermeable clayey horizons between the rivers Dora Riparia and Ticino.

In the western part the base of the fluvio-glacial deposits is sometimes represented by the "Ceppo" (a polygenic conglomerate, with a variable degree of cementation, locally stratified), which has, at its bottom, coarse fluvial or fluvio-lacustrine deposits (sands and gravels with clayey intercalations) of the Upper Villafranchian, passing downwards into clays with thin, continuous intercalation of fine lacustrine sands of the Lower Villafranchian. The pre-Quaternary formations, generally Piacenzian clays and Astian sands, occur near Turin and the Alpine margin.

In the central part of the northern sector of the Valley (from the Ticino to the Adda) the impermeable marine substratum is covered by alluvial deposits, which at times reach 300 m. It is possible to distinguish areas more or less favourable to the underground circulation and accumulation of water in deposits with considerably varying thickness and permeability.

The water-bearing materials of the first aquifer are, for the most part, Würmian fluvial and fluvio-glacial deposits. The thickness increases gradually from the upper Plain i.e. from the front of the moraine circles towards the middle Plain, and comes to considerable values (over 100 m) near the river Ticino. This material, which has an average thickness of 40 m near Milan, is constituted by gravels and sands in the north, while to the south it tends to pass gradually to sands; downstream from the line of the "fontanili"*, towards the lower Plain, it becomes fully clayey in the first 20 m.

The aquifers of medium depth correspond mainly to the Mindel and Riss deposits which occupy the surface in extensive areas of the upper Plain, stretching from the north towards the south and are more elevated than the Würmian deposits, while in the central and lower Plain they are to be found everywhere at depths varying from a few dozen metres to about a hundred metres. These deposits, slightly sloping towards the Po, are constituted by alternations of sands and gravels with thick and extensive clayey lenses. They contain confined aquifers fed by the more permeable materials of the upper Plain. The total thickness of these aquifers is generally from 10 m to a maximum of 25 m (near Brescia).

* "Fontanili" is the local name which is given in the Po Valley to the partly natural, partly artificial emergences of the phreatic water level. "Fontanili" occur on a continuous band 20 km wide, in average, from Piedmont to Veneto.

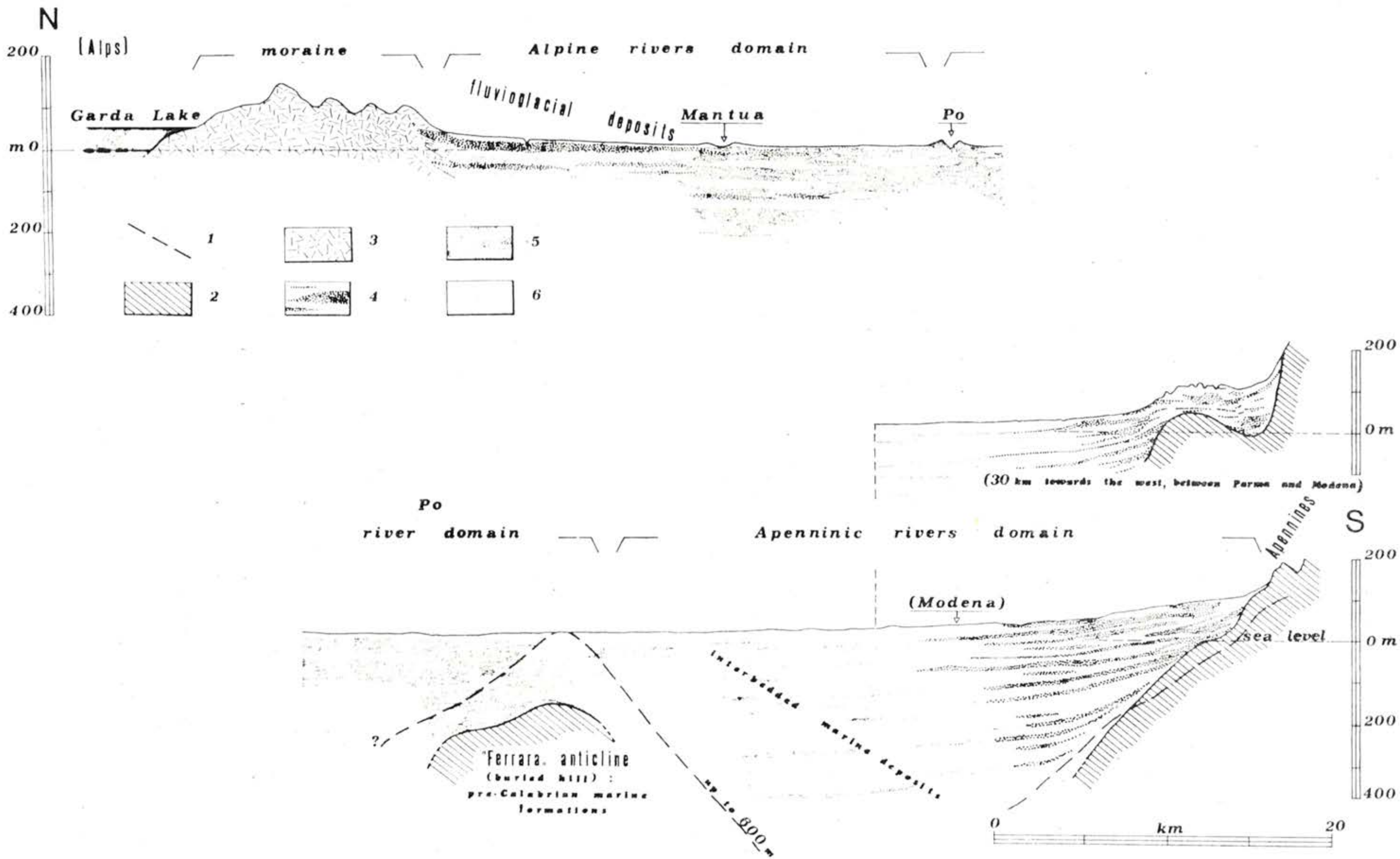

Fig. 4 Geological sections C—C and D—D: *1)* salt water—fresh water interface; *2)* marine formations (Calabrian and pre-Calabrian); *3)* moraine deposits; *4)* gravels and gravels with fluvial and fluvio-glacial sands; *5)* fluvial and fluvio-glacial sands, only locally of marine origin; *6)* fluvial and lacustrine clays, locally of marine origin (Holocene)

The deep aquifers, composed of sandy-pebbly layers contained in the peaty grey-coloured clays of the Villafranchian and in other Calabrian interbedded sands, have never been explored in a systematic way, with the exception of the Milan area. The Villafranchian aquifers are of special importance by their total thickness (5 – 15 metres) and notable areal extent. Thirty kilometres from Milan towards NW they have been found at a depth of 170 m and of 120 m to the east of Milan. South of the "fontanili" line they are not easily recognizable.

The aquifers contained in the Quaternary marine deposits are mainly sandy. It has been noted that these aquiferous beds are situated as deeply as below 500 m with a total thickness of over 30 m. It should be pointed out that near Milan the bottom salt waters are sometimes encountered at a depth of about 700 m.

On the southern margin of the sector between the Dora Riparia and the Mincio, especially to the east of the Ticino, there are clayey and sandy alluvial sediments with or without lenses of peat, which represent the finer materials of the Alpine rivers and of the Po. The lenses are greatly extended in an axial direction (west to east), in cases to dozens of kilometres; in some areas, especially near the river Mincio, till now they have been observed in water wells in a thickness of about 250 m.

Towards the west the sandy horizons may exceed 50% of the total thickness and generally form four principal aquifers: from Cremona to the river Mincio the thickness of the Quaternary alluvial cover reaches and even exceeds 300 m.

2. *Characteristics of the aquifers*

Near Milan, in the central area of the valley sector under consideration, the density of population is among the highest in Italy (over 200 inhabitants per km²). The numerous urban centres and, in particular, Milan (2,700,000 inhabitants in the whole province in 1961) are among the Italian towns with the greatest industrial activity. The number of the wells is therefore very high. In the course of the survey carried out for the study of the hydrogeology of the Po Valley, data of over 11,000 wells have been collected; 5000 out of these have contributed to the stratigraphical recording, too.

In the central part wells sunk in the first aquifer show a specific discharge of over 20 $l/s \times m$ and transmissivity up to 500 cm^2/s (river Ticino and Milan area). Elsewhere the transmissivity decreases by 2 to 5 times due to the reduced thickness of the aquifer; the yield, still being very high, is lower than that of the previous ones (from 5 to 10 $l/s \times m$).

Also in the central part of the Plain, the aquifers at an intermediate depth (Mindel and Riss) are not proved to be very good because of their smaller thickness and lower permeability. In fact, proceeding to the river Adda, the lenses of conglomerate become increasingly frequent when substituting gravels. In areas where these aquifers are developed, either because they are near the surface or because the first aquifer is polluted or exhausted, the specific discharge is at the most about 4 – 5 $l/s \times m$. The permeability is $10^{-2} - 10^{-3}$ cm/s.

The sandy aquifers of the Villafranchian give a similar but more varied yield; at the same time the permeability coefficient has values close to those of the previous ones.

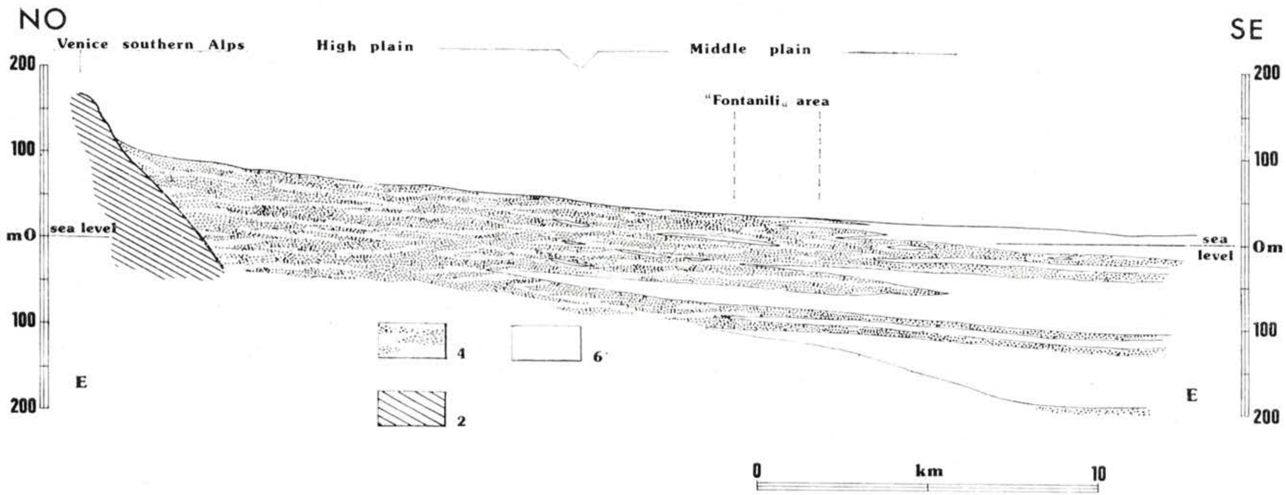

Fig. 5. Geological section E—E: *2)* marine sediments (pre-Calabrian); *4)* prevailing gravels; *6)* silty clays, sandy clays

With regard to the aquifers in the sandy marine deposits, the pumping tests show yields which are still lower. Here the specific discharge of the single aquifer very rarely exceeds one $l/s \times m$. In the middle Plain the water wells rarely reach the deeper levels of the continental series, unless the rocky substratum approaches the surface.

The most productive area is the Milan area, between the Ticino and the Adda (specific discharge of over 20 $l/s \times m$), while the yield is low or mediocre in the areas of the fluvio-glacial terraces (less than 5 $l/s \times m$) and in those to the east of the Adda, in which conglomerates are abundant. Intermediate values are encountered in areas where the upper aquifer is present, but it is not very thick (between Milan and the river Adda in the upper Plain).

In the lower Plain, towards the river Po, between the Adda and the Mincio, the specific discharges are fairly high (from 10 to 20 $l/s \times m$); the degree of permeability of the aquifers is nowhere very high, being around 10^{-2}–10^{-1} cm/s, however, their thickness and transmissivity are fairly high. In this area, where all the aquifers are under pressure, almost all the wells, in places 200 m deep, tap more than one aquifer with good yielding capacity. However, the excessive content of iron ions (up to 4.25 mg/l) is limiting the utilization of the waters stored in this band of the Plain.

Finally, it should be noted that in the region SE of Pavia (San Colombano hill) up to the river Adda, the yields are particularly scarce due to the presence of uplifted pre-Quaternary impermeable marine formations, against which even the deep aquifers of the surrounding areas are terminated.

For the same structural reason, here the interface between fresh waters and salt waters comes considerably closer to the ground level; likewise a direct communication between the phreatic aquifer and that under pressure takes place so that the former feeds the latter.

A widespread problem that has to be dealt with in almost all this sector of the Plain has been aroused by the strong alteration of the hydrological balance due to the increased lowering of the water level. In fact the recharge of the aquifers is considerably deficient compared with the quantities of water drawn up. Recharge waters are mainly secured by the infiltration from abundant rainfalls, never less than 1000 mm/year, through the highly permeable surface deposits in the upper and central Plain. Between the rivers Adda and Ticino, downstream from the moraine amphitheatres, water losses through streambeds acquire importance in recharging groundwaters.

In time the infiltration is steadily weakening for three principal reasons: the enlargement of the urban areas (doubled from 1950 till today) with the resulting impermeabilization of the ground; the reduction of the irrigated areas and the impermeabilization of the streambeds due to the accumulation of chemical substances released by industrial plants; the diversion of the fluvial waters before their outlet into the Plain.

The growth in production for water supply constitutes the second cause of the alteration of the hydrological balance: at present, in the city of Milan alone, the supply rate is 60 m^3/s, with loss by withdrawal of 150,000 m^3/year per hectare; in the central areas and in the industrial suburbs, the loss by withdrawal is about 5000 m^3/year per hectare. However it must be considered that the average loss in water by withdrawals per hectare is very high in Milan and the neighbouring centres, being mostly between 10,000 and

50,000 m^3/years per hectare. Similar values are displayed by all large industrial centres (Pavia, Cremona, Mantova).

In the Milan area the recharge of the aquifers cannot exceed 200 mm/year, equal to 2000 m^3/year per hectare, that is a value from 5 to 25 times inferior to the withdrawal. From this deficit in the hydrological balance derives the continuous and very rapid depletion of the piezometric surface in all the urban centres, at the rate of about 1 m/year. In the areas farthest from the urban centres the depletion is slower (10 cm/year), due both to the recharges from irrigation and the smaller production.

This progressive decline of the piezometric level has created vast areas of depression around the areas of highest withdrawal. E.g. near Milan the pressure level of 1950 has fallen by about 40 m till now, for the deep aquifers, and the area concerned takes up about 600 km^2.

The northeastern sector of the Po Valley: the Venetian Plain

1. *Geological and hydrogeological sketch*

The Venetian Plain is delimited by the Pre-Alps in the north, by the Lessini Mts and the Berici and Euganean Hills in the west, and by the river Adige and the Adriatic Lagoons in the south. The territory has its own distinct geographic individuality with peculiar stratigraphic and hydrogeological features in the Po Valley.

The Quaternary cover rests on a base mainly of Tertiary age. Its thickness, variable from zone to zone, reaches a maximum of 1000 m along the coastal strip (Venice Lagoon), decreasing appreciably to the NE down to values of about 200—300 m in the Lagoon area near the mouth of the river Piave.

The loose materials which constitute the cover are of fluvio-glacial origin along the upper part of the Plain, near the Pre-Alps, where the Quaternary is known to be as thick as 300—400 m. In the middle part, on the other hand, terrestrial deposits alternate with marine sediments in a thickness of several hundred metres. On the lower part, along the coast, the sediments are predominantly marine below thin continental deposits.

On the basis of stratigraphic data referring mainly to the first 200—300 m of depth, it has been possible to build up a fairly indicative picture of the upper part of the Quaternary cover which is currently subject to search for underground water. The following observations refer to depths of several hundred metres.

There are some stratigraphic and hydrogeological features that may determine, although only tentatively, different bands of the Plain. These bands with homogeneous characteristics follow each other in a succession from the north, from the Pre-Alps to the Adriatic Sea. They stretch almost over the entire Venetian Plain and run in a subparallel direction to the margin of the mountains and to the present coastline, perpendicularly to the water-courses.

In the upper Plain against the Pre-Alps where the rivers come out of the mountain basins, there is a strip about 5 to 20 km wide made up of gravelly alluvial deposits practically undifferentiated down to the rock substratum. The gravel thickness is about 300—400 m. The gravelly deposits always contain about 10 – 30% of sand as well as a fairly high percentage of pebbles. Somewhere thin lenticular mud-clay intercalations may be found, while in certain areas conglomerate horizons deriving from carbonatic cementation of the gravel are quite frequent.

Southwards from the zone described above, gravel gradually decreases in quantity and even runs out entirely after about ten kilometres, except for a few rare thin layers at great depth which extend in some cases under the Adriatic lagoons.

Lastly, the southern band 10 – 20 km wide of the territory examined running along the Adriatic coast, seems to be characterized by alternations of thick silty-clayey beds with rather thin sandy layers often fine-grained. Good-permeability sandbeds are not common, while gravel beds are extremely rare, occurring mainly to the east of the Tagliamento.

The cover of the upper Plain and most of that of the central portion of the territory is made up of a series of large contiguous, interpenetrating and partially overlapping alluvial fans which were deposited by large-scale floods of the rivers at the zones of their outlet into the Plain.

Along the piedmont strip the undifferentiated gravel mass extends practically without interruption in an E – W direction. This is due basically to the frequent changes in flow direction undergone by the Venetian rivers during the Quaternary as they had overflowed and deposited their gravelly sediments over wide areas.

2. *The characteristics of the aquifers*

Along the piedmont strip in the undifferentiated gravel cover a single phreatic aquifer occurs, whose lateral continuity is determined by the direct contact between coarse-grained permeable materials ($K = 10^{-1} - 10^{-2}$ cm/s) of the various alluvial fans. This is a thick, rich aquifer which is exploited intensively for the supply of important aqueducts. The specific discharge is estimated at approximately 10 – 30 l/s × m. The depth of the free surface usually attains its highest values closest to the Pre-Alpine mountains where it occurs at 50 – 150 m below ground level. The flow velocity, as estimated in several points by means of tracers, is rather high, i.e. in the order of up to 100 m/day. Towards the south the aquifer gradually approaches ground level until, after a path of 10 – 15 km, it comes to the surface spontaneously in the lowest points along a practically continuous strip lying between the upper and the central Plain. It is the line of "fontanili", characterized by the presence of numerous plain springs, which gives rise to a series of spring-fed watercourses, the largest of which, the Sile, has an average discharge of about 50 m^3/s.

In the middle Plain, where the various gravel fans have broken up into overlapping stratiform digitations of gradually diminishing thickness, usually immersed in fine-grained, very impermeable soils, a multiaquifer system is found, which consists of a relatively shallow phreatic aquifer (not always present, however) and of several confined aquifers, some of which are of artesian character. In some areas, for example in the Treviso and Vicenza areas where thick gravel beds occur as associated with larger and more advanced fans, the confined aquifer appears to be particularly rich (with specific discharge of up to even 5 – 10 l/s × m).

This hydrogeological system, as it is based on the alluvial fans of the Pre-Alpine rivers, is basically fed by the extensive water dispersion processes that occur in the streambeds at the outlets of the rivers into the Plain. Downstream from the outlet these rivers have carved their present beds into the permeable gravel cover deposited in older times. The beds of the Astico, Brenta, Piave, Cellina, Meduna and Tagliamento rivers are dispersing almost entirely their water of dry-season flow, thus they are dry for long periods during the year. At the time of high tide, the quantity of water discharged by the river into aquifers can take on enormous proportions, of the order of several hundred m^3/s. Experiments carried out on the Brenta and Piave rivers have revealed the occurrence of water losses to the aquifers of about one third of their total discharge estimated at $30-60$ m^3/s. The length of the river stretch along which dispersion may take place is about 10 km starting from the outlet into the Plain.

The close relationships between rivers and underground water circulation, including the confined aquifers located in the gravel intercalations downstream from the river stretches where dispersion occurs, are also demonstrated by the considerable similarities found between river regime and that of the aquifers. The fluctuations are shifted by about one month.

In some areas of the southern part of the Plain, there is a number of confined aquifers located in predominantly sandy materials; they have relatively low specific discharges (several $l/s \times m$). The recharge of many of these aquifers does not appear to be directly connected with the fluvial dispersion. The sand horizons almost never display any direct connection with the gravel structures of the alluvial fans. The natural recharge of these aquifers, which is certainly a complex process, seems rather to depend on very slow interaction processes between sand beds and surrounding silty-clayey sediments, as would appear to be indicated by the phenomena of slow soil subsidence along the edge of the Venetian Lagoons. In the lower part of the upper Plain a large contribution to the recharge of the aquifers is made by irrigation.

The southern section of the Po Valley: from the River Tanaro to the Adriatic Sea

1. *Geological and hydrogeological sketch*

The southern or Apenninic sector of the Po Valley is made up of the alluvial deposits of the right-bank tributaries of the river Po (Tanaro, Scrivia, Trebbia, Taro, Parma, Enza, Secchia and Panaro) and of the rivers of the Romagna Plain, east of Bologna (Reno, Ronco, Savio, Marecchia). The outlets into the Plain occur at an altitude which varies gradually from 150 m a.s.l. in the westernmost area to 30 m at Rimini on the Adriatic coast.

Lithostratigraphic evidence has been gathered for about 3600 wells, while other data (static levels, depths, flow-rates and chemical characteristics) have been collected for a further 15,000, although the total number of existing wells is estimated in 90,000.

The stratigraphic situation, highly differentiated by different flooding types and neotectonic movements, may be outlined as follows:

— The aquiferous alluvial cover becomes rapidly thicker in a south-northward direction, while the grain size of the permeable materials decreases in the same direction.

— In the Apenninic valleys, in the vicinity of the outlets onto the Plain, the alluvial cover (gravels) varies in thickness between 3 and 20 m and it lies on a pre-Holocene substratum usually made up of impermeable marine formations (Calabrian and Pliocene).

— In the upper Plain, the thickness of the alluvial deposits (gravel-sand alternated with mud-clay) attains 50 – 100 m as close as 2 km to the hill margin; further to the north the depth increases less quickly and at the extremity of the gravel fans laid down by the main streams (i.e. at a distance of 10 – 75 km from the hills) it reaches 400 – 500 m at least east of Parma; SE of the Reno the gravel fans are less extensive and near Rimini they dip under the Adriatic Sea. The number of gravel beds varies from 3 to 10 and their total thickness averages less than 30%. The substratum is made up of marine formations, mostly impermeable and generally dated Calabrian-Pliocene, although older sediments can be found locally (at Parma). The substratum is extensively affected by folds and faults.

— The Plain North of the Via Emilia (on the Parma – Rimini line and to the east of Parma), is mostly characterized by impermeable sediments with occasional, discontinuous sand layers which represent the finest and most distal suspended sediments of the Apennine rivers. The thickness of the alluvial cover is over 700 m.

— North of the latter part of the Plain, the Po River alluvial deposits occur; they consist of thick sand beds (up to 40 m) intercalated with clay.

— The plain of Alessandria has features that distinguish it from the rest of the Plain lying at the foot of the Apennines. A buried positive structure (Tortona – Valenza – Po Ridge) consisting of Tertiary marine formations, gives rise to a transverse threshold separating the Plain of Alessandria from the proper Po Valley thus dividing it into two parts: a southern part by the gravel fans of the Tanaro and Scrivia rivers, and the second, situated north of the former, by sediments of the River Po (sand and clay).

The configuration of the aquifers is almost in all places affected by the deep structures revealed by geophysical prospecting and oil research drilling; the Plain of the Apennine rivers has a substratum consisting of a series of folds and fold-faults ("Apenninic folds") inclined towards the north and with an axial trend running SE – NW affected by rather marked neotectonic movements in the upper Plain, near the boundary of the Apennines.

2. *Aquifer characteristics*

The irregularities in lithological setting governed by the torrential nature and the consequent rapid changes of course of the Apennine rivers, brings about an equally irregular pattern in the aquifers. So it is often difficult to stratigraphically correlate even wells put down closely to each other.

This situation, quite marked on the Apennine boundary of the Plain, tends to fade out in an eastward direction from Parma as getting close to the Po. Here sandbanks form hydrogeological units with an area of several tens of km^2 stretching out in an E – W direction.

As a rule the gravel fans area is characterized by a fairly high quantity of water and by aquifers which, in relation to the lenticular shape of the deposits, at least up to 20 years ago, displayed a high pressure head (up to 7 – 8 m over ground level*).

Beside the gravel fans of the main rivers there are fans composed largely of clay, with rare sand and gravel lenses, where water is less available. In the upper Plain, at the foot of the Apennines, there are a few sand or conglomerate intercalations (mostly with calcareous cement) contained in the marine clays of the Quaternary and the Pliocene.

The zone of the central Plain, east of Parma, is very poor in underground waters; to the north of it the aquifers begin to come under pressure as linked to a water circulation system of Alpine origin (the River Po and its left-bank tributaries).

As a general rule, water potential on the Plain at the foot of the Apennines is determined not only by the thickness of the water-bearing strata themselves but also by the distribution of salt waters. This usually follows the behaviour of the marine substratum even though the two almost never coincide. Occasionally, especially near Modena, the deep salt water near the Apennine boundary remains at an average depth of 100 m (even though there are also cases on record of superficial salt water) partly of the salt-bromium-iodine type, attaining a maximum depth of 600 – 650 m in the central Plain zone. However, in the central and lower parts of the Plain, between Modena and Ferrara, the buried positive structure ("Ferrara anticline") leads the salt water to the surface.

Exploitation of the aquifers is also limited by water hardness (up to 60 °F), as the gravels and sands are mainly calcareous and contain practically no dolomites, or by a too high iron content (more than 4 mg/l).

The exploitation carried out in an irrational manner, as in other sectors of the Po Valley, owing to the lack of public bodies responsible for its protection and control, satisfied all demand for water till not long ago. At the present time the trends in consumption and the impoverishment of the superficial strata has led to the utilization of more and more deep aquifers, with wells going down to 600 m i.e. to the freshwater limit. The depth, number and type of the wells depends on the hydrogeological conditions prevailing in the various zones. Thus in the upper Plain where gravels are present, the wells are usually shallower than 120 m (40 – 80 m on the average), occasionally exceeding 250 m (wells of large town water supply systems, irrigation consortia, large industries). Their density is very high, up to 10 or more wells per km^2 (Alessandria, Modena) with the result that, in a number of areas, a drop of over 35 m has been recorded in the static levels (near Modena and Parma) accompanied by aquifer exhaustion and a water balance deficit of the order of 80 l/s$\times$km^2. On the upper Plain, at least where the main fans are situated, the specific discharges are quite high, e.g. 12 – 25 l/s$\times$m at Alessandria.

In the central Plain area east of Parma, the well density is lower owing to the small number of surface aquifers. The low transmissivity makes it

* In this connection it is significant that the structural conditions producing aquifer under pressure differ radically from those of Artois, in France. In Italy, a distinction was made between them as early as 1700. The term "Modena Wells" is used for wells with water having a head deriving from the interstratified tongues of the gravel fans of water-courses.

necessary to attain the greatest possible number of aquifers, so that the wells commonly reach depths of 150 m (specific discharge of $1-5$ l/s$\times$m) although depths ranging to 624 m are also known.

The water balance deficit in the area of the Apennine piedmont fan is due not only to excessive consumption but also to other concurrent factors: the impermeabilization of the surface soil as a result of extensive industrial settlements built from 1955 on, e.g. the one on the River Secchia fan near Modena (more than 300 pottery factories in an area of just over 150 km^2). Attention should also be given to gravel excavation in riverbeds causing their level to drop in some places, at the head of the fans, by as much as $8-12$ m (between Parma and Bologna) thereby reaching the underlying impervious layers.

The diameter of the wells, which are driven in the upper Plain and drilled in the central and lower parts, varies mainly according to the use to which the water is to be put, i.e. from 2″ for domestic wells up to 30″ for those supplying large aqueducts. In general, the wells draw water from more than one aquifer and as the various aquifers intercommunicate facilitating thereby the spread of pollution and during dry periods the deep-seated aquifers supply the emptier, superficial ones.

* * *

Typical geological cross-sections of the various sectors of the Po Valley are presented in Figs 2, 3, 4 and 5. Their location is shown in Fig. 1.

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ÉTUDE DES NAPPES AQUIFÈRES PROFONDES DU BASSIN DE DOUALA (CAMEROUN)

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Avant-propos

L'étude qui est présentée ici a été conduite, à l'origine, dans une optique pétrolière. Il n'en demeure pas moins que les mesures de pression effectuées dans les forages pétroliers, ainsi que les diagraphies qui y sont enregistrées, permettent souvent de dégager les grandes lignes de l'hydrologie profonde d'un bassin. Tel est bien le cas, ici, où l'on s'adresse à une province particulièrement intéressante du point de vue hydrogéologique, puisqu'elle voit la cohabitation de deux systèmes fondamentalement différents: un système à hydrodynamisme par gravité, et un système à hydrodynamisme par compaction.

Cadre géologique du Bassin de Douala

Le bassin de Douala a une forme grossièrement triangulaire. Il est limité à l'Est et au Nord par le socle dont la bordure orientale est très sensiblement Nord–Sud (Fig. 1). A l'Ouest, les vastes épanchements basaltiques du Mont Caméroun le séparent du Rio del Rey. La côte atlantique de direction SE–NW est fortement entaillée par l'estuaire du fleuve Wouri, alors que plus au Sud le petit delta de la Sanaga dessine une avancée triangulaire. Les reliefs y sont faibles; les collines ne dépassent pas 80 m. La bordure orientale du socle culmine à 60 mètres, sans discontinuité notable dans la topographie. Le bassin est monoclinal, sans tectonique apparente. La série visible en affleurement est épaisse de 2400 mètres environ, mais les isobathes du socle magnétique indiquent près de 8000 mètres de sédiments dans la fosse de Kwa-Kwa. Cette série s'étend de l'Albo-Aptien au Quaternaire, la seule lacune notable étant celle de l'Eocène supérieur. Plutôt continentaux et détritiques au voisinage des affleurements, les faciès deviennent franchement argileux et marins vers le large.

Cadre hydrologique du Bassin de Douala

L'Oligo-Miocène (Fig. 2)

Les seuls renseignements que nous possédons concernent Souellaba. Le niveau pseudo-potentiométrique y est normal, puisqu'égal à +84 m. Les salinités de 20 g/l à 27 g/l ne présentent, elles non plus, rien d'anormal. Elles entrent très bien dans le schéma classique d'un bassin sub-hydrostatique.

L'Eocène (Fig. 3)

Les potentiels de Logbaba et de Kwa-Kwa sont des potentiels hydrostatiques (tableau 1). Le fait, cependant, que Logbaba présente un potentiel plus élevé que Kwa-Kwa suggère peut être un léger hydrodynamisme par gravité dirigé des affleurements bordiers vers la partie centrale du bassin. C'est également ce qu'indiquerait l'évolution des salinités qui passent de 0.2 g/l à 30 g/l, d'amont en aval. Par contre, *les puits de Souellaba ne font pas partie du système hydrogéologique précédent*. Les potentiels hydrauliques y sont anormalement élevés et les salinités faibles par rapport à celles de Kwa-Kwa (alors que la profondeur des réservoirs est supérieure).

Tableau 1

Puits	Niveau pseudo-potentiométrique (m)	Salinité (g/l)
Éocène		
Logbaba	+55 à +62	0.2
Kwa-Kwa	+22	30
Souellaba	+462 à +652	1 à 10
Paléocène		
Bomono	+118 à +194	0.2
Logbaba		0.2
Crétacé		
Maestrichtien		
Logbaba 3 bis		2.9
Logbaba 8		0.2
Logbaba 102		20
Logbaba 103	+70	18
Campanien-Sénonien inférieur		
Pibissou	+8	30
Logbaba	+1043 à +2344	10 à 25
Razel 1	+13	7.6
Razel 2		50
Grès de base		
Pibissou	+342	143
Razel 2	−29	45–50 à 120
Razel 1		30

Le Paléocène (Fig. 4)

A Kwa-Kwa l'entrée dans le Paléocène se marque par l'augmentation brutale de la pression de couche.

Nous retrouvons pour le Paléocène les deux systèmes précédemment décrits :

— un ystème sub-hydrostatique auquel se rattache Logbaba,
— un système auquel appartient Kwa-Kwa, où apparaissent des potentiels hydrauliques élevés.

Cette augmentation des potentiels hydrauliques est d'ailleurs liée à une chute importante de la résistivité des argiles comme le montre la figure 8.

Le Crétacé (Fig. 5)

1 *Maestrichtien.* Au niveau du Maestrichtien, LA 3 bis et LA 8 sont envahis par l'eau douce. Nous avons affaire ici à un système à hydrodynamisme par gravité le potentiel hydraulique de LA 103 étant d'ailleurs normal.

2 *Campanien-Sénonien inférieur.* Au niveau du Campanien et du Sénonien inférieur, Razel 1, Razel 2 et Pibissou font partie de la même entité hydrodynamique : les potentiels hydrauliques y sont subhydrostatiques. L'écart de potentiel entre Razel et Pibissou suggère un léger écoulement par gravité. Quant aux salinités elles évoluent normalement avec la profondeur. Logbaba, par contre, appartient à un système à hautes pressions.

Les « grès de base » (Fig. 6)

La formation des « grès de base » correspond à une épaisse formation continentale représentée à l'affleurement par des grès hétérogènes massifs, admettant de rares intercalations de marnes. La salinité de la nappe subit une évolution rapide en direction de la partie profonde du bassin et le potentiel de Pibissou est légèrement supérieur à la normale.

Cadre hydrologique général

Tous les réservoirs postérieurs aux « grès de base » subissent *une évolution identique* lorsque l'on se dirige des affleurements vers la partie profonde du bassin. On assiste d'abord à une augmentation progressive des salinités avec la profondeur, les pressions demeurant hydrostatiques. Il arrive alors un moment où les salinités se stabilisent, et même, dans certains cas, décroissent. On entre alors dans un système à *pressions anormales*. Ce phénomène est particulièrement net entre Kwa-Kwa et Souellaba au niveau de la nappe éocène.

On arrive ainsi à distinguer *deux systèmes hydrodynamiques :*

— un système à hydrodynamisme par gravité limité aux zones périphériques du bassin,
— un système à pressions anormales dont l'extension est plus ou moins grande suivant les unités auxquelles on s'adresse.

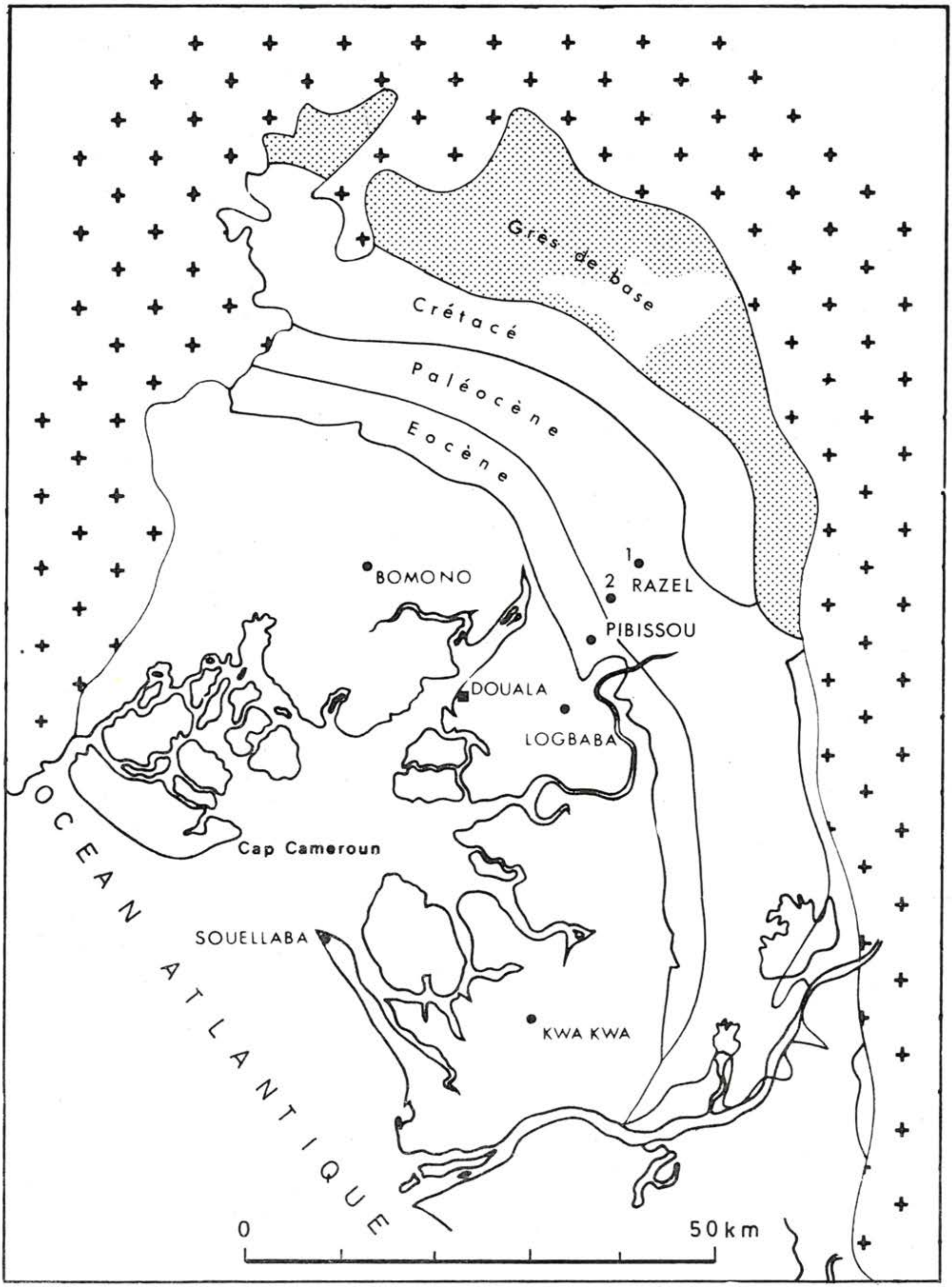

Fig. 1. Carte géologique simplifiée. Noms et emplacements des principaux forages pétroliers

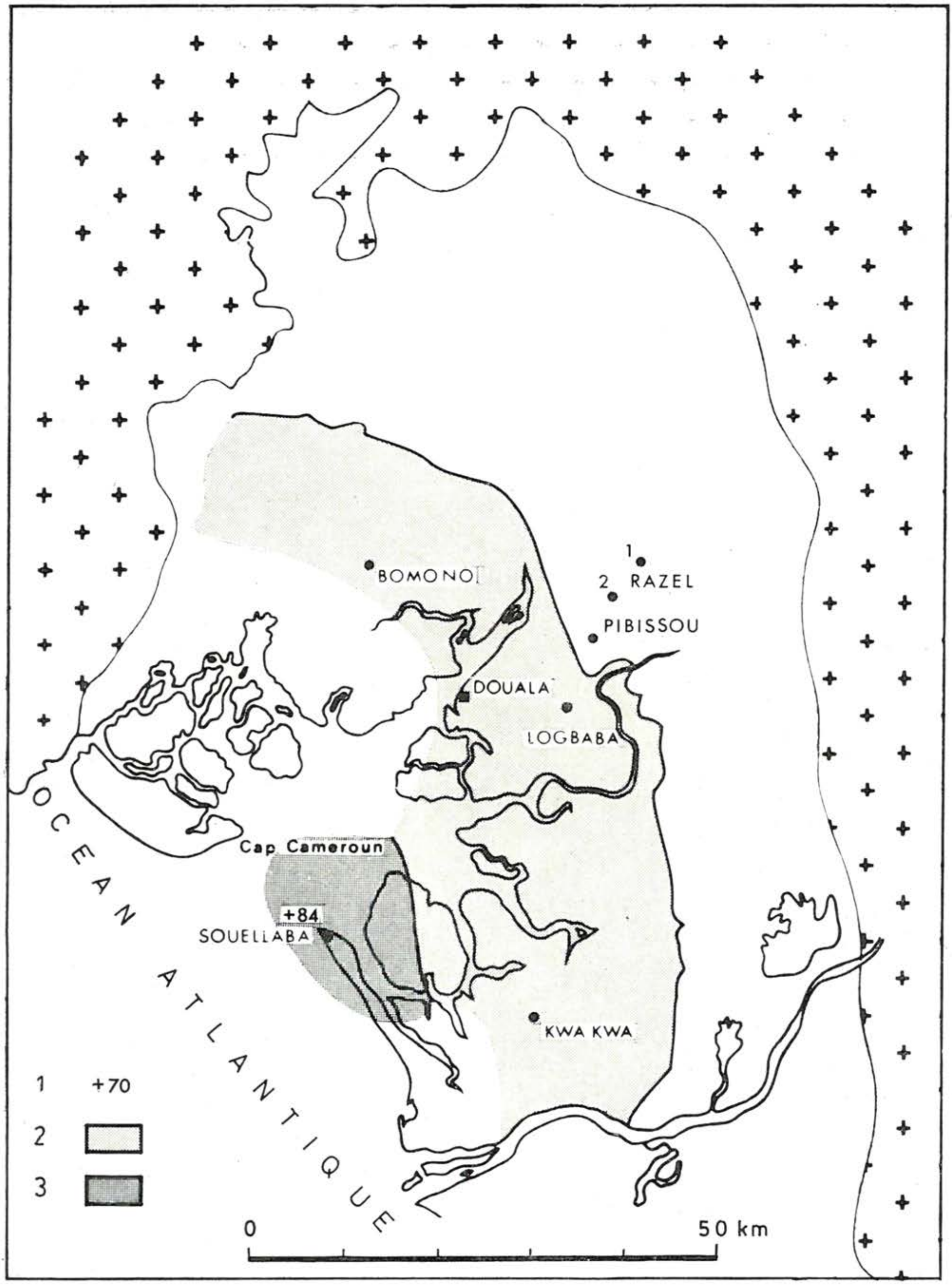

Fig. 2. Cadre hydrologique de l'Oligo-Miocène

1. Niveau pseudo-potentiométrique [m], *2.* salinités en g/l de 0 à 10, *3.* salinités en g/l > 10

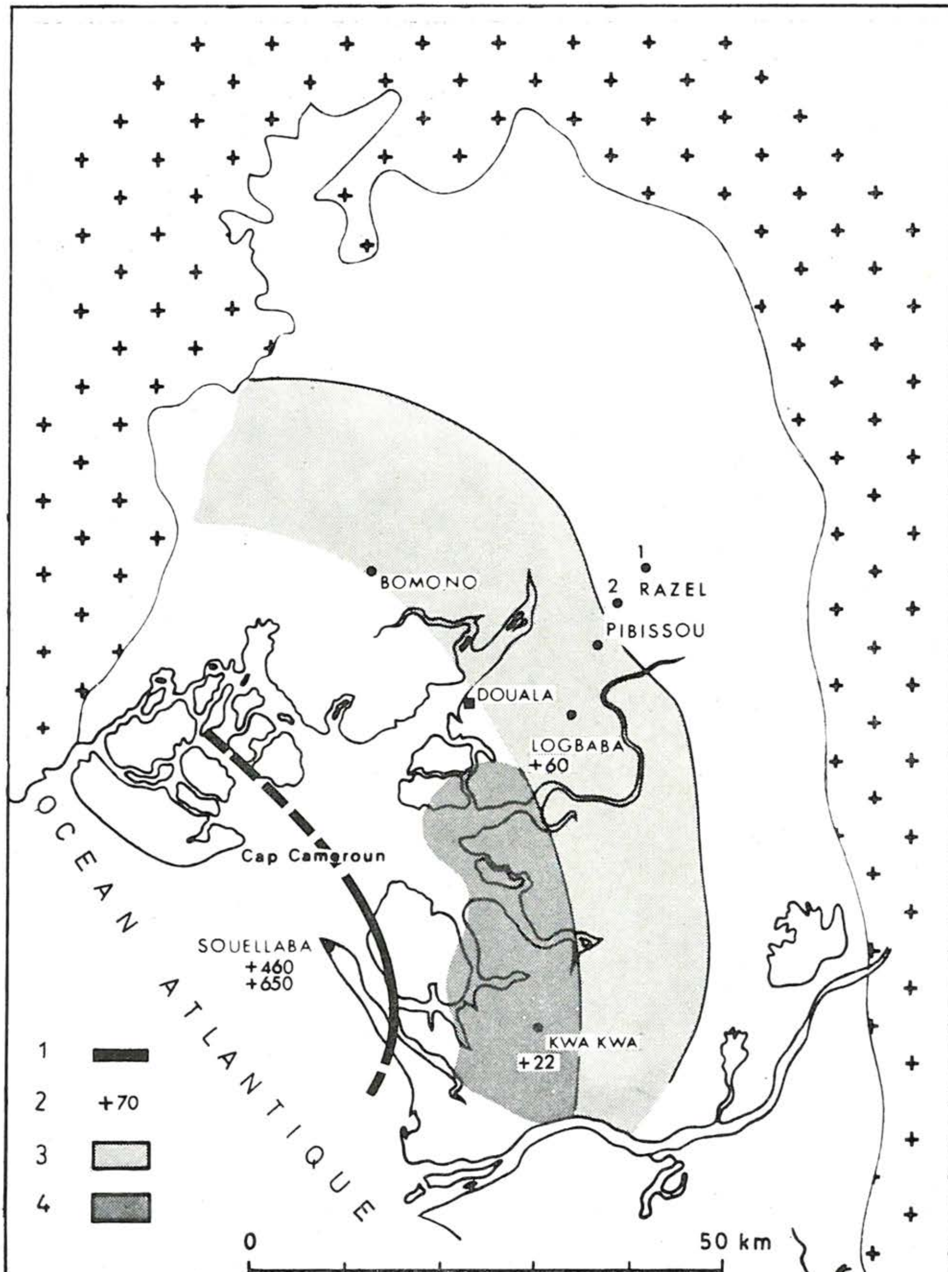

Fig. 3. Cadre hydrologique de l'Éocène

1. Front des pressions anormales, *2.* niveau pseudo-potentiométrique [m], *3.* salinités en g/l de 0 à 10, *4.* salinités en g/l >10

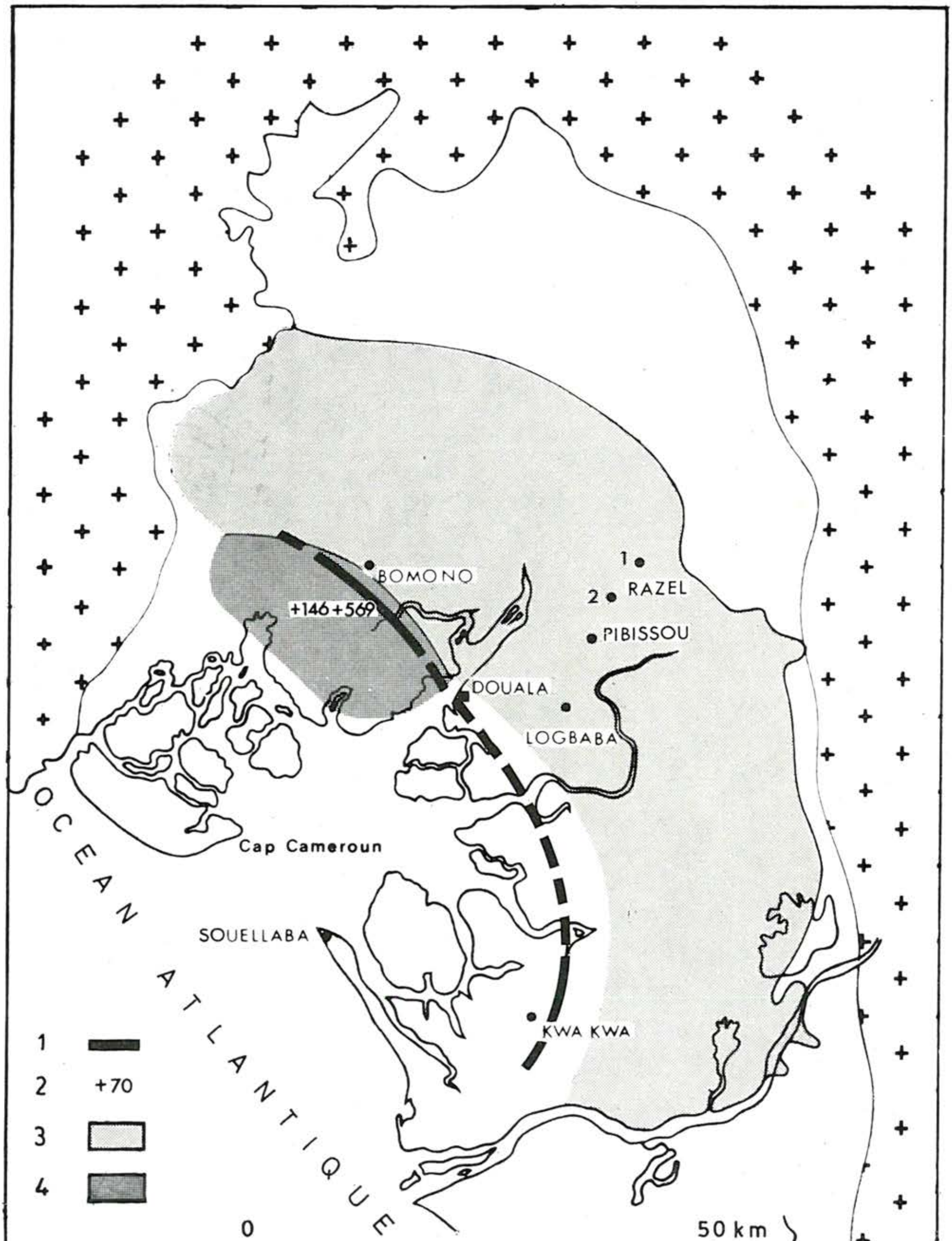

Fig. 4. Cadre hydrologique du Paléocène
Pour la légende voir Fig. 3.

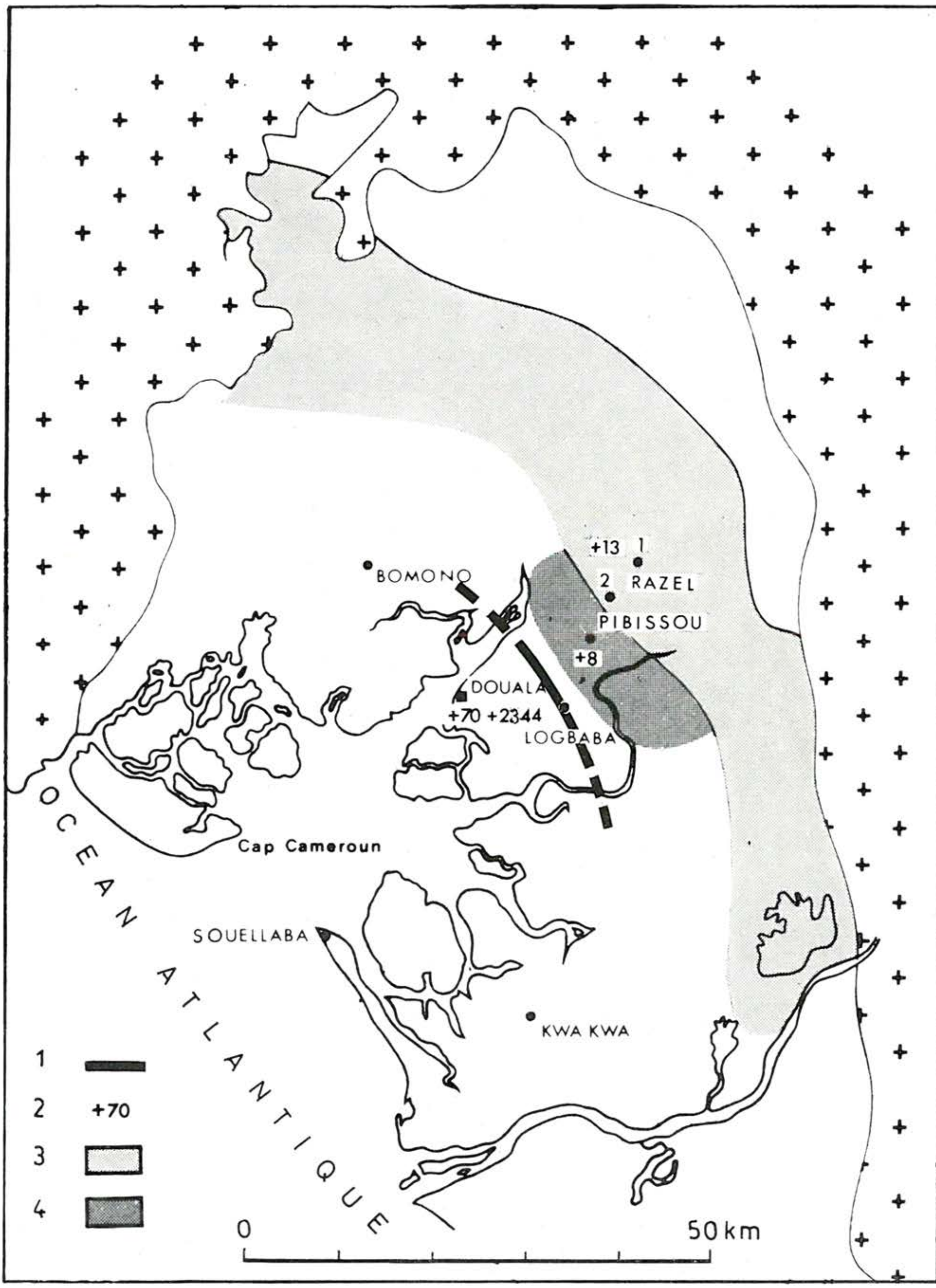

Fig. 5. Cadre hydrologique du Crétacé
Pour la légende voir Fig. 3.

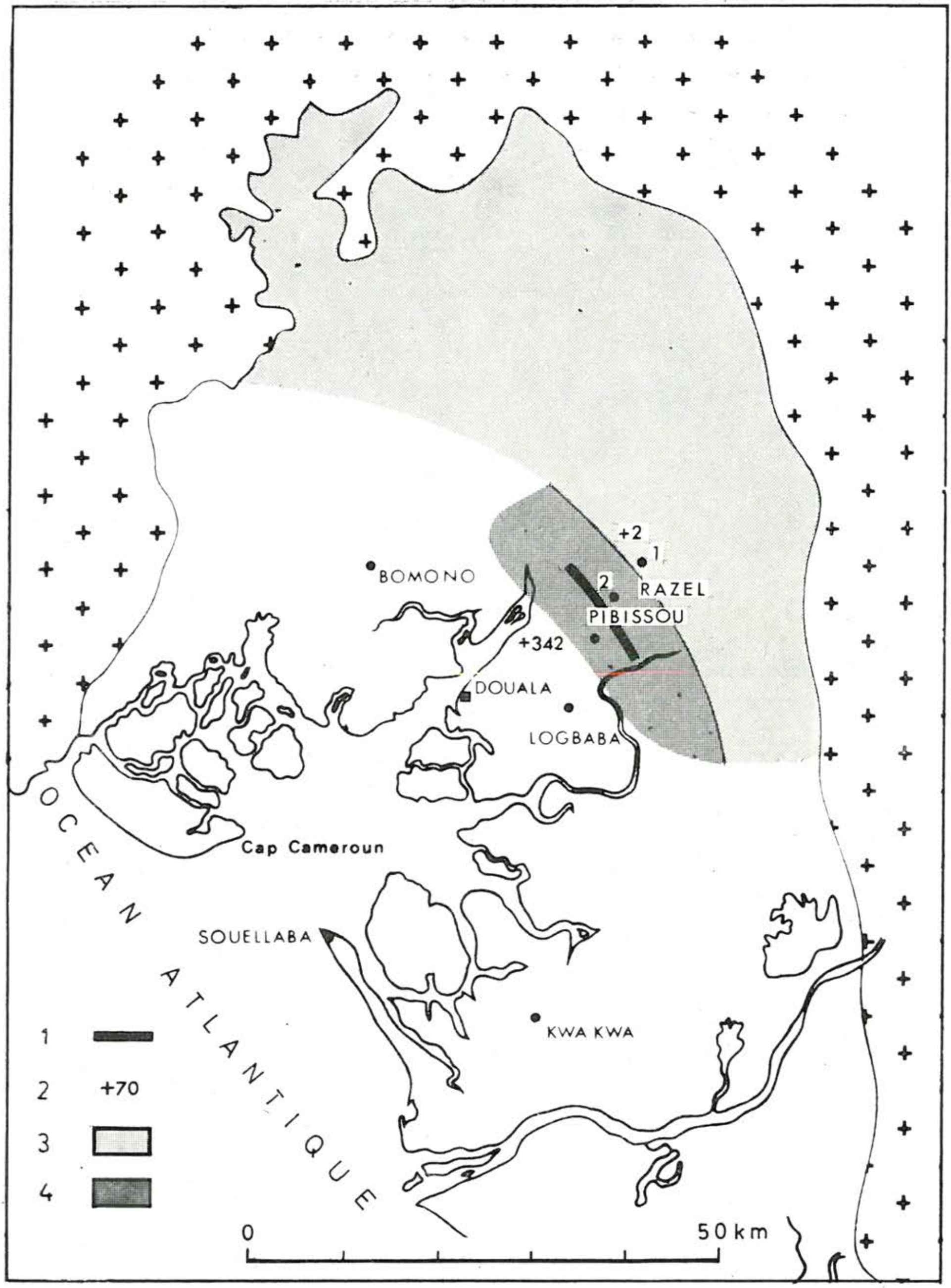

Fig. 6. Cadre hydrologique des grès de base
Pour la légende voir Fig. 3.

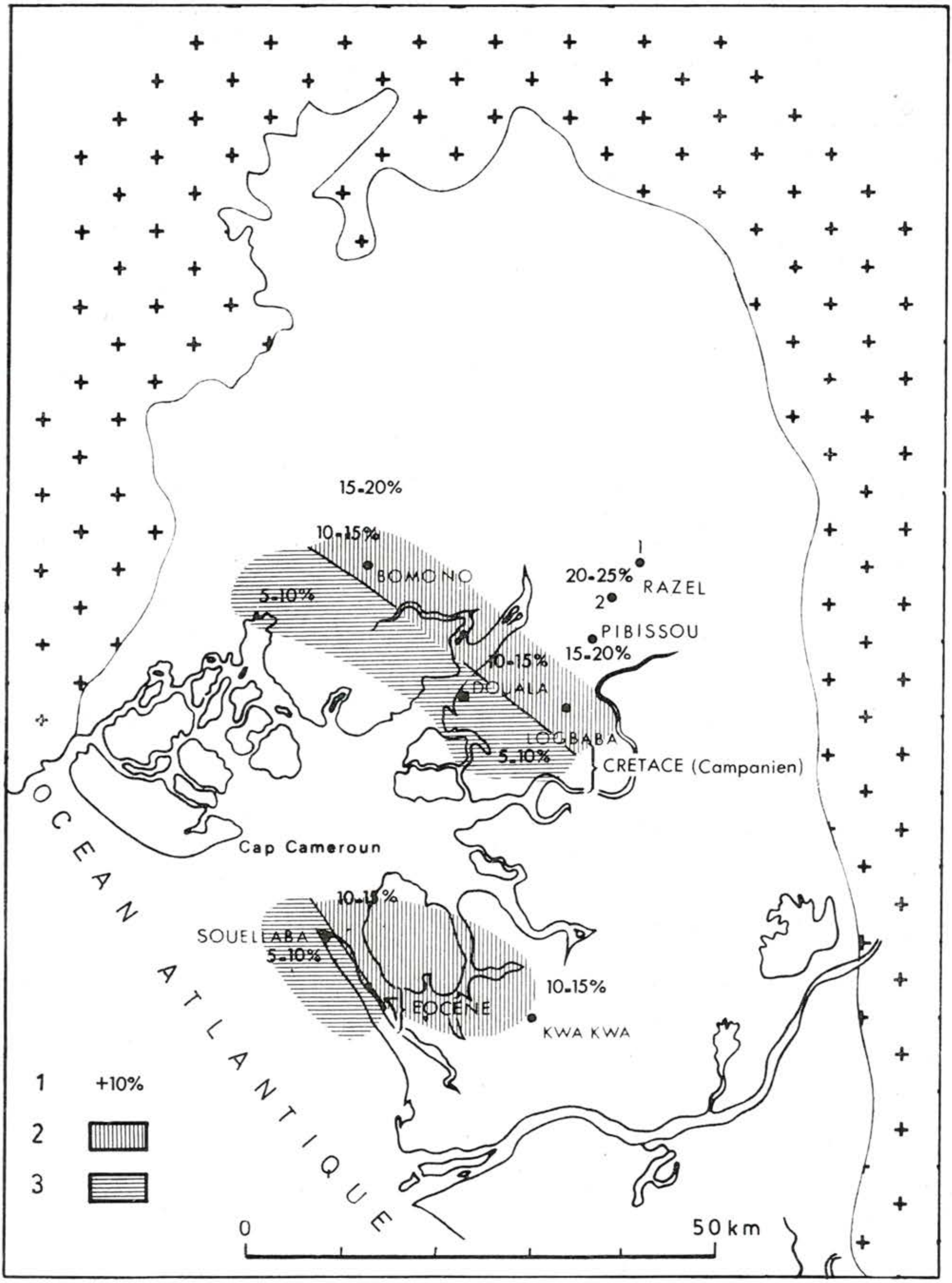

Fig. 7. Éocène et Crétacé. Pourcentage en détritique

1. Pourcentage en détritique, *2.* pourcentage supérieur à 10, *3.* pourcentage inférieur à 10

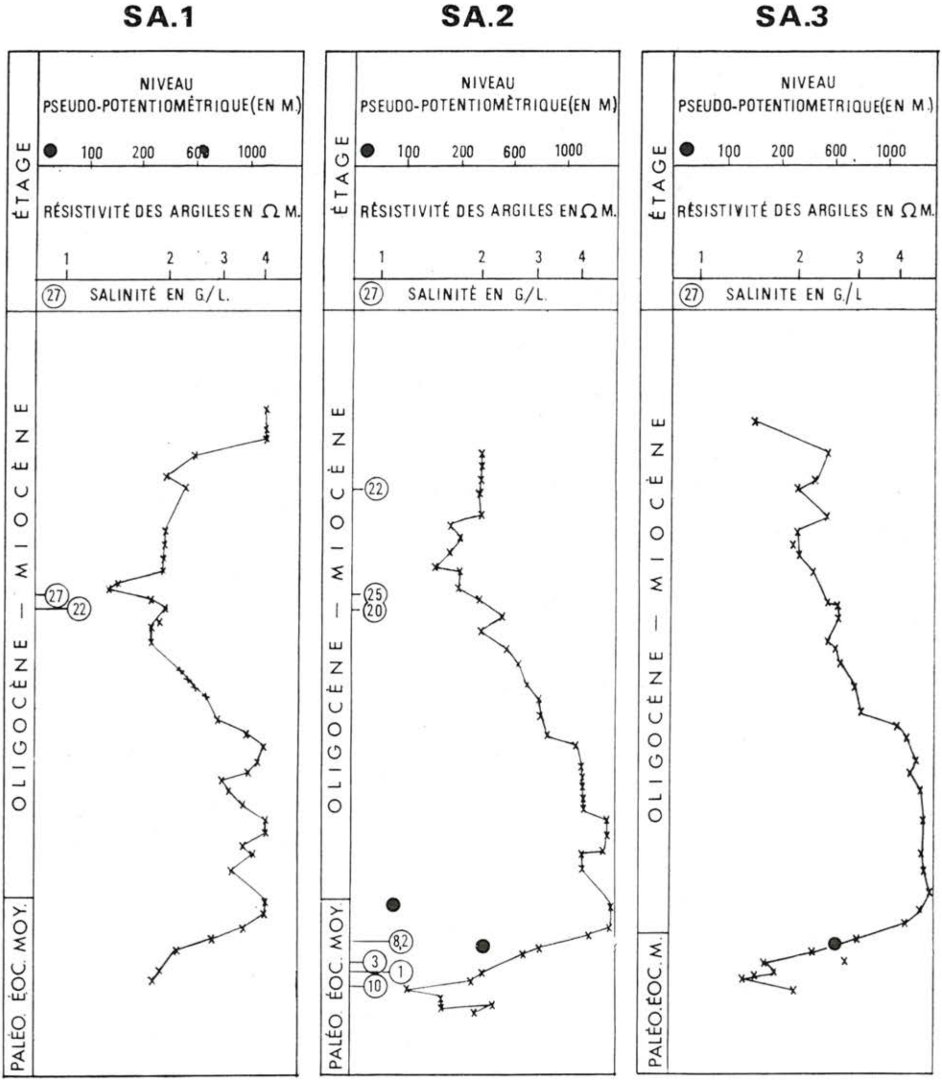

Fig. 8a Étude de la résistivité des argiles. Puits de Souellaba et de Kwa-Kwa

Les pressions anormalement élevées

Les pressions anormalement élevées, c'est-à-dire largement supérieures à la pression hydrostatique, ont généralement trois origines possibles : *géostatique, diagénétique, tectonique.*

Si au cours de la subsidence les fluides contenus dans les pores des sédiments argileux peuvent s'échapper librement, soit directement vers la surface, soit par l'intermédiaire de drains continus, la porosité du sédiment diminue régulièrement avec la profondeur (compaction), et la pression de pore tend, à tout moment, vers une valeur hydrostatique.

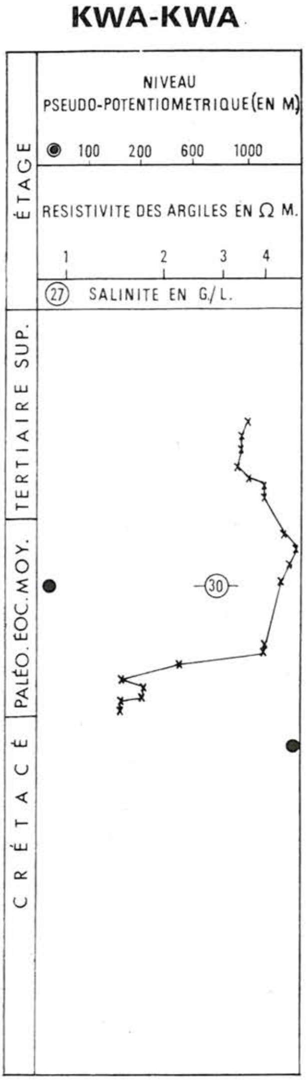

Fig. 8b

Au contraire, si ces fluides ne peuvent s'échapper que difficilement en égard aux conditions d'enfouissement, on assistera à une diminution anormalement faible de la porosité avec la profondeur, et à une augmentation anormale de la pression de pore (l'eau « en excès » supporte une partie du poids des terrains sus-jacents). Les argiles sous-compactées responsables d'anomalies de pression seront caractérisées par des vitesses acoustiques, des résistivités et des densités faibles.

Les anomalies de pression peuvent être également en liaison avec la gazéification de la matière organique présente dans les sédiments, et avec le craquage des hydrocarbures sous l'effet des accroissements de température en profondeur. L'augmentation du nombre des molécules consécutive à ces deux phénomènes – qu'il est difficile le plus souvent de distinguer – entraîne une augmentation de pression lorsqu'elle s'exerce dans un système clos, c'est-à-dire à volume constant. L'anomalie sera alors d'autant plus grande que le réservoir sera discontinu, et le produit hydrocarburé final léger.

Enfin les anomalies de pression peuvent être en rapport avec des contraintes tectoniques. C'est ainsi que la montée halocinétique du sel peut avoir comme conséquence une augmentation du potentiel hydraulique par compression tectonique des aquifères recoupés par le sel.

En ce qui concerne le *bassin de Douala* les fortes pressions apparaissent dans les zones profondes du bassin, où les séries sont les plus épaisses et où, par conséquent, l'enfouissement a été le plus rapide. La flexure de Douala, en particulier, marque une rupture hydrodynamique très nette au sein du Crétacé. Elles sont liées également à la disparition des faciès franchement gréseux, comme le montre la figure 7. L'entrée dans la zone des forts potentiels hydrauliques se marquant par l'apparition d'argiles à basse résistivité comme le suggère la figure 8, *tout milite, finalement, en faveur d'une origine géostatique des hautes pressions*, à laquelle peut venir s'ajouter localement une origine diagénétique.

On se trouve en présence d'un environnement hydrodynamique en beaucoup de points comparable à celui des séries deltaïques de Gulf Coast et du Nigeria.

Comme dans ces deux provinces, le toit des hautes pressions recoupe les surfaces isochrones puisqu'on le rencontre (Fig. 9) :

– dans les grès de base entre Pibissou et Razel 2,
– dans le Campanien entre Logbaba et Pibissou,

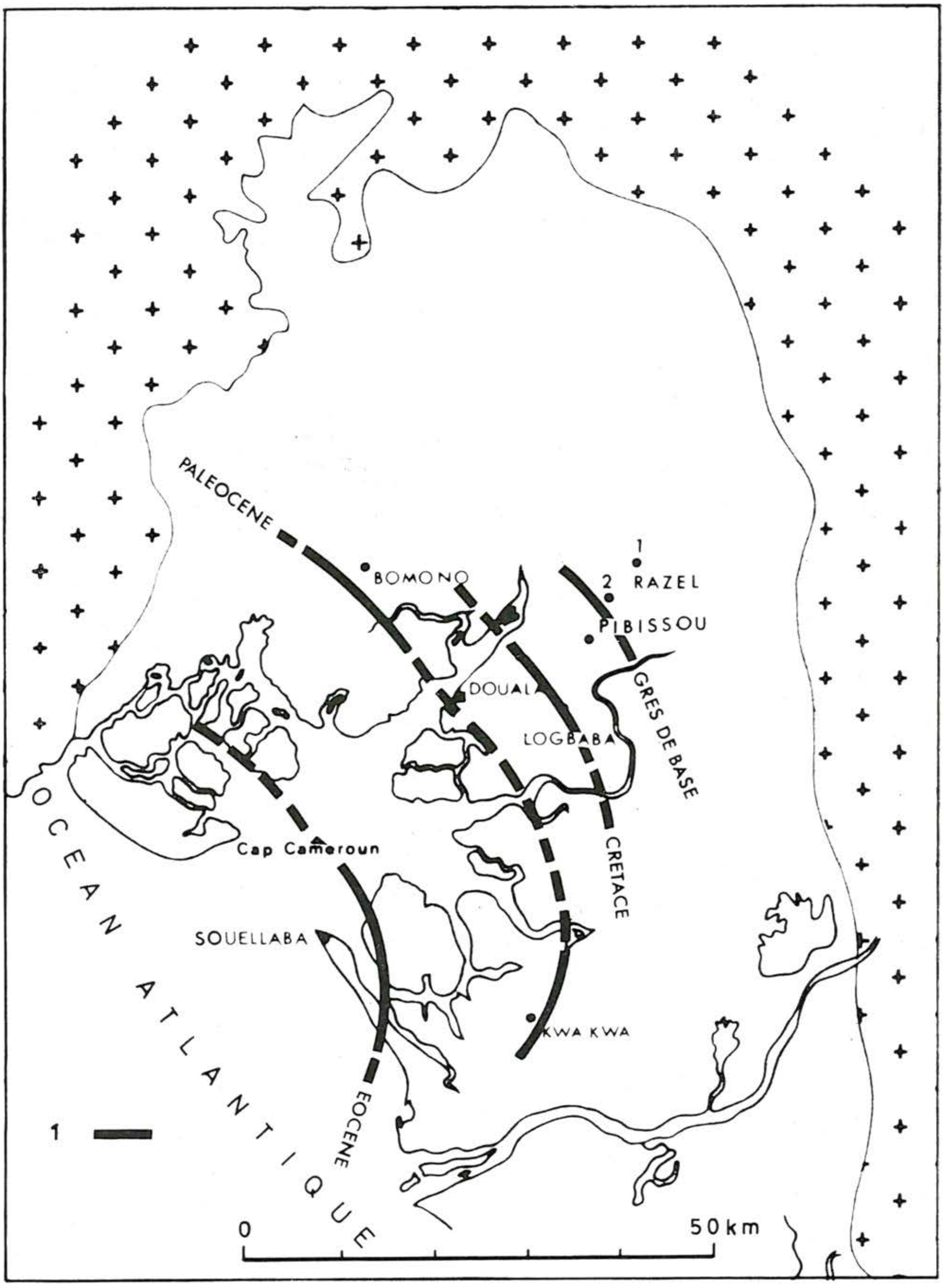

Fig. 9. Répartition stratigraphique du toit des hautes pressions
1. Toit des hautes pressions

— dans le Paléocène entre Bomono, Kwa-Kwa et Logbaba,
— dans l'Eocène entre Souellaba et Kwa-Kwa.

Les eaux peu salines rencontrées à grande profondeur ne font que renforcer l'analogie: ici encore on a affaire à des eaux dont la déminéralisation est le résultat d'une filtration au travers des argiles qui les abritaient primitivement, les ions les plus gros restant piégés du côté amont lors de l'écoulement par compaction. Nous ne possédons malheureusement pas d'analyse détaillée concernant ces eaux.

Conclusion

Le bassin de Douala a fait l'objet d'un certain nombre de forages pétroliers qui ont permis d'obtenir, entre autres, des renseignements intéressants sur les régimes de pression et l'évolution des caractéristiques chimiques des nappes aquifères profondes.

On aboutit à un schéma voisin de celui de Gulf Coast et du Nigeria. En bordure de bassin, c'est-à-dire dans les zones où les réservoirs sont prédominants, on a affaire à un hydrodynamisme par gravité. La salinité des nappes augmente alors de façon classique avec la profondeur. Dès que le drainage de la série devient mauvais, on entre dans un système à hautes pressions. Ce dernier est en liaison avec des phénomènes de sous-compaction au niveau des argiles, ce qui milite en faveur d'une origine géostatique.

On retrouve donc ici, à une échelle réduite, les phénomènes majeurs qui caractérisent l'hydrologie profonde des bassins récents à sédimentation rapide.

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HYDROGÉOLOGIE DE LA PLAINE DU MARQUESADO ET INFLUENCE DE SON DRAINAGE MÉRIDIONAL (DÉPRESSION GUADIX-BAZA, ESPAGNE)

HYDROGEOLOGY OF THE PLAIN OF MARQUESADO AND INFLUENCE OF ITS SOUTHWARD DRAINAGE

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ABSTRACT

The first results and conclusions of a hydrogeological study carried on in the Marquesado del Zenete region (Granada), are presented. This area belongs to the intramontane Guadix-Baza basin.

The surveyed part of this basin, is prevailing filled up with unconsolidated materials. It is located N of Sierra Nevada, in a tectonically complex area of the s. str. Betic Zone, where Nevado-Filábride and the Alpujarride Complexes are represented.

On the southern border of this intramontane basin an open pit mine working is located, the drainage of which plays an important role in the hydrogeology of the Marquesado Plain.

The hydrogeological surveying emphasizes the importance of the ground aquiferous surcharge in comparison with the whole basin.

Such a study also covers different hydrogeological aspects, from the hydraulic behaviour of materials up to the piezometry of the Quaternary detritic aquifer, as a part of the detailed research of the ground-water resources of the Marquesado Plain.

In this hydrogeological scope the influence of the permanent pumping required by the mining work is analyzed in detail. It is recalled that a conspicuous part of the big iron ore body was situated, originally, below the general piezometric level.

Cadre géologique

Les formations détritiques modernes, objet de notre étude, colmatent la grande dépression de Guadix-Baza, localisée dans la Cordillère Bétique, et, plus précisément dans la zone Bétique s. str. (Fig. 1). Nous allons nous limiter, exclusivement, à son extrémité méridionale, la Plaine du Marquesado, où se trouvent les exploitations minières de la « Compañia Andaluza de Minas, S.A. » (CAMSA).

Le substratum de l'aire qui nous occupe est constitué par des formations triasiques, paléozoïques et peut-être plus anciennes. Toutes ces formations ont été affectées par des structures alpines et, en différents degrés, par le métamorphisme alpin, aussi. Ces formations préorogéniques font partie de plusieurs unités structurales, correspondant à des nappes superposées, mises en place à une époque encore mal déterminée. Seulement, l'on peut bien assurer qu'elle se place entre le Crétacé inférieur et le Miocène supérieur ; il semble vraisemblable que cette mise en place ait eu lieu au Nummulitique, sans écarter

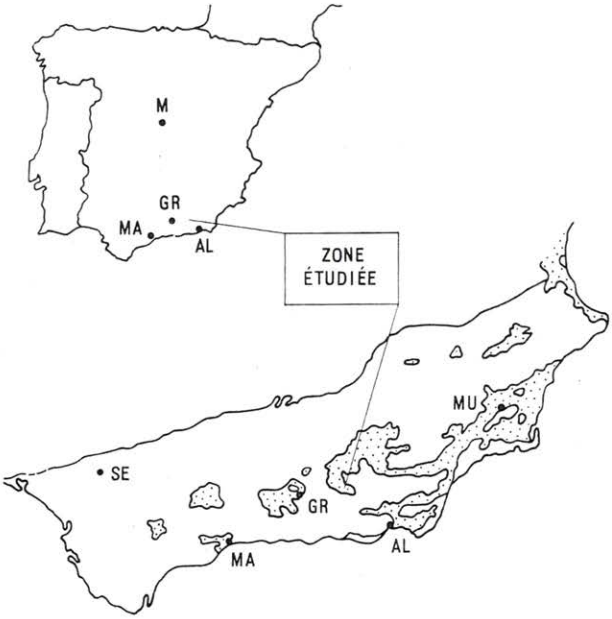


Fig. 1. Terrains postorogéniques des dépressions internes, d'après le schéma structural simplifié des Cordillères Bétiques (FONTBOTÉ, 1965)

que des déformations importantes aient eu lieu au Crétacé (FONTBOTÉ, 1970; id. et al., in litt.).

Dans l'aire étudiée on distingue deux complexes ou ensembles tectoniques de la zone Bétique :

— le complexe Alpujarride,
— le complexe Nevado-Filábride.

Le Nevado-Filábride est représenté par deux unités (PUGA, DIAZ DE FEDERICO et FONTBOTÉ, 1974), qui sont de haut en bas:

— la nappe du Mulhacén,
— la nappe du Veleta.

La nappe du Veleta correspond, à peu près, à ce que l'on nommait auparavant série de la Sierra Nevada (FALLOT, FAURE-MURET, FONTBOTÉ et SOLÉ, 1961), et Unité de la Sierra Nevada (PUGA, 1971). Elle est constituée, fondamentalement, des micaschistes foncés probablement paléozoïques (FONTBOTÉ, 1968), graphiteux presque sans exception. S'y insèrent fréquemment des intercalations lenticulaires de quartzites.

La nappe du Mulhacén est composée des formations décrites par PUGA (1971) sous les noms d' « Unité de la Caldera » et « Unité de las Sabinas ». Leur composition lithologique est variée (FONTBOTÉ, in litt.). A la partie inférieure de la nappe on trouve notamment des micaschistes graphitiques à grenats et/ou d'autres minéraux de métamorphisme, surmontés d'un passage, où abondent les micaschistes sans graphite, et plus riches en feldspaths. La partie supérieure se caractérise par la présence de roches métavolcaniques (amphibolites, gneiss) intercalées, entre des micaschistes, d'abord, puis de marbres ; ces derniers forment, presque exclusivement, la partie la plus supérieure de l'Unité.

Les marbres peuvent atteindre des puissances supérieures à 300 m, mais du fait de la tectonique et de l'érosion leur épaisseur est très variable ; ils peuvent manquer sur de grandes surfaces. C'est à ces termes carbonatés que se rattachent les importants gisements de fer d'Alquife.

Quant à l'ensemble Alpujarride, en nous bornant à l'aire qui nous intéresse, il est représenté, exclusivement, par le Trias et très probablement par le Permien subordonné ; l'on peut distinguer dans cet ensemble deux formations. L'inférieure est constituée, fondamentalement, de phyllites et de quartzites ; et la supérieure l'est de calcaires, de dolomies, et de quelques niveaux de calc-schistes vers le bas.

Les phyllites sont faiblement métamorphisées et, à leur base, sont limitées par une surface de glissement (de charriage ou simplement de chevauchement, suivant les cas) (Fontboté, 1968), oblique par rapport à la stratification et à la schistosité, ce qui explique les variations de puissance observées.

La formation carbonatée comporte des calcaires bien stratifiés à la base, tandis que vers le haut dominent les dolomies.

Dans l'aire objet de cet article, l'ensemble des formations décrites reste presque entièrement enseveli sous les terrains postorogéniques détritiques qui remplissent la partie méridionale de la dépression de Guadix-Baza (Fig. 2).

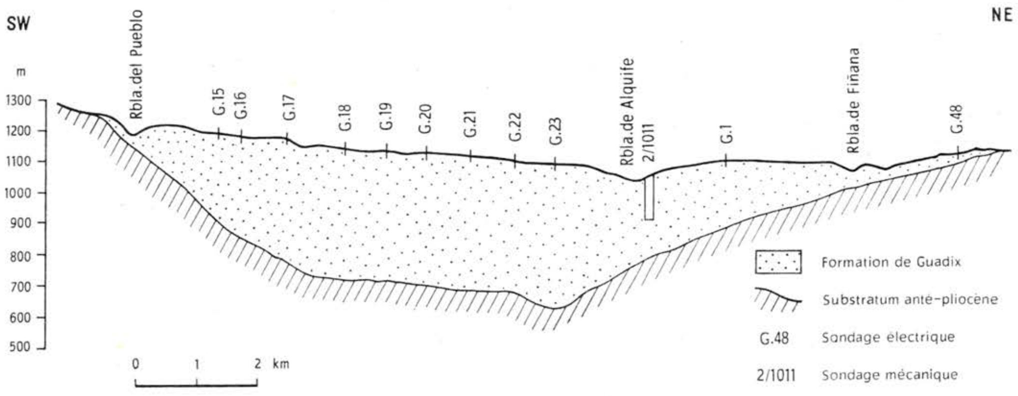


Fig. 2. Coupe de la base de la Formation de Guadix (FAO/IGME, 1972)

L'emplacement de cette dépression ne garde aucun rapport direct avec la tectonique de nappe, mais elle est conditionnée par la structure des plis de grande portée et par les failles, formés surtout dans le Néogène (Fontboté, 1968),

Ainsi, ces formations détritiques colmatent une cuvette, affectée par quelques failles longitudinales, de direction approximativement Est—Ouest ; elles sont limitées au Sud par la Sierra Nevada et au Nord par la Sierra de Baza, qui peut être considérée comme la terminaison occidentale de la Sierra de los Filabres.

Terrains postorogéniques

Les terrains postorogéniques qui, en discordance, remplissent cette partie de la dépression, se distribuent, de haut en bas en ces unités:

Dépôts modernes:

— alluvions
— cônes de déjection.

Formation de Guadix.

Certains auteurs (Pastor Mendivil, 1948) ont considéré à part un autre élément de ces formations, celui qu'on nomme *Rubial Eoceno.*

Il s'agit d'un conglomérat bréchoïde ayant les caractéristiques piémontaises. Très peu étendu, il recouvre les marbres et l'hématite d'Alquife. Des éléments de calcaire et d'hématite rouge, sont cimentés par une matrice calcareo-argileuse et ferrugineuse. La puissance est très variable (dans plusieurs sondages on a traversé plus de 40 m), et la continuité latérale est limitée.

Les études les plus récentes (Fontboté, in litt.), nous permettent, non seulement d'exclure l'âge éocène, mais de rattacher ce conglomérat à la base de la Formation de Guadix. Ses caractéristiques lithologiques spéciales seraient dues à l'abondance de matériel ferrugineux provenant du gisement d'Alquife.

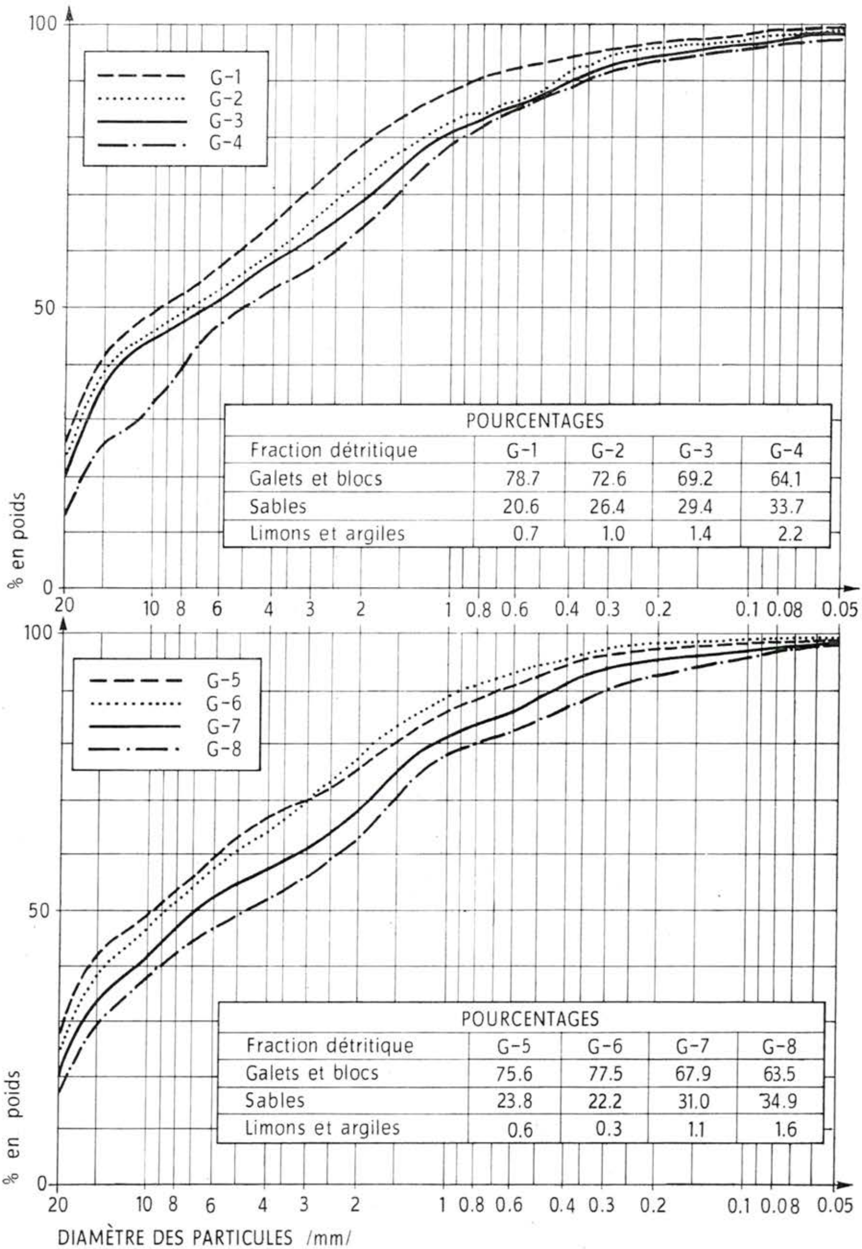
La Formation de Guadix se caractérise par une composition détritique hétérogène, avec de sensibles changements verticaux et latéraux, par sa matrice limono-argileuse, et par son faciès continental. Il existe plusieures études assez détaillées sur cette Formation; nous pouvons citer celles de Fallot, Faure-Muret et Fontboté (1967), et de Vera (1970b).

L'analyse de la granulométrie superficielle (Medina, 1975), réalisée avec des échantillons recueillis suivant deux coupes parallèles, de direction approximative Sud-Nord (Fig. 3), permet de mettre en évidence, des bords vers le centre de la dépression, une diminution de la proportion et de la taille des éléments détritiques grossiers. Près des bords montagneux prédominent les matériaux grossiers, qui se distribuent irrégulièrement suivant une hétérométrie évidente; vers la partie centrale, les niveaux de matériaux fins deviennent beaucoup plus abondants, et les conglomérats s'y trouvent en bancs intercalés. En beaucoup de cas on les voit remplir dans l'ensemble limoneux, d'anciens lits de torrents de formes irrégulières.

Cette Formation de Guadix remplit la plus grande partie du bassin, dont la surface de colmatation correspond à la Plaine du Marquesado. Cette surface se trouve à une altitude moyenne comprise entre 1000 et 1150 m; elle s'élève légèrement du Nord au Sud, et elle est découpée par le réseau de drainage.

Au Sud dominent les éléments (graviers, sables, etc.) qui proviennent du Complexe Nevado-Filábride, tandis que vers le Nord deviennent plus abondants les composants carbonatés, qui proviennent des matériaux alpujarrides de la Sierra de Baza, mais mélangés avec ceux du Complexe Nevado-Filábride qui affleure lui aussi dans cette Sierra.

Dans son ensemble, la Formation de Guadix apparaît comme un remplissage d'origine fluviatile (Vera, 1970a), comportant plusieurs épisodes, dans le temps et dans l'espace, en ce qui concerne l'énergie de transport; ceci a donné lieu à la variation d'épaisseur et à la position relative des graviers, des sables et des limons.



POURCENTAGES				
Fraction détritique	G-1	G-2	G-3	G-4
Galets et blocs	78.7	72.6	69.2	64.1
Sables	20.6	26.4	29.4	33.7
Limons et argiles	0.7	1.0	1.4	2.2

POURCENTAGES				
Fraction détritique	G-5	G-6	G-7	G-8
Galets et blocs	75.6	77.5	67.9	63.5
Sables	23.8	22.2	31.0	34.9
Limons et argiles	0.6	0.3	1.1	1.6

Fig. 3. Courbes granulométriques accumulées des matériaux détritiques

Les données disponibles sur les puissances des sondages réalisés par la CAMSA et par la FAO/IGME (1972) prouvent qu'en général, la puissance augmente des bords montagneux vers l'intérieur de la plaine, où elle peut dépasser 250 m. Cependant, le substratum de ces matériaux détritiques correspondant à un paléorelief montagneux, les puissances peuvent varier considérablement, même entre les points voisins.

D'après les datations faites dans d'autres secteurs de la dépression de Guadix (PENA, 1974, PENA et VERA, 1974), on admet qu'à l'extrémité sud-est du bassin la Formation de Guadix comprend certainement le Pléistocène, mais, en profondeur, on ne peut pas préciser sa limite avec le Pliocène, en supposant qu'il existe dans ce secteur-ci.

Les matériaux pléistocènes se trouvent couronnés par un niveau de 1 à 3 m d'épaisseur, de lithologie similaire au reste, mais avec des éléments plus gros et avec développement de croûtes d'exudation, qui peuvent correspondre déjà au Pléistocène supérieur-Holocène. Récemment on a constaté (CASAS, PENA et VERA, 1975) que la partie terminale de la Formation de Guadix correspond au Pléistocène supérieur, et concrètement, au Riss-Würm (PORTA, 1975).

Au bord sud de cette vallée, directement sur la Formation de Guadix, n note un remarquable développement et extension de cônes de déjection et de glacis de fanglomérats, formés par l'érosion de la Sierra Nevada, qui deviennent coalescents le long de cette bordure.

Les matériaux détritiques grossiers, de plusieurs décimètres au sommet de ces cônes, passent progressivement, vers la périphérie, à des matériaux beaucoup plus fins (limons, principalement), qui plus loin s'étendent en pellicule assez continue, sur la surface de colmatation.

Mis à part ces grands cônes, on en trouve d'autres de moindre envergure, formés au pied des torrents qui ravinent la Formation de Guadix.

En dernier lieu il faut signaler la présence des alluvions actuelles qui tapissent le lit des cours d'eau saisonniers ou temporaires, et sont constituées par de limons, des sables, des graviers et des galets.

Les alluvions forment les terrasses fluviales des cours d'eau les plus importants; celles-ci sont parfois d'une considérable largeur, mais toujours très allongées parallèlement au lit. L'épaisseur est variable selon les profils (le sondage 223/1011 réalisé près du Río Verde — FAO/IGME, 1972 — traversa 22 m de sable, de limon et de gros galets, correspondant au Quaternaire).

Dans la carte nous avons distingué deux niveaux de terrasses (MEDINA, 1975) quoiqu'en réalité on puisse en individualiser un troisième, actuellement en cours de formation, et qui correspond au lit d'inondation des plus grands cours d'eau, où aboutit un grand volume de matériaux solides durant les crues de caractère torrentiel.

Caractéristiques hydrogéologiques

Dans ce cadre lithologique et structural, le comportement hydrogéologique des matériaux décrits, selon des données déduites des sondages et des essais, peut être résumé comme suit:

L'ensemble Nevado-Filábride constitue le substratum général imperméable de la zone (Valle Cardenete, 1970, FAO/IGME, 1972), toutefois, il inclut localement des calcaires marbrés de la partie supérieure de la nappe du Mulhacén — Unité de Las Sabines (Puga, 1971) — qui forment un aquifère avec une conductivité hydraulique de l'ordre de 10^{-1} cm/s (BURGEAP, 1972), là se situent les sondages de drainage de l'exploitation minière à ciel ouvert, de la CAMSA. Cette haute conductivité hydraulique est la fonction fondamentale de l'intense fracturation tectonique des marbres.

De leur côté, les phyllites de l'ensemble Alpujarride constituent aussi un substratum imperméable local, pendant que les affleurements calcaréo-dolomitiques de la partie supérieure contribuent à l'alimentation souterraine latérale des dépôts détritiques postorogéniques. Leurs continuité et extension en profondeur ne sont pas bien connues.

Dans la zone d'Alquife, la base de la Formation de Guadix est fréquemment consolidée par un ciment carbonaté, localement cette couche est appelée rubial. Celle-ci a, dans l'ensemble, peu d'importance parce qu'elle se limite à une aire très petite. Elle présente un mauvais drainage à cause de la matrice argilo-limoneuse qui la ciment et de son manque de tectonisation, de telle façon que sa conductivité hydraulique est de l'ordre de $3 \cdot 10^{-6}$ à 10^{-5} cm/s (BURGEAP, 1972) ce qui permet de la considérer, avec l'hématite, comme un niveau de confinement *(Confining bed)* local (les sondages d'investigation minière ont mis en évidence sa non extension générale sur les marbres).

La Formation de Guadix se présente avec une grande épaisseur et extension, et, vers le centre du bassin, accuse une homogénéisation granulométrique notoire, ce qui implique une sensible amélioration de la conductivité hydraulique, qui, avec l'augmentation de son épaisseur, suppose un accroissement notoire de la transmissivité. On passe ainsi de valeurs de 10^{-4} m²/s (BURGEAP, 1972), au bord méridional, à 10^{-2} m²/s (FAO/IGME, 1972) au centre, et cela justifie le fait que les débits de pompage passent de 2 l/s dans les sondages faits en bordure, à 80 l/s dans les sondages de la zone centrale.

Plus au Nord, en nous éloignant de la source des matériaux détritiques, prédomine la fraction de limon et argile, et la conductivité hydraulique se réduit. Il se produit ainsi une relative fermeture hydraulique dans le secteur nord-ouest de la plaine du Marquesado, bien qu'on ne puisse marquer une limite précise, car il s'agit d'un changement granulométrique graduel avec une large imbrication d'horizons détritiques.

Dans le volume que l'on considère comme un aquifère, la conductivité hydraulique des matériaux détritiques de la Formation de Guadix, est conditionnée par la succession horizontale des niveaux aréno-conglomératiques et limoneux, et, à petite échelle, par la disposition sub-horizontale des galets aplatis de micaschistes, qui se sont disposés en concordance avec la surface topographique. Par conséquent, cette conductivité hydraulique est plus grande suivant cette direction, comme on peut l'apprécier des données suivantes:

Conductivité hidraulique (cm/s)		*Référence*
Horizontale	Verticale	
$5 \cdot 10^{-4}$ à 10^{-3}		FAO/IGME 1972
10^{-3}	$5 \cdot 10^{-3}$ à 10^{-4}	BURGEAP 1972
10^{-3}	10^{-4}	Gordillo 1974

Donc, vues ces caractéristiques hydrauliques, la Formation de Guadix, qui dans sa plus grande partie se trouve en conditions de saturation sous une surface piézométrique libre, constitue, dans son ensemble, un aquifère particulièrement bien adapté au stockage de l'eau et à sa transmission aux divers captages (VALLE CARDENETE, 1970, FAO/IGME, 1972).

Quant aux cônes de déjection, notons leur importance pour l'infiltration de l'eau de ruissellement, provenant de la Sierra Nevada.

Les alluvions, avec une conductivité hydraulique de l'ordre de $2 \cdot 10^{-2}$ à $3.8 \cdot 10^{-2}$ cm/s (FAO/IGME, 1972), jouent le rôle de collecteurs des eaux de la Formation de Guadix. Ceci explique que l'on ait creusé de nombreuses galeries de captage, en profitant de la faible profondeur de l'eau.

Schéma climatique et hydrogéologique

Le climat de la Plaine du Marquesado est continental semi-aride, il correspond à « Csa2 » suivant la classification de KÖPPEN, avec les caractéristiques moyennes suivantes:

Mois le plus froid: Décembre, Janvier (6, 8 °C)
Mois le plus chaud: Août (25,7 °C)
Mois le plus humide: Décembre (55,1 mm)
Mois le plus aride: Juillet (1,6 mm)

La pluviométrie annuelle moyenne est de 300 à 400 mm et les températures extrêmes enregistrées sont de 42.9° à −15.6 °C (FAO/IGME, 1972). Immédiatement au Sud, dans la Sierra Nevada et dans le même bassin versant, les précipitations annuelles moyennes augmentent graduellement de 450 mm à 2000 mm (FAO/IGME, 1972, MEDINA, 1975), Figure 4.

Dans le bassin détritique, la pluviométrie moyenne est de 373 mm/an (MEDINA, 1975). L'évapotranspiration réelle calculée par la formule de THORNTHWAITE avec une capacité de rétention fixée à 50 mm (FAO/IGME, 1972), est de 328 mm/an (MEDINA, 1975).

Le ruissellement sur ces matériaux détritiques étant pratiquement nul, l'infiltration directe est de 45 mm/an, 12 % de la précipitation (MEDINA, 1975).

Pour les matériaux imperméables du bassin versant de la Sierra Nevada, la pluie utile a été estimée à 45% de la pluviométrie, ce qui entraine une valeur moyenne de 500 mm/an (MEDINA, 1975). Cette eau de ruissellement, comme nous l'avons indiqué, une fois qu'elle arrive sur les cônes de déjection ou sur la Formation de Guadix alimente, en tête, l'aquifère, et seulement en période de pluies intenses l'apport superficiel arrive au Río Verde.

De la partie haute du cours des rivières, on dérive l'eau vers des bassins ou des « lacs collinaires » qui ont une capacité totale de 425,750 m^3, destinés à l'irrigation; par cette régulation des eaux superficielles on dérive 16.4 hm^3/an (MEDINA, 1975).

Comme il n'existe pas de station de mesure des débits dans la zone qui nous intéresse, pour calculer la pluie utile nous avons pris en compte les corrélations mises en évidence par l'étude de la FAO/IGME dans le bassin limitrophe du Rio Fardes (station de D. Diego). On arrive ainsi à une pluie utile théorique de 115 hm^3/an, où l'on inclut les dérivations en tête, par irrigation, et les possibles sorties souterraines vers la prolongation septentrionale du bassin hydrogéologique. A partir de la corrélation hydrométrique indiquée ci-dessus, on admet une valeur de 75 hm^3/an pour le ruissellement annuel total moyen du bassin (FAO/IGME, 1972).

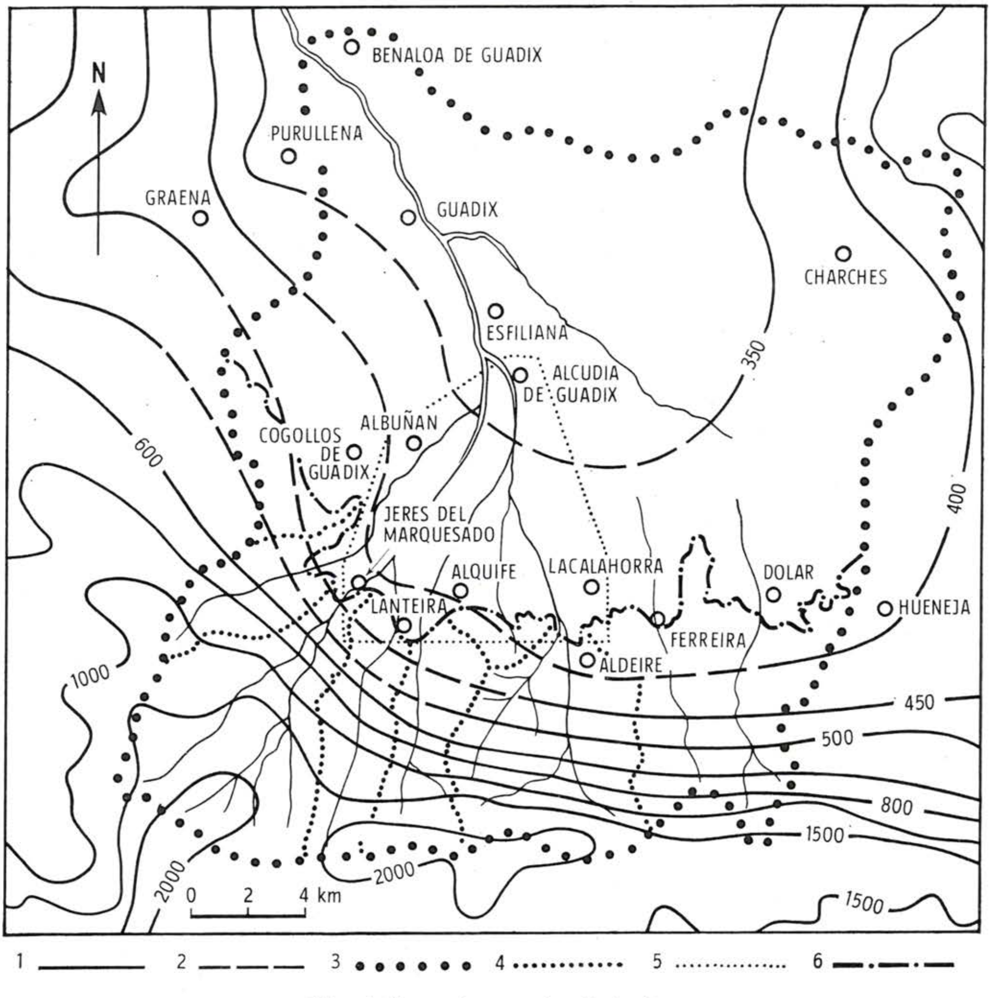


Fig. 4. Carte des courbes isohyètes

1. Courbes isohyètes (1942–1969) (FAO 1972), *2.* courbes isohyètes (1942–1973), *3.* bassin hydrographique du Río Verde, *4.* sous-bassins versants, *5.* aire étudiée, *6.* limite des terrains perméables

Fonctionnement hydrogéologique général

Dans la bordure Sud du bassin (Alquife), le niveau piézométrique primitif se trouvait près de la côte 1048 (Gordillo, 1974), et à partir de là il descendait suivant une pente douce vers le Nord. On peut penser, que dans des conditions non influencées par le pompage de la Mine du Marquesado, ce niveau est sensiblement le même dans la Formation de Guadix et dans les marbres Nevado-Filábrides (Gordillo, 1974, Medina, 1975). Une plus grande conductivité hydraulique de ces derniers pourrait entrainer une plus grande charge.

En observant les courbes piézométriques en état de non-influence par le pompage (Valle Cardenete, 1970, FAO/IGME, 1972), nous pouvons déduire que le réseau hydrographique jouait, par rapport au niveau aquifère, le rôle de *gaining stream* dans ses parties basses et de *losing stream* dans ses parties hautes. Mais l'actuel pompage de la Mine a entrainé des modifications par rapport à ce schéma, modifications que nous sommes en train d'étudier.

Le pompage de la Mine s'effectue dans l'aquifère des calcaires marbrés de la bordure sud. Ceux-ci, vu leur affleurement réduit et leur manque de continuité en profondeur, reçoivent une alimentation directe très faible, par contre il existe une alimentation *per descensum* et latérale à partir de l'aquifère détritique, rechargé lui-même par précipitation directe et par le ruissellement provenant des pluies torrentielles et de la fonte des neiges de la Sierra Nevada (Fig. 5).

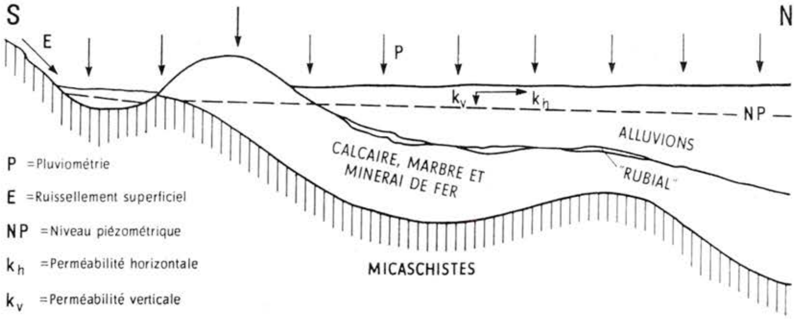


Fig. 5. Coupe hydrogéologique schématique

La recharge de l'aquifère détritique et la circulation subhorizontale qui lui fait suite, se produisent grâce aux cônes de déjection et aux matériaux détritiques grossiers, de sorte que le ruissellement superficiel en provenance de la Sierra Nevada finit, pour sa plus grande part, par alimenter l'aquifère détritique.

Fonctionnement hydrogéologique local

L'exploitation à ciel ouvert, réalisée par la CAMSA, constitue un point singulier, à partir duquel est pompée une importante quantité d'eau.

Le schéma hydraulique *A* (Fig. 6), correspond à la situation initiale, non influencée par le pompage. Quand les travaux miniers arrivèrent au niveau de l'eau, il devient nécessaire de pomper pour continuer l'exploitation en profondeur. On espérait ainsi une baisse du niveau piézométrique aussi bien dans la masse carbonatée, dans laquelle se trouve le gisement, que dans les terrains détritiques de recouvrement.

Mais les choses ne se passèrent pas ainsi car les deux formations étaient séparées par un écran de très faible perméabilité, formé de *rubial*, d'hématite ou de matériaux argileux.

Les caractéristiques hydrauliques favorables des calcaires marbrés permettent d'en extraire facilement des débits importants, ce qui produit un vaste cône de dépression dans cet aquifère; tandis que l'aquifère de la Formation de Guadix, reste perché à un niveau, sensiblement égal au niveau initial, et qui ne descend que très lentement (Gordillo, 1974), par drainance

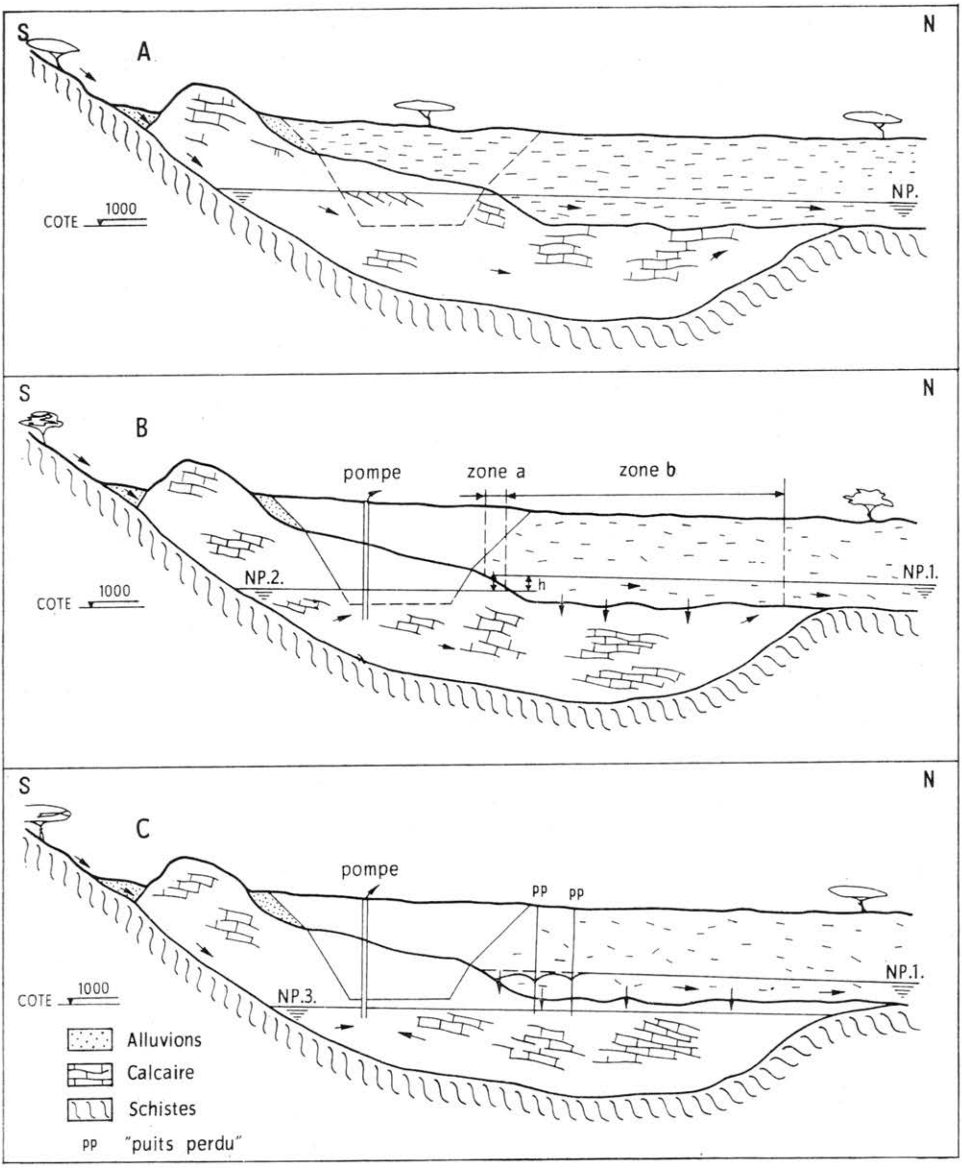


Fig. 6. Schéma du fonctionnement hydraulique de la mine du Marquesado

(« leakage ») étant donné la faible conductivité hydraulique verticale de ces matériaux.

Durant le pompage dans les calcaires, on a pu noter une corrélation linéaire entre le débit extrait et le niveau piézométrique dynamique. Ceci est dû au fait que la masse calcaire reçoit une alimentation induite provenant de la formation détritique, celle-ci est fonction de la différence de charge entre les deux aquifères.

Dans le schéma hydraulique *B* (Fig. 6) on distingue deux zones. Dans la zone « *a* », la Formation de Guadix se décharge sur les marbres secs, ce débit sera donc seulement fonction du niveau piézométrique *1:* comme le contact a un pendage assez fort dans cette zone, son extension et son importance sont réduites. Par contre dans la zone « *b* » l'eau qui percole sera fonction linéaire de la différence de la hauteur *h* entre les deux niveaux, mais comme le niveau supérieur est pratiquement constant, *h* est équivalent à l'abattement créé dans les calcaires marbrés.

Quand le niveau piézométrique *2* a atteint la situation qui apparaît dans le schéma hidraulique *C* (Fig. 6), cette proportionnalité sera rompue et à de petites augmentations de débit pompé, correspondront de fortes dépressions dans la masse carbonatée.

Le pompage a commencé dans la Mine en 1963 avec 100 l/s, et actuellement 9 sondages sont équipés de pompes et le débit extrait est de 495 l/s.

Pour le développement futur de la Mine vers le Sud, la présence de matériaux détritiques imbibés d'eau est un grave inconvénient, car il implique des talus de stabilité extrêmement tendus. Les essais de pompage dans les formations détritiques de ce secteur, ont donné des débits si faibles, qu'il serait prohibitif d'essayer d'accélérer la dépression naturelle du niveau piézométrique *1* par ce procédé. Pour cette raison la CAMSA a décidé d'installer un réseau de *puits perdus (wells points)* qui permettent le passage de l'eau depuis la Formation de Guadix vers les marbres.

Ce procédé, qui est encore une phase expérimentale, a buté sur quelques difficultés qu'il convient de signaler.

Au début, on a pensé utiliser les forages d'investigation minière en plaçant des filtres à la hauteur des niveaux détritiques mouillés et des carbonates, et une tuyauterie non crépinée au niveau de la zone imperméable. Mais comme les forages miniers se font par rotation et avec de la bentonite, les puits se colmataient rapidement; il a fallu donc abandonner le système et faire les puits perdus par battage. Dans ce cas il est préférable de crépiner tous les tubages qu'on utilise, pendant la perforation en dessous du niveau piézométrique, car sinon, ils risquent de rendre le puits inutilisable.

D'autre part, le colmatage des puits est fréquent étant donné la quantité de matériaux fins de la Formation de Guadix. Pour procéder à leur nettoyage (avec des polyphosphates, Giltex, etc ...) il faut employer un *packer* qui retient la solution à la hauteur désirée, au moins pendant vingt-quatre heures. Dans le cas de colmatage des marbres, le nettoyage est facilement réalisable avec du ClH.

Pour un contrôle effectif de ce drainage, il est nécessaire d'installer un dispositif de piézomètres avec une densité simiIaire à celle des puits perdus.

Étant donné les conditions hydrauliques de l'aquifère détritique de ce secteur, pour abaisser la nappe dans un rayon de 300 m, on pense qu'il sera nécessaire de recourir à une maille de puits perdus inférieure à 100 m.

Piézométrie

En fonction des données les plus récentes de l'inventaire des points d'eau, on a fait une interprétation piézométrique (Medina, 1975) qui met en évidence aussi bien les zones d'alimentation et de drainage, que l'effet piézométrique produit par le pompage et le drainage de la Mine.

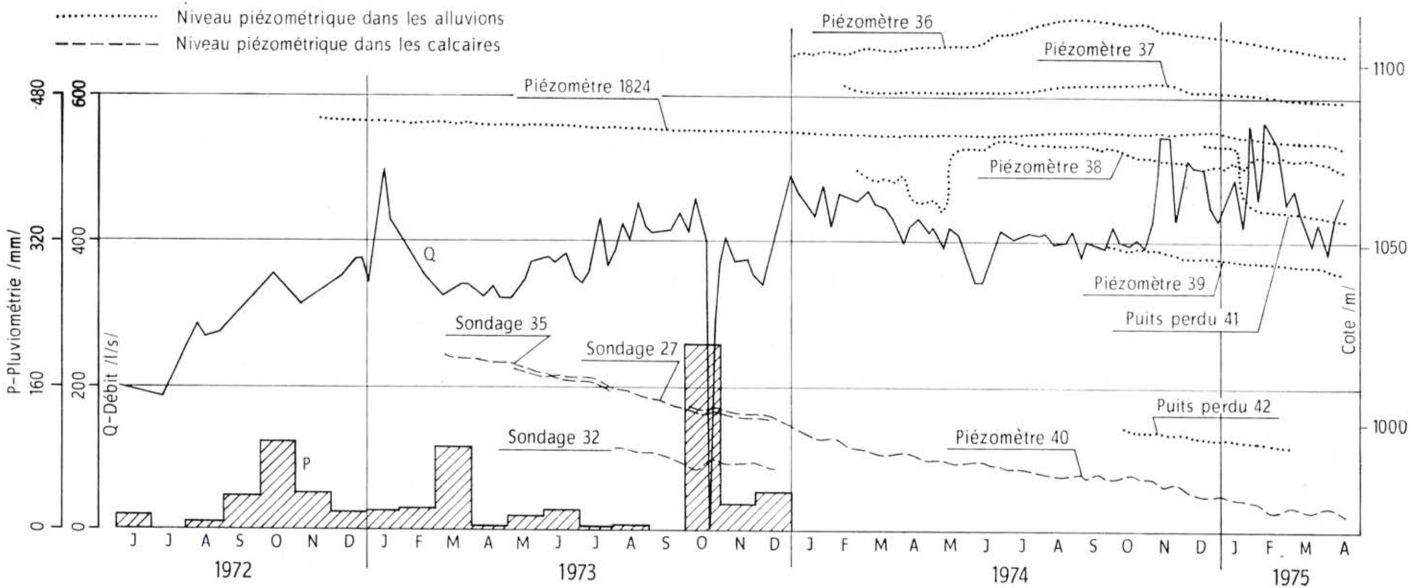


Fig. 7. Relation entre pluviométrie, pompage et niveau piézométrique

En observant le graphique de la hauteur piézométrique en fonction du débit pompé (Fig. 7), on peut constater que la saison des pluies n'influe pas sur la piézométrie dans les calcaires, et que son effet sur le niveau piézométrique dans les matériaux détritiques est d'autant plus intense que celui-ci est plus élevé.

Le temps que met cette fluctuation piézométrique à se manifester est fonction de la profondeur et pour les piézomètres étudiés il semble y avoir un déphasage de 4 à 6 mois (MEDINA, 1975), fluctuation qui dépend aussi du pompage et qui est conditionnée par l'anisotropie du milieu et par le processus de drainage.

Étant donnée la différence de pentes entre la topographie et la nappe aquifère, la profondeur du niveau piézométrique des matériaux détritiques, varie entre plus de 100 m, dans l'exploitation minière, et à peine quelques mètres, dans les terrasses fluviales septentrionales.

D'autre part, l'observation des mesures piézométriques mensuelles des points d'eau éloignés de la Mine (Fig. 8) met en évidence une fluctuation saisonnière peu accusée; toutefois celle-ci est plus intense vers le Nord de la dépression à cause de l'alimentation naturelle (MEDINA, 1975).

D'une manière générale, dans les conditions actuelles, le ravin de Benéjar semble avoir une fonction de « *losing stream* » et le Rio Verde, de « *gaining stream* », et entre les deux se situe la zone de drainage de la Mine, avec les incidences piézométriques antérieurement citées, qui modifient la circulation générale du Sud au Nord, au moins dans un rayon d'un kilomètre.

Ainsi le drainage actuel n'est pas seulement constitué par des suintements naturels ou ceux d'anciennes galeries, mais aussi par les pompages, dont le 97.5% appartient à la Mine, et le reste aux captations d'une ferme expérimentale. Le total de ces eaux est de 16 Hm³/an (MEDINA, 1975).

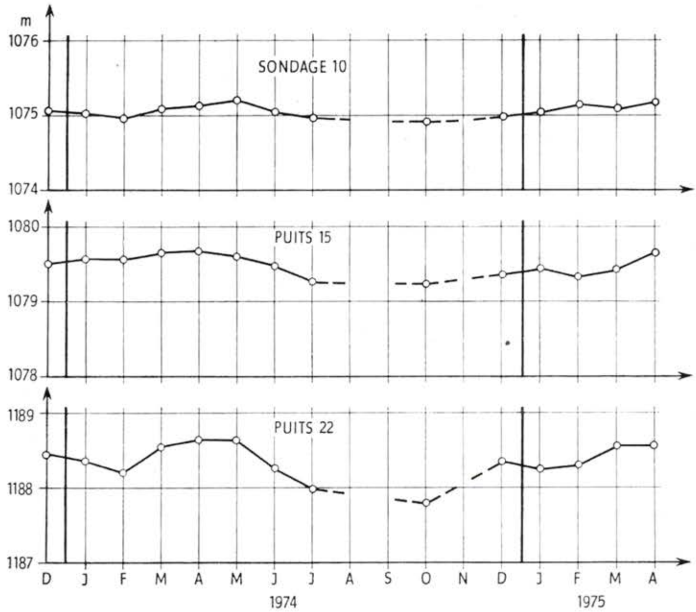


Fig. 8. Fluctuations piézométriques

Pour analyser l'incidence du pompage dans l'aquifère détritique et prévoir ses conséquences à long termes nous avons commencé la mise au point d'un modèle mathématique de l'aquifère, en régime non permanent (FERNÁNDEZ-RUBIO).

Nous voulons remercier au Prof. J. M. FONTBOTÉ pour sa révision des aspects géologiques.

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HYDRODYNAMIQUE ET HYDROCHIMIE DU COMPLEXE AQUIFÈRE DE HAUTE ET MOYENNE CAMARGUE

HYDRODYNAMICS AND HYDROCHEMISTRY OF THE SUCCESSION OF AQUIFERS OF HAUTE AND MOYENNE CAMARGUE

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ABSTRACT

Plaisancian marls which outcrop along the Costière and the Crau, constitute the impervious basement of the whole plain of the Camargue. They are overlain by Astian sandstones and sandy layers and by the gravels and conglomerates of the Villafranchian and old Quaternary, which represent a confined aquifer system, separated by peat beds and lenses from the phreatic aquifer system of recent alluvia.

The unconfined aquifer of recent alluvia shows that its hydraulic parameters are strongly influenced by the lithology which is related to the successively changing courses of the Rhône river. Thus, ground water flow is from high irrigated areas towards low closed depressions where water stagnates and is consumed by evaporation. This phenomenon has been particularly well shown by hydrochemistry, the zoneography of which corresponds with the piezometric surface, and by the study of seasonal fluctuations of salinity with depth.

The confined aquifer, whose recharge zones are located on the Costière, the Crau, and upstream of the river divergence, has a very small ground water flow range towards the south. Its hydraulic parameters are dependent upon the thickness of the aquifer body, whose upper part, that has been eroded, shows terraces and thalwegs. Hydrochemistry has shown an invasion by marine water and the presence of Fe^{3+}.

Finally, the results of tritium analyses together with the results of the water density study, which show some possible communications between the two systems, are mentioned in short.

Introduction

La Haute et Moyenne Camargue, limitée au Sud par l'étang du Vaccarès, représente la partie nord du delta du Rhône, et des terres gagnées sur la mer par accumulation des charges sédimentaires du fleuve. D'un point de vue paléogéographique, la transgression marine miocène se termine par un épisode pontique, connu par sondages uniquement. Au Pliocène, une nouvelle transgression, moins importante, envahit les zones érodées au cours de l'émersion pontique, et ce, jusqu'à l'Astien. Puis le régime continental s'établit de nouveau, caractérisé par les dépôts caillouteux. Pendant cette régression pré-flandrienne, le Rhône, grossi de la Durance, érode la bordure de la Costière et de la Crau ainsi que le cailloutis. Il y fait suite la transgression flandrienne, pendant laquelle la présence de marécages, en arrière de la zone littorale, est caractérisée par le dépôt de limons argilo-sableux, mêlés à des horizons de tourbe. L'érosion cesse et le Rhône comble de ses alluvions ses lits successifs.

Climatologie

Le climat est caractérisé par des précipitations irrégulières, à l'irrégularité du régime annuel se superposant celle, non moins importante du régime hyperannuel. La tendance semi-aride du climat est très marquée et les périodes estivales de sécheresse sont caractérisées par un déficit hydrique, l'apport d'eau par les précipitations restant inférieur à la consommation par évaporation et évapotranspiration.

Facteurs climatiques à La Tour du Valat (1944–67):

Température moyenne annuelle	14.8	°C
Précipitation moyenne annuelle	555.3	mm
Évaporation moyenne annuelle mesurée par évapomètre Piche	1571	mm

Quant à l'évapotranspiration réelle, étudiée sur des rizières expérimentales, pendant un laps variable de 120 à 138 jours, qui correspondent à la période de culture du riz de Mai à Septembre, elle atteint la valeur de 7682 à 8913 m^3/ha, soit 6.4 mm/jour.

Hydrologie superficielle

Remarque sur le cours du Rhône

Le Grand Rhône coule dans un lit limoneux, plus rarement sableux, et les poudingues de la Crau forment le seuil de Terrin, 10 km en aval d'Arles. Son débit maximal moyen est de 5200 m^3/s et son débit minimal moyen de 500 m^3/s (1961–70).

Quant au Petit Rhône, qui est en voie de colmatage, il a un débit représentant seulement 16% du débit du fleuve.

La riziculture

La riziculture s'est développée essentiellement sur les zones hautes des anciens bras du Rhône et la culture du riz dure cinq mois (de Mai à Septembre). On estime 4000 à 9000 m^3/an le volume d'eau qui s'infiltre durant toute l'immersion des rizières avec un apport extérieur de 21 000 à 23 000 m^3/ha, soit 1/3 des apports [8]. En Camargue, on peut estimer à $3.8 \cdot 10^8$ m^3/an les besoins en eau de la riziculture, à $3 \cdot 10^7$ m^3/an les besoins nécessaires à la submersion des vignes et à $6 \cdot 10^6$ m^3/an les besoins pour l'irrigation des prairies. 50 % au moins des apports ainsi introduits seraient éliminés par colature, et refoulés vers le Rhône [8].

Hydrodynamique et hydrochimie

La nappe phréatique

La série des alluvions modernes est caractérisée par une très grande hétérogénéité, due aux divers phénomènes ayant accompagné leurs dépôts.

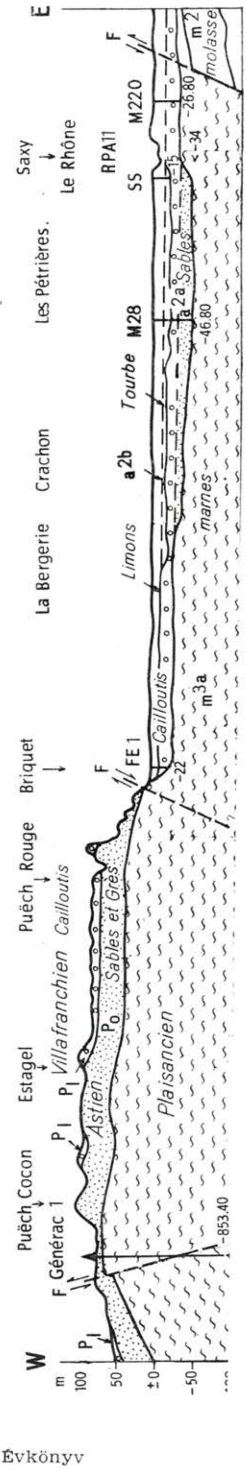


Fig. 1. Coupe W—E de Générac à Saxy

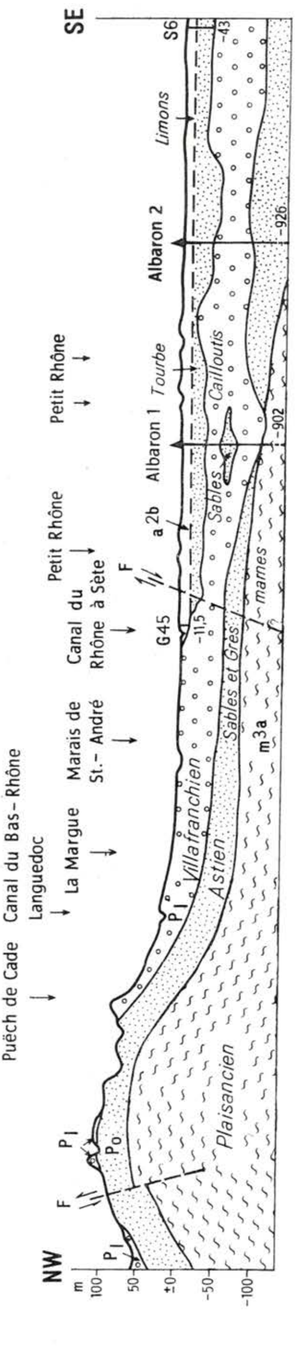


Fig. 2. Coupe NW—SE de Générac au Vaccarès

1. *Lithologie*

Le mur de la nappe est constitué de niveaux de tourbe imperméable s'imbriquant les uns dans les autres (Fig. 1 et 2).

Les anciens cordons littoraux présentent des galets à leur enracinement et sont constitués de sable à granulométrie caractéristique [5]. Quant aux bourrelets du Rhône, actuels ou anciens, ils montrent une texture allant de limon sableux (ou sable limoneux), dans leur partie haute, à des textures de plus en plus fines de limon argileux, argile limoneuse et argile vers les zones de marais qu'ils délimitent (Fig. 12).

2. *Caractéristiques hydrauliques*

Valeurs de K mesurées par la méthode du trou de tarrière [9] :

	Structures	
	K max. (cm/s)	K min. (cm/s)
Sables	$8 \cdot 10^{-3}$	$4 \cdot 10^{-3}$
Sables limoneux	$6 \cdot 10^{-3}$	$2 \cdot 10^{-3}$
Limons sableux	$4 \cdot 10^{-3}$	$9 \cdot 10^{-3}$
Limons	$3 \cdot 10^{-3}$	$2 \cdot 10^{-4}$
Limons argileux	$2 \cdot 10^{-3}$	$9 \cdot 10^{-5}$
Argiles limoneuses	$9 \cdot 10^{-4}$	0
Argiles limoneuses à argile	$1 \cdot 10^{-4}$	0

3. *Alimentation*

La nappe phréatique est une nappe à potentiel imposé dont le Grand Rhône et le Petit Rhône représentent une limite perméable. Cependant, l'enregistrement sur limnigraphe de leur niveau et le relevé des niveaux piézométriques de puits de plus en plus éloignés ont montré que l'influence du fleuve se limitait aux zones qui le bordent (Tableau 1). L'irrigation des cultures, le riz en particulier, est bien plus importante pour la morphologie de la nappe à cause des volumes d'eau ainsi introduits. C'est elle qui règle en définitive le régime des hautes et des basses eaux indépendamment du régime du fleuve et de la climatologie.

Tableau 1

Piézométrie comparée du mois d'Août 1970 (en m. N.G.F.)

Lieux	Éloignement du fleuve en km	09/08	14/08	22/08	30/08
Rhône au pont de St-Gilles	0	1.06	1.64	1.70	1.57
Mas-du Roi	0.1	3.69	4.05	3.78	4.09
L'Auricet	0.1	3.04	3.23	3.04	2.96
Mas-Neuf de Julian	1.6	2.33	2.38	2.35	2.30
Stanislas	4.2	2.08	2.14	2.06	2.07
Cabanne du Boulevard	3.1	0.96	1.01	0.81	0.82

4. *Écoulement*

Les courbes isopiézométriques (Fig. 3 et 12), espacées dans les zones hautes des bourrelets (où le gradient est de l'ordre de 0.05%), se ressèrent vers les zones basses des marais (où le gradient hydraulique est de l'ordre de 0.15%) par diminution de la perméabilité et des pertes de charge importantes. L'écoulement de la nappe se fait donc des zones hautes d'irrigation vers des bassins fermés, où la surface piézométrique est très près de la surface du sol.

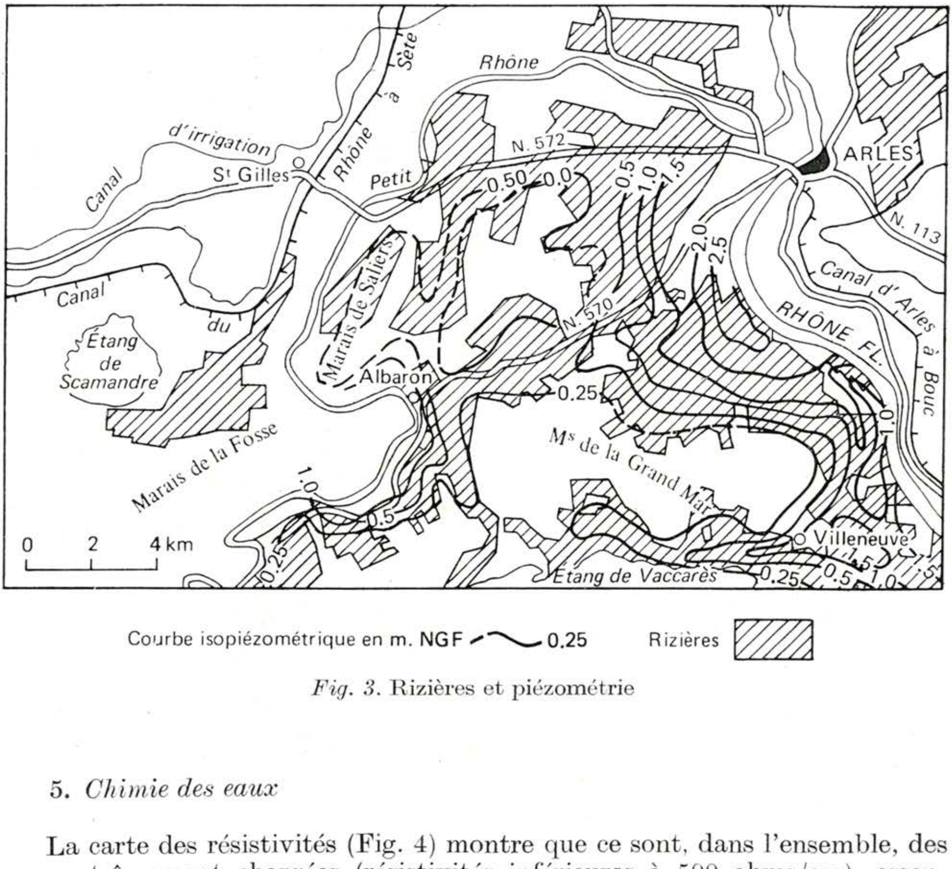


Fig. 3. Rizières et piézométrie

5. *Chimie des eaux*

La carte des résistivités (Fig. 4) montre que ce sont, dans l'ensemble, des eaux extrêmement chargées (résistivités inférieures à 500 ohms/cm), essentiellement réparties sur les zones basses. Les eaux de salinité moindre se trouvent de chaque côté des bras du fleuve, et sur l'emplacement des zones irriguées.

Ce sont des eaux chlorurées sodiques, saumâtres pour la plupart. On a étudié les variations de teneur en Cl^-, entre l'été et l'automne, du niveau supérieur (0 à -6 m) et du niveau moyen (-6 à -10 m) (Fig. 5 et 6). Si la zonéographie du niveau moyen ne montre guère de variation, le niveau supérieur montre une régression des eaux extrêmement salées (concentrations supérieures à 250 méq/l de Cl^-) et une extension des zones saumâtres (concentrations de 10 à 250 méq/l de Cl^-).

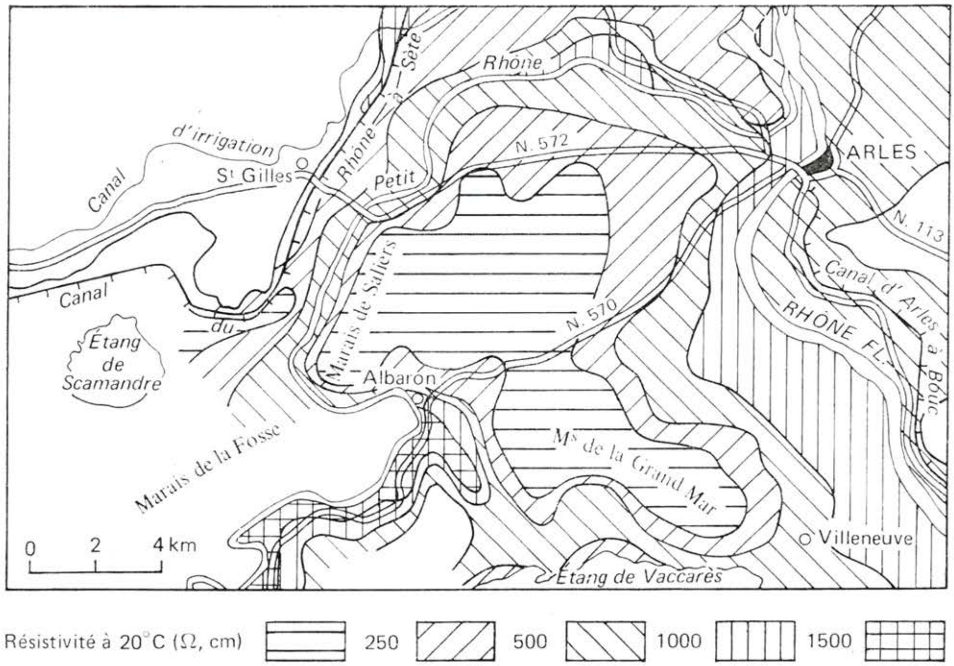


Fig. 4. Carte de la résistivité des eaux

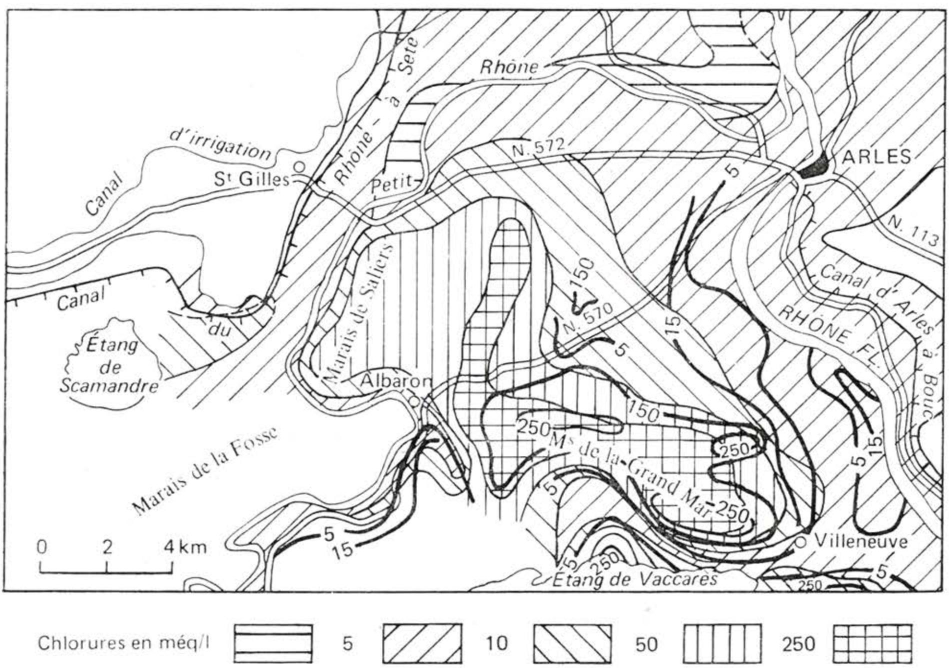
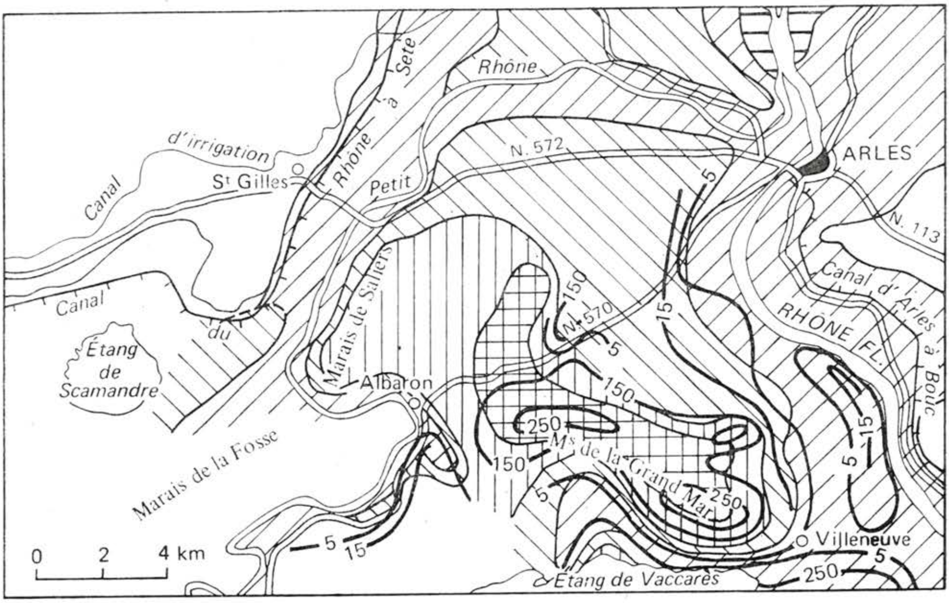


Fig. 5. Carte de la répartition du Cl⁻ (été)



Zonéographie du niveau supérieur (0 à 6 m)

Courbe d'isoteneur en Cl du niveau moyen (–6 à –10 m) en méq/l

Fig. 6. Carte de la répartition du Cl⁻ (automne)

En période estivale, sur les parties hautes irriguées, de l'eau douce surmonte des eaux saumâtres. Dans les zones basses, l'eau gagne la surface par capillarité puis l'atmosphère, évaporation, les concentrations augmentent considérablement et de l'eau salée surmonte de l'eau saumâtre.

En automne, l'irrigation cesse et il y a extension des zones saumâtres. En même temps, l'évaporation diminue et des précipitations ont lieu; on assiste à la régression des zones à forte salure. Le phénomène se poursuit au cours de l'hiver et l'on a partout des eaux moins salées en surface qu'en profondeur.

La nappe captive

Les marnes plaisanciennes qui affleurent le long de la Costière et de la Crau (Fig. 12) forment le soubassement imperméable de toute la plaine camargaise.

1. *Lithologie*

Cet horizon aquifère se compose du faciès sablo-gréseux astien, qui passe progressivement à un faciès plus marneux vers le sud, recouvert par le cailloutis sensu largo composé du Villafranchien et du Flandrien. L'Astien, s'il est absent au Nord de la carte à cause de l'érosion, a une épaisseur croissante vers le Sud, où sa puissance voisine 90 mètres. Quant aux cailloutis, leur épaisseur est variable (de 20 à 70 m).

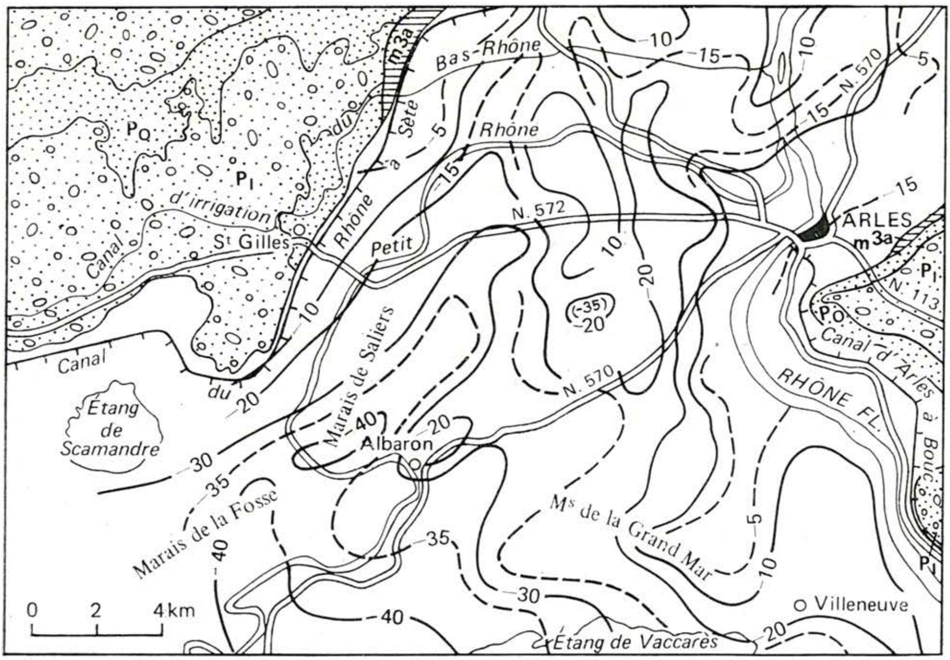
2. *Carte du toit du cailloutis s.l.* (Fig. 7)

Cette carte montre un net pendage vers le Sud (de -5 m au NW d'Arles à -40 m vers Albaron), ainsi qu'une topographie en « banquettes » et « thalwegs » dont le plus important longe la Costière. L'érosion a joué un rôle important dans le modelé de cette surface, mais il est vraisemblable qu'une tectonique plio-quaternaire l'a affecté également.

3. *Caractéristiques hydrauliques*

C'est une nappe sous pression et l'on observe partout un certain artésianisme, non jaillissant au Nord du Petit Rhône, jaillissant de façon temporaire au Sud de celui-ci, et jaillissant en permanence au Nord du Vaccarès. Les débits sont toujours faibles et de l'ordre de 0.2 l/s.

Les transmissivités, de l'ordre de 10^{-2} m^2/s le long des reliefs de la Costière, deviennent moins favorables vers le Sud, où leur valeur est de l'ordre de 10^{-4} m^2/s (Fig. 12), ce qui est dû à des passées de poudingues et de conglomérats dans le cailloutis et au changement de faciès de l'Astien.



Isobathe intermédiaire en m. —15   Isobathe principale en m. —30

Fig. 7. Structure de la surface du toit du cailloutis s.l.

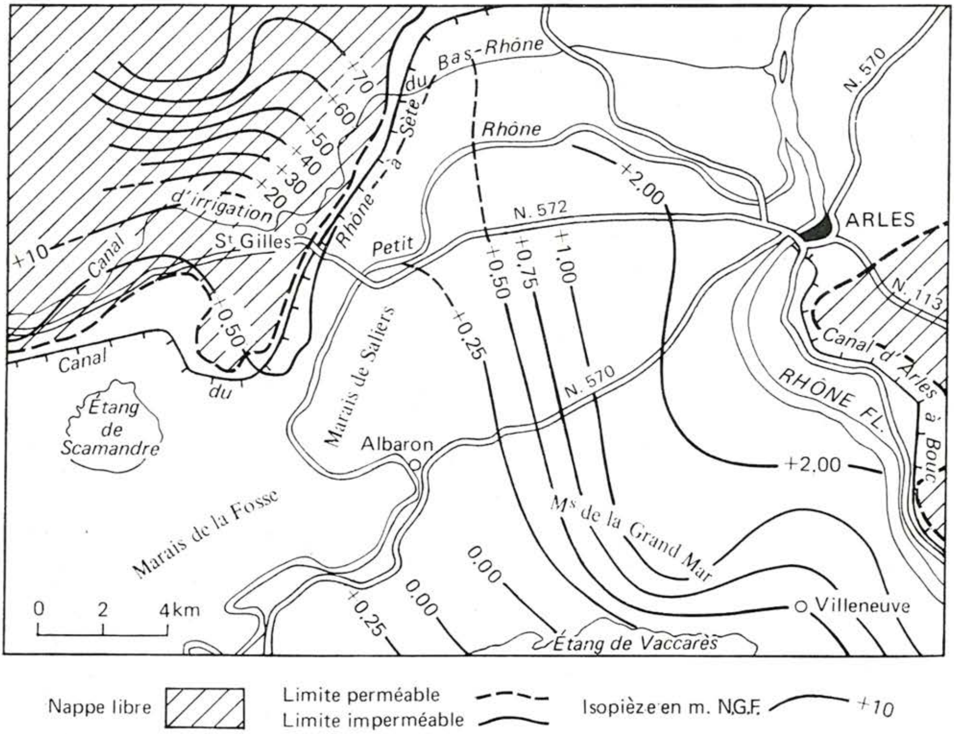


Fig. 8. Carte piézométrique de la nappe captive

4. *Alimentation et écoulement*

L'essentiel de l'alimentation de la nappe se fait en amont de la diramation du Rhône, par infiltration dans le lit du fleuve qui entaille les niveaux de tourbe [6]. Néanmoins, il existe également une alimentation latérale par la Costière et par la Crau avec lesquelles la nappe est en continuité. Son écoulement (Fig. 8) se fait en direction du sud-ouest, avec un gradient de 0.0013 %, et la surface piézométrique montre deux dépressions en correspondance avec les thalwegs de la surface du cailloutis.

5. *Chimie des eaux*

Les eaux sont ferrugineuses et de très mauvaise qualité. La teneur en fer est très variable, mais sa présence est constante partout. Bien que la présence de bactéries sulfato-réductrices ait été signalée [13], il semble qu'il faille plutôt attribuer la présence de fer à l'oxydation de la pyrite très fréquente dans les marnes plaisanciennes sous-jacentes [11].

Les eaux sont chlorurées sodiques, et les deux cartes, l'une des résistivités (Fig. 9), l'autre de la répartition des chlorures (Fig. 10), montrent une avancée en forme de doigt de gant, où les valeurs sont inférieures à 250 ohms/cm et supérieures à 50 méq/l de Cl^-. Le diagramme de PIPER (Fig. 11) traduit l'évolu-

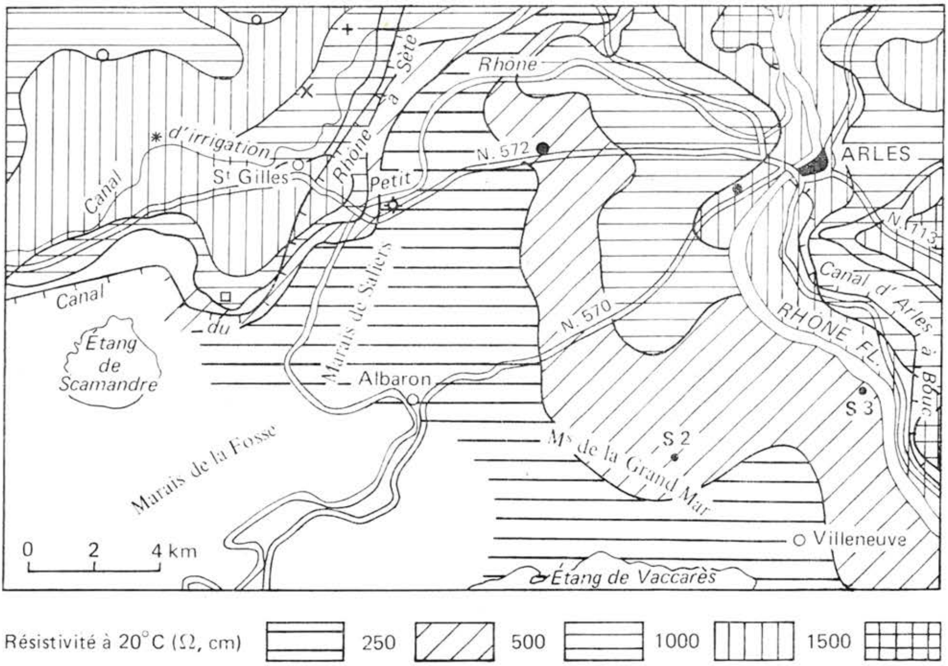


Fig. 9. Carte de la résistivité des eaux

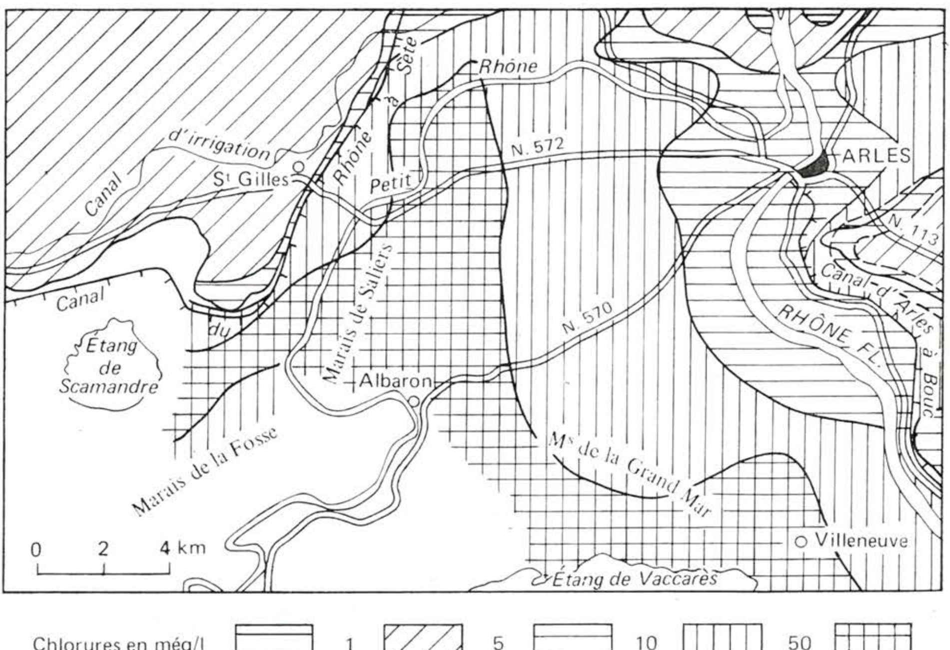


Fig. 10. Carte de la répartition du Cl^-

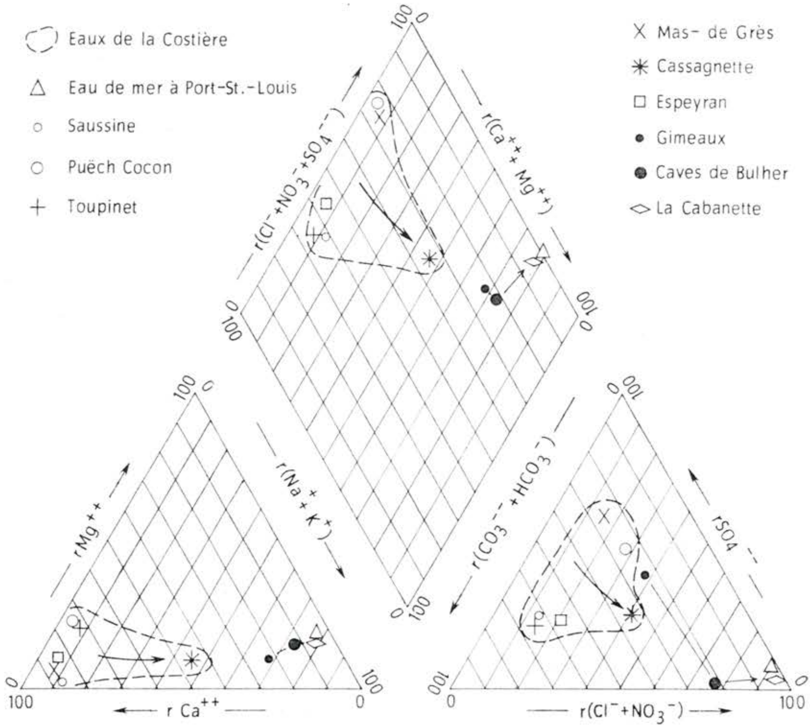


Fig. 11. Diagramme de piper. Évolution des éléments majeurs

tion de l'ensemble des ions majeurs pour les eaux de la Costière d'une part, et pour des points situés à l'est de cette avancée. On constate que, dans les deux sens de l'écoulement, il y a une évolution vers le pôle marin, plus prononcée cependant pour les eaux de la plaine puisque l'eau du forage de la Cabanette se confond pratiquement avec l'eau de mer à Port-Saint-Louis. Néanmoins, la comparaison des analyses chimiques (Tableau 2) montre qu'il ne s'agit pas d'eau de mer. Quant aux infiltrations des eaux du Rhône, elles sont particulièrement bien mises en évidence par une zone, où les résistivités sont comprises entre 1000 et 1500 ohms/cm qui s'étend sur les deux rives. Il en est de même avec les concentrations en Cl^-, avec des valeurs comprises entre 5 et 10 méq/l.

Relations entre les deux aquifères (drainance)

La structure en lentilles des dépôts alluviaux récents et la discontinuité des niveaux de tourbe ne s'opposent pas à l'idée d'une possible communication entre les différents aquifères.

Variation verticale de salinité

Certains piézomètres ont été crépinés dans le cailloutis seul, et on a prélevé des échantillons d'eau à différentes profondeurs. Pour les piézomètres S2 (Pont du Mas-de l'Ange) et S3 (Mas-de l'Ange) ainsi que pour deux autres

Tableau 2

Résultats des analyses chimiques (en méq/l)

	Résistivité ohms/cm à 20 °C	Ca^{++}	Mg^{++}	Na^{+}	K^{+}	Cl^{-}	SO_4^{--}	HCO_3^{-}	CO_3^{--}
1. Eau de mer à Pt.-St.-Louis	25.3	54.86	29.21	537.40	11.07	600.00	15.27	2.16	19.2
2. Saussine	680.0	10.10	2.90	1.20	0.03	4.15	6.00	3.50	0.0
3. Puëch Cocon	940.0	11.40	0.70	0.90	0.04	1.85	7.70	3.50	0.0
4. Toupinet	1110.0	8.65	0.80	0.60	0.03	1.95	2.08	5.70	0.0
5. Mas-de Grès	1460.0	6.25	0.05	0.75	0.00	0.85	1.66	3.90	0.0
6. La Cassagnette	1200.0	6.65	1.85	0.59	0.06	1.45	1.96	6.45	0.0
7. Espeyran	810.0	6.85	1.30	6.50	0.03	7.50	2.40	5.10	0.0
8. Gimeaux	830.0	5.00	2.42	16.48		9.40	9.40	5.60	0.0
9. Caves de Bulher	260.0	5.20	6.20	30.00		32.40	0.05	9.50	0.0
10. La Cabanette	40.0	38.00	66.00	330.00		360.00	27.40	8.00	0.0
11. Eau du Rhône à Saint-Gilles	3300.0	3.00	1.20	0.16		0.56	1.00	2.80	0.0

points hors du périmètre d'étude, la salinité s'est révélée plus forte au toit de l'aquifère.

Salinité de la nappe du cailloutis s.l. à différents niveaux:

Profondeur (m)	*Salinité* S2 (g/l)	*Salinité* S3 (g/l)
15		3.88
18		1.64
21		0.42
24	2.36	
27	2.32	
30	1.87	

Si l'on considère que S2 est implanté dans une zone, où la salinité de la nappe phréatique est très forte (résistivités inférieures à 250 ohms/cm), il est possible que ces variations verticales de salinité soient l'expression de communications entre les deux aquifères.

Variation verticale de densité

Des prélèvements d'échantillons ont été effectués dans des piézomètres couplés implantés, l'un dans la nappe phréatique (court), l'autre dans la nappe captive (long), et ce, à différentes profondeurs. La densité des différents échantillons, fonction de leur teneur en sels dissouts, a été mesurée et les résultats obtenus pour les piézomètres S18 et S21, situés au SE de St-Gilles, ont été reportés dans le tableau 3.

Les résultats obtenus, insuffisants pour être interprétés totalement, montrent cependant qu'en certains points les eaux du toit de la nappe captive sont à la fois moins chargées que celles de la nappe phréatique et que celles des couches sous-jacentes (S21). Ceci tendrait à prouver qu'il n'y a pas, en ce point, de communications entre les deux nappes et que les eaux de la Costière, moins chargées, forment un biseau en repoussant les eaux salines de la plaine (voir Fig. 9). Le fait que l'on trouve parfois, au toit de la nappe captive, des eaux à densité plus faible que celle des eaux de la nappe phréatique, mais plus forte que celle des eaux plus profondes pourrait être un indice de possibles communications.

Mesures de tritium

Le tritium, qui se trouve à l'état naturel dans l'atmosphère, pénètre dans le cycle hydrologique par le biais des précipitations. Cependant la méthode de datation des eaux par le tritium a été faussée par les apports dûs aux expérimentations thermonucléaires. En Europe occidentale, la concentration a montré un premier maximum de 600 UT en 1954, puis un maximum de 460 UT en 1958, et enfin le maximum le plus élevé en 1963 avec une concentration de 3300 UT. Naturellement, les teneurs en tritium sont maximales en été alors que les valeurs hivernales sont les plus basses.

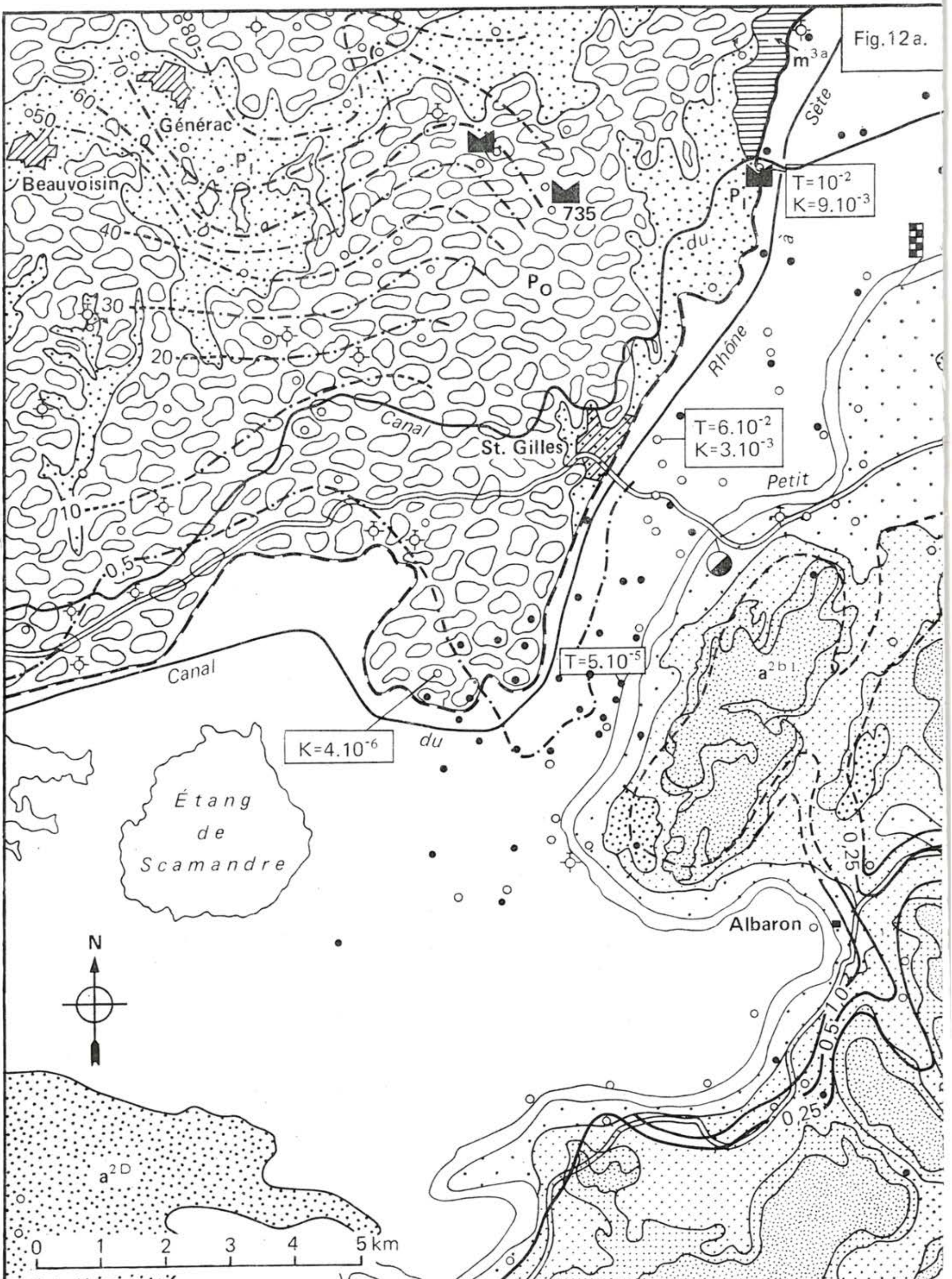
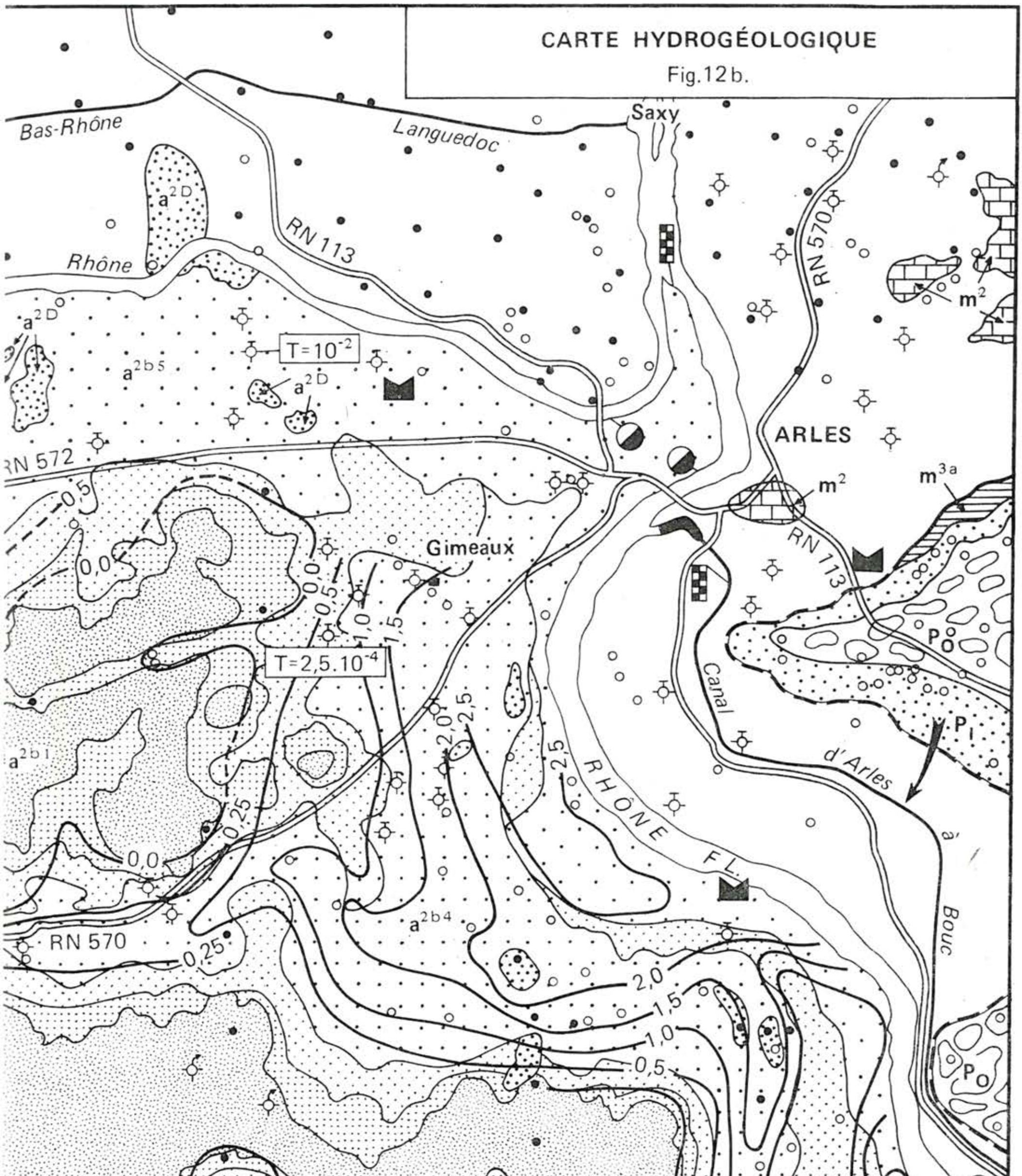


Fig. 12 a.
m^{3a}
Sète
Générac
Beauvoisin
P$_I$
P$_O$
735
T=10^{-2}
K=9.10^{-3}
du
Rhône
à
80
70
60
50
40
30
20
10
0,5
Canal
St. Gilles
T=6.10^{-2}
K=3.10^{-3}
Petit
Canal
T=5.10^{-5}
K=4.10^{-6}
du
a^{2bI}
Étang de Scamandre
Albaron
0,25
0,5
1,0
0,25
N
a^{2D}
0 1 2 3 4 5 km



CARTE HYDROGÉOLOGIQUE
Fig.12b.
Bas-Rhône
Languedoc
Saxy
Rhône
RN 113
RN 570
a^{2D}
a^{2b5}
T=10^{-2}
m^2
ARLES
RN 572
m^{3a}
Gimeaux
T=2,5.10^{-4}
Po
Pl
Canal
d'Arles
à
Bouc
RHÔNE FL.
a^{2b1}
a^{2b4}
RN 570
562
Villeneuve
Mejanes
Étang du Vaccarès
0,0
0,25
0,5
1,0
1,5
2,0
2,5

TERRAIN * (1 | 2 | 3)	Stratigraphie	Lithologie	Hydrogéologie
a^{2b5}	Bourrelet haut du Rhône actuel	Limon argileux à sable limoneux	Alimentàtion: Rhône et irrigation $K=6.10^{-3}$ à 9.10^{-5} cm/s
a^{2b4}	Zone haute des bourrelets anciens	Limon argileux à sable limoneux	$K=6.10^{-3}$ à 9.10^{-5} cm/s
a^{2b3}	Zone moyenne	Limon argileux à limon	$K=3.10^{-3}$ à 9.10^{-5} cm/s
a^{2b2}	Zone basse	Limon argileux à argile limoneuse	$K=2.10^{-3}$ à 0 cm/s
a^{2b1}	Bassin d'inondation (marais)	Argile limoneuse à argile	$K=9.10^{-4}$ à 0 cm/s
a^{2D}	Dunes et anciens cordons littoraux	Sable	Localisé et limité $K=8.10^{-3}$ à 4.10^{-3} cm/s
a^{2a}	Flandrien	Tourbe	Limite imperméable
a^{2a}	Flandrien	Cailloutis frais à matrice sableuse	Seul présent sous la tourbe au nord du Petit Rhône
P_I	Villafranchien	Cailloutis grossier à bancs de conglomérat	En continuité avec la Costière et la Crau
P_O	Astien	Sables grossiers et grès	Forme avec le Villafranchien une nappe captive
m^{3a}	Plaisancien	Marnes bleues	Imperméable
m^2	Burdigalien	Molasse calcaire	Extension limitée Aucun rôle

* 1:à perméabilité de fissure — 2: imperméable — 3:à perméabilité d'interstice

HYDROGRAPHIE

Canal temporaire

Station de jaugeage

Station pluviométrique

Échelle annonce de crue

POINTS D'EAU

Source

Piézomètre

Puits

Forage

Forage artésien non jaillissant

Forage artésien jaillissant

EAUX SOUTERRAINES

Limite imperméable

Limite perméable

30 Courbe isopiézométrique (en m. N.G.F.) (nappe du „cailloutis")

0,5 Courbe isopiézométrique (en m. N.G.F.) (nappe phréatique)

Direction de l'écoulement

Fig. 12c

Les apports en tritium des eaux de pluie au fleuve et/ou à la nappe représentent la recharge annuelle ε_n en relation avec la moyenne pondérée mensuelle des différences $P-E$, où

$P=$ module pluviométrique mensuel,
$E=$ évapotranspiration potentielle mensuelle « Thornthwaite ».

Tableau 3

Densité des eaux de la nappe phréatique et de la nappe captive à différentes profondeurs pour les piézomètres S18 et S21

Piézomètre		Profondeur	Densité
S18 (1re série)	Court	−0.84	1.005
		−1.34	1.010
		−3.34	1.013
		−4.34	1.013
	Long	−13.32	1.005
		−24.32	1.003
S18 (2e série)	Court	−0.84	1.008
		−1.34	1.007
		−3.34	1.010
	Long	−13.32	1.008
		−18.32	1.008
		−23.32	1.006
		−24.32	1.006
S21	Court	−3.40	1.002
	Long	−12.96	1.000
		−22.96	1.000
		−24.46	1.005

On a ainsi calculé l'âge des eaux du Rhône à Pierrelatte [2] 100 km en amont d'Arles. Les teneurs en tritium des eaux de pluie sont connues depuis 1964, grâce à quoi il a été possible de calculer ε_n.

Teneurs en tritium des eaux de pluie à Pierrelatte (moyennes annuelles) :

1964	*1965*	*1966*	*1967*	*1968*	*1969*	*1970*
1444	511	223	155	357	333	444

D'après le modèle établit par P. Hubert et al. (1970), la valeur théorique tritium N'_n de la nappe et/ou du cours d'eau s'établit :

$$N'_n = \sum_{p=0}^{p=\infty} \cdot \alpha \cdot (\mathrm{I}-\alpha)^p \cdot \lambda^p \cdot \varepsilon_{n-p},$$

ou α = coefficient de mélange (rapport du mélange entre la recharge de l'année et les eaux des années précédentes),
λ = constante radioactive (= 0,0547 pour le tritium),
ε_n = recharge de l'année n.

La comparaison des valeurs théoriques N'_n et des teneurs réelles du Rhône sur trois ans amène à choisir la valeur $\alpha = 0.45$ (Tableau 4).

Tableau 4

Comparaison des valeurs théoriques et mesurées du Rhône à Pierrelatte [2]

Période	UT mesurée	UT $\alpha=0.40$	UT $\alpha=0.45$	UT $\alpha=0.50$
1968	217	250	230	210
1969	195	200	190	170
1970	158	170	160	150

L'âge moyen t de l'eau est fonction du coefficient de mélange α (PH. OLIVE, 1970):

$$t=\frac{2-\alpha}{2\alpha}$$

D'où l'âge de l'eau du Rhône, compris entre 1 an et demi et 2 ans. Ces valeurs sont à rapprocher des teneurs en tritium des eaux de la Crau, du Rhône en Arles et du forage du Mas-de l'Ange en période estivale (voir Tableau 5). Les analyses reportées dans ce tableau montrent que la nappe captive de la Haute

Tableau 5

Teneur en tritium des eaux de la Crau et de la Camargue (période estivale)

	Crau	Camargue	Rhône
Lieux	Castelflaure 325	S1 286	> 500
	Chanoines 185	S2 < 50	
	La Crosse du Couchant 300	S3 < 50	
		S4 < 50	
		Aubépines 50	

et Moyenne Camargue est alimentée par le fleuve, au moins au niveau du seuil de Terrin.

D'autre part, les faibles teneurs trouvées au nord du Petit Rhône et dans la région du Vaccarès prouvent que l'écoulement y est très lent. La période du tritium étant égale à 12.3 ans, il faudrait environs 35 ans pour que l'influence des eaux du Rhône arrive en S2 (Pont du Mas-de l'Ange).

Conclusions

Il existe en Camargue deux systèmes aquifères bien distincts : la nappe alluviale phréatique et la nappe captive du cailloutis s.l. La nappe phréatique, alimentée essentiellement par l'eau d'irrigation, a un écoulement vers des zones basses, où la perméabilité des sols est plus faible. L'eau stagne, gagne l'atmosphère par évaporation et ont lieu des phénomènes de concentration en sel, qui se traduisent parfois par des dépôts.

La nappe captive est alimentée par le Rhône et la Crau, mais son écoulement est très lent. De ce fait, les eaux marines n'ont pas été entièrement lessivées. Les eaux, qui sont par ailleurs de qualité médiocre, sont aussi chargées en fer.

Enfin, il est possible que des communications entre les deux systèmes existent, sans que ces phénomènes affectent l'individualité de chaque aquifère.

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EXPERIENCES OF THE OBSERVATION OF CONFINED AQUIFERS AT DUNAÚJVÁROS

EXPÉRIENCES D'OBSERVATIONS D'AQUIFÈRES CAPTIFS A DUNAÚJVÁROS

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RÉSUMÉ

Dans la région de Dunaújváros, il y avait des tâches spéciales en ce qui concerne l'observation et l'analyse des nappes phréatiques, parce que les hauts bords des rivières formés de « loess » subissent des mouvements fréquents, cela met en danger l'agglomération, et rend nécessaire des mesures de précaution préliminaires. Pour l'analyse des nappes phréatiques, il fallait d'abord déterminer punctuellement les niveaux poreux dans les diverses couches formées d'alluvions à grains fins, où il fallait organiser un système d'observations du changement de pression des nappes phréatiques. Cette tâche est à accomplir sur les hauts bords de la rivière, où les nappes phréatiques sont disposées à 20—50 mètres en-dessous du niveau du terrain, et sur le bord du Danube, où il fallait déterminer le mouvement des nappes phréatiques près du niveau du terrain.

Les données rendent possible de déterminer les caractéristiques géohydrologiques de la région, les changements de pression de nappes phréatiques ont des tendances contraires si l'on observe les puits sur les hauts bords et près du Danube. En analysant les séries de données ainsi régistrées, on peut tirer la conclusion que les différences de pression des nappes phréatiques atteignent plusieurs mètres et ces différences de pression sont en relation étroite avec les conditions extérieures, surtout avec le niveau du Danube. Ce rapport entre les changements de pression des nappes phréatiques et le niveau du Danube peut être observé près du bord du fleuve et la haute rive, malgré qu'il y a une tendance inverse entre la profondeur des nappes phréatiques et leur niveau normal.

Comme les relations entre la pression des nappes phréatiques et des conditions extérieures est plus proche comme nous l'avons supposé, il faut généraliser les éléments observés pour mieux connaître les conditions d'économie hydrologique de la région.

Introduction

On the high banks along the Danube belonging to the study area landsliding hazard influences the development of the towns and villages near to the bank. It can damage the existing structures, it disturbs the traffic and the ice movement on the river.

In order to eliminate the movements of the high bank of the Danube and to choose the most expedient sorts of the necessary bank protecting structures and works, in 1965—66 widespread explorations were made in the area of Dunaújváros.

General engineering geological and hydrogeological characteristics

The frequent movements of the high banks of the Danube can be considered as recent denudation processes caused by different natural effects. This natural process can be influenced considerably by human activity. The investigations proved the bank movements are caused by several factors among which the hydrogeological features have a decisive role. Therefore the investigations referred mostly to the knowledge of the hydrological conditions and regularities.

According to the observations and explorations in the geological structure of the bank section Upper Pannonian and Pleistocene sediments take part. The thickness of the Pleistocene formations – which build the actual high bank – varies between 30 and 50 m. The series of layers consists mostly of loess and from its modified variations. In the loess building up the high bank ground water can be found. In the proximity of the edge of the bank the ground water level is strongly inclined and the water reaches the surface in the form of springs at the base of the bank. During the last years the waters escaping from the public works caused a ground water level increase of several metres.

The explored section (50 – 60 m) of the underlying Upper Pannonian strata complex consists of the alternation of sand, clay and silt layers and it contains confined waters. Down to the explored depth depending on the geologic conditions even 3 – 6 water bearing sand layers can occur.

Exploration method and completion of the observation wells

It was necessary to bore a lot of test- and observation wells in the area of Dunaújváros. The hydrogeological investigations covered an area of about 5 km length and of 3 – 4 km width, within which the borings and the observations concentrated to the section between the embankment and the Danube. Moreover on single critical sections the observation wells had to be bored in a dense pattern (Fig. 1).

The borings were made by the dry method and in consequence of the alternating series of layers a manifold casing string became necessary. The dry boring gave a reliable sampling as result, which was made complete by the dense, undisturbed sampling. Beyond the location of the layers and the stratification in the research borings every important water bearing layer was separately controlled by running in a test screen and with different methods the permeability was determined too. The formation tests were made parallelly with the drilling. The test drilling of small diameter made the reliable exploration of the formation boundaries possible. In the exploratory boreholes 2 – 4 formation tests were made in each. When reaching the projected depth of the borehole, parallelly with the pulling out of the pipes the positioning of the single members of the observation group of wells started, as well as the permanent separation of the porous layer with filling up and cementing (Fig. 2).

According to the investigations in starting soil creepings both the ground water and the confined waters presenting themselves in more successive layers have a considerable role. Therefore observation wells were completed so that all water bearing layers could be observed.

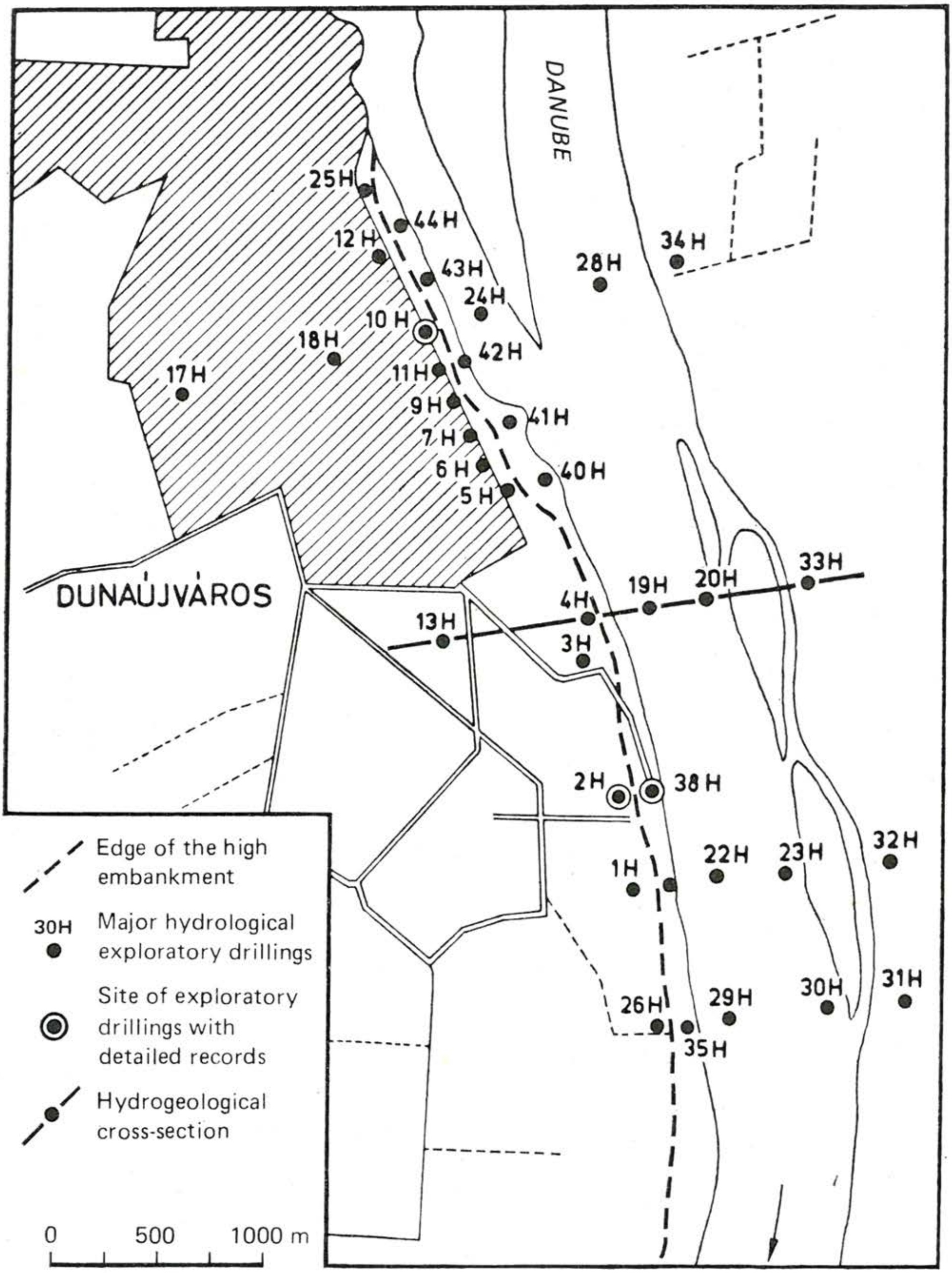

Fig. 1. An overlook drawing about the research area in Dunaújváros

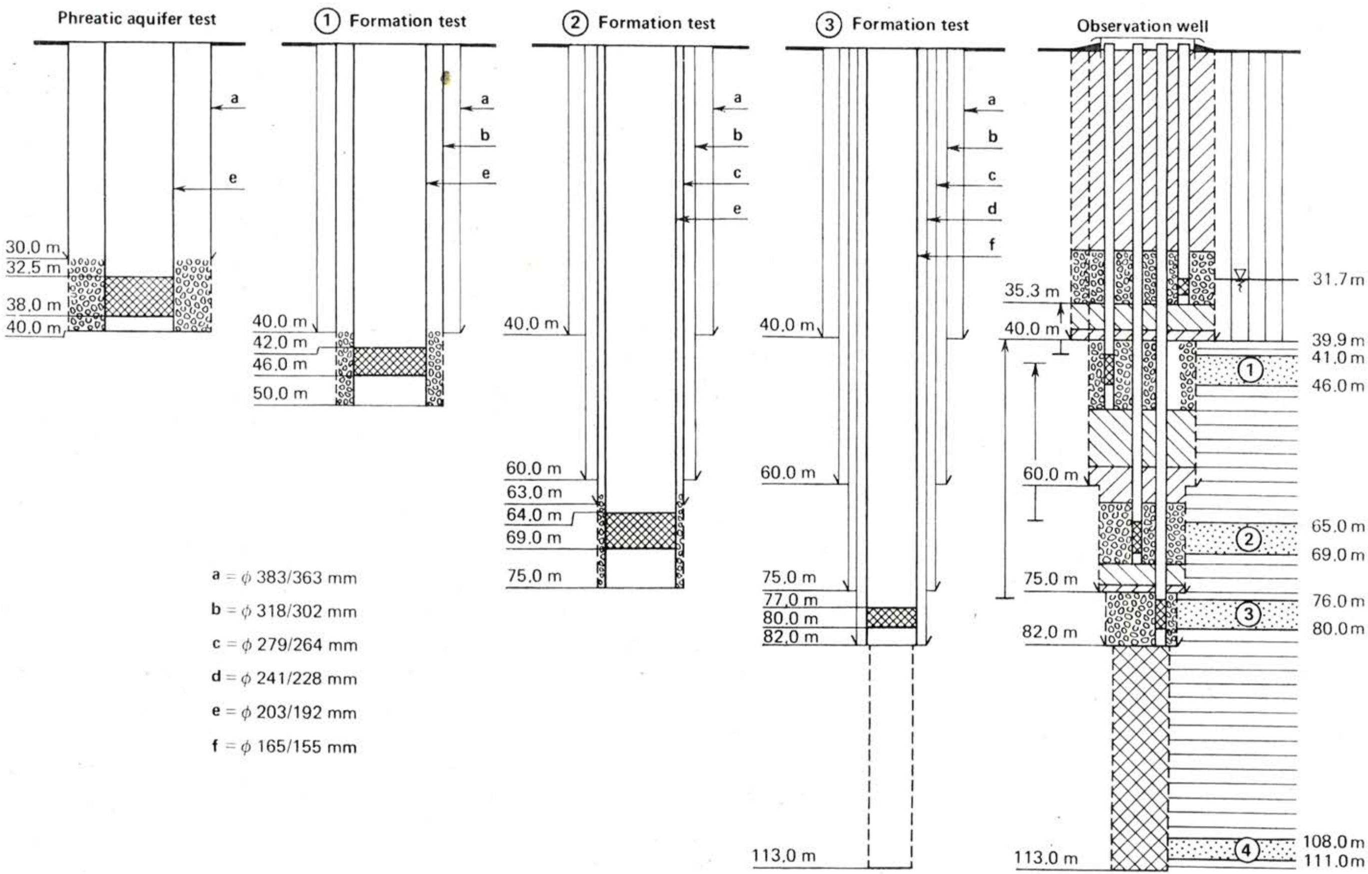

Fig. 2. Method of the construction of test and observation wells

In the interest of the manifold formation test and the completion of the observation wells (70 – 100 m) the emplacement of casings of greater diameter, which was necessary, had to be ensured. It was necessary because of the graveling in order to hinder the block up of the screen in the layer consisting of very fine-grained sands around the screen.

In the interest of the reconstruction of the natural pressure state of the single sand layers, as well as for the elimination of the effect of other layers the desired effect could be reached with a cementing of 2 – 3 m thickness. This solution could be applied only then, when between two layers the distance was minimum 4 m.

The single water bearing sand layers were cased by zinc-plated tubes of 2″ diameter with a proper screen installation according to the thickness of the layer. Each observation tube was placed on a cement base to hinder later movements.

Contrary to the high coast with the observation wells on the Danube riverside little modifications had to be made. Here because of the water levels above the ground sometimes a protecting tube longer as 6 m was built in and at the observation pipes the water level fluctuation caused by the Danube had to be taken into consideration too. For measuring of the static water levels above the ground surface on the protecting tube rungs of ladder were welded.

The above informed exploration method and the completion of an observation well type suitable for the observation for more layers worked well in the area of Dunaújváros. The more than 100 wells supplied undisturbed and safe all those data, for which they were bored.

Exploration and observation results

From the exploration data we had the picture that the porous Upper Pannonian layers under the high bank extend without any important change towards east. The undisturbed occurrence of the porous layers indicates that young tectonic movements did not disturb these layers and did not cause considerable modification, or gap of continuity. In the development of the piezometric surfaces a special tendency was expressed, with a different character on the high bank and on the riverside. On the sliding areas in many cases the transitory character became evaluable (Fig. 3).

With the revaluation of the hydrogeological data the related elevation of the explored layers and of their water level shows well the changing direction of the pressure change under the high bank and the riverside of the Danube, further the transitory character of the pressure change on the rubble slope. In evaluating the overpressure of the confined waters in the exploratory boreholes of the plateau and of the riverside of the Danube it could be stated that the formation pressure is not primarily determined by the reservoir pressure, but by other circumstances. Among the influencing circumstances we have to mention the local effect of landslides closing the single layers (Fig. 4).

The observation of the processes concerning the confined waters was based on the recognition that the connection of the confined waters and of the waters on the surface is unambiguous on this area. The attention for the necessity of this was called by the experience of the confined water observation

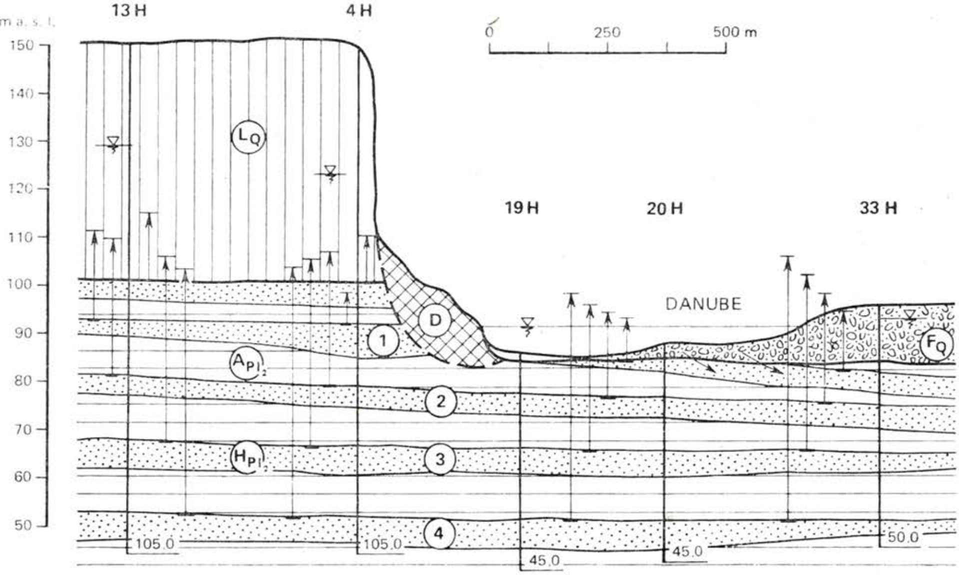

Fig. 3. Characteristic hydrogeological section in the area of Dunaújváros

effected in the frames of the Hungarian Geological Institute, from which it was known that even in closed systems a considerable change of pressure can be observed in confined waters and this can be related to the outer conditions. The observations of this kind of the HGI were completed by newer results, among others with the control of the hydrodynamic system of the Pannonian Basin.

The data series of 8 years of the observation of the ground and confined waters in the area of Dunaújváros shows a considerable fluctuation of the piezometric surface of the confined waters. There is a close connection between the water stage of the Danube and the piezometric surface of the confined waters (Fig. 5). In the observation well on the *high bank* the piezometric surface of the confined water No 1 showed an extreme variation of 2.65 m, while during this time the difference of the maximum and of the minimum water level reached similarly 2.45 m. In the layer No 1 the greatest change of the water level during 1 – 1 period reached 1.45 m, while in the layer No 2 it reached 1.95 m. During this time with the layer No 1 the number of the water level changes exceeding 1 m and occurring within one period – considering the elevating branches – is 3, the same could be found in the case of the layer No 2 four times (Fig. 6). The evaluation shows the change of the ground water level too. As a reason for the landslide on the bank a considerable increase of the ground water level occurred until 1965, so the bank protection elaborated a solution for the lowering of the ground water level too. The 8 years formation of the ground water level shows unambiguously that the slow subsidence of the water level occurred on this area. This increased values of the pressure change of the confined waters were supplied by the observation well system on the *riverside of the Danube* (Fig. 7). The change is in a close connection here too with the water stage of the Danube. On this place the maximum change of the water level of the layers No 1 and 2 – of course in an always different interval – proved to be close to 3.95 m. Within a period – examining the elevation side too – the greatest change of the water level was 3.95 m with the layer No 1 and 2.75 m in the layer No 2. The water level elevation more than 2 m could be observed in the layer No 1 once and in No 2 four times (Fig. 8).

Conclusions

a) The observations effected in the last decade in connection with the pressure change of confined waters show unambiguously that their change exceeds the previously supposed measure on the one side and can be related to the outer circumstances on the other side.

b) The engineering geological researches in the area of Dunaújváros supplied informations needed for the protection and they gave a picture about the piezometric surfaces of the confined waters.

c) With regard to the complex tasks such an exploration method could be developed, which ensured the necessary informations and at the same time it made economically possible to transform the exploration boreholes to a group of wells for water level observation.

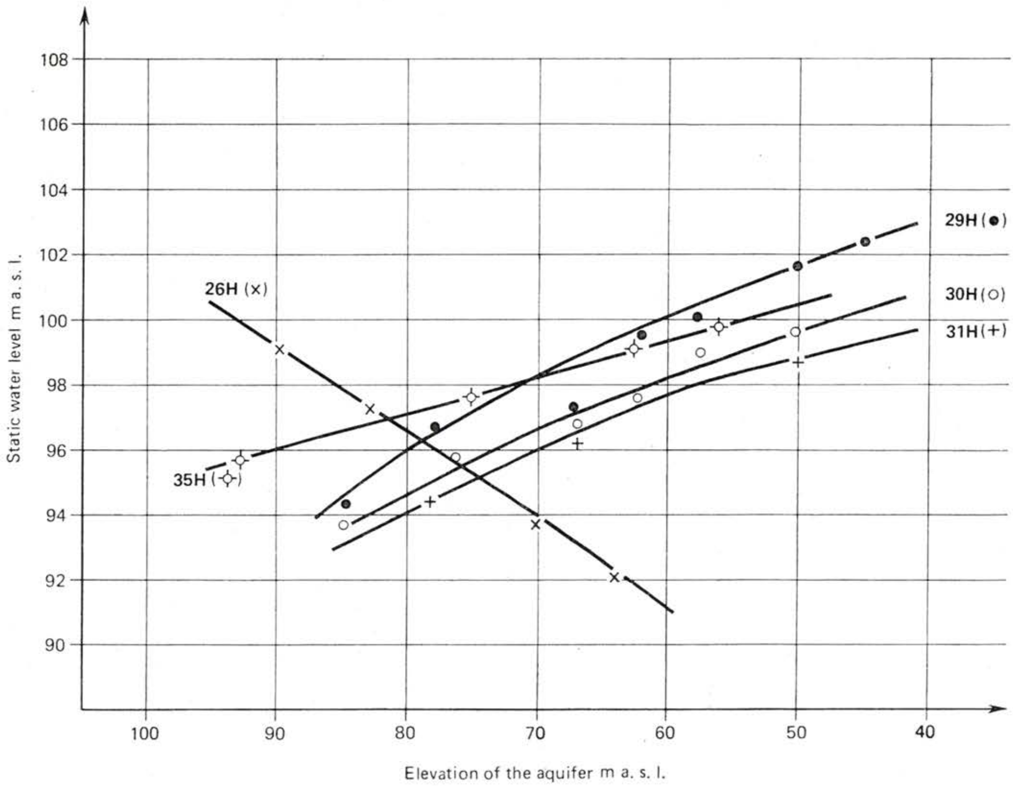

Fig. 4. Connection between the depth position of the water bearing layers and the piezometric surface

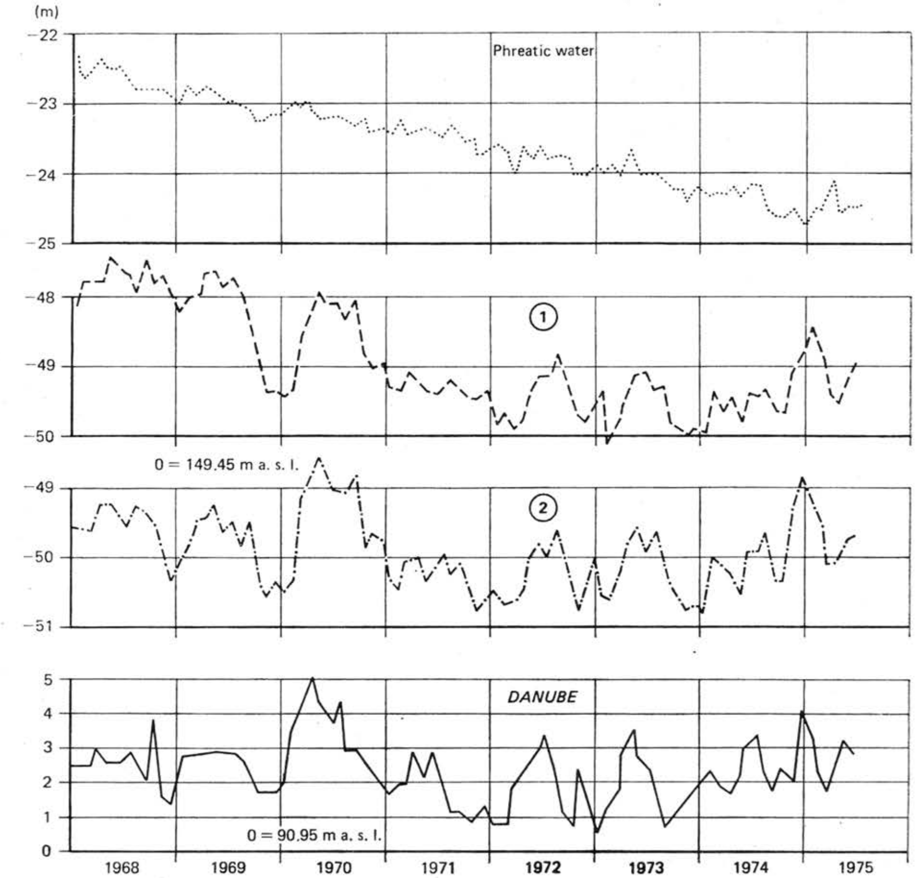

Fig. 5. Observation data line of a high bank observation well

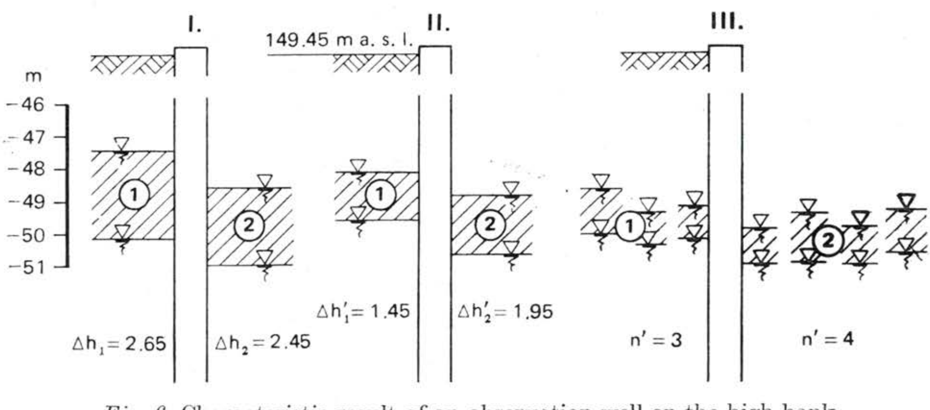

Fig. 6. Characteristic result of an observation well on the high bank

1, 2 = water bearing layer, I = the maximum of the variation, II = maximum variation of a period, III = the fluctuations and their number (n) beyond 2 m

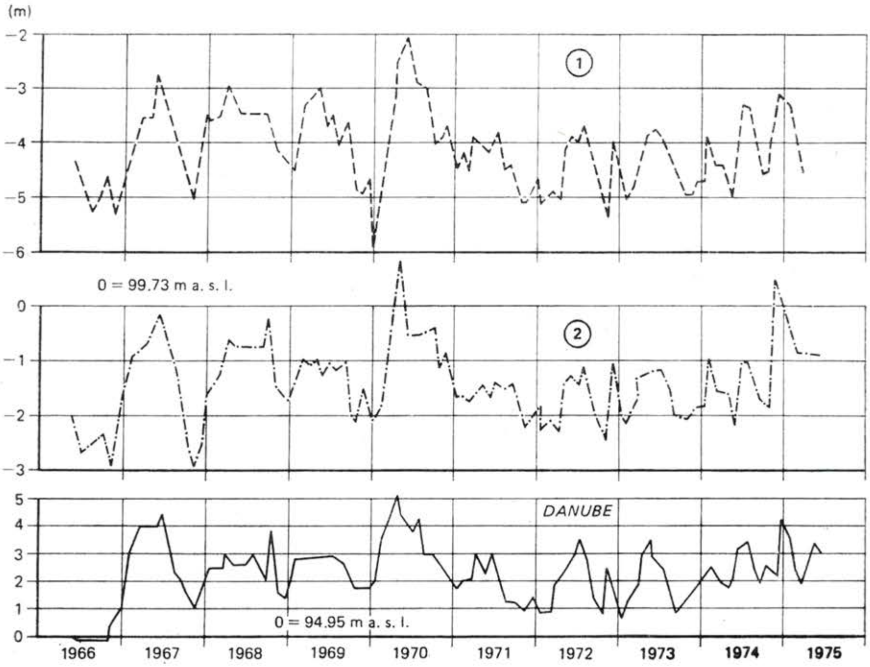

Fig. 7. Water level fluctuation of an observation well on the riverside of the Danube

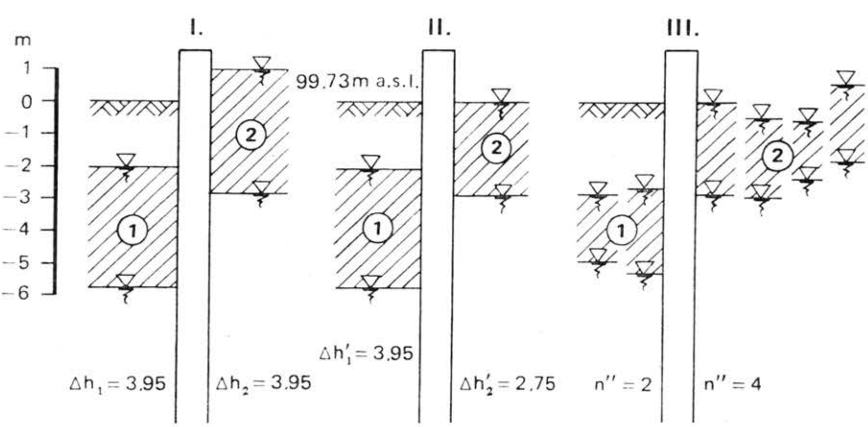

Fig. 8. Characteristic result of an observation well on the riverside of the Danube

d) The observation wells thus completed followed the change of the formation pressure with corresponding sensibility during the last 8 years.

e) The observation of the water level on the given area is more difficult than usually. The task is to observe water levels, which are 20 – 50 m under the ground surface on the high bank and which are of transitory character on the Danube riverside.

f) From the results of the water level observation during 8 years it can be stated that in the Upper Pannonian confined waters a change of water level exceeding several meters may exist. The piezometric surface of these confined waters shows a close connection with the variation of the stage of the Danube. The change of the water level extends both to the observation wells on the Danube riverside and on the high bank in spite of the fact that at the time of the starting investigations and in the observation period the pressure value of the confined waters shows an opposite tendency on both areas.

g) A comprehensive generalization of the observations concerning the pressure change of the confined waters would be highly desirable. As a result, the outer circumstances influencing the pressure change will be shown much more reliably, also the possibility of the effective and of the potential water recharge as a final result.

CARACTÈRE DES EAUX MINÉRALES AU NORD-EST DE LA POLOGNE

CHARACTER OF MINERAL WATERS IN THE NORTH-EASTERN PART OF POLAND

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ABSTRACT

The north-eastern part of Poland belongs to different structural elements. The most evident ones are: the elevated part of Precambrian platform, the Kujavian-Pomeranian elevation and a depression dividing these elevated structures from the Peri-Baltic syneclise. The horizon separating the mineral and fresh waters is found in this area at depths from 0 m in the SW to more than 1200 m in the SE. This hanging wall occurs in different stratigraphic units: in the Quaternary, Tertiary, Cretaceous, Jurassic and in the SE as low as in Cambrian formations. The mineralization of water grows below the separating surface rather quickly, both at shallow position of mineral waters and at deeper one. The shallowest occurrence of mineral waters is connected with tectonically disturbed salt deposits. Their deepest position is met in limited spaces, within the basin structures.

On the basis of the frequency of certain chemical features of the mineral waters closed to the surface the typical content is determined, the deviation from it for the particular ions being marked on the map. The Mg content did not reveal noticeable deviations. Taking into consideration low water mineralization, typical values were chosen at large intervals, and so: Cl — 50—90, SO_4 — below 20, HCO_3 — 10—50, Na — above 75, Ca and Mg below 20% mvals.

At a general predominance of Cl-Na waters, local considerable deviations are noted, which can be connected with gypsum- and sulphide deposits (SO_4), or with the differentiation in the intensity of water circulation. Some anomalies may be treated as the reflexions of present and past tectonic processes.

Introduction

La situation du « toit » des eaux minérales, c'est-à-dire de la surface qui les sépare des eaux douces devient la question qui gagne dernièrement plus d'importance, surtout en ce qui concerne l'évaluation des ressources en eaux souterraines et leur protection. Le caractère des eaux minérales dans la zone de leur surface constitue un des éléments de la prévision de qualité des eaux minérales plus profondes. Il peut permettre d'indiquer la présence de structures « cachées » et les lignes des dislocations.

Les valeurs de la minéralisation au dessus desquelles on considère les eaux comme minérales varie entre 1 et 2 g/l. Nous admettons plutôt la seconde limite, particulièrement là, où l'ion Cl^- ne prédomine pas distinctement parmi les anions.

Afin d'établir la profondeur de la surface des eaux minérales sur le territoire étudié on a utilisé 63 points de base (forages) sans compter les endroits, où la salinisation des eaux est observée à proximité du niveau du sol. D'entre ce nombre de points de base, seuls 47 ont fourni des données se rapportant au contenu des eaux. En déterminant le caractère des eaux et la profondeur de leur minéralisation originale on a appliqué l'inter- et extrapolation.

Conditions géologiques

Les plus grandes différences de la profondeur des eaux minérales en Pologne ont été constatées dans sa partie nord-est. La superficie sur laquelle porte ce communiqué est d'environ 100.000 km².

Au point de vue de la géomorphologie, ce sont des terrains à faible relief, un peu variés grâce aux collines morainiques de la dernière glaciation. On y trouve quelques unités tectoniques de premier ordre auxquelles du côté nord-est contiguë la vaste plate-forme précambrienne. Sa partie élevée (Fig. 1, I) se distingue par une mince série sédimentaire dont la puissance n'est pas quand même inférieure à 400 m. Autour de cette élévation le socle cristallin est situé à une profondeur plus importante. Vers le Nord c'est la synéclise péribaltique, vers l'Ouest le synclinorium marginal, en dehors duquel s'étend la plate-forme paléozoïque. De celle-ci fait partie, sur notre territoire, un fragment de l'élévation de Kuyavie-Poméranie (Fig. 1, II) provenant de l'orogénèse laramienne. Une structure du même âge, celle du bassin de Masovie, recouvre en partie les deux plates-formes précitées et le synclinorium marginal. Le socle cristallin s'abaisse à la limite du synclinorium à une profondeur allant jusqu'à 5 km, et ne cesse pas de s'approfondir vers le Sud-Ouest. De cette façon tout notre territoire, sauf la partie élevée de la plate-forme précambrienne, appartient aux structures du type de bassin.

Les unités structurales de différent âge ne se manifestent pas à la surface du sol, parce qu'elles sont couvertes de formations cénozoïques incohérentes. Ce sont les sédiments sableux et argileux du Tertiaire, d'épaisseur de 0 à environ 250 m, et les dépôts quaternaires — glaciaires, fluvio-glaciaires, fluvio-lacustres et fluviatils, dont l'épaisseur varie de quelques dizaines à plus de cent mètres. Sur la carte (Fig. 1) on a indiqué les terrains, où la couverture du Quaternaire dépasse 100 m. On y peut trouver aussi la répartition continue des argiles du Pliocène. Cette couche argileuse sépare les aquifères du Quaternaire de ceux du Tertiaire; elle ne constitue pas cependant la limite des eaux douces.

« Toit » des eaux minérales

La surface des eaux minérales se situe sur le territoire en question à une profondeur de 0 m à partir du niveau du sol (environ +250 m par rapport au niveau de la mer) jusqu'à plus de 1200 m (−1000 m), ce qui constitue une des plus grandes profondeurs en Pologne. Au Sud-Ouest, les deux extrémités se rencontrent à une distance de 20 km à peine; au Nord de Lódź, les eaux salées apparaissent tout près du niveau du sol, et autour de cette grande ville leur surface s'abaisse pour y atteindre une profondeur de plus de 1000 m (−700 m). La zone de grande profondeur des eaux minérales se prolonge vers l'Est. Au Sud-Est, l'épaisseur des eaux douces dépasse 1200 m dans la cuvette de Podlasie. Au Nord-Est, une certaine région est entièrement dépourvue d'eaux minérales, dans la couverture sédimentaire.

La répartition des eaux minérales indique sur notre territoire, particulièrement au Sud-Ouest, leur surface (« toit ») très variée; à l'exclusion de cette région on rencontre les eaux salées en principe à une profondeur de 200 à 700 m (de −400 à −900 m par rapport au niveau de la mer).

La différenciation de la profondeur du « toit » des eaux minérales est liée aux divers facteurs géologiques, bien que l'interprétation valable ne soit

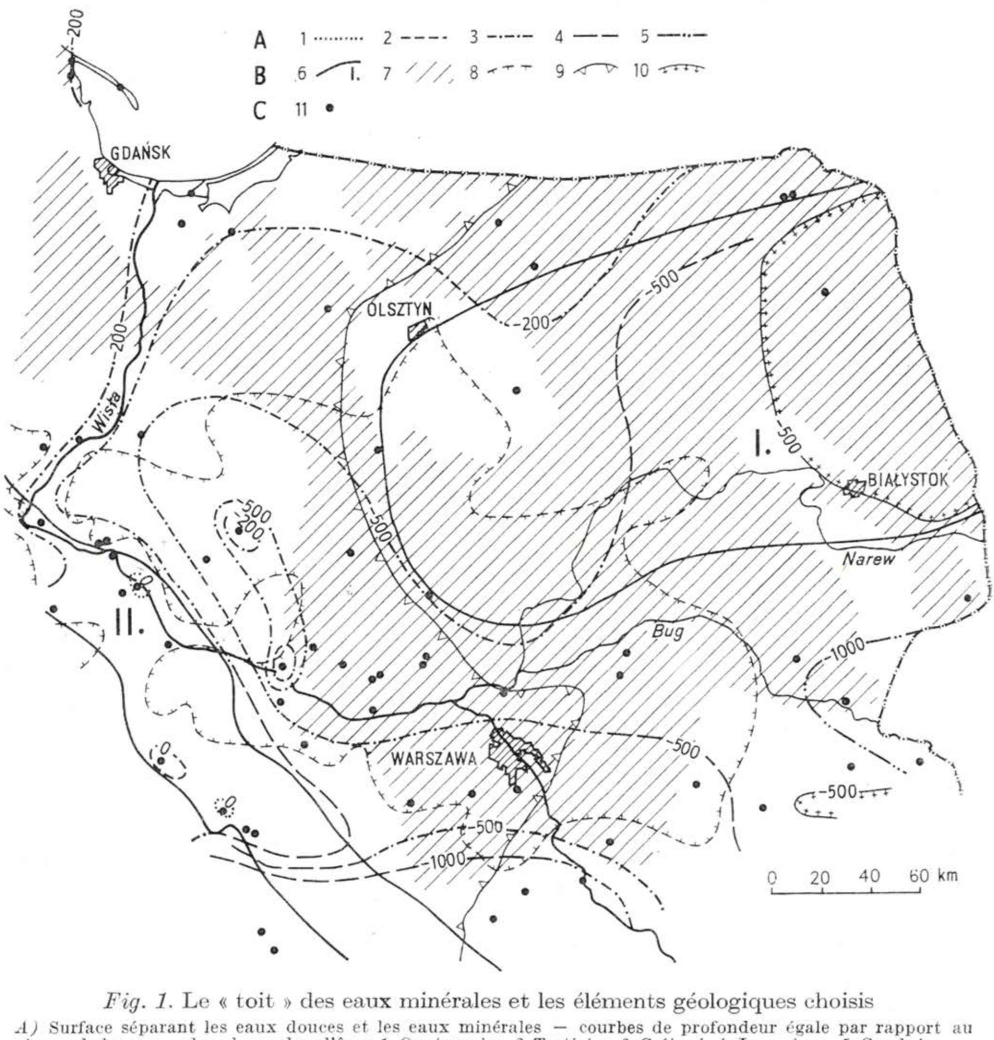

Fig. 1. Le « toit » des eaux minérales et les éléments géologiques choisis

A) Surface séparant les eaux douces et les eaux minérales — courbes de profondeur égale par rapport au niveau de la mer — dans les roches d'âge: *1.* Quaternaire, *2.* Tertiaire, *3.* Crétacé, *4.* Jurassique, *5.* Cambrien. — *B) 6.* Structures élevées: I — élévation de Mazurie-Suwalki, II — anticlinorium de Kuyavie-Poméranie. *7.* Terrains où la puissance des sédiments du Quaternaire est supérieure à 100 m. *8.* Limites de la formation argileuse du Pliocène, qui sépare les aquifères du Quaternaire de ceux du Tertiaire. *9.* Limite des roches salifères du Zechstein. *10.* Terrain, où le socle cristallin se trouve à une profondeur supérieure à 500 au dessous du niveau du sol. — *C) 11.* Points (forages) de base pas toujours possible. Ce sont les régions de formations salifères chahutées du Zechstein qui jouent ici un rôle important (Fig. 1). Il est nécessaire de mentionner que les roches salifères d'autre âge n'y sont pas reconnues. La désalinisation de certaines parties des bassins résultent probablement des processus paléohydrogéologiques, en liaison avec la paléotectonique. Comme les zones, où la différence de la profondeur de la surface des eaux minérales est de l'ordre de 1000 m, se trouvent l'une près de l'autre, on peut constater l'influence d'une récente activisation de la circulation des eaux souterraines, c'est-à-dire du « rinçage » de certains terrains, même dans le voisinage des écrans hydrochimiques qui séparent ces régions de celles, où on observe la salinisation près du niveau du sol.

Changement vertical de la minéralisation des eaux

L'accroissement de la minéralisation des eaux au dessous de leur surface est souvent assez rapide, comme cela se manifeste dans les diagrammes (Fig. 2). Les points indiqués au moyen des numéros 1, 5, 6, 7, 9 peuvent y servir d'exemple. La minéralisation de l'eau augmente jusqu'à 10 g/l plus lentement aux points 2, 3, 4, 10. Le dernier point est une exception, car ce n'est qu'à la distance de 1000 m, où la teneur monte de 1 à 10 g/l. Le diagramme no 11, caractérise une situation pareille et une réduction instantanée de la minéralisation de l'eau dépendamment de la profondeur (« inversion » hydrochimique) n'y est past probablement un phénomène naturel.

Écarts de la chimie des eaux minérales

En disposant des données hydrochimiques de 47 points on a essayé de déterminer un type chimique des eaux minérales à leur surface. On s'est décidé à admettre les vastes intervalles pour les eaux typiques, vu que les résultats des analyses ne concernent pas souvent la minéralisation considérée strictement en tant que limite des eaux minérales. Il ne faut pas non plus oublier de l'instabilité chimique qui caractérise les eaux naturelles à faible concentration.

Pour les ions particuliers on a admis les intervalles suivants: Cl^- — de 50 à 90 % miliéquivalents, bien que les valeurs les plus fréquentes s'enferment dans les limites 50 à 75 %; SO_4^{2-} — au dessous de 20 % mé (le plus souvent au dessous de 10 %); HCO_3^- — de 10 à 50 %, les valeurs au dessous de 10 % n'étant pas rares. Parmi les cations, Na^+ se manifeste par des valeurs typiques au dessus de 75 % mé, le plus souvent jusqu'à une cinquième de points contient 50 à 75 % de cet ion. Les ions Ca^{2+} et Mg^{2+} ne dépassent pas en principe 20 % et souvent même 10 %; Ca^{2+} est sporadiquement observé dans l'intervalle 20 à 50 %.

Pour les ions principaux on a indiqué sur la carte (Fig. 2) les écarts qui dépassent ces vastes intervalles. On peut facilement constater que les ions Cl^- et HCO_3^- se distinguent par des écarts allant dans deux directions c'est-à-dire au dessus et au dessous des intervalles susmentionnés, SO_4^{2-} alors que tous les trois cations — seulement dans un sens unique.

La répartition des eaux minérales « typiques » au sens conventionnel de ce mot, et de celles caractérisées par les écarts, n'est past régulière au point de vue régional. Plus grandes valeurs de Cl^- apparaissent aussi bien dans les zones de « toit » des eaux minérales profond que là, où leur salinisation commence à se manifester tout près de la surface du sol. L'ion SO_4^{2-} est lié généralement aux couches gypsifères et aux couches abondantes en sulfures. La teneur inférieure à 50 % est constatée à deux points; les

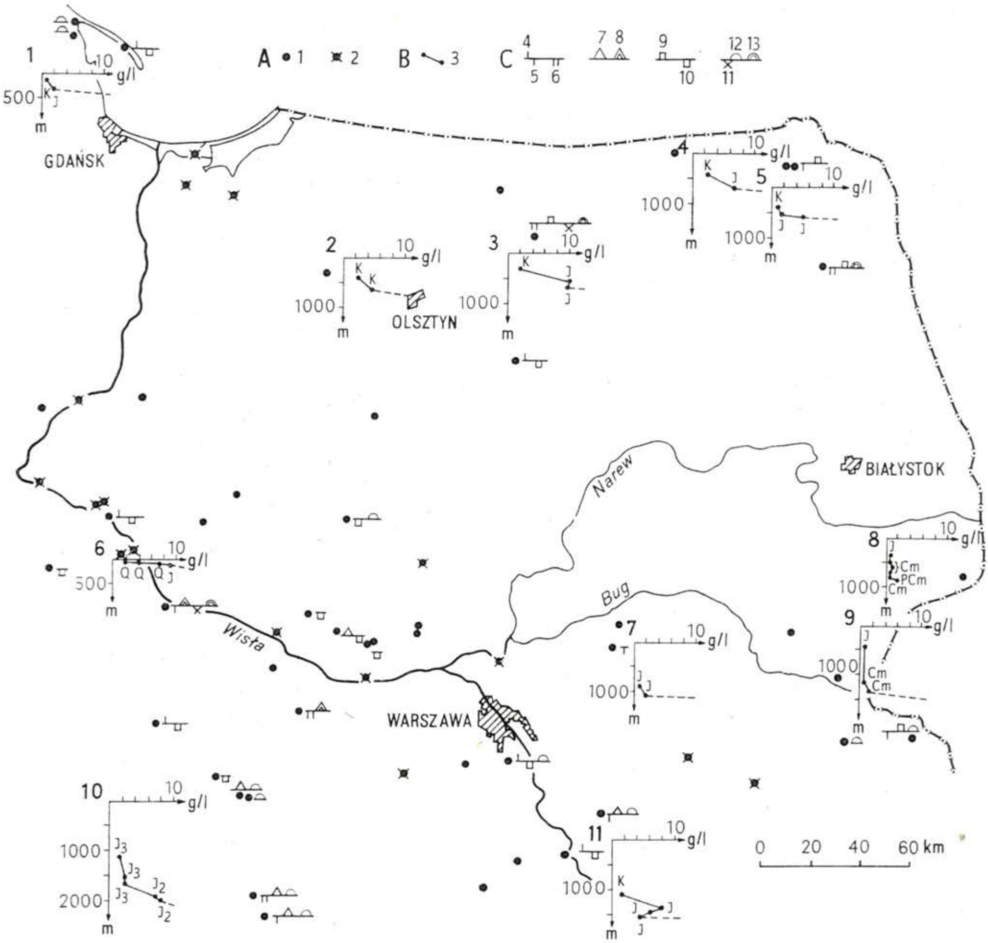

Fig. 2. Géochimie des eaux minérales près de leur « toit »

A) Points de base: *1.* Avec les analyses chimiques de l'eau. *2.* Sans données concernant la qualité de l'eau. — *B) 3.* Diagrammes — changement de minéralisation des eaux dépendamment de la profondeur; symboles stratigraphiques indiquant l'âge des aquifères: Q — Quaternaire, K — Crétacé, J_3 — Jurassique supérieur, J_2 — Jurassique moyen, J — Jurassique indivisé, Cm — Cambrien, PCm — Précambrien. — *C)* Quantité des éléments particuliers différant des valeurs considérées comme typiques, % miliéquivalents: *4.* Cl^- — au dessus de 90. *5.* Cl^- — 20 à 50. *6.* Cl^- — 10 à 20. *7.* SO_4^{2-} — 20 à 50. *8.* SO_4^{2-} — 50 à 75. *9.* HCO_3^- — 50 à 90. *10.* HCO_3^- — au dessous de 10. *11.* Na^+ — 20 à 50. *12.* Ca^{2+} — 20 à 50. *13.* Ca^{2+} — 50 à 75.

sédiments cités s'y manifestent. La teneur en ion HCO_3^- très réduite est due à l'influence de la salinisation chlorurée. Les cas des écarts « positifs » de cet ion, par ex. au dessus de 50 et même de 75 %, principalement vers le Nord-Est et l'Est, ne sont pas nombreux. La quantité de cet élément diminue dépendamment de la profondeur du « toit » des eaux minérales très vite.

Le sodium (Na^+) se fixe toujours au dessus de 50 % mé, sauf les eaux sulfatées (SO_4^{2-}). Les valeurs supérieures à 75 % sont observées vers le Sud-Ouest et le Sud-Est, même à des grandes profondeurs de la minéralisation. Le calcium (Ca^{2+}) apparaît en quantité supérieure à 20 %, sporadiquement dans les zones de diverses profondeurs du « toit » des eaux minérales. Il y est accompagné d'ordinaire par les valeurs élevés de HCO_3^-. Les eaux à SO_4-Ca constituent un type spécial. La teneur en magnésium (Mg^{2+}) ne dépasse pas 20 %, pour la plupart elle est inférieure à 10 %. Cet élément produit les différences spatiales importantes seulement dans les zones des eaux plus concentrées.

Interprétation des écarts

Deux facteurs se posent comme primordiaux dans l'interprétation des écarts observés. Tout d'abord, la valeur élevée de SO_4^{2-} est l'effet, comme on l'a déjà dit, de la répartition des sédiments riches en combinaisons de soufre. En dehors de ces régions on n'observe pas de zone d'eaux sulfatées, qui devrait être, selon les schémas classiques, intermédiaire entre les zones du type de HCO_3 plus proches du niveau du sol et celles de Cl^- — plus profondes.

Le second facteur, moins concret, est l'intensité de la circulation des eaux souterraines, c'est-à-dire leur échange sous l'aspect hydrochimique. L'activité de cet échange conduit vers l'augmentation de la teneur en HCO_3^- — principal élément des eaux souterraines non profondes qui gardent le contact avec l'atmosphère. Parmi les cations c'est Ca^{2+} qui y domine. Dans la région étudiée on constate les anomalies de HCO_3^- même en cas de « toit » profond des eaux minérales là, où il serait difficile de supposer que la circulation des eaux est actuellement active. Il faudrait chercher une explication sérieuse de ce fait dans les conditions paléohydrogéologiques. On y pourrait voir l'effet de la désalinisation ancienne des eaux dans les bassins originellement salés. L'éloignement des eaux salées était le résultat du processus tectonique du type d'élévation. Les phénomènes contraires de la salinisation des eaux souterraines non profondes, sont aussi liés aux mouvements tectoniques, entre autres les dislocations halocynétiques (diapirisme).

L'influence décisive de la tectonique sur la formation du « toit » des eaux minérales autorise à suivre l'interprétation contraire et à attacher certains écarts et anomalies hydrochimiques à une évolution tectonique concrète de la région. Cette évolution peut se développer par divers étapes, y compris l'étape contemporaine. Cela exige cependant des matériaux analytiques plus abondants et en même temps des recherches plus complexes dans les régions choisies. Actuellement des documentations hydrochimiques régionales sont déjà utilisées par les géologues pour identifier les structures constatées et hypothétiques. L'interprétation de caractéristique des eaux minérales à proximité de leur surface y joue également un rôle considérable.

THE INFLUENCE OF NEOTECTONIC JOINTS UPON THE HYDROGEOLOGICAL CONDITIONS OF SOUTH UKRAINE

INFLUENCE DE LA FISSILITÉ NEOTECTONIQUE SUR LES CONDITIONS HYDROGÉOLOGIQUES DU SUD D'UKRAINE

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RÉSUMÉ

On a realisé l'étude des singularités des variations spatiales des paramètres hydrodynamiques des sédiments carbonatées et terrigènes du Mésozoïque et Caïnozoïque dans la partie méridionale d'Ukraine. Ce travail a été réalisé à l'aide des méthodes géophysiques (prospection séismique, radiométrie etc.) et d'interprétation des données de forage et des pompages d'essai. On a établi que les territoires de perméabilité la plus forte formaient des zones linéaires dont la majorité avaient dans la Crimée la direction NE 50—60° (la direction NO 300—330° avait une importance secondaire), et en Prichernomor'e occidentale NO 330—345°. Les données basées sur des faits montrent que ces zones représent des bandes de l'épaississement des fissures tectoniques qui ont surgi pendant la dernière phase de l'orogénèse et après son achèvement. La révélation de la situation des zones fissurées a un grand intérêt pratique pour toutes les espèces des investigations hydrogéologiques et ingenieur-géologiques, y compris la récupération des eaux souterraines.

This work was undertaken by the authors in 1973—1975 with a purpose of studying the regularities of the variation in space of infiltration properties of carbonate to terrigenous sediments belonging to the Meso-Cenozoic group in South Ukraine, especially the Upper Jurassic and the Neogene that include the most important aquifers. The main attention was given to the recognition, exposure and contouring of the zones of high permeability rocks, because many practical problems arise from this high permeability.

The investigations were carried out in the central and eastern part of the Crimean Mountains (where the underground waters are stored in karst aquifer i.e. in faulted carbonate rocks of Upper Jurassic age, an area of mountain relief with temperate—moist climate), moreover in the western Crimea Plains (mostly with karst water, the aquifers are of carbonate to terrigenous sediments of the Neogene, Paleogene and Cretaceous, valley-ravine relief, semi-arid climate) and, lastly in some zones of NW Prichernomoje (the Black Sea Region), (aquifers in Neogene limestones, valley-ravine relief, semi-arid to arid climate).

The study included the rocks above the first water table, with a thickness ranging from 1.5—2.0 m (in the flood plain of the Dnieper river) to 500—600 m (the Crimean Mountains). In the western part of the Crimea Plains the investigations could be extended to the water-bearing horizons in the whole, because of the dense network of boreholes.

The research was comprehensive and adapted for the changing features of different regions. Geophysical methods as well as seismic survey, radiometry (especially work-out method), electrical logging and radiowave method, were widely used in all regions.

The estimation of permeability of rocks in well-drillable areas was made by means of drilling data, hydrogeological testing of the boreholes and special hydrogeologic studies. In the Crimean Mountains, there are many surface and subsurface karst forms and the exposure of very high permeability zones in the Upper Jurassic limestone of this region included the study of data of the space accommodation and nature of karst features (by means of especial geomorphological maps of the distribution of sinks, dry valleys, caves and streams, moreover by contour maps showing equal depths of karst wells and deep karst channels, equal volumes of cavities and equal yields of karst streams, diagrams on directions of horizontal and inclined karst cavities).

The conducting works allow to clear up the spread of zones of exceptionally high permeability, high permeability, average permeability and low permeability of rocks and also the areas where the rocks are practically impervious. The differences between this areas are very high.

There is a great difference in filtration properties of the rocks of these zones. So, for the limestone of Neogene age of the Novoselov elevation (Crimea Plains), the filtration coefficient k exceeds 100 m/day and specific yield q is greater than 50 l/sec for the zones of exceptionally high permeability; for the zones of high permeability $k=15-100$ m/day, $q=10-50$ l/sec, for the zones of average permeability $k=0.5-15$ m/day, $q=0.3-10$ l/sec, for the zones of low permeability $k<0.5$ m/day and $q<0.3$ l/sec.

No doubt, these zones of good permeability are the zones of thickening of the linear tectonic jointing. First of all, this is confirmed by data of a careful study of joints in the exposures, showing the increase of the jointing coefficient in high permeability areas to be five-ten times of the areas characterized by low permeability. Secondly, the directions of these zones coincide with the general direction of tectonic jointing in the given region. Thirdly, there is a linear parallelism regarding the zones on the whole or their linear elements. It is emphasized that there are two directions in the morphology of most zones and, in the majority of cases, these directions are perpendicular or crossing at angles of $45-60°$ that also characterizes the tectonic type of joints.

An overwhelming majority of the revealed high permeability zones have the complex but clearly expressed linear configuration (Fig. 1). The consistency and extent of zones and the distance between them are different. Most of the large zones are very consistent, surpassing in length $20-40$ km and in width $0.5-3$ km. For example, the complex Karaby-Chambey jointed karst zone is about 70 km long; the southern part of this zone is illustrated in Fig. 2.

Facts are evidencing that these zones of high permeability are the belts of thickening of the recent linear neotectonic jointing, formed at the last stage of folding and affecting the whole geological sequence from the oldest deposits up to the youngest ones. So, in the Crimea Plains and in western Prichernomorje the jointed karst zones of high permeability are extend from the upper sediments of the Cenozoic group to the lower beds of the Mesozoic at the depth of 3 to 4 km. It is important that the revealed zones cross the tectonic elements of different folding stages, as being proved by our works

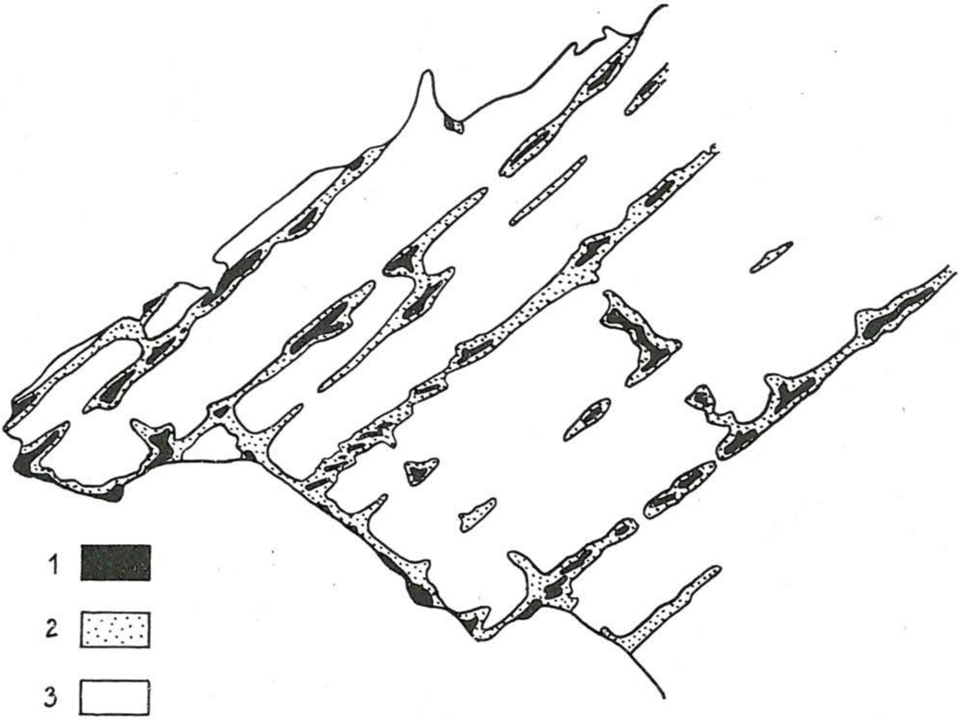

Fig. 1. The linear-parallel jointed zones of Tarhankut
1. Areas of exceptionally high permeability, *2.* areas of high permeability, *3.* areas of very low permeability

in the western (the Geracley peninsula) and eastern (the Kerch peninsula) periclinal structures of the Mountain of the Crimean anticlinorium, and also on its northern slope (the region of the Belogorsk). The transition of these zones manifested by a grading from the area of Alpine folding into the area of Hercynian folding is often observed.

In that way, the considerable jointed zones may be taken for a "superimposed" jointing (KUSHNAREV, LUKIN, 1960) that arose in the last stage of tectogenesis or after its end when the structural pattern had already been formed.

It is significant that the direction of parallel joints in zones of the Crimea Plains and in western Prichernomorje coincides with the direction of Alpine folding in the Crimean Mountains. Most of the zones of Crimea have a general orientation of NE 50–60° (Fig. 1); the direction NW 300–330° plays a secondary role. In western Prichernomorje a direction of NW 330–345 predominates.

Limestones and also loose deposits (sands, loam with loess, sandy loams, aleurites) are the densely jointed rocks with a higher permeability within the involved zones. In the central parts of the largest jointed zones the permeability tends to increase even in clays.

Most jointed zones are indicated also by the younger relief. There are many small ravines, valley lineaments (Donuslav, Chatirlyk, Voronsovka)

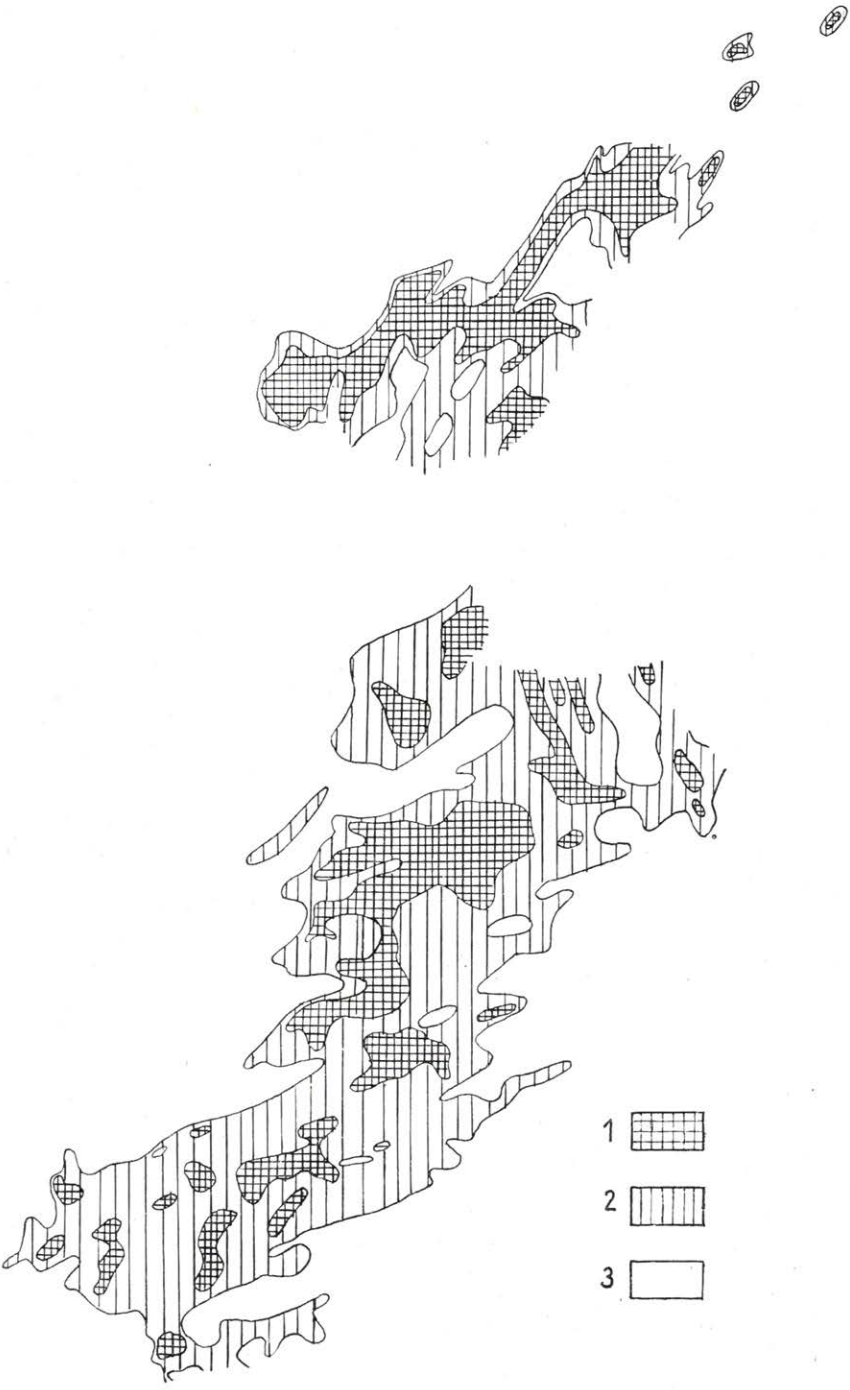

Fig. 2. The Karaby-Chombey jointed karst zone

1. Areas of exceptionally high permeability; *2.* areas of high permeability, *3.* areas of very low permeability

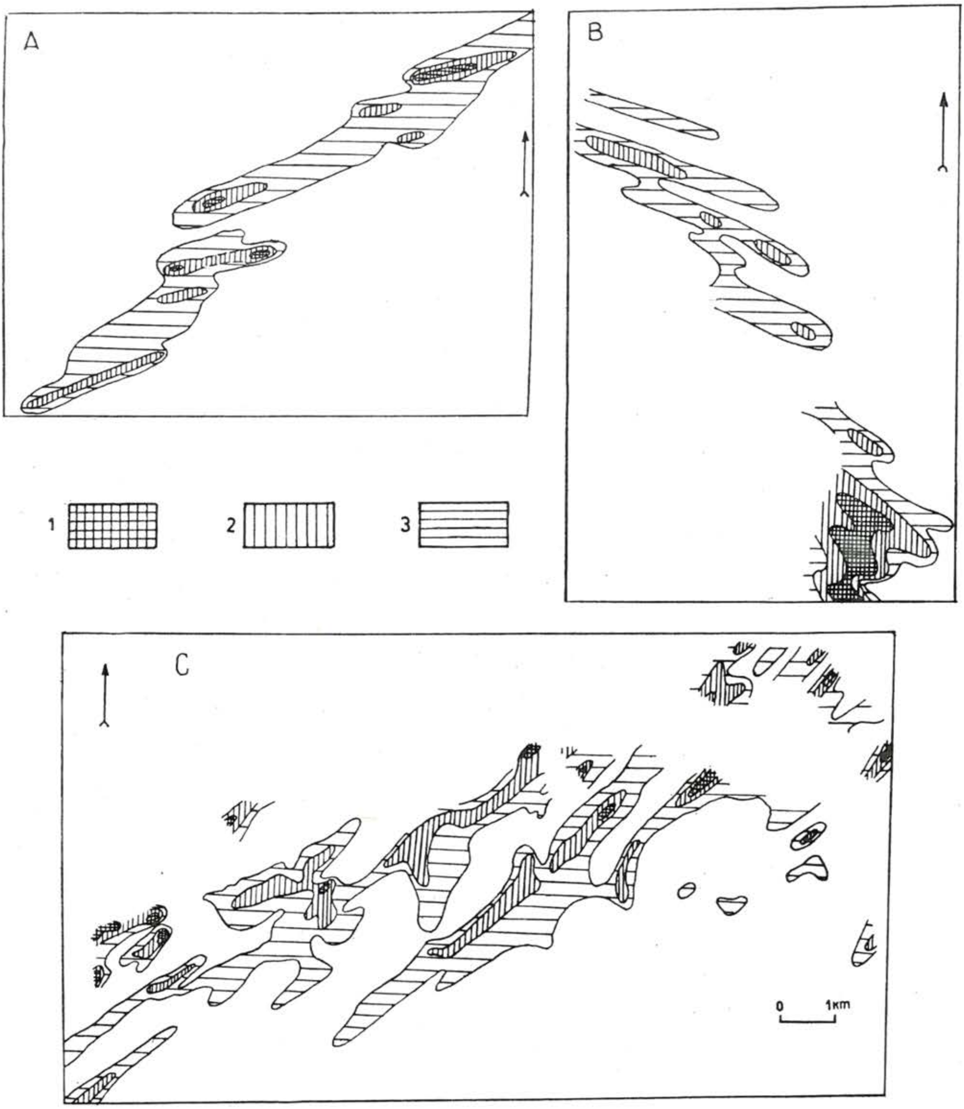

Fig. 3. Character of the coulisses in jointed karst zones

1. Areas of exceptionally high permeability, *2.* areas of high permeability, *3.* areas of average permeability

and rivers (Salgir, Karasu etc.) in these zones. The majority of the slopes in the Crimea Mountains are formed on these jointed zones (e.g. the eastern slope of Chatyrdag, the western slope of the Dolgorukov massif, and many places of the south scarp of Karaby-jajla).

An important feature of the configuration of all reconnoitred zones is the coulisse-like development of the structural components. For example, the Donuslav zone (Fig. 3, *A*) has the main direction NE 60°, but the parallel sets branching in form of coulisses have a direction of NE 75 – 80, and each unit joins the other in a sinistral direction (left hand rule). Coulisses in a row tending to the right are in the West-Dolgorukov zone; these coulisses have a lineation according to NW 300° with main direction given by a zone directed NW 320° (Fig. 3, *B*). The coulisse-type tectonic development is reflected by the morphology of other jointed zones, too (Geracley zone, Fig. 3, *C*).

Most authors (Radkevich, 1960, Melnikov, 1960 etc.) consider the coulisses as a distinctive peculiarity of the jointing, though there is an opinion that it can be explained by the vectorial expansion of the principial tangent stresses by compression, on the basis of the scheme of strain ellipsoid of G. Becker.

The discovery and mapping of joint zones of different permeability is of great practical interest to all aspects of the hydrogeological and engineering-geological investigations and especially to ground-water recovery.

LE RÔLE DES MOUVEMENTS NÉOTECTONIQUES DANS LA CRÉATION DE RÉSERVOIRS D'EAUX SOUTERRAINES DANS LA PARTIE NORD-EST DU BASSIN DE VIENNE

THE ROLE OF NEOTECTONIC MOVEMENTS IN THE DEVELOPMENT OF GROUND WATER RESERVOIRS IN THE NORTH-EASTERN PART OF THE VIENNA BASIN

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ABSTRACT

The Quaternary neotectonic movements conditioned the origin of a vast collapse ditch covering the territories of Austria and Czechoslovakia. It is hydrogeologically an important graben extending from Wiener-Neustadt in Austria through Moravské pole well in to Slovak territory in the Záhorská plain. It is 106 km long, from 2 to 8 km large and filled with hydrogeologically favourable Quaternary sediments, thick 50 to 150 m. It is a marked supply reservoir of ground waters in the territories of both countries.

This hydrogeologically important graben which was developed under the influence of Quaternary subsidence tectonics is divided by lateral horsts into three graben units filled with Quaternary sediments, thus creating 3 vast reservoirs of Quaternary ground waters.

The southermost ground water reservoir is the Mitterndorf graben, 40 km long, 2 to 8 km wide and a thickness of Quaternary sediments up to 150 m (Küpper 1954). The middle ground water reservoir is the "Lasser Senke" graben in the Moravské pole, equally with important thicknesses of Quaternary sediments. The northern reservoir is the Zohor graben, 36 km long, 7 to 8 km wide and Quaternary sediments hydrogeologically favourably 50 to 100 m thick (Kullman 1966).

An extensive geological and hydrogeological research carried out during the 10 recent years made possible a detailed evaluation of the hydrogeological conditions of this northernmost partial graben. Quaternary land subsidences in it attain 100 to 150 m (Banacky—Harcár—Sabol 1965).

Moreover, the detailed evaluation pointed out the presence of complicated neotectonic influences inside this partial graben, caused by an unequal subsidence of the individual partial blocks of the Neogene basement, separated by lateral faults. Therefore an important role, along the longitudinal faults between which the Zohor graben was subsiding, was played here also by lateral faults which divided it into unequally subsiding subsections.

In relation to hydrogeology this caused in the Zohor graben the creation of three partial Quaternary reservoirs more or less independent, divided mutually by horsts. The favourable situation of the Zohor graben made possible its gradual filling with proluvial sediments hydrogeologically favourable from the Little Carpathian mountain range from south-east, with the eolian sands from north-west and north and with alluvial deposits of the Morava river in the southern to south-western part of the graben. The co-action of Quaternary tectonics and the sedimentation of hydrogeologically favourable Quaternary sediments made possible in the Zohor ground water reservoirs the development of important dynamic supplies of Quaternary ground waters at an approximate rate of 1200 l/s and the accumulation of 1.46 milliard m^3 of ground water.

Introduction

Les mouvements néotectoniques influencèrent dans une mesure substantielle la création de réservoirs d'eaux souterraines importants dans les plaines de la Slovaquie. L'une des régions les plus importantes, dans laquelle la jeune tectonique, surtout quaternaire, avait rendu possible la création d'importants réservoirs d'eaux souterraines, est la plaine Záhorská nižina dans la partie nord-est du bassin de Vienne. Sur le territoire de la plaine Záhorská la partie la plus mobile à l'époque quaternaire fut le graben sousmontagnard des Petites Carpates, s'étendant sur une longueur d'environ de 106 km sur le territoire de l'Autriche et de la Tchécoslovaquie.

Un aperçu géologique

Le graben sousmontagnard des Petites Carpates, appelé aussi le graben de Zohor, forme la partie nord d'un vaste fossé d'effondrement, de largeur variante entre 2—8 km. Il part de Wiener-Neustadt en Autriche et passe par Moravské pole pour entrer sur le territoire de la Slovaquie dans la plaine de Záhorská nižina.

Ce vaste fossé d'effondrement très important au point de vue hydrogéologique, est divisé par de horsts latéraux en 3 unités de grabens indépendants remplis de sédiments quaternaires. De ces trois grabens deux s'étendent sur le territoire de l'Autriche et le troisième — le plus au Nord — sur le territoire de la Tchécoslovaquie. Celui situé le plus au Sud est le graben de Mitterndorf qui débute au Sud à Wiener-Neustadt, s'étend par Mitterndorf jusqu'à Wienerherberg. Il est limité par la faille de Sollenau-Moosbruner et de Goldberg, le long de laquelle il s'effondra et forma un graben long de 40 km, large de 2 à 8 km, rempli de graviers pleistocènes dont l'épaisseur va jusqu'à 150 m (Küpper 1954).

Entre le Danube et la Morava dans la continuation du graben de Mitterndorf au Nord-Est on trouve le second graben indépendant important sur Moravské pole, qui est le graben « Lasser Senke » similaire au précédent, effondré également le long des failles et rempli également de sédiments quaternaires (Fink 1955).

Le graben « Lasser Senke » devient moins profond vers Marchegg et à l'intérieur du fossé d'effondrement dans la région au Sud-Ouest de Marchegg il passe en graben qui est documenté par la terrasse de Gänsendorf du Würm (voir J. Fink 1955). Il s'agit probablement de horst latéral peu étendu, car à Marchegg s'étend déjà la bordure sud du troisième graben quaternaire indépendant celui de graben de Zohor, qui est un graben sousmontagnard des Petites Carpates et se trouve pratiquement par toute son étendue sur le territoire de la Tchécoslovaquie. Dans ce rapport nous présenterons une caractéristique géologique et hydrogéologique générale de ce graben situé le plus au Nord, rangé parmi les régions les plus importantes en Tchécoslovaquie du point de vue de l'économie de l'eau.

Le graben de Zohor fait partie du graben néogène de Zohor-plavecká depresia. Vers la fin du Néogène la partie du graben plavecký (la continuation nord de graben de Zohor) se stabilisa déjà et seulement le graben de Zohor continuait à s'affaisser. Les recherches géologiques et hydrogéologiques effec-

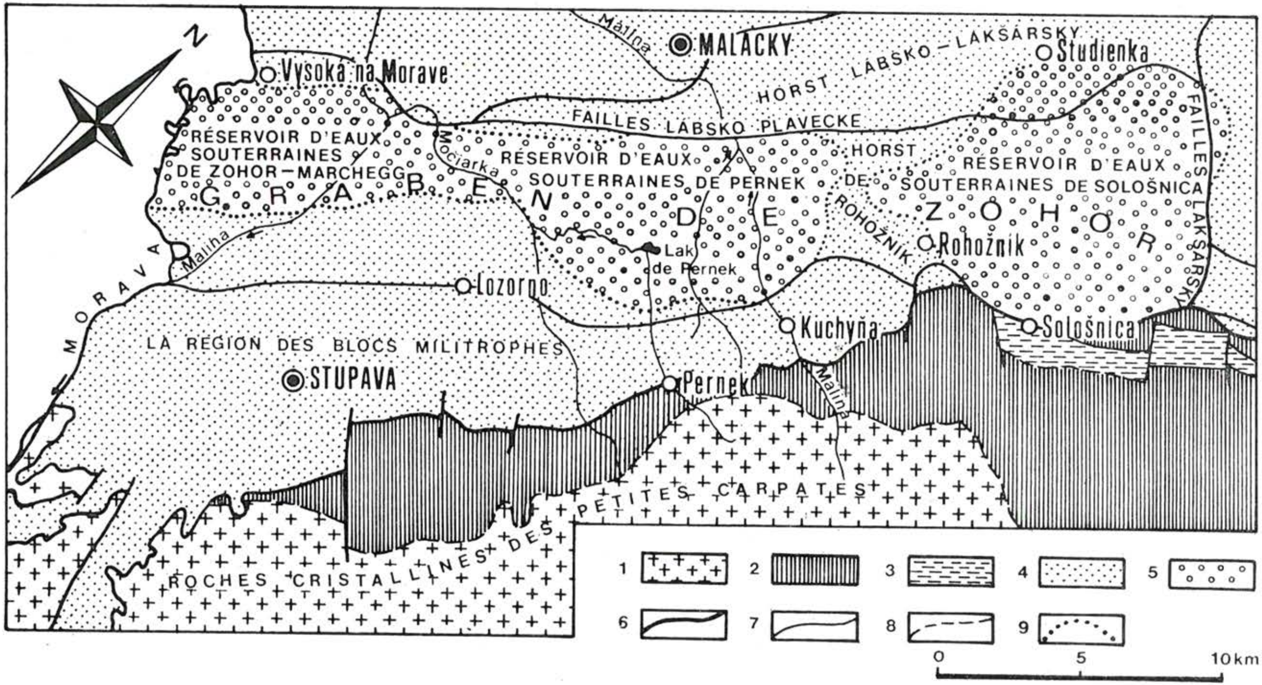

Fig. 1. Carte schématique des ensembles tectoniques des Petites Carpates à Záhorská Nízina avec l'indication des réservoirs d'eaux souterraines quaternaires

1. Roches cristallines, *2.* Mésozoïque, *3.* Paléogène, *4.* Néogène (y inclus le Quaternaire susjacent hydrogéologiquement moins significatif), *5.* le Quaternaire hydrogéologiquement important des réservoirs d'eaux souterraines, *6.* bord des Petites Carpates, *7.* limites des unités tectoniques individuelles identiques aux lignes principales tectoniques, *8.* frontières géologiques, *9.* bornage des réservoirs quaternaires

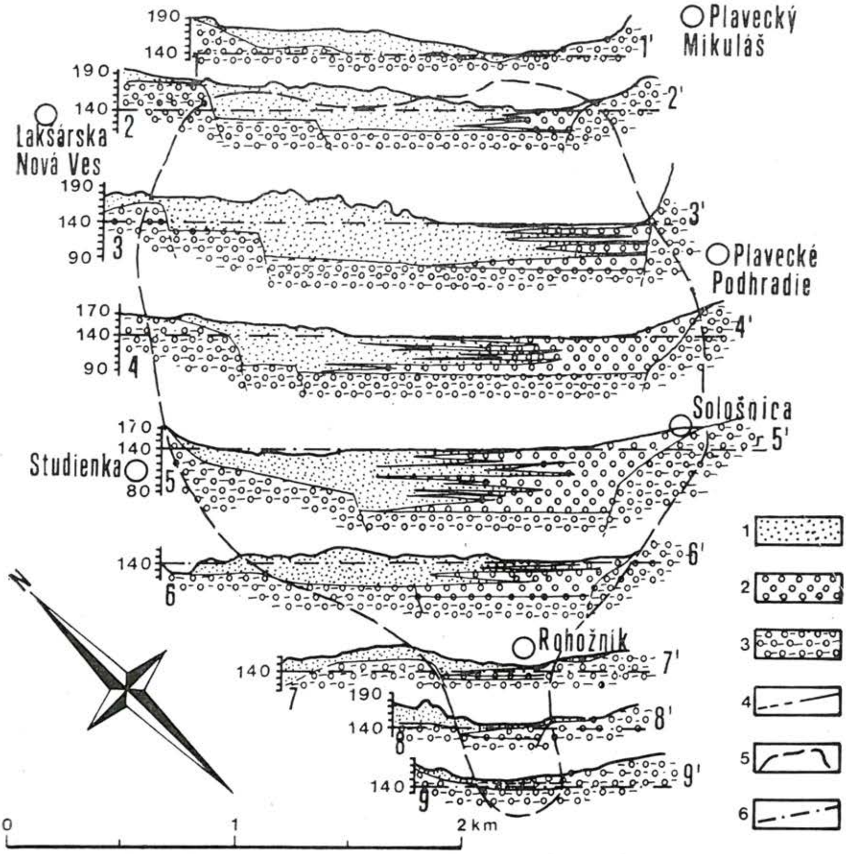

Fig. 2. Carte de profil du réservoir d'eaux souterraines de Sološnica à Záhorská Nižina
1. Sables (prédominamment mouvants), *2.* graviers proluviaux par endroits limoneux (*1—2* Quaternaire), *3.* argiles, argiles sableuses (Néogène), *4.* failles, *5.* bord du réservoir, *6.* ligne profil (l'axe de renversement du profil)

tuées sur le graben de Zohor au cours de la dernière décennie permettent de présenter son évaluation plus détaillée.

Le graben de Zohor s'étend au SO-NE dans le graben sousmontagnard des Petites Carpates, de Marchegg jusqu'à Plavecký Mikuláš, où il est limité au Nord par la faille latérale de Lakšáre. Les limites nord-ouest ainsi que sud-est sont formées également par de failles. Au nord-ouest il est limité par les failles lábsko-plavecké envers les horsts lábsko-laksárske. La limite sud-est est formée par les failles limitrophes des Petites Carpates. Le graben, lui-même, au Nord se situe tout près de la chaine des Petites Carpates. Plus loin dans la direction sud se trouve entre lui et la chaine de montagne la banquise néogène limitrophe des Petites Carpates. Similaire aux deux grabens précédents en Autriche, c'est un graben limité par les failles et d'affaissement marquant, long environ 36 km et large de 7 à 8 km.

Le graben de Zohor, lui-même, ne forme point un ensemble uniforme. A son développement contribuèrent aussi les mouvements le long des failles latérales (de direction NO-SE), qui conditionnèrent à l'intérieur la division du graben de Zohor en d'autres horsts et grabens partiels. Par deux horsts expressives, celui de Rohožnik et de Lozorno, le graben de Zohor est divisé en grabens quaternaires parties de Sološnica le plus au Nord, le graben central de Pernek et le graben de Zohor—Marchegg le plus au Sud.

L'affaissement quaternaire dans le graben partiel de Sološnica représente environ 100 m; dans le graben partiel de Pernek il atteint 150 m et dans le graben partiel de Zohor—Marchegg environ 100 m.

Le développement des sédiments quaternaires dans ces grabens partiels fut influencé d'une façon expressive par la tectonique d'affaissement quaternaire qui, dans le graben sousmontagnard des Petites Carpates, exerçait une influence en premier ordre sur le développement des cônes d'alluvionnement. Tandis que sur les horst partiels les cônes se présentent en petite épaisseur (2 à 3 m), dans les grabens partiels l'affaissement a produit l'accumulation des matériaux proluviaux et la formation des séries de couches épaisses de sédiments clastiques quaternaires. Dans le graben de Sološnica l'épaisseur est d'environ de 70 m. Les dépôts proluviaux remplissaient les grabens de Sud-Est. De l'Ouest et de Nord-Ouest ceux-ci furent remplis de sables mouvants qui remplissaient les parties ouest des grabens jusqu'à une épaisseur de 50 à 80 m. D'ici l'alternance des sédiments proluviaux et éoliens dans les parties centrales.

Un caractère différent présentent les sédiments quaternaires dans le graben partiel de Zohor—Marchegg, dans lequel on trouva l'épaisseur des graviers et des sables fluviaux de 85 à 100 m.

L'hydrogéologie du graben de Zohor

Dans le graben de Zohor nous pouvons distinguer 3 réservoirs importants d'eaux souterraines quaternaires (Kullman 1966), séparés par les horsts partiels.

Le réservoir d'eaux souterraines de Sološnica le plus au Nord s'étend entre la faille de Laksáre, la faille limitrophe des Petites Carpates, la partie supérieure des failles lábsko-plavecké et le village de Rohožník; sa superficie est de 85 km^2. Il est rempli d'une épaisse série de sédiments quaternaires hydrogéologiquement favorables. A l'arrière de la faille limitrophe près du bord SE de réservoir, il y a les affaissements de la plus grande amplitude, là sédimentait le matériel trié des cônes de cours d'eaux carpatiens. Les sédiments quaternaires dans cette partie du réservoir peuvent être caractérisés comme un complexe puissant de sédiments proluviaux, de fragments de roches aux angles grossiers aigus, en partie couverts de limon, et sables éoliens peu épais. L'épaisseur totale de sédiments quaternaires dans cette partie du réservoir va de 50 à 85 m. Le coéfficient d'infiltration de ces sédiments proluviaux se range entre $7.1 \cdot 10^{-5}$ m/s et $3.0 \cdot 10^{-4}$ m/s et le coéfficient d'accumulation entre 0.20 et 0.22. Vers la plaine à partir des Petites Carpates les conditions géologiques changent substantiellement dans le réservoir de Sološnica et par là aussi les conditions hydrogéologiques. Plus loin de la faille limitrophe en direction du centre de réservoir, les fractions des sédiments de gravier et de sable deviennent plus fines, leur recouvrement de limon s'augmente ainsi que l'épaisseur des sédiments éoliens. Vers le NO de la bordure du réservoir prend lieu une transition graduelle en sédiments éoliens purs — sables mouvants,

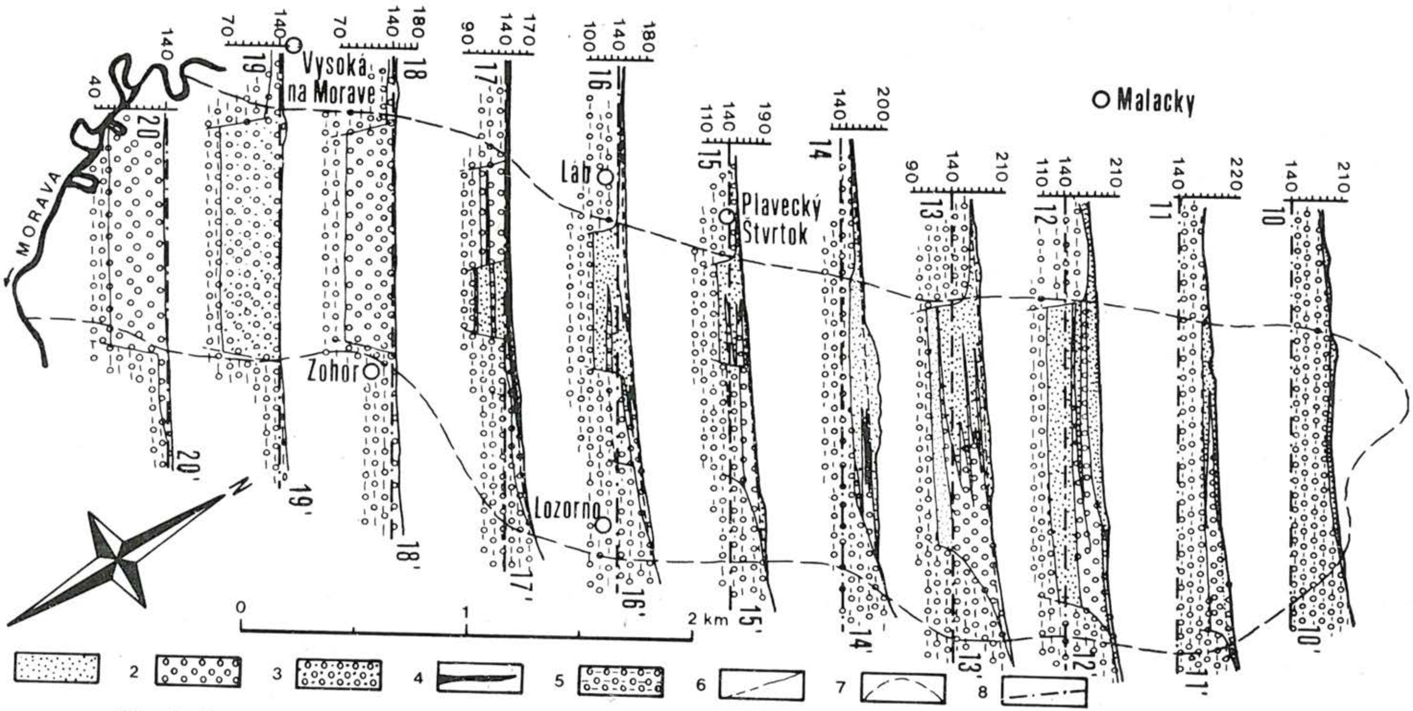

Fig. 3. Carte de profil réservoir d'eaux souterraines de Pernek et de Zohor—Marchegg à Záhorská Nižina

1. Sables (prédominamment mouvants – Quaternaire), *2.* graviers proluviaux par endroits limoneux – Quaternaire, *3.* graviers et sables avec intercalations d'argile (dans le réservoir de Zohor – Marchegg – Quaternaire), *4.* limons entrainés, tourbes, *5.* argiles, argiles sableuses, sables argileux, *6.* failles, *7.* bord du réservoir, *8.* ligne profil (l'axe de renversement du profil) formés prédominamment de grains de sable fins et atteignant une épaisseur maximum de 70 à 80 m. Leur coéfficient d'infiltration varie entre $8.0 \cdot 10^{-5}$ m/s et $1.1 \cdot 10^{-4}$ m/s. Le coéfficient d'accumlation d'eau calculé par plusieurs méthodes (au terrain et au laboratoire) va de 0.09 à 0.20. Le réservoir est alimenté par le passage direct des eaux souterraines des roches carboniques des Petites Carpates, ainsi que par l'infiltration des eaux de surface, des cours d'eaux des Petites Carpates, et aussi par des eaux atmosphériques du réservoir propre. Les recherches effectuées jusqu'à présent dans cette région documentent en moyenne l'infiltration de 3.6 – 4.1 l/s/km² sur le territoire du réservoir.

Le drainage des réserves dynamiques d'eaux souterraines du réservoir se fait par le cours d'eau de surface de Rudava sur la barrière des failles lábsko-plavecké.

Entre les villages de Rohožník et Kuchyňa le réservoir d'eaux souterraines de Sološnica est séparé d'un autre réservoir d'eaux souterraines quaternaires par le horst partiel de Rohožník. Ce dernier est couvert de sédiments quaternaires peu épais, sans importance hydrogéologique.

Un autre réservoir d'eaux souterraines nommé de Pernek est le réservoir central du graben de Zohor. La limitation tectonique sud-ouest et nord-est du réservoir forme des failles identiques au réservoir précédent. La limitation nord-est est formée par le horst latéral de Rohožník et celle de sud ouest par le horst latéral de Lozorno limitant le réservoir seulement partiellement. Le réservoir quaternaire couvre une surface de 69.4 km². Il est rempli de puissantes séries de couches de sédiments déluviaux-proluviaux et éoliens. La partie est du réservoir dans le massif montagneux des Petites Carpates est remplie de graviers proluviaux de grains petits et gros avec mélange de sable au grains moyens et fins, alternant avec les couches de sables éoliens aux grains fins. L'épaisseur moyenne de sédiments quaternaires dans cette partie de réservoir atteint 60 à 70 m, avec le coéfficient moyen d'infiltration $2.2 \cdot 10^{-4}$ m/s. La partie centrale et ouest du réservoir est comblée par des couches de sables éoliens aux grains fins jusqu'à moyens avec des couches individuelles de limons, limons sableux, de graviers et de sables. La partie inférieure de la série de couches est formée de limons et couches fortement limoneuses de graviers et de sables. L'épaisseur maximum des sédiments quaternaires dans la partie centrale et est de réservoir est de 120 m. La partie supérieure et centrale du complexe est formée de sables éoliens à une profondeur d'environ de 70 m. Leur coéfficient d'infiltration moyen est de $1.4 \cdot 10^{-4}$ m/s et le coéfficient d'accumulation 0.10 – 0.16. Les eaux souterraines du réservoir sont alimentées tant par l'infiltration des cours d'eaux des Petites Carpates dans les cônes d'alluvionnement de réservoir, ainsi que par l'écoulement direct des eaux souterraines des Petites Carpates et aussi par l'infiltration des eaux atmosphériques sur la surface du réservoir. Le drainage principal des eaux souterraines est similaire au réservoir précédent. C'est au total 120 à 230 l/s d'eaux souterraines qui s'écoule du réservoir. Le drainage secondaire des réserves dynamiques de ce réservoir quaternaire se fait par un fossé étroit à travers le réservoir de Zohor – Marchegg situé plus au Sud.

Entre le réservoir central de Pernek et celui de Zohor – Marchegg émerge le horst latéral de Lozorno avec le sous-sol imperméable élevé. Entre ces réservoirs il existe, cependant, une connexion hydraulique par ce horst à travers une banquise affaissée (large environ 2 km), qui passe latéralement par le horst de Lozorno et forme un effondrement étroit, rempli de sables et graviers quaternaires d'une épaisseur de 45 à 60 m. Ce canal permet le passage direct des eaux souterraines de réservoir de Pernek dans celui de Zohor—Marchegg. Par ce canal s'écoulent en moyenne 100 l/s d'eaux souterraines (KULLMAN 1966).

Le réservoir de Zohor—Marchegg situé plus au Sud est limité au Nord-Ouest et Sud-Est par de failles identiques aux réservoirs précédents. La limite Nord-Est est formée par le horst latéral de Lozorno et au Sud-Ouest par le horst émergeant au Sud-Ouest de Marchegg. Près du bord OSO la rivière de Morava passe par ce réservoir. Il occupe une surface de 37.7 km^2 sur le territoire de la Tchécoslovaquie. Le remplissage quaternaire de ce réservoir est formé par des alluvions de la rivière de Morava, et en partie aussi par de sables éoliens près de la bordure NO et N. Dans la région à Sud-Ouest du réservoir de Pernek l'épaisseur moyenne de sédiments quaternaires est environ de 48 m, plus loin elle augmente à 56—62 m et finalement près de Morava elle atteint 85 à 100 m. Dans la majeure partie du réservoir les sédiments quaternaires sont formés de graviers sableux et couches d'argiles sableuses. Le coefficient d'infiltration de ces sédiments est de $5.7 \cdot 10^{-4}$ m/s à $8.55 \cdot 10^{-4}$ m/s. Avec l'augmentation du composant sableux la valeur du coéfficient d'infiltration diminue à $3.6 \cdot 10^{-4}$ m/s jusqu'à $5.1 \cdot 10^{-4}$ m/s. Dans la partie nord-est du réservoir dans les formations de sables éoliens, les coéfficients d'infiltration diminuent jusqu'à $1.2 \cdot 10^{-4}$ m/s. La valeur du coéfficient d'accumulation est 0.14 à 0.20. L'alimentation de réservoir se fait d'un côté par les eaux souterraines provenant de réservoir de Pernek, par les eaux infiltrées des cours d'eau surface et de l'autre côté par les eaux atmosphériques. Le drainage des réserves dynamiques des eaux souterraines se fait par leur écoulement direct dans la rivière de Morava.

L'évaluation des réserves d'eaux souterraines du graben

Le graben de Zohor situé au nord fut évalué en détail au cours des 10 dernières années en utilisant une série de méthodes géologiques et hydrogéologiques ainsi que de nombreux forages. Les résultats de ces travaux permettent de donner une évaluation précise des réserve dynamiques des eaux souterraines ainsi que des quantités des eaux souterraines accumulées dans les trois réservoirs quaternaires du graben de Zohor. Se basant sur ces investigations réalisées dans les trois réservoirs partiels du graben de Zohor on avait documenté la moyenne des réserves dynamiques des eaux souterraines dans une quantité de 1200 l/s. L'accumulation totale des eaux souterraines dans ces trois réservoirs d'eaux souterraines du graben de Zohor représente 1.46 milliard m^3 (E. KULLMAN 1966, A. SUBOVÁ 1973, E. KULLMAN 1974).

Dans le graben de Mitterndorf occupant la partie sud du graben, long de 40 km, large 2 à 8 km, où l'épaisseur de sédiments quaternaires est 50 à 150 m, on avait effectué une documentation d'orientation de l'accumulation totale des eaux souterraines d'une quantité de 2.3 milliards m^3 (H. KÜPPER 1955).

L'évaluation détaillée avait fait remarquer les influences néotectoniques compliquées à l'intérieur du graben de Zohor causées par l'affaissement non uniforme de blocs individuels séparés par les failles latérales.

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HYDROGEOLOGICAL DATA FROM SE TRANSDANUBIA AS A PART OF MARGINAL AREA OF THE GREAT HUNGARIAN PLAIN AND DRAVA BASIN

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Geological review of SE Transdanubia

The area of SE Transdanubia is geologically divided into four parts according to the structure of the pre-Tertiary basement and thickness of covering Neogene deposits. On the pre-Tertiary basement contour map (Fig. 1) – which can be regarded as a Neogene isopach map – the following distinct parts are recorded:

a) *The Neogene Depression of Drava Basin.* Thickness of the Neogene sequence increases from 1000 m up to 3500 m showed a fairly uniform SW trend. The basement of this area is generally composed of Precambrian metamorphic rocks. This depression – as a great basin – is the subject of our detailed investigation.

b) *The area of Mecsek – Villány Anticline.* Thickness of the Neogene deposits on the highland areas goes from 0 m to 500 m but on the southeastern and northwestern flanks of the anticline they enlarge from 500 to 1000 m thickness. This situation of the anticline is motivated by the trench of the "Mecsekalja Tectonic Belt" filled up by more than 500 m thick Neogene deposits. The core of the anticline consists of Precambrian metamorphic rocks, granites and Late Paleozoic sedimentary formations. The bottom part of the flanks consists of Mesozoic cavernous carbonate rocks (see Fig. 2).

c) *Neogene trench-like area between Liget- and Ellend Basin.* Thickness of the Neogene sequences of both basins surpasses at least 1000 m. The bottom part is generally composed of karstic carbonate rocks. This Neogene tectonic zone is located to the area of so-called "Villány – Szalatnak Paleozoic Deep Fracture Zone" and it is regarded as a Neogene renewal of the Late Paleozoic movements. This area is isolated from the great basin, therefore further details are omitted.

d) *The area located to the east from the Villány – Szalatnak Paleozoic Deep Fracture Zone* where some strips of the rocks can be found with particular geological structures, bordered by fractures striking NE – SW direction. This area is also separated from the great basin.

Concluded from the mean-thickness distribution of the Neogene deposits

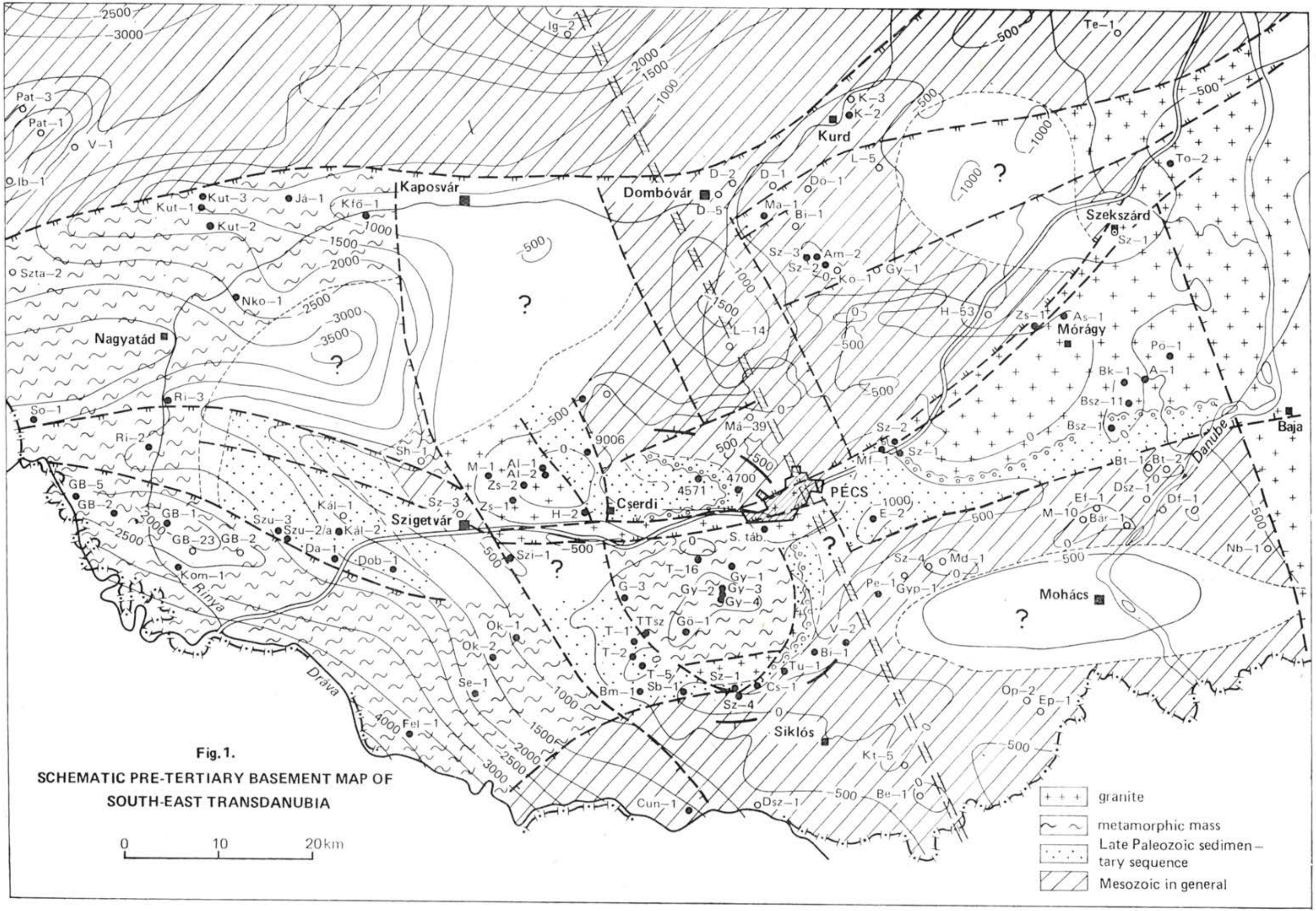


Fig. 1.

SCHEMATIC PRE-TERTIARY BASEMENT MAP OF SOUTH-EAST TRANSDANUBIA

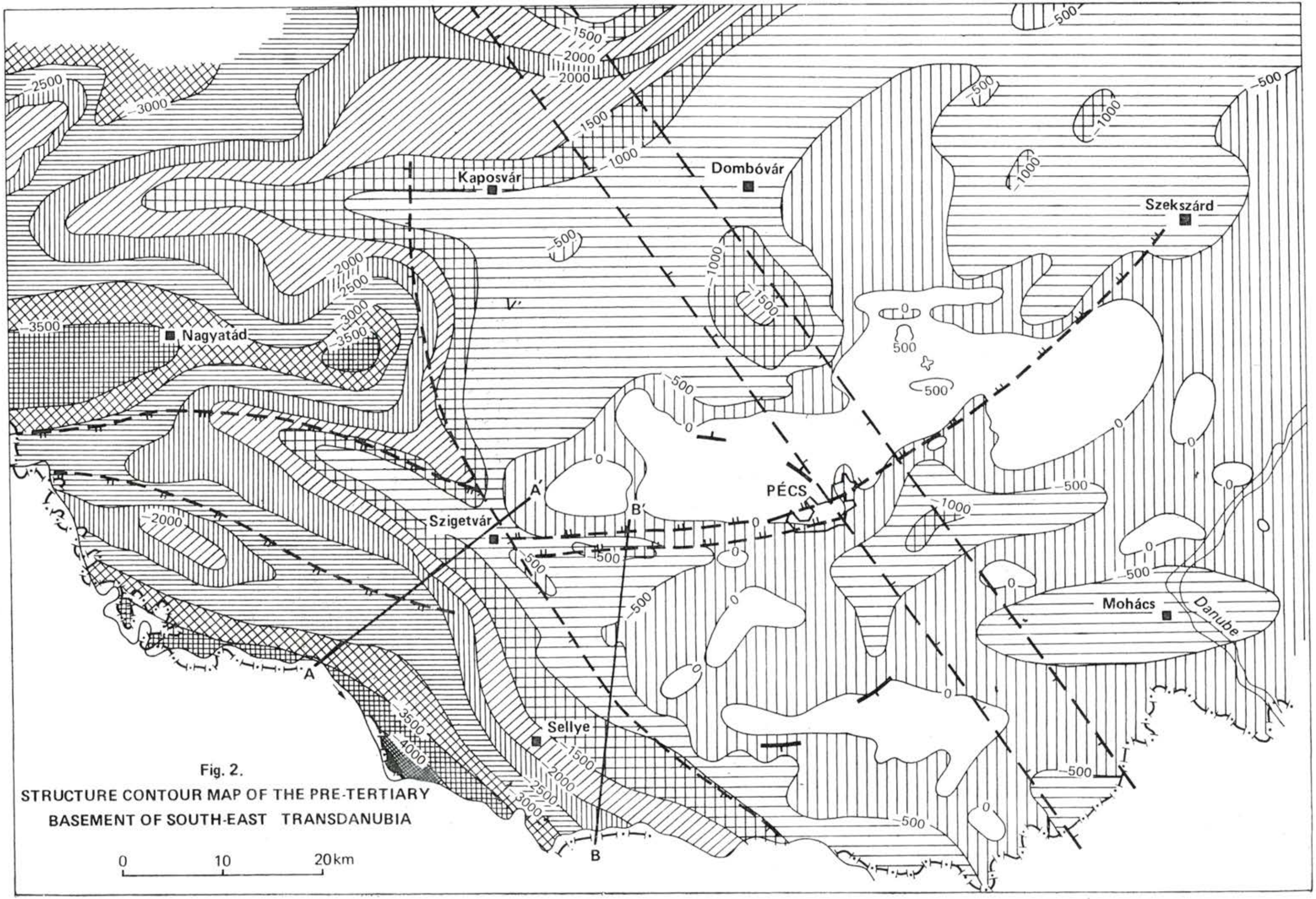


Fig. 2.
STRUCTURE CONTOUR MAP OF THE PRE-TERTIARY BASEMENT OF SOUTH-EAST TRANSDANUBIA

as well as tectonic inference, the main tectonic zones of the investigated region are the following ones:

1. NW–SE striking fracture zone at the eastern border of Drava basin, which is a notable water-flow zone too. This gives a possibility for fundamental inferences related to the investigation of great basins.

2. Tectonic zones with NE–SW trend represented in the area of Mecsek–Villány anticline (Villány Mts. and northern side of Mecsek Mts) as well as trench-depression of "Mecsekalja Tectonic Belt".

3. NW–SE striking Neogene tectonic zone coinciding with "Villány–Szalatnak Paleozoic Deep Fracture Zone" having some tectonic elements which are deeply wedged into pre-Tertiary basement.

4. Tectonic zones having a NE–SW trend which are located to the eastern part of the area.

Hydrodynamic investigations of the deposits of Drava "great basin"

Neogene deposits of the young depressions of the examined area are rather uniform. The basement is overlying by predominantly impermeable Middle Miocene clayey-marls, clayey-calcareous sandstones, tuffs and lavas, as well as a Lower Pannonian sequence; and all these at several localities have some separated sandstone lenses with concentrated salty-water content.

The oil-prospecting boreholes explored Badenian "Leitha limestone" lenses above the uplifted basement frequently containing waters of extremely high pressure.

In loose and porous sandstones of the Upper Pannonian complexes there are utilizable thermal waters as reservoir fluids with low salinity. In the investigated area such an aquifer is drained by the well of Pettend, village supplying artesian water of 39.5 °C temperature. Similar water quality has the thermal water of Sellye, with 48 °C temperature.

According to many signs near to the surface, the Upper Pliocene and Pleistocene formation waters and phreatic ones form a single, continuous porous water-bearing formation. Salt content of the Upper Pannonian layers is the same as the sea-water, however, in the lower part of the Upper Pannonian stage brackish-water can be found, during the upflow it changes gradually into fresh-water (B. Kleb 1973). According to the geoelectric soundings carried out on the area, the "regionally impermeable floor" of the coarse-grained complexes near the surface is more or less on the level of the lignite beds of Upper Pannonian complex, which at the same time is the boundary between salty and fresh-water. (See on the hydrogeological cross-section Fig. 3. designated with "$A-B$", marked by a resultant line.)

Relationship among aquifers occurring in different depths was examined for the first time by method of continual approaches after E. Almássy (1962). It was established that static pressure of the Quaternary, Levantian and several Upper Pannonian beds of the area shows regular alternations which have relationship with morphology of the "impermeable floor" of the porous complex. Above the uplifted basement the static water levels of the deeper wells are lower-, and on the depression they are higher than the static water-levels of the shallow wells. The static level setting in at different points of the aquifer complex, or formation pressure-values are belonging to the hydrostatic pressure state. The vector sum determined is the so-called "piezometric gradient", analogically with the hydraulic gradient, derived from the potential theory of the seepage. When the formation pressure is measured the vertical component of the piezometric gradient is

$$[\overrightarrow{\text{grad u}}]_z = \frac{10 \cdot \Delta p}{\gamma \cdot \Delta z} \frac{\gamma_v}{\gamma_0}$$

where Δp is the measured pressure difference in [atm], z is the difference in drainage elevation [m], v is the mean specific gravity of the water head [p/cm³], γ_0 is equal 1.0 [p/cm³] of the specific gravity of the water on 4 °C temperature. When the specific gravity of the water is unit, then this value is directly determinable from the piezometric surfaces:

$$[\text{grad u}]_z \approx [\text{grad}]\ h_z = \frac{\Delta h}{\Delta z}\ .$$

If in some chosen horizontal planes (in the so called "drainage horizons") the values of the piezometric surfaces reduced to the water having an unit specific gravity are available (see in Fig. 4), then it can be obtained average values of the vertical components of the piezometric gradient regarding certain depth intervals as a differential quotiens of the contour lines of the suitable piezometric surface maps. Values of the horizontal components can be directly measured from the individual water level contour maps, and average values referring to the intermediate depth parts can be determined by arithmetical mean of two adjacent h/x or h/y gradient maps (I. LORBERER-SZENTES and Á. LORBERER 1976).

The Fig. 5. shows the differential quotiens of contour maps, namely the average values of the vertical components of the piezometric gradients between ± 0 and $+50$ m above Adriatic sea. On the highlands and hilly elevated areas, forming the main reproduction bases of the ground waters, the values of the gradients are negative, in the depressions they are positive, referring to the decisive horizontal flow and descendent as well as ascendent water movement.

If the surface or near surface derived colder waters are descending, then the rocks which are connecting with them will be cooling down and due to the heat-extraction they will be converted into thermal waters. If ascendent water flow occurs then a heat emission will take place. Consequently, a vertical water movement is followed by local negative as well as positive geothermal anomalies due to the convective heat transport. On the map of the apparent geothermal gradients (Fig. 6) the areas of negative heat anomaly (larger than 20 m/°C value patches) generally coincide with areas of negative vertical piezometric gradients. On the area of positive heat anomalies with 10 m/°C or less values, the vertical components of the piezometric gradients are also positive.

The most conspicuous common feature of the piezometric gradient and geothermal as well as water quality maps (Figs. 5–7) is the definite structural orientation.

The piezometric pressure profiles (Fig. 3a and 3b) represent not only potential distribution of the investigated area, but they also give data regarding universal characteristics of the events which had taken place within the basin

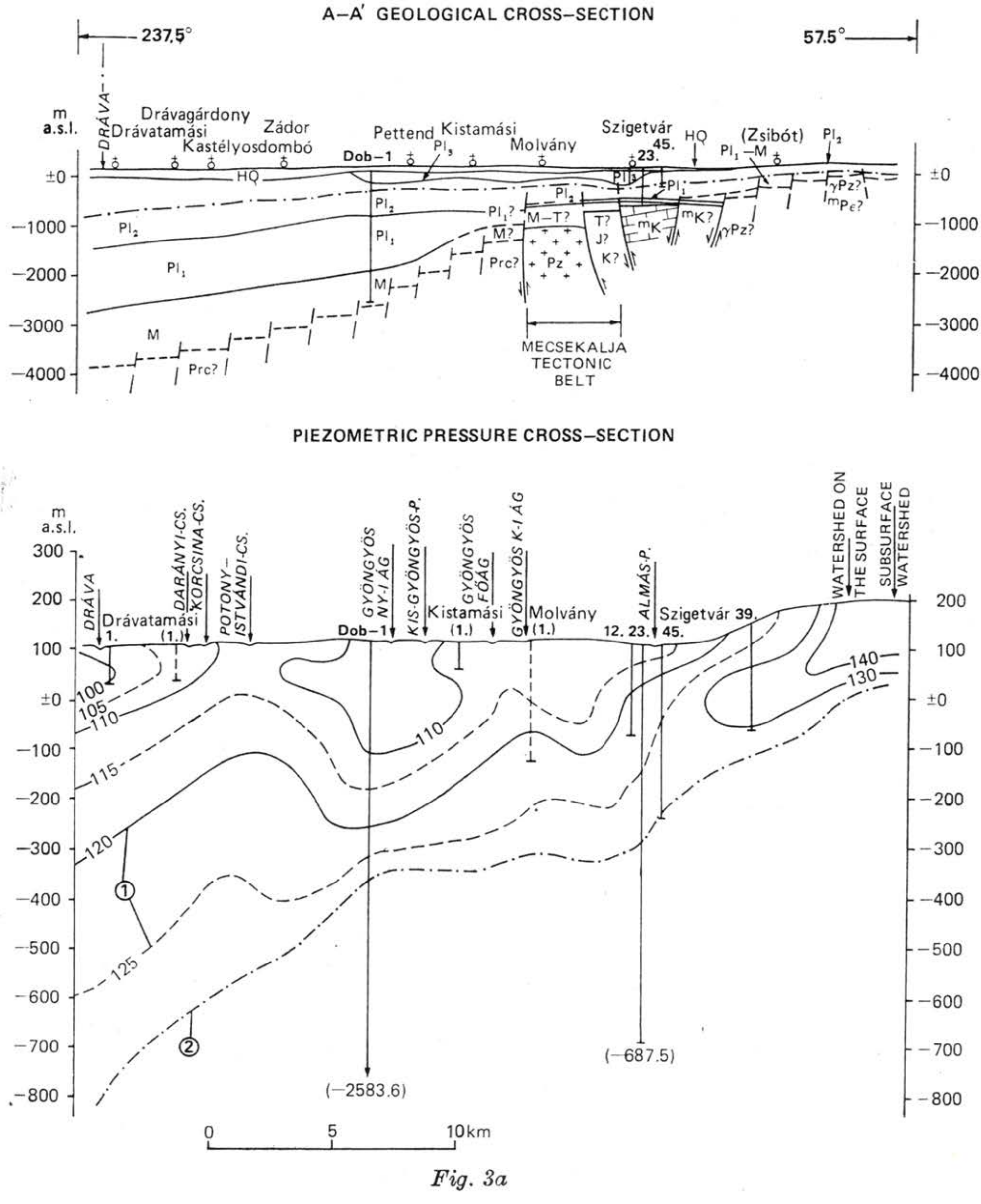
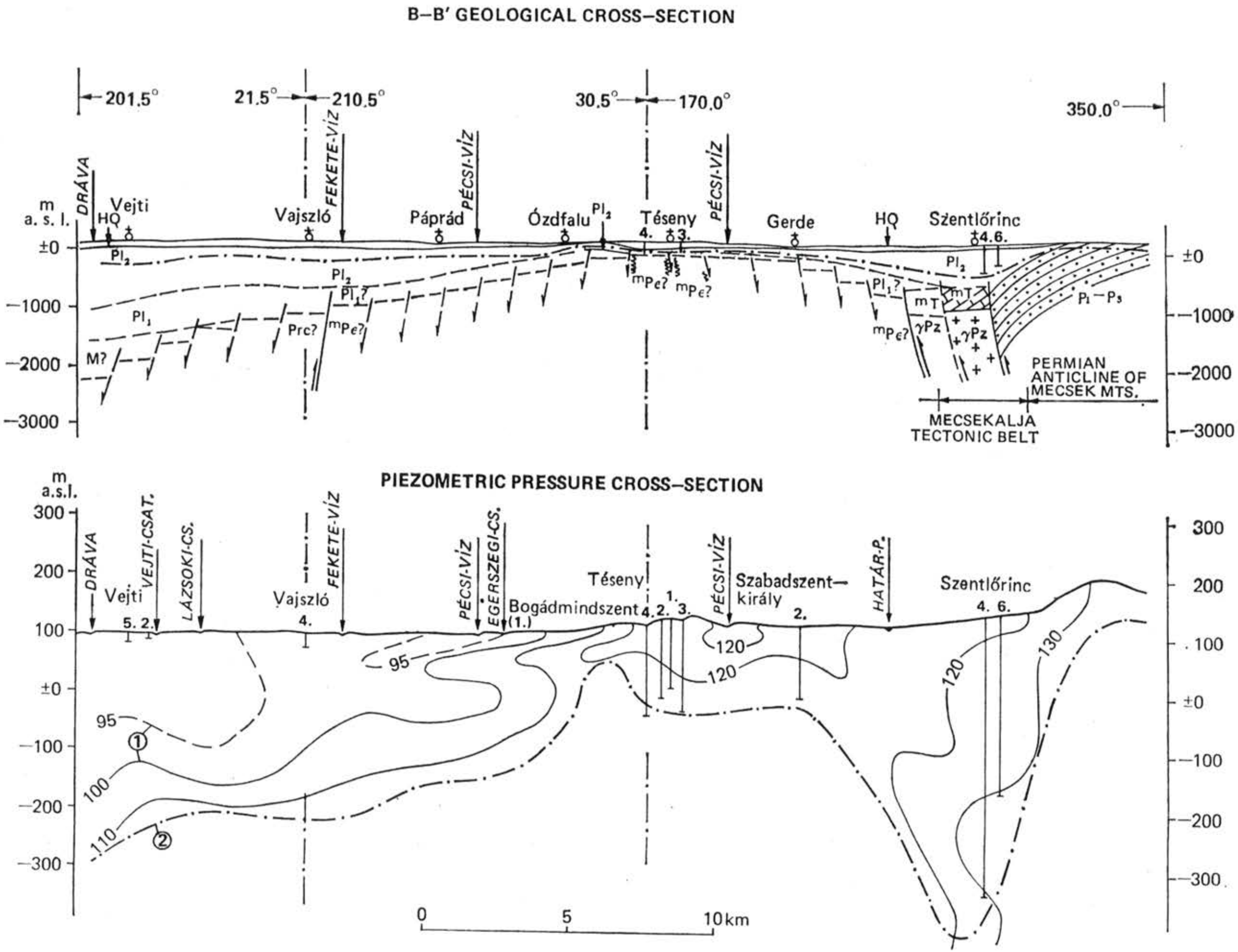

A–A' GEOLOGICAL CROSS–SECTION
237.5°
57.5°
DRÁVA
Drávagárdony
Drávatamási
Zádor
Kastélyosdombó
Pettend
Kistamási
Molvány
Szigetvár
(Zsibót)
Dob–1
MECSEKALJA
TECTONIC
BELT
PIEZOMETRIC PRESSURE CROSS–SECTION
DARÁNYI-CS.
KORCSINA-CS.
POTONY–
ISTVÁNDI-CS.
GYÖNGYÖS
NY-I ÁG
KIS-GYÖNGYÖS-P.
GYÖNGYÖS
FŐÁG
GYÖNGYÖS K-I ÁG
ALMÁS-P.
WATERSHED ON
THE SURFACE
SUBSURFACE
WATERSHED
(–2583.6)
(–687.5)
0
5
10km

Fig. 3a


B–B′ GEOLOGICAL CROSS-SECTION
201.5°
21.5°
210.5°
30.5°
170.0°
350.0°
m a. s. l.
DRÁVA
HQ
Vejti
Vajszló
FEKETE-VÍZ
Páprád
PÉCSI-VÍZ
Ózdfalu
Téseny
Gerde
Szentlőrinc
PERMIAN ANTICLINE OF MECSEK MTS.
MECSEKALJA TECTONIC BELT
±0
−1000
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−3000
PIEZOMETRIC PRESSURE CROSS-SECTION
m a.s.l.
VEJTI-CSAT.
LÁZSOKI-CS.
EGERSZEGI-CS.
Bogádmindszent
Szabadszent-király
HATÁR-P.
300
200
100
−100
−200
−300
0
5
10km

Fig. 3b

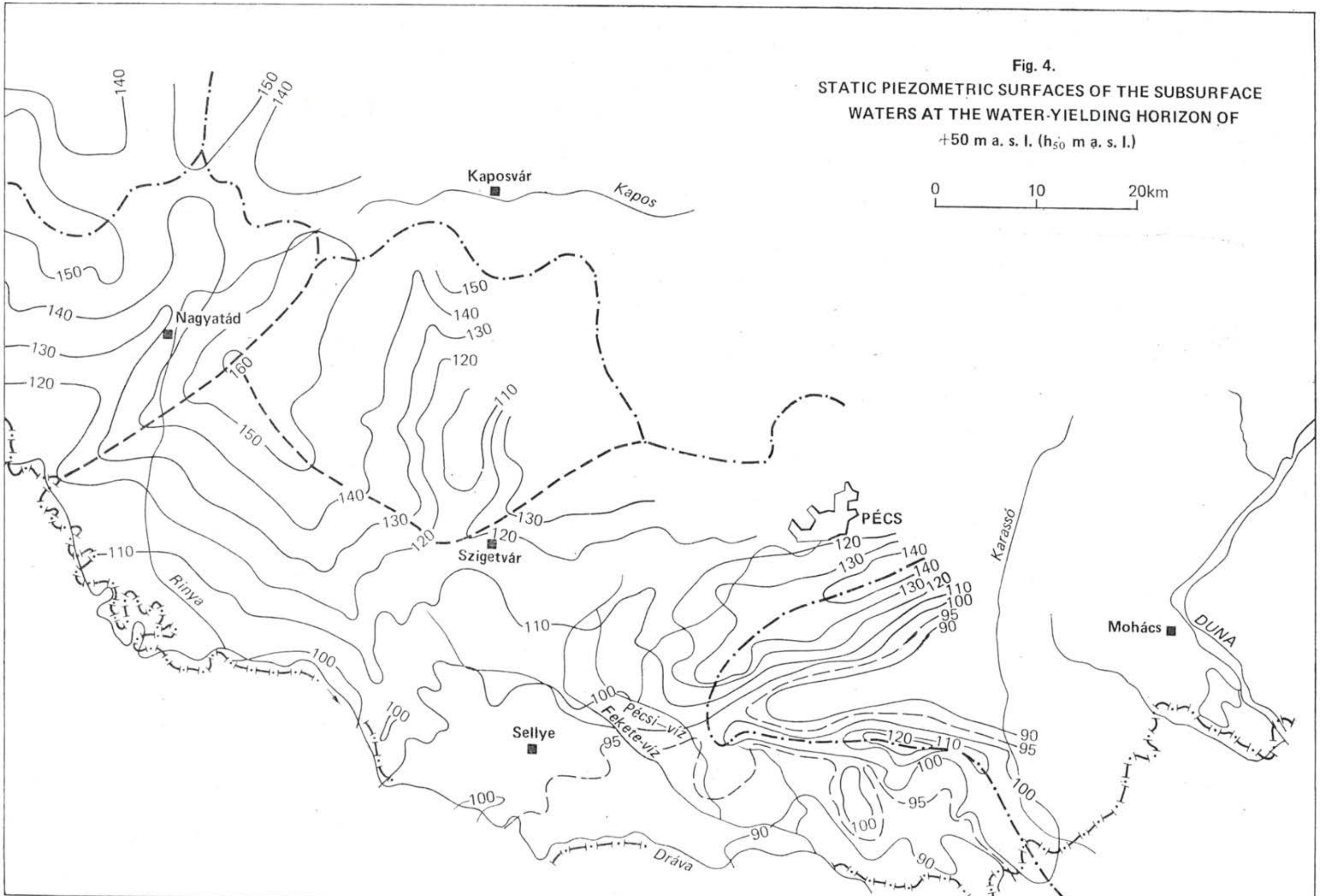

Fig. 4.

STATIC PIEZOMETRIC SURFACES OF THE SUBSURFACE WATERS AT THE WATER-YIELDING HORIZON OF +50 m a. s. l. (h_{50} m a. s. l.)

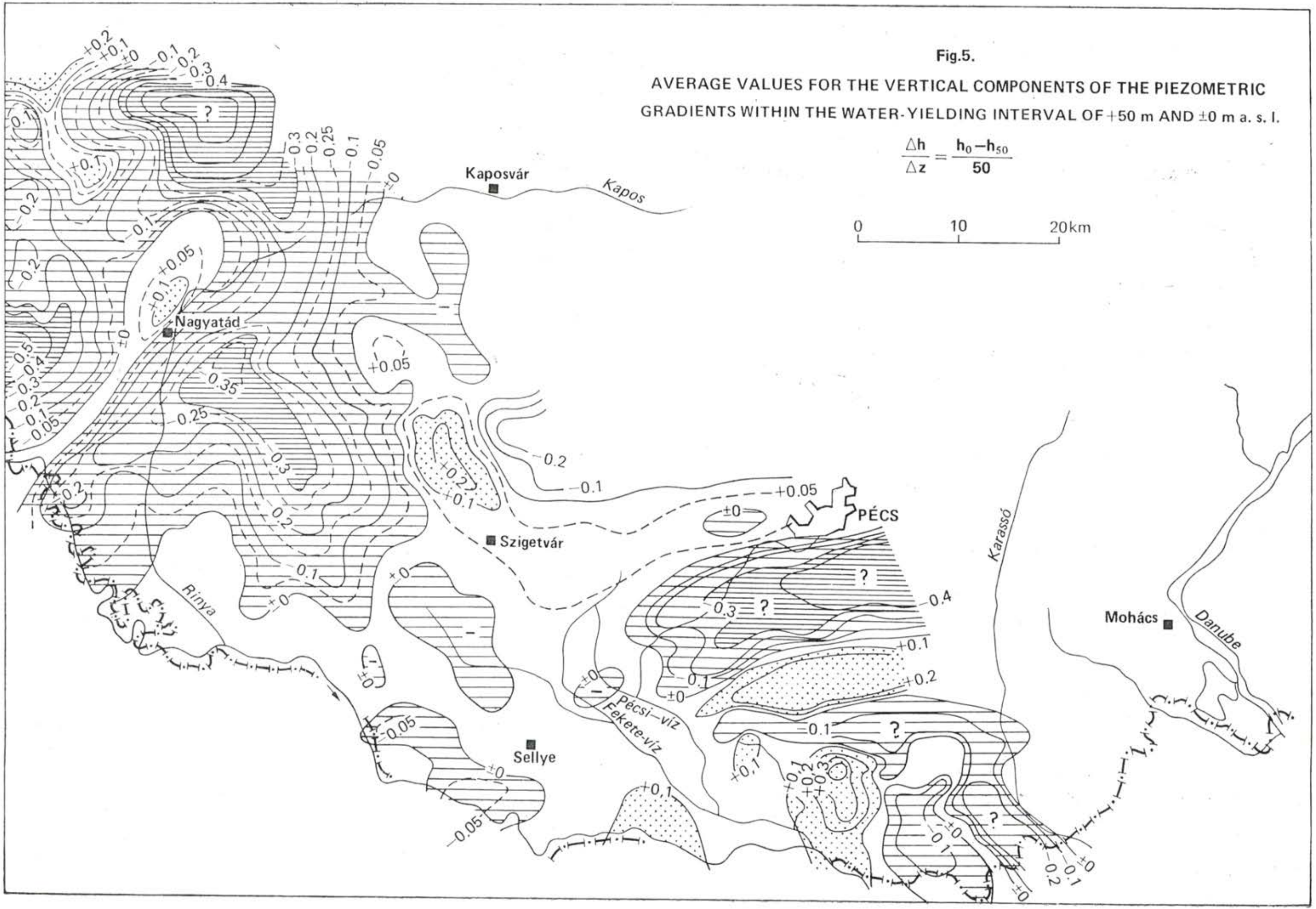

Fig.5.

AVERAGE VALUES FOR THE VERTICAL COMPONENTS OF THE PIEZOMETRIC GRADIENTS WITHIN THE WATER-YIELDING INTERVAL OF +50 m AND ±0 m a. s. l.

$$\frac{\Delta h}{\Delta z} = \frac{h_0 - h_{50}}{50}$$

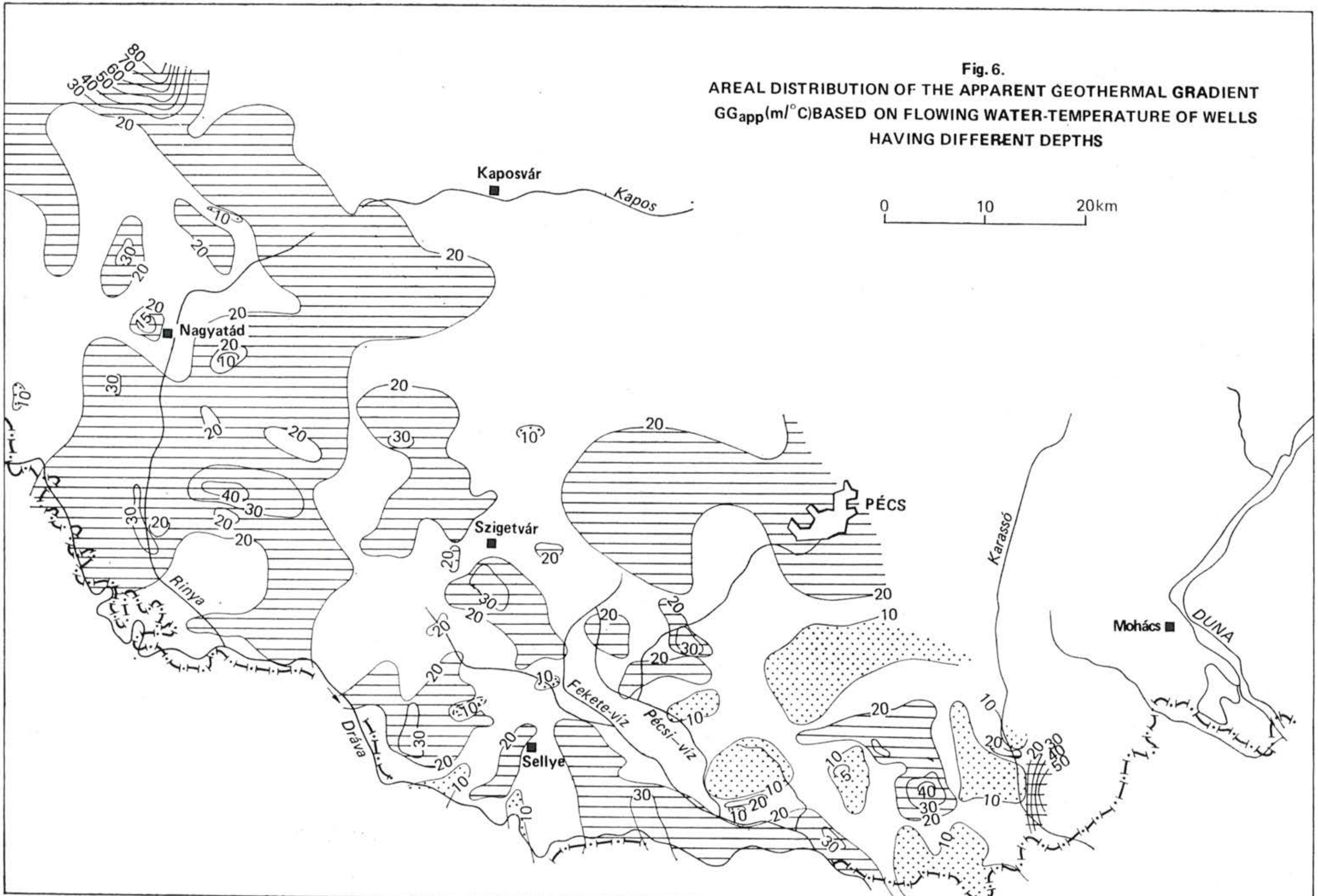

Fig. 6.
AREAL DISTRIBUTION OF THE APPARENT GEOTHERMAL GRADIENT GG_{app}(m/°C) BASED ON FLOWING WATER-TEMPERATURE OF WELLS HAVING DIFFERENT DEPTHS

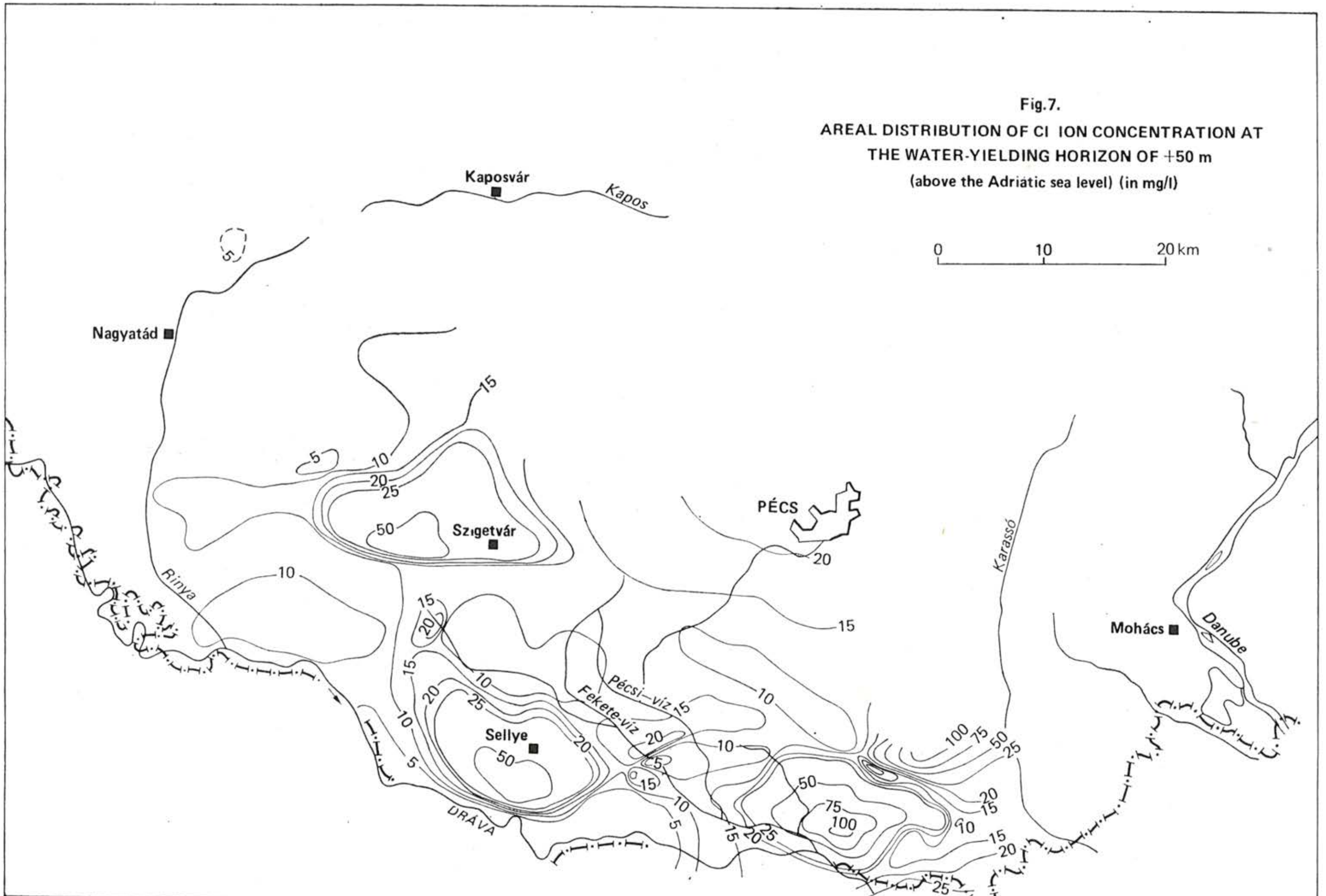

Fig.7.

AREAL DISTRIBUTION OF Cl ION CONCENTRATION AT THE WATER-YIELDING HORIZON OF +50 m (above the Adriatic sea level) (in mg/l) deposits. Since the multiannual average value of the natural subsurface flow conditions can be regarded constant, the potential distribution represented by water level contour maps and piezometric pressure profiles, corresponds to the permanent seepage condition. With respect to the flow conditions, most decisive proofs are provided by local geothermal anomaly which can be exclusively explained by convective heat-transport. That is ought to observe a negative heat anomaly in the depression having a larger thick deposit mass, on the basis of the heat-flow conductivity, — as it was indicated by heat-flow measurements made in the deeper horizons by T. BOLDIZSÁR (1964) and L. STEGENA (1973). We have not sufficient data available to determine the seepage velocity on the basis of the change of the heat-flux (I. LORBERER-SZENTES—Á. LORBERER 1976); — in this respect the investigated area is too complicated.

Connection of the ground water resources of the basin deposits with the waters of the deeper reservoirs

According to the investigations, along several main tectonic lines the local pressure-, temperature and hydrochemical anomalies can only be explained by hydrodynamic communication between basin deposits and deeper thermal water reservoirs.

It is well known that in a flow system the effect of the local draining and the surface relief have less and less scale (M. ERDÉLYI—GY. KOVÁCS et al. 1972). Since a considerable part of the water-mass flows in subsurface layers having a better transmissibility, therefore approached to the "impermeable floor", the absolute values of the piezometric gradients also decrease.

On the other hand, under descending conditions (e.g. Szigetvár region), the positive values of the vertical components of the piezometric gradients become higher and parallelly with this phenomenon, the geothermal gradients which can be calculated from deeper wells are decreasing; thus the heating effect does not decrease but it parallelly increases with the depth. It was conspicuous that change of the Cl^- ion concentration in several wells (along with 120—350 mg values) drilled on Pannonian layers, which proportionally was heightening by the increase of drawdown. These wells are located in a small zone of the southern part of the town about 1.5 km distant from the thermal well of Szigetvár (which is situated in the northern part of the town) could not exert influence on the latter, because they were constructed earlier. Therefore it is acceptable that the Albian limestone—which comprises the basement below Szigetvár—with the connecting deeper porous thermal water reservoir, also under natural condition provides a supply basis for the basin deposits (L. RÓNAKI—T. SZEDERKÉNYI 1966). Fig. 3a shows this upflow along the fracture, which is not too considerable under natural condition, but during the intensive water production from the subsurface layers, destructive changes can be taken place in the water quality of the "productive layers" because of the vertical water flow (Á. LORBERER 1975).

The extreme positive heat anomaly is restricted on the buried carbonatic mass of Harkány.

Based on the current system of the karstic waters of the Villány Mesozoic zone, it can be supposed that the waters derived from Villány Mts. were descended on a nine-times larger area as a free infiltration one. The concentrated upflow of the heated karstic waters proves a fundemental hydrogeological role of the Villány–Szalatnak Paleozoic Deep Fracture Zone (M. KASSAI 1972).

It is proved by above mentioned reasons that on the marginal area of the Great Hungarian Plain and Drava Basin the hydrodynamic connections between basin deposits and deep thermal water reservoirs are linked to the main marginal structural lines of great basins, and further research trends are determined by this fact.

Conclusions

1. On the area of the great Neogene depressions of Drava Basin, the piezometric gradient maps and water-quality maps drawn to different drainage horizons, as well as the areal distribution of the geothermal gradients show an appreciable tectonic orientation which first of all proves a decisive effect of the neotectonic fracturing on the flow-relations of the ground waters.

The large tectonic zone striking NW–SE direction, appearing at the eastern wedging boundary of the basin deposits can generally be characterized by ascendent waters and positive geothermal anomalies (Szigetvár, Harkány).

2. It can be directly proved the connection between basin deposits and deeper reservoirs which is realized by marginal fractures of the basin. The proofs are sudden changes in pressure, temperature and water quality.

3. According to the results of these investigations, thermal water yield may be most successful and economical on the marginal parts of the great basin.

4. In generalizing the above-mentioned statements it can be said that in the great basins thermal water research possibilities are connected with marginal fracture zones of the basins, if the basement is a suitable thermal water reservoir.

5. The research referring to the area of SE Transdanubia as a part of marginal area of the Great Hungarian Plain and Drava Basin proves a feasibility of the methods involving applied hydrogeology and regional hydrogeological mapping, if we have suitable geological maps and aspect.

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CARACTÉRISTIQUES HYDROGÉOLOGIQUES DE LA PLAINE DE NEGOTIN (BORDURE DU BASSIN ARTÉSIEN DU DANUBE)

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Introduction

La région étudiée s'étend sur les parties extrêmes à l'Est de la Yougoslavie, encadrant la plaine de Negotin, laquelle représente, au point de vue hydrogéologique, une partie très éloignée à l'W du bassin de Dacia.

Il est probable qu'il n'existe pas en Serbie d'autres régions si riches en eaux souterraines dans la nappe phréatique. Pourtant, la ville de Negotin et un grand nombre de ses colonies, sont touchées par la pénurie de l'eau potable (les eaux de la nappe phréatique provenant d'une grande partie de la plaine, sont contaminées, tandis que celles artésiennes sont assez profondes ayant un faible débit).

Cependant, les parties plus basses de la plaine sont inondées en hiver et au printemps, tandis qu'en été les terrains deviennent tellement secs ce que touche considérablement la production agricole.

Caractéristiques géomorphologiques de la plaine de Negotin

Les recherches les plus récentes du paléorelief et du relief d'abrasion de Timočka Krajina prouvent la présence des terrasses d'abrasion assez basses, situées au niveau topographique de 343 à 360 m, développées pendant la phase marine-lacustre à partir du Méotien durant tout le Pliocène. Au temps présent, les terrasses sont disparues ; elles sont détruites par une érosion latérale très intense du Danube, ce qui explique la formation d'une falaise actuelle s'étendant vers le fond de la plaine Vlaska — Pont. Les traces de la coupure latérale de la bordure du bassin ancien, sont mises en évidence par les terrasses du Danube dont la hauteur rélative dépasse 65 m. Dans la plaine de Negotin leur hauteur absolue est de 100 à 110 m environ. Les sédiments tertiaires se sont déposés sur un grand plateau prolongeant entre les chaînes de montagne (à l'W Miroč, Deli Jovan et d'autres) et les fleuves Danube et Timok à l'E, puis de Sip au N jusqu'aux environs de la ville Zaječar au S. On peut constater sur place quelques affleurements des roches primaires schisteuses du Paléozoïque et des serpentines redressant de cette série des roches non-consolidées, consolidées et demi-consolidées.

La formation du Danube a été prédisposée par la régression de la mer pliocène. Sur un grand plateau au fond du bassin, de Djardap vers l'Est et le Sud, le Danube avait déplacé son lit, en érodant et transportant les sédiments détritiques du Néogène. Les variations climatiques très fréquentes avaient influencé beaucoup sur l'évolution du lit, sur l'écoulement et le niveau du Danube, ce qui favorisait certainement le développement intensif de l'érosion fluviatile et l'accumulation. Ainsi, l'érosion avait emporté les roches moins résistantes, celles du Miocène et Pliocène (Dunavski Ključ et la plaine de Negotin).

Géologie

La plaine et sa bordure se différent essentiellement, aussi par leur composition géologique.

La bordure de la plaine est représentée, en quelque part, par des schistes cristallins précambriens (gneiss, micaschistes, amphibolites), souvent percés par des filons de granite et de gabbro. La région au versant de Ravna Reka et Vratna est constituée de diabases et de phyllitoïdes; les massifs des montagnes: Djerdap, Miroč, Veliki Greben et Deli Jovan, sont formés de conglomérats liassiques, de grès, argiles et calcaires. Le Jurassique moyen est représenté par des calcaires, conglomérats et grès, tandis que ce sont les calcaires récifaux qui représentent le Jurassique supérieur. Les sédiments du Crétacé constituent, en général, une plus grande partie de la bordure de la plaine de Negotin.

Les sédiments du Tertiaire sont transgressifs. Cette série comprend les couches prolongeant du Méditerranéen II jusqu'au Pliocène plus jeune, sans aucune observable discontinuité de sédimentation. Toutes les couches du Tértiaire ont un pendage vers l'Est; les couches anciennes sont plus inclinées que celles plus jeunes. Les couches du Méditerranéen II sont les plus anciens sédiments du Tértiaire constaté en Serbie NE. Elles reposent transgressivement sur les formations anciennes et sont recouvertes à leur tour par des sédiments sarmatiens.

Leur transition est graduelle, marquée par les argiles des couches de Buglovka. Le Méditerranéen II est représenté par des argiles, grès, calcaires de Leitha. La bordure de la plaine renferme des couches du Sarmatien dont la répartition horizontale est très grande, tandis que celle verticale est assez réduite. Le Sarmatien inférieur abonde en argiles souvent sablonneuses et micacées; les sables de différents grains et les calcaires représentent le Sarmatien moyen, tandis que les sables aux intercalations de calcaires sablonneux représentent le Sarmatien supérieur.

La plaine de Negotin. Après avoir fait un profond forage à Prahovo, on a constaté un développement complet du Néogène prolongeant à partir du Méditerranéen II jusqu'au Quaternaire d'une épaisseur de 1000 m. Le mur du Néogène aussi même que toute la plaine, est constitué des sédiments de la « Série de Sinaïa ». Les sédiments du Méotien, Pontien et ceux du Pliocène supérieur et Quaternaire en forme de terrasse se sont déposés au-dessus d'une complète série du Miocène, identique à celle de la bordure de la plaine. C'est-à-dire le Méotien est formé des argiles, calcaires, sables et grès; le Pontien est constitué des argiles et celles sablonneuses et enfin, ce sont les graviers, les sables et ceux argileux qui entrent dans la constitution du Pliocène supérieur et du Diluvium. L'Alluvium est représenté, en général, par des grès et des sables.

Tectonique

Dans cette région de la Serbie, les mouvements tectoniques très remarquables sont liés à trois phases.

I. Après la sédimentation des couches de la « Série de Sinaïa » qui s'est faite à la fin du Crétacé supérieur et au Paléogène, le flanc occidental de la faille s'est baissé profondement en direction de la bordure de la plaine; ainsi, les sédiments crétacés se trouvent écroulés plus de 900 m en profondeur.

II. Les mouvements tectoniques survenus dans le Sarmatien supérieur et le Méotien (la phase attique) ont fissuré les couches du Sarmatien en les faisant inclinées vers l'Est.

III. Les couches du Méotien et Pontien sont bouleversés et inclinés vers l'E par la troisième phase, la rhodanienne.

Caractéristiques hydrogéologiques de la plaine de Negotin

La bordure de la plaine

D'après la porosité des roches, déterminée lithologiquement, et d'après le caractère de la nappe, on y peut distinguer trois types de nappe:

— Le nappe de fractures dans les couches de la « Série de Sinaïa »;
— La source d'éboulis présenté dans les sédiments sarmatiens
— La nappe composée dans les sédiments sarmatiens d'une porosité fissurée et intergranulaire.

La nappe de fractures est développée dans les sédiments de la « Série de Sinaïa » (calcaires, grès, marnes et argiles) qui sont bien cassés et décomposés par les mouvements tectoniques. Bien que, dans les parties superficielles de ces roches mêmes, existent certains systèmes de fissures et de failles, ce sont elles qui représentent un mur qui est le plus souvent imperméable pour les couches aquifères du Sarmatien. Dans les vallées raidement creusées par les cours des eaux superficielles, dont les lits se trouvent coupés dans les roches à la profondeur de 30 m, on peut remarquer rarement des zones humides situées 2 à 3 m sous le contact de ces mêmes roches et des couches aquifères du Sarmatien. Dans les couches de la « Série de Sinaïa » l'accumulation des eaux souterraines est tellement petite qu'on peut à peine trouver une source dont le débit dépasse plus de 0.05 l/sec. Bien que ces roches abondent en sources, en été elles deviennent sèches.

La source d'éboulis est développée dans les roches du Sarmatien (graviers, sables, conglomérats et grès) dont la porosité est intergranulaire. Ce type de nappe est dominant dans la bordure de la plaine, et par conséquent, il est le plus important au point de vue de son débit. Dans la vaste bordure de la plaine (plus de 2 km de la largeur) cette nappe est représentée par des graviers sarmatiens et des sables dont l'épaisseur dépasse plus de 30 mètres.

Dans les parties plus profondes constituées des graviers a été formée une unique source d'éboulis. En outre que les caractéristiques hydrogéologiques y sont très favorables (composition granulométrique, porosité, perméabilité et d'autres) on peut dire que cette nappe n'a pas un grand débit, mais qu'elle est pauvre en eaux souterraines. Les sédiments qui constituent la zone d'aération de cette nappe sont assez perméables, d'où résultent une percolation plus vite des eaux superficielles et leur écoulement. Cependant, de minces intercalations et des lentilles d'argiles divisent les couches de la zone d'aération à un nombre de petites nappes, c'est-à-dire des nappes suspendues. Dans la pente de la plaine où s'apparaissent les argiles à la surface de nombreuses tranchées, les nappes situées au-dessus de ces terrains donnent la naissance aux sources dont le débit dépasse souvent plus de 5 l/sec. Telles sont les conditions hydrogéologiques dans la zone d'aération de la principale nappe des graviers sarmatiens et des sables au fond desquels se trouvent les couches imperméables de la « Série de Sinaïa ». C'est la raison d'une alimentation très faible de cette nappe-ci. C'est pourquoi qu'à la bordure constituée des couches gréso-sablonneuses sarmatiennes affleurent, en général, des couches argileuses du Sarmatien d'une petite épaisseur, dont le mur imperméable est presque complètement aplani, composé aussi des couches de la « Série de Sinaïa » (Fig. 1). De nombreuses sources situées dans la bordure de la plaine drainent une grande quantité des eaux de cette nappe.

Le mur imperméable s'élève au Nord (les couches de la « Série de Sinaïa »), d'où résulte une épaisseur réduite des sédiments perméables du Sarmatien. La présence des intercalations et des lentilles d'argiles est assez rare dans cette série. Le drainage des eaux souterraines se fait dans le mur imperméable par la source ou le long de la faille en alimentant des aquifères profondes (artésiennes).

Dans ce complexe des sédiments sarmatiens au Sud prédominent des argiles; c'est pour cela qu'on y trouve de plusieurs nappes superposées, drainées par un grand nombre de sources à très faible débit. Répandus sur un grand espace à l'W, les sédiments sarmatiens ont une épaisseur assez réduite.

La nappe composée s'étend sur une grande partie de la bordure norde de la plaine. Les couches perméables de cette nappe sont constituées des grès, calcaires et conglomérats du Sarmatien. A l'Est de la faille qui fissure ces couches à la ligne de la tranche, cette nappe très puissante est barrée par des argiles du Mio-Pliocène. Une petite quantité de ses eaux est drainée par la source Bolbolos, tandis qu'en profondeur la nappe est reliée aux sables des horizons artésiens trouvés à l'autre côté de la faille, — dans la plaine de Negotin.

Les sédiments sarmatiens de cette région se caractérisent par l'existance d'une unique nappe composée qui renferme des roches fissurées dont la porosité est intergranulaire. Dans la zone où cette nappe se trouve au contact direct avec les horizons aquifères artésiens s'était faite une liaison hydraulique.

La plaine de Negotin

La nappe phréatique. Après avoir fait une analyse des données de forage, on a constaté le concours des graviers dans la couche aquifère plus de 70 % ainsi que certaines régularités dans leur sédimentation. Au-dessus des argiles néogènes se sont déposés des graviers, ensuite des graviers sablonneux, et des

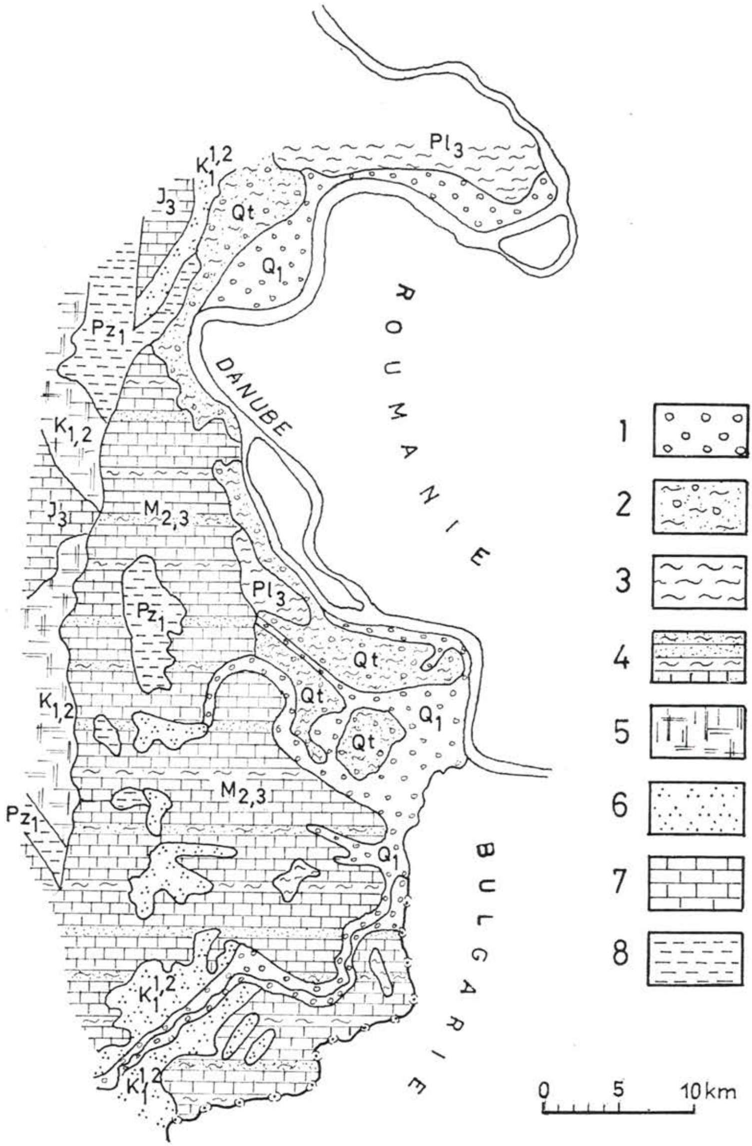

Fig. 1. La carte géologique de la plaine de Negotin et de sa bordure

1. Pléistocène: gravier, sable et argile, *2.* les sédiments terrassiques: gravier, sable, argile, loess et sable boulant, *3.* Pliocène: argile (marneuse, sablonneuse, graveleuse), *4.* Miocène: calcaire, argile, conglomérat et grès, *5.* Crétacé: calcaire massif, *6.* les couches de Sinaïa: marne, calcaire, grès et argile, *7.* Jurassique: calcaire, *8.* Paléozoïque: micaschiste

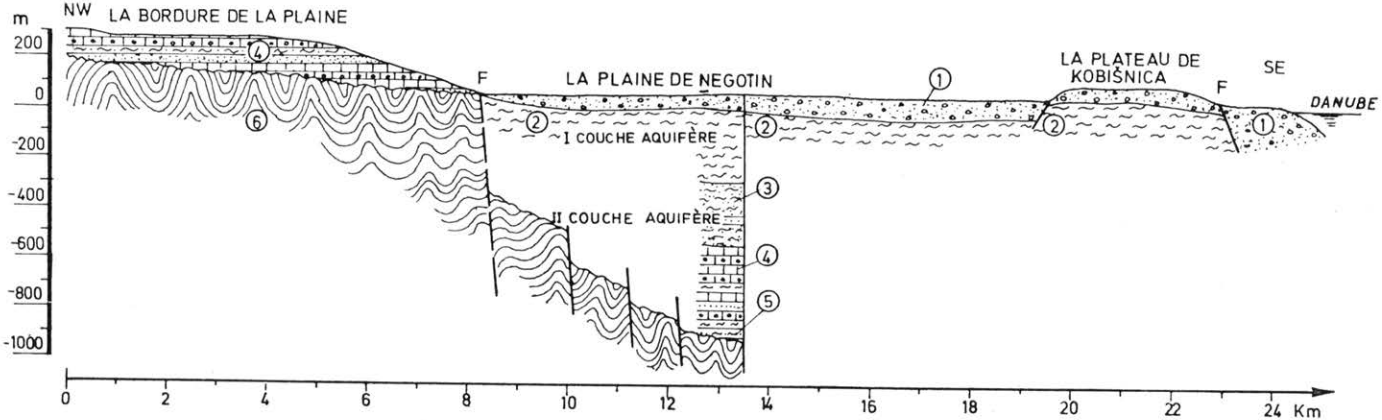

Fig. 2. Coupe hydrogéologique de la plaine de Negotin et de sa bordure

1. Quaternaire: sable, gravier et argile, *2.* Pontien: argile (marneuse, sablonneuse, graveleuse), *3.* Méotien: sable, argile, sables argileux et gravier, *4.* Sarmatien: calcaire, conglomérat, argile, grès et sable, *5.* Méditerranéen II: argile, *6.* les couches de Sinaïa: marne, calcaire, grès et argile sables, et au sommet des argiles sablonneuses et des sables argileux. L'épaisseur des graviers et des graviers sablonneux au centre de la plaine dépasse souvent plus de 20 mètres; le plus souvent elle varie de 10 à 12 mètres. Vers la bordure occidentale de la plaine l'épaisseur se réduit, tandis que les couches aquifères des graviers touchent au SW la bordure de la plaine. Au Nord et à l'Est de la plaine les graviers sont entaillés par des cours d'eau (le Danube près de Prahovo) ou ils sont traversés par des cours des fleuves (le Danube près de Radujevac et Timok à l'embouchure du Danube). Dans les parties plus basses de la plaine on peut remarquer les variations fréquentes de la composition lithologique et granulométrique des sédiments. En effet, toute la plaine est constituée des graviers et des sables d'Alluvium où s'était formée une unique nappe phréatique dans les roches d'une porosité intergranulaire.

Les coefficients de filtration des sédiments de la couche aquifère ont été déterminés par des essais de pompage (62 essais); les valeurs obtenues se trouvent entre 1.81. m/jour et 387 m/jour. La valeur moyenne du coefficient de filtration est 82.7 m/jour.

Régime des eaux souterraines

Par rapport aux relations réciproques des eaux souterraines et superficielles de la nappe phréatique, on peut distinguer deux types du régime des eaux:

— un régime des eaux souterraines reliées, au point de vue hydraulique, aux cours des eaux superficielles;

— un régime des eaux souterraines indépendantes, au point de vue hydraulique, des cours des eaux superficielles.

1. *Le régime des eaux souterraines reliées*, au point de vue hydraulique, aux cours des eaux superficielles a été formé dans les environs de grands cours d'eau (Danube, Timok, Jasenička et Dupljanska reka). Durant presque toute l'année, ces cours d'eau drainent tout au long de la nappe phréatique, mais pendant la crue ils représentent les retenues en alimentant ainsi la nappe bien au loin des cours d'eau. Les eaux des petits affluents du Danube s'infiltrent continuellement le long des cours d'eau en alimentant cette nappe peu profonde dans la plaine. Leur participation dans l'alimentation est variable, qui dépend certainement du niveau des cours d'eau, du niveau de la nappe et de la perméabilité des sédiments aquifères.

Le niveau des eaux souterraines varie dans l'espace et dans le temps. On a constaté en général une remontée et un abaissement périodique du niveau des eaux au cours de l'année. Du mois janvier jusqu'à l'avril et mai la remontée du niveau est assez visible, suivie par une période d'un lent abaissement. Cependant, dans la région près du Danube, sauf une amplitude assez grande, la régularité déjà citée est troublée seulement dans le cas où les lignes de la remontée et l'abaissement du niveau sont variables. Les variations positives et négatives du niveau des cours d'eau sont assez fréquentes ce qui a un grand influence sur le niveau des eaux souterraines. On peut y ajouter une perméabilité assez grande des couches aquifères. L'amplitude du niveau des eaux du Danube au cours de l'année est souvent 9 mètres, tandis que celle moyenne varie de 7 à 8 mètres. A l'exception de la zone étroite près du Danube (environ

150 mètres) où les amplitudes du niveau du fleuve et de la nappe sont approximatives, le niveau des eaux varie entre 1 et 4 mètres dans la zone où les eaux du Danube ont une grande influence aux eaux souterraines à la distance de 2 km. En vue de la retenue sur l'espace près de Radujevac on peut constater une grande influence des eaux du Danube à la distance de 2 km et plus de 3 km de ses côtiers.

Dans certaines régions assez éloignées du Danube, les cours superficielles — les affluents du Danube — tiennent leur part à l'alimentation et au régime du niveau de la nappe. Une perméabilité assez grande des sédiments au fond du lit des affluents liées aux graviers de la nappe, ainsi qu'une petite inclinaison des lits longitudinaux, permettent une infiltration très intensive dans la nappe. En général, la variation du niveau des eaux souterraines dans le temps correspond aux périodes de la crue. Au printemps, les niveaux sont les plus élevés, tandis qu'en hiver ils sont les plus bas. L'amplitude la plus grande dépasse 4 m.

2. *Le régime des eaux souterraines indépendantes*, au point de vue hydraulique des cours des eaux superficielles est développé dans la bordure de la plaine et sur la plateau de Kobišnica.

Sur le plateau de Kobišnica les eaux souterraines coulent du centre vers la bordure dans tous les sens, et étant donné que les plateau est incliné vers le Sud, les niveaux des eaux souterraines sont d'une profondeur très variable de 10 à 20 mètres. Dans la partie plus extrême du Sud, où le plateau se transforme doucement à une plaine élevée, la profondeur qui atteint le niveau des eaux souterraines varie de 2 à 5 mètres, c'est-à-dire 10 mètres.

La fluctuation du niveau des eaux souterraines n'est pas si expressive ce qui est à cause de l'alimentation efficace (y compris seulement la précipitation) et des propriétés hydrogéologiques des sédiments dans la zone d'aération et dans la nappe.

Le drainage de la nappe du plateau de Kobišnica se fait par de nombreuses sources pérennes de débit fort, concentrées au fond de la falaise au Nord de la plaine, par des sources intermittentes (pendant la crue) dans la falaise vers l'Est, ainsi que vers le Sud et l'Ouest où les eaux de cette nappe coulent dans la plaine élevée, en faisant les parties plus grandes de la plaine marécageuses.

Caractéristiques qualitatives et quantitatives des eaux souterraines

Les variations de la température des eaux souterraines au cours de l'année dans les parties un peu plus profondes de la nappe phréatique sont de 9.9 à 11.8 °C.

Le débit de cette nappe est très variable:

— dans les parties périphériques vers l'W le débit des puits varie de 1 à 2 l/sec;

— en allant à l'E de cette nappe, le débit spécifique est variable en dépassant le plus souvent d'environ 10 l/sec pour un puits.

Le débit maximum des puits dans la nappe des sédiments alluviaux et terrassiques est aussi variable en oscillant de 0.5 à 36.0 l/sec. En ce qui concerne le chimisme, les eaux souterraines de la nappe basse près du Danube et dans les terrains plus bas, sont d'une bonne qualité, en général, tant pour les habîtants que pour l'agriculture. On les classe au groupe des eaux dont la minéralisation varie de 0.300 à 0.862 gr/l. La valeur de pH varie de 7.0 à 8.4. La dureté varie de 1.04 à 46.16° dH; donc les eaux souterraines sont assez tièdes et d'une dureté modérée.

Parmi les anions prédominent les ions d'hydrocarbonate dont la teneur varie de 13.74 à 75.09 %r. Dans tous les autres anions, le pourcentage de sulfates est un peu plus élevé, de 31.95 à 54.10 %r.

La teneur en Na + K est assez élevée, mais sans variation. Elle se trouve entre 50.00 et 70.00 %r.

Les ions de Ca sont représentés en teneur plus élevée que celle de Mg, bien qu'on puisse avoir des cas contraires. La valeur du Ca varie de 4.57 à 42.59 %r.

Sur la base des relations réciproques et des composants d'anions et cations prédominants, on peut distinguer certains types des eaux souterraines:

— le type hydrocarbonato-alcalin est le plus répandu;
— le type hydrocarbonato-calcium-alcalin trouvé plus proche de la bordure de la plaine.

La nappe artésienne

Les eaux de la nappe artésienne dans les sédiments néogènes de la plaine de Negotin ne sont pas bien recherchées, quoi qu'on les exploite de deux couches aquifères par un grand nombre de forages pour alimenter en eau des colonies; la première couche se trouve à la profondeur de 155 à 180 mètres, la seconde de 320 à 332 mètres.

Sur la base des données incertaines obtenues par le forage, on a construit la coupe lithologique des parties plus profondes de la série des sédiments néogènes dans la plaine de Negotin (le forage 3, au sud du village Miloševo), comme suit:

de la côte +52,00 á +44,00 m dépôts argileux,
+44,00 à −11,00 m argiles griso-bleus,
−11,00 à −13,00 m sables argileux,
−13,00 à −33,00 m conglomérats argileux,
−33,00 à −108,0 m argiles sablonneuses aux intercalations des conglomérats argileux,
−108,0 à −123,0 m sables aquifères.

Si l'on peut accepter les données authentiques, ainsi que le fait du forage toujours arrêté à la profondeur du premier horizon artésien, on peut alors supposer qu'à la distance identique du rivage de la mer pliocène (environ 2 km), la sédimentation s'était faite sous les conditions très uniformes. Par conséquent, à la distance du rivage de la mer pliocène, de 2 km, la première nappe se trouve à la profondeur de 155 à 180 mètres.

A Negotin, avant la I[e] guerre mondiale, ont été forés trois puits artésiens par lesquels on a découvert la couche aquifère à la profondeur de 320 à 332 m, étant sous la pression. Son faible débit (de 0.2 à 0.3 l/sec) influençait beaucoup à la solution de l'approvisionnement en eau de la ville de Negotin.

A Prahovo, en 1956, par un forage de reconnaissance, on avait déterminé l'épaisseur totale des sédiments néogènes et leur substratum – la « Série de Sinaïa ». La coupe du forage est la suivante:

de la côte + 59,00 à	+ 22,00 m	sédiments terrassiques (quaternaire)	Pontien
+ 22,00 à	+ 59,00 m	argiles graveleuses et sables griso-marneux	
– 59,00 à	– 66,00 m	argiles griso-bleues	
– 66,00 à	– 79,00 m	argiles griso-graveleuses	
– 79,00 à	– 305,00 m	argiles grises sablo-marneuses aux intercalations de minces couches de sables	
– 306,00 à	– 441,00 m	argiles sablo-marneuses	Méotien
– 449,00 à	– 528,00 m	argiles sablo-marneuses	
– 528,00 à	– 556,00 m	sables microgrenus	Sarmatien Tortonien Méditerranéen
– 556,00 à	– 991,00 m	calcaires, grès, graviers, conglomérats, argiles et sables	
– 991,00 à ...	m	calcaires, grès, marnes et argiles	« Série de Sinaïa »

Le débit de la nappe artésienne dans la plaine de Negotin varie au cours des ans. Aussi a été constatée la variation de la température des eaux des puits artésiens. Au cours des années 1964 et 1965 le maximum des variations saisonnières des forages artésiens dépasse 1 l/sec et la température de l'ordre 0.8 °C. Le débit des forages artésiens varie de 1 l/sec à 4.5 l/sec et la température de 14 à 15.1 °C. La composition chimique des eaux artésiennes est la même comme celle de la nappe phréatique. La minéralisation des eaux artésiennes est moins de deux tiers de celle constatée par l'analyse des eaux de la nappe phréatique. Les eaux du forage artésien dans le village Miloševo, dont le débit le plus grand est – 4.5 l/sec, ont la suivante teneur des principaux ions et cations: $Cl = 11.51$; $HCO_3 = 67.00$; $CO_3 = 6.94$; $SO_4 = 14.56$; $Ca = 7.05$; $Mg = 16.18$; $Na + K = 76.77$ % et le résidu sec 0.447 gr/l.

D'après le concours des composants chimiques, toutes ces eaux et les autres aussi analysées, appartiennent au type des eaux hydrocarbonato-alcalines ce qui est le cas des eaux de la nappe phréatique dans les terrains plus bas des terrasses et dans les couches alluviales.

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CHARACTERISTICS OF THE DEVELOPMENT OF GROUND WATER FLOW AND REGIONAL ESTIMATION OF NATURAL GROUND WATER RESOURCES IN OPEN-TYPE ARTESIAN BASINS

(Exemplified by the Moscow artesian basin)

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Studies of the regularities of ground water flow and estimation of fresh ground water resources, widely used for water supply, are of great scientific and practical importance to areas with platform-type large artesian basins where densely populated industrial regions are frequent.

An example of such a basin is the Moscow artesian basin in the centre of the European part of the U.S.S.R., a complicated multi-layer aquifer system confined to Mesozoic-Cenozoic and Upper Paleozoic sediments. The main fresh ground water resources over the largest area of the artesian basin are formed in Carboniferous and Devonian sediments composed largely of carbonate rocks. The studies, carried out in the Moscow artesian basin (Hydrogeology of the U.S.S.R., Vol. 1, 1966) allow to draw some conclusions concerning the characteristics of hydrodynamic conditions and the methods of regional estimation of natural ground water resources in similar artesian basins.

Dense and deep erosional ruggedness of the relief, facies-lithologic nonuniformity of water-bearing rocks, irregular distribution and sharply varying thickness of clay layers both in the strata of main aquifer systems of Carboniferous and Devonian age and in the overlying Mesozoic and Quaternary sediments, the presence of lithologic "windows" and zones with a disturbed structure are typical of the basin. The distinctive features of geomorphology and geological structure of the basin contribute to the intensive and deep infiltration from precipitation, moreover facilitate the hydraulic interconnection of separate aquifers and aquifer systems and their connection with surface streams and water bodies, and create favourable conditions of ground water recharge, flow and discharge. This allows to state that the Moscow artesian basin is in the same class with the type of open artesian basins. The open character of the basin, in turn, determines the nature of the aquifer interrelationship, the characteristics of ground water movement, the recharge and discharge conditions moreover the ground water composition and resources formation.

Head variation in separate aquifers is characteristic of artesian basins of the given type: in interfluve areas the piezometric levels fall with depth, in river valleys water head increases with depth. Such a regularity is disrupted only in the vicinity of large water outputs where water levels of the separate aquifers are appreciably lowered due to development. In the Moscow artesian basin, the difference in the water levels of principal Carboniferous aquifers and overlying Mesozoic and Quaternary sediments averages 10–15 m and more rarely 20–30 m (commonly in the case of insignificant thickness of separating clay layers) in watersheds. In river valleys, reverse level difference is observed as coming to minus 10–minus 15 m and sometimes to minus 30 m. The vertical hydraulic gradients vary from 0.2–0.5 to 3–5 in clay layers. Therefore, within the basin, the variation of piezometric levels of aquifers in watersheds govern the possibility of ground water leakage from upper aquifers to lower ones, and in valleys discharge from lower aquifers to upper aquifers and river channels.

Ground water recharge through infiltration of precipitations occurs all over the area of the artesian basin. The recharge areas are not continuous and confined only to the peripheral parts of the artesian basin, they are separate areas confined mainly to interfluve areas and having contours of the latter. The recharge areas alternate with discharge areas in river valleys. In addition, discharge occurs artificially through numerous water outputs.

The presence of clay rocks in the water-bearing strata does not exclude the possibility of ground water recharge down to deep-seated aquifers within watersheds and ground water discharge in river valleys. Dome-like elevations of the piezometric surface, distinctly observed in interfluve areas, testify to the existence of the recharge of deep aquifers through the strata of semipermeable clay sediments. Similarly, the drainage influence of large river valleys affects the character of the piezometric surface of the aquifer in areas of its deep occurrence and when containing thick clay strata in its upper part. However, clay layers with an appreciable thickness hamper ground water recharge and discharge that influences the ground water flow amount. As a rule, the most favourable recharge conditions are in watersheds, within the peripheral part of the aquifer where clay strata in the aquifer's top are relatively thin or even absent. Aquifer recharge conditions deteriorate toward the centre of the basin and with the increase in the total thickness of overlying clay layers.

The process of vertical seepage occurs on the background of ground water movement along the aquifer, i.e. horizontal flow. The main factor governing the direction of ground water movement is the draining effect of river systems. The ground water flow is directed from recharge areas outlined by the watershed to discharge areas, river valleys. In the Moscow artesian basin valleys of major rivers, the Volga and Oka, which are regional areas of the discharge of the artesian basin, produce the strongest draining effect. The direction of ground water movement to these rivers from the west, south-west and south to the north, north-east and east conforms itself to the structural and orographic pattern of the artesian basin. The main direction of water movement to regional drains is disrupted by the draining effect of the local erosional system and, in separate areas, by intensified ground water development. The depth of the draining effect of small and medium rivers (local erosional system) is largely governed by the character of the relief and the depth of the erosional downcutting. On a flat watershed with a young river system (downcutting of 50–75 m) and over vast bogged lowlands even with large rivers cutting down as deeply as 25 m only, the drainage depth does not exceed 100 m. The drainage effect of the rivers extends down to a depth of 150–200 m on the slopes of upland areas with an intensively developed erosional system cutting down to 100—125 m. A drainage depth of over 200 m is typical os the regional drains of the basin, Volga and Oka valleys, which are the centref of discharge of deep ground water that is not drained by the local erosional system.

The principal regularities of the distribution of the thickness of the fresh-water zone are connected with regional hydrodynamic characteristics of the basin. For the Moscow artesian basin it averages 200—250 m, varying from 0—50 to 300—350 m, which depends on the areal distribution of recharge and discharge areas. The largest thickness of the freshwater-bearing strata is confined, as a rule, to most elevated watersheds where deep-seated aquifers are intensively recharged. Relatively poor conditions of water exchange and a thickness of the freshwater-bearing strata of no more than 100 m, are characteristic of the lowland areas.

Thus in an open-type artesian basin a distinct dependence of ground water recharge, movement and discharge on the distribution and characteristics of the orographic elements of the area, moreover the present-day drainage system formed in the course of the long history of the geological development of the region, can be observed.

The ground water flow of the artesian basin is formed within two upper hydrodynamic zones (or stages). The upper zone of intensive ground water flow (intensive water exchange) includes the upper part of the freshwater-storing strata and extends downwards to a depth of 150—200 m (in the Moscow artesian basin). These are relatively shallow unconfined ground waters or waters with local head and interlayer confined waters. They are recharged in interfluve areas all over the area of the artesian basin. The ground water flow is formed under the draining effect of the local erosional system, the regime of the flow is closely connected with local physiogeographical factors.

The other zone of deep (slow) ground water flow includes lower freshwater aquifers and aquifer systems and, in addition, strata bearing slightly salty water with a dissolved solids content of 10 g/l at the most. Deep-seated confined aquifers are mainly recharged within most elevated watersheds of the principal river systems, and ground water flow is directed towards the largest rivers, the regional discharge areas of the artesian basin. Here, the recharge conditions and the formation of deep ground water resources are somewhat hampered as compared with the upper hydrodynamic zone that influences the resources regarding the amount and rates of the deep ground water exchange. The thickness of the deep flow zone in the Moscow artesian basin is 100—150 m, and it extends to a depth of 350—400 m.

The different hydrodynamic characteristics and conditions of the ground water resources govern the selection of methods usable for regional estimation of the upper and lower hydrodynamic stages of the artesian basin.

The method of separating the hydrograph of the total river runoff and distinguishing the subsurface components is applied to estimate natural ground water resources. (Natural ground water resources are indicators of the ground water recharge and show the ground water availability of a region, so they constitute the necessary basis for the solution of problems of the rational ground water use.) The quantitative estimate of the deep flow zone may be obtained by compiling the average long-term water balance of the area and by determining the amount of present-day recharge and discharge of the deep ground waters. Both methods include an obligatory consideration of the hydrogeological conditions of the region and characteristics of the ground water flow (Kudelin, 1960). These methods are most suitable for the application in regional estimation of the natural ground water resources of such a large region as an artesian basin.

The ground water flow of the zone of intensive water exchange is distributed over the area of the basin in keeping with climatic, hydrogeological, geomorphological and other natural conditions (the extent of forests, shrub, etc.).

Climatic factors, mainly precipitation and evapotranspiration, govern ground water recharge. The distribution of ground water flow values is influenced both by latitudinal zonation characteristic of climatic elements and local climatic factors such as larger amount of precipitation in uplands, increased evaporation in areas with extensive swamps and lakes, etc.

Topography plays a major role. A larger amount of precipitation ("orographic precipitation") and deep erosional ruggedness, a dense river system, steep topographic and ground water level slopes are typical, as a rule, of the upland areas. All this contributes to the recharge and discharge of aquifers and intensifies ground water flow. Within vast, often swamped lowland areas the river system is poorly developed, erosional downcutting is insignificant, the drainage influence of rivers extends to a relatively small depth (down to 100 m), and evaporation from swamps is appreciable. The ground water flow is, as a rule, weak.

The effect of hydrogeological conditions of the area is very complicated and diverse. The lithologic composition, fissuring and karstification of water-bearing rocks determining the storage properties of water-bearing strata, their thickness and conditions of occurrence, the composition and thickness of semipermeable confining layers retarding seepage and recharge, the hydrodynamic situation, the relative position of levels of separate aquifers and their interrelationship, an interaction with surface streams and aquifer discharge, etc. – in each individual case all these conditions govern the amount and distribution of the ground water flow.

All the above natural factors act interdependently in the complex process of ground water flow formation supplementing or interfering with each other to such an extent that it is often difficult to choose the main factor. The influence of the whole physiogeographic and geological–hydrogeological situation is reflected on various values of the ground water flow and regularities in its areal distribution.

The following average ground water flow values of the intensive water exchange zone are characteristic of the Moscow artesian basin: the average annual ground water flow discharge is about 2 l/sec/km^2, runoff layer is up to 60–70 mm/year, minimal discharge is 1 l/sec/km^2, and the ground water flow coefficient is 10% of the normal precipitation. The most favourable conditions of ground water flow formation are within the Valdai and Middle Russian uplands. Here, the average annual ground water discharge amounts to 3–3.5 l/sec/km^2 (90–100 m/year), the minimum discharge is over 1.5 l/sec/km^2, the coefficient of ground water flow is up to 15%. The smallest values of ground water flow are observed in the upper Volga and Mologa-Sheksna lowland areas, the average annual ground water discharge here is less than 1 l/sec/km^2, the minimum is up to 0.5 l/sec/km^2, the runoff coefficient is about 5%. The total amount of ground water flow from all aquifers, confined to sediments of different age and drained by the local erosional system, is distributed over main aquifer systems according to their thickness, coefficients of permeability and flow slopes. Carboniferous aquifers have the largest natural resources, Quaternary and Devonian aquifers have appreciable resources.

The deep ground water flow in the Moscow artesian basin has average values of deep infiltration (deep water recharge), on the larger part of the area, amounting to 20—30 mm/year (the average ground water flow discharge is 0.9 l/sec/km²). High values of deep recharge (up to 60—80 mm) are observed in the northern and north-eastern parts of the basin where, as it was noted above, the upper zone of the intensive ground water flow is poorly developed, and the precipitation mainly recharges deep ground waters. The deep water discharge within the Volga and Oka valleys (regional drains of the artesian basin) amounts to 40—60 mm/year and 80 mm/year in separate areas.

Regarding the deep component, the total ground water flow (mainly freshwater flow) within the two upper hydrodynamic zones of the Moscow artesian basin, it is estimated at 3 l/sec/km² (90 mm/year). This amounts to 40% of the total water resources of the region and 15% of the precipitation. The ground water flow of the zone of the intensive water exchange accounts for about 70% of the total value of ground water recharge, more than 30% of infiltrating precipitation forms the deep artesian flow. Comparing the discharges of the upper and deep ground water flow, their relative discharges (related to thickness unit) and the rates of water exchange are 1.5—2 times lower in the rate of deep flow as compared with the flow to rivers.

The estimates of ground water flow values for separate areas of the region, made for refining and checking the regional estimates obtained by other methods (modelling, using low river discharge increments, hydrodynamic calculations of flow) have shown close results (LEBEDEVA, 1972).

The quantitative estimation of ground water flow and the character of its distribution over the area, objectively reflecting the hydrogeological, geostructural and physiogeographical conditions of the region, allow to separate more validly areas of ground water recharge and discharge, and they are well-correlated with the regional picture of the hydrodynamic situation of the artesian basin.

It is also possible to determine more exactly the boundaries of the artesian basin. The boundaries of the platform-type artesian basin should be drawn along the hydraulic divide of deep ground water flow (of the zone of retarded water exchange), position of which is determined by the reciprocal position of recharge and discharge areas, connected in turn with the orographic and geostructural characteristics of the area. Therefore the hydrodynamic boundary of the artesian basin traced along the hydraulic divide of deep flow, as a rule, coincides with the principal positive geological structures and with the orographic divides of main river systems of the area.

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CONTRIBUTIONS TO THE KNOWLEDGE OF STRUCTURES WITH THERMAL WATERS IN THE EASTERN PART OF THE PANNONIAN DEPRESSION (ROMANIA)

CONTRIBUTIONS A LA CONNAISSANCE DES STRUCTURES DES EAUX THERMALES DE LA PARTIE ORIENTALE DE LA DÉPRESSION PANNONIENNE

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RÉSUMÉ

Dans le présent ouvrage les auteurs font l'analyse des facteurs géologiques, hydrogéothermiques et des données géophysiques concernant le secteur roumain de la Dépression Pannonienne, qui s'étend entre les vallées du Someș et du Mureș. En déchiffrant les caractères structuraux des formations prénéogènes, on a pu mettre en évidence des compartiments tectoniques affaissés, plus fréquents dans la moitié nord de la région. L'existence de certains secteurs de subsidence néogène dans l'aire de ces compartiments a rendu possible la formation de structures aquifères. Les éléments structuraux tectoniques sont illustrés dans des cartes et coupes géologiques.

Les données de température enregistrées par de nombreux sondages ont mis en évidence une série de zones anomales géothermales, à amplitudes et fréquences maxima dans la région située au nord du Crișul Repede. Ces zones anomales ont été partiellement étudiées à l'aide des forages hydrogéologiques, ce qui a permis de constater, dans certains secteurs, la présence d'accumulations d'eaux à températures montant jusqu'à 90 °C.

La corrélation des conditions géologiques structurales de la région, du degré de perméabilité des formations pliocènes et des anomalies géothermiques identifiées nous a conduit à établir les possibilités de mettre en évidence, dans ces formations-là, quelques régions de perspective pour les eaux thermales.

On présente aussi, dans cet ouvrage, les particularités géologiques et hydrogéologiques du secteur Oradea-Toboliu, où on connaît des eaux thermales localisées dans les carbonatites mésozoïques.

Introduction

The utilization of new energy sources in the economical circuit constitutes today one of the major concerns of researchers. Within this broad framework an essential part is incumbent on the practical application of thermal water resources whose existence has been pointed out in boreholes drilled over the territory of Romania. For this reason during the two last decades the western part of Romania constituted the object of geological, hydrogeological and geophysical investigations. To the advances achieved as to the geological and hydrogeological knowledge of the Romanian territory, within the Pannonian Depression, there have essentially contributed the works of several specialists. A particularly valuable contribution was yielded by the studies carried out by following authors: PAUCĂ (1954), LITEANU et al. (1965), ISTOCESCU and IONESCU (1969) ICHIM et al. (1971)*, BLEAHU et al. (1971), NECHITI

* Report in the archives of the Ministry of Mines, Petroleum and Geology.

et al. (1971)*, PATRULIUS et al. (1972)*, PARASCHIV and CRISTIAN (1973), PARASCHIV et al. (1975)*, VASILESCU and OPRAN (1975), BANDRABUR et al. (1975)*, etc.

The deciphering of deep-seated geological structure has been performed by seismic, magnetometric, airmagnetometric, geoelectric, gravimetric, more recently geothermic surveys whose results are recorded in the works of GAVÁT et al. (1963), CRISTESCU et al. (1965–1966)*, FOTOPOLOS et al. (1969)*, PROCA et al. (1967)*, ALI-MEHMED et al. (1969)*, SAVU et al. (1970)*, ANDRIESCU and MIHAIL (1971)*, DANANAU et al. (1969–1973)*, RĂDULESCU et al. (1974)*, VELICIU (1974)*, VISARION et al. (1975)*, etc.

This paper is aiming at the elucidation of some problems connected with the structural evolution, hydrogeological and geothermic factors which have favoured the important accumulations of thermal waters in the Mesozoic and Neogene formations.

Lithostratigraphic data

The crystalline basement of the Pannonian Depression is built up of meso- and epimetamorphic schists, which locally comprise essential bodies of eruptive rocks.

The study of magnetic properties of crystalline formations has revealed that they are quite weakly magnetized. Consequently the regional magnetic anomalies plotted on the area of the depression, which are sometimes genetically associated also with gravity anomalies, reflect infracrystalline predominantly basic magmatite bodies. Whereas in the Zădăreni-Arad Zone the mapped anomalies are produced by metabasites from the Păiușeni Series (SAVU, 1962), in other sectors their interpretation is more difficult since a direct extrapolation of data relating to the western border of the Apuseni Mountains is not feasible. In this last category an important part is played by conjugated gravity and magnetic anomalies in the Chișineu Criș Zone, largely developed on surface, and whose apex is located within the Hungarian territory.

The sedimentary formations from the basement of the depression belong to a stratigraphic interval comprised between the Triassic and the Lower Cretaceous inclusively; they have been encountered in boreholes of the Oradea Zone.

The Triassic developed in the Oradea Zone is transgressively resting on the crystalline basement. The sequence comprises a basal conglomerate including crystalline elements being overlain by a quartzitic sandstone level (Seisian). There follow red and green argillaceous schists (Lower Campilian), dolomites and bituminous limestones (Upper Campilian-Anisian), and this sequence is closed by Ladinian massive limestones (PATRULIUS et al. 1972).

The Jurassic extending only in the Oradea Zone, consists of a sequence of formations in calcareous facies which resemble the one cropping out east of the border of the depression, within the Bihor Autochthonous.

The Lower Cretaceous is developed only in the Oradea area being represented by a sequence of limestones with pachyodonts overlain by

* Report in the archives of the Ministry of Mines, Petroleum and Geology.

glauconitic sandstones and limestones that are laterally grading into reddish argillaceous schists. The Lower Cretaceous closes the Mesozoic pre-Alpine sedimentary cycle.

The pre-Neogene sedimentary formations which arose subsequently to the emplacement of the Alpine Codru Nappes belong to the Upper Cretaceous and the Paleogene.

The Upper Cretaceous, the first term of the depression, is characterized by a wide-range lithofacial variation and variable thicknesses (200 – 1300 m) covering more extended surfaces in the region north of the Crişul Repede river. As a whole its structure allows to distinguish a detrital facies with gradings into marly-calcareous or argillaceous facies.

The Paleogene showing variable thicknesses (600 – 1500 m) is developed only in the Maramureş Trough where it is represented by a marginal detrital facies (conglomerates and massive limestones) and by a flysch facies with rhytmical sequences of sandstones and marls.

The Neogene sedimentary formations, mostly widespread within the depression, pertain to the Tortonian, the Sarmatian and the Pannonian.

The Tortonian, whose thickness amounts to 650 m, consists in its lower part of a conglomeratic complex with intercalations of tuffs, and in its upper part marls with intercalation of marl-limestones and sands do predominate.

The Sarmatian is represented by a monotonous series of sandstones with intercalations of marls and marl-limestones.

The Tortonian and Sarmatian formations are not uniformly spread within the depression as they are lacking on some elevated compartments of the pre-Neogene.

The Pannonian is widely developed and also shows an appreciable thickness exceeding 2500 m in the depression zones. The intricate paleogeographical picture of the sedimentary Pannonian Basin, characterized by a relief displaying a marked uneveness and branches, is reflected both in the facies of the Pannonian and the spread of its various terms.

On the regional scale, as a function of the lithological and paleontological constitution of the Pannonian, two main sequences have been distinguished:

– The Lower Pannonian consisting prevailingly of marls and argillaceous marls, and displaying a transgressive character;

– The Upper Pannonian which represents the largest part of the Neogene filling in the depression, and is built up of a thick sand pile (maximum 1800 m) with intercalations of clays, sandy clays, coals and grading at its upper part. Corresponding to some clearcut facial changes, the boundary among these terms appears distinctly on electrical well-logs and sometimes may be also evidenced on seismic profiles.

Major structural elements

The main informational data on the deep-seated limits of the crust have been yielded by seismic surveys carried out on the last segment over the Romanian territory of the 11th international profile (Radulescu et al. 1974). The characteristic feature of the Mohorovičič discontinuity over this profile consists in its regional uplift; the latter is located in the vicinity of the frontier with Hungary at a depth of approximately 27 km in accordance with the results obtained by MITUCH, POSGAY (1972).

The existence of a zone showing a marked thinning of the crust on the scale of the entire structural unit is proved by a gravity anomaly, which points out a major rise of the values exceeding 40 mgal from east to west, corresponding to a non-uniform Mohorovicič discontinuity with amplitudes of some 6—7 km.

The structural features of the Pannonian Depression have been chiefly disentangled by the complex interpretation of geophysical and geological borehole data. The seismic investigations carried out by numerous specialists working in the petroleum and gas industry (DANANAU et al. 1969—1973; ICHIM et al. 1971, etc.) played an essential part in these researches.

On the time sections of seismic profiles there may be separated two regimes of reflected events with clearly differentiated characteristics.

The upper part of sections corresponding to Pannonian formations is characterized by the presence of numerous distinct reflectors with relatively slight dips and rare interferences. Some reflectors may be traced over appreciable distances of the order of ten km, without, however, wearing the aspect of characteristic reflectors, fact that indicates the absence of some significant velocity contrasts within the Pannonian.

In the lower part of the section, within the pre-Pannonian formations, the number of reflectors decreases, and hence tracing becomes more difficult because of frequent interferences with multiple, diffracted or other disturbing waves.

In areas with an uplifted basement towards the lower part of the Pannonian a characteristic reflection may be followed; it is produced through the strong contrast of acoustic properties occurring between the crystalline formations and the overlying sedimentary rocks. This reflection may be also stated in zones wherein the Neogene formations are overlying the Mesozoic and Paleogene ones, however, its characteristic aspect is attenuated. Taking into account the migration of the reference reflector on various geological formations a structural map of the basement of the Neogene was plotted (Fig. 1) in accordance with geological borehole data. The tracing of major ruptural accidents requires the consideration also of other geological data, particularly the gravity ones (VISARION et al. 1975).

The structural map reveals the existence of a "tectonique cassant", consisting of major fractures whose movements on vertical extent involve both the crystalline basement and the overlying sedimentary formations. These fractures led to the separation of some uplifted and sinked compartments of varied size, with different functions in the sedimentary process which they have essentially influenced. The evidenced fractures show two preferential trends: NW—SE and NE—SW. The fractures of the first category, characterizing the middle zone of the depression, continue also east of the border of the depression delimiting or affecting the central sectors of the Neogene basins around the Apuseni Mountains. In the northern area of the depression prevail fractures oriented NE—SW, characteristic of the Pannonian Domain, which have arisen or resumed in the Pannonian.

The pre-Neogene relief of the depression presents an intricate configuration, which determined quite varied thicknesses of the overlying formations from some metres in the uplifted compartments in the vicinity of the border, and up to 4000 m in some sinked compartments. The delimitation of the main areas of the thickened Neogene formations has a particular significance as it permits the elucidation of at least the structural factors which control the thermal aquiferous complexes in the Pannonian.

The eastern border of the Pannonian Depression which marks the limits among the depression proper, the mountain frame and the Neogene intramountain basins, is considered to be of a tectonic nature, separating two domains with a different geological evolution in the time interval between the Miocene and the Quaternary.

The chief structural elements of the Fig. 1, significant for the Neogene tectonics, have been rendered in the Fig. 2. In addition the authors have represented the pre-Neogene formations which are overlain by recent sediments.

One may notice how the spread area of the pre-Neogene sedimentary formations corresponds to the sinked zones of the crystalline basement, and the more uplifted zones of the crystalline basement are directly overlain by Pannonian sediments.

The main uplifted compartments on which the crystalline formations come into direct contact with the Pannonian are located in the vicinity of the eastern border of the depression, and namely in the extension of the Bîc Mts *(A)*, in Tinca-Homorog sectors *(B)*, Tăuteu *(C)*, Satu Nou-Cermei *(D)*, Chișineu Criș-Chereluș *(E)*, as well as in the Arad-Turnu sector *(F)* representing the continuation towards the west of the Highiș Mts. A similar situation is encountered in the Mihai Bravu uplifted zone *(G)* whose apex is located on the territory of Hungary.

The Neogene cover is underlain in other four sectors by the Paleogene or Mesozoic formations whose chief features will be described hereunder.

The Maramureș Trough *(H)* with a filling composed of Paleogene formation in the flysch facies may be traced between the northern sector of the Romanian frontier and the Hungarian one and the regional, eventually bearing a crustal character, fracture known under the name of the Dragoș Vodă Fault. The Paleogene formations with an important thickness are intensely folded, and rest on the crystalline basement divided into compartments with different spatial positions.

The Cadea-Andrid graben *(I)* is bordered by a system of fractures trending NE–SW which separate it from the adjacent uplifted zones.

Some aspects of interest are presented by the Oradea compartment *(J)* delimited by important peripheral fractures where a thick pile of Triassic, Jurassic and Lower Cretaceous formations, overlying the generally sinked crystalline basement, is preserved.

The borehole data evidenced that these formations belong to the westward extension of the Bihor Autochthonous in the substratum of the depression. In relation to adjacent zones, where the crystalline basement directly underlies the Neogene formations, the uplifted relief of the Mesozoic in the Oradea compartment is noticed. The Socodor-Grăniceri Depression Zone *(K)* with a large development in the Hungarian territory (8000 m) (Nemesi, 1973), does probably also comprise Mesozoic formations beneath the Neogene cover.

These data prove that besides the uplifts of the crystalline basement the high reliefs may be also constituted of Mesozoic formations (Oradea compartment) or even of Paleogene formations.

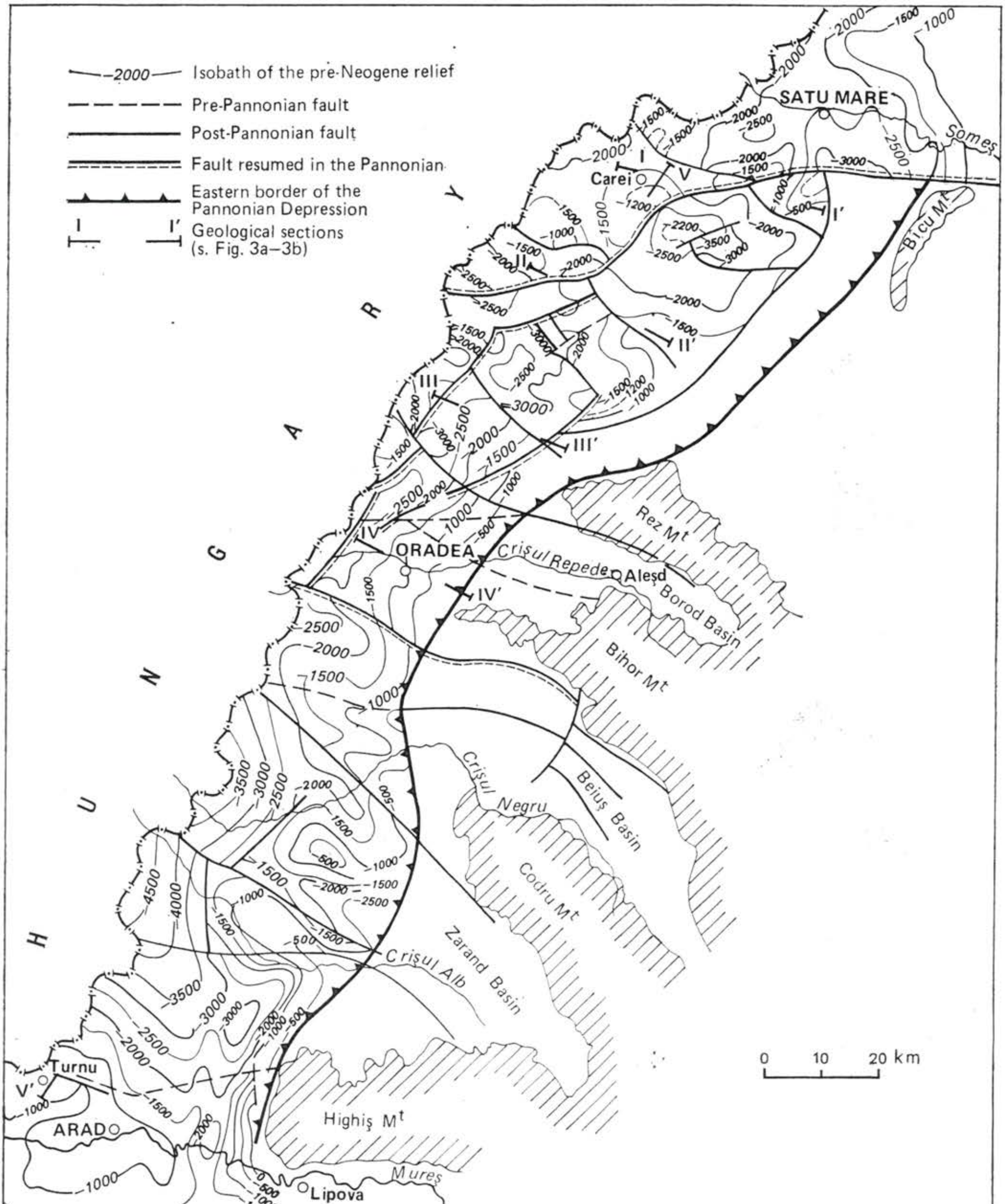

Fig. 1. Structural map of the pre-Neogene relief

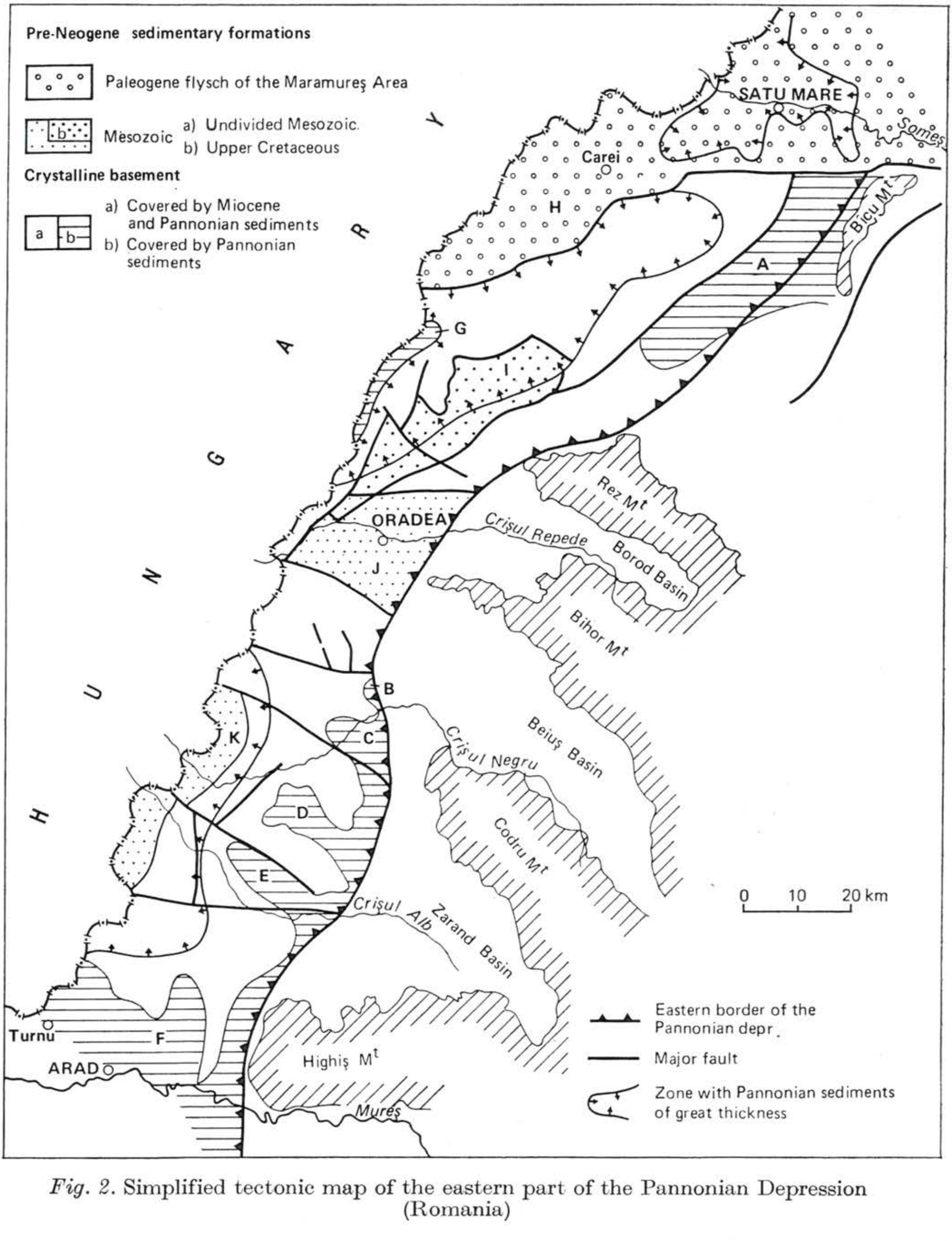

Fig. 2. Simplified tectonic map of the eastern part of the Pannonian Depression (Romania)

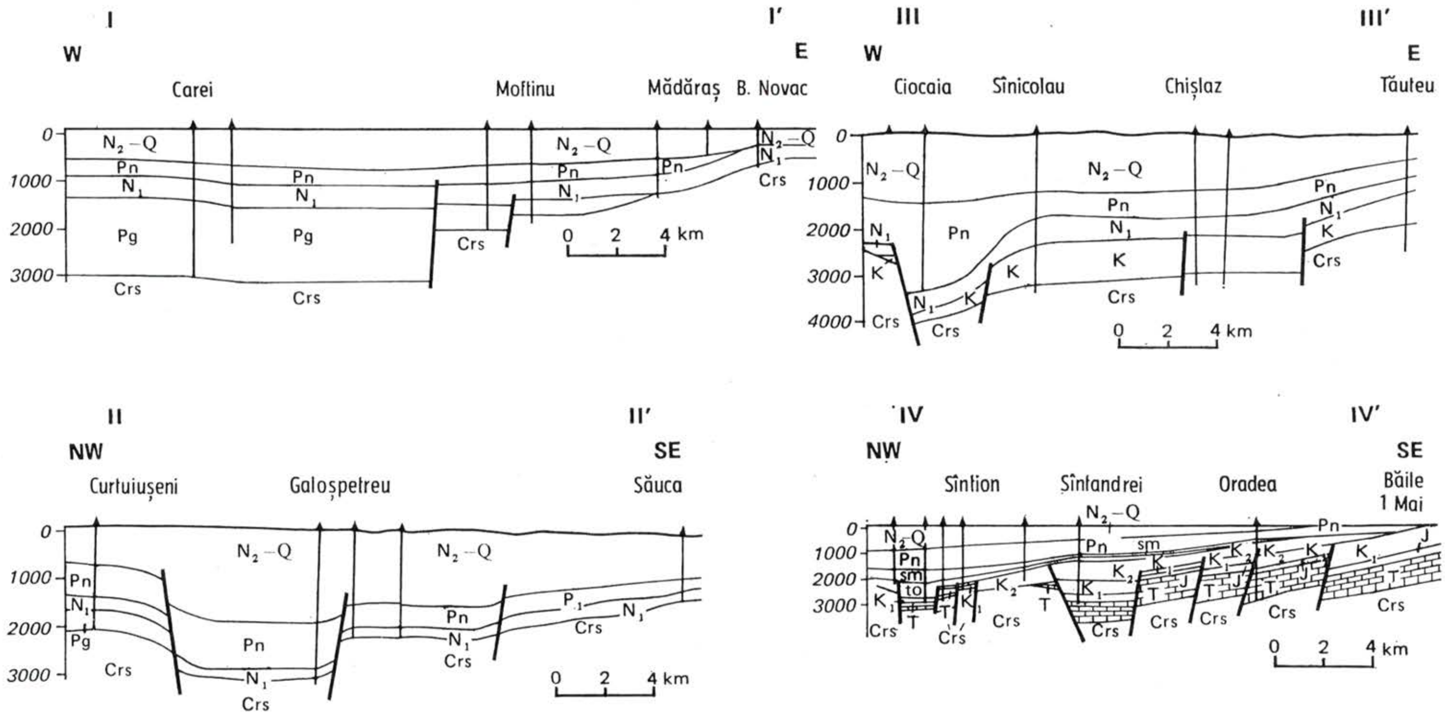

Fig. 3a Geological cross sections through the eastern part of the Pannonian Depression

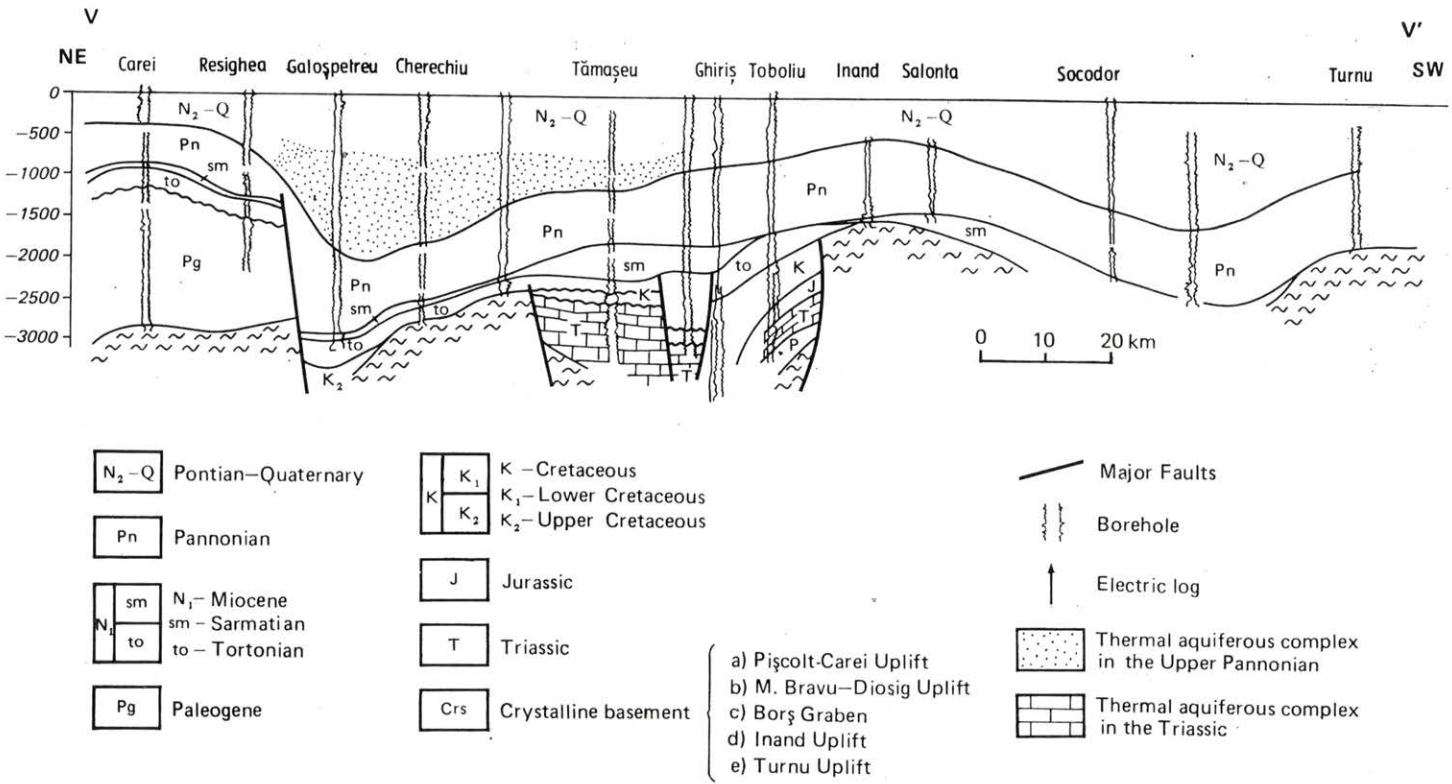

Fig. 3b Geological longitudinal sections through the eastern part of the Pannonian Depression

Within this structural edifice there have been distinguished the Satu Mare, Galoşpetreu-Mecenţiu and Socodor-Grăniceri Sectors characterized by the active subsidence in the Pannonian (Fig. 2).

The structure of the Pannonian formation is prevailingly determined by the relief of the pre-Neogene formations. Whereas in the most part of the depression the former shows a quasi-horizontal settlement, on uplifted compartments the seismic sections point to a more frequent existence of some structures of the hemi-anticlinal type in the vicinity of the border, and of the brachyanticlinal one within the depression.

A series of representative geological sections plotted through the interpretation of seismic and borehole data, illustrates the characteristic features of the depression and its substratum (Fig. 3a—b).

Geothermic particularities

The presence of thermal waters in the western part of Romania was known for a long time due to the fact that in some sectors they could reach the surface in the form of hot springs (Băile Felix — 1 Mai, etc.).

The wells which were drilled for hydrocarbons and the special ones, however, less numerous, that aimed at identifying deepseated aquiferous layers, have cleared some specific features of the geothermic regime in the eastern part of the Pannonian Depression (Paraschiv, Cristian, 1973; Paraschiv et al. 1975).

The investigations carried out in recent years have yielded new elements which allowed us to reach a more detailed knowledge of the geothermic particularities of the depression. They traced the systematization of temperature data measured at various stratigraphic levels in wells drilled for hydrocarbons, and directly of waters in hydrogeological wells. The areal distribution of results is relatively non-uniform, as the chief informational data proceeded from wells drilled for hydrocarbons, to be found over restricted prospecting zones, and which are especially referring to Pannonian formations.

The plotting of a synthesis map with geoisotherms has implied the introduction of some corrections which might render the measurements comparable.

A first correcting factor is referring to time intervals at which the measurements of water temperatures in hydrogeological wells have been carried out. As it results from the following table, partially compiled by Nechiti et al. (1971), the temperatures measured subsequently to a time interval exceeding 15 days since artesian flowing started, are higher in comparison with the initial ones, the differences amounting to 20—25 °C. This is due to the fact that the thermal equilibrium is not reached within the drilled zone.

The second essential factor is referring to the stabilization conditions of the thermic regime around the boreholes. The boring entails two disturbing factors which affect the steady state of thermic regime of the borehole: the drilling process during which an amount of heat is released, as well as the circulation of the drilling fluid which modifies the initial temperature of rocks. The studies in this field proved that in order to establish the temperature of rocks in situ it is necessary to carry out the measurements subsequently to about a 3 month period after the end of the drilling. The differences between the stabilized and non-stabilized temperatures range from 5° to 25 °C depending

Table 1

Temperature measurement well	Depth of the aquifer m	Temperature (°C) measurements		
		Test period	>15 day period from end of the test period	$\Delta T°$
Carei		42	49	+7
Satu Mare		65	72	+7
Tămășeu	1043–1168	55	60	+5
Marghita	985–1376	59	64	+5
Oradea	1100–2813	60	80	+20
Oradea	2052–2700	62	87	+25

upon the geological characteristics of the zone wherein the wells were drilled and also upon the depth at which the measurements have been performed (Paraschiv, Cristian, 1973; Demetrescu, 1975).

The application of the correcting factors allowed the construction of a map with geoisotherms, plotted at the isobath level of –1000 m. This corresponds, in the most part of the depression, to the middle depth of the aquiferous complex in the Upper Pannonian, the main reservoir recognized both over the territories of Romania and Hungary (Fig. 4). Although the obtained picture is partially approximate owing to the non-uniform distribution of boreholes on the surface, nevertheless this map localizes a series of positive anomalies with high values.

In the northern part of the depression, in the Săuca–Mecențiu–Moftinu sector an anomaly with values exceeding 80 °C is outlining. Westwards this anomaly is extending to Carei where the temperatures amount to 90–95 °C. Structurally the anomaly is predominantly localized on a sinked compartment wherein the sedimentary pile shows a thickness of some 3100 m. Southwards high temperature values are noticed over the area of the Pișcolț–Curtuiușeni structure, and this tendency is more accentuated in the vicinity of the frontier with Hungary.

A less extended anomaly with temperature values exceeding 80 °C covers the Abrămuț sector. High temperature values (75–85 °C) are also stated westwards being indicative of a local anomaly in the Săcueni Zone. Although the geoisotherms are also indicating a decrease of temperatures southwards, one may still trace a less ample anomaly in the Chișlaz-Sălard-Tămășeu sector.

In the northern part of the Biharea horst a local anomaly outlined by geoisotherms of 80 °C was recognized.

In the region comprised between the Crișul Repede and Mureș rivers the information on the temperature distribution in depth is limited. The available data offer a regional image of temperature variations at the depth of the selected reference plane, characterized by relatively high values exceeding 65 °C west of the Miersig–Tăutu–Chișineu Criș alignment.

In the southern part of the region–Turnu Structure Zone–an anomaly is revealed by isotherms of 75 °C and 80 °C. It represents an eastward extension of the major Szentes–Szeged anomaly, detected by temperature measurements in numerous boreholes. In this zone the above anomaly was checked by

Fig. 4. Distribution of temperatures in the eastern part of the Pannonian Depression

determinations of the geothermic gradient in steady state regime, being performed in boreholes at a depth of 20 m and respectively 30 m (VELICIU, 1974).

The geothermic particularities of the Pannonian Depression on the regional scale are assigned by most researchers to the essential thinning of the Earth's crust (BOLDIZSÁR, 1966; CERMÁK et al. 1968; PARASCHIV, CRISTIAN, 1973, etc.). Locally the above described geothermic anomalies are connected with a series of structural and hydrogeological factors. In the whole it is estimated that the distribution of anomalies and likewise their trend are chiefly controlled by structural factors, especially by fractures which have determined the differentiated movements on vertical extent of diverse previously described compartments.

Hydrogeological considerations

The region under study is of a particular interest owing to the presence of some thermal aquiferous complexes within the Triassic, Cretaceous and Upper Pannonian formations. Accumulations of thermal waters are also encompassed in other sedimentary formations, and even in the crystalline basement, however, because their high mineral content and the low discharges they do not present any economical interest.

Triassic thermal aquiferous complex. The Triassic formations with a limited extent have a most complex development in the Oradea Zone where their thickness amounts to some 1000 m. The lithostratigraphic sequence, correlable with the outcropping encountered in the Bihor Autochthonous includes thick series of limestones and dolomites. In fissured and karstified dolomites and limestones (Upper Campilian-Anisian) the Oradea and Toboliu boreholes tapped a thermal aquiferous complex which has yielded artesian discharges up to 800 m^3/day with temperatures ranging from 87 °C to 90 °C. The hydrostatic levels have values comprised between +60 m and +150 m, and the dynamic ones of +2 m (NECHITI et al., 1971). The aquiferous complex extends to north of Toboliu in the Borș sector where in a series of boreholes thermal waters have been pointed out. However, they are of a minor interest because of their mineralization degree which amounts to 14 g/l.

The temperature measurements of rocks allowed us to work out of a hydrothermic sketch, at a level located at the top of the Triassic dolomites (Fig. 5). It is not out of question that beneath this level the water temperatures should be higher.

Lower Cretaceous aquiferous complex. The presence of thermal waters in Lower Cretaceous limestones was for a long time known at Băile Felix and 1 Mai resorts, located in the vicinity of the eastern border of the depression.

In the Băile Felix boreholes, the Lower Cretaceous limestones have been tapped within the interval of 100–1000 m. In the upper part of the limestone complex (Băile Felix borehole in the depth interval of 110–280 m), the artesian discharge capacity amounts to 17,000 m^3/day with temperatures about 49.5 °C. At lower levels (1100 m) the artesian discharge is of 260 m^3/day, with temperatures of some 34 °C.

In upper horizons the reduction of the discharge capacity has been assigned to the differentiated degree of fissured limestones, more pronounced in the upper part of the geological section and more reduced with depth. The diminu-

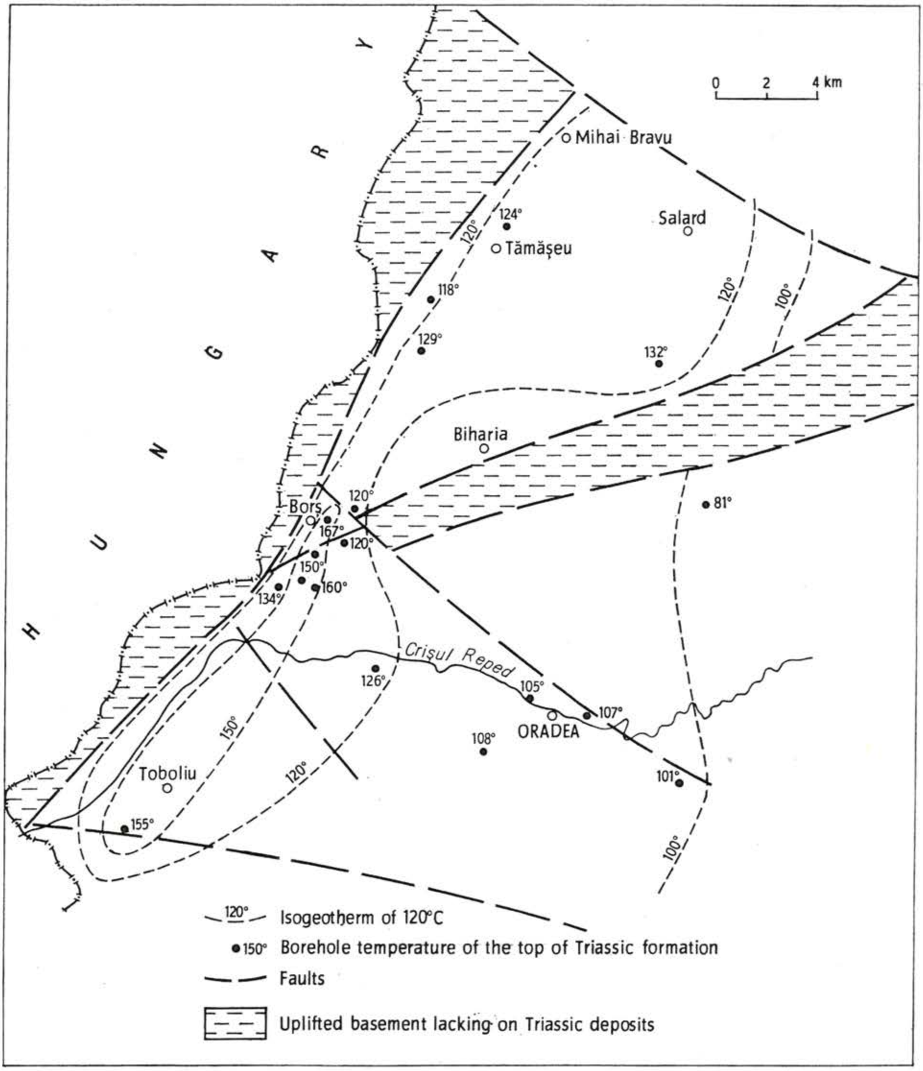

Fig. 5. Hydrogeothermal zonation of the Triassic aquiferous complex

tion of the water temperatures, from surface to depth contravening the geothermic principle, would have been determined by an essential supply of hyperthermal waters from depth, transported to the surface through some major fractures. The large fissuration of the upper part of limestones facilitates an intense discharge of hot waters whereas towards the lower part of weakly fissured carbonate sediments the infiltration of waters is reduced (Nechiti et al. 1971).

Upper Cretaceous aquiferous complex. The Upper Cretaceous formations, characterized by frequent lithofacial variations, are storing only minor accumulation of thermal waters. In some boreholes from the Sîniob, Abrămuţ, Borş, Toboliu and Sînandrei sectors, the fissured limestones contain thermal waters with an ascensional level and surficial temperatures between 37 °C and 60 °C. The artesian discharges are generally less than 0.5 l/s.

Upper Pannonian thermal aquiferous complex. In accordance with results yielded by wells drilled over the territory of Romania and the territory of Hungary, this complex encompasses the main reservoirs of thermal waters in the Pannonian Depression.

Electrical well-logs have evidenced a sequence consisting of sandy strata, which alternate with relatively thin clay and sandy clay horizons. The thickness of this lithologic complex is variable ranging from some ten metres in sectors nearby the border to over 1500 m in some subsidence zones.

The permeable intervals vary from 2 m up to 28 m as a function of the emplacement of boreholes. Their thickness is most frequently comprised between 2 and 8 m. The total thickness of permeable horizons is likewise quite varied (20 – 300 m) depending on the emplacement of boreholes.

The determinations of permeability carried out on sands indicated values from 80 to 100 m D, and the average porosity amounts to 35% (ICHIM et al. 1971).

The Upper Pannonian permeable horizons encompass important accumulations of thermal waters whose temperature varies between 50 °C and 96 °C. According to our current knowledge the waters with highest temperatures (up to 96 °C) have been encountered in boreholes north of the Crișul Repede river, on the eastern slope of the Diosig-Curtuiușeni uplift.

The thermal aquiferous complex in the Pannonian shows mostly an artesian flowing presenting values of the piezometric level between +1 m and +60 m, and of the dynamic level between −21 m and +2 m.

The discharge capacity varies as a function of the grainsize of permeable horizons. In sectors with coarse grain-sized sands discharges up to 3000 m^3/day (Săcueni) have been obtained at a drawdown of 58 m, while in zones with fine sands discharges occasionally decreased under 1000 m^3/day.

The water supply of the aquiferous complex is obtained from atmospherical precipitations in the vicinity of the contact between the depression and the mountain border where the Upper Pannonian crops out.

The correlation of structural and geothermic data with specific features of sandy horizons from the Upper Pannonian allows a zonation of prospecting areas for thermal waters. On the regional scale, three main prospecting regions (Fig. 2), which represent concomitantly subsidence zones in the Pannonian, may be distinguished.

In the northern part of the depression, in the Satu Mare region, a zone of thermal waters was stated; at the level of the Upper Pannonian their temperatures could amount to 80 °C.

According to our current information the most important prospecting zone is located in the Cadea-Galoșpetreu – Mecențiu Graben, where the permeable complex at moderate depths has an appreciable thickness. Temperatures measured in boreholes at the lower part of the aquiferous complex show values exceeding 140 °C.

In the region south of Crișul Repede temperature measurements are rare still indicating in the Socodor Zone values of 100 °C and in the Tăutu – Salonta sector values higher that 60 °C. The existence of some favourable factors, the considerable thickness of the Pannonian as well as the significant results obtained over the territory of Hungary have led to the delimitation of the third prospecting zone in the region under study (Fig. 2).

In reference to the chemical composition of thermal waters (Bandrabur and Craciun, 1975), it is specified that in the Mesozoic sediments, due to the active circulation and the obvious hydrogeological connection with the recharge zone, the thermal waters show a reduced mineral content (0.5 – 1.3 g/l); they are to be assigned to the hydrogeochemical facies as follows: the Ca-HCO_3 type (Cretaceous waters) and the Ca-SO_4 one (Triassic waters). The waters from the Triassic of the Borș area are to be excepted due to intricate tectonics, which has substantially reduced the hydrodynamic connections with the eastern recharge zone; their mineral content is higher (up to 15 g/l), and they are referred to the Na-Cl type.

The waters accumulated in the Upper Cretaceous, Miocene and Lower Pannonian formations are intensely mineralized (5 – 32 g/l) as a result of slow dynamics and an advanced degree of the hydrogeological closing. They are chiefly of the Na-Cl type, and only subordinately of the Cl-HCO_3-Na or Ca type.

The waters from the Pannonian which circulate through permeable formations and benefit by an active recharge are characterized by a reduced mineral content (0.5 – 3.0 g/l), and belong to the Na-HCO_3 type.

Over zones with oil structures in the Pannonian, Miocene and Upper Cretaceous, the thermal waters may be referred to the Cl-Mg and Cl-Ca types.

Conclusions

The integrate study of geophysical data and those obtained from geological boreholes allowed us to establish significant criteria, which ensure the delimitation of prospecting zones for thermal waters, along the eastern border of the Pannonian Depression.

The corroboration of results obtained through surface geophysical exploration and borehole investigations with informational data, yielded by well drilled in the depression area, led to the localization of the main subsidence zones in the Pannonian, whose upper sandy section proves to encompass the chief thermal accumulations.

The practical application on the scale of the whole unit of results obtained from temperature measurements in boreholes permitted the construction of a map with geoisotherms at the isobath level of – 1000 m. On this map there are figured anomalies with intensities of 70 – 80 °C in the region north of Oradea, and of 60 – 70 °C in the southern sector, where the bulk of data is even more reduced.

The main recognized aquifers are included in the Triassic carbonate formations within the Oradea – Toboliu – Sălard Zone and the Pannonian sandy sediments in the Satu Mare, Cadea, Galoșpetreu – Mecențiu and Socodor – Vîrșand sectors.

The data on thermal waters yielded by hydrogeological boreholes are variable, indicating discharges of 80 – 17,000 m³/day with temperatures between 50 °C and 90 °C from Mesozoic carbonates, and of 1000 – 3000 m³/day with temperatures ranging from 60 °C to 96 °C in the Pannonian sands.

The complex research methodology applied in our work has facilitated the elucidation of hydrogeological and hydrochemical factors which condition the accumulation of thermal waters in reservoirs of economical interest.

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LE GRAND BASSIN HYDROGÉOLOGIQUE DU MARANHÃO – BRÉSIL PERSPECTIVES SUR L'EXPLOITATION

THE GREAT HYDROGEOLOGICAL BASIN OF MARANHÃO – BRAZIL PERSPECTIVES OF EXPLOITATION

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ABSTRACT

In Brazil 20 basins or groups of sedimentary basins can be identified. The Maranhão Basin could be considered as 3rd on this list, with a surface of approx. 700,000 km^2. It is a circular basin with a maximum of 3000 m of depth filled with sediments of more or less continuous sedimentation of the Devono-Silurian until the Cretaceous, where clastic and little consolidated sediments are predominant.

The succession of formations in order of permeables and little permeables with a geometry of "bottom of plate", assure the conditions of artesianism in almost 1/3 of the ground waters of the basin.

Actually, the hydrogeological maps with the scale of 1/500,000 cover practically all of the Maranhão Basin where the sought aquifers can be found. Also there are indications for the calculation of well depth, the possible discharge and the probable composition of the waters.

The regulator reserves are evaluated in $3 \cdot 10^9$ m^3/year. The permanent reserves are $3 \cdot 10^{12}$ m^3 and the explorable resources accessible with the wells of 1000 m of depth during a period of 50 years, are evaluated in $1.4 \cdot 10^{12}$ m^3.

In respect of the chemical quality, the vertical zonality shows that the waters situated below a depth of 1500 m, are very rich in salts.

Until a very recent period the exploration of the ground water in the Maranhão Basin (almost 2000 wells) was realized with insufficient technical and economical methods. The most recent wells furnish a spontaneous discharge of 500 m^3/h in average; and the maximum obtained was around a thousand of m^3/hour.

The only solution can be, a combined study of water with double aspects: technical and economical.

The results of the models of management have to furnish objective bases for the choice of the best management. The realization of a financial study will permit the definition of the economical limit from which production expenses of the water will be superior to the benefices which they can produce.

Introduction

Le Grand Bassin Hydrogéologique du Maranhão, couvrant une superficie d'environ 700,000 km^2, se classe 3^e dans la liste des vingt bassins ou groupes de bassins sédimentaires du Brésil qui occupent environ 3,165,000 km^2, soit, 37 % de la superficie du Pays (Fig. 1).

Ces bassins sédimentaires sont géologiquement bien connus, principalement grâce aux travaux détaillés de Petrobrás, organisme chargé de la prospéction et de l'exploitation des hydrocarbures au Brésil.

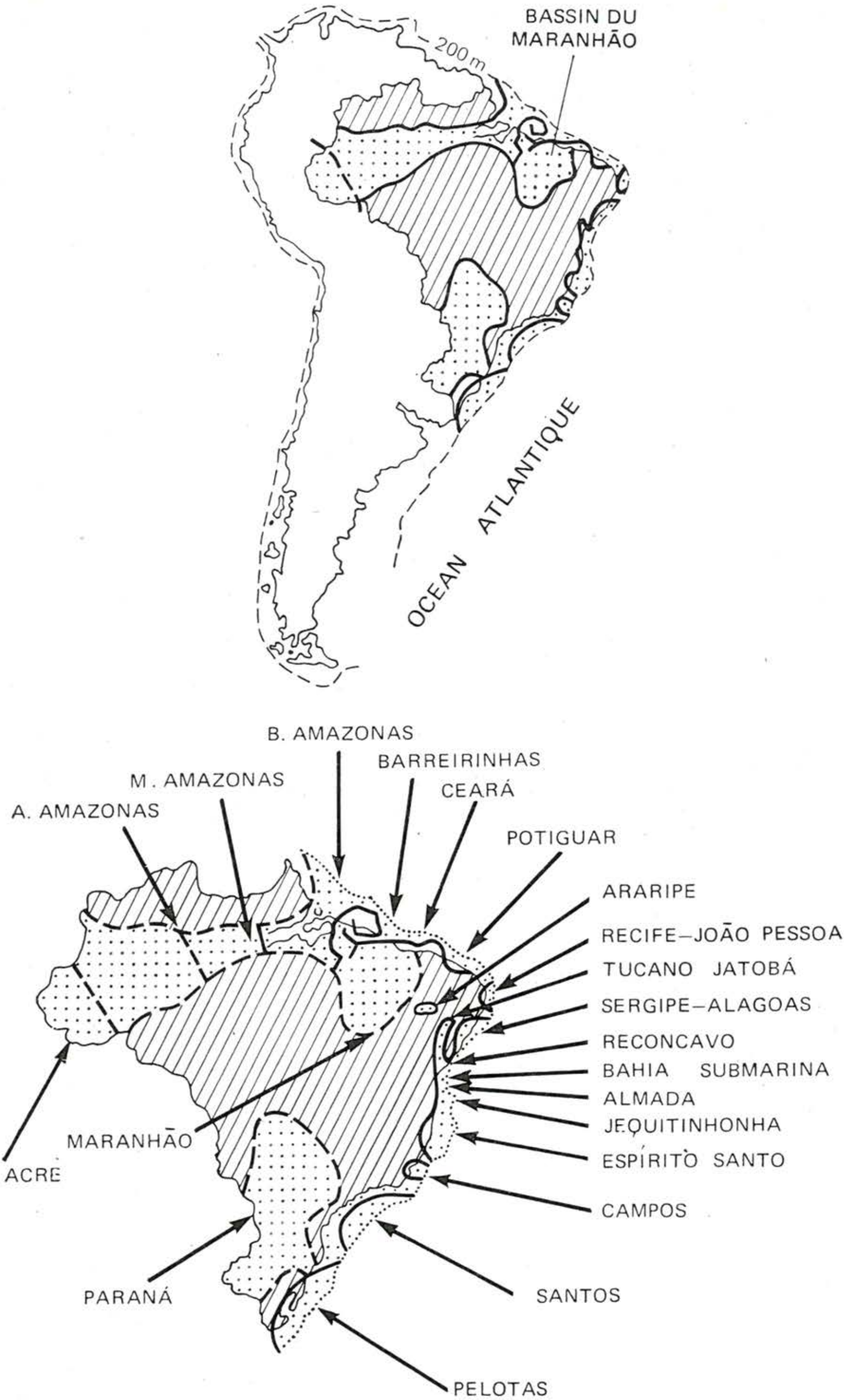

Fig. 1. Les bassins sédimentaires Brésiliens

Leur structure permet de les grouper en deux catégories:

— Les synéclises (Amazone, Maranhão et Paraná) formant un ensemble de 2,700,000 km^2, peu faillé et à sédimentation presque continue depuis le Silurien.

— Les bassins d'effondrement (graben), beaucoup plus réduits, sont limités par des failles en « touche de piano » aux rejets souvent considérables, et comblés par une sédimentation fort discontinue, et d'âge essentiellement crétacé. Ils sont de deux types: intérieurs, en bassin fermés dans le substratum métamorphique, et côtiers, en bande étroit (30 km) parallèle à la côte, au dessin de laquelle ils sont tectoniquement liés.

Du point de vue des eaux souterraines, les conditions géologiques se traduisent de la manière suivante:

a) L'alternance de formations perméables et peu perméables dans des structures généralement en cuvettes (ou monoclinales) assure à tous les bassins des conditions de mise en charge et souvent d'artésianisme.

b) Dans les bassins intérieurs d'effondrement, les formations plus anciennes, rejetées en profondeur par le jeu de failles, affleurent très peu ou pas de tout, et sont, de ce fait, très peu réalimentées par l'infiltration pluviale.

c) Les bassins côtiers présentent tous des problèmes d'invasion marine; souvent même, du fait de la petite extension de leurs parties émergées, de leurs grandes approfondissements sous la mer et de la faible altitude des zones d'alimentation.

Les réserves exploitables jusqu'à une profondeur de 1000 m, ayant 1/3 productif, soit $80 \cdot 10^{12}$ m^3, couvrent largement la consommation domestique espérée sur une période de 50 ans et pourraient, en plus, servir localement à une irrigation d'appoint, dans des cas à définir, en fonction des conditions naturelles et des besoins à desservir (Reboucas 1975).

Les ressources en eaux souterraines des bassins sédimentaires du Brésil sont souvent celles dont la mise en oeuvre est la moins onéreuse et la plus immédiatement rentable. Cependant, bien qu'on se soit préoccupé depuis le début du siècle dernier (le premier forage a été exécuté en 1831), l'exploitation a été, jusqu'à présent, menée avec des moyens techniques insuffisants ou inadaptés; captages mal conçus, diminuant immanquablement les débits. En conséquence, les bassins sédimentaires, zones à réserves importantes, ont été en général négligés et sont tous sous-exploités.

Cadre physique et climatique (Fig. 2)

Naturellement les cycles d'érosion n'ont pas sculpté une masse homogène tout au contraire. En conséquence, on observe des formes très variées telles:

— plate-formes subhorizontales de grès terminées en falaises d'éboulis pardessus des collines des formations sédimentaires inférieures;

— succession de « cuestas » dans les domaines périphériques, dues aux effets d'une érosion différentielle.

Les bassins hydrographiques des rios Parnaíba et Itapecuru-Mearim, principaux éléments de drainage, offrent l'aspect d'une cuvette inclinée vers l'océan, où toutes les surfaces d'érosion convergent et s'abaissent graduellement jusqu'au niveau de la mer.

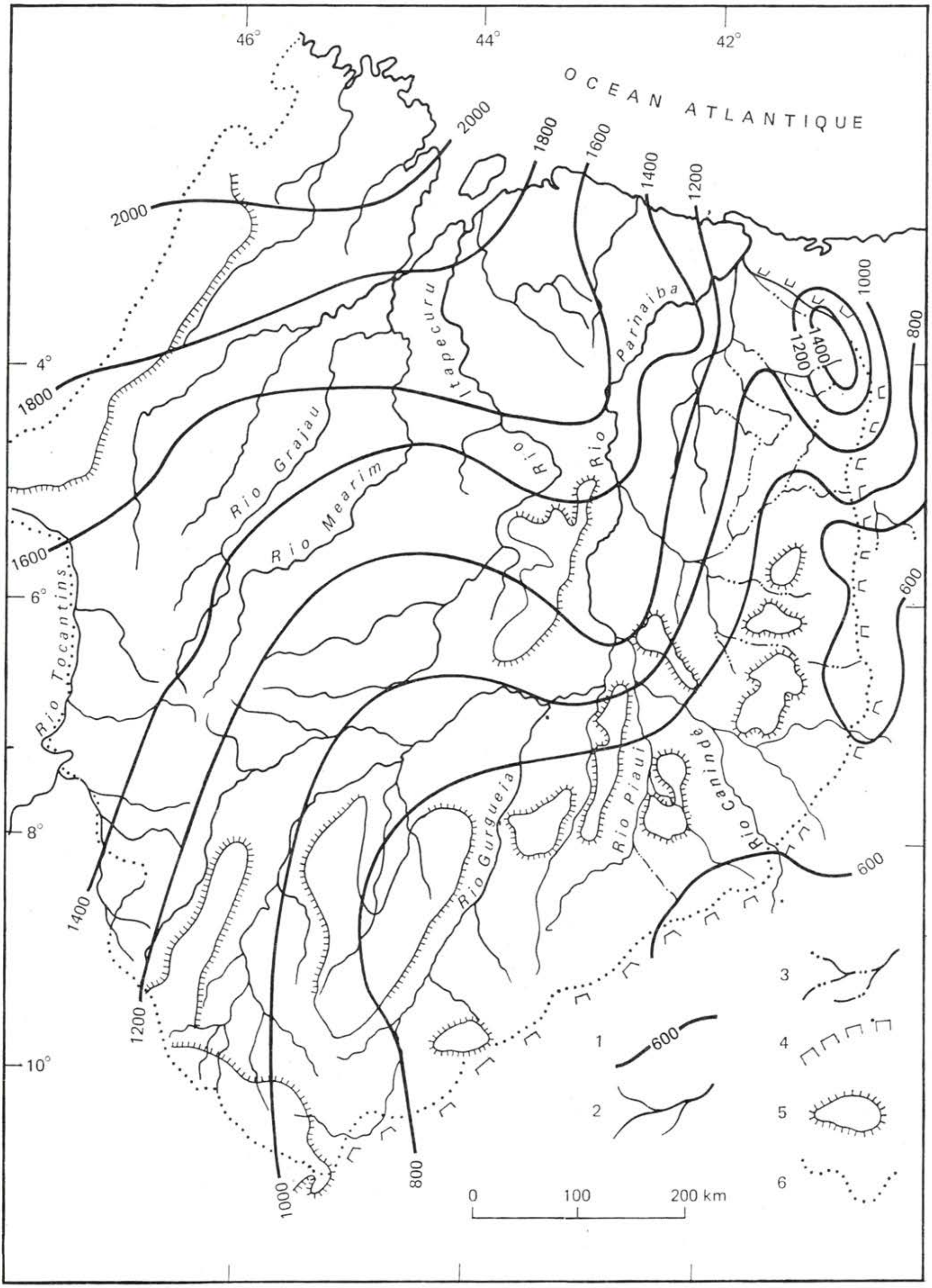

Fig. 2. Bassin du Maranhão. Cadre physique et climatique

1. Isohyète annuelle (mm), *2.* cours d'eau permanent, *3.* cours d'eau temporaire, *4.* cuesta, *5.* plateau, *6.* limite géologique du bassin

L'influence de la variété de faciès morphologiques modelés par l'érosion, sur les conditions hydrogélogiques existe, mais il paraît impossible de l'apprécier à l'échelle d'une analyse générale.

Les éléments orographiques majeurs (700—800 m) sont représentés par les plateaux de grès.

Le Bassin Géologique du Maranhão s'inscrit entre les parallèles Sud 3° et 10° et entre les méridiens 41° et 47° Ouest.

Alors que, par sa position géographique, le domaine devrait jouir d'un climat de type équatorial, le mouvement des masses d'air lui impose un régime pluviométrique très particulier, sujet à des écarts élévés d'une année sur l'autre, conduisant ainsi le climat à revêtir un aspect tropical voir semi-aride, dans son bord Est.

La pluviométrie moyenne annuelle se situe entre 2000 et 1000 mm dans la moitié Ouest et entre 1000 et 600 mm/an dans son bord Est. Ce comportement est engendré par le déplacement des masses d'air équatoriales, originaires de l'Amazonie.

La saison de pluie aborde le domaine par le Sud-Ouest, vers octobre et se déplace en direction de la côte septentrionale qu'elle atteint cinq mois plus tard. Le régime pluviométrique est du type tropical à une seule saison de pluies concentrée sur 3 à 4 mois avec un maximum unique.

Le régime des vents dominants découle du mouvement des masses d'air équatoriales.

La température de l'air à l'échelle mensuelle, oscille entre un minimum moyen de 24 °C et un maximum moyen de 29.7 °C.

L'humidité relative moyenne journalière oscille au rythme des saisons sèche et pluvieuse et très régulièrement entre un minimum inférieur à 50 % et un maximum égal à 80 %.

Les hauteurs annuelles évaporées varient d'une station à l'autre sous l'influence des divers facteurs climatiques. L'évaporation suit le mouvement annuel de l'humidité et de la température de l'air. Ainsi, elle oscille entre 1000 mm/an dans la moitié Ouest et 2500 mm/an du côté Est.

Ces différents éléments climatiques combinés font apparaître :

— une évapotranspiration très forte durant toute l'année, spécialement pendant la saison humide qui est une saison chaude;

— le prélèvement de l'évapotranspiration sur les pluies étant prioritaire par rapport à l'écoulement superficiel et à l'infiltration, il en résulte pour les eaux souterraines, que: les quantités d'eau annuellement disponibles pour l'infiltration sont en général très faibles, d'autant plus que les zones d'affleurement des principaux aquifères se trouvent là où la pluviométrie est moindre; l'infiltration potentielle est marquée, comme la pluviométrie par une forte irrégularité annuelle et interannuelle.

Toute exploitation intensive des eaux souterraines doit tenir compte des répercussions sur l'aquifère d'une éventuelle série d'années sans réalimentation.

Cadre géologique (Fig. 3)

Le Bassin du Maranhão (700,000 km^2), cuvette à peu près circulaire de 3000 m de profondeur en son maximum, faillée sur son bord Ouest, est comblée par une sédimentation à peu près continue du Dévono-Silurien au Crétacé avec de sills de diabases, très étendus et de dikes très nombreux.

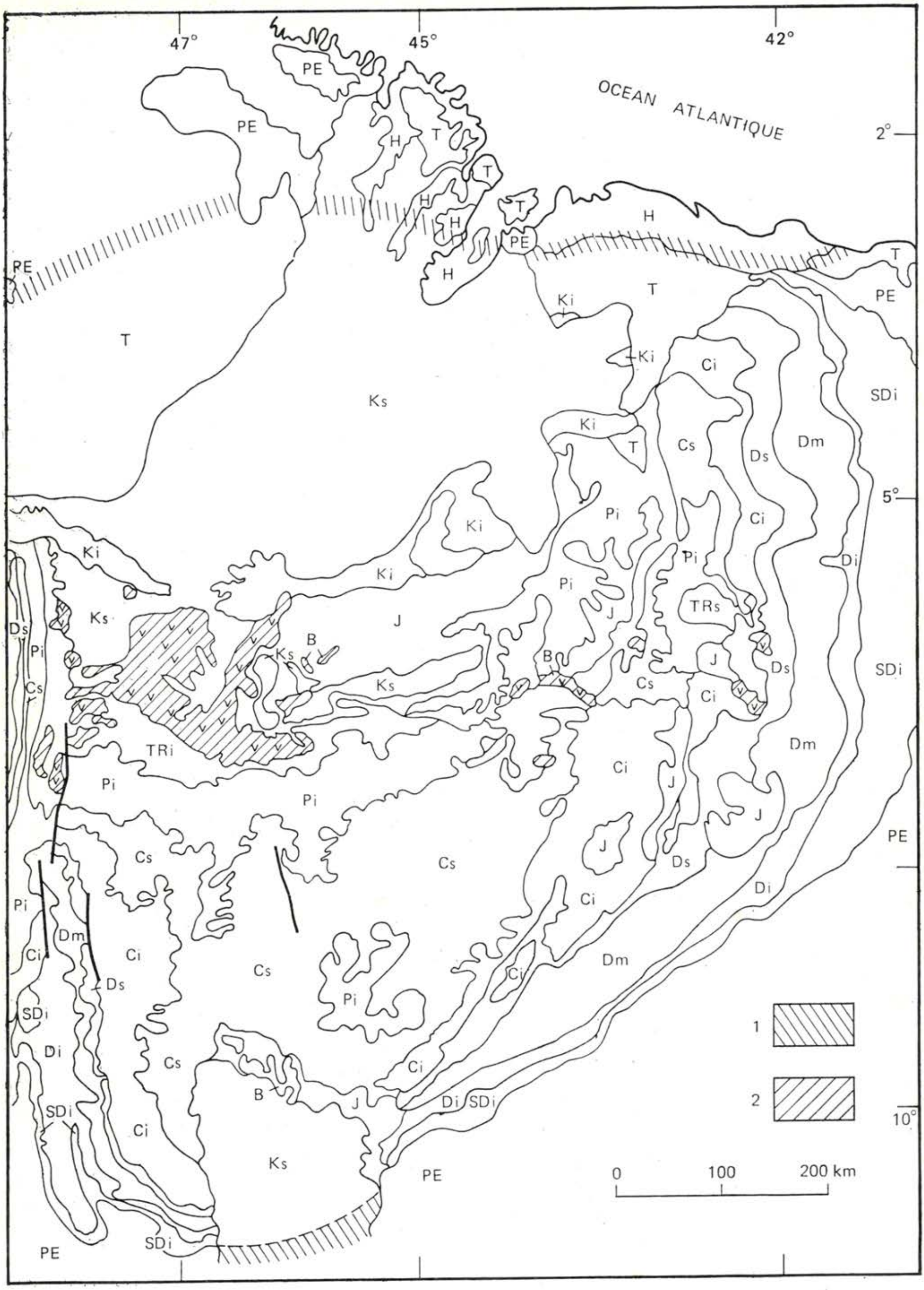

Fig. 3. Bassin du Maranhão. Cadre géologique. (Extrait Carte Géologique Brésil, 1971)

1. Cristallin peu profond, *2.* basalte et associées. Dm = Formation Cabeças, Di = F. Pimenteiras, SDi = F. Serra Grande, PE = substratum cristallin, TRs = F. Pastos Bons, TRi = F. Sambaiba, Motuca, Pi = F. Pedra de Fogo, Cs = F. Piaui, Ci = F. Poti, Ds = F. Longá, H = alluvions, dunes, T = F. Barreiras, Ks = F. Itapecuru, Ki = F. Codo, Grajau, J = F. Corda

Pour bien situer le cadre de notre exposé, où la géologie n'est qu'une source d'information, signalons que de multiples articles ont déjà été publiés sur le Bassin du Maranhão.

Depuis 1950 la Petrobrás a exécuté sur le domaine des compagnes géologiques-géophysiques (sismique, gravimétrie) et une vingtaine de sondages profonds de contrôle. Ces travaux ont permis d'établir les grandes lignes de la structure du Bassin du Maranhão. Les faciès lithologiques traversés par les forages se sont révélés identiques à ceux des affleurements.

Mesner et Wooldridge (1964) ont identifié trois grandes séquences sédimentaires, bien caractérisées par le climat et par des schémas téctoniques de déposition différents. Elles sont:

1. La séquence inférieure, essentiellement clastique, d'âge Néo-Silurien, Dévonien et Mississipien;

2. la séquence moyenne constituée par des sédiments de couleur rouge, dépôts d'anhidrite, dolomites calcaires et grès fins continentaux, d'âge Pennsylvanien, Permien et Triasique;

3. la séquence supérieure constituée par des sédiments argilo-sableux d'âge Crétacé.

Entre les séquences 1 et 2 il y a une discordance avec angularité de 3 à 5° et entre les séquences 2 et 3, il y a un hiatus pendant lequel des diabases et basaltes ont été injectés et ont formé des sills ou coulées plus ou moins importants. Sur le bord Est du bassin les terrains affleurants sont plutôt paléozoïques, dévono-siluriens et carbonifères, à prédominance clastiques, plus ou moins sableux, continentaux à la base et marins vers le sommet à partir du Dévonien inférieur.

Sur le bord Ouest les terrains paléozoïques sont recouverts, en discordance érosive, par les sédiments du Mésozoïque. La lithologie est à prédominance silt-argileuse. La persistence des composants argileux vers le sommet, ne doit pas être écartée.

Les conditions géologiques (alternance de couches perméables et peu perméables, géométrie en « fond d'assiette ») alliées à une superficie topographique qui descend, en marche d'escalier, vers le centre, assurant des conditions d'artésianisme des eaux souterraines sur près de 1/3 du domaine hydrogéologique.

Conditions hydrogéologiques (Fig. 4)

Depuis quelques années, l'importance des eaux souterraines s'impose à tous et d'autant plus que le coût moyen du stockage, traitement et transport des eaux de surface devient tellement élevé que, pour des raisons économiques, il est de plus en plus fréquemment recommandé aux collectivités brésiliennes de ne se tourner vers celles-ci que si les eaux souterraines sont insuffisantes ou inutilisables. Un effort à l'échelle régionale fut donc développé pour que, partout, des renseignements précis puissent être offerts à chaque utilisateur.

En ce qui concerne le problème de l'eau, Small (1914) soulève la nécessité de faire appel aux eaux souterraines qui doivent y exister face à l'absence de sites convenables pour la construction de barrages-réservoirs.

A l'heure actuelle, des cartes hydrogéologiques à l'échelle de 1/500,000 couvrent pratiquement tout le Bassin du Maranhão. Chaque feuille (2° lat.

par 3° long.) comprend les éléments essentiels, dont la représentation a été faite selon les normes internationales.

Dans le cadre géologique précedemment analysé, 5 systèmes aquifères principaux apparaissent : l'aquifère Serra Grande, l'aquifère Cabeças, l'aquifère Poti-Piaui, l'aquifère Sambaiba et le système Corda-Grajaú.

Les systèmes aquifères

L'aquifère Serra Grande est constitué par un mélange en proportion variable de grès grossiers feldspathiques, d'origine continentale, avec des intercalations de textures plus fines et souvent même schisteuses. Le passage entre le faciès inférieur, plus grossier et le supérieur, plus fin, est graduel.

Avec ses 300 m d'épaisseur moyenne et 400,000 km² de superficie, la Formation Serra Grande constitue un des plus importants systèmes aquifères du Brésil. Le substratum de ce système aquifère est le cristallin. Les eaux souterraines y sont mises en charge par les faciès argileux de la Formation Pimenteiras. Cet ensemble est d'origine marine de couleur rouge prédominante. Une des caractéristiques est la grande hétérogénéité de faciès dans le sens horizontal aussi bien que vertical. Les passages plus gréseux offrent parfois de débits importants, mais à l'échelle régionale la Formation Pimenteiras constitue le niveau confinant du système aquifère Serra Grande et le substratum du système Cabeças.

La Formation Cabeças, ayant une épaisseur moyenne de 200 m et près de 300,000 km² de superficie, constitue le deuxième système aquifère plus important du Bassin du Maranhão. Elle est caractérisée par un lithofaciès clastique du grès fin, mais pouvant être également du type moyen, couleur gris-foncé et rouge.

La Formation Cabeças d'âge Dévonien moyen (Kegel 1953), d'origine marine, repose en concordance sur la Formation Pimenteiras qui constitue, du point de vue hydraulique, un aquitard.

L'écoulement régulier vers le centre du bassin fait que les eaux y emmagasinées soit sousmises à des pressions croissantes sous l'aquiclude Longá. Cette Formation (150 m) est constituée par des schistes et silts bitumineux.

Le troisième système aquifère du Bassin du Maranhão est représenté par les Formations Poti (200 m) et Piaui (150 m). Cet ensemble est caractérisé dans sa majeure partie, par des grès blanchâtres, continentaux, d'âge Carbonifère. Cette couche est récouverte par des sédiments à prédominance argileux surtout sur les niveaux moyen et supérieur de la Formation Piaui, et par les couches permiennes des Formations Pedra de Fogo et Motuca. Ces Formations atteignent une épaisseur de l'ordre de 400 mètres et sont essentiellement composées d'argiles, silts, dolomites et anhidrites.

L'aquifère Sambaiba est représenté par des grès peu consolidés, de granulométrie fine à moyenne avec des stratifications croisées. Le système est confiné par les coulées basaltiques (F. Mosquito) d'âge Jura-Trias moyen (Aguiar 1969). L'étanchéité est encore augmentée par les schistes bitumineux et argiles feuilletées de la F. Pastos Bons.

Finalement, le système aquifère Corda-Grajaú d'âge Jurassique-Crétacé. Cette unité est constituée par des grès rouges à ciment carbonaté. Des coulées basaltiques (F. Sardinha), des schistes bitumineux, avec des intercalations

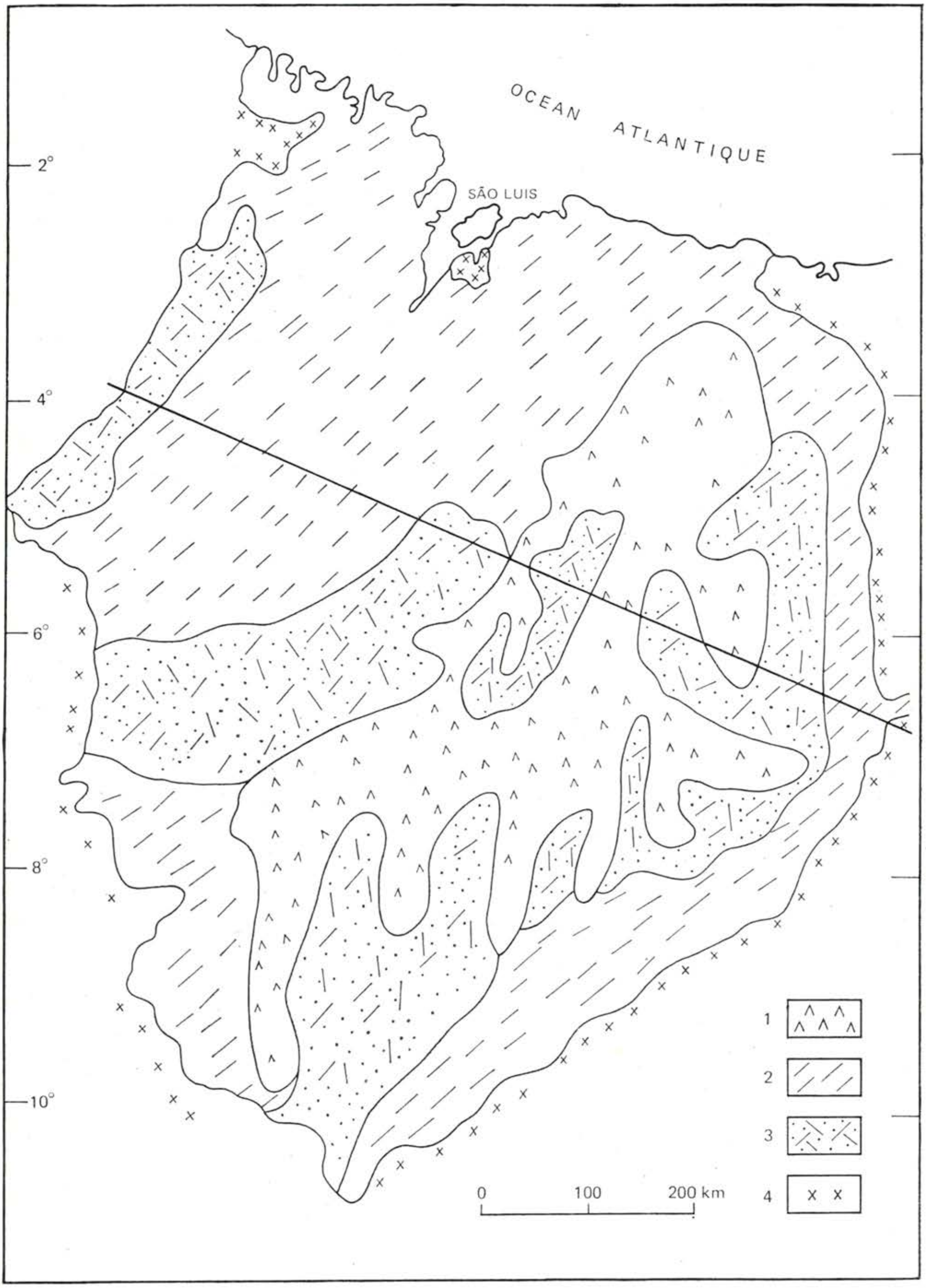

Fig. 4a Bassin du Maranhão. Conditions de gisement des eaux souterraines

1. Eaux artésiennes, *2.* eaux peu profondes <50 m, *3.* eaux profondes 50—100 m, *4.* limite géologique du bassin

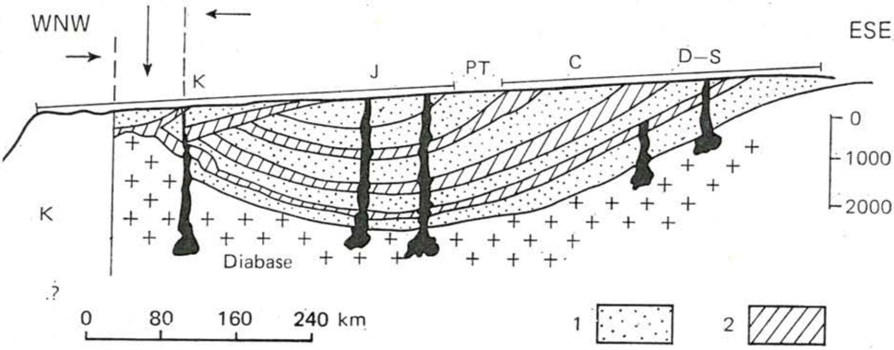

Fig. 4b Bassin du Maranhão. Conditions de gisement des eaux souterraines
1. Perméable, *2.* imperméable

de calcaire et anhidrites de la F. Codó, et des sédiments très argileux (F. Itapecuru) dont l'épaisseur peut atteindre près de 1000 mètres, constituent le recouvrement de ce dernier système aquifère.

Les paramètres hydrodynamiques des différents systèmes aquifères ont été calculés à partir des données des pompages d'essai, selon les méthodes classiques préconisées par JACOB (1950), HANTUSH (1956) et WALTON (1970) d'après des temps de pompages d'au moins 24 heures.

Les valeurs moyennes sont de $3 \cdot 10^{-3}$ m^3/s pour la transmissivité et de $5 \cdot 10^{-4}$ pour le coefficient d'emmagasinement. En zone libre S est de $5 \cdot 10^{-2}$ en moyenne.

Les conditions d'alimentation et de décharge

La seule source d'alimentation est représentée, en cette occurrence, par la fraction d'eau de pluie qui parvient à s'infiltrer. L'allure des courbes isopiézométriques démontre que, en dehors des courtes périodes d'inondation, les cas des cours d'eau qui alimentent les nappes aquifères ne sont qu'exceptionnels.

La piézométrie montre un écoulement de direction variable, mais dont la résultant coïncide avec le sens de plongement des couches. Les perturbations dans ce comportement général sont attribuées aux influences de la topographie, anisotropie et conditions structurales localisées.

Ainsi, les flux sont vraisemblablement dirigés vers le centre du bassin et ils doivent participer, par endroits, à l'alimentation des aquifères voisins à travers les niveaux peu perméables intermédiaires.

L'augmentation de l'épaisseur des couches aquifères dans le sens des flux serait l'origine de si faibles gradients. Il faut tenir compte en plus, des conditions aux limites de cet écoulement, notamment sous la couverture des sédiments crétacés.

Localement, les conditions d'écoulement sont influencées par la présence de dikes et sills de roches basiques interstratifiés et surtout abondants sur le bord Ouest du bassin.

En ce qui concerne les conditions aux exutoires sur le bord Nord, notons que les écoulements des niveaux captifs doivent être partiellement bloqués par le seuil imperméable du substratum cristallin peu profond (voir Fig. 3) et ne se produisent que vers le haut à la faveur d'accidents ou à travers des passages relativement plus perméables des aquicludes ou aquitards sus-jacents.

Ces aspects des conditions générales d'alimentation et décharge des aquifères, vont se traduire par les faibles valeurs des coefficients de rénouvellement (rapport de l'écoulement annuel aux réserves permanentes), 10^{-3} en moyenne, et par le fait que ce n'est, en général, que dans les parties superficielles des aquifères sus-jacents aux seuils, dans la partie qui a été lavé par les eaux d'infiltration d'origine météorique, que ces terrains contiennent de l'eau douce (Reboucas, 1973).

Évaluation des réserves

Les calculs des réserves régulatrices, volume infiltré transitant chaque année dans les aquifères libres et en charge, ont été faits surtout à partir du réseau de courbes isopiézométriques qui, par son importance fut considéré comme représentatif du fonctionnement de l'ensemble des nappes. D'autre part, nous avons étendu à chaque système aquifère les caractéristiques hydrodynamiques *(T, S)* qui y avaient été localement déterminées.

Les caractéristiques dimensionnelles des réservoirs ont été déduites des éléments géologiques en notre possession. Pour les réserves permanentes nous avons adopté une profondeur limite de 1000 mètres. Sur le plan technique les ressources exploitables d'un système aquifère quelconque peuvent être évaluées à partir de leur paramètres hydrodynamiques: coefficient de transmissivité *(T)* et coefficient d'emmagasinement *(S)*.

Ainsi à partir des valeurs des coefficients hydrodynamiques qui ont été reconnues comme représentatifs de l'ensemble du domaine, nous avons appliqué la méthode d'évaluation préconisée par Theis (1935). En fait, on peut raisonnablement envisager l'exploitation étalée sur une période de cinquante ans (deux générations) de la partie des réserves accessibles par forage pendant cette période. Nous avons adopté pour leur estimation, une profondeur limite de 1000 mètres, un débit constant et un niveau de pompage à ne pas dépasser les 100 m (Reboucas 1972).

Aucune hypothèse quant aux limites n'a été faite, c'est-à-dire, le domaine d'exploitation est infini, homogène et non influencé. De toute manière, il a été ménagé dans tous les calculs des marges de sécurité systématiques, de sorte que les évaluations qui suivent peuvent être considérées comme une approche pessimiste du problème.

Écoulement [m³/an]	
libre	$3 \cdot 10^9$
en charge	$1 \cdot 10^9$
Réserves permanentes [m³]	
saturation	$3 \cdot 10^{12}$
sous pression	$90 \cdot 10^9$
Exploitables à 1000 m et 50 an [m³]	$1.4 \cdot 10^{12}$

Qualité chimique des eaux

L'étude la plus complète et intéressant pratiquement l'ensemble de la région Nord-Est du Brésil, a été faite par Cruz et Melo (1968).

Cette analyse a été reprise ici en grande partie, en tenant compte des nouvelles données, près de 600 analyses. La zonalité horizontale des types chimiques montre que dans le bassin hydrogéologique du Maranhão les types mixtes et bicarbonatés coïncident d'une façon remarquable avec les zones d'affleurement des aquifères profonds. Les eaux moins salines (RS), inférieur à 100 mg/l sur 80 % de cas appartiennent aux types mixtes. Les types hydrocarbonatés sont plus minéralisés ayant, sur 75 % des cas, un résidu sec (RS) supérieur à 450 mg/l. La fréquence des eaux hydrocarbonatées à Ca, Mg et Na est de 43 %.

Le troisième type plus fréquent est représenté par les eaux chlorurées dont la fréquence est de 22 %, le résidu sec est déjà supérieur à 900 mg/l sur 75 % des cas. Les autres types hydrocarbonatés chlorurés, hydrocarbonatés sulfatés, sulfatés et sulfatés chlorurés ont des fréquences très faibles, soit de 15 % au total.

Il faut souligner que la majeure partie des échantillons d'eau ont été prélevées dans des forages à la fin du nettoyage et développement du trou. Les profondeurs des forages sélectionnés se situent entre 100 et 150 mètres.

Dans les forages plus profonds quoique les aspects constructifs ne soient pas très satisfaisants pour assurer un bon échantillon des eaux appartenants à chacun des systèmes aquifères traversés, on identifie une zonalité verticale.

En outre, il faut souligner que, dans les domaines où les nappes aquifères sont superposées et séparées par des couches peu perméables, on doit tenir compte de ce qu'elles sont hydrauliquement en communication entre elles. Cette circonstance est d'une grande importance pratique pour les régimes d'exploitation. En fait, les nappes exploitées peuvent être alimentées par l'apport d'eau d'autres nappes, à la suite des pompages de longues durées.

Un nombre croissant des cas révelant des modifications de la qualité des eaux des forages constitue des preuves qui illustrent sans doute les conditions générales d'interchange entre les différentes systèmes aquifères.

Les zones hydrochimiques verticales dans lesquelles les eaux souterraines se différencient par leur composition chimique sont les suivantes:

— Une zone supérieure dont l'épaisseur est de l'ordre de 100 à 150 m. Les eaux souterraines y sont à potabilité bonne et d'ordinaire hydrocarbonatées calciques.

— Ensuite, en règle générale, vient une zone des eaux mélangées, bicarbonatées-sodiques, bicarbonatées chlorurées et sulfatées-sodiques et calciques, pour des profondeurs très variables. Ce comportement semble être lié en partie à la présence des assises argileuses étendues qui contiennent de sels et qui renferment les systèmes aquifères plus importants.

— A partir de 1000 à 1500 m, les eaux deviennent spécialement riches en sels. Dans quelques cas la teneur en chlorures de sodium dépasse 150,000 ppm.

D'ailleurs, les logs éléctriques des sondages plus profonds (2000 à 3000 m) faits par les pétroliers indiquent des valeurs très faibles de la résistivité (1 à 2 ohm.m) à partir de 1000 mètres. Notons que le pétrolier a un intérêt spécial par les structures fermées. La possibilité de trouver de l'eau douce dans des forages plus profonds ne doit pas être écartée, suivant les conditions plus favorables de circulation.

En appréciant les données des forages, tenant compte des conditions hydrogéologiques des secteurs aquifères captés, on notera que le géochimiste des eaux souterraines est spécialement lié au régime hydraulique des réservoirs, notamment en ce qui concerne les conditions d'assainissement.

Perspectives de l'utilisation

Jusqu'à une période très récente l'exploitation des eaux souterraines du Grand Bassin Hydrogéologique du Maranhão avait été menée avec des moyens techniques insuffisants. Si les sondeuses à percussion (battage) se sont montrées adaptées aux terrains cristallins, leur utilisation dans ce domaine sédimentaire est le plus souvent impraticable. On peut multiplier les exemples de forages que se sont arrêtés au toit des niveaux en charge ou artésiens, faut de n'avoir pu y pénétrer (Reboucas et Gaspary 1966).

La cartographie hydrogéologique à l'échelle de 1/500,000 menée par les équipes de la Surintendance pour le Développement du Nord-Est (SUDENE) a inventorié près de 2 mil points d'eau.

Les forages plus récents, convenablement construits donnent des débits spontanés de plusieurs centaines de m^3/heure; le maximum atteint est de l'ordre d'un millier de m^3/heure. Il s'agit évidemment d'un cas exceptionnel.

L'étude des aquifères profonds, en charge et de grandes dimensions, doit porter en priorité sur la connaissance des caractéristiques hydrodynamiques et dimensionnelles. Ces données serviront à évaluer les ressources exploitables par l'établissement de modèles de gestion.

L'écoulement devra évidemment être étudié et le débit estimé, mais on pourra se contenter, dans ce domaine, d'une connaissance schématique qui aura pour but de préciser les conditions aux limites et le régime hydraulique des aquifères en cause.

Notons que la question couramment posée par les exploitants, à savoir s'il est prudent ou non de puiser dans les réserves permanentes des aquifères à faible taux de rénouvellement, se présent comme un faut problème. En effet, toute mise en exploitation d'un aquifère se traduit par une baisse de niveau d'eau, donc par la diminution inéluctable de ses réserves. La seule réponse possible à une telle question consiste à réaliser la planification à long terme de cette exploitation, en fonction de toutes les données y compris les données économiques des projets.

L'analyse à laquelle nous sommes parvenus, constitue un véritable outil de travail et de réflexion pour la détermination des priorités d'allocation des ressources disponibles. Il faut toutefois les utiliser avec le maximum de profit pour l'économie générale du pays et sans oublier les intérêts des artisans de l'augmentation de production. L'eau est l'un de ces moyens, et chercher à atteindre cet objectif signifie que l'on doit chercher à assurer par avance des ressources adaptées, en qualité, quantité et prix, à la structure socio-économique visée.

La simulation utilisée comme outil prévisionnel représente alors une aide précieuse dans la tâche difficile qui consiste à prendre des décisions.

Conclusions

Les paramètres essentiels du climat font apparaître que l'eau d'abord très abondante dans les régions situées à l'Ouest, se rarifie au fur et à mesure qu'on se déplace vers l'Est, par un phénomène d'aggravation où l'évaporation s'intensifie en même temps que les précipitations deviennent plus faibles et irrégulières.

Les eaux souterraines du bassin hydrogéologique du Maranhão, situées à l'abri des irrégularités pluviométriques peuvent satisfaire, dans le cas général, aux demandes domestiques et, en plus, servir localement à une irrigation d'appoint.

La solution ne peut être qu'un compromis, étudier le problème de l'eau sous le double aspect: technique et économique. Les résultats des modèles de gestion doivent fournir les bases objectives de choix de l'aménagement optimum, où les disponibilités en eau sont exprimées en termes monétaires et non plus en volumes emmagasinés. La réalisation du bilan financier permet alors de définir le seuil économique d'exploitation au-delà duquel les coûts de production de l'eau sont supérieurs aux bénéfices que provoque cette ressource.

La tâche des planifications consistera à trouver le juste équilibre entre les diverses solutions possibles et à en établir l'ordre de priorité. Elle est indispensable pour sortir de l'empirisme actuel qui a conduit à une utilisation restée au total très faible. La disproportion décevante entre les efforts remarquables déployés pour capter l'eau et les maigres bénéfices jusqu'à présent obtenus constituent une bonne preuve de cet état de fait.

En définitive, la solution du problème de l'eau dans le grand bassin hydrogéologique du Maranhão dépend largement d'une politique d'utilisation rationnelle de ses ressources.

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CARACTÈRE HYDROGÉOLOGIQUE ESSENTIEL DE LA GRANDE PLAINE HONGROISE

HYDROGEOLOGICAL PARTICULARITIES OF THE GREAT HUNGARIAN PLAIN

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ABSTRACT

Fluviatile Quaternary and lacustrine Pliocene sedimentary sequences, which have a thickness of about 1500 – 2000 m in the Hungarian Basin, include a high number of aquifers containing fresh-water. The study of these water-bearing layers indicates lots of anomalies manifesting in the size and in the horizontal and vertical distribution of hydrostatic and rock pressure as well as in the temperature and chemical composition of the subsurface waters.

Thousands of observation wells built on the phreatic water table and a considerable number of check-wells built on the artesian aquifers of different depths, going to 1100 m, give wide-ranging informations on all kinds of hydrogeological problems.

Data gathered from stationary level-recording instruments enable us to set apart the effects of the terrestrial ebb and tide from those of the atmospheric pressure and from the fluctuations induced by the water supply in the subsurface water levels. These latter being of prominent importance from the point of view of underground water budget.

Beyond the seasonal fluctuations there are long-term, rising and lowering tendencies too in the subsurface level changes brought about by longer humid and dry climatic periods. The fluctuations caused by the natural water supply in the subsurface water-bearing layers are manifested all over the Great Plain and at any depth to about 500 m in a few days or weeks. This is the speed of propagation of the pressure impulses. The velocity of the seepage is far more slower and could be evaluated only by the determination of the absolute age of the underground waters in the water-bearing layers situated at different depths.

Juste au milieu du continent de l'Europe, entourée par les chaînes des Carpathes est située la Grande Plaine hongroise, nommée Alföld. Cette grande dépression tertiaire et quaternaire s'étend sur une surface environ 100,000 km^2 et la puissance des couches meubles pliocènes et quaternaires qui la remplissent atteint ensemble les 4000 m.

Ce grand bassin rempli de sédiments marins, lacustres et fluviatiles avec plusieures centaines de couches aquifères, qui montrent des anomalies surprenantes quant à la porosité et la transmissibilité des matériaux, au débit des couches aquifères, à la formation de la pression dans les profondeurs différentes, à la température et la composition chimique des eaux, ce bassin est vraiment un modèle excellent créé par la nature pour faire des études hydrogéologiques.

Le fond du bassin dans une profondeur de 3000 à 4000 m est traversé par des montagnes paléozoïques et mésozoïques qui se lèvent du pied du bassin

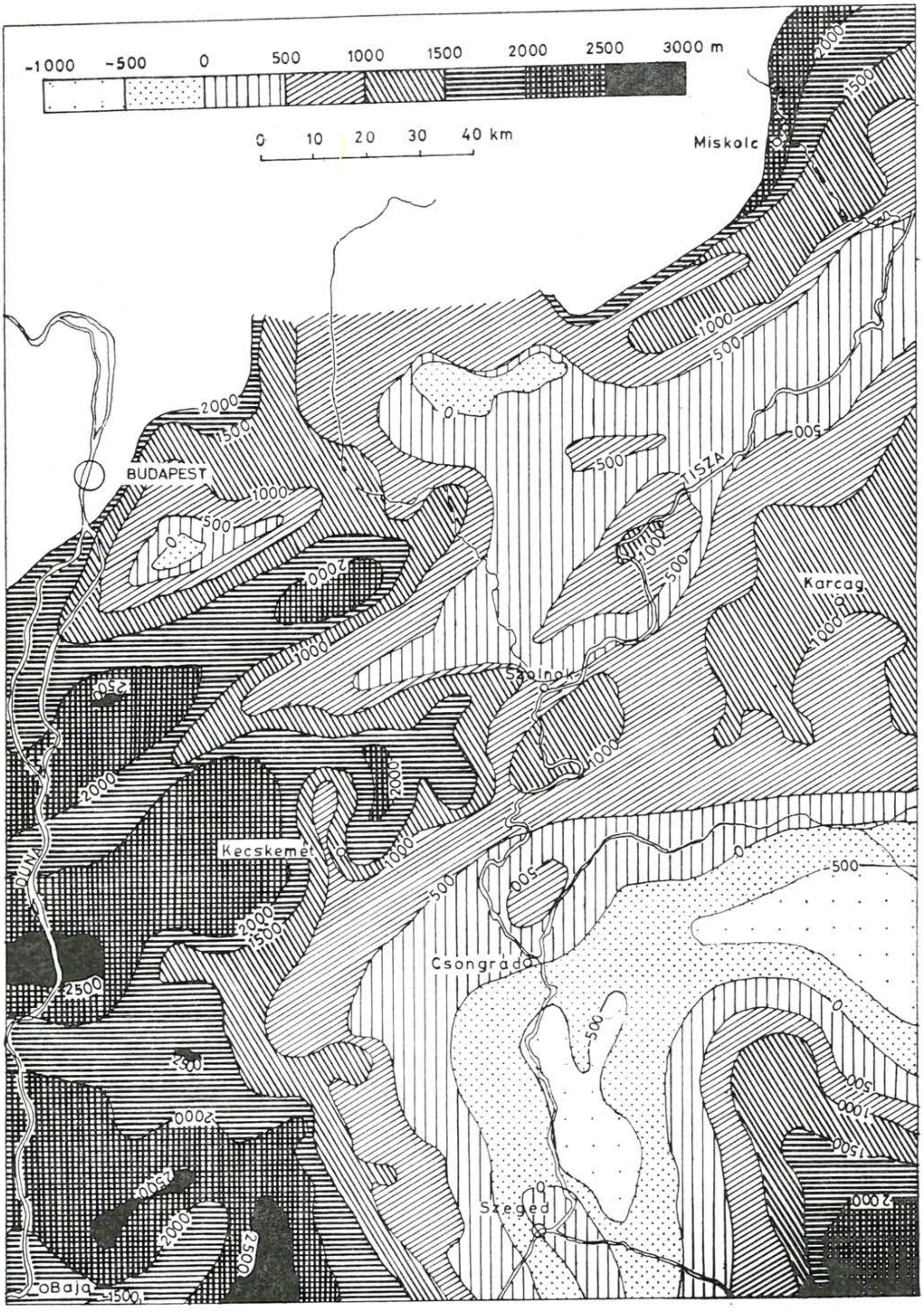

Fig. 1. L'orographie du substratum de l'Alföld. Le niveau de base c'est la profondeur de – 3000 m au-dessous du niveau actuel de la mer

à une altitude relative de 1500—2000 m. Leur allure tend de SW à NE. Une partie du substratum paléozoïque et mésozoïque du bassin est couverte des argiles paléogènes et volcanites tertiaires, les autres tout directement par des couches pliocènes meubles.

C'est la morphologie accentuée du substratum qui a créé un nombre des bassins locaux, différant l'un de l'autre au point de vue hydrogéologique. Les mouvements tectoniques survenus dans le Pliocène et après, ont morcelé le territoire en blocs séparés encore davantage.

La couverture meuble du substratum on peut détacher en 4 parties suivantes.

1. Les couches marines du Pannonien inférieur. Conglomérats, grès et sables, argiles et marnes. Elles couvrent le territoire presqu'entier sous une épaisseur de 1000 à 2000 m. La série est très consolidée, contenant de l'eau salée.

2. La série des couches lacustres, datée comme Pannonien supérieur. Limon et sable densement stratifié, beaucoup plus meuble que la série précédente. Ces couches couvrent le territoire entier en puissance de 500 à 2500 m. Elles contiennent un grand nombre de couches aquifères excellentes. La base de cette série est riche surtout en eaux thermales. L'eau est douce, de caractère d'hydrocarbonate de sodium.

3. La partie troisième est une série d'un côté argileuse et de l'autre caillouteuse et sableuse. C'est le terme final de la sédimentation pliocène lacustre, en même temps il représente déjà les premières couches fluviatiles dans certains affaissements locaux. Cette partie a une puissance de 300 à 400 m, mais elle ne couvre pas le territoire entier du bassin. Tous les deux faciès, celui l'argileux et le sableux sont caractérisés par une granulométrie mixte. Les grains fins de différente dimension composent chez les argiles un assortiment tellement solid, ce qui rend cette formation la plus imperméable. A l'encontre, dans la partie sud de l'Alföld, où les couches du Pliocène final sont caillouteuses et sableuses, nous trouvons des aquifères excellentes.

4. Au-dessus de tous ces successions nous trouvons les couches fluviatiles quaternaires sous une puissance de 200 à 600 m. La sédimentation fluviatile cyclique motivée par les changements climatiques et par des affaissements en escalier produisait des séries consécutives qui se composent de gravier, de sable, de limon et de l'argile, la série se répétant plusieurs fois. On trouve les meilleures couches aquifères parmi ces sédiments.

A l'étude des conditions hydrogéologiques de cette vaste plaine nous disposons d'une bien grande quantité d'information tant relative à l'eau phréatique qu'aux nappes artésiennes. Les recherches peuvent s'appuyer sur un recensement des puits effectué de 1950 à 1954, qui a cartographié et mesuré à la Grande Plaine 770,000 puits, ainsi qu'environ 1000 puits qui ont été établis depuis 1950 à la Grande Plaine pour observer la fluctuation de niveau des eaux phréatiques.

Les couches artésiennes sont étudiées par les forages et puits artésiens dont 41,000 sont enregistrés jusqu'à 1975. Depuis quelques années la fluctuation des niveaux artésiens est aussi observée sur 46 puits dont la profondeur varie entre 30 et 1050 m.

En ce qui concerne la situation du niveau souterrain de l'eau phréatique deux phénomènes sont à relever et à expliquer. L'un c'est la grande diffé-

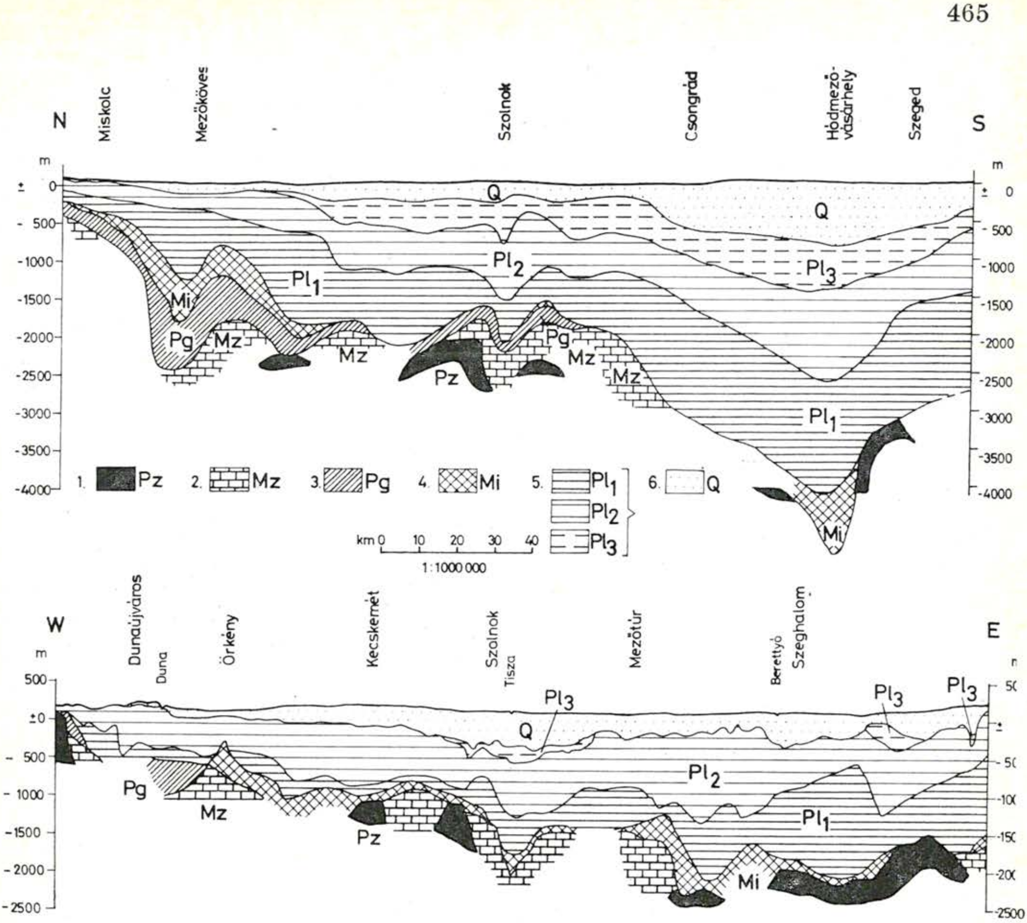

Fig. 2. Coupes géologiques N—S et W—E au milieu de l'Alföld

1. Roches paléozoïques et plus anciennes, *2.* r. mésozoïques, *3.* r. paléogènes, *4.* r. miocènes, *5.* r. pliocène inférieur, moyen et supérieur, *6.* r. quaternaires

rence qui se manifeste dans la profondeur du niveau de la première nappe de l'eau au-dessous du terrain. Ni le relief ni le matériel des couches superficielles n'expliquent ces différences. C'est la structure profonde qui engendre le niveau de balance entre la nappe de l'eau atmosphérique qui s'infiltre vers le bas et de l'eau artésienne qui tend de dessous vers le haut.

Le deuxième phénomène à expliquer c'est le mouvement horizontal très faible et lent de l'eau phréatique le long des inclinaisons du terrain vis à vis de la filtration des eaux profondes qui se bougent sous pression. Ce qui explique p.e. la situation entre les fleuves Danube et Tisza, où dans les collines de sable qui s'élèvent à une hauteur de 60—70 m au-dessus des alluvions des deux fleuves, la nappe phréatique reste 1 à 3 m au-dessous de la surface et ne filtre pas — même pas dans le sable — vers les vallées des deux fleuves situées beaucoup plus bas.

Le manque de mouvement horizontal dans les aquifères phréatiques est démontré aussi par le caractère chimique des eaux. Nous trouvons une minéralisation extrêmement variée et des types chimiques différentes dans les puits

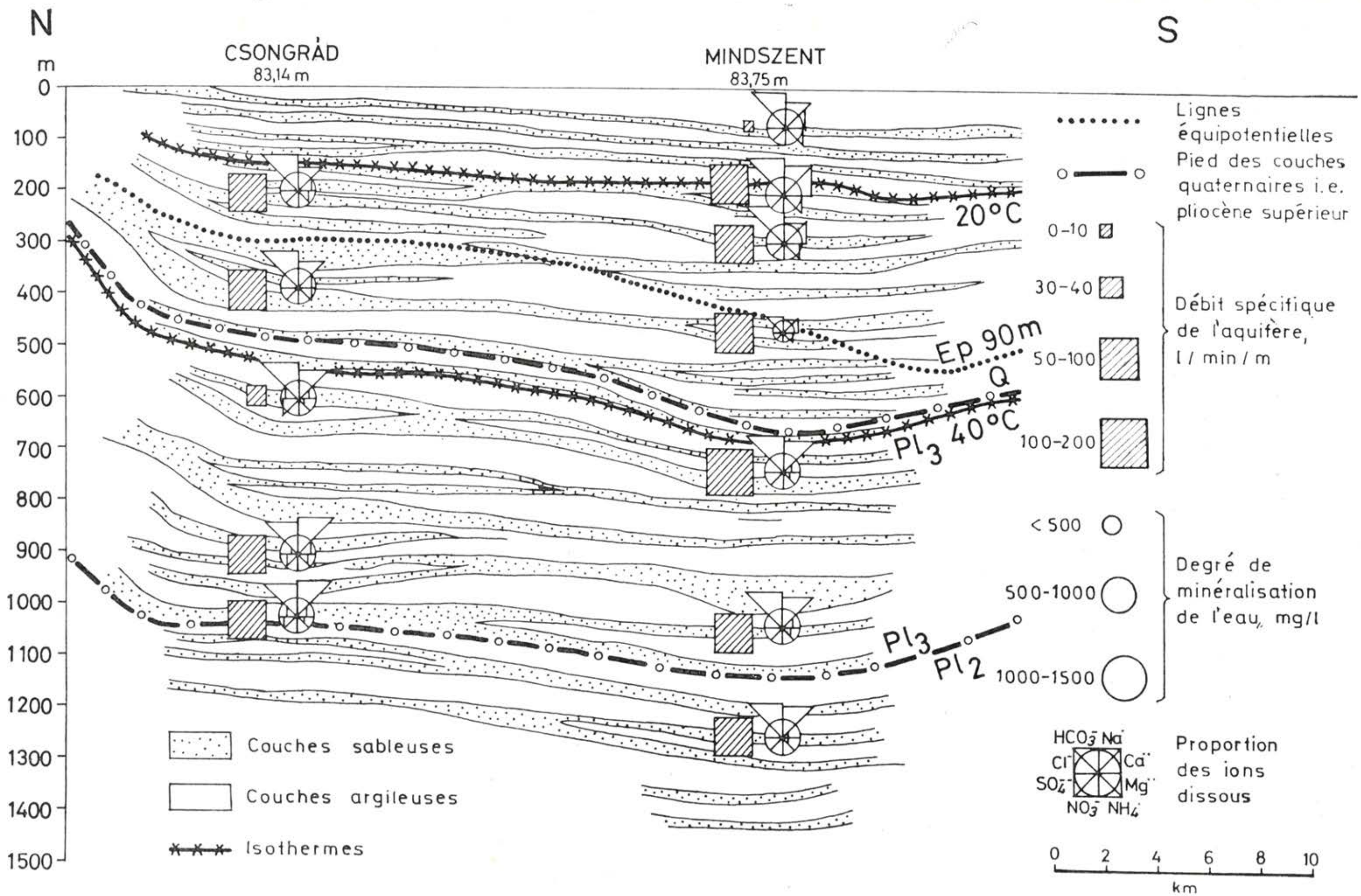

Fig. 3. Détail de la coupe géologique de N—S

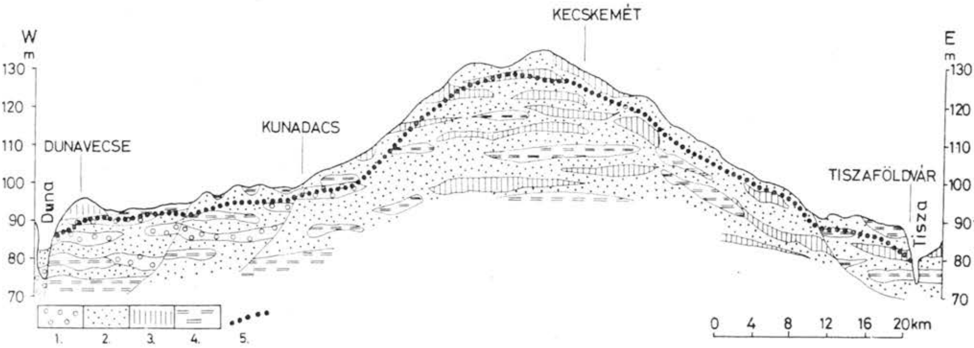

Fig. 4. La position de la nappe phréatique entre le Danube et la Tisza
1. Gravier, *2.* sable, *3.* loess, *4.* limon, *5.* niveau de la nappe phréatique

et forages voisins. Le lessivage et l'échange de matière entre le sol et l'eau se procède plutôt dans la zone de la fluctuation de la nappe. Le résultat dépend des conditions locales, de la série des assises affectées par la fluctuation de la nappe sur le lieu. Et parce qu'il n'y a pas de mouvement horizontal qui pourrait mélanger et homogénéiser les différentes solutions, la minéralisation des eaux phréatiques reste très variée sur des endroits prochaines.

Le lessivage est local, mais il ne l'est pas le rythme de la fluctuation. Celle-ci est influencée par l'ensemble du climat qui prédomine dans le bassin Carpathique.

C'est ce qui explique l'identité des mouvements saisonniers et multiannuels sur le territoire entier de la Grande Plaine.

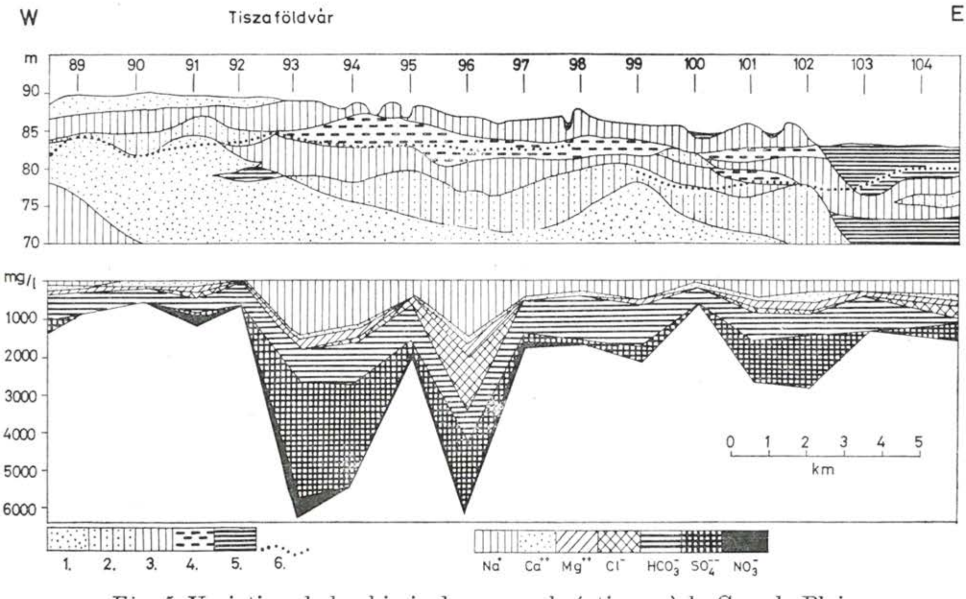

Fig. 5. Variation de la chimie des eaux phréatiques à la Grande Plaine
1. Sable, *2.* sable loessique, *3.* loess, *4.* limon, *5.* argile, *6.* niveau de la première nappe de l'eau

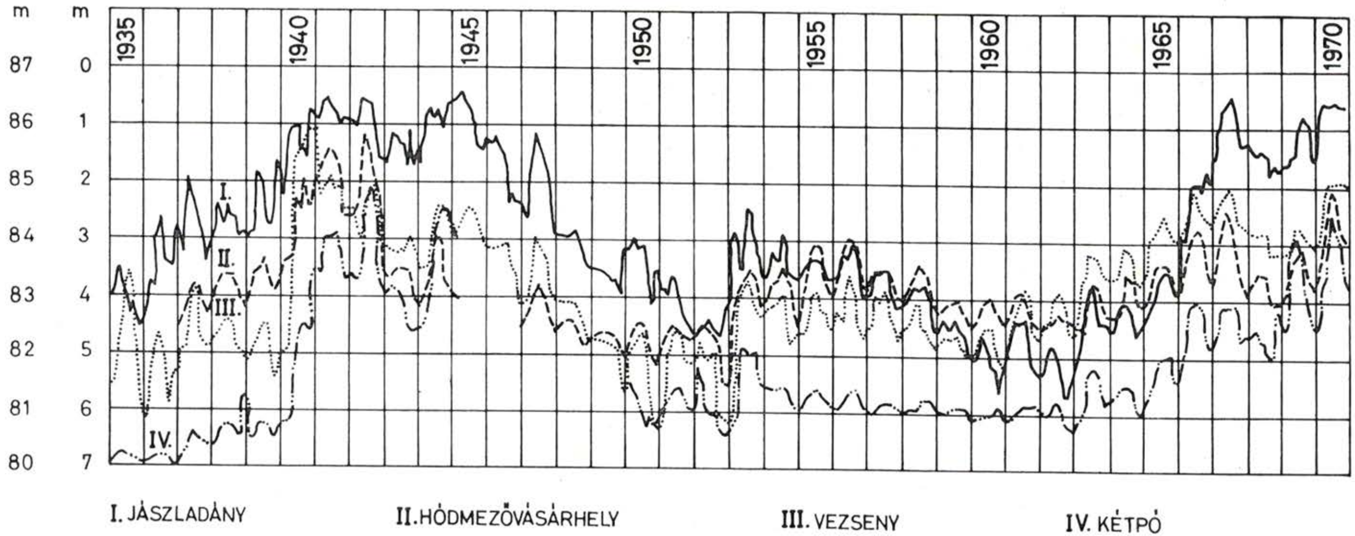

Fig. 6. Fluctuations multi-annuelles de la nappe phréatique à différents points de la Grande Plaine

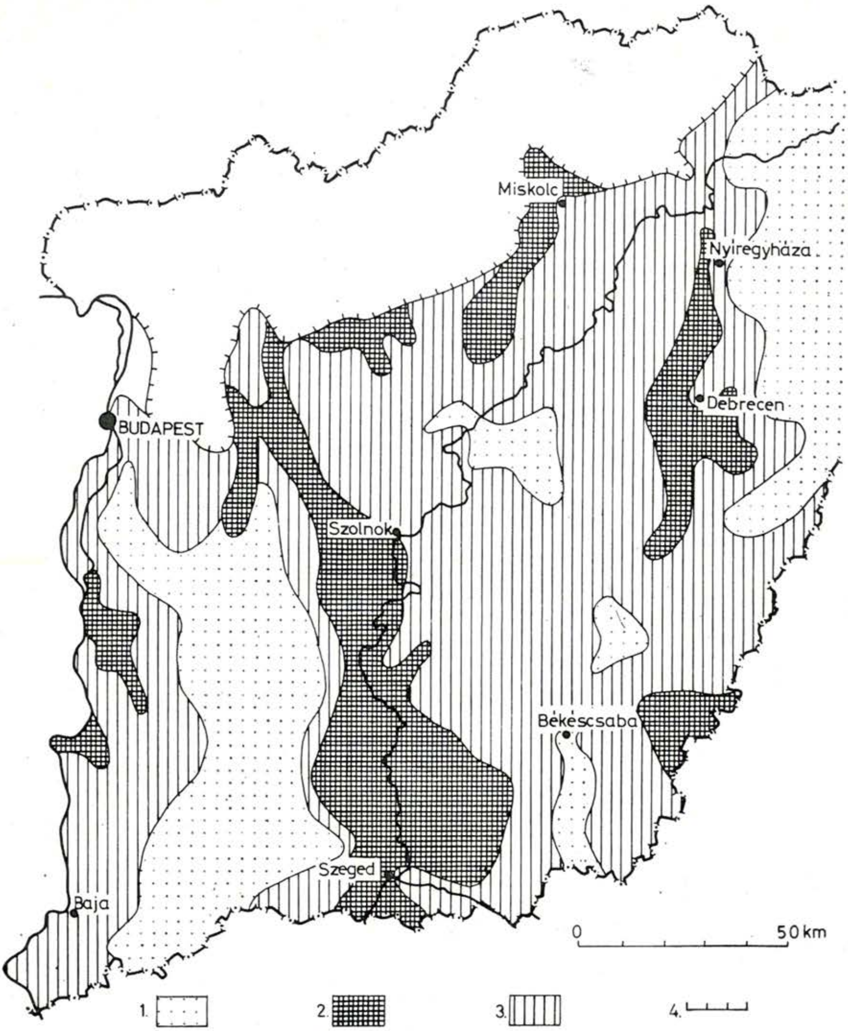

Fig. 7. Variation de la pression dans les aquifères artésiennes

1. Le gradient de pression se diminue vers le bas. *2.* Le gradient de la pression augmente vers le bas. *3.* Territoires transitoires. Conditions hydrostatiques. *4.* Limite des montagnes du Nord

Au-dessous de la nappe phréatique parmi les sédiments fluviatiles quaternaires, ainsi que dans les successions lacustres pliocènes un grand nombre de couches aquifères sont emprisonnées, détachées l'un de l'autre par des couches limoneuses. La filtration de l'eau à travers et entre ces couches quelquefois lenticulaires, quelquefois étendues sur des territoires vastes est bien compliquée. Il est soumis tant d'influences, que les conditions hydrogéologiques montrent une variation excessive tant horizontalement que verticalement. Malgré tous ces différences on a observé des mouvements souterrains qui viennent des montagnes périphériques, des roches karstiques et fissurées, des terrasses de graviers et des cônes de déjection des rivières et qui conduisent l'eau vers le milieu de la Grande Plaine à une profondeur de plusieures centaines de mètres. D'ici du fond du bassin, l'eau poussé par la pression hydrostatique et par la pression des couches, filtre vers le haut à travers toutes les couches perméables et semi-perméables jusqu'à la surface, c'est-à-dire jusqu'à la nappe phréatique. Parmi les forces qui font monter l'eau, outre la pression hydrostatique et la pression des couches, c'est la température augmentée qui joue un certain rôle, ensuite l'eau gazeuse venante de plus grande profondeur ainsi que l'élasticité accrue des formations dans cette profondeur.

Le mouvement de l'eau se dirigeant vers le centre et en bas d'une part, l'ascension vers le haut au milieu de la plaine, d'autre part, créent une circulation permanente souterraine. Les gradients du changement de la pression vers la profondeur découvrent les lieux, où le mouvement de l'eau tend vers le bas et ceux, où elle s'infiltre en haut. Les niveaux piézométriques des eaux se rangent de plus en plus bas dans les aquifères sous-jacentes sur les lieux, où l'eau ruisselle en bas et ils se situent de plus en plus haut dans les couches, où l'eau monte vers le haut.

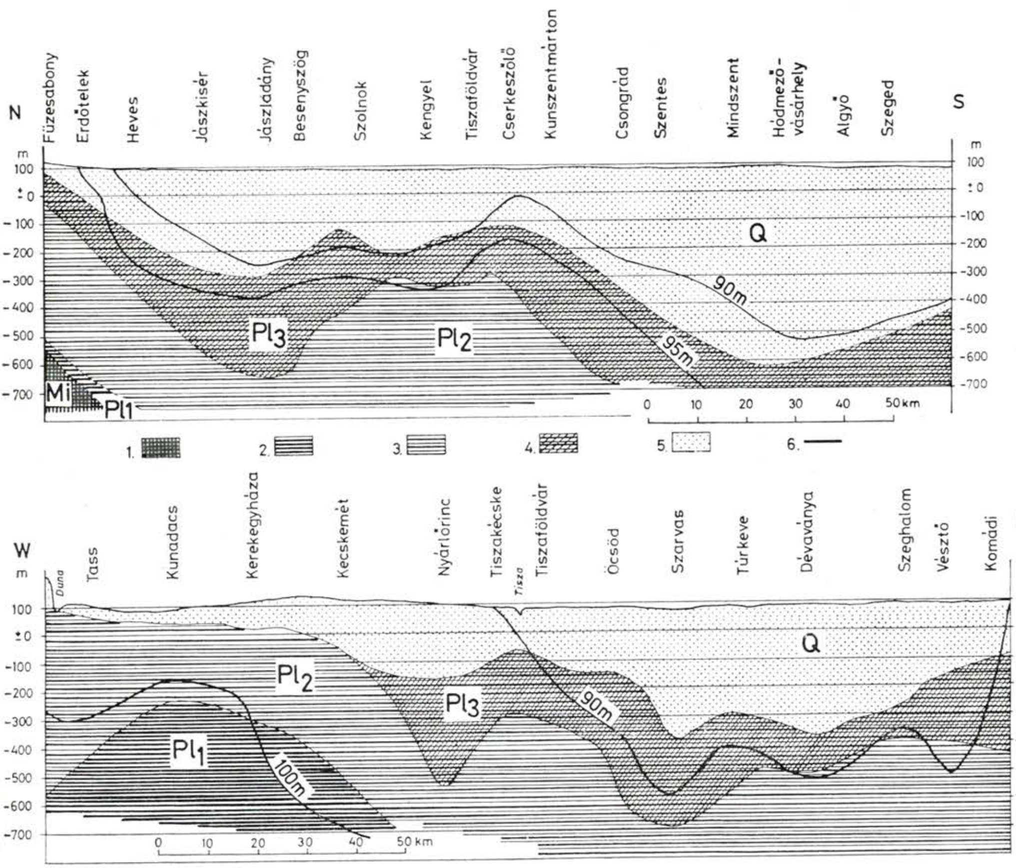

Fig. 8. Variation de la pression dans les nappes artésiennes le long d'une coupe géologique N–S et W–E

1. Miocène, *2.* Pannonien inférieur, *3.* Pannonien supérieur, *4.* Pliocène final, *5.* Quaternaire, *6.* lignes équipotentielles d'où l'eau monte jusqu'à 90–95 m (resp. 90–100 m) au-dessus du niveau de la mer

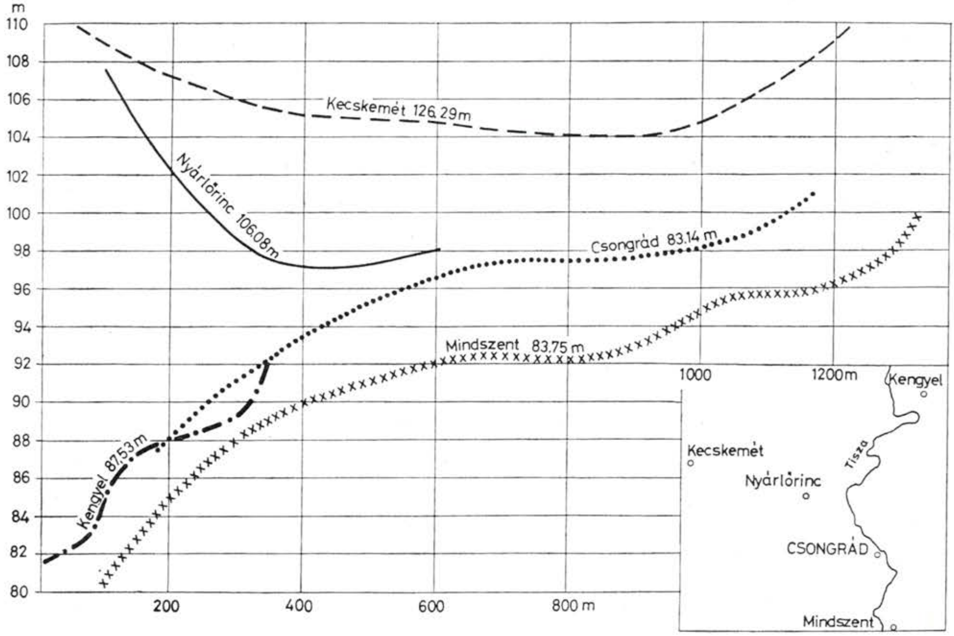

Fig. 9. Relation entre la profondeur de la nappe et le niveau piézométrique de l'eau

Ces mouvements ne sont pas réguliers non plus. Les lignes équipotentielles nous donnent des informations plus minutieuses. Ces lignes s'accomodent à l'orographie de la surface, à la structure géologique, à la succession des couches perméables et imperméables, à la pression hydrostatique combinée avec la pression des couches. L'un des plus importants des facteurs qui précisent la situation des lignes équipotentielles c'est la distance des régions d'alimentation élevées. La position prise par les lignes équipotentielles nous indique les lieux et la profondeur d'où il est à prévoir que l'eau sautera au-dessus du terrain.

L'augmentation et la diminution du gradient de la pression n'est pas uniforme parmi les couches souterraines. Les différences obtenues sont causées premièrement par des différences lithologiques locales.

Le gradient négatif de la pression c'est-à-dire les niveaux piézométriques baissant vers le bas se présentent seulement jusqu'à une certaine profondeur dans les couches souterraines. Ils s'inversent dans une profondeur donnée et en pénétrant encore plus bas, les niveaux commencent à augmenter. Le lieu de ces tournants se trouve à des profondeurs différentes. Juste au-dessous du terrain nous trouvons quelquefois dans les forages une section plus ou moins longue où le gradient de la pression est zéro, c'est-à-dire les niveaux piézométriques restent les mêmes en allant vers le bas. La pression ici est normale hydrostatique. Au-dessous de cette section le gradient vient d'être positif, les niveaux commencent à augmenter.

Ce point de tournant du gradient de la pression indique la profondeur jusqu'à laquelle l'infiltration peut aller vers le bas. Cette profondeur peut atteindre 600–800 m à la Grande Plaine hongroise. Il faut souligner que cette filtration n'est pas un mouvement vertical, mais une progression de l'eau le long d'une pente plus ou moins raide, qui vient des périphéries, quelquefois d'une distance considérable.

On peut suivre la voie de l'eau vers la profondeur entre les sédiments non-consolidés aussi à l'aide des analyses chimiques. L'eau atmosphérique possédante de CO_2 devient, en général, bicarbonatée à calcium et magnésium juste après la percolation. Elle conserve ce caractère dans les couches à grains grossiers, dans les graviers et dans le sable grossier. En arrivant parmi les couches limono-argileuses, le caractère chimique se change, l'eau devient hydrocarbonatée à sodium. Nous trouvons ce dernier type de minéralisation presque sur tout le territoire de l'Alföld et dans les couches situées au delà d'un profondeur de 100 à 200 m jusqu'à 1000 à 1500 m. L'eau à hydrocarbonate de calcium progresse seulement dans quelques cônes de déjection caillouteux jusqu'à une profondeur de 300 à 400 m.

La littérature hydrogéologique parle d'une corrélation inverse entre le débit des couches aquifères et la minéralisation des eaux souterraines. A ce

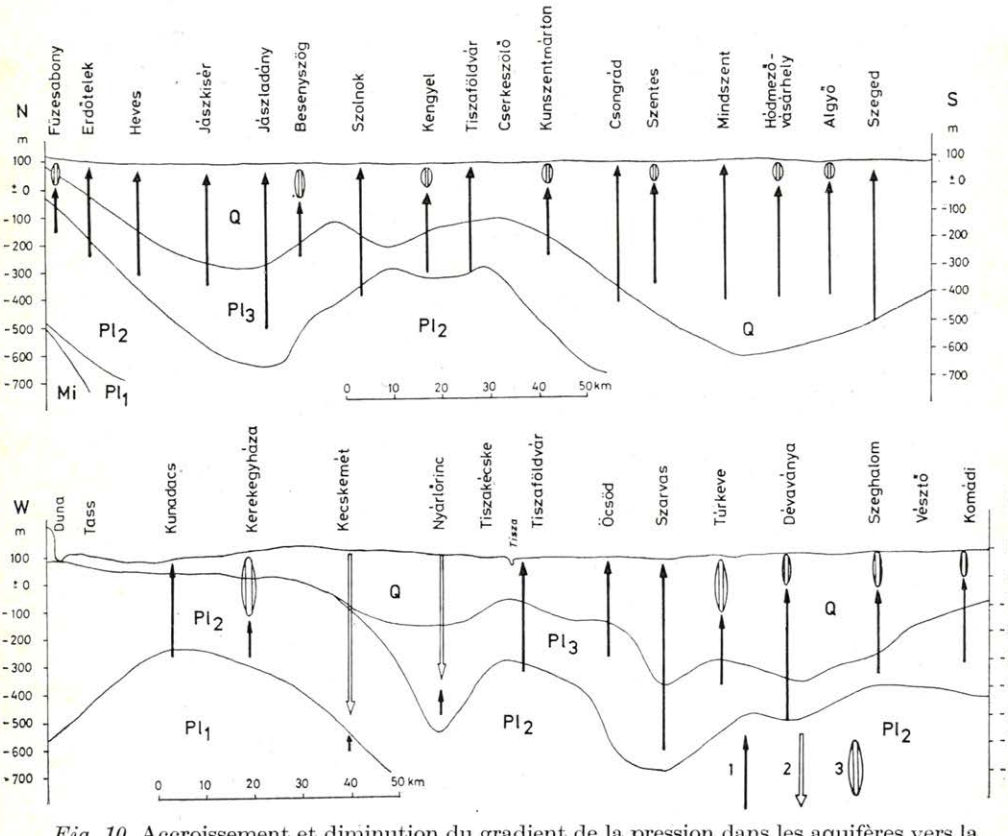

Fig. 10. Accroissement et diminution du gradient de la pression dans les aquifères vers la profondeur. Coupe N–S et W–E

1. Le gradient de pression augmente vers le bas. *2.* Le gradient de pression diminue vers le bas. *3.* Le gradient de pression ne change pas vers le bas

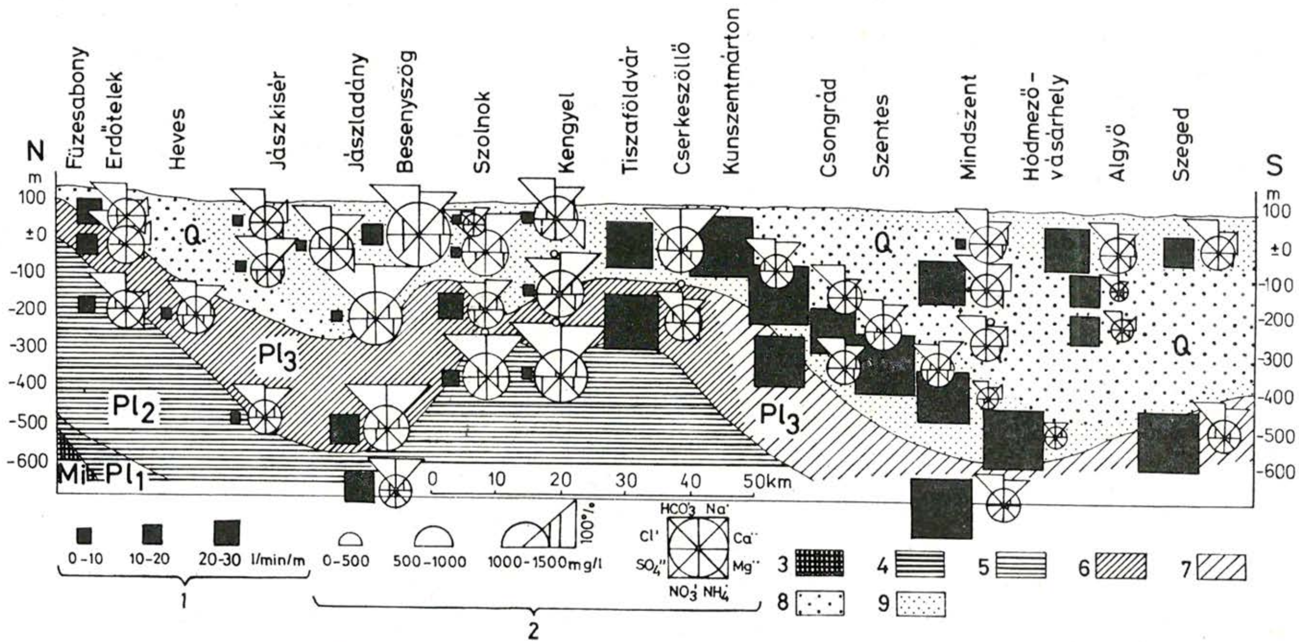

Fig. 11. Coupe géologique N—S indiquant le débit et la composition chimique des eaux souterraines

1. Débit spécifique. *2.* Résultat des analyses chimiques. La dimension des cercles est proportionnée au poids des sels dissous dans l'eau, celle des triangles à l'équivalent pour cent des ions. *3.* Miocène. *4.* Pannonien inférieur. *5.* Pannonien supérieur. *6.* Pliocène supérieur argileux. *7.* Pliocène supérieur sableux. *8.* Quaternaire à grains grossiers. *9.* Quaternaire à grains fins

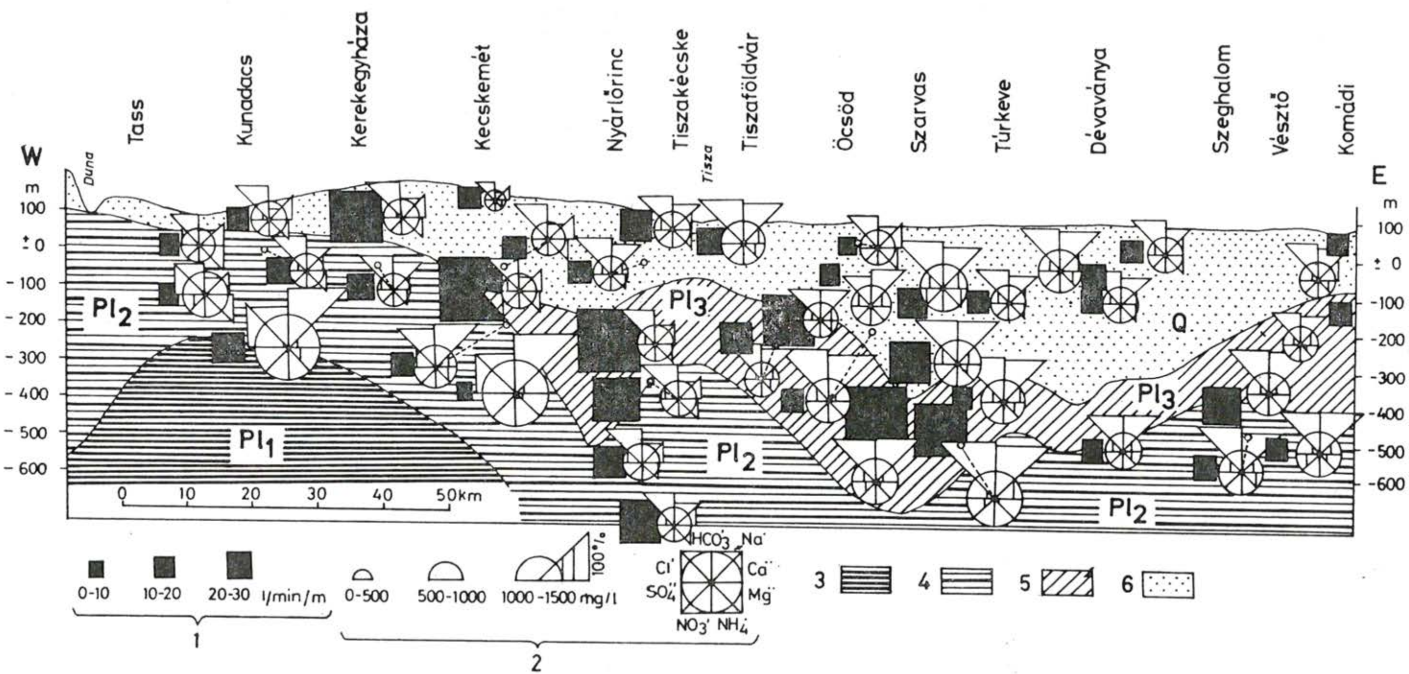

Fig. 12. Coupe géologique W—E indiquant le débit et la composition chimique des eaux souterraines
1. Débit spécifique, *2.* figure de l'analyse chimique, *3.* Pannonien inférieur, *4.* Pannonien supérieur, *5.* Pliocène supérieur, *6.* Quaternaire

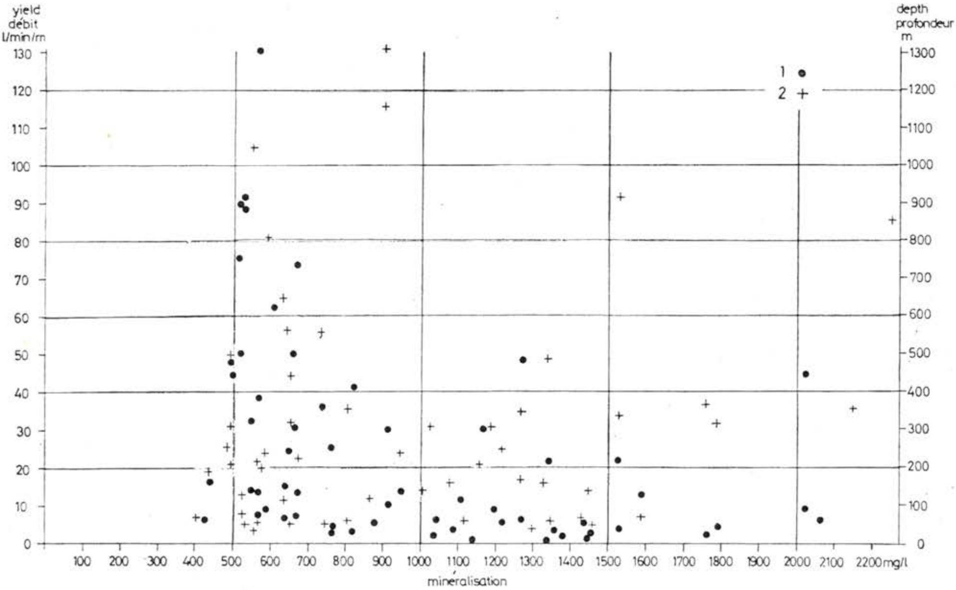

Fig. 13. Rapport entre le débit et la minéralisation *(1)* et la profondeur et la minéralisation *(2)*

sujet nous trouvons des preuves affirmatives à la région de Jászság et de Szeged. Mais les exemples opposés sont aussi nombreux.

C'est du pareil au même, si nous cherchons la corrélation entre la profondeur des couches aquifères et la minéralisation des eaux. Dans beaucoup de cas il y a une augmentation légère dans le poids des sels dissous dans les couches successivement de plus en plus profondes. C'est le cas, où la filtration tend vers le bas. Mais on peut observer des changements inverses. Si la direction de mouvement est opposée, c'est-à-dire l'eau filtre vers le haut, la minéralisation augmente aussi en approchant de la surface.

Il faut parler d'une autre corrélation bien connue ce qui dit qu'il y a une relation stricte entre la granulométrie des couches aquifères et leur débit. Néanmoins, les estimations sur le débit présumé faites sur la base de la grossièreté des grains sont fréquemment loin de la réalité. C'est souvent le cas où la composition granulométrique est mixte. Il est difficile de trouver un coefficient qui caractérise proprement ce mélange et qui peut servir comme base aux calculs. L'incertitude est bien grande surtout dans un bassin remplis de sédiments fluviatiles, où la granulométrie varie dans chaque petite strate à quelques centimètres d'épaisseur.

La figure 14. sert comme exemple combien elle est compliquée la granulométrie de deux couches de sable dans deux puits d'observation. Le champs des courbes granulométriques est tellement vaste que le moyen des valeurs ne nous indique pas la transmissibilité réelle. Tous les calculs basés sur l'une ou l'autre de ces courbes peuvent produire des erreurs considérables.

L'anomalie qu'on observe dans la température des eaux souterraines en Hongrie est bien connue. Vis à vis de l'augmentation de la température par un Celsius grade à chaque 30 à 35 m distance vers la profondeur, qu'on tient

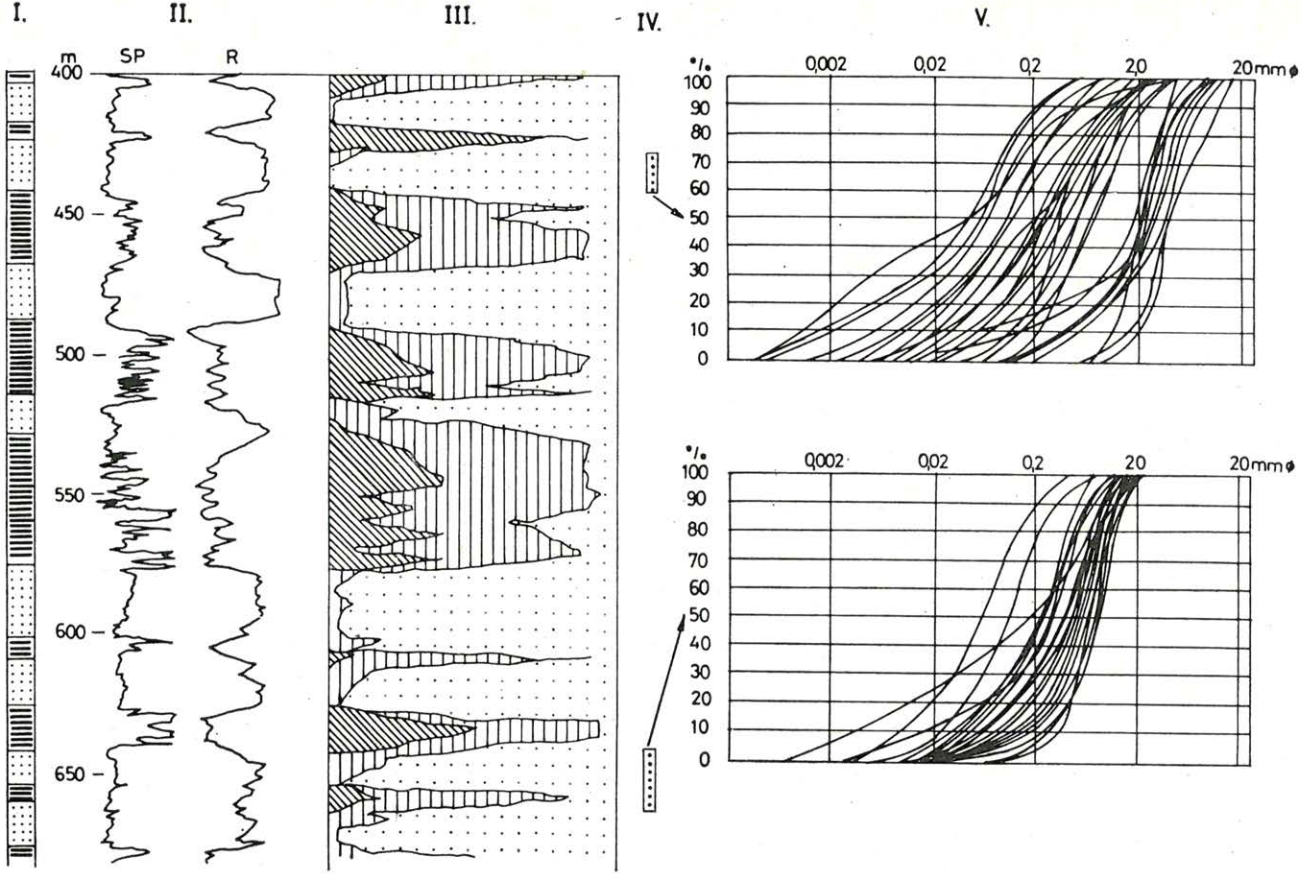

Fig. 14. Hétérogénéité de la composition granulométrique des couches fluviatiles aquifères

I. Profil lithologique du forage. *II.* Log géophysique. *III.* La proportion des différentes fractions de grain. *IV.* Lieu du filtre dans les puits. *V.* Courbes granulométriques des échantillons pris de la couche filtrée

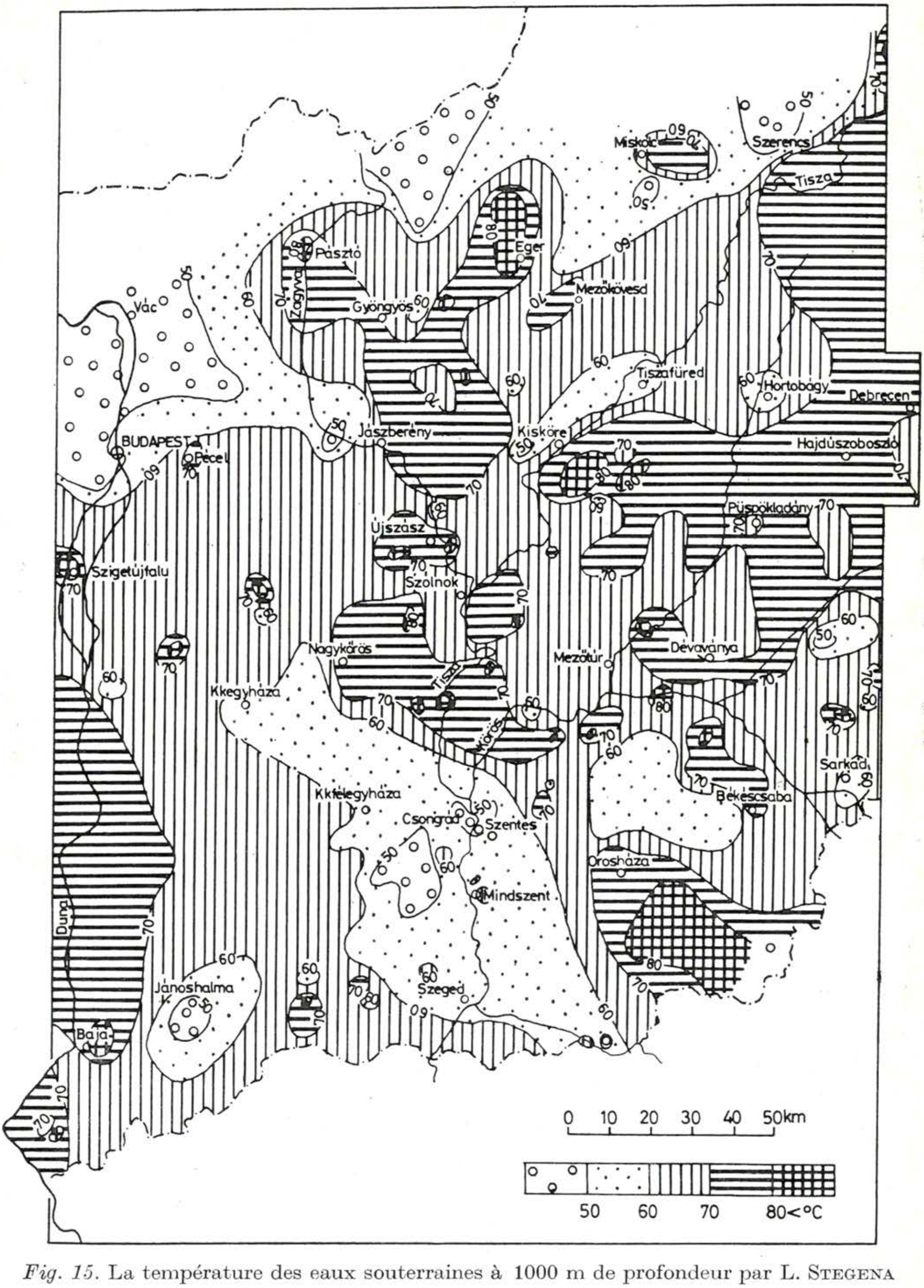

Fig. 15. La température des eaux souterraines à 1000 m de profondeur par L. STEGENA

comme normal en dehors du Bassin Carpathique, cette augmentation se produit en moyenne par chaque 18 m, avançant vers le bas dans la Grande Plaine hongroise. Cette moyenne se compose des valeurs qui vont de 7 m/°C jusqu'à 28 m/°C. Grâce à cette anomalie de la température souterraine nous produisons des eaux thermales 100–120 °C chaudes d'une profondeur environs de 2000 m. Comme ces eaux jaillissent au-dessus du terrain jusqu'à une hauteur de 20 à 30 m, en général, il ne faut pas les mettre au jour avec energie artificielle, il faut simplement les amener et distribuer.

Le gradient géothermique change non seulement horizontalement mais aussi verticalement. C'est pourquoi les cartes qui représentent les conditions géothermiques sur un territoire vaste doivent se rapporter toujours à une profondeur donnée.

On peut démontrer dans les coupes géologiques la variation horizontale et verticale de la température des eaux, qui dépend de la conductibilité thermique des couches, de leur stratification, du voisinage des territoires d'alimentation et des conditions tectoniques. On peut plutôt noter ces anomalies que les expliquer.

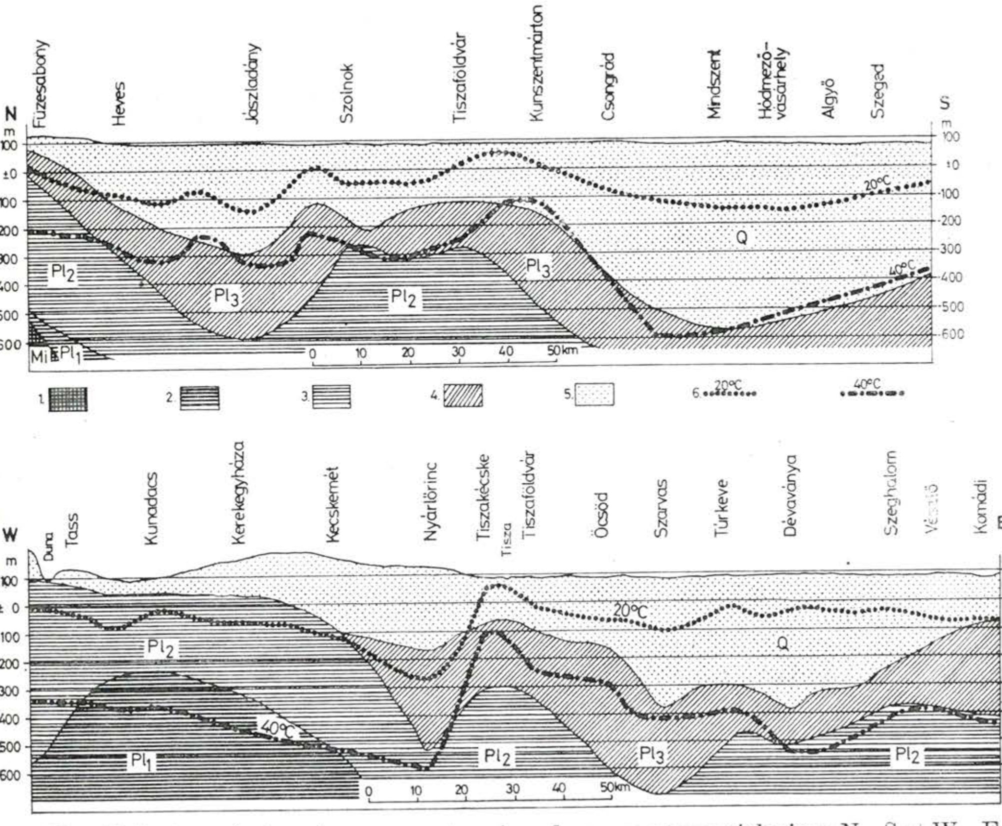

Fig. 16. La température des eaux souterraines dans une coupe géologique N–S et W–E de l'Alföld

1. Miocène, 2. Pannonien inférieur, 3. Pannonien supérieur, 4. Pliocène final, 5. Quaternaire. 6. Isothermes à 20 °C et 40 °C

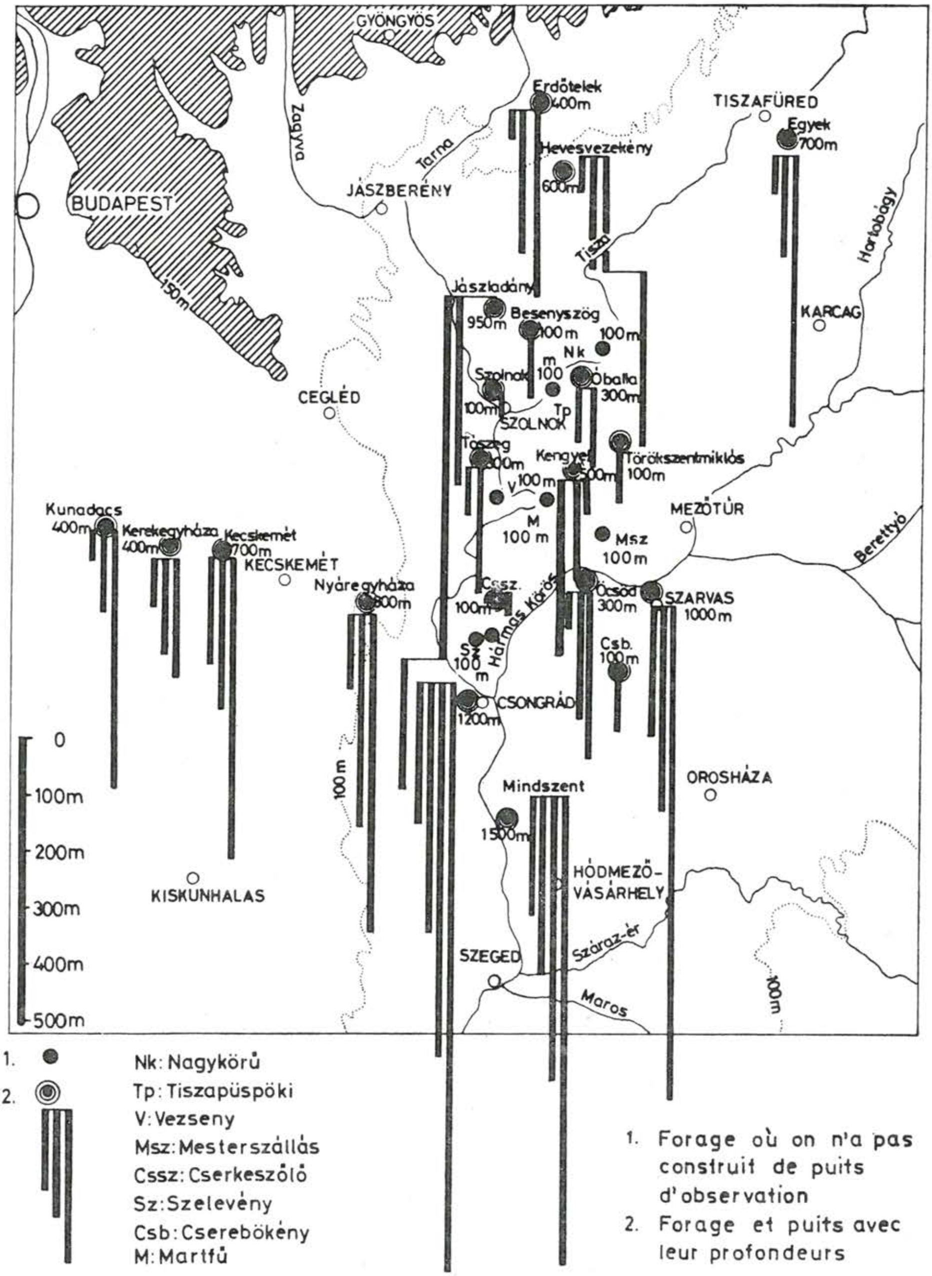

Fig. 17. Distribution des forages fondamentaux de l'Institut Géologique de Hongrie et les puits d'observation construits jusqu'à 1975

Un problème qui reste encore à traiter c'est la quantité profitable des eaux souterraines, les réserves statiques et dynamiques.

Environs 2.3 milliards m³ c'est la quantité d'eau que l'on produit par an par des sources souterraines artésiennes à la Grande Plaine hongroise. Ce qui signifie 51,000 m³ par an à chaque km² et 400 m³/an à chaque habitant. Au début de ce siècle il fonctionnait déjà plus de 300 puits artésiens à la Grande Plaine. Aujourd'hui leur nombre est 41,000. La quantité de l'eau exploitée des sources souterraines au territoire de la Grande Plaine pendant les dernières cinquant années est tout ronde 40 milliards m³. Ce qui correspond à une colonne d'eau d'une hauteur de 90 cm répartie sur le territoire total et à une couverture d'eau de 1.8 cm par an.

Vu cette production, le niveau de la nappe des eaux souterraines n'a baissé considérablement que dans quelques grandes villes, comme par ex. à Debrecen et à Kecskemét. L'affaissement du terrain dû à l'exploitation des eaux souterraines — n'est considérable non plus que dans la ville Debrecen, où il atteignait 40 mm pendant 25 années, 1.5 mm par année.

Ce qu'on peut conclure des susdites c'est que la production de l'eau souterraine — toute grande qu'elle est — n'a pas atteint considérablement jusqu'ici les réserves non renouvellables. La question surgit maintenant, jusqu'à quel point peut on aller avec l'augmentation de la production, combien il est le ravitaillement naturel des eaux souterraines.

Pour les besoins de la cause l'Institut Géologique de Hongrie a exécuté une série de puits d'observation le long de l'axe N-S de l'Alföld et une autre traversant le milieu de la Grande Plaine à l'W-E.

En 1975 le nombre des puits artésiens faits pour l'observation scientifique était 46. La profondeur de ces puits varie entre 30 et 1050 m. Il y a des lieux, où il n'y a qu'une seule aquifère enregistrée, mais il y a beaucoup d'autres,

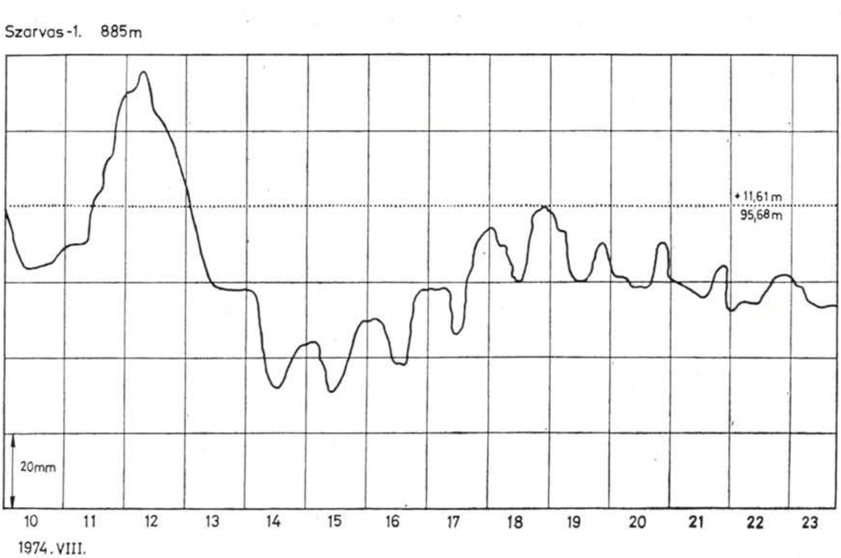

Fig. 18. La fluctuation de la nappe artésienne dirigée par les marées terrestres et atmosphériques

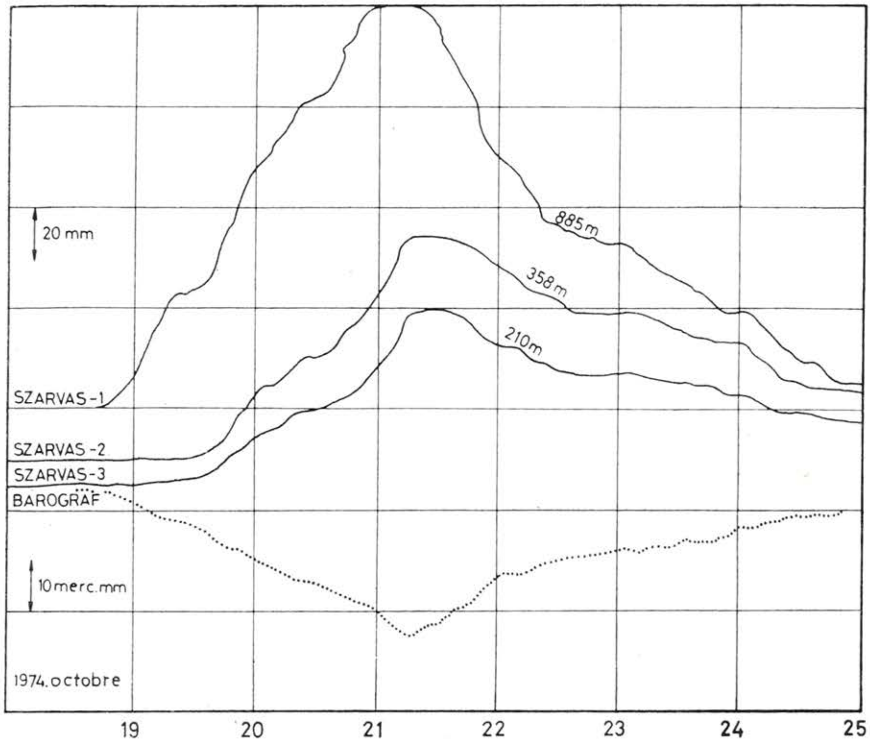

Fig. 19. La fluctuation des niveaux piézométriques des eaux artésiennes de différentes profondeurs par rapport aux oscillations de la pression atmosphérique

où 2, 3 ou 4 aquifères sont observés l'un au-dessous de l'autre par des puits distincts. Sur 15 puits l'observation se continue déjà depuis 9 années et nous avons installé des limnographes automatiques dans 7 puits.

Les résultats trouvés jusqu'ici sont les suivants :

1. Les plus petites fluctuations périodiques du niveau des eaux sont celles causées par le flux et reflux terrestre et la marée de l'atmosphère. Leur dimension est 2 à 3 cm, se répétant régulièrement par 6 et 12 heures. On peut observer ces fluctuations jusqu'à 1100 m en profondeur.

2. Une oscillation régulière, mais pas périodique, est causée par la variation de la pression atmosphérique. Cette pression provoque des fluctuations plus grandes dans les puits profonds (800 à 900 m) et plus petites dans les aquifères situées plus proche de la surface. Leur dimension varie, selon nos données enregistrées jusqu'ici, entre 6 et 20 cm.

Au delà de ces fluctuations régulières de la nappe, il y a une tendence d'augmentation et abaissement saisonnière et multi-annuelle, qui correspond aux périodes climatiques de 7 à 14 années observées dans le Bassin des Carpathes.

La dimension jusqu'ici observée de ces mouvements va jusqu'à 2 et 3 mètres. Ce qui est frappant c'est que les périodes d'augmentation et d'abaissement saisonnières, multi-annuelles se correspondent dans tous les puits, qu'ils soient situés au bord du nord de la Grande Plaine, au milieu de celle-ci ou bien dans la partie du Sud. Le cycle des mouvements et les points de tournant

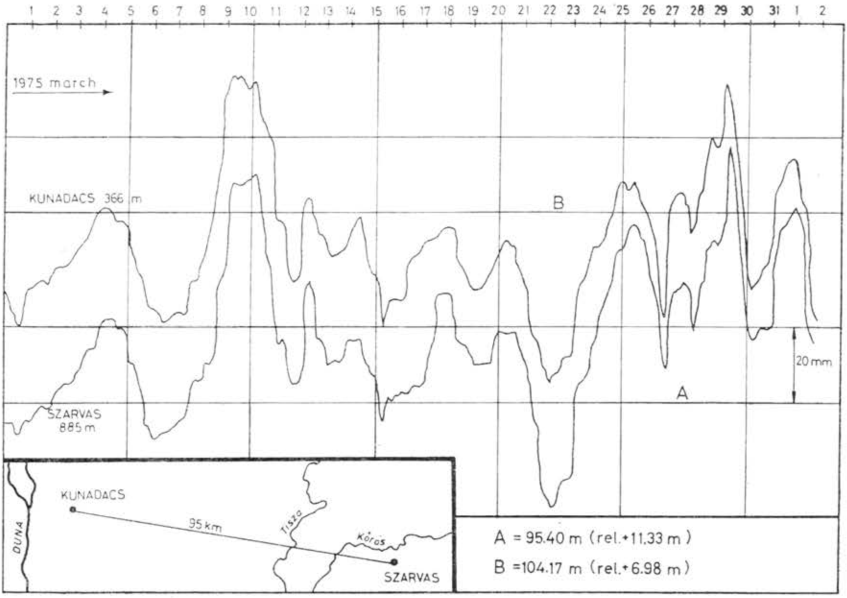

Fig. 20. La fluctuation des niveaux piézométriques causés par la pression atmosphérique dans deux puits situés à grande distance l'un de l'autre et de diverses profondeurs

coïncident dans toutes les profondeurs, jusqu'à 500 m. Des plus grandes profondeurs nous n'avons pas encore de données suffisantes.

Ces phénomènes ne peuvent être expliqués que par le fait que la fluctuation multi-annuelle des nappes artésiennes est dirigée par le climat du Bassin Carpathique entier et pas par les conditions locales. Ensuite, il est évident que l'augmentation ainsi que l'abaissement du niveau des nappes souterraines causée par l'alimentation sont provoquées par la progression de l'impulse de la pression et pas par la filtration de l'eau, qui est beaucoup plus lente. C'est la propagation, l'avancement rapide des ondes de pression dans les couches auquel il est à attribuer que les changement survenus dans l'alimentation se présentent dans les puits partout et environ dans le même temps dans tous les coins de la Grande Plaine et dans tous les profondeurs.

Il faut continuer l'observation dans les puits au moins 10 à 15 années afin que nous puissions déterminer la mesure des fluctuations causées par les cycles climatiques en dégageant des courbes celles causées par les marées et par la pression atmosphérique.

En connaissant les différences du niveau causées par l'alimentation, ainsi que l'épaisseur des couches aquifères, leur composition granulométrique, leur débit spécifique nous possèderont les bases pour le calcul du rechargement naturel des nappes d'eaux.

La vitesse de l'infiltration dans les couches aquifères — qui a une importance primordiale, — peut être étudiée aussi par des isotopes et par la détermination de l'âge absolu des eaux souterraines aux différents lieux et dans des différentes profondeurs. A ce sujet c'est aujourd'hui qu'on fait les premiers pas. Nous entenderons des résultats obtenus jusqu'ici à ce sujet au cours de notre conférence.

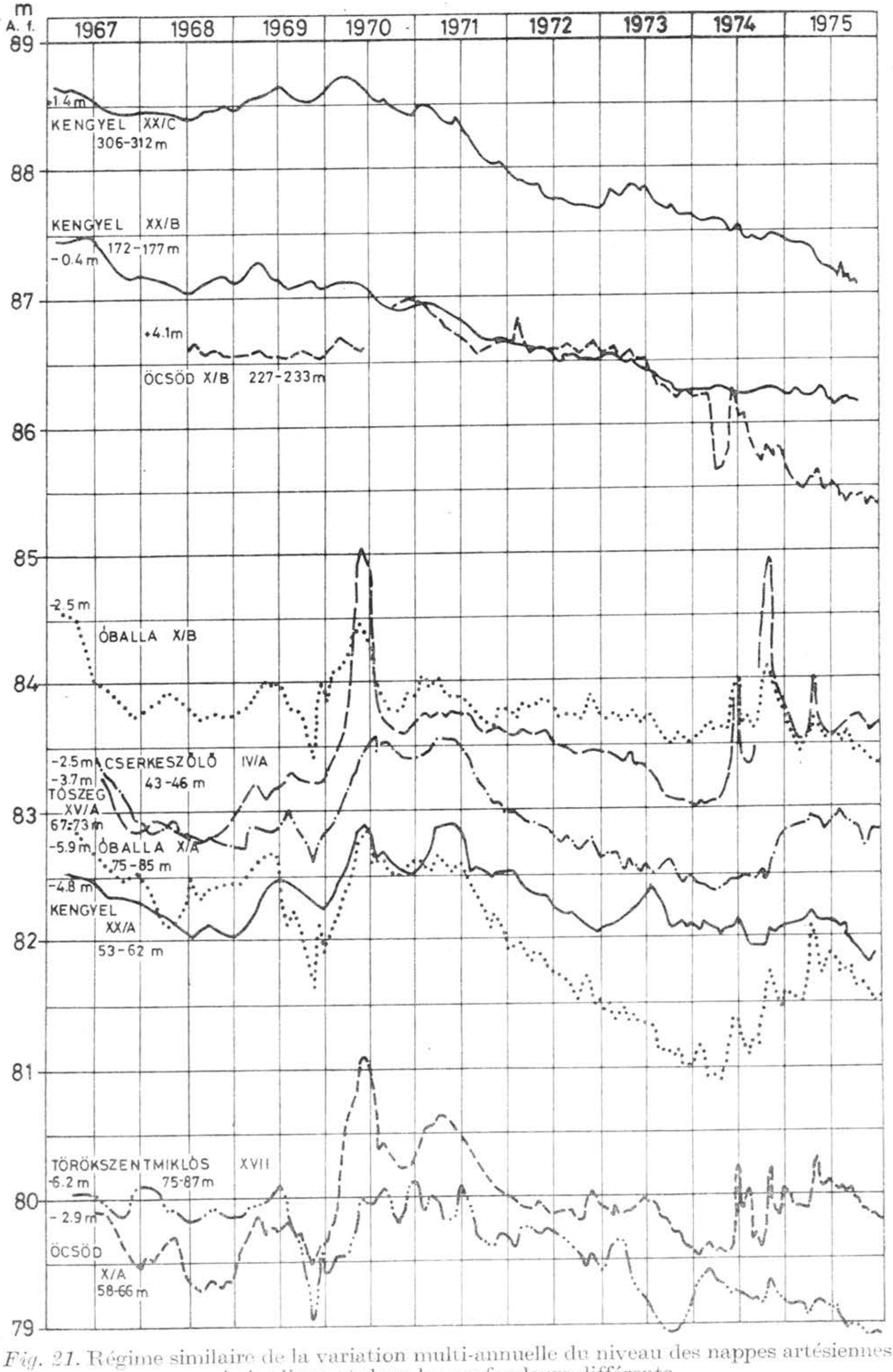

Fig. 21. Régime similaire de la variation multi-annuelle du niveau des nappes artésiennes à des lieux et dans les profondeurs différents

HYDROLOGIC AND HYDROGEOLOGIC ASPECTS OF THE REGENERATION OF LAKE PALIĆ

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Man's life is tightly connected with nature – with his environment. In converting materials and producing waste he interferes with the state of nature, and often drastically changes and pollutes it. This holds especially true of water in nature, in the first place, of rivers and lakes. Lake pollution is a common phenomenon that is constantly getting worse. Damage caused by it will soon be irreparable.

Lake Palić, the subject of the present study, has also become a victim of pollution. The sewerage system of the town of Subotica had for years polluted the waters of the lake, with inevitable and well-known consequences. The lake, once famous for its beach, gradually decayed. The water of the lake slowly changed its colour, turning into an intense green and, on the bottom of the lake fine, dark silt rich in organic matter accumulated. High concentrations of organic matter intensified the bacterial life in the silt. The bacteria used up enormous quantities of oxygen creating thus anaerobic conditions in silt and partly in water above it. The lack of oxygen brought about the death of many kinds of vegetation in the lake, which in its turn brought about the death of fish. At this point it became obvious that it was necessary to treat the lake water. However, it was already too late. Since it was impossible to preserve the lake in its former condition, it was decided to clean it.

After a few years of study a project for the regeneration of the lake was worked out, consisting of the following stages:

1. Emptying of the lake by means of gravitational flow as well as by pumping, with a subsequent removal of the silt.

2. Treatment of the liquid waste being emptied into the lake.

3. Refilling of the lake.

The subject of the present study is the possibility of the refilling of the lake – the actual stage of the regeneration of Lake Palić (Fig. 1).

The conditions under which the refilling of Lake Palić can take place depend upon hydrologic and hydrogeologic properties of the lake and its vicinity. In the present study data accumulated in the course of earlier studies, as well as data gained in the recent years, have been used.

Lake Palić is of eolian origin. Its surface covers 5.6 km². The part of the lake used for recreation takes up 3.9 km². The lake is situated in a depression

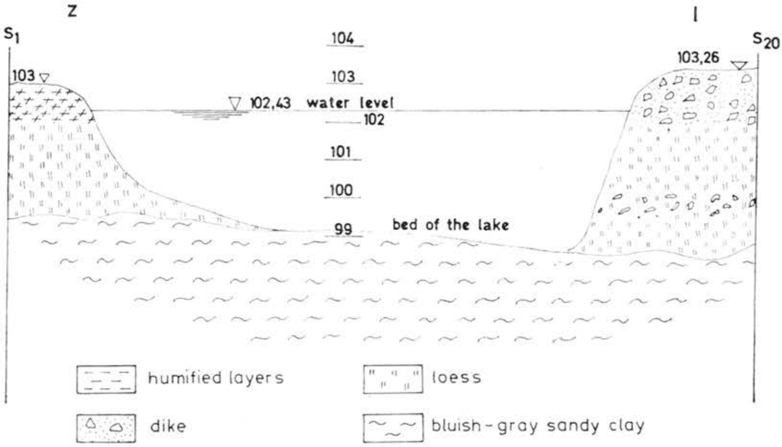

Fig. 1. The hydrogeologic profile of Lake Palić

bordering on sandy soil to the north and on a loess platforms towards Subotica. The bottom of the lake had been formed before sand and loess were deposited, and consists of semi-permeable grayish to bluish clay. The lake was formed in the latest geological period, the Quaternary i.e. Alluvium.

The results obtained by chemical analyses of the lake, phreatic and artesian water, as well as thermal-mineral water, are given in graphic representation. According to the hydrochemical character, the water of Lake Palić could be described as being of carbonate-bicarbonate nitrite type. In recent times, especially after the World War II, the chemical composition of the lake water changed due to anthropogenic influence, to such an extent that at the moment of emptying the lake its water was of sulphurous-nitric-magnesian type.

The relationships among the ions in the lake water can be expressed in the following manner:

$$Na > Mg > Ca$$

$$SO_4 > Cl > HCO_3$$

The lake water was extremely hard, its dH value attaining to 45°, water reaction was alkaline, as average pH being 8.5.

Due to the presence of humic acids, the colour of the water was yellowish-green.

Since the amount of dissolved solids was about 2000 mg/l, the lake's water could be considered as mineral water. The differences in chemical composition of the water (Fig. 2) depended on the meteorological conditions and on the amount of the liquid waste being drained into the lake.

Water mineralization originated from surface accumulation. The water balance is characterized by a yearly deficit of about 220 mm, in extreme cases even by 460 mm. With the exclusion of other factors contributing to the water balance, the deficit by loss from the water surface is expected to come to 25% only. Major characteristics of the water balance can be seen in Fig. 3.

The level of the phreatic water is, as a rule, dependent on climatic conditions. In accordance with the water balance influenced by climatic conditions, the ground-water table rises during the winter period (December to March)

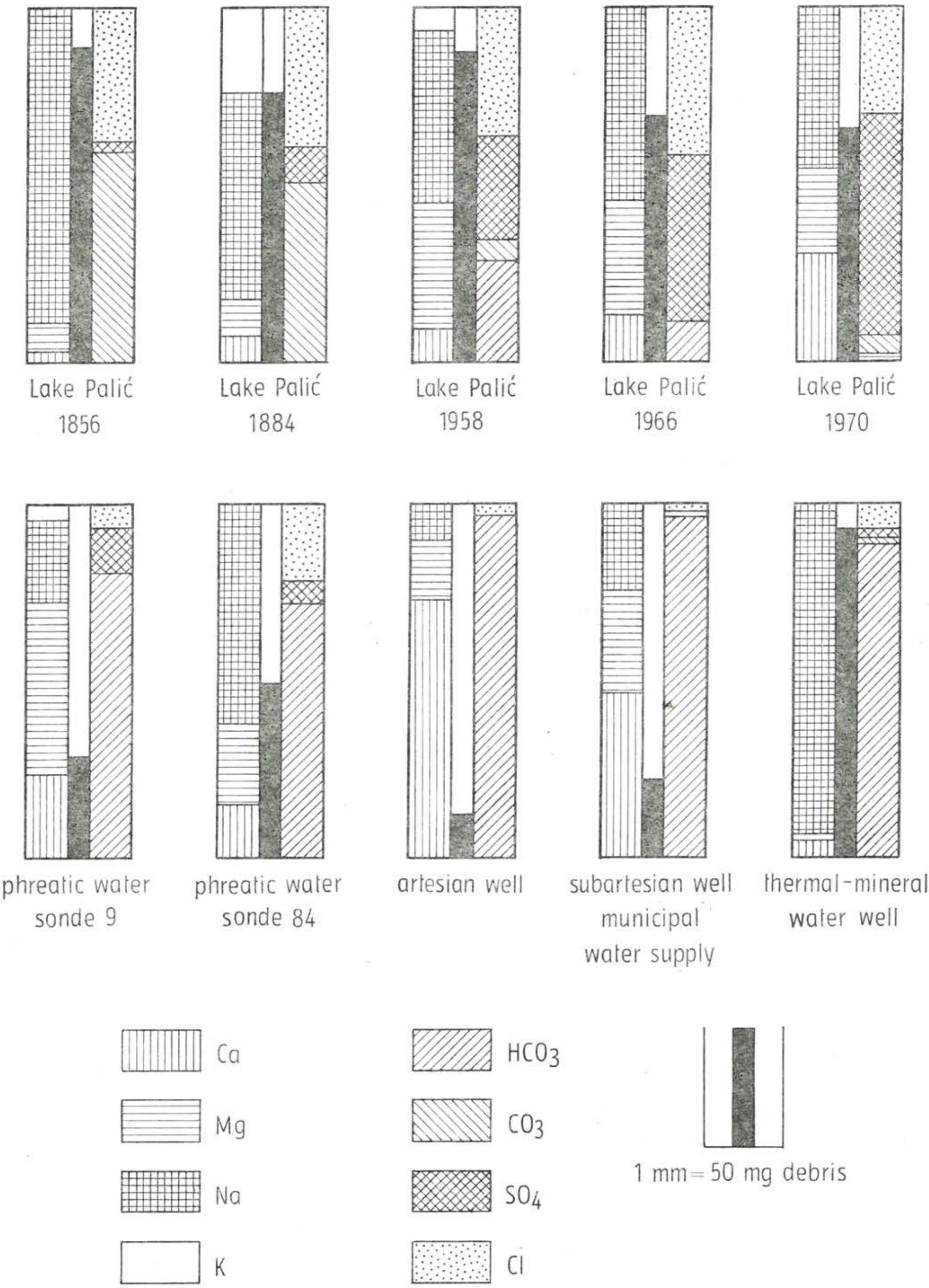

Fig. 2. Results of the chemical analyses of lake water

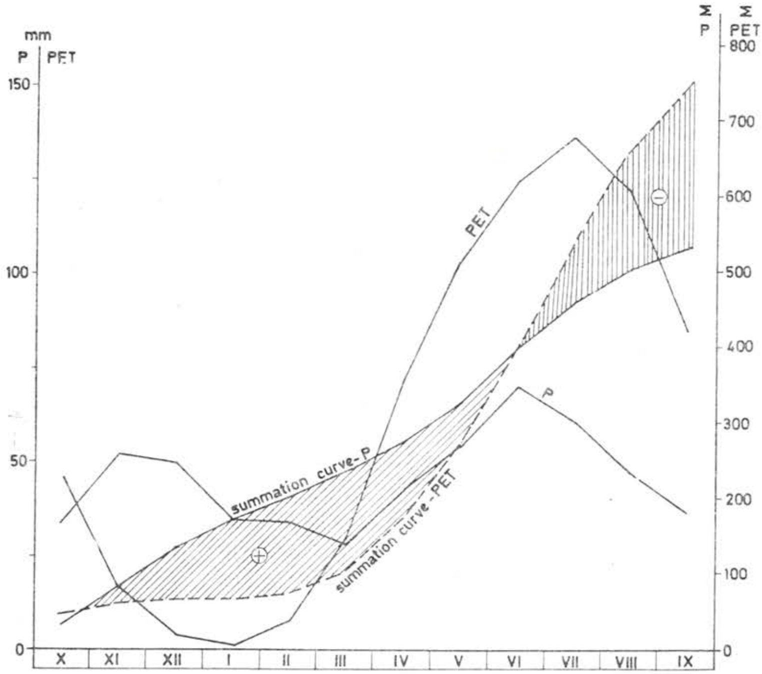

Fig. 3. Rainfall (*P*) and potential evapotranspiration (Turc) (*PET*). Average values for the period 1950/51 to 1973/74

and falls during the period of vegetation (April to September), as can be seen in Fig. 4 (two adjacent observation wells). The average level of the phreatic water table in the vicinity of the lake is 102.00 m. The planned level of the lake will be at the same height.

A significant relation has been found between climatic conditions and the phreatic water level.

This correlation renders possible to determine the influence of the lake water level upon the level of phreatic water, i.e. their interaction. The phreatic water level shows a yearly fluctuation of about 100 cm. The rise of the lake level in winter period is greater than its fall in the vegetation period, a fact that points to a constant influx of the phreatic water to the lake, with the involvement of a broad region around it.

The speed of the underground phreatic water flow is very low (about 8 cm per day). Most water accumulated in the lake through underground flow comes from the sandy soils lying north of the lake. On the basis of observations performed it can by far be expected that the flow into the lake will reduce the fall of the lake level by about 20%. Approximately 40 mm of the water deficit will be regained by phreatic water influx in a year of average weather conditions.

This correlation shows that in winter periods the rise of the lake level exceeds that brought about by the amount of rain fallen during the same period. This rise, however, is correlable with the quantity of the rainfall (Fig. 6).

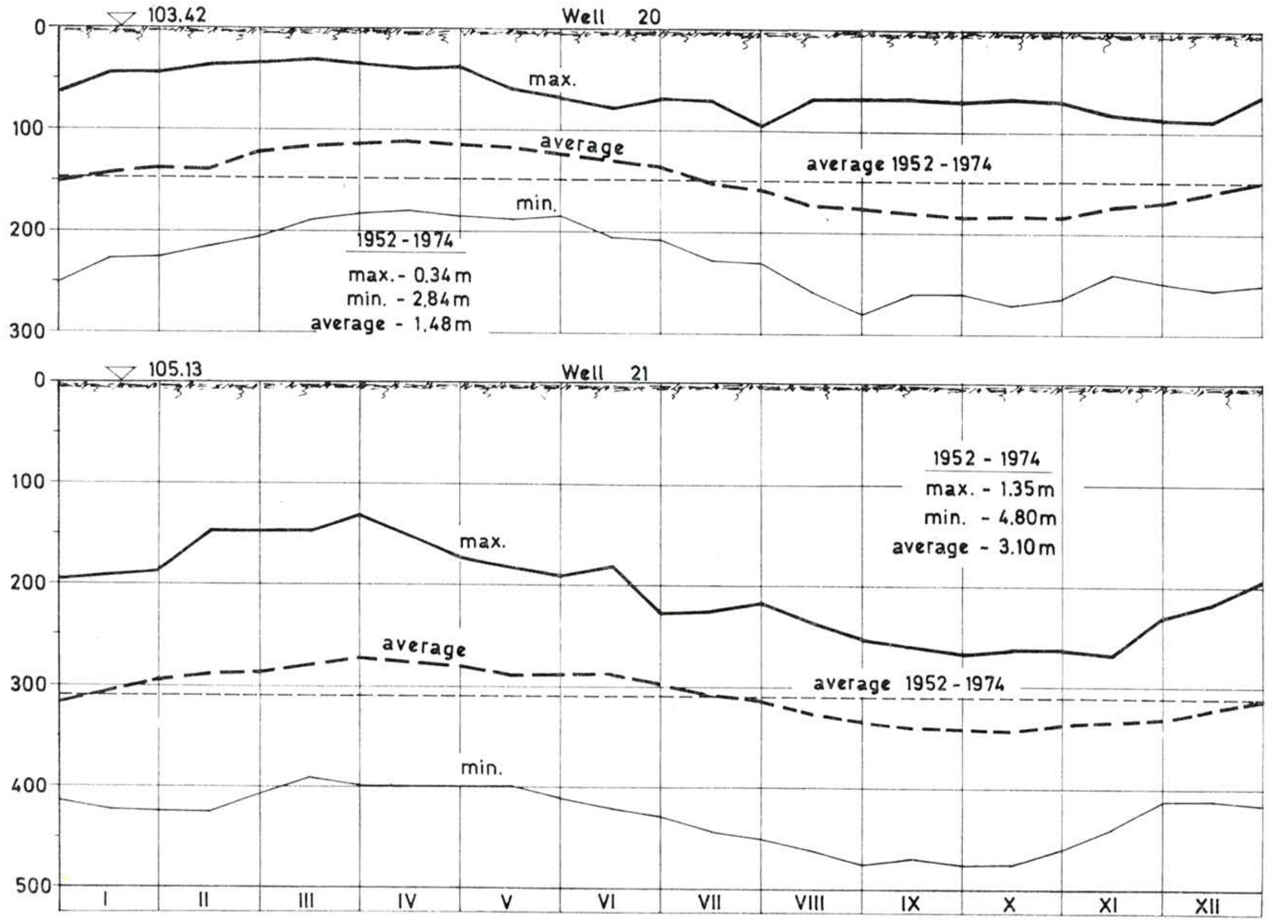

Fig. 4. Extreme and average phreatic water levels between 1952 and 1974, Lake Palić (two localities)

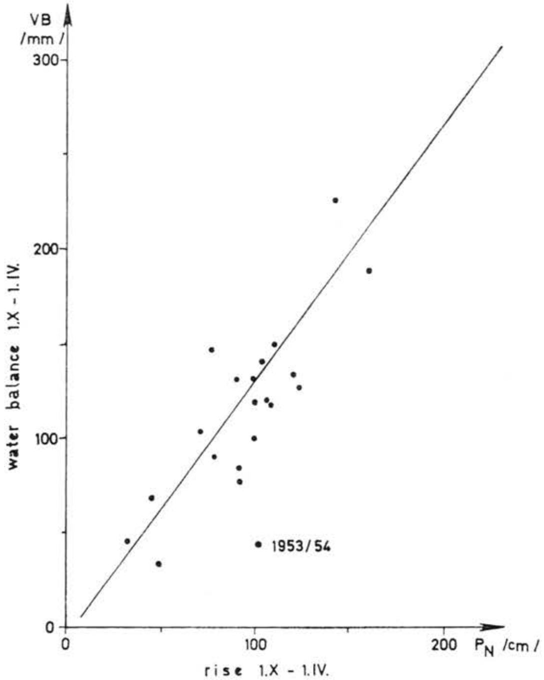

Fig. 5. Correlation of the average phreatic water table rise with water balance from 1952 to 1974 (October to April), Lake Palić

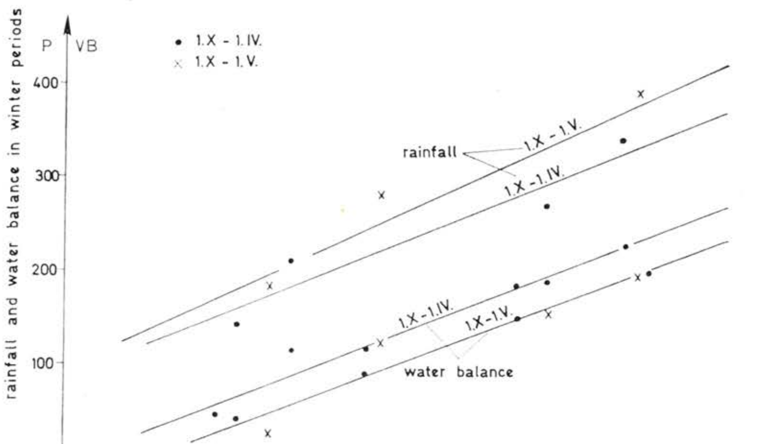

Fig. 6. The rise of the lake level in response to rainfall and water balance in winter

This has helped to determine the following values:

- $PJ = -270 + 3P$ (where PJ is the rise of the lake level in relation to winter rainfalls (December to April, Fig. 7), and P represents the total rainfall in that period in mm);
- The immediate watershed of Lake Palić is 70 km^2;
- The coefficient of runoff (K_0) from the watershed in winter periods can be as little as 0.04 as corresponding to the minimum seasonal rainfall and attaining to a maximum 0.20, as shown in Fig. 8.

It is possible to refill Lake Palić with water in several ways. The quickest way would be feasible by catching up waste waters from the town of Subotica. In a year abundant in rainfall (frequency of 1%), the average influx by liquid

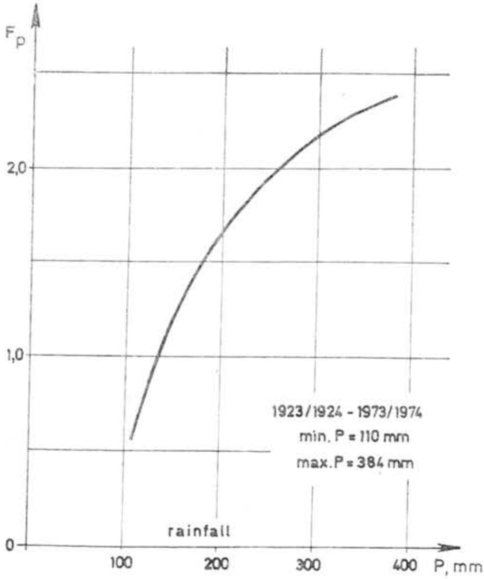

Fig. 7. Factor of the rise of the lake level (F_p) and its dependence on rainfall in winter periods (December to April) (P)

waste would be 0.25 m^3/sec. This would take about 450 days to fill up the whole lake (560 ha), or 325 days for the recreation part only (387 ha). In a period with medium rainfall this would take about 600 days for the whole lake, and about 420 days to fill up that part of the lake reserved for recreation.

To compensate the loss of water in summer months, water of highest quality could be supplied from artesian wells. Eight wells would be needed for the whole lake, and six to feed with water the recreation part only. In an extremely dry year 16 i.e. 12 wells would be needed, respectively.

Climatic and hydrologic factors alone are not sufficient to enable the refilling of the lake. Under most favourable conditions similar to those experienced during three and a half hydrologic years prior to the spring of 1956,

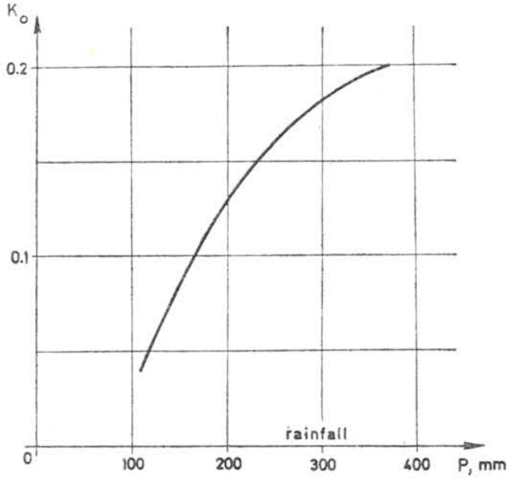

Fig. 8. The coefficient of runoff (K_0) of the Lake Palić catchment area in correlation with the total rainfall in the winter period (October to April) (P)

it would just take three and a half years to fill up the basin of lake. Under average conditions this would last as long as ten years.

The utilization of thermal-mineral water is also unfeasible because of the unfavourable conditions under which this type of water would be available.

The optimum solution for the problem of the refilling with water of Lake Palić is supposed to be found in supplying water from the Danube river. However, in order to have a better economic justification for such a project, it ought to be included the solution of the problem of water supply not only for Lake Palić, but also for the whole northern Bačka.

ON THE THERMAL WATERS OF THE GIPPSLAND BASIN AND THE PROBLEMS ASSOCIATED WITH THE STUDY OF HIGH TEMPERATURE AQUIFERS

LES EAUX THERMALES DU BASSIN DE GIPPSLAND ET LE PROBLÈME SE POSANT LORS DE L'ÉTUDE D'AQUIFÈRES A TEMPÉRATURE ÉLEVÉE

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RÉSUMÉ

Les nappes thermales superficielles des districts de lignite du Gippsland sont générées par le flux géothermique, à cause de l'effet isolant du lignite épais qui recouvre les sables aquifères plus conductifs.

Les mesurages expérimentaux publiés des conductivités thermiques des sables saturés, comparés avec ceux du lignite, indiquent un rapport d'à peu près quatre à un. Ceci est en harmonie avec les mesurages de chantier des rampes thermiques. Les augmentations locales dans le flux géothermique et la perméabilité aquifère doivent aussi être considérées dans l'étude de ce problème.

La théorie moderne des nappes de surface a été principalement désignée à expliquer des régimes de basse température, et assume que le volume et viscosité relatifs de l'eau ne varient que d'une façon insignifiante. Ces présomptions ne peuvent être appliquées aux agents aquifères thermiques communément trouvés dans des bassins sédimentaires, et des modèles mathématiques plus rigoureux sont requis si l'on veut expliquer les vastes réserves de nappes superficielles et d'énergie thermique de basse teneur concernées dans ces systèmes.

The development of the Gippsland sedimentary basin as with all the post-Palaeozoic basins located along the southern margin of the Australian continent, commenced with the formation and subsequent infilling of an extensive system of rifts that heralded the breakup of Gondwanaland. Sedimentation, which is predominantly non-marine in the more easterly rifts, ceased by the Albian.

A new sedimentary regime commenced in subsidiary basins that formed across the main zone of crustal weakness as the result of an irregular pattern of downwarping of the southern continental margin. Horsts of Mesozoic and Palaeozoic sediments isolate the individual basins.

The Gippsland Basin is the most easterly of these and occupies 6500 km^2 of the mainland and extends offshore beneath Bass Strait for over another 45,000 km^2. The northern limits of this basin are bounded and partly underlain by Palaeozoic sediments and granites while the central part of the basin is underlain by the arkosic rocks that occupy the early Mesozoic graben (Fig. 1).

Broadly, the sedimentary sequence consists of a deltaic complex of Upper Cretaceous and Tertiary terrigenous sands, clays and incalated volcanics reaching a thickness of 4000 m offshore which are overlain by a thick sequence of Tertiary limestones and marls.

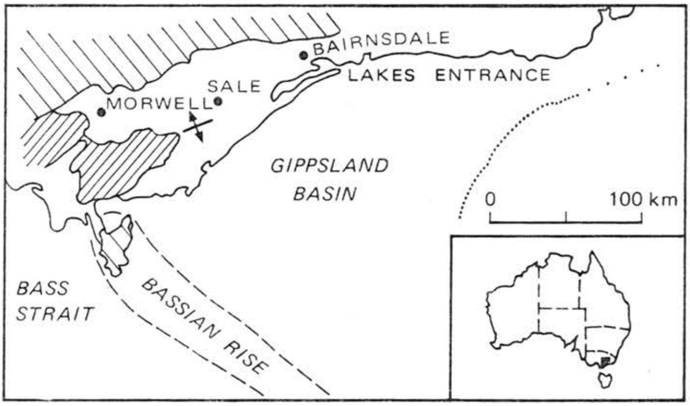


Fig. 1

Onshore the Upper Cretaceous is poorly developed, the first sequence of any consequence consisting of terrestrial sediments with associated volcanics considered to be of Eocene age. This is overlain by a suite of sands, clays and thick brown coal deposits referred to as the Latrobe Valley Coal Measures. The higher temperature waters in question occur within and beneath this formation. The marine sequence is well developed in some onshore areas but is not discussed here.

The stratigraphy and structure of the brown coal fields have been described by Thomas and Baragwanath [29], Gloe [8] and others [9], while the offshore sequence has attracted considerable interest because of the discovery of oil and gas within it [22].

The Latrobe Valley Coal Measures occupy an east-west trending graben modified by a number of anticlinal, synclinal and monoclinal features which play an important role in the mineable location of thicker sequence of coal (Fig. 2). The ancient zone of crustal weakness that marks the northern limit of the Mesozoic rift system also passes somewhere along the centre of Latrobe Valley although the limited deep drilling in the area indicates that this feature could be complex. In all, the area is structurally complicated by the superposition of a number of intersecting regional trends and is a zone of ancient volcanicity as well as a tectonically active area.

The major coal seams recognised are the Yallourn Seam – 90 m thick, the Morwell Seams – up to 150 m thick, and the older Traralgon Seam which attained 150 m in thickness although the seams do not reach their maximum thickness in the same areas.

Pressure waters are located in interseam beds of coarse quartz sand and the piezometric levels exhibited are such as to necessitate the completion of a complex dewatering scheme in the open cut areas.

The water levels in the vicinity of the Morwell mine have been lowered by some 100 metres by allowing bores within the open cut to initially flow at rates totalling 1 m^3/sec, the water temperature averaging about 50 °C. The water has a low salinity of around 400 ppm and is predominantly a soft, sodium chloride-bicarbonate type having a low pH, and a relatively high ferrous iron content and contains traces of CO_2 and H_2S. The aquifer sands,

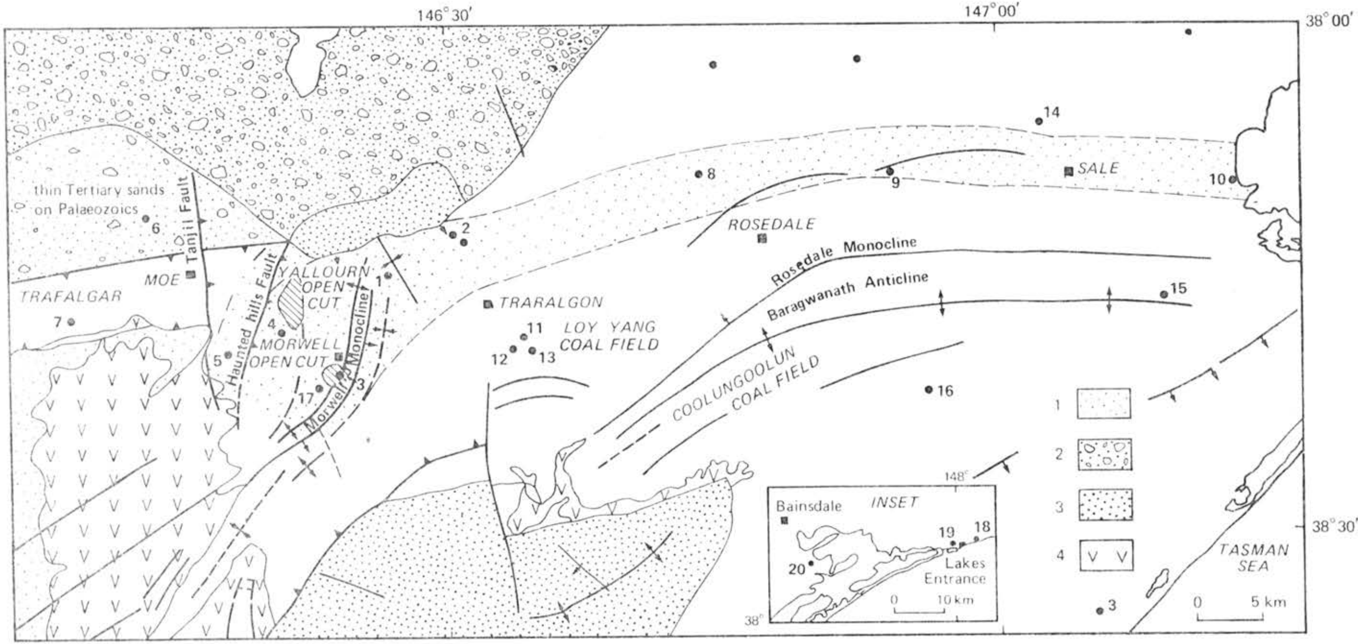


Fig. 2. Bore locations and coal fields

1. Thermal groundwaters, *2.* palaeozoic outcrop, *3.* mesozoic outcrop, *4.* older basalts, *5.* bore locations (see Table 1)

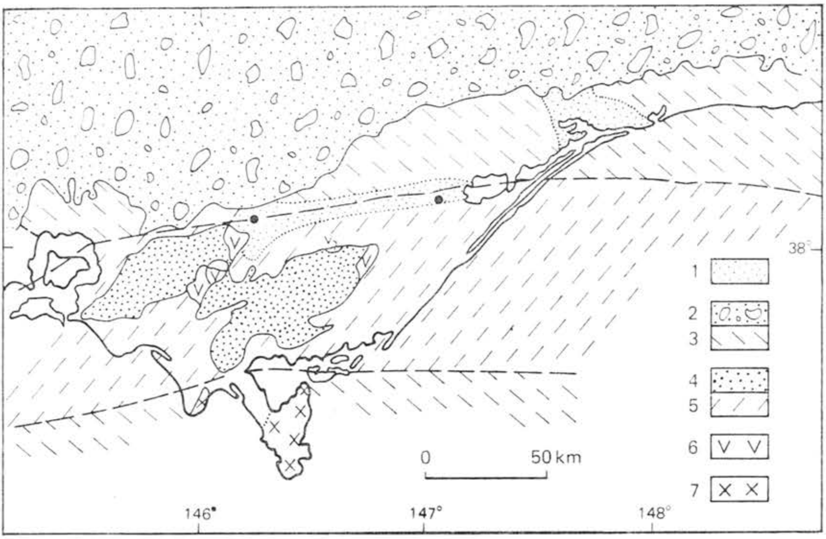


Fig. 3. Outline of the geology of the Gippsland Basin

1. Thermal waters, *2.* palaeozoic outcrop, *3.* palaeozoic basement, *4.* mesozoic outcrop, *5.* mesozoic basement, *6.* older volcanics, *7.* granite

particularly those beneath the Morwell 2 Seam, are coarse grained and have a high transmissibility of between 0.02 and 0.08 m^2/sec. [5, 30].

The location of thermal waters as indicated by recent drilling is given on Figure 3 and location, depth, temperature and bedrock type are indicated. A summary of measurements are presented in Table 1.

The origin of high thermal gradients in Gippsland

High temperature groundwaters have been encountered in two main areas in the on-shore Gippsland basin. An extensive area of high temperature aquifers is located associated with the Morwell brown coal fields and elsewhere in the Latrobe Valley graben, and another limited area occurs some 150 km to the east along the northeastern margin of the basin near Bairnsdale and Lakes Entrace (Fig. 1).

In both areas thermal gradients of 1 °C per 10 metres are common while gradients of 1 °C per 5 metres are reached within the coal measures of the Morwell area. The average gradient expected in a sedimentary sequence is 1 °C per 20 to 30 m depth of burial.

The association of high temperature groundwaters and brown coal deposits in the Latrobe Valley and the well known observation that oxidation of brown coal occurs rapidly on exposure to the atmosphere lead Gloe [8] to suggest that partial oxidation of the coal could be initiated by the presence of dissolved oxygen in the surface waters in the intake areas of the aquifer system thus heating the aquifer waters.

However, since the dissolved oxygen content of water at the local ambient temperatures reaches saturation at about 60 parts per million, the potential for heat generation is very slight. It is more likely that the dissolved oxygen would rapidly combine with iron and enter solution as ferrous iron.

Other explanations involving chemical reactions of minor constituents such as pyrite can also be discounted because of very large energy involved in the continuous generation of hot waters.

Another answer has been suggested by STUTZER [28] following the study of a similar phenomena observed in the Bohemian brown coal fields. He encluded "that heat of coal deposits is generated by coal formation" a contention he supports with some theoretical estimates of heat released by chemical reactions and carefully measured temperature gradients within the coal bodies studied.

This process appears to be contrary to what might be expected from the well known relationship between coal rank and regional thermal history. KARWEIL [11], for example, has stated that the temperature gradient is the major factor in coal metamorphism, time is the second most important factor while pressure is of less importance. A similar conclusion was reached by BROOKS [7] concerning temperature and pressure on the rank of coals within the offshore Gippsland section.

Although relatively high temperatures are recorded in the aquifers of Gippsland there is no evidence to suggest that the coal is still undergoing metamorphism particularly in view of the shallow depth of burial and generally low temperature regime that has prevailed for at least some millions of years.

The magnitude of the thermal energy being transported from the area via the high temperature water within the aquifer systems is such that a significant depletion of the coal deposits would occur within a short period of geological time if this energy was being generated by chemical reactions alone.

It also should be noted that relatively lower thermal gradients have been observed in other areas, for example Loy Yang and Coolungoolun, where thick deposits of coal are located, while in east Gippsland high thermal gradients are recorded where only thin coal beds are involved. In each case however the close overall relationship of depth of burial and temperature is still maintained.

Thus there is no evidence that a uniform chemical change is occurring within the coal which is capable of generating heat to be transferred to the aquifers and this hypothesis must be rejected.

Geological considerations suggest that in relatively low grade regimes, the extent of coalification reflects the thermal history of the coal body rather than the reverse, and it is considered that apart from a brief initial period of biogenetic alteration the formation of brown coal is basically an endothermic process.

The effect of variations in thermal conductivity

It is proposed that the thermal gradients recorded by STUTZER and the thermal waters of the Gippsland coal fields can be explained in terms of variations between the thermal conductivity of the coals and the sediments provided that allowance is made for the possibility of changes in the earth's thermal flux as a result of structural and bedrock differences.

The average thermal gradient in Bohemia as quoted by STUTZER [28] is 1 °C per 20 metres depth. Exact temperature measurements in the brown coal fields of northern Bohemia show this gradient down to the coal, the gradient then increases to 1 °C per 5.25 m through the coal beds. Similarly the temperature gradient above the coal beds at Karlsbad are stated to be 1 °C per 12.2 metres, increasing to 1 °C per 5.03 metres in the coal. It was stated that gradients within coal beds of increasing rank are considerably less.

Similar gradients occur in the Latrobe Valley coal fields across the coal intersections. It should be noted that values given in Table 1 are the gradients from the surface, to the bedrock or to the major aquifer sand and therefore represent the overall gradient.

The thermal gradients within the coal measures at Morwell have been described as highly linear in an S.E.C. Rept. DD73. [24]. In this the relationship between temperature and depth, which has a regression coefficient of 0.998, is given by $T=12.8+0.186\ D$. T=Temp °C, D=depth in m.

On inspection it can be assumed that the 12.8 °C represents the mean annual surface temperature (since the mean annual air temperature at Yallourn is 13.6 °C) [21]. So the thermal gradient through the coal as given by this relationship is 1 °C per 5.38 metres.

The marked similarity between the thermal gradients of the brown coals of Bohemia and Victoria is in fact no more than a measure of the thermal conductivity of this type of coal. That is, brown coal is only one quarter as conductive to heat energy as the average sedimentary rocks sequence which exhibit a thermal gradient of 1 °C per 20 to 30 metres depth of burial [4, 13].

This conclusion is borne out by the limited experimental data available.

Somerton [25] gives the thermal conductivity of saturated unconsolidated sand as $\lambda_t=0.0043$ cal/cm/sec °C which is in agreement with values given by Luikov.

Allardice [1] arrived at a figure of between $\lambda_t=0.001$ and 0.0008 cal/cm/sec °C for the Gippsland brown coal, the variation being dependent on moisture content. This figure is also supported by measurements on extruded Morwell brown coal carried out at the Chemical Engineering School of Melbourne University where a figure of $\lambda_t=0.523$ W/m K was obtained for coal with a moisture content of 63% (B. Stanmore pers. comm.).

Stutzer [28] also presents a similar value with his statement that "the ability of coal to conduct heat is very small, its absolute value is 0.00063 cal/sec/cm °C" but no description of the type of coal is given. His observation that coal of higher rank contains "less heat" is readily explained in this context since the higher ranked denser coal has a greater thermal conductivity and therefore exhibits a lower thermal gradient through it.

It is proposed that the presence of thermal waters in the vicinity of the coal fields is due to the insulating effect of the coal beds on local areas of high geothermal flux, the presence of artesian aquifer systems within the coal sequence modifying the simple thermal conductivity relationships.

Thermal gradients through a sequence of brown coals and aquifer sands

The presence of aquifer systems beneath or within a sequence of insulating brown coal beds being subject to a geothermal flux averaging 1.5 micro cal/cm^2/sec, [13] results in the absorption of part of this energy which is subsequently removed from the area by the aquifer.

Table 1

Thermal gradients in Gippsland

Average annual surface temperature 13 °C

	Index No. (Fig. 2.)	Location or bore name	Temperature °C	Depth in metres	Average gradient per 1 °C	Bedrock type		Remarks
West Gippsland Coalfields	1	Maryvale 1	55 – 69	470 – 500	10 m	Aquifer	Thin Mesozoic on Palaeozoic	Water temperature
	2	Boola Boola 1	64	590	9	Aquifer	Mesozoic	Water temperature
	3	Morwell Open Cut	M1 48 M2 52	300 320	8	Logged	Palaeozoic	Not to bottom observation bore
	4	Narracan 3294	40	300	11	Logged	Palaeozoic ?	Observation bore
	5	Narracan 3283	21	165	21	Logged	Mesozoic	Observation bore
	6	Wesbury 8	21	125	15	Logged	Palaeozoic	Observation bore
	7	Trafalgar 40	21	200	25	Logged	Mesozoic	Observation bore
	8	Toongabbie Sth 31	65.6	715	14	Aquifer	Palaeozoic	Water temperature
	9	Dennison 53	52.2	863	21	Bedrock	Mesozoic	Observation bore
	10	Sale 13	62.5	1050	21	Aquifer	Mesozoic	Water temperature
	11	Loy Yang 1177	31	156	8	Aquifer	Mesozoic	Water temperature
	12	Loy Yang 1151	28	146	10		Mesozoic	Water temperature

	13	Loy Yang 1156	28	149	9	Aquifer	Mesozoic	Water temperature
	14	Sale 16	48	480	14	Logged	Mesozoic	
	15	Glencoe Sth 8001	23	506	50	Aquifer	Mesozoic	Water bore
	16	Holey Plain 66	23	123	12	Aquifer	Mesozoic	Water bore
	17	Hazelwood 1228	38	127	5	Aquifer	Mesozoic	Water bore (blocked) Near Morwell open cut
East Gippsland	18	Colquhoun 1	35	308	14	Aquifer	Granite	Water bore
	19	Colquhoun 12	37	422	17	Aquifer	Granite	Water bore
	20	Bairnsdale 6	58	570	13	Aquifer	Granite	Water bore
World average		Great Artesian Basin			30 6		Sediments, crystalline basement	
		Winnindoo 45 Dennison 54		413 328			Palaeozoic	Average temperature
					20 to 30		Sediments	
		Bohemia		300	10 – 20 5		Sediments, brown coal	
		Morwell			15 – 20 5		Sediments, brown coal	

Thus the temperature gradients will be dependent upon the hydraulic characteristics of the aquifer sands as well as the thermal conductivity of the sediments and coals making up the sedimentary pile.

Continuous temperature and delta temperature logs were made on wells penetrating the east Gippsland sequence. The equipment consisted of a standard Gearhart Owen wire-line temperature probe in which a thermistor generated pulse signal varies in frequency, in response to temperature variations. Standard chart recorders give a readable accuracy of about 0.5 °C. The bores were also logged with γ ray and electric probes to delineate brown coal and sand beds. The examples in Fig. 4 include the results of logging from two types of wells – groundwater observation bores and mud filled bores drilled further out in the basin and logged soon after completion.

Fig. 4b shows the temperature gradient exhibited by Narracan 3294, a steel lined groundwater observation bore within the coal fields.

The thermal gradient through the combined Morwell 1B and Morwell 2 seams is approximately 1 °C per 5 metres with a rapid increase in temperature being recorded in the aquifer sands beneath the coal body as indicated by the delta temperature log, although the boundary between these zones could be somewhat higher than expected in the observation bore because of conduction in the steel casing and convection in the water filled bore.

The deeper rotary wells drilled further out in the basin (Fig. 4) feature an unusual drop in temperature through the coal beds compared with the aquifer sands. This probably indicates the lack of thermal equilibrium in the bores due to invasion of drilling fluids. However the delta temperature logs do show that the aquifer sands contain high temperature waters. Variations in these temperatures could give a measure of the particular aquifer's transmissivity.

Stallman [27] has suggested that horizontal temperature gradients in aquifer systems could be used for this purpose but in the above case what is suggested is a comparison of one aquifer with another in the particular area. The effect of thermal conductivity variations on the temperatures within the aquifers would have to be ignored. (See also Bredehoeft and Papadopulos [6].)

Another interesting feature is the narrow width of the higher temperature zone along the outflow direction as indicated by the drilling program. The zone is limited to the south by the presence of the Baragwanath anticline which brings the aquifer sands to the surface. The total width of the high temperature outflow appears to be no more than 5 km.

The continuity of the higher temperature zone and brown coal beds from the Morwell area to Lake Wellington, where they occur beneath some 600 metres of Oligocene and Miocene marine sediments, indicates a stratigraphic continuity and places the coal measures as pre-Oligocene in age, a point at present in dispute on palaeontological grounds [19].

Variations in the geothermal flux

If the average geothermal flux is assumed to be 1.5 micro cal/cm^2/sec [13] the energy generated per square kilometre of earth's surface amounts to 1.5 $\times 10^4$ cals per second (63 kW).

Using an average transmissivity of 0.02 m^2/sec and gradient of 0.1% to represent the main aquifer system and since the thermal waters appear to be restricted to a 5 km wide zone in which the water temperature is some 47 °C above the average annual intake temperature of 13 °C, then the thermal energy being transferred by the aquifers amounts to approximately 4.5×10^6 cals per sec (18 megawatts) thus indicating that some 300 square kilometres are involved in generating this energy, provided radiation losses are ignored.

Although the main coal fields cover an area of this order the rapidity with which the groundwaters reach higher temperatures (Fig. 2.) suggests the possibility that high geothermal gradients could be encountered in the vicinity of the monoclinal structures over which the groundwater traverses such as the Yallourn Monocline and Haunted Hills Fault. These are expressions of basement faults which have probably been intermittently active throughout the Cainozoic.

Since it is known that vertically dipping beds are more conductive to the geothermal gradient than horizontal beds [7], it seems reasonable to assume that deep seated structures such as the Yallourn Monocline could influence the local thermal gradient. Such variations in the geothermal flux have been recorded in the Rhine graben, a feature somewhat similar in structure and age to the Latrobe graben. Variations in thermal flux of between 1.5 and 4 μ cal/sec/cm^2 occur, and extremely high thermal gradients are recorded along the fault zones bordering the graben (Doebl et al.).

The differences in "bedrock" type could also influence heat flux. The older Lower Cretaceous sediments which underlie the southern part of the area (see Fig. 3) can be expected to have a lower heat flux because of the greater depth of burial of the crystalline basement beneath these rocks. This variation may explain the general lack of high temperature waters in the Loy Yang and Coolungoolun coal fields.

The thermal waters of eastern Gippsland so far only mentioned briefly, can be explained by the predominance of granite basement in this part of the basin, and the location of aquifer sands close to the basement beneath a thick cover of Tertiary marls and limestones. The influence of coal beds is not significant in this area.

Hind and Helby [10] have attributed the presence of high temperature waters over large areas of the Great Artesian Basin of New South Wales and Queensland to higher than normal geothermal flux rates from crystalline basement compared to other areas underlain by thick sedimentary sequences.

Accurate measurements of the geothermal flux could, as they suggest, lead to a better understanding of "basement" geology and structure. Similar studies of thermal gradients could result in regional estimates of groundwater parameters but some basic assumptions made in the development of groundwater mathematical models preclude the use of these models in the deeper basins where high temperatures are encountered.

Groundwater hydraulics and heat energy transfer

The importance of aquifer elasticity in determining the yield of water from an artesian aquifer by Theis [33] in 1935 marks the beginning of modern groundwater theory.

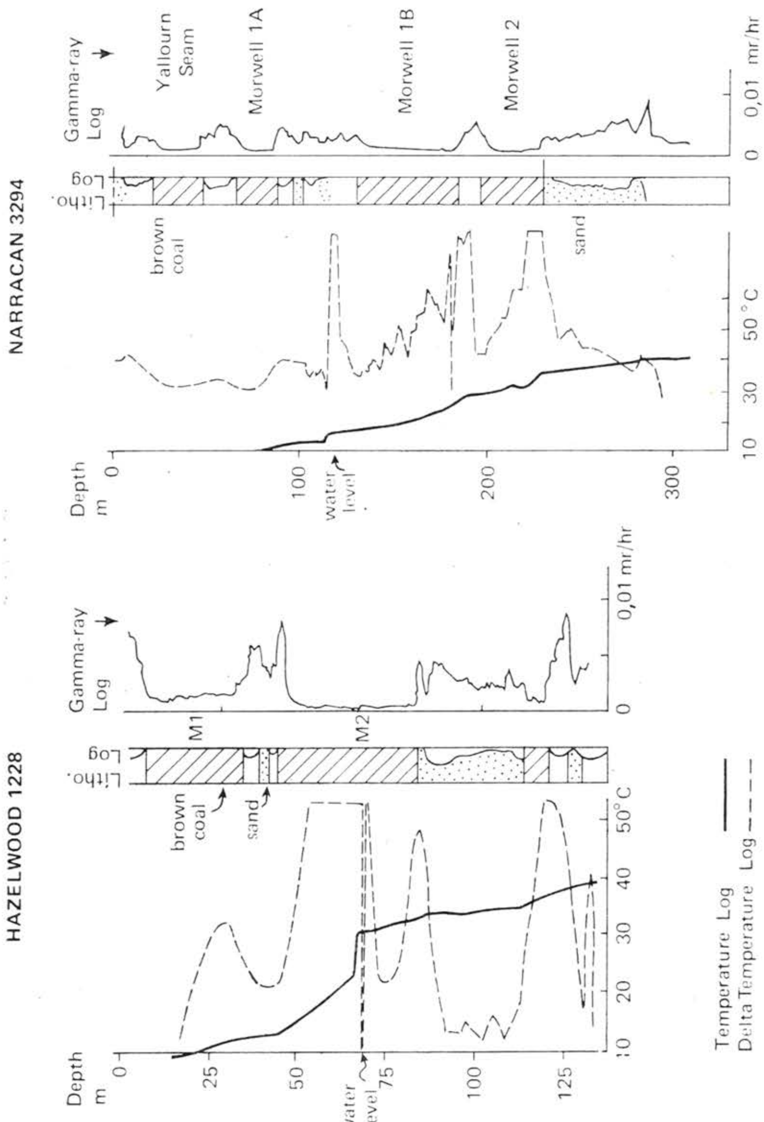
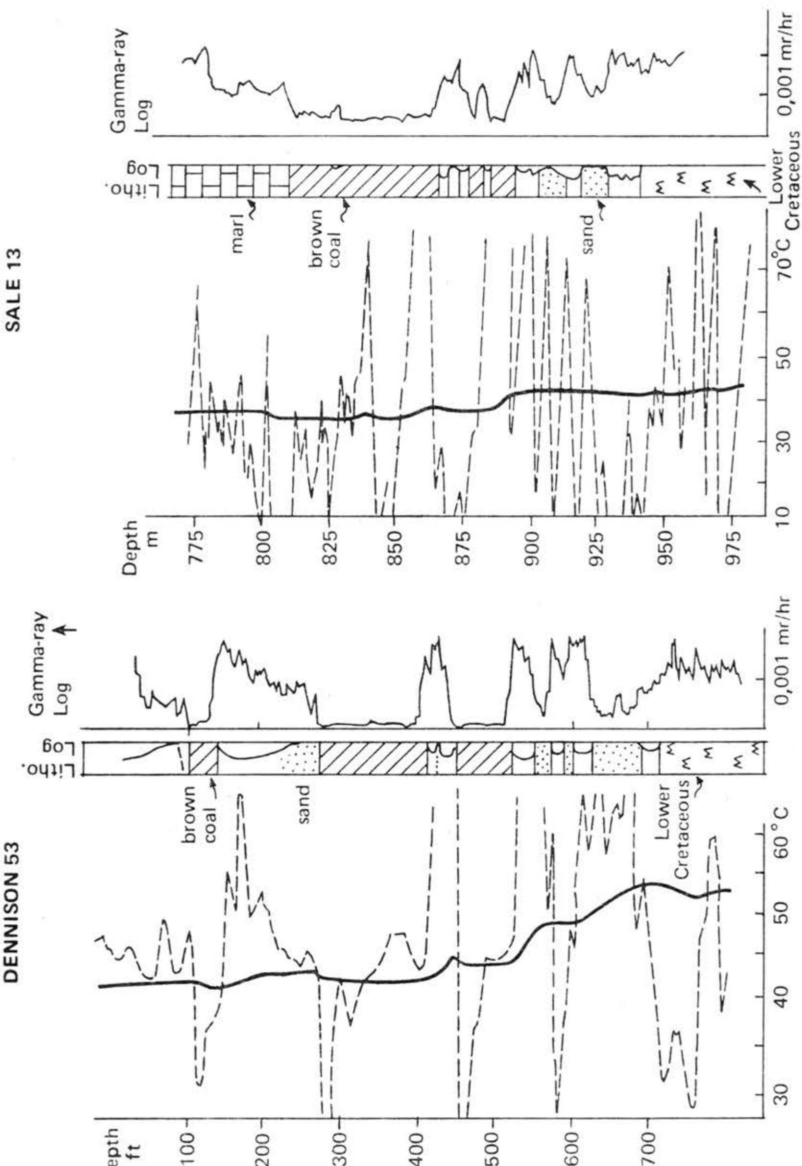

HAZELWOOD 1228
Depth m
water level
Litho. Log
brown coal
sand
M1
M2
Gamma-ray Log
0,01 mr/hr
50°C
NARRACAN 3294
Yallourn Seam
Morwell 1A
Morwell 1B
Morwell 2
0,01 mr/hr
50 °C
Temperature Log
Delta Temperature Log

Fig. 4a—b


DENNISON 53
Depth
ft
2100
2200
2300
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2500
2600
2700
30
40
50
60 °C
brown
coal
sand
Lower
Cretaceous
Litho.
Log
Gamma-ray
Log
0,001 mr/hr
SALE 13
Depth
m
775
800
825
850
875
900
925
950
975
10
30
50
70°C
marl
brown
coal
sand
Lower
Cretaceous
Litho.
Log
Gamma-ray
Log
0,001 mr/hr

Fig. 4c—d

Subsequently, mathematical models which successfully describe many field problems have been developed by the adoption of energy transfer equations from thermodynamic theory (WALTON [31] lists these early developments).

However, since groundwater studies are mainly concerned with yields, the solutions are described in terms of volume transfer rather than mass transfer for convenience. The assumption is made that density is constant on the basis that most groundwater provinces remain at relatively low temperatures (BEAR [2], WALTON [31]).

The problems of thermal energy transfer are treated in more detail in the thermodynamic literature but again the problem is simplified by considering it under four special cases (LUIKO and MIKHAILOV) [16, 15], the two cases applicable to groundwater problems assume either constant temperature and variable pressure or constant pressure and variable temperature, an approach usually taken by other investigators.

However, although constant temperature and thus constant density and volume can safely be assumed for many groundwater problems, both the pressure and the temperature can vary considerably in deeper basins and this assumption can lead to inconstancies.

As an example, consider the well Sale 13 near Lake Wellington (Fig. 1) which was developed to serve as an observation well for an artesian aquifer at a depth of 1000 metres. This flows at 0.75 m^3/sec — the water reaching a temperature of 62 °C after some hours. The shut in pressure exhibited by the aquifer, which is about 12 atmospheres at the surface, is dependent upon the density (thus the temperature) of the 1000 metre of water column above it. This will reflect the average temperature of the sedimentary column above the aquifer and will be at a considerably lower average temperature than the aquifer water. On the other hand, if the well is allowed to flow so that the temperature of the water column reaches that of the aquifer the time involved is such that a significant drop in pressure at the well head will have already occurred.

Thus, it is not possible to measure the piezometric level of such an aquifer at the surface at the temperature exhibited by the aquifer and shut in pressures would have to be determined.

This follows a somewhat similar conclusion to LUIKOV's [15] statement that at the molecular scale, it is impossible to measure the velocity of a fluid from density differences causing the motion.

Variations in the density of the water column result in considerable changes in the value of the piezometric level. In the example given a theoretical difference of about 20 metres is involved. Using standard solutions for the groundwater equations these changes would probably be interpreted as boundary conditions.

Other inaccuracies also arise in the determination of Storage Coefficient and Transmissibility of higher temperature aquifers.

Aquifers as energy transfer systems

An aquifer can be envisaged as a process by which gravitational and thermal energy is being transported by fluid motion to the point of outflow where this energy is then released.

The gravitational energy is involved from the point of entry of the water at the intake area, the thermal energy is added to the system as the aquifer passes through the geothermal flux of between 1 °C for 20 to 30 metres depth of burial.

The gravitational energy is released as the rate of mass transfer at the outflow zone and the thermal energy absorbed is indicated by the temperature of the outflow waters. However at any point within the aquifer the energy involved will be partly expressed as changes in viscosity, density and compressibility of the aquifer matrix and fluid as well as temperature and flow rate.

In groundwater theory the hydraulic characteristics of an aquifer are basically described by two parameters. Firstly, its permeability defined by DARCY's Law as the rate of flow of a unit volume per unit cross-sectional area per unit time under a unit hydraulic gradient. The value of permeability is dependent on both the fluid density and viscosity and is usually measured at 20 °C. A related field term the Coefficient of Transmissivity is defined as the volume of water transmitted per unit time through a unit width of the aquifer under a unit hydraulic gradient at the field temperature (WALTON etc.). Thus the field term does not include corrections for temperature variations by definition although tables are give to calculate permeability at 20 °C.

The other basic parameter is the aquifers coefficient of storage which is defined as volume of water released from the aquifer per unit surface area per unit change in head and is given by:

$$Sr = Sy + \gamma\Theta\beta m\left(1+\frac{\alpha}{\Theta\beta}\right)$$

where Sy = specific yield
β = bulk modulus of elasticity of water
α = bulk modulus of the aquifer skeleton
m = thickness of the aquifer
Θ = porosity
γ = unit weight of water

i.e. $S_{\text{total}} = S_{\text{due to drainage}} + S_{\text{due to compressibility}}$

The specific yield is not an energy term in that it is a measure of the quantity of water that can be drained from the aquifer thus infers that pumping is undertaken but the elastic storage is an energy term.

For water table aquifers the Storage Coefficient, which is equal to the specific yield in this case, varies between 0.01 and 0.30, while for artesian aquifer the elastic coefficient can be between 10^{-3} and 10^{-6} depending on the aquifer thickness.

It is a dimensionless parameter but refers to volume and not mass and thus takes no account for variations in relative volume due to temperature variations. These can in some cases be of the same order as the elasticity factor.

The change in relative volume of pure water with temperature increase is given on Fig. 5. It can be seen that the expansion of water, unlike the expansion of solids is non-linear, and increases with increased temperatures and is a measure of the gradual disruption of the intermolecular bonds between water molecules. At atmospheric pressure the increase amounts to 4% up to 100 °C while a value of about 1.5% would apply in the case of the Gippsland waters.

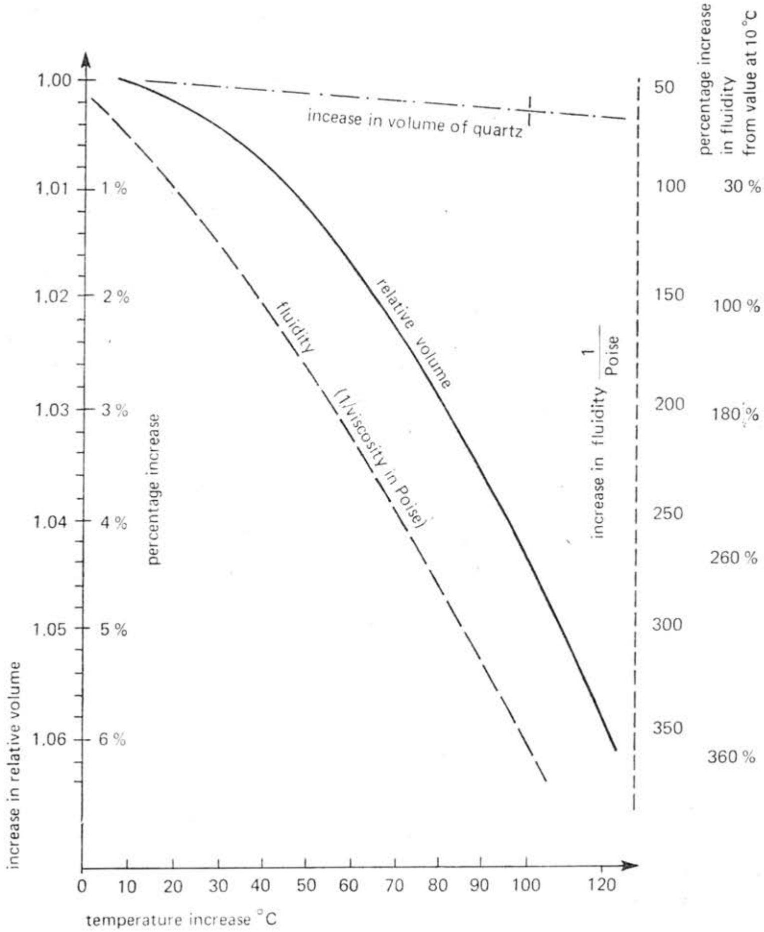

Fig. 5. Relative increase in volume and fluidity of water with increase in temperature at 1 atmosphere

Thermal energy and aquifer parameters

The effect of increased temperature on an aquifer can be considered in relation to the aquifer parameters.

In terms of volume the specific yield would show a relative increase due to the increase in temperature which, in the temperature range between 50 and 60 °C, would amount to $S = 3 \times 10^{-4}$ per °C for an aquifer with a porosity of 30%.

The effect on the elastic storage of a pressure aquifer would be due to a change in the porosity because of an expansion of the aquifer skeleton, and a possible change in the bulk modulus of elasticity of water which shows an increase with increasing temperatures above 60 °C at atmospheric pressure.

The capacity of an aquifer to transmit a fluid as defined by DARCY's Law is dependent on its viscosity and will increase substantially with increases in temperature.

This relationship as applied to a porous medium is given by the REYNOLD's Number which can be written as [31]

$$V = \frac{\mu R}{\varrho dm}$$

where R = Reynold's Number
V = bulk velocity
μ = dynamic viscosity
ϱ = density
dm = mean grain diameter

Thus the velocity will increase in proportion to the increase in fluidity and the decrease in density with increasing temperature.

The change in fluidity (l/viscosity in poise) with increased temperature is plotted on the same graph as increase in relative volume but on unrelated scales (Fig. 5). It can be seen that the fluidity value (l/poise) increases by some 360% over the range between 10° to 100 °C. This represents an increase in value of between 2.3 to 3.6% per degree C.

RORABRAUGH [23] comments on the change in infiltration rates in response to viscosity changes due to temperature variations, and suggests that the relationship is directly proportional. Tables are given by other authors to enable calculations of permeability to be made in terms of the specific permeability at 20 °C.

If transmissibility was defined in terms of standard temperatures an increase of some 60 to 80% would be applicable to the Gippsland parameters because of the increase in temperatures involved.

When it is considered that the piezometric level of the deeper high temperature aquifers cannot be measured with any certainty, and that temperature variations cause changes in basic aquifer characteristics, it can be seen that mathematical models which neglect these factors can no longer be used with any confidence however successful they are in describing low temperature groundwater systems.

A decrease in temperature due to pumping

These concepts suggest that tne release of pressure with an aquifer by removal of water from it will result in a decrease in temperature since the aquifer temperature is an expression of the thermal conductivity, the flow rate through the aquifer and the geothermal flux rate.

In most cases this is expected to occur very slowly. For example there are reports of flowing bores in the Great Artesian Basin that have maintained temperatures above 100 °C for over 50 years. It has also been noted that heat loss from porous zones that have been subject to artificial injection of high temperature fluids occurs very slowly (SQUIERS et al. [33]).

The Morwell open cut dewatering programme is an unusual case in that water at 50 °C was pumped at the rate of 1.4 m^3/sec from the main aquifers for four or five years, – the present rate is about 1 m^3/sec. This represented a thermal energy loss at the rate of some 80 megawatts from a relatively small area.

The dewatering bores show considerable fluctuations of temperature but the SEC report [24] quoted above states that "the trend in both the Morwell 1 and 2 aquifers is for slighly cooler discharge – this is in the vicinity of 2 °C in almost all bores between 1972 and 1973".

The aquifers are not fully dewatered and there is no evidence of intermixing with recently induced recharge waters as indicated by the radio-carbon age determinations quoted in this report – M1 waters gave an age of 20,000 years BP and M2 waters 13,800 BP.

Conclusion

The recent development of more accurate temperature probes will enable heat flow studies of sedimentary basins to be carried out in far greater detail than has been possible in the past.

This technique is likely to be of considerable value in delineating the stratigraphy and the hydraulic capacity of deeper aquifers as well as resolving some of the basic problems of accounting for variations in the geothermal flux.

The study of the Gippsland thermal waters indicates that both aspects must be considered together in sedimentary basins since aquifers can account for a large proportion of the heat energy balance of a system.

However, some of the basic assumptions made in the development of groundwater theory are no longer fully applicable in the study of such aquifers, and modifications must be made, perhaps by returning to the mass transfer theory again, if the vast reserves of both water and low grade thermal energy are to be studied in quantitatively.

Acknowledgements

The wells mentioned in this report were either drilled as a part of the basin investigation being undertaken by the Geological Survey of Victoria or by the SEC as coal bores. MR. C. GLOE of the SEC has been kind enough to bring some of the data included in this paper to my attention; MR. B. STANMORE of Melbourne University supplied a thermal conductivity value for coal and has been helpful in discussion. The study was undertaken to be included as part of a Ph D project, and I wish to thank the Director of Geological Survey and Prof. J. F. LOVERING of Melbourne University for their encouragement.

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GROUND WATER OF BANGKOK METROPOLIS, THAILAND

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Introduction

The city of Bankok, Thailand, is one of the fastest growing capitals in the South East Asia. It has been also rapidly increased in population, the present total of which is about 4.2 millions, so that the adequate water supply has not met the demands. The water supply problem is due mainly to the insufficient distribution system and treatment plants as well as the raw water. The major source of untreated water at present is the Chao Phraya River. Another source which contributes about one-third of the total is from the aquifers underlying Bangkok. Moreover, during the past fifteen years as the industrial activities have been promoted, additional considerable quantity of water has been developed from the underground reservoir. Domestic water supply for the expanding housing projects and individual households in the suburbs of Bangkok Metropolis is also from the aquifers.The pumpage is so high at present that the permanent decline of water level could be observed and the salt water encroachment into the aquifers occurs. Moreover, large number of wells has not been technically properly constructed. The sealing off the undesirable water of poor quality, generally the saline water, is not adequately and properly conducted, so that the salt water contaminates not only the well but also the aquifer. The consequence is an abandonment of a large number of the production wells at present.

Geography

The Bangkok Metropolis consists of the cities of Bangkok, Thon Buri, Nonthaburi, Samut Prakan and Pathum Thani covering an area of about 750 km^2 (see Fig. 1.) The Bangkok Metropolitan area is located at the latitude 13°30′—13°55′ N and longitude 100°25′—100°45′ E, and is situated on the delta and flood plain of the Chao Phraya River which traverses the Lower Central Plain of Thailand. The Plain, also known as the Lower Chao Phraya Basin, extends about 200 km from north to south and about 175 km from east to west. It is bounded on the east and the west by mountain ranges and by the Gulf of Thailand to the south. To the north, it is bordered by a series

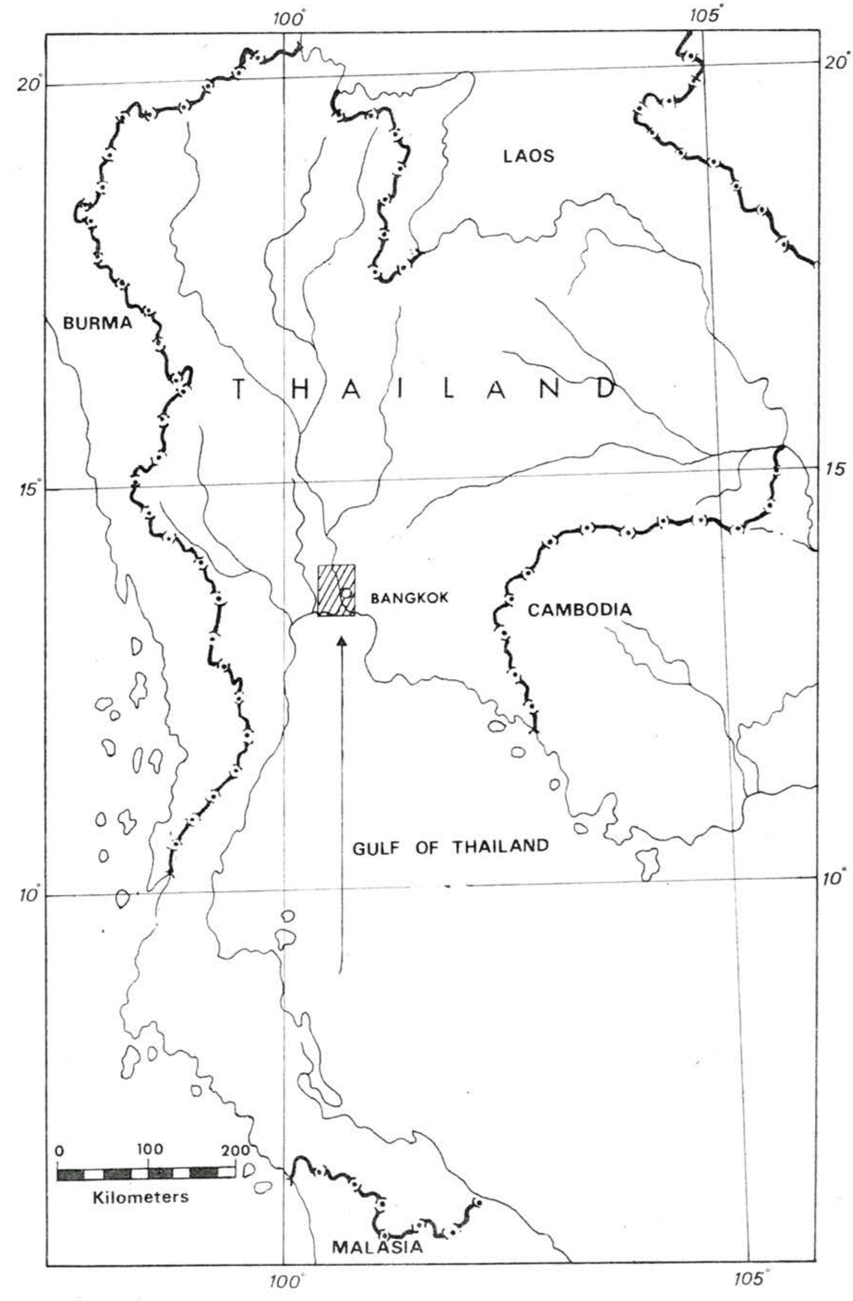

Fig. 1. Index map of Thailand. Shaded quadrangle = study area

of small hills deviding it from the Upper Central Plain, with the Chao Phraya River as an interconnection. The drainage systems of the Pa Sak River, the Bang Pakong River, the Tha Chin River, and their tributaries are also included in the Lower Chao Phraya basin, having the total drainage area of about 55,000 km^2.

The topography of the basin is relatively flat, with an average flood plain altitude of 1.5 meters above mean sea level in Bangkok and 15 meters from the same datum at 150 km north. The formation of soils in the basin has been resulted from the geologic processes and the topographic influences. The soils in the flood plain can be differentiated into three types: the soft marine clay, the brackish water soils and the fresh water soils. The soft marine clay over which Bangkok Metropolis is constructed is generally occurring along the Gulf shore-line extending from the Province of Chachoengsao in the east to north of Bangkok then swinging through the Province of Nakhon Phathom in the west until Petchaburi at the south-southwest. The brackish water soils, clayey in composition, occur continuously inland from the marine clay in semicircular arch starting from south of Prachin Buri to Nakhon Nayok, Ayuthya, Suphan Buri and Nakhon Phathom. The fresh water soils of mostly sandy clay composition occur in flood plains of the Pa Sak River, the northern reach of the Chao Phraya River and the Suphan Buri River. Soils that form the higher and the lower terrace areas on eastern and western sides of the basin are also included in the fresh water type but their composition is sandy to gravelly.

Average rainfall in the basin is about 1300 mm per year. The total water flow into the basin through the Chao Phraya River is estimated to be 24,100 million cubic meters per year. Out of this figure, about 6,200 million cubic meters is utilized for irrigation and 4,700 million cubic meters is for counteracting sea water intrusion in the Chao Phraya River (JARATWATHANA, 1971). The Metropolitan Water Works Authority (MWWA) withdraws about 200 million cubic meters per year from the Chao Phraya River for the public water supply. The rest (about 12,800 million cubic meters a year) discharges into the Gulf of Thailand.

Geology

Basement complex

The Lower Central Plain was evidently formed on the fault/flexure depression filled with clastic sediments. Aeromagnetic data indicate that the depression is underlied by various type of bedrocks at the depth from 400 to 3,500 meters. The basin floor generally slopes toward the central axis which is located more or less along the Chao Phraya River course and inclines southward to the Gulf of Thailand (ACHALABHUTI, 1975). In the vicinity of Rungsit just north of Bangkok the exploration hole encountered the quartzite bedrocks at the depth of 582 meters. A wildcat well at Phasi Charoen district striked granite gneiss basement at depth about 1,831 meters while another well at Ban Laomu southeast of Bangkok (803953, Fig. 2) reached quartzite and gneiss basements at the depth of 436 meters.

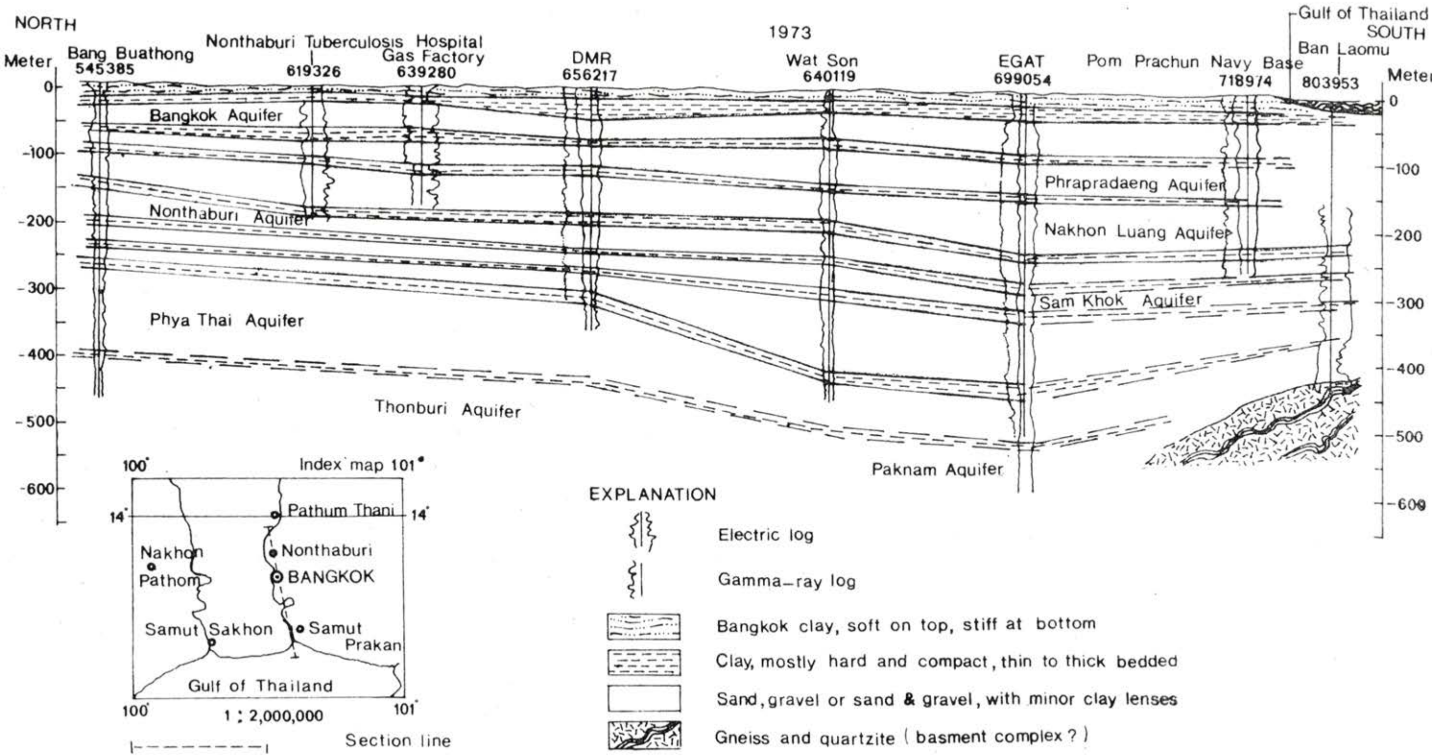

Fig. 2. Hydrogeologic N—S section of the lower Chao Phraya Delta showing principal aquifers of Bangkok Metropolis (correlated by electric and gamma-ray logs). (By CH. PIANCHAROEN—S. BUAPENG)

Unconsolidated and semi-consolidated deposits

Overlying the basement complex are unconsolidated and semi-consolidated sediments aged from the Tertiary to the Quaternary. During these periods the depositions were probably occurred under the fluviatile and deltaic environments with occasional shallow sea sedimentation. Total thickness of the formations ranges from about 400 to more than 1,800 meters. According to ACHALABHUTI (1975), sediments of the Chao Phraya basin and the Gulf of Thailand were laid down at the same time. He stated that oil wells in the Gulf penetrated sedimentary sequences ranging in age from Recent to Oligocene. Water well drillings in Bangkok Metropolitan area at the depth exceeding 600 meters indicate that there are at least three major breaks in depositions as evidenced by formation composition, geoelectrical properties and water quality. Each major break could be also subdivided into minor horizons.

The topmost sediment of Bangkok is a soft to stiff dark gray to black clay, known as the "*Bangkok Clay*" (MUKTABHANT, 1963), ranged in thickness between 20 and 30 meters. Underlyings the Bangkok Clay until the first break at the depth of about 100 meters are two sequences of medium to thick sand and gravel layers with minor clay lenses, separated by a distinct clay bed. The formations consist of poorly to moderately sorted medium to coarse sand with a typical thick sand-and gravel-bed very coarse grained at the lowest part. The carbonized woods or logs were always found in this particular horizon. The colour is generally light gray but in places yellowish due to oxidation. Underlying this formation is a yellowish brown compact clay bed which marked the first break mentioned above. The occurrences of the Bangkok Clay and the underlying clay (at the depth of about 100 meters) might suggest a hypothesis on the depositional environments which effect the ground water quality. Since the Bangkok Clay is saturated by brine or salty water with Recent shell fragments, the deposition might have been taking place probably during the transgression of sea water toward the basin in the post-Glacial time. This is also proved by the transgressive pattern of SP-curves (PIRSON, 1970) obtained from various boreholes in the area. If the hypothesis is correct, the sand-gravel formation at the depth of 20 or 30 to 100 meters should have been accumulated during the Upper Pleistocene. The oxidizing character and the presence of both salty and fresh water suggest that the deposition occurred under the fluviatile and deltaic environments during the transgression of the sea. The presence of carbonized woods or logs within the formation might be also a clue for correlation to other Pleistocene beds in other parts of the country (BURAVAS, 1969).

Electric and gamma-ray logs of drilled wells indicate the second shift of bases at the depth of 250 meters in the northern parts of Bangkok and gradually increased in depth to about 400 meters near the Gulf. These changes of the patterns are probably due to the break of sedimentation under prevailing environments, and might be corresponding to the close of Tertiary Orogeny (Late Pliocene) which formed the present physiographic features of the country (BROWN et al., 1951). The sediments at the depth of 100–400 meters seem to have more or less the same characteristics. Of the total thickness ratio between clay/sand-gravel is about 1:3. The total thickness of clay is about 70 meters and that of the sand-gravel is about 230 meters. The clay of this sequence is light brown to yellowish brown, usually stiff, normally sandy, gravelly in places. Fragments of shells are present in some clay lenses. Beds of sand-gravel are generally thick, ranging from 10 to 25 meters. The sand-gravel is characteristically light brown to dirty brown, becoming white in clean and well sorted layers, subangular to subrounded, generally moderately to well sorted but poorly sorted at the depth of about 260–300 meters. Based on fresh water and locally brackish to salty water contained by the formation, with their geoelectrical properties as well as the distinct break at the depth 350–400 meters mentioned, the whole of the sediments were probably accumulated under subaerated fluviatile environments during the Lower-Middle Pleistocene period. This is confirmed by Woollands and Haw (1976) who postulated that the regional deposition of the Gulf after Pliocene took place under paralic-terrestrial environments. The SP curves from the most of the wells penetrated this formation show the patterns of transgressive shoreline, and the brackish-salty beds normally show a bell shape and barrel shape SP curves. These patterns are an indication of distributary channel-distal bar deposits (Person, 1970). The correlation of SP curves indicate that during the Pleistocene period the Chao Phraya delta probably shifted back and forth at about 15 km north of its recent position, but its tributary channels seemed to be active in the southwest area.

Sedimentary sequences accumulated during the Pleistocene period have been found to be thickened southward and south-westward. In Pathum Thani area the depth to the sediments started at about 256 meters while in the Phra Pradaeng and Nong Khaem areas they can be found at the depth of about 420 meters (Fig. 2).

Beneath the second distinct break, sediments of different character could be envidenced by cutting samples and electric logging characteristics. The deposits are composed mostly of well sorted medium to coarse sand with occasional gravel. Beds of sand are commonly thick. The intercalated clay is generally sandy, having characteristics of claystone instead of ordinary clay. Colour of the whole sequences, from the depth of about 420 meters down to about 600 meters, is pinkish brown to grayish brown. Exception is made for the clay-stones found in the hole bored in area of Samut Prakan having gray to olive gray glauconitic colour which probably belongs to the fluviomarine environments. The system might be corresponding to the shaly sand, dark shale and red beds penetrated in the Gulf of Thailand which occurred during the sea transgression in the Pliocene (Woollands and Haw, 1976). The system consisting of reddish brown sand, pinkish to reddish brown clay, and olive gray clay was probably deposited in the fluviatile–fluviomarine environments.

Electric logs and gamma ray logs of three wells drilled beyond the depth of 650 meters indicate another shift of bases at about 600 meters. Informations are, however, inadequate for geologic interpretation other than that the ground water contained is brackish to salty.

Ground water

Ground water has been exploited for domestic supplies in Bangkok since the past six or seven decades. The first consumer of ground water in Bangkok is not known or recorded. One well driller who is now almost 65 years of age witnessed the drilling of a well, in his childhood, by a Chinese driller using the bamboo rig, and the completed well was of artesian nature. Another source discloses that the first European-made drilling rig, powered by steam engine, was operated at Bangkok Dock some 30 years ago. The first governmental development of gound water for public water supply in Bangkok was marked by the drilling of 10 production wells by the Department of Public and Municipal Works during 1954–1955 while the Department of Mines, now Department of Mineral Resources, has been undertaken the ground water exploration in the Northeast. In 1957–1959, the Department of Public and Municipal Works drilled another 52 production wells in Bangkok and Thon Buri. Also in 1958–1960, the Police Department installed another 80 wells for fire protection, and turned over to the municipality later for public water supply. In addition, USOM sponsered the drilling of 20 wells in Bangkok and Thon Buri during 1958–1960.

Since the establishment of the Metropolitan Water Works Authority (MWWA) in 1967, more than 100 wells were drilled to date. Of more than 200 wells mentioned, only about 170 wells are still in operation for municipal water supplies for Bangkok and suburbs, not including more than 100 production wells drilled for government complexes. The pumping rate of those wells ranges from 50 to 300 m^3/hr. The present total pumpage for public supply is about 340,000–350,000 m^3/day.

As already mentioned, the probable first utilization of ground water was done by private owners. The total number of private wells is not known to date. The constant users are industries, hotels, apartments, government complexes, hospitals, universities, housing projects, and private homes. The industries, especially the brewerage and paper factories, consume a large quantity of ground water. The locations and actual pumping rates are still doubtful. By a reasonable guess, there are more than 1,500 private wells existing, with the pumping rate ranging from 2 m^3/hr up to 400 m^3/hr. The history and records of drilling of these private wells are mostly not available for references. The CAMP, DRESSER and MCKEE (1969) estimated that the total pumpage of private wells was about half of the total public supplies. Due to the sharp declining of water levels at present, such estimate was probably too low for today pumpage. With all available evidences, approximation of 75% or the pumpage of 255,000 m^3/day would be more logical. The present total pumpage altogether would then be something like 600,000 m^3/day.

The aquifer

From the geological, hydrological and geophysical studies, the aquifer system of Bangkok can be differentiated into 8 aquifers, from top to bottom, namely:

Bangkok Aquifer (50 m zone),
Phra Pradaeng Aquifer (100 m zone),
Nakhon Luang Aquifer (150 m zone),
Nonthaburi Aquifer (200 m zone),
Sam Khok Aquifer (300 m zone),
Phaya Thai Aquifer (350 m zone),
Thon Buri Aquifer (450 m zone),
Pak Nam Aquifer (550 m zone).

The Bangkok Aquifer is immediately overlied by the Bangkok clay which is the topmost sediment in Bangkok. It consists of a sequence of thin to thick layers of sand and gravel with many clay lenses. The depth of this aquifer is 30 meters as an average. Total thickness is about 50 meters. The aquifer is unused due to the occurrence of salty water in it.

The Phra Pradaeng Aquifer is separated from the Bangkok Aquifer by a clay bed of about 10 – 15 meters in thickness. The aquifer can be reached at an average depth of 80 – 90 meters, and is about 50 meters thick. The aquifer consists of gravels and sands with clay lenses. The sand and gravel are easy to identify for they are mostly coarse grained, white colour, with carbonized woods or logs in many places. The deposits were undoubtedly laid in the mouth of the river or the very shallow sea on the erosional surface of hard and compact older clay which is now underlying them. Thickness of the aquifer decreases to the north and increases to the south. Ground water is heavily produced from this aquifer in areas south and southwest of Bangkok, but not many wells were drilled in the north because salty water is present.

The third, *the Nakhon Luang Aquifer*, is very heavily exploited since ground water has been utilized for water supply in Bangkok. It is also highly suffered from the overpumping at present. Nakhon Luang is the shallowest aquifer that produces water of excellent quality for drinking purpose. The aquifer also consists of permeable sands and gravels with some clay lenses, and leaky clay beds. Depth to the aquifer is about 120 – 150 meters and average thickness is about 50 meters. The thickness is slightly decreased to the south where it is now contaminated with salty water. The aquifer is hydraulically separated from the above aquifer by the hard and compact clay already mentioned.

The Nonthaburi Aquifer has been utilized since not many years but is now popular to drillers who require larger productive wells. The aquifer was not developed in the old days because it is too deep to drill economically, and the quantity of water of the above aquifers is normally sufficient to meet the specifications which require not more than 150 m^3/hr per well. At present many wells are specified to pump 300 m^3/hr. It is then unavoidable for drillers to tap this aquifer for additional water.

Depth to the aquifer is at an average 200 meters, and the thickness is fairly uniform for about 50 meters. Geologically, the aquifer is similar and conform to the Nakhon Luang Aquifer, i.e. it composes of sands and gravels with minor clay lenses, and can be divided into at least 3 subaquifers, separated by leaky clay beds.

The deepest aquifer penetrated by normal drilled wells is the *Sam Khok Aquifer* which is located at the depth of about 300 meters; down to the bottom of the aquifer at the depth of about 350 meters. The deepest production well penetrating this aquifer is 305 meters in depth and is located at Pathum Thani, the town previously called Sam Khok. The aquifer consists of alternating layers of sands or gravels and clays. Clays are generally brown to yellow and moderately to highly compacted. Sands and gravels are generally medium to well sorted but clay lenses always intercalated.

The aquifers at a depth exceeding 350 meters are too deep to reach by normal production wells. They consist of thin to rather thick sand-gravel layers with the water bearing properties similar to those of the Sam Khok Aquifer.

Aquifer characteristics

The Department of Mineral Resources started collecting data of ground water in Bangkok by way of relaxation, bit by bit, since 1966 until now. And more data are still needed.

As already observed, the production wells in Bangkok are tapping the Phra Pradaeng Aquifer, Nakhon Luang Aquifer and Nonthaburi Aquifer; with not many wells penetrating the Sam Khok Aquifer. The individual sand or gravel beds are up to 15 meters thick. They consist of poorly to well sorted mixture of sands and gravels which are very permeable and yield water readily to the properly constructed wells. The individual aquifer is fairly uniform in thickness and extends a long way out of Bangkok to the recharge areas. Most wells properly drilled and developed will yield at least 100 m^3/hr, although the pumping rate of the municipal wells and some industrial wells is as high as 300 m^3/hr.

The specific capacity of each aquifer is generally similar, ranging from about 15 m^3/hr per meter of drawdown to 40 m^3/hr per meter of drawdown depending on the type of screen, screening section, and completeness of development. The poorly constructed and developed wells rarely yield less than 45 m^3/hr with specific capacity less than 6 m^3/hr per meter of drawdown. This figure does not include those of old wells suffered from clogging of sand or incrustation or corrosion of well screens.

In 1968, Department of Mineral Resources worked together with the CAMP, DRESSER and McKEE in conducting the longterm pumping tests for MWWA. Recently, Department of Mineral Resources has conducted much more controlled pumping tests for the Bangkok, the Phra Pradaeng and the Nakhon Luang Aquifers, the results of which are shown in Table 1.

From the figures in Table 1, the aquifers are evidently moderate to highly permeable and very similar in characteristics. For other aquifers, reliable controlled pumping test has not been undertaken. Since the geological and geophysical examinations reveal the similar nature of depositions and characteristics, high capacity wells can be expected.

Table 1

Aquifer characteristics

Aquifer	Location of tested well	Coefficient of transmissibility [m^2/hr]	Permeability [m/hr]	Storage coefficient
Bangkok	Bang Pun	160	3.40	1.00×10^{-4}
Phra Pradaeng	Pom Phra	110	3.74	–
	Chun Navy Base Phra Pradaeng	70	2.38	–
	Wat Phai Ngoen	120	2.38	1.00×10^{-4}
Nakhon Luang	Wat Phai Ngoen	65	2.21	1.00×10^{-4}
	Lum Phini Park	100	3.40	2.00×10^{-4}
	Pak Kret	110	2.55	–
	Bang Bua	125	3.45	2.2×10^{-3}
	Dept. of Mineral Resources	50	2.65	2.60×10^{-4}

Much has been asked also about the proper well spacing to avoid interference effect. By using the aquifer constants and mathematical calculation (Table 1), two wells producing an average rate of 150 m^3/hr from the same aquifer can be spaced 500 meters apart, or wells producing average rates of 300 m^3/hr being 1500 m apart would not cause very severe interference effects. (Interference effects are about 0.75 and 0.90 meters in each case, respectively.)

Water level

As previously mentioned, uses of ground water in Bangkok can be dated back at least 70 years. Records of water levels of those old days were not available except the statement of the flowing conditions of wells or the water levels near ground surface. At any rate, the water level or piezometric head should undoubtedly be only few meters from ground surface since the pumps installed were mostly the centrifugal pumps or reciprocating pumps whose suction head can not exceed 8 meters. This type of pumps were replaced by the deep well pumps some 20 years ago.

Record of water levels were not adequate until 1958 – 1959 when there were data enough to prepare a water level map. By 1958 – 1959, water levels have already been declined and cones of depression, actually called cone of pressure surface, developed in many areas of heavy pumpage. The deepest water level as shown in Fig. 3 was about 12 meters (40 feet) in the center of Bangkok, while the deepest in suburbs was about 9 – 10.5 meters (30 – 35 feet).

Since 1958 – 1959, the Phra Pradaeng and Nakhon Luang aquifers were heavily pumped. The regional decline of water levels was resulted. Many isolated cones of depression joined each other to form a large cone as also shown in Fig. 3 for 1968 – 1969 period.

By 1968 – 1969, the deepest water level in the heavily pumped area in the heart of Bangkok was 25.3 meters (83 feet). In the suburbs, where water level in 1958 – 1959 was 4.5 meters (15 feet) the level declined to the depth of 12 meters (40 feet). At least 2 large cones of depression developed in 1968 – 1969. The new one was located near Phra Pradaeng where industrial wells were dominant.

In the 10 years period of 1959 – 1969, the water in both aquifers dropped at almost 12 meters (40 feet) in the heart of Bangkok and 7.6 meters (25 feet) in the suburbs. The decline of this 10 years period cannot be computed as rate of decline per year, since the actual heavy pumpage occurred in the later part of 1966, as shown by the hydrographs of observation wells in the areas.

Since 1966 – 1967, ground water have been heavily utilized in the suburbs, especially in areas east of the Chao Phraya river, while in the center of Bangkok no great increase of pumpage has been observed. The result is that water level in the heart of Bangkok declined only gradually, but that in the suburbs, especially in areas southeast, east, and northeast of Bangkok is markedly changed. South of Bangkok at Sam Rong, water level in 1969 was 10.7 meters (35 feet), 25.3 meters (83 feet) in late 1972 and 32.3 meters (106 feet) in 1974; a decline of 21.6 meters (71 feet) in 5 years. Just north of Amohoe Bang Kapi, water level in 1969 was 10.7 meters (35 feet), 22.6 meters (74 feet) in late 1972 and 26.8 meters (88 feet) in 1974; a decline of 16.1 meter (53 feet) in

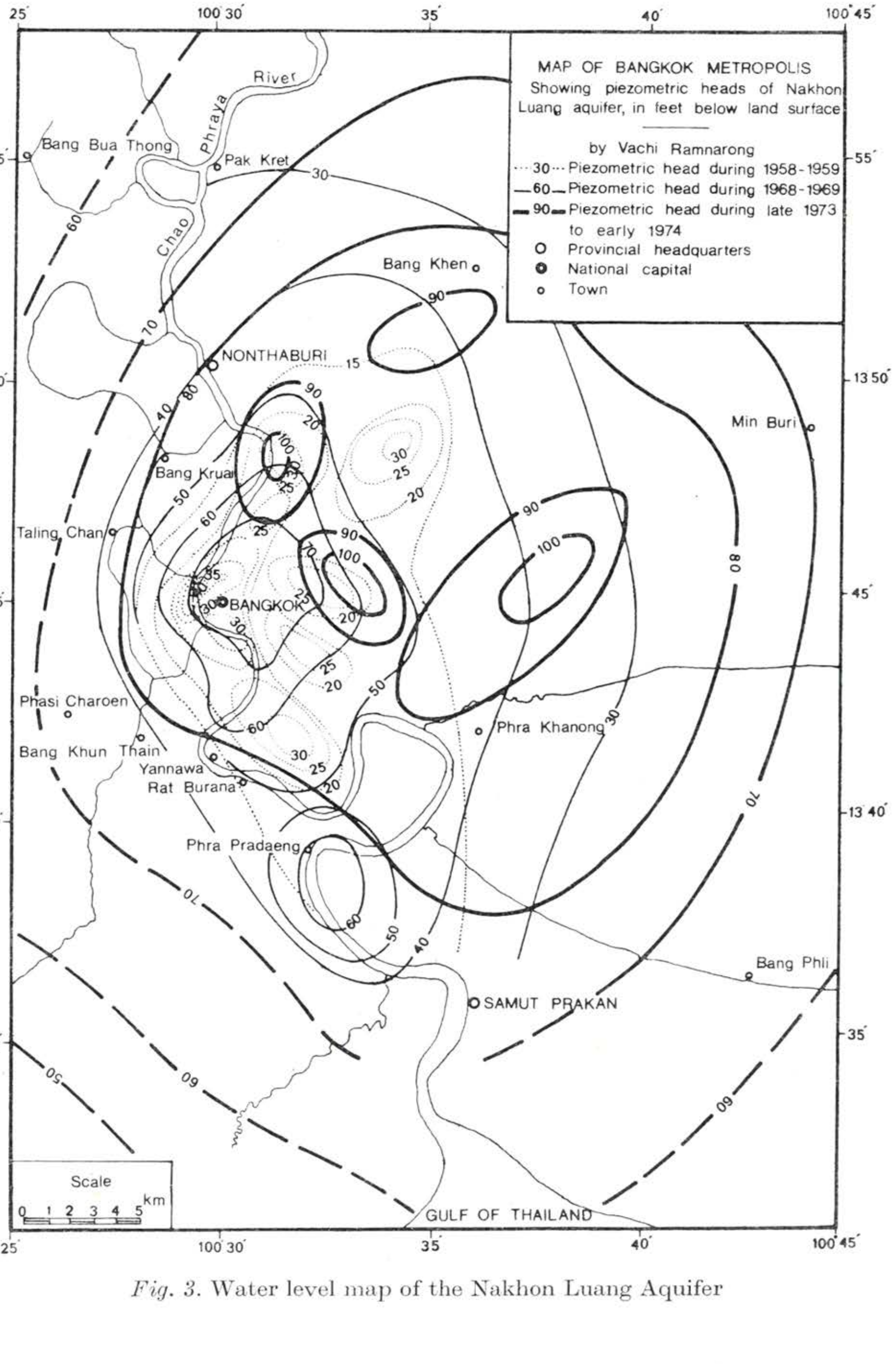

Fig. 3. Water level map of the Nakhon Luang Aquifer

5 years. At Bang Khen, water level in 1969 was 9.2 meters, 24.0 meters (79 feet) in late 1972 and 29.6 meters (97 feet) in 1974 (30 feet); a decline of 20.4 meters (67 feet) in 5 years.

Within these overpumped areas, water is withdrawn mostly from the Nakhon Luang aquifer and partly from the Nonthaburi aquifer. The rate of declining to day is 3 – 4 meters per year and the pattern of cones of depression is shown in Fig. 3. Since the Phra Pradaeng aquifer is no more popular in this region, the rate of declining seems to be less, i.e., 2.7 m/yr at Department of Mineral Resources observation well, and 1.8 m/yr at Pak Kret observation well.

Ground water recharge

Since the uppermost geologic formation of the Bangkok Metropolitan area is a very thick clay bed, the direct percolation of rainfall or influent seepages from streams into the ground water reservoirs are nil. The ground water recharge is primarily from the lateral ground water inflow from the periphery of the basin as shown in Fig. 4. The tentative hydrologic balance studies of the Chao Phraya Basin (PIANCHAROEN, 1976) indicate that the direct percolation through the soil zone as well as seepage from streams to become ground water recharge are only 6% of the total annual rainfall, or about 4,212 million cubic meters. This recharge is regarded as the safe yield of the ground water system or the perennial ground water supply for the Chao Phraya Basin under the existing climatic and hydrologic conditions. Since the Bangkok Metropolitan area could share only a fraction of the safe yield, the heavy pumpage of individual aquifer would result in the overdraft. The pumpage at present is about 1.38 million m^3/yr (about 1 mgd) exceeding the safe yield.

Water quality

The major cause of inferiority of the ground water quality in Bangkok is NaCl, Hardness, Fe and Mn. For the Bangkok aquifer, analysis of many wells reveals high concentration of the following constituents:

Cl	2000 up to 6500 ppm
Fe	as high as 2.0 ppm
Hardness	as high as 920 ppm
Mn	as high as 1.5 ppm

In the Phra Pradaeng aquifer, Cl is less than 10 ppm at Phra Pradaeng and Bang Na, up to 500 ppm in Central Bangkok and 1000 ppm at Pak Kret. The highest is 1952 ppm at Min Buri. The hardness is less than 100 ppm in Phra Pradaeng, up to 836 at Pak Kret. Fe is less than 0.3 at Phra Pradaeng, up to 4.2 at Pak Kret while Mn is generally 0.00 – 2.00 ppm.

In Nakhon Luang and lower aquifers, water quality is generally ranged as follows:

Cl	5 – 50 ppm
Fe	0.16 ppm (average)
Hardness	40 – 180 ppm
Mn	0.03 ppm (average)

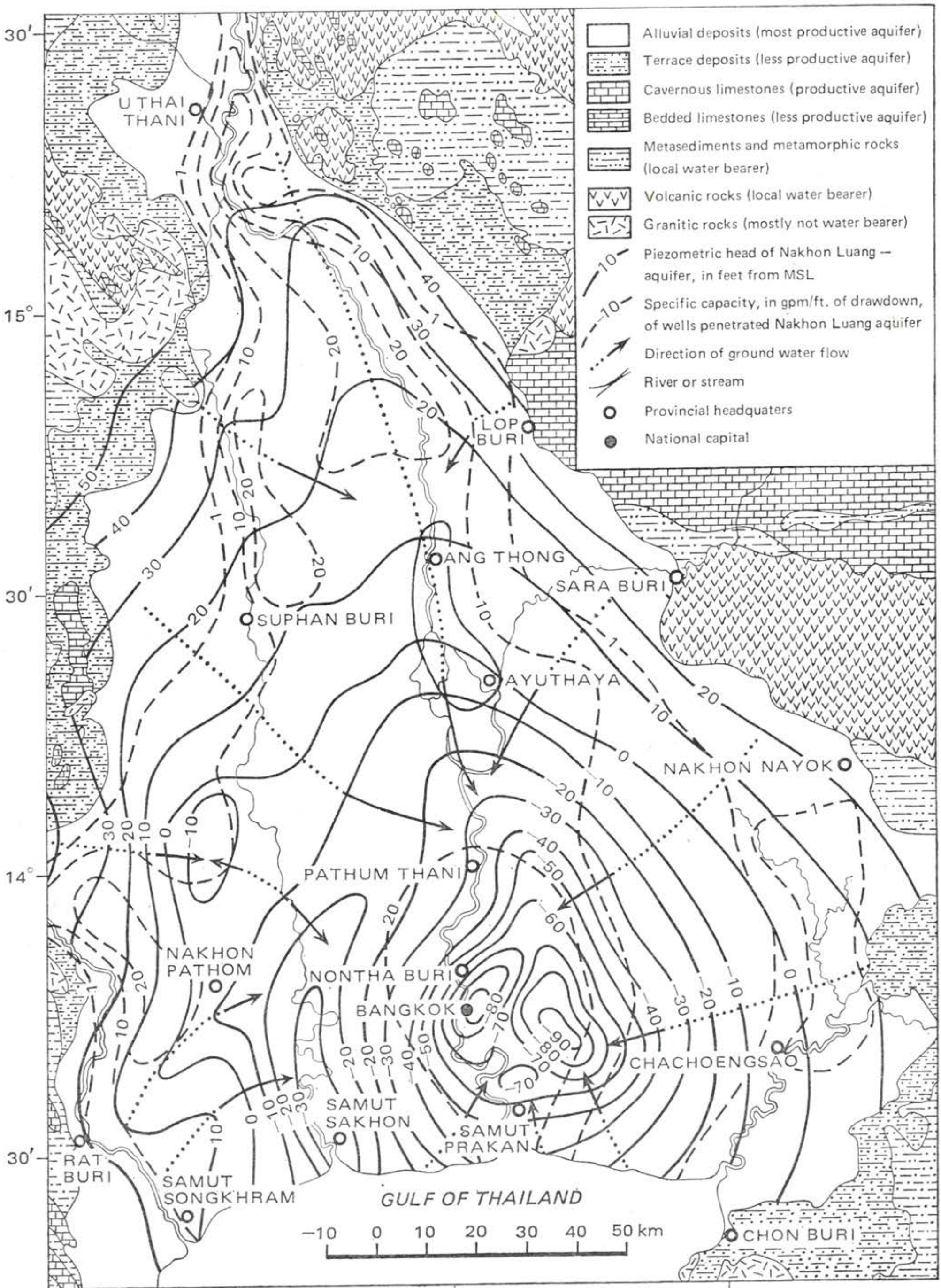

Fig. 4. Hydrogeologic map of the Lower Central Plain

In general, water quality does not change very much with time of uses. Exception is made for the pattern of isochlors which indicate salt water intrusion as shown on the map. More informations are referred to in Table 2.

Salt water intrusion

The salt water intrusion had not been observed until the last 15 or more years when many municipal wells in Thon Buri and Southern Bangkok yielded brackish or salty water. Since then more and more number of wells, especially along the bank of the Chao Phraya river southward from Central Thon Buri and Southern Bangkok, have been abandoned due to salty water. The mechanism of salt water intrusion into the aquifers is not yet clarified. Observations and interpretations of chemical character of water from those abandoned wells revealed that actual sea water intrusion occurs only in areas near the shore; such as at the Pom Phra Chun Navy Base near the mouth of the Chao Phraya River. Landward, no trace of sea water can be interpreted because the concentration of chloride in the water is much less and the ratio of calcium to magnesium is much higher than those of sea water. The chemical characters of such salty water are presumably those of connate water or water entrapped in pore spaces of sediments at time of their accumulation in the shallow sea of brackish to salty environments. In recent time, as the sea retreated or landmass was uplifted and delta extended southward, salty water was gradually flushed out by fresh water derived from infiltration in the west, north, and east. Flushing effect was almost completed in areas northward not many kilometers from the present Gulf coast, leaving some less permeable sediments soaked with brackish to salty water as discovered today. At all time, the hydrologic system was adjusting so that fresh-and salt-water bodies were more or less in equilibrium with each other, and at the same time in equilibrium with the sea water body. With these equilibriums, the general direction of ground water flow was presumably from the north, west and east toward the sea.

When ground water has been exploited in Bangkok and Thon Buri, the hydrologic balances between fresh water/salt water/sea water bodies were disturbed. The pattern of ground water flow was shifted. The disturbance was severe when the cones of depression due to heavy pumpage were developed. Ground water flows from all directions toward the center of cones of depression. This will bring about the reverse flow of ground water of areas south of the cones from the original north to south direction to south to north toward the heavy pumpage areas. This reverse flow has already brought the salty water to move northward and encroached back into the fresh aquifers.

The rate of movement of salt water is relatively low in comparison to fresh water movement toward cones of depression. This is not due to the high or low permeability of salt water zones, but it would rather be due to the higher proportion of fresh water flowing into the cones of depression when the pumpages are still not too great to accelerate the rate of salt water movement. Therefore, the rate of salt water intrusion differs in different parts of Bangkok and Thon Buri. In areas of heavy pumpages, the rate of movement was estimated for about 0.5 kilometer per year while 0.1 kilometer per year

Chemical analyses of ground water

Aquifer	Location code	Depth (meters)	pH	Ca	Mg	Na	K	Fe (total)
Bangkok	707473	47	7.3	308	193	1610	29	–
Phra Pradaeng	670169	120	7.5	19	7	–	–	0.1
	698992	106	7.5	142	51	338	1.2	4.4
	661092	88	9.0	12	9	90	6.5	0.1
Nakhon Luang	658145	152	7.2	72	75	248	7.0	0.2
	649326	198	8.4	15	11	66	4.7	0.1
	671206	158	8.3	25	78	98	0.0	0.3
Nonthaburi	629369	149	7.3	2	2	251	2.3	0.2
	639280	176	7.4	38	14	47	5.5	0.3
	722293	198	7.2	28	10	93	3.1	0.0
Sam Khok	659426	268	7.7	27	7	–	–	0.1
	545385	274	8.2	–	–	–	–	2.1
Phyathai	545385	335	8.0	–	–	–	–	1.0
Thon Buri	475147	487	7.7	46	27	16	9.8	0.2
	545385	457	8.0	–	–	–	–	2.1
Pak Nam	560158	640	7.4	16	19	175	16	20
	699054	560	7.4	32	10	76	5.5	0.4

was a figure for areas of less pumpage. At any rate, the stronger decline of water levels would undoubtedly accelerate the rate of movement.

As already mentioned, Central to Southern Thon Buri and Southwestern Bangkok are already encroached by salt water. The upper limit of encroachment in the Nakhon Luang aquifer can be drawn somewhere from the vicinity of Amphoe Bangkok Yai to south of Bangkok Noi railway station, crosses the Chao Phraya River toward east-southeast reaching Amphoe Pathum Wan, then swings southwest to Amphoe Samphanthawong, runs down south until Amphoe Yan Nawa, curves up east-northeast to Bangkok Technical College, then swings down south and southeast through the east of Amphoe Phra Pradaeng, Samrong and Amphoe Bang Phli. There is a tendency of further movement to the east and northeast directions in the vicinity of Pathum Wan, Khlong Toei, and Samrong areas.

Land subsidence

The possibility of land subsidence in Bangkok due to the effect of deep well pumping has been postulated by many soil scientists but no significant evidences or scientific proof could be offered. There is also no systematic investigation or observation leading to a reliable quantitative expression on subsiding behaviour to date. The soil engineers are, however, of the opinion that extracting of water from the aquifers is causing declines in the piezo-

Table 2

from selected wells in Bangkok

Mn (total)	Cl	CO_2	CO_3	NO_3	F	HCO_3	SO_4	Total solids	Hardness as $CaCO_3$	Aquifer
113	3075	28	0	0.0	0.1	350	366	–	1560	Bangkok
0.0	14	3	0	0.2	0.1	60	1.4	121	76	Phra Pradaeng
0.6	773	13	0	1.4	0.0	253	5.0	–	567	
0.0	33	1	16	0.0	0.1	227	0.4	–	66	
1.0	578	27	0	1.1	0.0	270	2.8	–	490	Nakhon Luang
0.0	3	2	6	0.0	0.0	251	5.2	–	82	
0.0	17	3	0	0.0	0.1	341	4.0	–	94	
0.0	198	24	0	0.0	0.1	304	1.6	–	14	Nonthaburi
0.0	2	19	0	0.0	0.2	307	2.8	–	152	
–	0	38	0	0.4	0.1	376	1.2	383	110	
0.0	9	14	1	0.4	0.1	449	0.0	463	97	Sam Khok
–	6	–	–	0.004	–	–	–	404	106	
–	90	–	–	0.006	–	–	–	476	128	Phyathai
0.0	4	8	2	0.0	0.8	248	1.2	–	226	Thon Buri
–	580	–	–	0.0	0.5	258	54	–	120	
0.8	78	27	12	0.0	1.2	423	6.4	–	118	Pak Nam
0.0	33	16	0	0.0	0.5	258	54	–	120	

metric pressure in the normally consolidated Bangkok Clay. HALEY and ALDRICH (1969) conducted the consolidation test and reported that the Soft Bangkok Clay is somewhat overconsolidated. By calculation, assuming the past pressure of 3 – 5 ton per square meter (tms) greater than the existing effective stress, the increase in load of 3 – 5 tms as a result of decline of piezometric pressure would bring the stress level to crisis and result in the settlement from re-compression of 0.05 – 0.07 m, respectively. Further from this critical stress, additional 3 – 5 tms stress would have the magnitude of settlement about 10 times the re-compression settlement. Under the tested soil property, the rate of drawdown of 0.3 m per year in the clay would take 10 – 15 years to increase the average stress level by 3 – 5 tms, respectively. BRAND and PAVEENCHANA (1971) computed the settlement of Bangkok Clay as a function of water level drawdown for four sites by two modes of assumption, either the water level dropping but the clay still being saturated or the water level remaining unchanged but piezometric level in the underlying sand layer declined. The maximum subsidences of 0.82 – 1.04 m were calculated for each mode of assumption as a result of consolidation in the Bangkok Clay alone. The computation also indicated that the subsidence of 10 cm at each site would be due to the drawdown of 0.15 – 1.3 m for the former mode and 1.9 – 4.7 m for the latter. Other consolidation tests at two sites as reported by SITTHICHAIKASEM (1975) also revealed that under external loading of 3 – 5 tms, the maximum compaction of the Bangkok Soft Clay are 0.40 – 0.56 m at one place and 0.85 – 1.24 m at another, respectively. Since the permanent external loads at present are earth-fills and the engineering constructions without piles or with short piles, the maximum subsidence of 1.06 m is predicted for the residential area as a result of load from earth-fill and buildings alone.

The observation and measurement of three piezometers by the Ground Water Division indicated that the rates of lowering of water level in the Bangkok Clay and the underlying sand are 6.5 and 106 cm per year for the last three years, respectively. The significant relative surface settlement in the vicinity of a deep well site within the area observed is 6 cm. The cause of subsidence is far from conclusive as a result of pumping since many new building constructions were also commenced. This type of land settlement occurs in many areas, the maximum of which is measured 15 cm.

Since Bangkok could not tolerate any land subsidence, the Government has accepted in principle to issue a Ground Water Act to control the pumping in Bangkok, and the National Environment Board is planning to implement three inter-related scientific programs in Bangkok: the levelling for investigation of land subsidence, the investigation of land subsidence caused by deep well pumping and the development and management studies of ground water resource. These programs are aimed for completion within 4 years with the total estimate budgets of about 1.5 million dollars.

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Discussion

E. SOBOTHA donne quelques annotations au sujet des méthodes employées et de l'importance de recherches concernant les bords et le sous-socle d'un bassin et l'utilisation des données météorologiques et des données sur les eaux souterraines. Il présente plusieurs diapositives.

A. C. REBOUCAS parle des bassins sémi-arides du Brésil. Il expose la zonalité du chimisme des eaux souterraines. Jusqu'à 300 m de profondeur il y a des eaux carbonatées, de 300–500 m, des eaux sulphatées, audelà 500 m, des eaux chlorées et très salées. Quant au mouvement des eaux souterraines, le cadre géologique est décisif : la configuration des roches cristallines, des couches argileuses, la situation des dykes. Aujourd'hui la demande en eau est très faible par rapport aux ressources. Dans le bassin en question il y a 7 aquifères superposés jusqu'à 300 m. L'élément le plus important est la qualité de l'eau. Maintes fois la température est très élevée, il faut refroidir l'eau. L'auteur présente plusieurs diapositives.

T. BANDRABUR répond aux communications de M. STEGENA. Il illustre les conditions géologiques et hydrogéologiques dans la partie de l'est de la dépression pannonienne. Plusieurs coupes géologiques démontrent le grand graben au Nord d'Oradea. C'est une depression plus de 4000 m profonde. L'auteur fait connaître les données de débit et de la temperature des sources et des puits thermales et les anomalies géothermiques.

H. M. VAN MONTFRANS : "The Netherlands are situated at the margin of the North Sea Basin. Since Early Tertiary time, a subsidence of some 1500–2000 metres has developed. The many seismic surveys that have been made for the oil companies allow the unravelling of the Tertiary and Quaternary tectonic movements. Two different neo-tectonic 'provinces' are present : a salt-tectonic area in the eastern and northern part, and an area of general tension in the middle and the south of the country. In these two areas, faults connected with the presence of salt in the subsurface and downthrow-type faults respectively, are found. Apart from the presence of faults there are distinct differences in subsidence. In general, the neo-tectonic movements have clearly influenced the pattern of occurrence of the lithostratigraphic (=hydrogeological) units, as is shown by the following examples.

In the Central Graben in the southern part of the country, coarse-grained Middle Pleistocene sands of the river Rhine are present. In the tectonically higher parts on both sides of this graben these sediments are absent. The sands form a very good aquifer. They are covered by Upper Pleistocene fine-grained sands and loams of low permeability. These sediments are much reduced in thickness to the sides of the graben. Farther down the section, coarse-grained Lower Pleistocene and Pliocene sediments are found. Due to the graben structure great differences in thickness are present in these sediments. These beds too contain fresh water down to a depth of a few hundreds of metres, and are therefore of great importance for ground-water exploitation. Summarizing, it can be said that in this area there is a clear relationship between tectonic movements in Tertiary and Quaternary time on the one hand, and the pattern of occurrence of the lithostratigraphic units on the other.

Another example is the occurrence of the so-called potclay deposits in the salt-tectonic area in the northern part of the Netherlands. These deposits are now thought to have probably formed in melt water lakes, during Elsterian Glacial time. Locally there is a relationship between the occurrence of these clays and the salt-tectonic movements in the subsurface. The presence of the potclay, with thicknesses of sometimes well over a hundred metres, is an important factor for the geohydrology of the area.

A third example of the relationship between tectonics and lithostratigraphy is the situation near Manderveen in the eastern part of the country. Here a NNW–SSE striking salt-tectonic grabenlike structure was found with the aid of seismic reflection profiles. Coarse-grained Lower- and Middle Pleistocene sediments are present in this graben. During the Saalian glaciation the structure probably was a topographic low, judging from the observation that tills and other glacigenic sediments are found in the structure. On one side there is a large ice-pushed ridge. Ground-water exploitation in the area has led to a lowering of the contour lines of the phreatic ground-water, giving rise to damage claims. For this reason, a detailed knowledge of the geological setting is needed.

Also on the scale of the whole country an insight into the neo-tectonic history is of importance. For a long time it was thought that the coarse-grained and the fine-grained fluviatile sediments represented sediments from glacial and interglacial times respectively. Now it appears to be more probable, at least in the case of the Lower Pleistocene sediments in the SE Netherlands, that these differences have mainly stemmed from the balance between the tectonic activity in the source areas and in the area of deposition itself."

T h e m e 3

HYDROGEOLOGICAL MAPPING, GEOPHYSICAL AND GEOCHEMICAL DATA

T h è m e 3

CARTOGRAPHIE HYDROGÉOLOGIQUE, DONNÉES GÉOPHYSIQUES ET GÉOCHIMIQUES

GENERAL REPORT

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15 papers have been presented in our section according the Thematical Group No. 3 on

— hydrogeological mapping,
— geophysical and geochemical data.

Altogether 15 papers have been prepared by 21 authors from 6 different countries. There are 7 authors from the Soviet Union and the same number from Hungary, 3 from Czechoslovakia, 2 from Poland, and 1—1 from Yugoslavia and Romania. — 11 manuscripts were written in English and 4 in French.

Looking at the subject of the group of theme concerned, it could be thought that reports of only measured and described facts would be offered as material for our discussion. But, on the contrary; being the Nature one and indivisible, and so is the science itself too, we have received excellent analyses and interesting conclusions. So we have a lot of links to the other thematical groups, such as those of the subsurface water movements, thermal and mineral waters, tectonics etc.

There are only 2 papers dealing definitely with mapping, and the largest number of the studies, namely 10 are operating partly or wholly with chemical data and 5 with results of geophysical investigations. 3 papers summarize the results of regional exploration, and there are 4 ones from which interesting methodical conclusions can be drawn.

ERDÉLYI in his paper "*Chemical aspects of groundwater-flow regions of the Hungarian basin*" tells us a few examples how hydrochemical data have helped to explain hydrodynamic conditions. Precipitation water enters the soil on the topographic highs, these are the recharge areas of the groundwater flow. The bicarbonate-type groundwater moves down and away from the recharge areas, meanwhile the warmer, saline deep-situated groundwater moves upwards from greater depths. A zone of chemical equilibrium has been formed between the ascending and descending waters, where the result is an intermediate sort of chemistry. In this transition zone water is characterized by a lower hardness and higher salinity. In the basinward direction the salinity is growing.

There are areas of low salt concentration, lying also farther away from the border of the basin. This seemingly anomalous situation is explained by

ERDÉLYI as taking for basis the sand percentage of the Quaternary sedimentary sequence and the existence of many sub-basins with separate hydrodynamic centres. The characteristics of these centres are given in the paper.

KARÁCSONYI and SCHEUER consider a smaller area in the northern peripheries of the Hungarian Great Plain, in their paper entitled "*Physical and chemical characteristics of confined waters in peripheric territories*". This study presents a full comparison between two smaller areas. Local lithological changes in the Upper Pannonian sequence are reflected by varied hydrogeological characteristics. The mixed chemical character of water is one of the evidences of a potential recharge. The paper is really based on manysided work. Water chemistry, lithological data, temperature and pressure conditions and radiometric dating have helped the authors to summarize their conclusions.

FERU deals with the artesian waters of the Upper Pliocene to the north of Oradea. In the paper "*Particularités chimiques des eaux artésiennes du nord d'Oradea*" the chemical differences between phreatic, artesian and artesian-mineral waters are represented. He discusses the role of the clay minerals, especially that of the montmorillonite in the change of the chemistry of water. He also found that, with the depth of the aquifer, the chemical character changes from Ca-Mg bicarbonate to sodium bicarbonate type with increasing CO_2 contents.

IVANOV and KARASSIEVA in their paper "*Gites des eaux carboniques des bassins artesiens profonds*" give the main types of chemical constitution of the artesian waters in different basins of Siberia, the Caucasus and the Russian territories.

Now there are 2 further papers approaching the problems from a chemical aspect, dealing with one single chemical constituent of water.

EREMEEV—KOROBEINIK and YANITSKY offer a very interesting study on "*The nature of water exchange in artesian basins, according to the helium surveying*".

A general pattern of the underground water migration is discussed and, by the way, water exchange in the artesian basins is clearly manifested. As helium is constantly born in the deeper parts of the Earth, and is dissipating from the atmosphere into the space, a steadily growing concentration can be expected in the Earth's crust. Highly accurate and effective methods of measurement as well as results of the concerned surveying in the Soviet Union, are reported. The concentration of helium was always connected with deep faults in the Earth's crust, meanwhile the order of the measured concentrations is connected with the position i. e. depth of the crystalline basement, and not with the age of the rocks.

As along the tectonical and hydrogeological openings, higher helium concentrations were observed, the authors deduce that there should be present an underground water migration there. To prove this statement they present data showing anomalous chemical composition of water where helium appears in a maximum concentration. A 4-year-period measurement has shown to be values of concentrations periodically changing up to peaks of 25 times higher than normal. The observations have proved the existence of a dynamic exchange of waters in all parts of the hydro- and lithosphere. It is but questionable whether helium is moved by water or air.

MAJKA-SMUSZKIEWICZ describes a local but really very interesting problem aroused by the sulphurdihydrogen content of waters around a sulphur mine in Poland (*"Problèmes hydrochimiques dans l'exploitation du soufre de la région de Tarnobrzeg"*).

There are 3 further papers dealing with the chemical constituents of water.

In the paper *"Geochemical regularities in the distribution and concentration of alkaline elements of the carbonated waters in Georgia"* ZAUTASHVILI matters over the changes in the chemical composition of waters taken from Mesozoic aquifers.

Approaching the active tectonic zones, the migration order of the rare alkaline elements changes and there is formed a new sequence led by lithium and followed by caesium, sodium, potassium and rubidium.

Further on the author shows a remarkable correlation analysis, and demonstrates interesting relations between the concentration of different elements. The

ratio of sodium to potassium is about	4,
potassium to rubidium grows up to	83,
potassium to lithium	0.5, etc.

All these relationships are probably connected with postvolcanic metasomatosis.

A question significant for health protection is treated by DOBOS, in the paper *"Examen des eaux potables et thermales principales de Hongrie pour la détermination de leur teneur en bromide, iodide et fluoride"*. The critical and necessary concentrations of halogenic elements are given, from the viewpoint of health protection, and the measured concentrations are compared to an example taken from the Hungarian Great Plain. The results are given in maps.

DOWGIALLO's paper entitled *"Stable isotopes as indicators of the origin and zonal affiliation of deep-seated groundwaters in sedimentary basins"* tries to solve the very complicate problem of the origin of groundwater in deep basins, by the help of mass spectrometry measurements of the oxygen and hydrogen isotopic composition of water. The results indicate the marine, atmospheric or mixed origin of the water and, by the way, the vertical zones of their distribution. Presuming that the isotopic composition of the water of the oceans has not changed since the Late Palaeozoicum, the author propounds to use a ratio given by the portion of sea water found in the water, expressed in per cents and named as isotopic relictivity indicator. The factors which may complicate the interpretation are discussed in the paper, and some theoretical considerations are taken for the Polish Lowland.

In the eastern areas of the Great Hungarian Plain the gas content of waters in the Pannonian Basin was investigated by KARÁCSONYI and SCHEUER. Their experiences are summarized in the paper *"Relationship between the gas content of the water-bearing strata and the hydrogeological regions"*.

The comparison of the theoretically possible and the actually observed gas contents of the water shows that waters containing the highest amount of gas can be found at a depth of 50 to 100 m.

According to the authors, in this region the gas contents of the water-bearing layers are composed of "in situ"-formed natural gases. Looking also at the chemical data, it seems that this region of maximum gas content corresponds to the transition zone, discussed in the paper of ERDÉLYI, and so it could be connected with the movement of water masses.

There have been presented studies based upon geophysical methods (see e.g. KARÁCSONYI and SCHEUER) on the physical and chemical characteristics of confined waters in the marginal zones of the Hungarian Basin. BUCSI SZABÓ and SZLABÓCZKY deal with the relationship between the changing relief of the buried compact basement and the uncommon pressure variation in the shallow, relatively loose i.e. unconsolidated Pannonian aquifer system. It is supposed that the vertical decrease of the pressure gradient is closely related to the structural position of the basement. The paper "*Hydraulic interconnection between the relief of a highly dense basement and the near-surface loose porous groundwater-bearing sediment sequence*" discusses four areal units of the Great Hungarian Plain showing regularities in the pressure gradient distribution. The negative change in the gradient may be an important indicator of an efficient recharge and its recognition may result in a considerable spare of energy costs in context with a high groundwater production. When reviewing the paper "*A contribution of tritium analyses to the study of groundwater flow in the sediments of Zitný Ostrov*" by POSPISIL, I have to refer to the Thematical Group No. 4, with the study of J. DEÁK. The author of this paper, using radioactive measurement, analysed the influence of the groundwater flow in the pervious sediments of the island to the east of Bratislava. He could define the main flow directions, the depth of the flow and the speed of the water movement.

KRULC and KOVACEVIC have reported in their paper "*The role of geophysics in the hydrogeological investigations of great sedimentary basins*" on a successful drilling carried out after a previous accurate geophysical measurement had been made. This work has led to the finding of abundant artesian water resources. In addition, they render a short description of the SW marginal zone of the Pannonian Basin.

Two excellent studies have been presented on mapping. SZLABÓCZKY ("*Preparation of maps of regional hydrogeological units*") reports on the preparation of the 1:100,000-scale hydrogeological map of a geologically rather complicated territory of about 10,000 km². The paper deals with hydrogeological classification, map symbols and with every necessary detail.

Finally, we should mention the excellent report by CERMÁK, LUBIMOVA and STEGENA under the title "*Geothermal mapping in Central and Eastern Europe*", which is expected to be subject of a further discussion for synthesis.

HYDRAULIC INTERCONNECTION BETWEEN THE RELIEF OF A HIGHLY DENSE BASEMENT AND THE NEAR-SURFACE LOOSE POROUS GROUNDWATER-BEARING SEDIMENT SEQUENCE

COMMUNICATION HYDRAULIQUE ENTRE LE RELIEF D'UN SOUBASSEMENT DE HAUTE DENSITÉ ET LA SÉRIE SÉDIMENTAIRE DE LA NAPPE PHRÉATIQUE

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RÉSUMÉ

Quatre exemples bien connus sont présentés de l'intérieur de la Grande Plaine Hongroise dans le bassin Carpathique en ce qui concerne la relation entre la surface alternante du socle à forte densité composé de Paléozoïque et Mésozoïque et de Miocène et les changements anormales du gradient de pression des aquifères meubles peu profondes.

Dans la région d'étude la pression statique n'augmente pas avec la profondeur des zones crépinés des couches de sable dans certains puits, mais elle va diminuer. Il est supposé que la diminution des gradients verticaux de pression est en relation avec les socles profonds. Les études ont des caractères descriptifs, car la preuve de la théorie a besoin d'une prospection détaillée avec des forages exécutés jusqu'au socle. Il est à remarquer que ce problème peut être expliqué par un changement de tension au point de vue géotechnique similaire au charge vertical développé autour des ponceaux tubulaires déposés dans le fond des digues contre les inondations à cause des tensions des ailes latérales.

Le changement négatif de gradient de pression comme un « cas extraordinaire » peut être une indication importante de l'alimentation dynamique et il peut résulter une économie considérable et raisonable des frais d'énergie de la production d'eau.

Problem

It was often found in the Hungarian subsurface water explorations that in the inner lowland areas of the Carpathian basin underlying by certain buried basement depression the pressure (head) is abnormally decreasing downward within the formation water-bearing Pannonian sand beds occurring about a few hundreds meter deep. In some wells different subsequent sand formations were screened and it was found that the pressure heads of the various aquifers were diminishing versus depth instead of increasing. The same was found within sand beds belonging to the same stratigraphical horizon having an anticlinal form when the potentiometric surfaces were compared at different points in such an area which was not disturbed by groundwater withdrawal.

Its theoretical model is shown in Fig. 1. This paper is dealing with some places where abnormal pressure condition of this kind was found in Hungary (Fig. 2). Before the study of this problem by means of test-wells any theoretical investigation is premature. The relation between the basement form (relief) and the hydraulical interconnection of the overlying sediment sequence is dealt extensively with references [6].

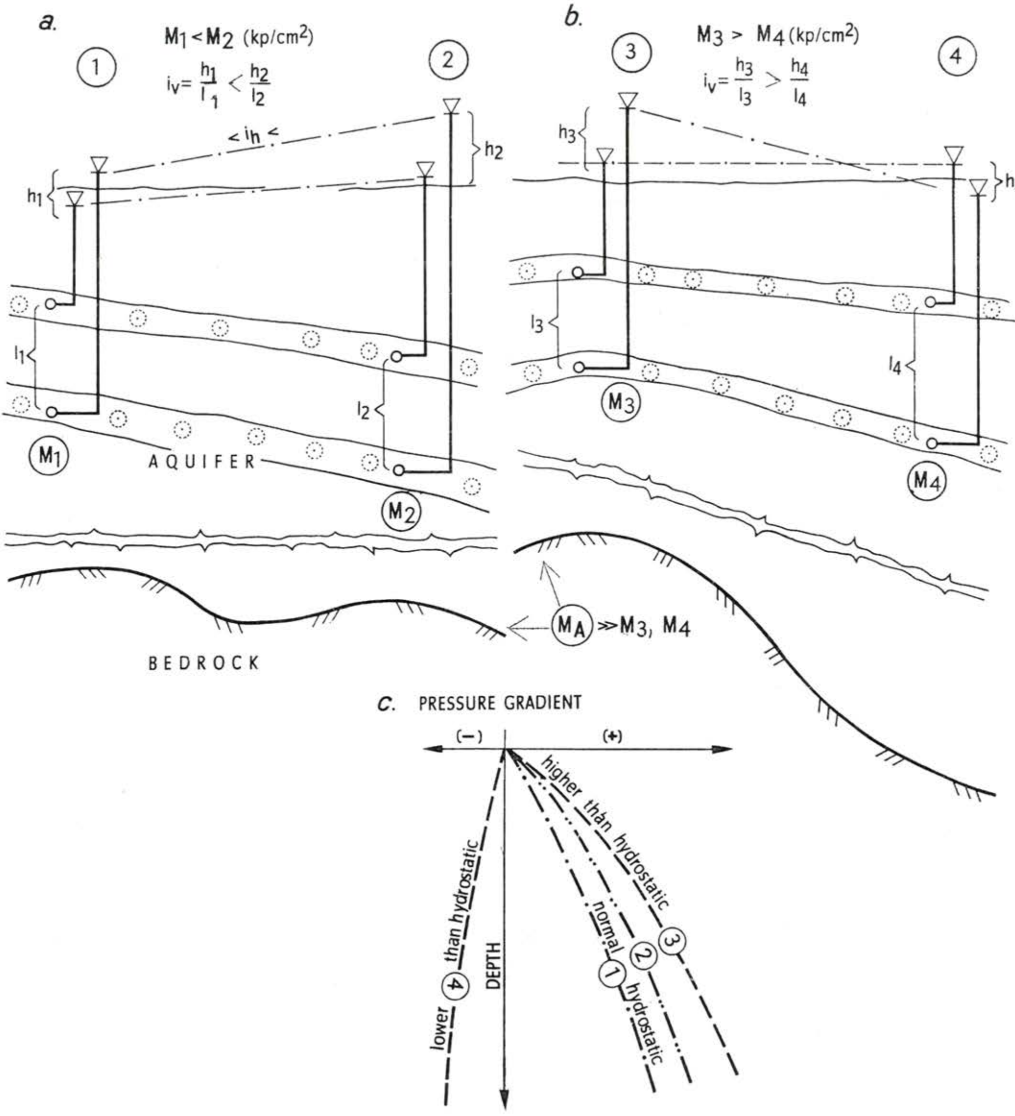

Fig. 1. Structural and hydraulic relation of the basement of high density and of the shallow aquifer system

Study made in the groundwater observation well network of the Great Hungarian Plain monitored by the Hungarian Geological Institute

The following tendency of pressure head variation was observed in the groups of well [5] spaced 8 to 10 kms apart in the central part of the Lowland where individual wells are screened in different zones, usually at three intervals (Fig. 3).

In the group of well near to Kunadacs located on an underlying high of basement a normal, relatively high pressure gradient was measured from the static water levels of the aquifer screened in different depth intervals. Eastward in the group of well at Kerekegyháza located on the border of the downwarped basement pressure gradient was dropped. The static water levels of the aquifers screened at different depth intervals remained nearly common. On the contrary, in the group of well at Kecskemét located on the steep slope of the buried basement a reverse pressure gradient was observed. The pressure heads of the ever deepening aquifers were more and more lower.

The structural section of the basement was constructed from gravitational and seismic geophysical data, based on the results of oil prospecting wells too. The minimum and maximum depths of the basement surface shown in Fig. 3 represent the interpretation of the gravitational measurements. The basement relief drawn by continuous line was obtained from "filtered" gravitational measurements and we consider it to be more exact.

In Fig. 4 can be seen the static water level measured in the three groups of well as well as the variation of the vertical pressure gradient (Fig. 5). As a result, a comparison is made between the basement relief and the vertical pressure gradient.

Above the maximum of the basement a relatively high positive vertical pressure gradient whereas towards the minimum of the basement a negative one was observed within the Quaternary-Pliocene unconsolidated water-bearing sediment sequence within the depth interval from 50 to 500 meters.

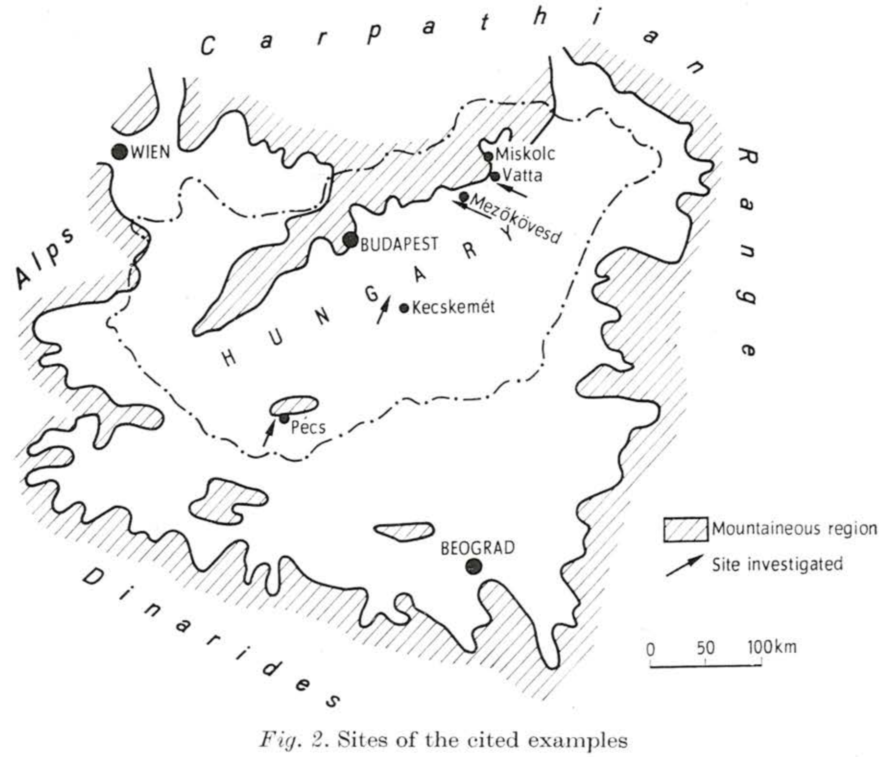

Fig. 2. Sites of the cited examples

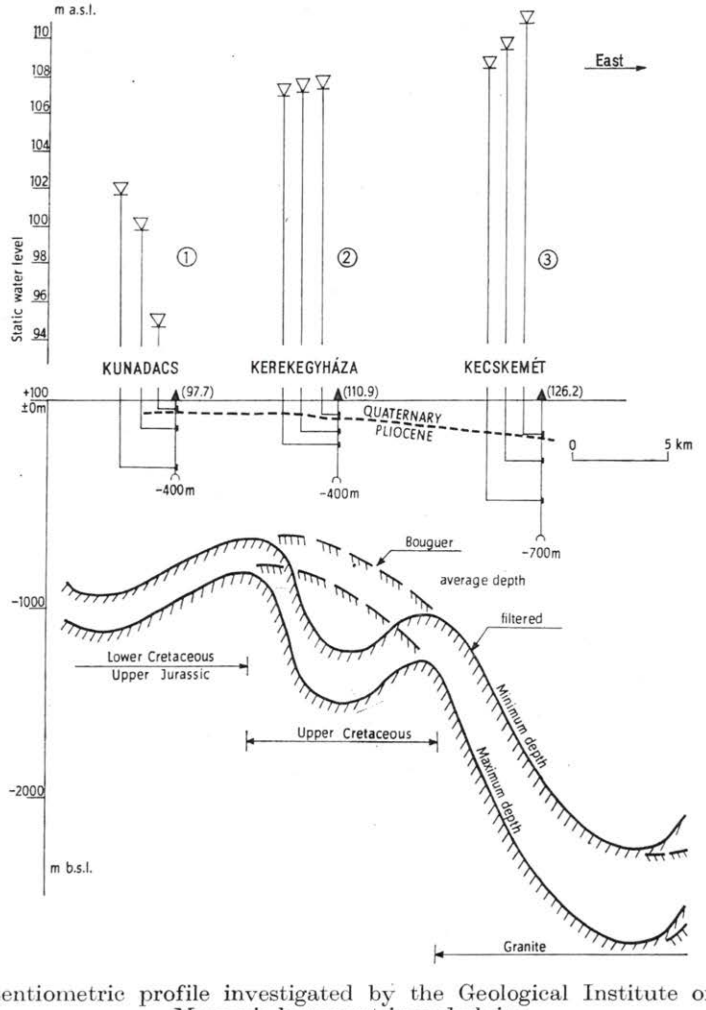

Fig. 3. Potentiometric profile investigated by the Geological Institute of Hungary. Mesozoic basement is underlying

Regional pressure variation observed during the explorations made in the northern part of the Great Hungarian Plain

Along the northern border of the Great Hungarian Plain a widely extended thick sand formation was revealed by more than 1000 exploratory drillings and nearly 50 groups of well screened at different depth intervals. This Pannonian sand formation has a general SE-dip.

A mildly dipping anticline was revealed in the north-eastern part which is intersected also by numerous faults and the sand formation is downwarped into the so-called Vatta – Maklár graben [2] which is situated in the foredeep of the Bükk Mountains. The anticline was developed likely due to the intrusion of the Miocene lava flow and here the highly dense basement rock is represented by this eruptive igneous rock.

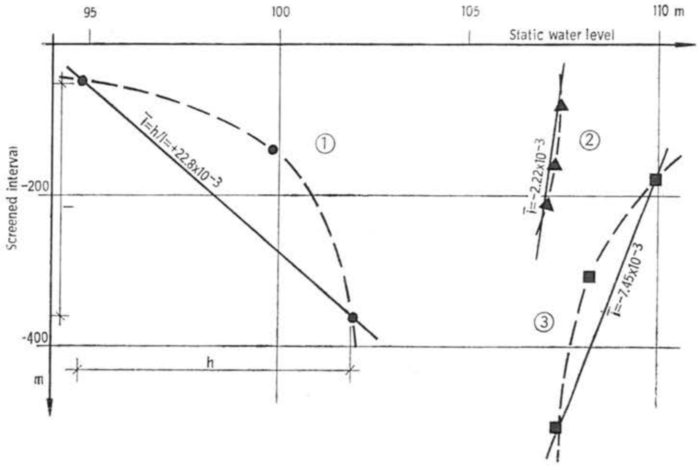

Fig. 4. Vertical pressure variation of a well network

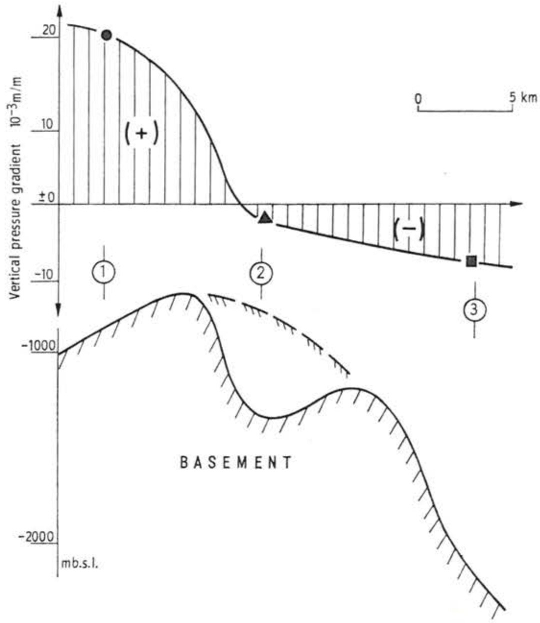

Fig. 5. Variation of the groundwater pressure gradient

In the case of normal change in pressure head it would be anticipated that the static water level would be increased proportionally with the depth of the sand formation which tend to be greater or lower in SE direction. On the contrary, a reverse situation was found on ground of numerous data: the static pressure was decreasing in SE direction. It can be seen clearly in the water level contour map in Fig. 6. On the one hand, the static water level

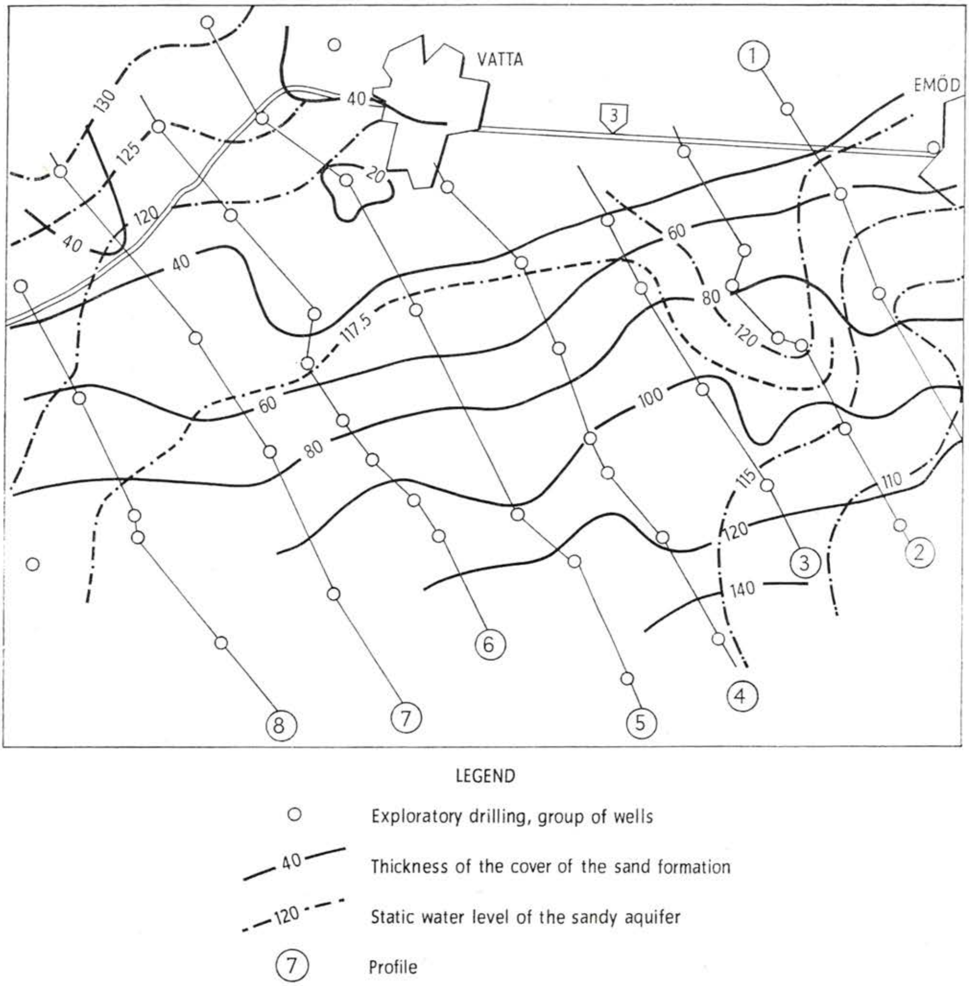

Fig. 6. Structural and hydraulic condition of a Pannonian sand formation (thickness and water level in m a.s.l.)

related to the 40 m thick cover layer is situated about 120 m a.s.l. near the crest of the anticline in the NW part of the region, on the other hand, in the SE part, however, the static water level occurs only about 110 to 115 m a.s.l. where the thickness of cover sediments amounts to 120 m. The variation of the horizontal static water level of the related sand formation is shown in Fig. 7. The screened interval and the static water level of the tested sandy aquifer in the different groups of well is varying according to the Fig. 8 along geologic sections down their dip as shown in Fig. 6.

It should be noted that in the same Pannonian sediment sequence about 20 km apart from here in NE direction near Miskolc, a normal pressure head variation was found. The decreasing potentiometric surface in the direction of the inner part of the Great Hungarian Plain has some practical importance

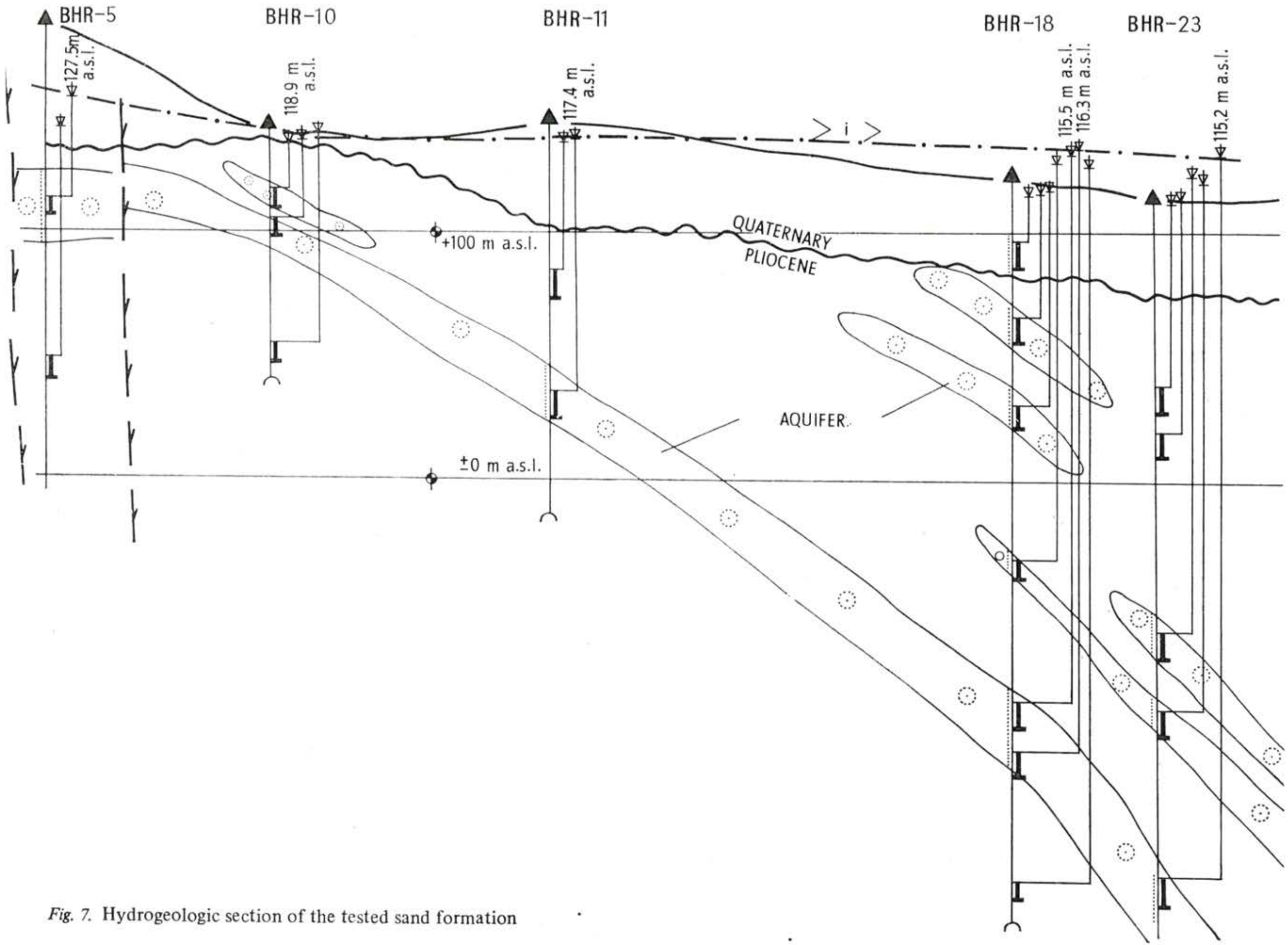

Fig. 7. Hydrogeologic section of the tested sand formation

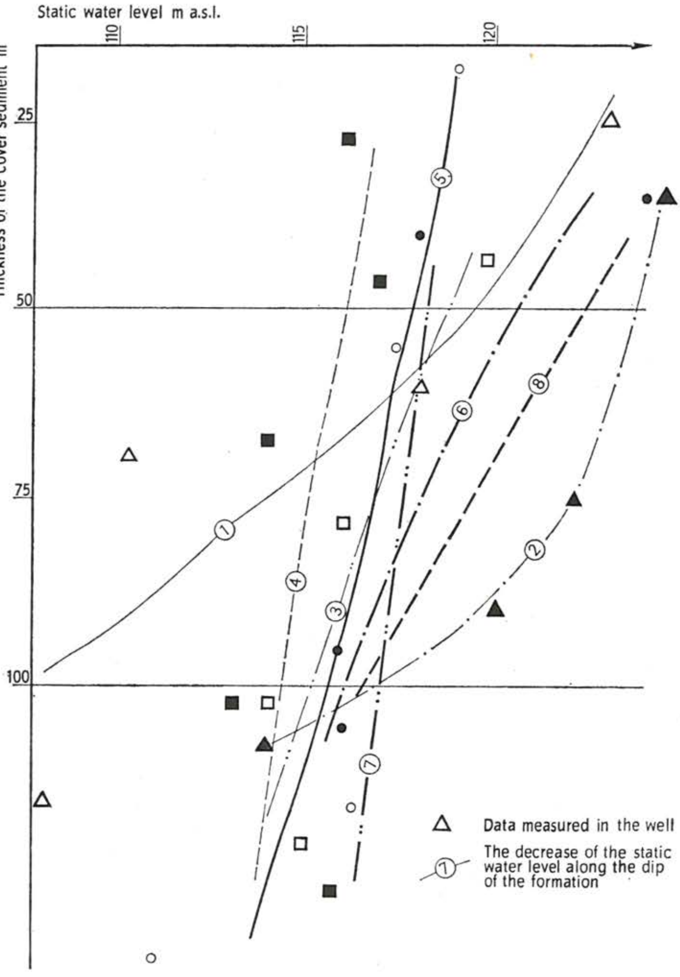

Fig. 8. Change in static water level versus depth measured in the geological section down the dip of the tested sand formation

from the aspects of recharge since according to hydraulic estimations [1] a certain leakage is likely through the basement rocks from the karstic mountains of carbonate formations—situating 10 kms northwards as outcrop—towards the Pannonian sandy aquifer system.

Pressure difference observed in two thermal water wells in the northern border zone of the Great Hungarian Plain

About 10 kms distance westwards from the above mentioned area near to town Mezőkövesd (Fig. 9) in a Pannonian sandy aquifer at depth of 500 m underlying by uplifted Triassic basement rocks explored by petroleum exploration [2], a very high pressure gradient ($4.8 \cdot 10^{-2}$ m/m) was measured accompanied by a static water level of 20 m above the ground surface. On the contrary, above a deep-seated (downwarped) basement block, in the direction of the afore mentioned Vatta—Maklár graben in a well located on the same aquifer the static water pressure was lower by 2 atm ($=9.7\cdot10^{-3}$ m/m) when water temperature, gas content and dissolved solid content were nearly the same. The electrical logs of both well spaced about 350 m shows clearly that until 300 m depth the correlated formations are situated in the same horizon whereas at about 540 m depth the corresponding correlated layers were in deeper structural position (by 20 m) in wells located on the depression than in the wells occurring on the structural high.

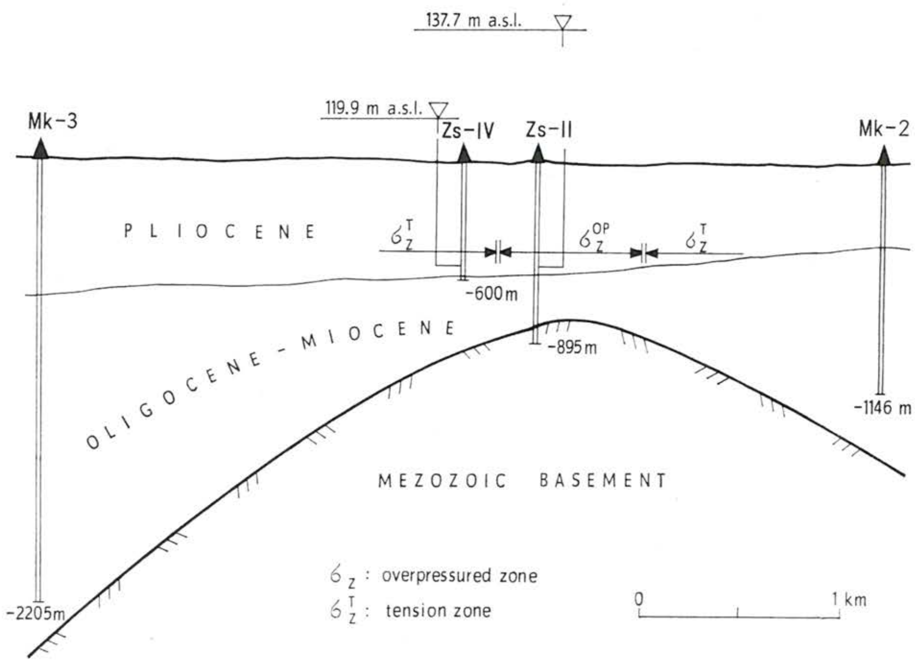

Fig. 9. The uplift of the basement at the thermal bath near Mezőkövesd and the pressure variation of the groundwater

Pressure variation in the groundwater production well network located on the local sedimentary basin of Pécs—Pellérd

Within the groundwater production wells located on the Pannonian sandy aquifers of high water-yielding capacity (with a specific yield of nearly 100 litres/minute.meter) in the southern foredeep ot the Mecsek Mountains are characterized by generally negative pressure gradients. Among the long screened intervals and multiple-zone screened sections (40 to 80 m length), the static water levels of the deeper water-yielding sections are often lower than those of screened at higher horizons. The horizontal trend of such reversed pressure variation are shown by arrows among the pairs of well in Fig. 10.

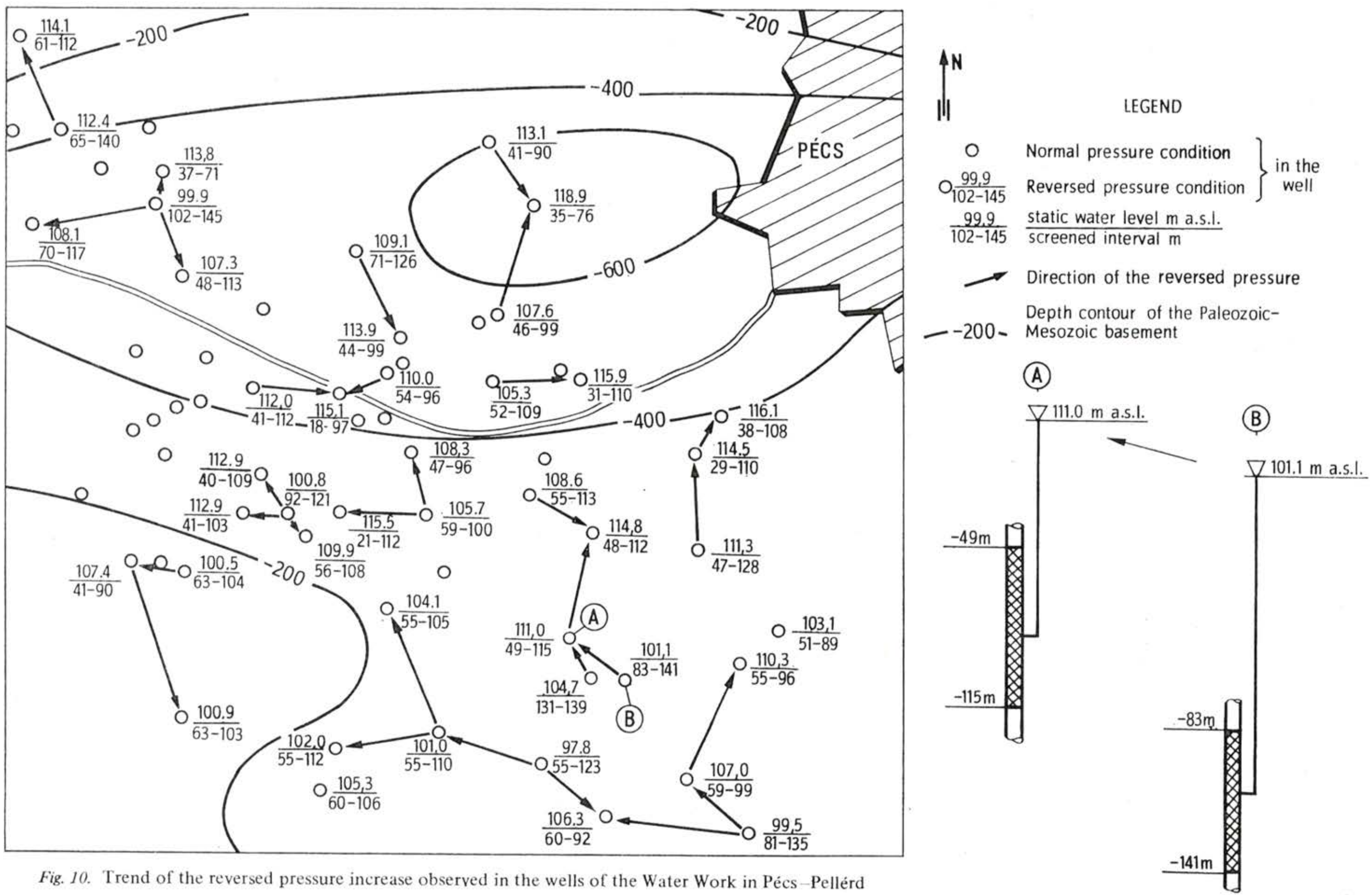

Fig. 10. Trend of the reversed pressure increase observed in the wells of the Water Work in Pécs–Pellérd

A more exact analysis of the subsurface pressure variations is not possible due to the lack of selecting zonal pressure studies and surveys and to the mutual overlapping of long screened sections within the neighbouring wells.

Evaluation and interpretation

The above mentioned examples have proved and indicated that within the shallow unconsolidated water-bearing formations such areas of negative pressure variation can be often encountered where the static water level or static pressure head is diminishing abnormally with depth. A possible reason of it may be the relative position and occurrence of the surface of the basement of high density. Above structural highs sedimentary formations have excess pressure conditions whereas in depressed areas deficient pressures can be found. It is assumed that in depressions the modulus of compressibility (M, kp/cm^2) is decreasing in the layers since a progressive consolidation of the sediments filling the depositional basin will result a pseudo-folding phenomenon and contracted zones will be developed. In the consolidating and thicker sediment sequence within the subsidence a greater subsidence can be expected due to the differencial compaction then in areas of uplifted basement blocks. The contraction of the subsiding flanks will increase, however, the normal weight of the overburden. The formations remained at higher positions will be stretched and pulled over the uplifted blocks due to the inner cohesion of the subsiding and compacting rock mass within the depressed area. This phenomenon is similar to the well known state of soil mechanical process when over the pipe culvert built into the bottom of the levee a higher weight will be exerted than that of the overburden of the earth mass because of the subsidence of these earth masses of greater thickness near to the pipe culvert [4].

This phenomenon will be even increased if there is a difference of the vertical movement between the uplifted and downwarped part of the basement, that is, the uplift of the basement is rising with proper velocity in relation to the consolidation of the formation.

No doubt that the afore mentioned pressure variations may be resulted from other sources [3], such as, the structural conditions, the occurence of a gas cap, recharge along faults, however, after a thorough analysis of these examples will exclude any possibilities of such causes.

The studied abnormal changes in the groundwater pressure heads has also practical implications, such as

a) The specific cost of energy for a unit discharge in the groundwater production will be highly affected by the site of the water withdrawal in relation to the changing pressure conditions (in water works of widely extended well network or in mining water-level subsidence).

b) The mapping of the negative pressure variations may provide important data for a possible natural water household or budget within an aquifer system, for the hydraulic communication with other aquifer systems having higher transmissivities (e.g. karstic formations, alluvial gravel deposits) and ultimately, for the dynamic recharge.

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EXAMEN DES EAUX POTABLES ET THERMALES PRINCIPALES DE HONGRIE POUR LA DÉTERMINATION DE LEUR TENEUR EN BROMIDE, IODIDE ET FLUORIDE

EXAMINATION OF HUNGARY'S MOST IMPORTANT POTABLE AND THERMAL WATERS FOR BROMIDE, IODIDE AND FLUORIDE

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ABSTRACT

Interstice waters of Pleistocene and Upper Tertiary (Upper Pliocene and Upper Pannonian) porous formations, most significant in the country's water supply, were analyzed for the three haloid elements (bromide, iodide and fluoride). A total of about 2000 analyses were taken into consideration.

Such an elaboration was motivated by the important physiological function of the haloids and by the necessity for the discovery of geological-hydrogeological relationships. Serial chemical analyses for elements including haloids are still compulsory to carry out exclusively for mineral and medicinal waters, though potable waters have also been the subject of analyses of this kind in recent years.

With a view to possible consumption in the form of drinking cure, the requirements stipulated by potable water standards have been taken into consideration in the case of thermal waters as well. The admissible limiting values considered were: 0.2 mg/l for bromide, 0.01 to 0.14 mg/l for iodide, 1.0 to 1.5 mg/l for fluoride: all figures being borrowed from experiments by L. I. EL'PINER, J. B. SAFIROV et al.

The three haloid elements were mapped separately for potable waters on one hand and thermal waters on the other. In some diagrams the haloid element content of the Pleistocene and Upper Tertiary formations was plotted as a function of depth.

In the Pleistocene fluviatile sandy sediments the accumulation of any of the haloid components is mostly controlled by local circumstances. The main level of haloid enrichment is the lower part of the Upper Pannonian formations. This level in the individual subbasins has developed at different depths (between 1000 and 2000 m), in dependence on the rate of subsidence.

As a result of the progressive establishment of a freshwater regime the concentration of haloids will usually decrease from the lower horizon of the Upper Pannonian upwards. In the near-surface strata the accumulation of fluoride can be found mostly there where the aquifer is constituted by fine-grained, eventually slightly clayey, sediments. On the national scale, the haloid content of potable waters does not attain, as a rule, the required quantity.

Ce sont deux complexes géologiques qui prennent la plus grande importance dans l'alimentation en eaux potables et thermales du pays: celui de Pléistocène et du Pannonien supérieur, et par endroits celui du Pliocène supérieur (Levantin) sous faciès sableux. A l'Est du Danube, à la Grande Plaine de Hongrie (Alföld), l'acquisition de l'eau potable se fait presqu'exclusivement dans des couches pléistocènes sableuses, épaisses de 300 à 400 m en moyenne, tandis qu'à l'Ouest du Danube, en Transdanubie, ce sont, en prédominance, les niveaux supérieurs des formations pannonien supérieur qui représentent les couches aquifères, sous un mince complexe pléistocène.

Parmi les éléments haloïdes c'est le chlorure que l'on a étudié au plus tôt. De même on analyse depuis longtemps la teneur en iodure et bromure des eaux minérales et thermales.

L'iode joue un rôle grand et important dans des organismes vivants, par suite l'étude de la teneur en iodure de l'eau potable est extrêmement importante.

L'importance du brome est aussi connue depuis longtemps. Le système nerveux central et la thyroïde présentent la sensibilité la plus élevée, mais la quantité du brome, introduite dans l'organisme vivant, influence considérablement aussi le réflexe conditionnel.

A notre pays, c'était K. Than (1880) qui a présenté la première étude sur le fluorure. Il a analysé l'eau du puits thermal de Városliget n° I. (Bois Municipal n° I.) y démontrant 0.069 mg/l de fluorure.

Dans les années de 1940 on a commencé les analyses en nombre plus élevé. En 1957 Sz. Papp a publié, 58 résultats d'études, mais qui ne concernent que les eaux minérales et médicinales, exclusivement. Dans les trois dernières dizaines d'années par l'apparition et l'emploi des micro-méthodes, l'analyse est devenue plus simple, et les valeurs obtenues approchent mieux de la réalité.

Dans les années de 1940 et 1950 les médecins ont constaté que la carie entaire est considérable à certains territoires, tandis qu'ailleurs n'est pas du dout démontrable. En même temps on a pu circonscrire à certains endroits t'émail des dents tacheté, aussi. Ces observations obtenues à ces territoires-ci lurent comparées à la teneur en fluorure de l'eau potable, et on a montré fque la carie dentaire résulte du manque en fluorure et l'émail des dents tacheté dérive de la quantité de fluorure en excédent.

L'Institut National de la Santé Publique (Országos Közegészségügyi Intézet = OKI) a aussi commencé, à la fin des années de 1940, l'étude de la teneur en fluorure des eaux de robinet.

A partir des années de 1960, l'alimentation en eau potable de l'habitation du pays fut de plus en plus résolue par des puits artésiens, explorant les eaux des profondeurs, et parallèlement l'établissement des usines de distribution d'eau a aussi pris son commencement. Mais, hors du chlorure la détermination des autres composants haloïdes n'est obligatoire qu'exclusivement pour les eaux minérales, médicinales et thermales.

Quelques institutions analysent systématiquement l'eau des puits débitant de l'eau potable, et d'après ces analyses, ainsi que sur la base des analyses chimiques obligatoires, nous avons construit les cartes concernant les eaux potables et thermales et les diagrammes illustrant la variation des trois composants en fonction de la profondeur.

Lors la détermination des valeurs de limite des composants, en chaque cas nous avons mis en considération les prescriptions du standard légal des eaux potables. Les prescriptions hongroises étaient directives pour le fluorure et l'iodure. Pour le bromure, d'après les études de L. I. El'piner et de B. Safirov Jn. et al. nous avons pris la quantité de 0.2 mg/l, tolérable dans l'eau potable. Valeurs de limite tolérables pour l'iodure: 0.01 à 0.14 mg/l, pour le fluorure: 1.0 à 1.5 mg/l.

Le bromure et l'iodure dans les eaux potables et thermales

La discussion commune des deux composants est motivée par les théories de genèse, enrichissement et de migration de ceux-ci.

Les hydrocarbures développés dans le complexe sédimentaire lacustre pannonien inférieur déterminent la propriété des eaux interstitielles. Les formations lacustres du Pannonien supérieur s'adoucissent progressivement, et le sédiment lacustre est remplacé par ceux de lac peu profond, deltaïques-fluviatiles, puis fluviatile, alluvial et marécageux. Conformément au processus d'adoucissement, la propriété des eaux interstitielles change, aussi. Dans la zone de passage entre le Pannonien inférieur et supérieur apparaissent des teneurs en bromure et iodure un peu moindres que dans les eaux interstitielles du Pannonien inférieur, mais en chaque cas nous les trouvons en concentration plus élevée par rapport aux niveaux plus supérieurs.

Les cartes (Fig. 3, 4) démontrent que dans les puits implantés aux niveaux supérieurs des couches du Pannonien supérieur (dépression de Jászság, Csongrád et ses environs), le bromure et l'iodure apparaissent en concentration très basse (bromure: 0.02 mg/l, iodure: 0.03 mg/l). Là, où on a atteint ou approché de la limite entre le Pannonien inférieur et supérieur [entre-fleuve du Danube et de la Tisza, dorsales de la Grande-Coumanie (Nagykunság) et Hajdúság] dans les puits profonds de 1 000 à 2 000 m, les teneurs en bromure et iodure ont brusquement augmenté et peuvent approcher de la quantité de 13 mg/l (Hajdúszoboszló).

La sédimentation du Pannonien supérieur transdanubien présente des différences. Certes, les teneurs en bromure et iodure montent à la limite entre le Pannonien inférieur et supérieur ou à la proximité de celle-ci, mais pas en telle mesure qu'à la Grande Plaine de Hongrie. C'est plutôt le complexe pannonien inférieur, où l'on peut observer du changement plus important, le bromure y atteignant même 22.0 mg/l et l'iodure 19.0 mg/l, par rapport aux teneurs moyennes en bromure de 3.0 mg/l, resp. en iodure de 2.5 mg/l des niveaux inférieurs du Pannonien supérieur.

D'après les constatations actuelles, l'oxydation est nécessaire pour le dégagement du bromure et de l'iodure qui pourrait être résultée de l'eau marine riche en oxygène, déjà lors de la sédimentation ou ultérieurement de l'eau descendante de la surface vers les profondeurs. Cette dernière possibilité-ci soit réelle non seulement dans le cas des eaux proches de la surface, mais aussi dans celui des couches aquifères situées plus profondément. L'analyse chimique de l'eau — provenant des formations pliocène supérieur et pannonien supérieur d'un puits thermal (Algyő) de la région sud de la Grande Plaine de Hongrie — semble le justifier :

	Br^-	J^-
Débit d'eau bas	0.05 mg/l	0.33 mg/l
Débit d'eau plus élevé	0.05 mg/l	0.20 mg/l
Débit d'eau maximum	0.10 mg/l	0.21 mg/l

Après le puisage durable — peut-être pluriannuel — on peut supposer l'installation de l'équilibre. L'exigence de temps plus élevé est aussi motivée par ce qu'il s'agit de l'ouverture des couches de sables situées en profondeurs de 1 200 à 1 300 m.

Dans le cas de certaines valeurs plus élevées apparaissant dans quelques lambeaux plus petits, on peut déceler une relation structurale envers des

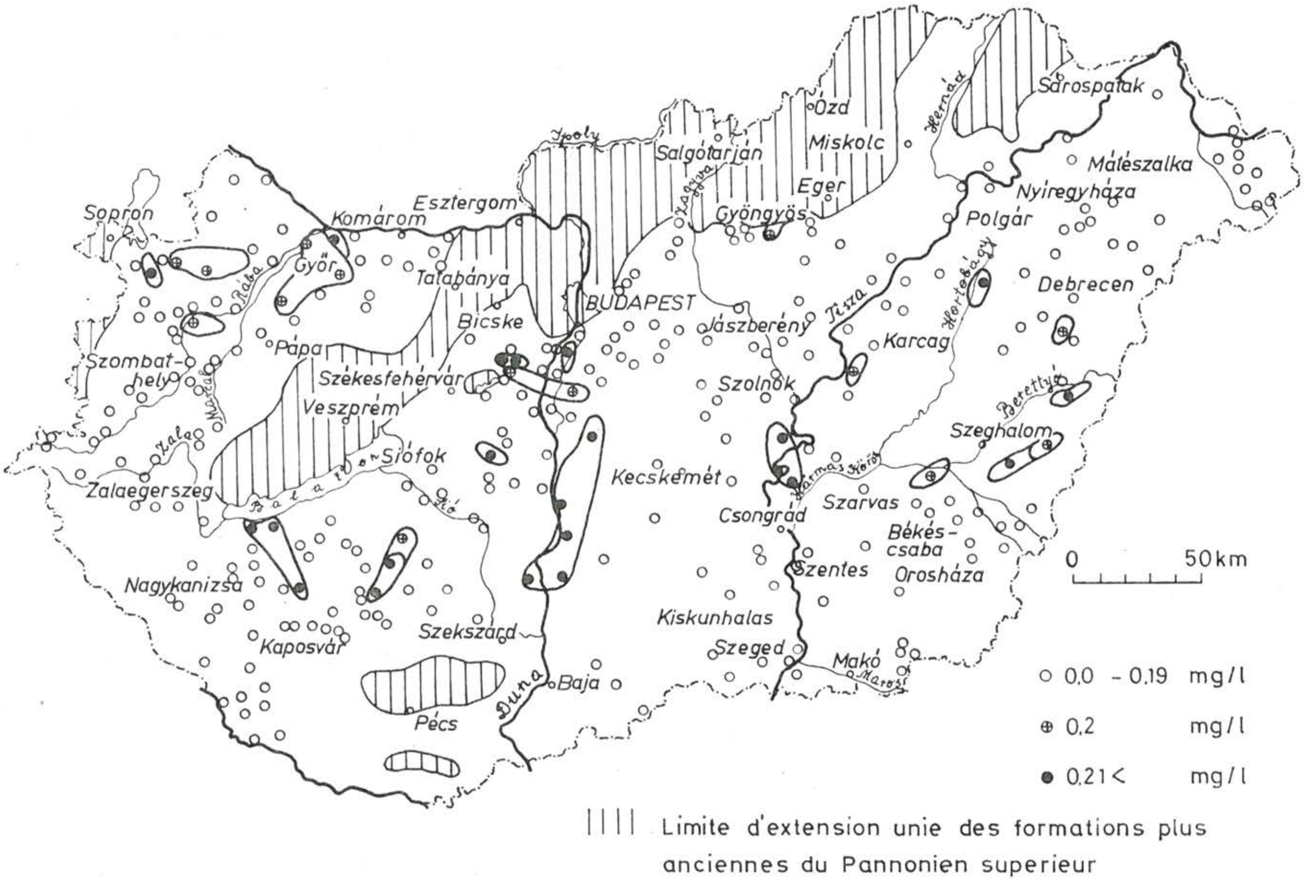


Fig. 1. La teneur en bromure de l'eau potable, dans les couches pléistocènes et pannonien supérieur

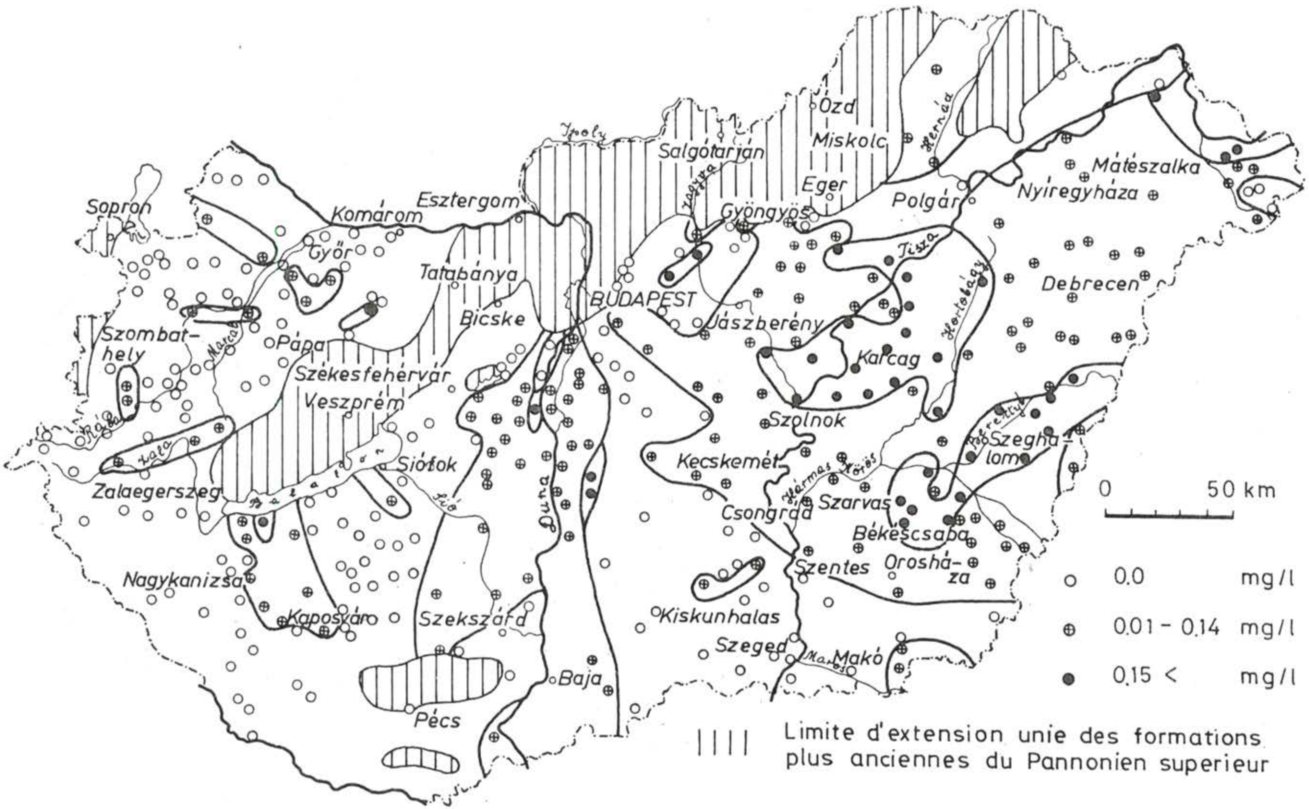


Fig. 2. La teneur en iodure de l'eau potable, dans les couches pléistocènes et pannonien supérieur

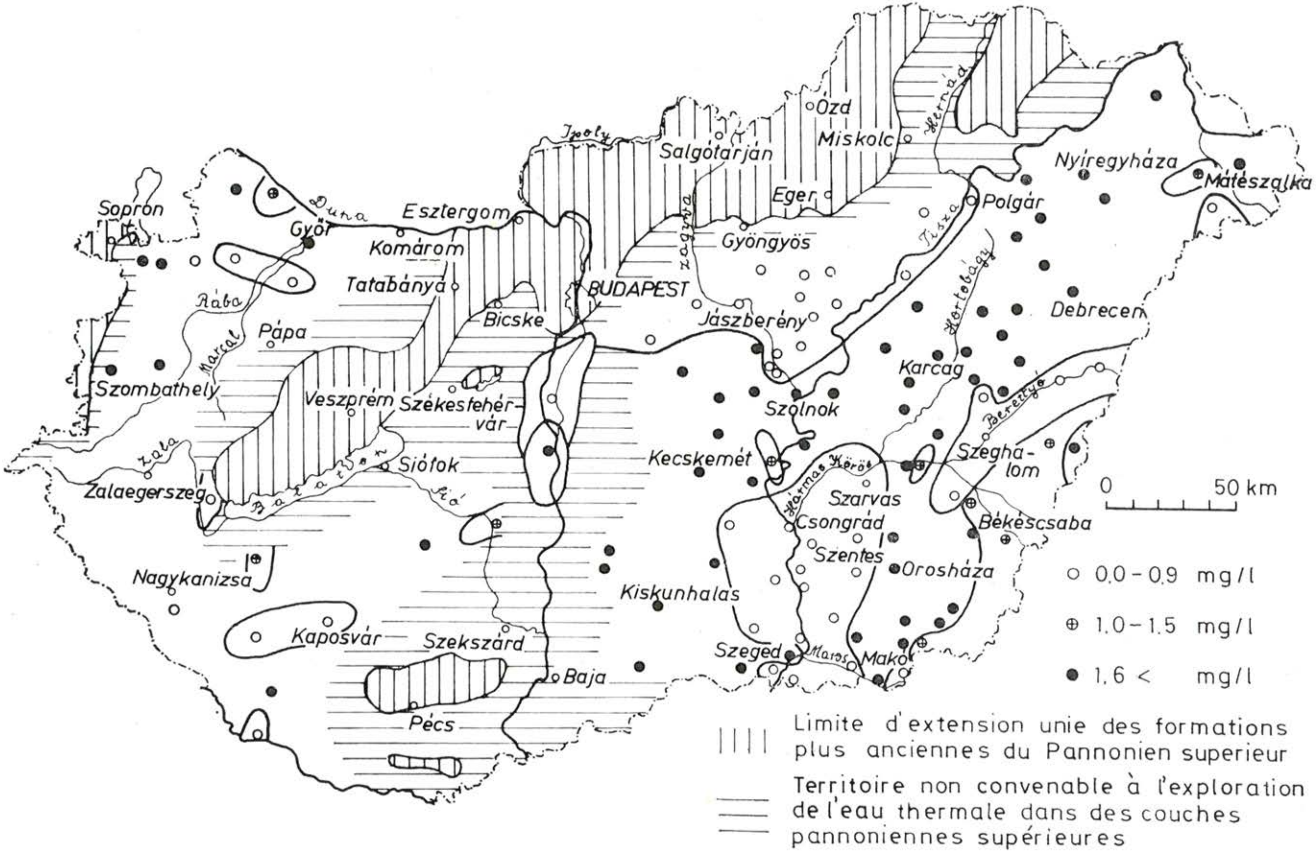


Fig. 3. La teneur en bromure de l'eau thermale, dans les couches pannonien supérieur

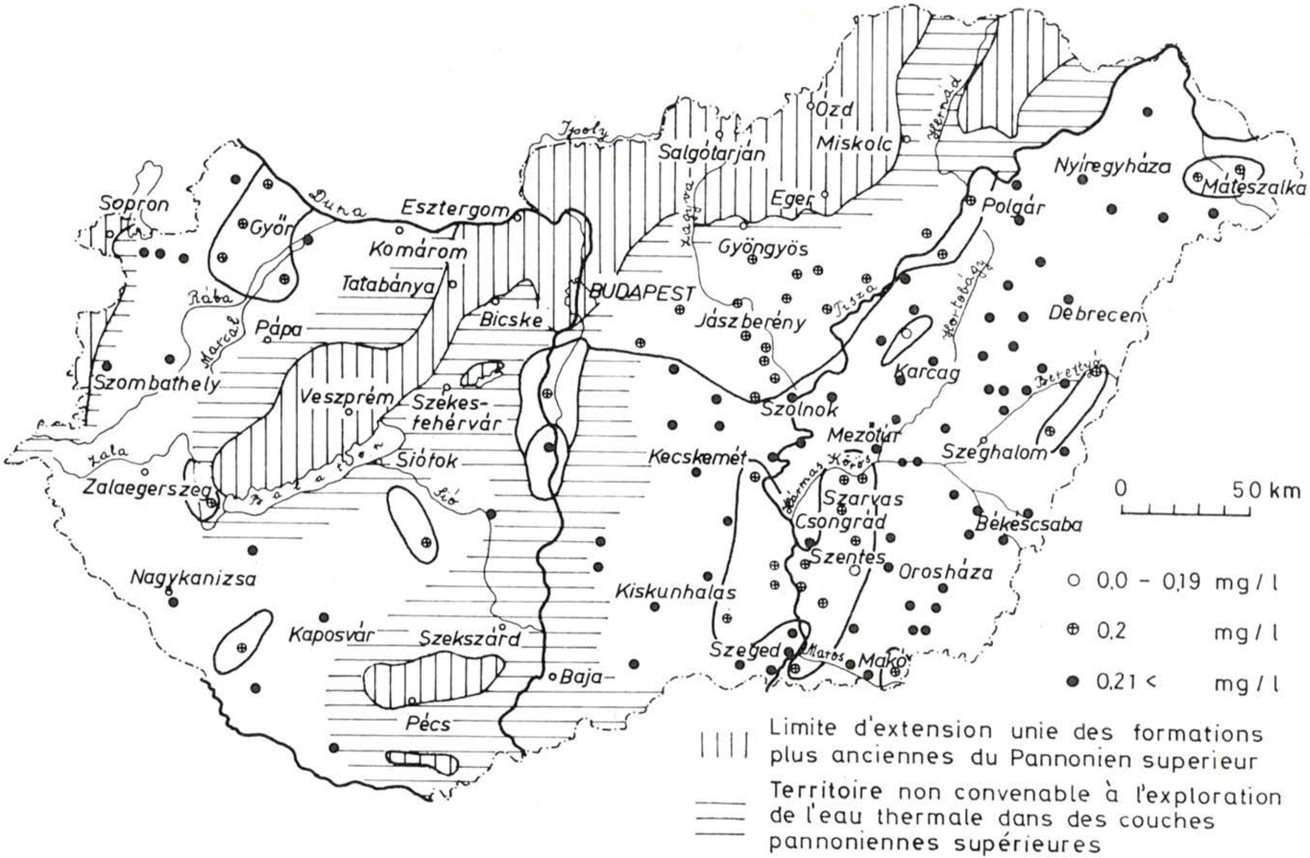


Fig. 4. La teneur en iodure de l'eau thermale, dans les couches pannonien supérieur

couches situées plus profondément. Ce fait est aussi appuyé par la quantité de chlorure plus élevée, le long du Danube. Aux environs du lac Balaton et de la Montagne de Velence, on peut supposer des conditions analogues (Fig. 1).

A la Grande Plaine de Hongrie, la teneur en iodure des nappes d'eau pleistocènes évolue assez contradictoirement. A la région méridionale de la Grande Plaine à conditions hydrogéologiques très favorables, où le ravitaillement (sur-oxydation) pourrait obtenir son maximum, la teneur en iodure ne dépasse pas 0 mg/l. Aux autres territoires (dorsales et dépressions), on doit tenir compte en partie de l'effet de la sur-oxydation, en partie de la migration le long des lignes structurales.

En Transdanubie, la teneur en iodure plus élevée de 0.15 mg/l, apparaissant en lambeaux, pourrait être aussi en relation avec la migration. A la proximité de Szombathely, la teneur en iodure de l'eau potable est, peut-être, due aux gîtes de lignite (Fig. 2).

Le fluorure dans les eaux potables et thermales

Parmi les haloïdes, l'hydro-géochimie du fluorure est le moins connue et éclaircie. Au cours de nos études nous avons constaté que la teneur en fluorure des eaux thermales – pareillement à celles du bromure et de l'iodure – dépend premièrement de la situation microstratigraphique des formations pannonien supérieur. En général, dans les nappes d'eau – emmagasinées dans des formations à position stratigraphique plus inférieure – la teneur en fluorure dépasse 1.5 mg/l. Parallèlement à l'adoucissement des nappes d'eau supérieures, leur teneur en fluorure diminue sous 1.5 mg/l (Csongrád et ses environs, dépression de Jászság). D'après les données sporadiques de la Transdanubie, nous sommes arrivés à la même conclusion (Fig. 6).

La teneur en fluore des eaux potables n'atteint qu'exceptionnellement la quantité optima. Ce sont les communes Békéscsaba, Kardoskút et Csanádalberti qui se distinguent par la teneur en fluorure particulièrement élevée (2.0 à 2.5 mg/l). En considérant que les composants argileux sont capables à l'adsorption de la majeure partie du fluore, on peut bien supposer que le fluore passe facilement en solution, car la vitesse de flux de l'eau est relativement petite dans ces couches.

Dans les couches pléistocènes – en prédominance à grains moyens, la teneur en fluorure est : 0.0 ou 0.3 à 0.4 mg/l (Szeged, Hódmezővásárhely).

En Transdanubie, nous trouvons une valeur marquante au S de la Montagne de Velence, à Seregélyes, où nous avons ouvert – au but de puisage d'eau potable – la couche pannonien supérieur, située entre 430 et 480 m. Ici, la teneur en fluorure a atteint la quantité de 4.0 mg/l. La fluorine de la Montagne de Velence proche pouvait aussi contribuer à l'enrichissement. Deux localités proches pourraient encore appartenir à ce type génétique (Fig. 5).

Coupes typiques de quelques territoires

Malgré les données de 2000 environ, du point de vue des haloïdes l'étude des couches aquifères du Pannonien supérieur et plus récents ne peut être complète. Juste à ce but-ci, nous avons choisi quelques-unes de telles localités,

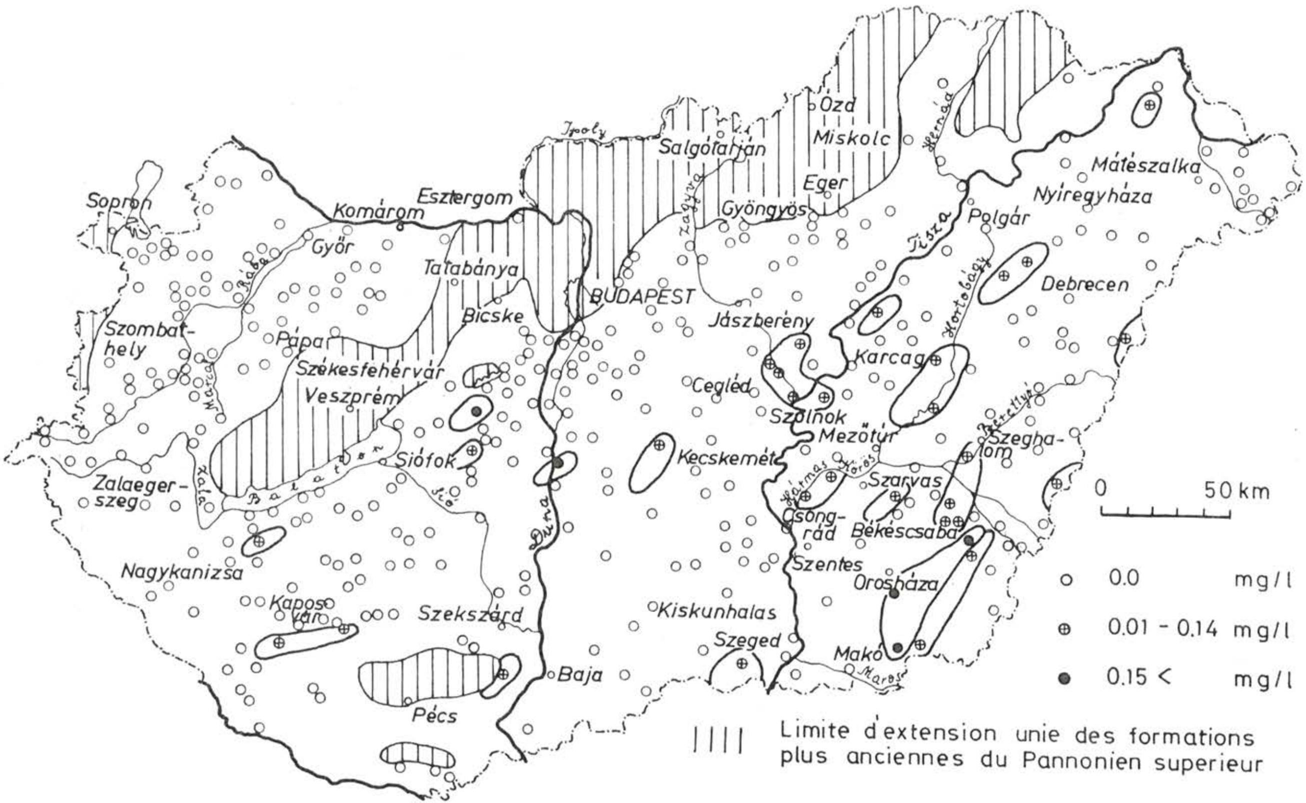


Fig. 5. La teneur en fluorure de l'eau potable, dans les couches pléistocènes et pannonien supérieur

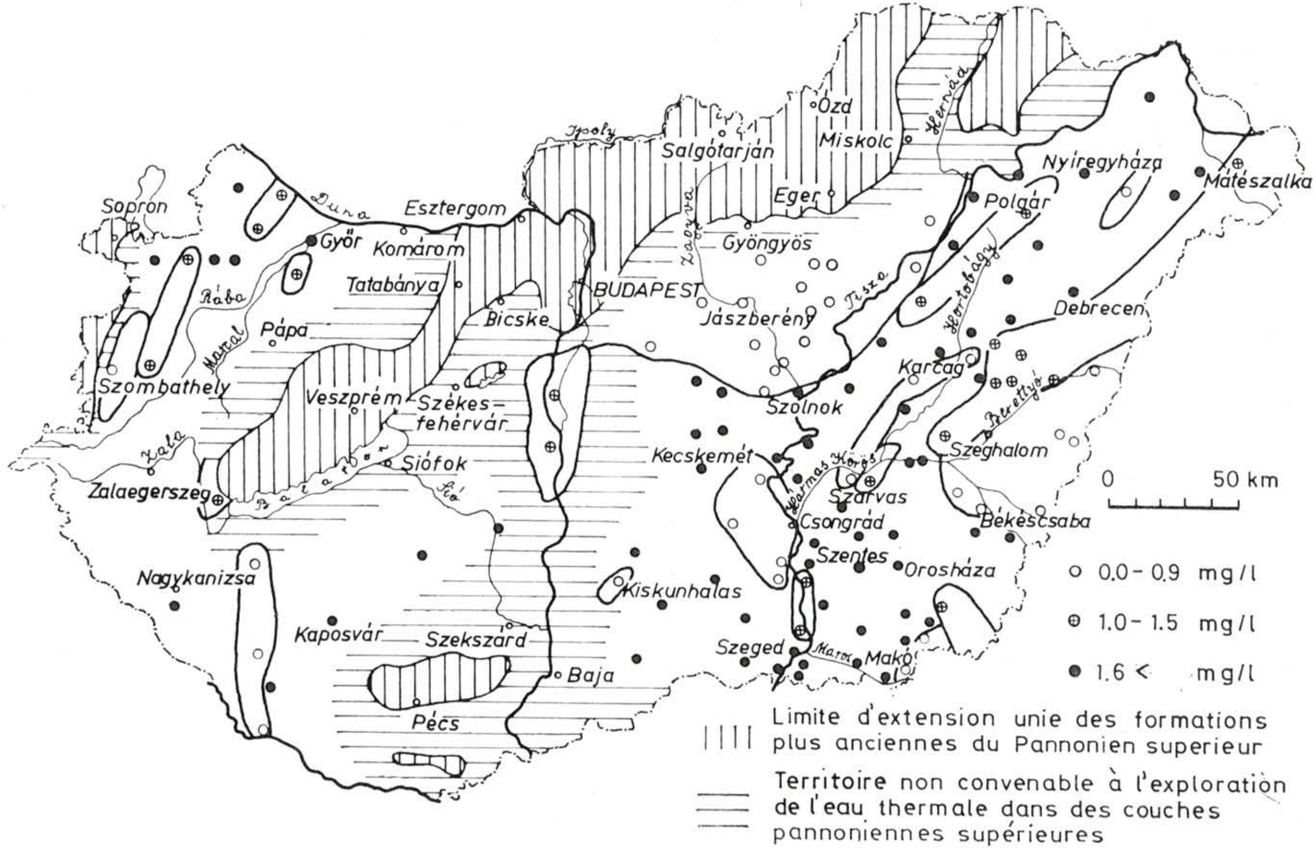


Fig. 6. La teneur en fluorure de l'eau thermale, dans les couches pannonien supérieur

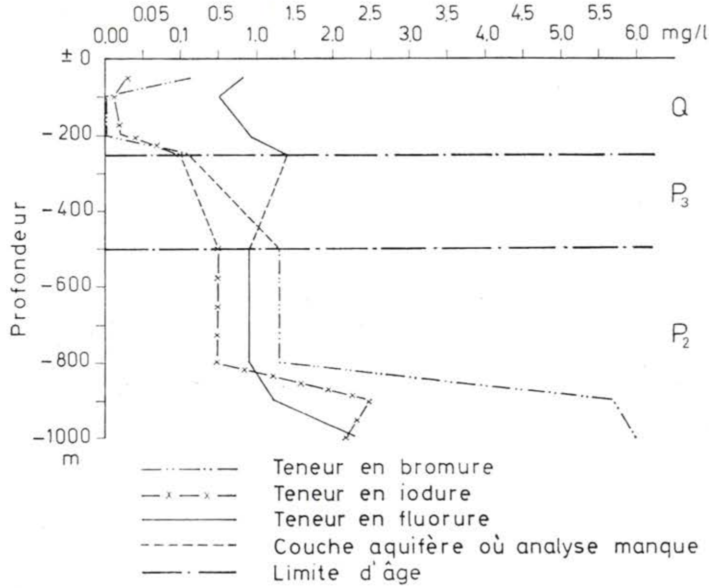


Fig. 7. La teneur en bromure, iodure et en fluorure des eaux potables et thermales de Debrecen

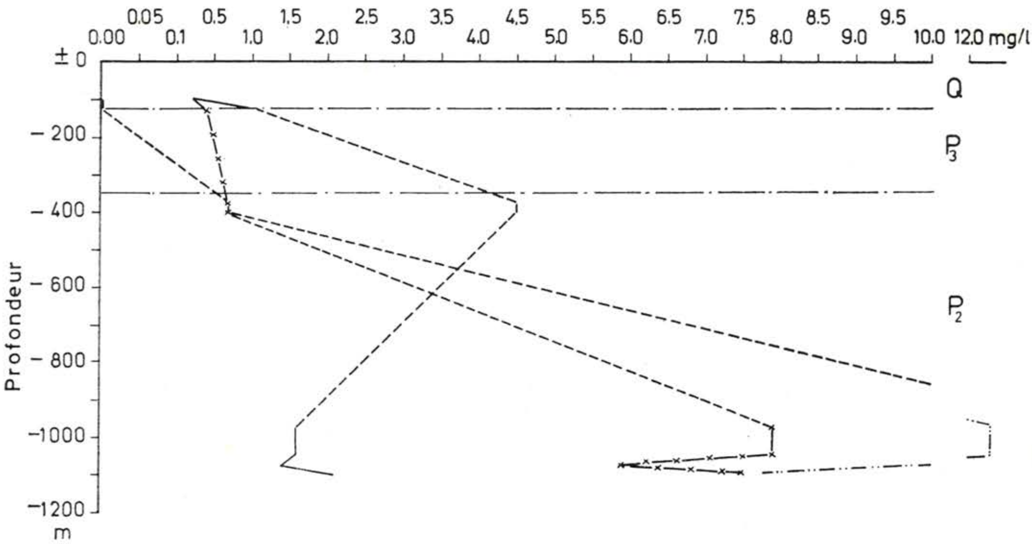


Fig. 8. La teneur en bromure, iodure et en fluorure des eaux potables et thermales de Hajdúszoboszló

Voir la légende de Fig. 7

d'où nous disposons des données — provenant de la plupart des couches à eaux potables et thermales — pour pouvoir illustrer la concentration des trois composants en fonction de la profondeur.

Au territoire de la dorsale de Hajdúság, à Debrecen, au-dessous de la formation quaternaire épaisse de 200 à 250 m environ, jusqu'à la profondeur de 500 m les couches pliocène supérieur (levantines) — à la plupart sous faciès d'argile — surmontent le complexe pannonien supérieur contenant des couches aquifères favorables jusqu'à 1 000 m environ. Dans le Pléistocène, le fluorure et l'iodure présentent des propriétés identiques. Dans la couche ouverte pour l'alimentation (entre 100 et 200 m), la teneur en bromure et le fluorure et l'iodure n'atteignent ou n'approchent pas de la quantité optima. Jusqu'à 800 m, dans le complexe pannonien supérieur les trois composants restent les mêmes. Puis à partir de cette côte la quantité du bromure dépasse plusieurs fois celle du fluorure et de l'iodure (Fig. 7).

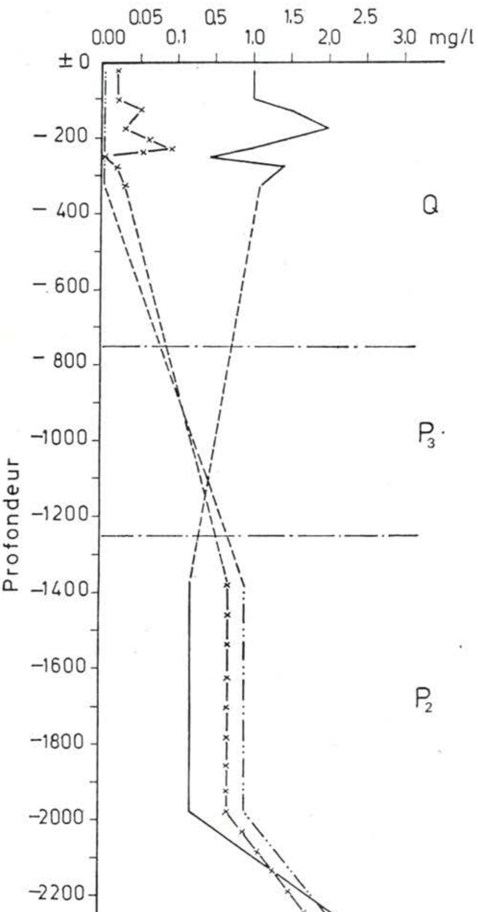


Fig. 9. La teneur en bromure, iodure et en fluorure des eaux potables et thermales à Békéscsaba

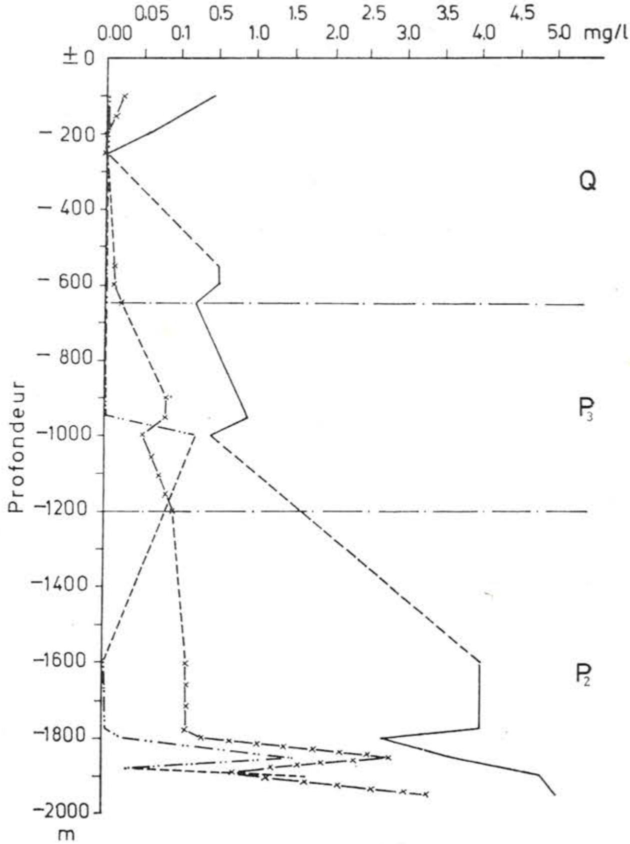
Voir la légende de Fig. 7

Hajdúszoboszló est situé également à la dorsale de Hajdúság et en même temps au territoire de l'un des plus grands champs d'hydrocarbures du pays. Malgré la petite distance, la structure géologique n'est pas identique à celle de Debrecen. L'épaisseur des formations quaternaires est seulement de 100 à 150 m. L'épaisseur du Pliocène supérieur est aussi plus mince à faciès plus sableux, et atteint la profondeur de 330 m, au-dessous de la surface.

A partir du Pléistocène la quantité de tous les trois composants augmente regulièrement, et c'est seul le Pannonien supérieur, où la teneur en fluorure retombe de 4.5 à 1.6 mg/l, autour de 950 m environ, puis elle remonte de nouveau. On peut observer cette tendance inverse aussi dans le cas de l'iodure et du bromure, dans la zone entre 1 000 et 1 100 m (Fig. 8).

A la dépression de Békés — caractérisée par l'affaissement continu et lent — des couches aquifères — à la plupart à grains fins et à petite épaisseur — sont aussi caractéristiques. Dans le Pannonien supérieur, la quantité de tous les trois composants augmente régulièrement (Fig. 9).

Fig. 10. La teneur en bromure, iodure et en fluorure des eaux potables et thermales de Szeged
Voir la légende de Fig. 7



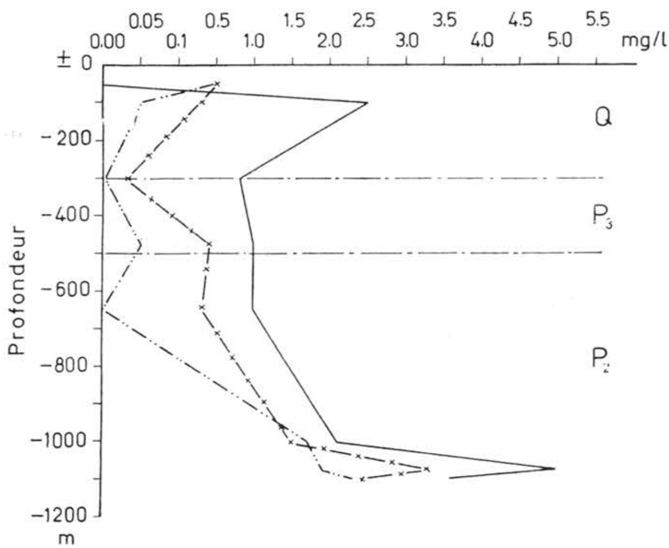


Fig. 11. La teneur en bromure, iodure et en fluorure des eaux potables et thermales de Szolnok
Voir la légende de Fig. 7

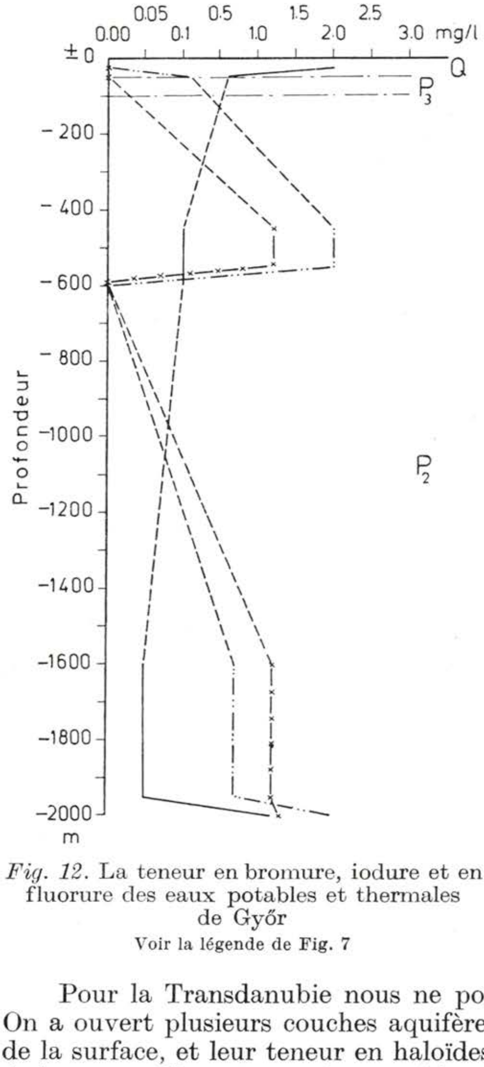


Fig. 12. La teneur en bromure, iodure et en fluorure des eaux potables et thermales de Győr

Voir la légende de Fig. 7

C'est la coupe de Szeged qui présente l'une des plus complètes des séries d'analyses. Ici, tant le complexe quaternaire que ceux pliocène supérieur et pannonien supérieur sont extrêmement sableux. Dans le diagramme il est visible que la teneur en fluorure et aussi celle en iodure augmentent, en général régulièrement, de 250 m environ jusqu'à 1 900 m. A la limite entre le Quaternaire et le Pliocène supérieur, la quantité du fluorure redouble et celle de l'iodure augmente huit fois plus.

La teneur en bromure est 0.0 mg/l aussi dans les couches aquifères pléistocènes que dans la zone supérieure du Pliocène supérieur. A 950 m elle monte brusquement à 0.2 mg/l, puis — dans la zone de 1 600 à 1 800 m du Pannonien supérieur — elle diminue de nouveau à 0.0 mg/l, et ne commence à remonter — ensemble avec l'iodure et le fluorure — qu'à partir de 1 800 m (Fig. 10).

A Szolnok, à l'aide du changement quantitatif du fluorure, bromure et de l'iodure, on peut bien désigner les limites entre le Pléistocène et le Pliocène supérieur, resp. entre le Pliocène supérieur et le Pannonien supérieur. La réhausse quantitative brusque apparaît à 1 000 m (Fig. 11).

Pour la Transdanubie nous ne pouvons présenter que la coupe de Győr. On a ouvert plusieurs couches aquifères dans le complexe de graviers proche de la surface, et leur teneur en haloïdes est très variée. La teneur en haloïdes de la nappe d'eau pannonien supérieur — située entre 450 et 600 m — est à peu près identique à celle de la nappe la plus inférieure (entre 1 950 et 2 000 m). Cette dernière-ci représente le niveau le plus inférieur du Pannonien supérieur (Fig. 12).

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STABLE ISOTOPES AS INDICATORS OF THE ORIGIN AND ZONAL AFFILIATION OF DEEP-SEATED GROUND WATERS IN SEDIMENTARY BASINS

ISOTOPES STABLES COMME INDICE DE LA GENÈSE ET DE L'APPARTENANCE AUX ZONES HYDROCHIMIQUES DES EAUX PROFONDES DANS LES BASSINS SÉDIMENTAIRES

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RÉSUMÉ

Le présent travail concerne les eaux souterraines (pour la plupart les saumures à concentration diverse) qui se trouvent dans les zones profondes des bassins sédimentaires. Le problème bien compliqué de la genèse de ces eaux et de leur composition chimique peut être résolu, au moins partiellement, à l'aide de déterminations à spectromètre de masses de la composition isotopique de l'hydrogène et de l'oxygène dans l'eau. Les résultats permettent de distinguer les eaux marines, atmosphériques ou mixtes (hybridales) et, en même temps, indiquent la zone verticale hydrogéologique dans laquelle elles se trouvent. En se basant sur l'équation générale pour les eaux météoriques ($\delta D = a\,\delta^{18}O + b$) et admettant que la composition isotopique des eaux océaniques n'a pas changé, au moins depuis le Paléozoïque supérieur, on peut utiliser la notion proposée ici du degré de résidualité des eaux souterraines. Elle peut être comprise comme la part de l'eau marine résiduelle présente dans l'eau examinée et exprimée en pourcent comme l'indice isotopique de résidualité (IIR).

Une série de facteurs qui peuvent compliquer l'interprétation des mesures isotopiques doit être prise en considération. Les plus importants parmi eux, comme l'évaporation, l'échange isotopique entre le roche et les eaux, teneur en cations et l'ultrafiltration, sont discutés dans le présent article.

Les considérations théoriques sont illustrées par les conclusions des résultats des déterminations isotopiques dans les eaux des bassins sédimentaires de Pologne. On peut observer ici, comme règle, l'augmentation de la teneur en isotopes lourds parallèle à l'augmentation de la salinité (tous les deux augmentant avec la profondeur). Ces résultats confirment l'idée de la zonalité verticale et en même temps démontrent que le degré de résidualité augmente avec la profondeur.

The origin of waters occurring in deep sedimentary aquifers is one of the most discussed problems of hydrogeology and the publications dealing with it, both in regional and theoretical sense may be counted by the hundreds. In most of them the chemical composition of waters is considered as an all-important tool to answer the above-mentioned question. Only during the recent few years, the rapidly developing isotope techniques have indicated new directions for the research and have become the most effective method as far as the problem of the origin of the so-called formation waters is concerned.

The basic assumption on which the following considerations are based is the existence of two principal genetic types of ground waters. The first type includes waters of atmospheric origin, participating in the present hydrological cycle or precipitation having infiltrated during past geological periods (palaeo-infiltration waters). The latter group may be conventionally related to Pleistocene or pre-Quaternary infiltration (J. DOWGIALLO 1971). The second type includes waters of marine origin, contained within the interstices of sedimentary layers since their deposition onto the sea bottom (strictly speaking: connate waters) or sea waters having entered consolidated formations during transgression periods expulsing from there the possibly present ground waters of atmospheric origin.

According to the above-presented classification all ground waters except those taking part in the actual hydrological cycle, i.e. having infiltrated during the latest 10,000 years, may be designated as fossil or relict waters, coming also under the category of connate waters as re-defined by D. E. WHITE, 1965. Though waters of atmospheric origin may, in certain regions, play an important role within this group, marine waters seem to be here the most widespread genetic species. Taking into account the possible diversity of meaning of the terms "connate" and "fossil", the notion of relict water will be used in this paper as far as ground waters of marine origin or atmospheric waters of pre-Holocene infiltration are concerned. Instead, the term of meteoric water will be reserved for waters participating in the contemporary hydrological cycle.

The existence of pure relict waters, the majority of them being undoubtedly of marine origin is hardly possible except, may be, waters in young (Cainozoic) sediments of depression basins or in the deepest part of the whole sedimentary sequence. The major part of deep-seated ground waters is probably mixed, containing both meteoric and relict waters in all possible proportions, according to the geological and hydrological features of a given region. Thus, a third genetic type must be distinguished, namely hybrid ground waters. With a certain simplification due to leaving out of account palaeo-infiltration waters, the hybrid type may be defined as being composed of marine waters and meteoric waters of the Holocene age.

Relict waters in sedimentary basins are almost always saline, their mineralization increasing with depth. A great number of investigators have tried to establish chemical indicators permitting to state, whether the salinity originates from the salt content of the ancient sea water or from leaching of evaporites by meteoric waters. It has to be noted that these endeavours have not yield satisfying results. The reason of it is not only the high complexity of the problem but, to a certain degree its incorrect methodological setting. As it has been recently observed by SELECKIJ et al., 1973, the notion of "the origin of the chemical composition of ground water" is often incorrectly replaced by the notion of "the origin of ground water". The latter notion must be limited to the solvent (H_2O) which during its residence within, or migration through the rock body acquires its chemical composition as a result of the whole complex of lithologic, thermodynamic and hydrodynamic factors. Thus, the chemical composition of ground waters used very often as a basis for defining of their genesis has to be considered as a deceptive criterion. Instead, the isotopic lebelling of the water molecule allows, at least in certain cases, to define its atmospheric or marine origin and as far as hybrid waters are concerned gives an idea of the proportion of both genetic types forming the liquid considered.

As observed by P. A. DICKEY, 1966, the increase of water's mineralization with increasing depth is a world-wide phenomenon. It had been explained in many different ways and the dispute among the representatives of various theories (e.g. leaching of evaporites, diffusion, membrane filtration etc.) is still being continued.

The increase of mineralization with depth is accompanied by the tendency to changes in the chemical composition of water according to the SULIN–IGNATOVITCH sequence (H. SCHOELLER, 1959) based on the anions prevailing may be schematically presented as:

$$HCO_3—HCO_3+SO_4—SO_4—SO_4+Cl—Cl$$

This sequence is the basis of the vertical hydrochemical zonality theory, according to which the underground hydrosphere may be divided into three zones: the upper zone of intensive circulation and exchange (HCO_3 and SO_4 waters), intermediate zone of hindered exchange (SO_4 and Cl waters) and the lower zone, where stagnant or extremely slowly moving Cl waters occur. Using the above defined terms it may also be assumed that within the upper zone only meteoric waters occur, the intermediate zone is characterized by the occurrence of hybrid waters, while within the lower zone relict waters are dominating.

The above-mentioned hydrochemical sequence by no means answers the question concerning the origin of waters occurring in the particular zones except the upper one, where the atmospheric origin of ground water is evident. Brines occurring within the intermediate and lower zones may be as well marine connate waters as meteoric ones, the latters owing their concentration to the leaching of evaporites. The universally observed unification of the chemical composition of waters at considerable depths is due to thermodynamic factors within the rock-water system and to some specific phenomena such as ion exchange and membrane filtration. The latter process may lead to a far-reaching assimilation of the chemical composition of membrane filtered marine water and membrane concentrated meteoric waters (both terms proposed by D. E. WHITE, 1965). In such a case only isotopic studies may give an appropriate way for deciphering the history of the fluid.

The deuterium and oxygen-18 content of water are commonly expressed as the deviation in per mills of this content in the international standard SMOW (Standard Mean Ocean Water), defined by H. CRAIG (1961a). All precipitation waters and the majority of continental surface waters have δ values negative in relation to SMOW, which by definition equals O. There occurs here a close relation between δD and $\delta^{18}O$ which for the northern hemisphere is expressed by the equation: $\delta D = 8\delta^{18}O + 10$ (H. CRAIG, 1961b), and when generalized has the form $\delta D = a\,\delta^{18}O + b$. The temperature at which the vapour condensation occurs is the chief factor controlling the fractionation of oxygen and hydrogen in precipitation waters. As the temperature is correlated with altitude and latitude, particular regions, even not large ones are characterized by a typical mean isotopic composition of precipitation, strictly related to the mean local annual temperature (W. DANSGAARD, 1964).

During infiltration into the ground, there does not occur any isotopic fractionation of oxygen and hydrogen, and ground waters within the upper zone of intensive exchange have the same isotopic characteristics as the precipitations within the catchment area. Thus, the meteoric genetic type of water may be easily distinguished. Additional methods which may corroborate the participation of ground water in the actual hydrologic cycle are measurements of the tritium and ^{14}C content (the latter one in bicarbonates).

The deeper parts of the underground hydrosphere are a zone of mixing of meteoric waters with relict ones. Taking into account that the former type has isotopic characteristics similar to local precipitations, while the latter one should be characterized by δ values close to 0, the share of the particular genetic types may be found graphically (Fig. 1), connecting on the plot δD versus $\delta^{18}O$ the precipitation point with the SMOW point and obtaining thus the mixing line (B. HITCHON, I. FRIEDMAN, 1969). The position on the mixing line of the point corresponding to the water considered shows the share of the meteoric and relict (SMOW) waters in the compound.

The above presented reasoning does not take into account the possible presence of palaeo-infiltration water as a component of relict water. Therefore, it may be valuable only in such cases, in which a detailed palaeohydrogeological analysis excludes the presence of this component. Obviously, it has to be assumed, too, that the isotopic composition of waters in open seas of the geological past did not differ fundamentally from the SMOW. This assumption seems to be justified at least as far as Cainozoic and Mesozoic seas are concerned (A. P. WINOGRADOW, 1967).

The perfect fit of a point corresponding to a given formation water to the mixing line indicates that there were no processes leading to a differential enrichment in oxygen-18 or deuterium. In the former case the point would lie under the mixing line, in the latter one — above it. Differential enrichment in oxygen-18 considered by introducing a correction for the cations content (Z. SOFER, J. R. GAT, 1972), is a particular common feature of waters occurring in deep sedimentary aquifers. It has been reported among others from the USA (R. N. CLAYTON et al., 1966), Canada (B. HITCHON, I. FRIEDMAN, 1969), Poland (J. DOWGIAŁŁO, 1973) and the USSR (W. G. TRATCHUK et al., 1975). This phenomenon may be due to the isotopic exchange between water and marine carbonates, occurring even at moderate temperatures. In certain circumstances it may also be due to the presence of connate waters of marine evaporites, since the evaporation processes enrich the oxygen-18 content of surface waters. Besides, ultrafiltration processes may produce an enrichment of membrane-concentrated waters in oxygen-18 compared with the deuterium as shown recently by T. COPLEN and B. B. HANSHAW (1973).

The major factor of differential enrichment of water in oxygen-18 i.e. the isotopic exchange with the water-bearing rock stratum does not play such an important role as far as deuterium is concerned. The enrichment in deuterium may be caused by the exchange with H_2S if occurring in considerable quantities or, eventually, with hydrocarbons. Ultrafiltration may also play here a certain role. However, the enrichment processes concerning deuterium occur on a rather small scale, and for our reasoning the deuterium content of a marine relict water may be considered as such undergoing changes caused only by mixing with meteoric water. Thus, contrasted with the oxygen-18 content easily shifting towards positive values, the deuterium content of water can be used as a mixing-reference quantity. Therefore, if the water considered is enriched in oxygen-18 and the corresponding point is situated below the line of proportional mixing, the share of relict water may be found graphically, by connecting this point with the mixing line by a straight line parallel with the oxygen-18 axis (Fig. 1). This practice cannot be used in the

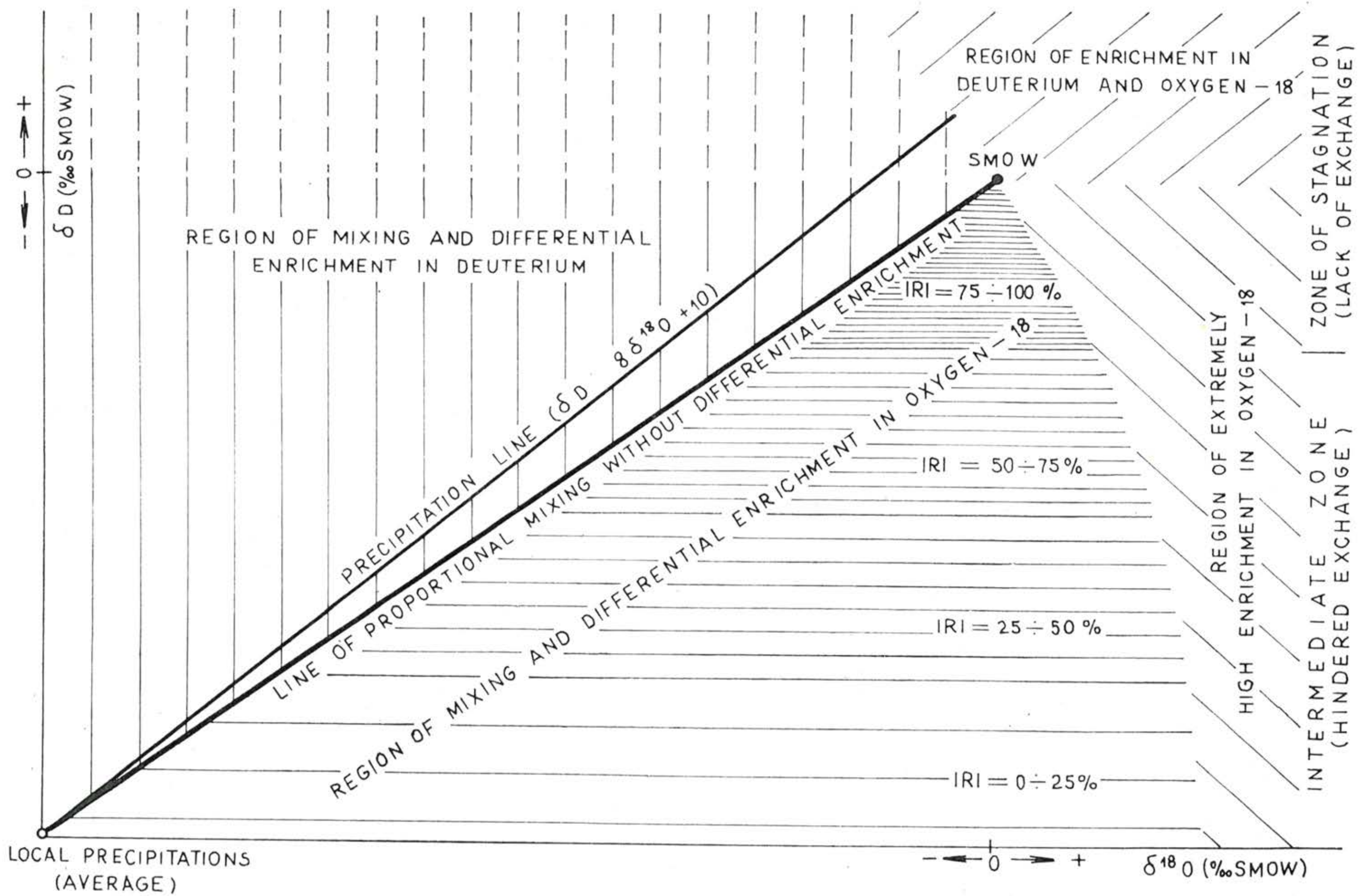

Fig. 1. Schematic diagram indicating the IRI values in connection with the vertical zonality of ground waters (see text for details)

case of a differential enrichment in deuterium, because it is not known to what a degree the oxygen-18 content has increased simultaneously.

The share of relict marine water in the mixed water may be expressed in per cent by means of a simple equation:

$$M=\left(1-\frac{\delta D_f}{\delta D_p}\right)\cdot 100\%$$

where M is the share of marine water, δD_f is the δD value of the formation water considered and δD_p is the δD mean value of the atmospheric precipitations in the region under consideration. The condition of applicability of this formula is $\delta D_f \leq 0$.

The value M is proposed here to be called the *isotopic relictivity indicator* (IRI). It may be used for comparative studies on waters sampled in different parts and aquifers of sedimentary basins, of a rather simple tectonic structure, where marine sediments distinctly prevail over the continental ones and where the average isotopic composition of precipitations, recharging particular aquifers may be easily determined.

Once a quantitative expression for the degree of relictivity is admitted, the IRI may be easily connected with the vertical zonality of ground waters. For the upper zone its value is 0, for the intermediate zone it is assumed arbitrarily: $0 > \mathrm{IRI} \leq 75$, for the lover zone: $75 < \mathrm{IRI} \leq 100$. These ranges, of course, have to be established individually for each basin on the basis of their hydrogeological features and of the available hydrochemical data.

It has to be mentioned here, that an attempt to evaluate quantitatively the proportions of the relict and meteoric compounds in deep-seated waters has been done by W. S. Brezgunow et al. (1966). Taking into account the salinity and deuterium contents, these authors have distinguished in the Lower Cretaceous strata of the Azow-Kuban depression (U.S.S.R.) 3 vertical zones within which the content of marine relict water would be 70%, 30% and 0% respectively. Though the isotopic results obtained by means of the densimetric method might be dubious, the idea of attributing the increase of deuterium with depth to the increasing share of the relict compound remains valuable. Also A. E. Babiniec et al. (1968), have tried to evaluate the content of relict marine water in thermal brines of artesian basins taking into account the $^{18}O/^{16}O$ ratio and using the Br/Cl ratio as an indicator of the primary salinity of sea water in the accumulation period of connate waters. As the oxygen-18 content of water may be considerably influenced by the isotopic exchange with rocks, the applicability of this method is probably very limited.

Studies on the isotopic composition of deep ground waters in the sedimentary platform basins of the Polish Lowland have shown that the following general rules may be established if the isotopic composition of oxygen and hydrogen is considered in relation to the depth of the aquifer and to the degree of the mineralization in its waters:

a) The characteristic points of the waters considered are located either on the precipitation line or on the mixing line, the oxygen isotopic shift being negligible or small, but higher in deeper aquifers and more saline waters.

b) The concentration of the heavy isotopes of oxygen and hydrogen increases with the depth of aquifer.

c) The concentration of the heavy isotopes of oxygen and hydrogen increases linearly with the increase of water mineralization.

These features have led the writer (J. DOWGIALLO, 1971) to the conclusion, that the Mesozoic formations of the Polish Lowland contain hybrid waters occurring in various proportions. This opinion is different from that prevailing up to now, according to which the salinity of waters in these formations is due to the leaching by waters infiltrating into the Zechstein salts, which underlie the Mesozoic strata and form salt domes, sometimes percolating through younger sediments. Conclusions on the presence of waters of mixed origin have been drawn also by A. RÓZKOWSKI and K. PRZEWLOCKI (1974) with regard to the Palaeozoic formations of the Lublin coal basin. The results obtained in Poland are in good agreement with the isotopic-chemical zonality scheme, discussed above.

Concluding, the author would like to emphasize that the proposed isotopic relictivity indicator is to be considered as a semi-quantitative feature of ground water. The possibility of differential enrichment of water not only in oxygen-18 but also in deuterium as a result of isotopic exchange with hydrogen-containing substances or of membrane filtration makes it necessary to use this indicator with extreme caution. Every attempt of evaluation of the relict marine water share in a hybrid fluid has to be accompanied by a penetrating analysis of the hydrogeological features of the aquifer concerned and of the chemical characteristics of water. It has also to be pointed out that the IRI may be used only in selected hydrogeological units. Large artesian basins seem to be the most appropriate objects for such analysis.

The practical usefulness of the IRI consists, among others, in the possibility to establish the zonal affiliation of water in a given aquifer. The possible revealing of the lower zone is of importance for oil prospecting, as its deposits are often connected with such zones.

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PARTICULARITÉS CHIMIQUES DES EAUX ARTÉSIENNES DU NORD D'ORADEA (ROUMANIE)

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Le présent ouvrage s'occupe de la zone de plaine (représentée par la marge orientale de la dépression pannonienne) limitée par le ruisseau Crișul Repede au Sud et la vallée de Ier au Nord, ainsi que comprise les basses-plaines de certaines vallées qui traversent la zone de colline limitrophe, dont la constitution géologique renferme des dépôts quaternaires, tertiaires et mésozoïques, au-dessus d'un soubassement cristallin.

Dans les dépôts susmentionnés les forages exécutés ont fait ressortir l'existence de quelques couches et complexes aquifères, qui présentent un caractère ascensionel, même artésien, dans la plupart des cas. Parmi ceux-ci, nous rapellons : le complexe aquifère des formations mésozoïques et les couches aquifères des dépôts miocènes et pliocènes supérieurs.

Dans le réseau carstique des dépôts mésozoïques, le complexe aquifère accumulé présente un aquifère d'importants débits et une haute thermalité. Cet aquifère est bien connu, grâce aux recherches hydrogéologiques entreprises par E. Liteanu et al. (1965), E. Liteanu et Gh. Vasilescu (1967), Gh. Vasilescu et Gr. Nechiti (1968), à la suite de l'exécution de quelques forages à grande profondeur.

A leur tour, les couches aquifères dans les dépôts miocènes et pannoniens ont un débit le plus souvent réduit et un chimisme non adéquat pour des eaux potables.

Dans la partie ouest de la dépression pannonienne, bien des puits artésiens exploités pour l'alimentation en eau, sont emplacés dans les horizons perméables des dépôts pliocènes supérieurs; ces derniers ont constitué l'objet de nos recherches.

Caractéristiques hydrogéologiques générales

Un horizon prépondérant sablonneux, de 100-150 m d'épaisseur, constitué de sables fins et moyens à intercalations d'argiles et de marnes repose au-dessus d'un complexe lithologique pannonien prédominant marneux, qui dans la région de la ville d'Oradea arrive à des épaisseurs de 350-600 m et dépasse souvent 1000 m dans la région de Borș et de Biharia. La succession se continue avec un horizon argileux charbonneux, formé d'argiles et de marnes fréquemment sablonneuses à intercalations de lignites et de sables argileux avec une épaisseur de 80 à 150 m et s'achève avec des dépôts pséphitiques épais de 40-60 m.

Les recherches effectuées sur la base de nombreux forages à petite profondeur nous ont porté à des conclusions que tant l'horizon sablonneux que celui argileux charbonneux reviennent au Pliocène supérieur — qui correspondrait au Lévantin s.l., selon la littérature de spécialité hongroise, ou à l'intervalle dacien-lévantin de la zone extracarpatique — renfermant également

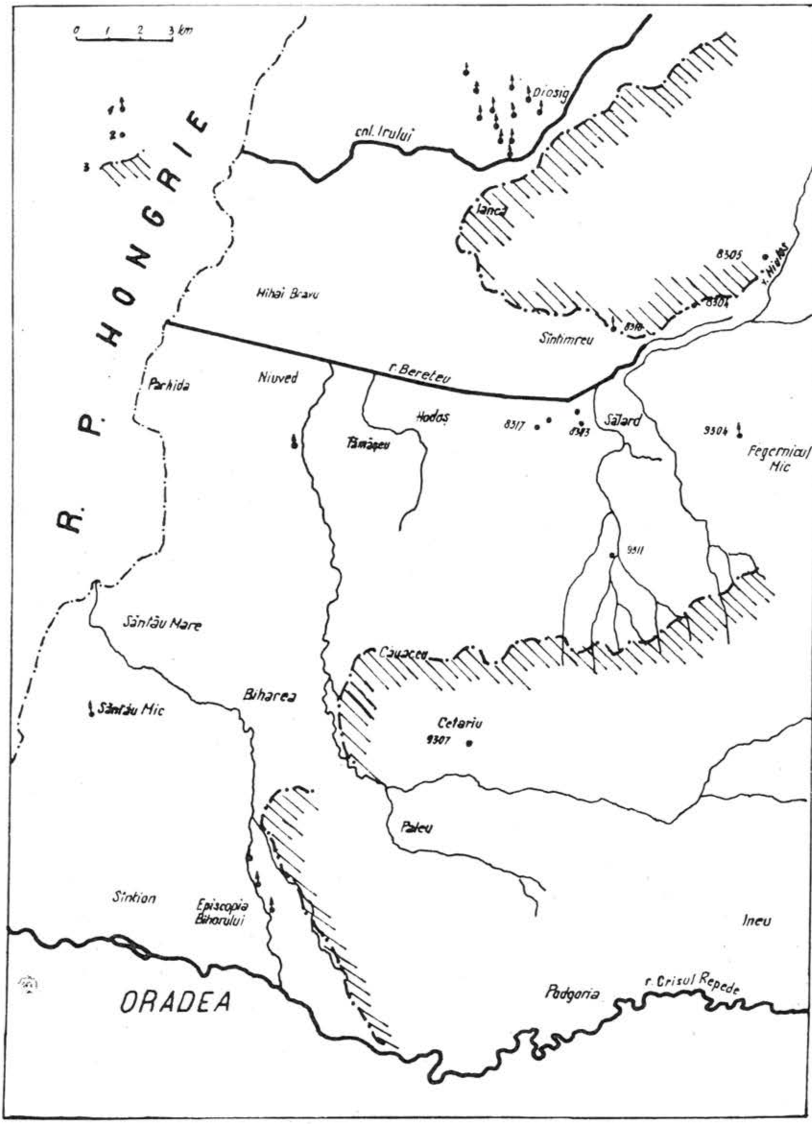

Fig. 1. Schéma de la position des puits artésiens de la plaine du nord d'Oradea

1. Puits artésien en fonction; *2.* puits artésien fermé; *3.* limite de la zone de collines une partie du Pléistocène inférieur, suivant notre opinion. Les dépôts pséphitiques représentent des anciens cônes de déjection pléistocène du Barcău et du Crișul Repede.

Les intercalations perméables de l'horizon sablonneux permettent l'accumulation des importantes couches aquifères de profondeur, qui présentent dans la plupart des cas un caractère artésien, révélées tant par les forages hydrogéologiques exécutés par nous que par des forages antérieurs, quelques-uns de ces forages fonctionnant depuis le siècle passé.

Tableau 1

Répartition des eaux artésiennes accumulées dans les dépôts pliocènes supérieurs de la zone de plaine entre Oradea et Diosig

No.	Localité	Profondeur m
1	Diosig	100 – 150
2	Fegernicu Nou	112
3	Sălard	100 – 155
4	Sălard (4 km au sud)	119
5	Sîntimreu	environ 150
6	Tămășeu	330
7	Santău Mic	> 100
8	Episcopia Bihor	environ 250
9	Oradea	100 – 300

Les données concernant l'extension des eaux artésiennes accumulées dans les dépôts pliocènes supérieurs de la zone de plaine d'entre le Crișul Repede et le Ier (Oradea-Diosig) aussi bien que les eaux de la zone de la basse-plaine du Barcău le long du secteur de Săniob-Tămășeu (Fig. 1) ont été représentées dans le tableau no. 1.

Particularités chimiques

Les résultats de l'étude des caractéristiques chimiques des eaux artésiennes, sont données par le tableau no. 2 et par la figure 2. Cette étude nous permet de faire l'observation : les eaux artésiennes sont des eaux bicarbonatées alcalines, en se distinguant clairement des eaux phréatiques, qui sont des eaux bicarbonatées calciques-magnésiennes. Les exceptions à la règle sont les eaux minérales rencontrées par nous dans un forage exécuté à Sălard, dont la composition chimique ressemble à celle des eaux phréatiques.

Un grand nombre des analyses de différentes eaux des forages de prospections, qui ont fait l'objet d'un ouvrage antérieur [3], démontre qu'à mesure que l'on descend à l'échelle stratigraphique (des dépôts plus récents aux formations plus anciennes), les eaux perdent peu à peu leur caractère bicarbonaté calcique-magnésien, devenant des eaux bicarbonatées sodiques. Cette modification vers des eaux bicarbonatées sodiques s'effectue par l'entremise des eaux à caractère mixte, rencontrées surtout dans les intercalations perméables de la partie supérieure de l'horizon argileux charbonneux ; les eaux de la partie inférieure de cet horizon présentent d'évidentes affinités avec les eaux alcalines de l'horizon sablonneux.

En comparant, dans leur totalité, un peu de caractéristique commune des eaux accumulées dans les dépôts pliocènes supérieurs est révélée, à savoir : plus le caractère alcalin, déterminé par l'enrichissement en ions de Na, s'accentue en même temps que la profondeur, plus la minéralisation totale, diminue au fur et à mesure que l'on intercepte des couches aquifères situées à des profondeurs plus grandes. Le phénomène est plus frappant dans la partie nord du périmètre examiné.

C'est ainsi que les données du tableau no. 3 sont concluantes. Nous avons figuré les ions caractéristiques et la minéralisation totale des eaux rencontrées

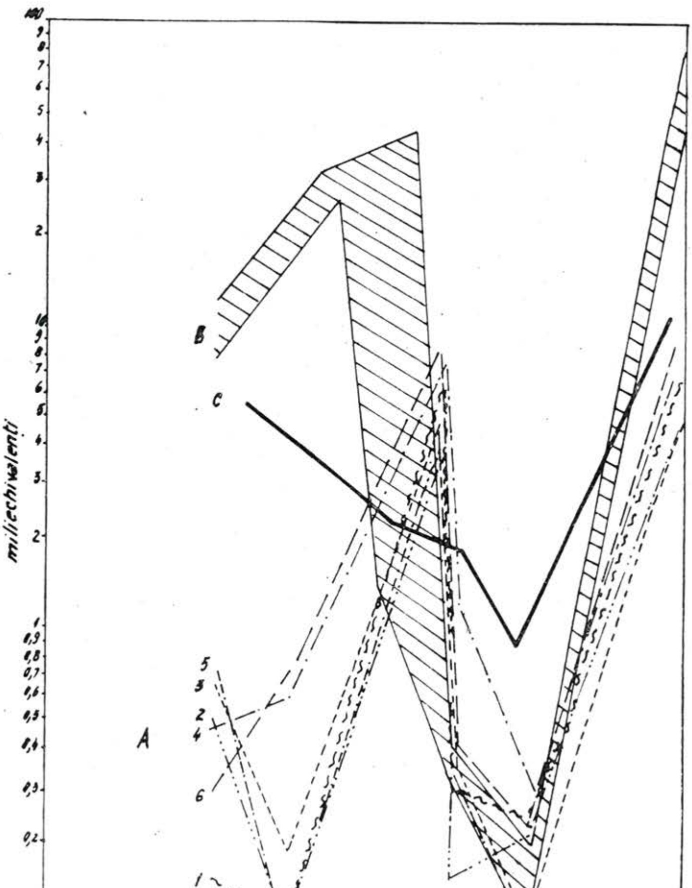

Fig. 2. Représentation graphique du chimisme des eaux artésiennes de la plaine du nord d'Oradea. (Diagramme semi-logarithmique d'après SCHOELLER)

A. Eaux artésiennes: *1*. Diosig; *2*. Fegernic; *3*. Sălard; *4*. Santău Mic; *5*. Episcopia Bihor; *6*. Oradea. *B*. Domaine des eaux artésiennes minérales de Sălard, Sîntimreu et Tămăşeu. *C*. Chimisme moyen des eaux phréatiques

Tableau 2

Composition chimique moyenne des eaux artésiennes de la zone de plaine entre Oradea et Diosig

Localité	Anions						Cations						H_2SiO_3 mg/l	CO_2 libre mg/l	Min. tot. mg/l
	Cl^-		SO_4^{--}		HCO_3^+		$Na^+ + K^+$		Ca^{++}		Mg^{++}				
	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%			
Diosig	10.6	2.0	11.5	1.6	414.8	46.4	164.0	48.6	3.4	1.1	0.5	0.3	23.3	abs	628.1
Fegernicu Nou	8.8	1.9	1.9	0.2	454.5	47.8	131.7	38.8	21.1	6.5	11.7	4.7	7.7	—	670.6
Sălard*	9.9	0.6	3.3	0.1	1338.5	49.2	47.6	4.7	225.1	21.5	191.0	23.7	103.4	616.0	1907.4
Sălard, 4 km au sud	12.0	2.6	17.1	2.9	390.0	44.3	88.2	27.2	45.1	15.1	12.5	7.1	15.6	14.9	548.0
Sîntimreu*	14.2	0.3	9.6	0.1	4684.8	40.6	804.2	22.5	236.0	7.6	373.1	19.8	93.2	1284.8	7504.0
Tămășeu*	14.2	0.3	7.7	0.1	4819.0	49.6	992.3	26.9	158.7	5.0	347.8	18.0	62.1	932.8	7339.4
Santău Mic	45.0	7.8	14.6	1.8	402.7	40.4	162.1	43.2	9.2	2.8	7.3	3.7	20.1	abs	662.9
Episcopia Bihor	5.3	1.6	11.5	2.5	268.4	45.7	93.7	42.1	15.2	7.9	traces	—	12.9	abs	407.8
Oradea	14.2	2.2	9.0	1.0	518.6	46.4	190.1	44.1	6.0	1.6	9.6	4.3	38.8	abs	790.2

* Eaux minérales carbogaseuses.

Tableau 3

Caractéristiques chimiques principales de quelques couches aquifères localisées dans les dépôts pliocènes supérieurs du nord d'Oradea

No. du forage et la localité	Profondeur de la couche aquifère m	Caract. de la couche aquifère	Min. tot. mg/l	Anions						Cations					
				Cl^-		SO_4^{--}		HCO_3^-		Na^+		Ca^{++}		Mg^{++}	
				mg/l	mg e%	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%	mg/l	mg e%
8304 Sîntimreu	11.8–13.6	ascens.	947.5	49.6	5.7	92.2	7.8	549.0	36.5	80.7	14.2	105.8	21.4	41.3	13.8
	66.0–67.5	ascens.	475.8	17.7	4.3	17.3	3.0	305.0	42.7	110.1	40.8	11.6	5.0	5.3	3.7
	95.8–96.2	ascens.	564.8	21.3	4.3	49.9	7.3	329.4	38.4	138.8	42.9	11.6	4.1	4.1	2.4
8305 Siniob	27.6–30.2	ascens.	579.4	17.7	3.5	30.7	4.3	366.0	41.9	91.5	27.8	56.1	19.6	2.4	1.4
	94.2–95.4	ascens.	473.6	10.6	2.6	21.1	3.8	305.0	43.6	106.7	40.4	12.8	5.6	3.4	2.4
9304 Fegernicu Nou	17.6–19.2	ascens.	658.6	7.1	1.3	46.1	6.0	414.9	42.7	64.3	17.6	59.9	18.8	24.7	12.8
	84.1–84.5	artésien	437.6	10.6	2.9	1.9	0.4	292.8	46.7	105.6	44.7	10.4	5.0	abs	—
9307 Cetariu	30.7–33.2	ascens.	719.7	66.0	9.9	35.0	3.9	410.0	36.0	32.2	7.5	83.2	22.3	41.9	18.7
	62.4–65.1	ascens.	581.2	74.0	13.3	46.5	6.1	290.0	30.2	66.9	18.6	57.0	18.2	20.8	11.1
9311 Sălard 4 km au S	20.6–22.6	ascens.	677.4	14.0	2.5	25.1	3.3	424.0	44.2	52.0	15.3	72.8	21.2	20.7	10.9
	39.2–41.3	artésien	572.6	10.0	2.0	21.8	3.3	370.0	44.6	67.2	21.5	49.3	18.2	14.6	8.8
	117.6–118.3	artésien	451.9	12.0	2.8	4.5	0.7	287.0	44.0	101.5	40.0	13.2	6.0	2.2	1.7

à divers niveaux du Pliocène supérieur dans quelques forages exécutés par nous.

On peut donner une explication de ces phénomènes sur la base de l'existence des argiles bentonitiques mises en évidence par nous, à divers niveaux, dans les deux horizons lithologiques du Pliocène supérieur. Nous jugeons que la diminution de la minéralisation totale des eaux de profondeur est due surtout aux propriétés du montmorillonite (dans les argiles bentonitiques en proportion de 55-75 %) de fixer les sels dissous en eau. Cet aspect résulte clairement de l'examination du tableau no. 3, où on peut observer que la diminution de la teneur en ions de HCO_3, SO_4 et Cl est en rapport avec la profondeur.

Le changement du caractère chimique des eaux avec la profondeur est influencé aussi par les argiles montmorillonitiques, dont la capacité d'échange ionique est bien connue. Tel qu'il ressort du même tableau, le phénomène a lieu avec une retenue des ions de Ca et Mg et une élimination des ions de Na, à la suite des quelles les eaux deviennent alcalines.

Certaines petites variations par rapport au cadre général, notons celles du forage 8 304 ou celles des autres forages non figurés sur le tableau, sont dues aux conditions locales et surtout au caractère discontinu des intercalations d'argiles bentonitiques, à cause desquelles la durée et la surface de contact entre les eaux souterraines et les argiles montmorillonitiques diffèrent non seulement d'une couche aquifère à l'autre, mais aussi dans le cadre de la circulation d'une seule couche aquifère.

* * *

Nous avons vu, en effet, que les eaux artésiennes de la zone de plaine comprise entre Crişul Repede et Barcău sont des eaux bicarbonatées alcalines, qui se caractérisent par une minéralisation très réduite, qui ne dépassent pas 0.700 g/l et que cette caractéristique est déterminée par le bicarbonat de sodium, présent en proportion de 70-88 % du total des sels composants de la minéralisation totale.

La minéralisation totale 10 fois plus grande (tableau 2) et la teneur élevée en CO_2 libre (1-2 g/l), totalement absente aux autres eaux artésiennes, sont les caractéristiques hydrochimiques qui mettent en relief les eaux artésiennes de la zone de Sîntimreu-Tămăşeu, en leur imprimant le caractère d'eaux minérales. En outre, on constate que ces eaux, à côté du bicarbonat de sodium, contiennent presque en parties égales du bicarbonat de magnésium et des quantités appréciables de bicarbonat de calcium, sels présents en quantités négligeables aux autres eaux artésiennes. Rappelons également que, tandis que la minéralisation totale des eaux artésiennes diminue à mesure qu'on intercepte des couches aquifères plus profondes, les eaux minérales se caractérisent, tout autrement, par une augmentation de la minéralisation avec la profondeur, due aux quantités de plus en plus grandes de teneur en CO_2, qui détermine l'augmentation considérable de la capacité de dissolution des sels des roches environnantes.

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PHYSICAL AND CHEMICAL CHARACTERISTICS OF CONFINED WATERS IN PERIPHERIC TERRITORIES

PROPRIÉTÉS PHYSIQUES ET CHIMIQUES DES NAPPES AQUIFÈRES CAPTIVES DANS DES ZONES PÉRIPHÉRIQUES

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RÉSUMÉ

La présente étude traite du dépouillement complexe des résultats disponibles dans l'espace des villes Gyöngyös et Hatvan, dont l'enseignement ne se restreint pas à la reconnaissance complexe des caractéristiques locales mais amène aussi la possibilité de la déduction des conséquences généralisables. Les expériences ont révélé uniformément qu'entre autres les couches profondes de 300 mètres d'un des territoires n'ont qu'une alimentation très limitée, contrairement aux couches exploitées dans l'autre part du territoire, où la balance de l'alimentation et de la production demeure encore intacte et le caractère mixte des nappes phréatiques prouve la possibilité d'une alimentation potentielle.

Introduction

Upon ground of geological and hydrogeological characteristics overwhelming part of the literature appoints the south peripherial part of the North Middle Mountain (Északi Középhegység) (Cserhát—Mátra—Bükk) as the recharge area of the deep aquifers of the Hungarian Great Plain.

It is the centrally positioned south part of the Mátra Mountain which plays in this respect an important role. In consequence, the further analyse of the hydrogeological characteristics of this tract as well as the manifold evaluation of the data being at disposal can also contribute to the clearing up the recharge circumstances of the water-bearing strata. In the south foreland of the Mátra Mountain the areas of the towns Hatvan and Gyöngyös offer two suitable tracts for a comparative analysis, relating to which on basis of the data being at disposal with FTI appropriate inferences can be drawn.

General hydrogeological characteristics

Important differences can be found between the geological structure of these two areas.

At Gyöngyös the thickness of the Pleistocene strata does not exceed 20 m and is directly underlying by upper Pannonian strata. In the Hatvan space according to explorations — the Pannonian sediment series is situated at different depths. Northerly of the town—in the environs of Lőrinci—it follows under the 10—15 m thick Quaternary river deposited strata complex, whereas southward to the city the Quaternary strata are thickening and the upper Pannonian strata series begin only at the depth of 150 m. In consequence, also the hydrogeological characteristics are different.

In the Gyöngyös area only the sand strata disclosed in the upper Pannonian strata series are suitable for water supply. Earlier formations—lower Pannonian and Tortonian strata—are unsuited for water supply purposes. The lower Pannonian series consists of clay and clayey marl, the Tortonian series of sandstone, clayey marl, andesite tuff and agglomerate formations. North of the town towards Mátrafüred the upper Pannonian strata series is still of little thickness (60—100 m) and of predominantly clayey development. It is expanding southwards and the sand strata become more and more frequent in the sediment series. Around Gyöngyöshalász the depth of the sediment series is appr. 350 m, southwards at Atkár it exceeds even 450 m. The most favourably developed spots as to granulometric composition and thickness can be displayed between Gyöngyös and Gyöngyöshalász (Fig. 1). Southward the thickness of sand layers are diminishing and the more subtle fraction is prevailing. The water-bearing strata branch off in southward direction and disintegrate into more layers. The sand layers incline gently towards the Hungarian Great Plain and are covered with Quaternary sediments of various thicknesses which are storing ground water.

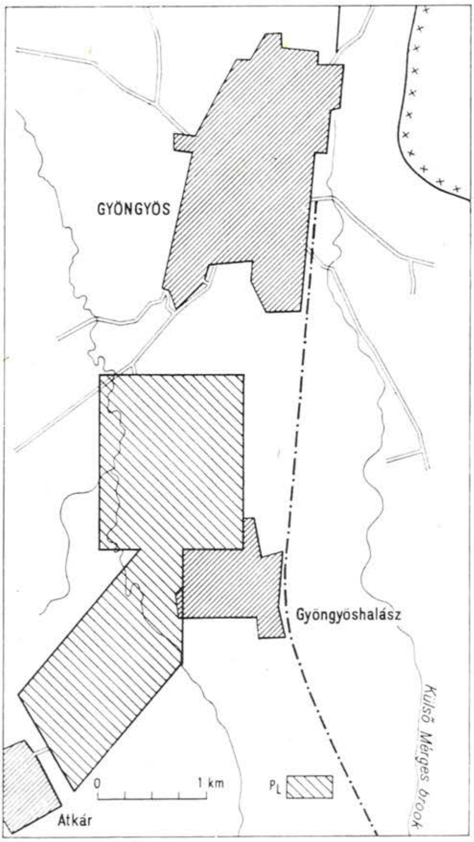

Fig. 1. Site plan of the Gyöngyös water supply area

In the Hatvan area the sand, gravelly sand and gravel layers of the upper Pannonian and Quaternary strata series can be taken into consideration for water supply purposes (Fig. 2). North of the town, in the Zagyva valley partly the ground water of the river deposited gravel layer situated near the ground surface, partly the ground water of the upper Pannonian sand layers settled at different depth (until 350 m deepness) are suitable for water supply. The town and its industrial plants utilize the waters of the aquifers situated near the ground surface for industrial purposes and use the waters of Pleistocene aquifers as drinking water. The appr. 300 m thick superior section of the upper Pannonian strata complex contains 4—6 water-bearing sand layers. The water-bearing strata situated near the ground surface

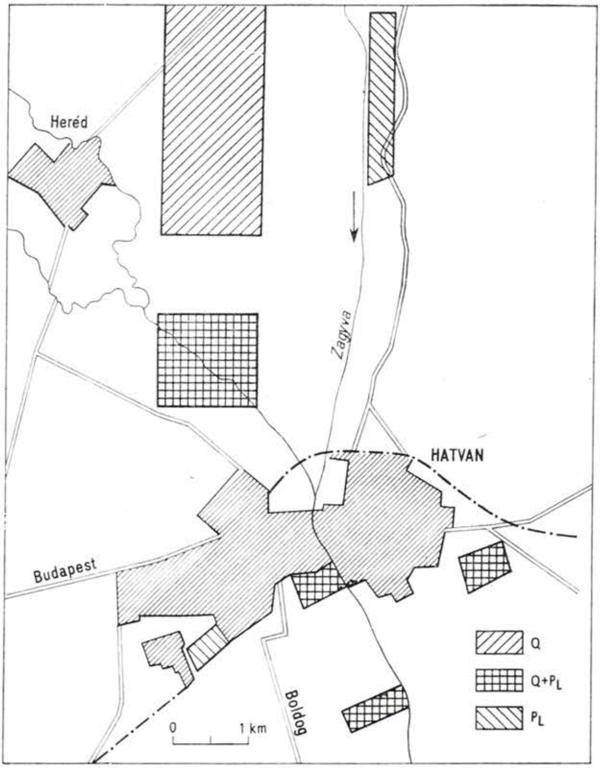

Fig. 2. Site plan of the Hatvan water supply area

north of the town are sinking gradually southwards in the depth and Quaternary strata of increasing thickness are accumulated on them. Near Hatvan the depth of the Quaternary strata series exceeds 100 m. This strata series contains appr. 2–3 gravelly and sandy water-bearing strata.

Near the ground surface in the Zagyva valley and in the valley of the brook Heréd good water-bearing sandy gravel strata are available at the depth of 3–30 m. According to explorations the water which can be recovered is generally of bad quality and unutilizable as drinking water.

Sand strata suitable for water supply constitute the middle part of the upper Pannonian series. In this formation period lignite, clayey sand and clay strata were deposited beside the sand strata. Similarly to the surroundings of Gyöngyös the sand strata explored and utilized for water supply were formed in the middle upper-Pannonian formation period too, in consequence the upper Pannonian water-bearing layers of both explored areas were formed in an approximately identical period.

Hydrochemical characteristics

Beyond the hydrological and hydrogeological explorations done continuously during several decades both areas treated above have been subject to precise and reliable hydrochemical examinations, too. By help of these the water types and all those changes which are ensueing in horizontal and vertical direction in the chemical characteristics of the waters can be demonstrated.

Nearby Gyöngyös the dissolved solid content of the ground water in the Quaternary strata is increasing from the mountain border to the south. The ground water of the aquifers situated near to the ground surface can be ranged among the Ca-Mg and bicarbonate containing waters. The water-pollution effect of the town prevails to a high degree.

The quality of the waters of upper Pannonian strata is variable both horizontally and vertically. Depending on the depth the dissolved solid content varies not essentially but with the cations a definite regularity can be demonstrated (Fig. 3). In the waters of the upper strata the prevailing Ca-Mg content is diminishing towards the depth, whereas the Na content is increasing. In compliance with this the waters of the upper strata can be ranged still among the Ca-Mg bicarbonate-containing waters. The waters contained by the strata between 130—240 m are already mixed, thus containing Ca-Mg-Na bicarbonate, while with those under 250 m, sodium will be the prevailing cation and the waters become containing only sodium bicarbonate. Together with the variation of the Ca-Mg quantity the hardness of the water is also

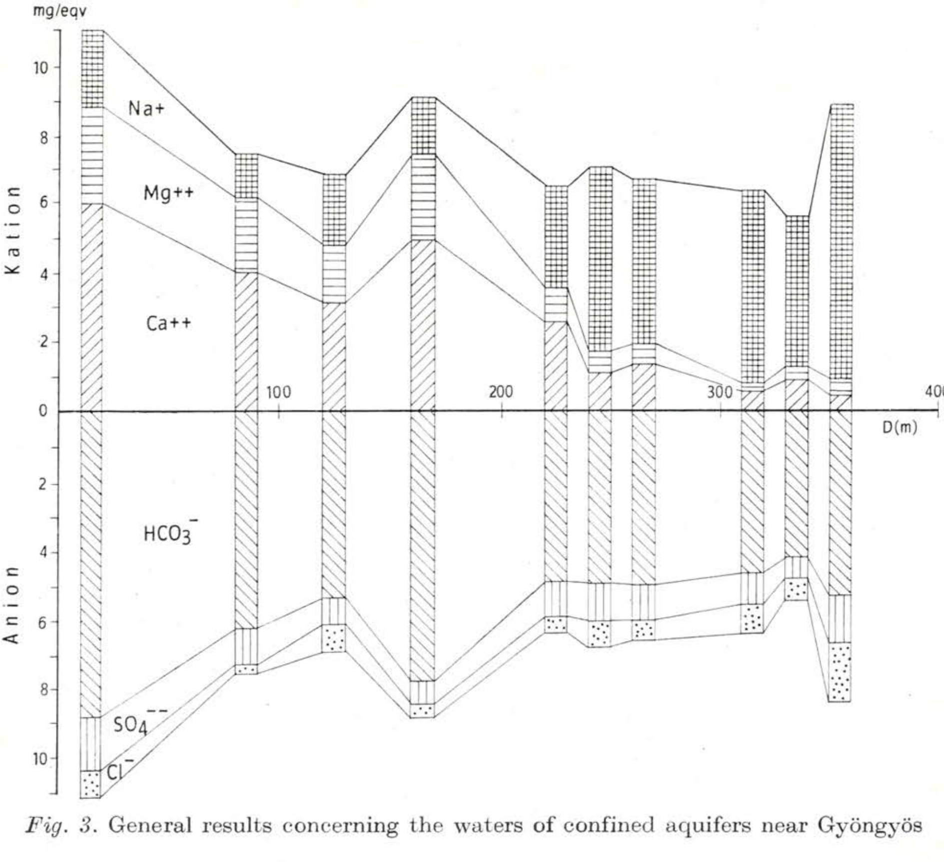

Fig. 3. General results concerning the waters of confined aquifers near Gyöngyös

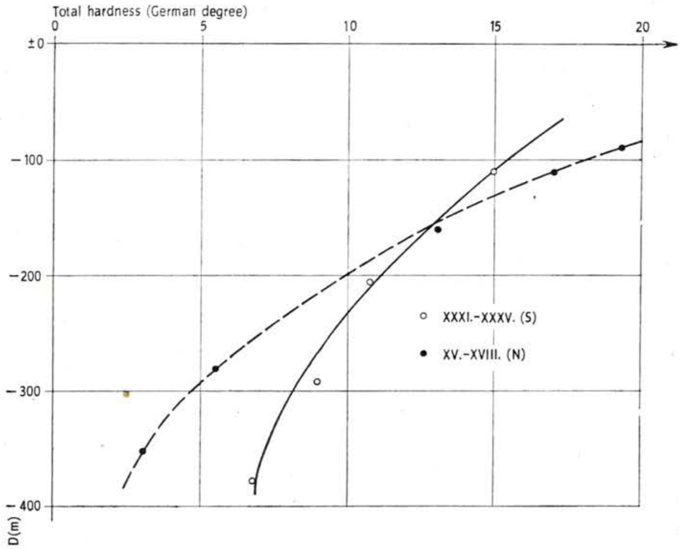

Fig. 4. Characteristic values of total hardness of the waters contained in confined aquifers at Gyöngyös

decreasing with the depth. The hardness of the upper strata shows a value of 15−17 German hardness degree, that of the lowest strata a value of 2−3 German hardness degree.

In horizontal direction we can observe a tendency that the hydrochemical variations spoken of previously do not prevail so vigorously. With the most southward well group being at the farthest from Gyöngyös, nearby Atkár, the Ca-Mg content and the hardness are not decreasing in the same measure −depending on the depth−as on the upper and middle section of the water-bearing area (Fig. 4). At the same time we can observe a characteristic picture on the variation of the total hardness upon ground of data being at disposal on the northern and southern well groups of the area.

In the surroundings of Hatvan the quality and the chemical composition of the ground water in aquifers situated near the ground surface is varying according to the area units. This can be explained by local effects. Especially in the Zagyva valley and southwards from Hatvan these waters have a very high dissolved solid content consequently they cannot be utilized as drinking water. But in the Nógrád brook valley and in the higher situated western terraces of the Zagyva they are of drinking water quality, although they must be treated in consequence of their high iron and manganese content. These waters can be ranged generally among the waters containing Ca-Mg bicarbonate. In the surroundings of the strongly polluting centers this composition varies and enrichment in SO_4 can be observed.

The chemical composition of the waters in Pleistocene strata are favourable. The hardness is varying between 11−15 German degree and is exceeding this value only rarely (see Fig. 5). The dissolved solid content increases with the depth only in a smaller extent. The waters can be ranged among those containing Ca-Mg bicarbonate, though it is worth while to note that the quantity of sodium is relatively high and its value increases with the depth. Sulfate can be found only in traces and the quantity of chloride is also a very subordinate one. In consequence, bicarbonate prevails among the anions. The Mg equivalent of cations and anions does not go beyond 10.

In the chemical composition of the upper Pannonian strata such kind of variations can be observed in the explored appr. 350 m depth out of which we are able to conclude to the origin and derivation of the waters (Fig. 6). The dissolved solid content is in all the cases higher than in the Pleistocene confined aquifers and increases with the depth. Also the quantity of calcium

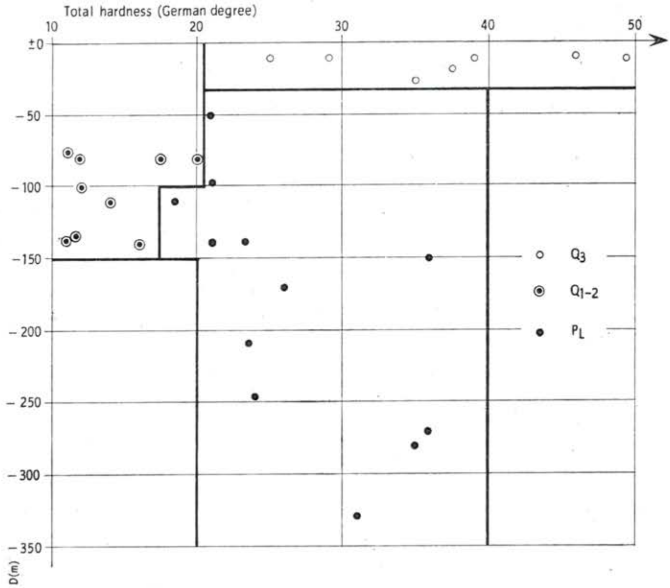

Fig. 5. Relationship between the total hardness of subsurface waters and the depth of aquifers in the surroundings of Hatvan

and magnesium is increasing but sodium manifests itself beyond this measure and therefore becomes a prevailing cation already under 200 m. Similarly to its sodium content, its chloride-ion content also increases with the depth (Fig. 7), and the water of the upper strata containing Ca-Mg bicarbonate is passing into a type of water containing sodium, Ca-Mg bicarbonate and chloride. The water in the upper strata contains Ca-Mg bicarbonate, referring to natural recharge. The water stored in the lower strata contains in addition to its mixed composition Na-Cl, i.e. common salt. In consequence of such a chemical composition of the waters it seems probable that the waters contained by this strata conserved to a certain extent their original characteristics so they must be in some extent contemporaneous, notwithstanding their mixed character is a proof of being mixed up with other water types, i.e. waters of aquifers situated near the ground surface. It means at the same time that the water recharge of the lower strata is limited and motivates the fact that the complete water exchange did not take place until our days. Essential differences can be demonstrated between the chemical composition of the

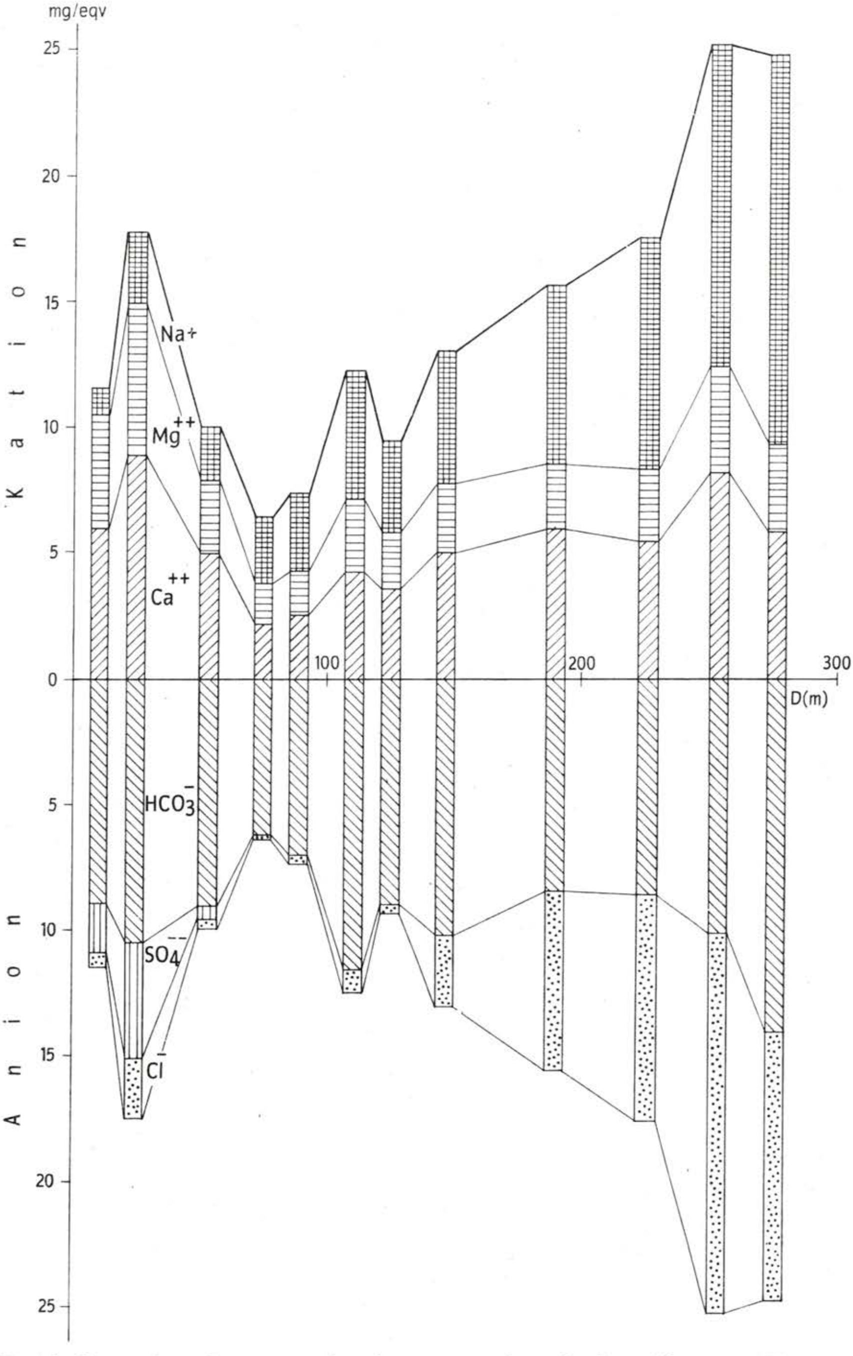

Fig. 6. General results concerning the waters of confined aquifers near Hatvan

upper Pannonian confined aquifers in the surroundings of Hatvan and Gyöngyös, respectively.

a) The dissolved solid content in the confined aquifers is increasing with the depth at Hatvan, which circumstance cannot be observed at Gyöngyös. Furthermore, the quantity of the dissolved solids in the waters is much more higher at Hatvan than at Gyöngyös.

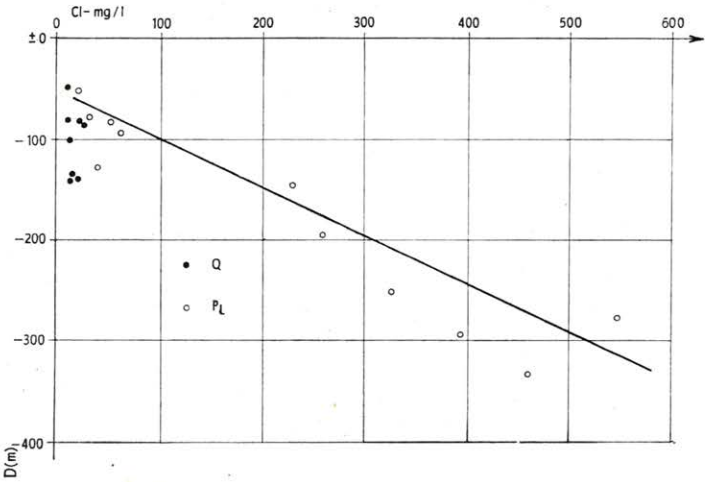

Fig. 7. Chloride-ion content in the waters of Hatvan

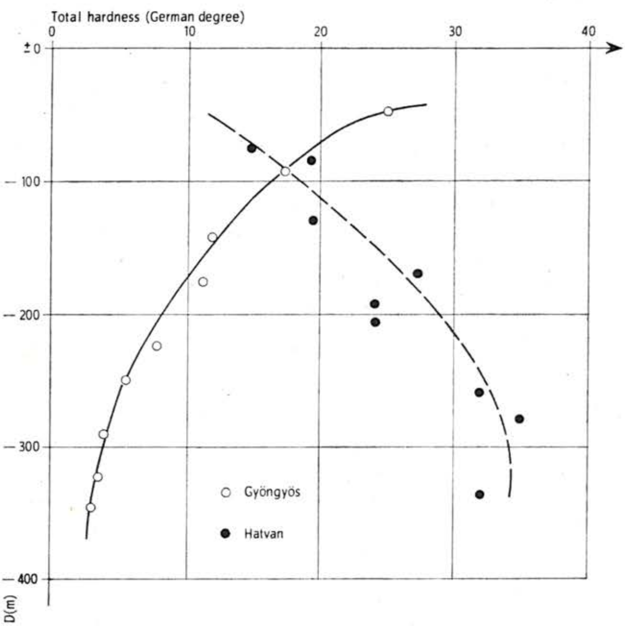

Fig. 8. The total hardness of confined aquifers in the Gyöngyös and Hatvan areas

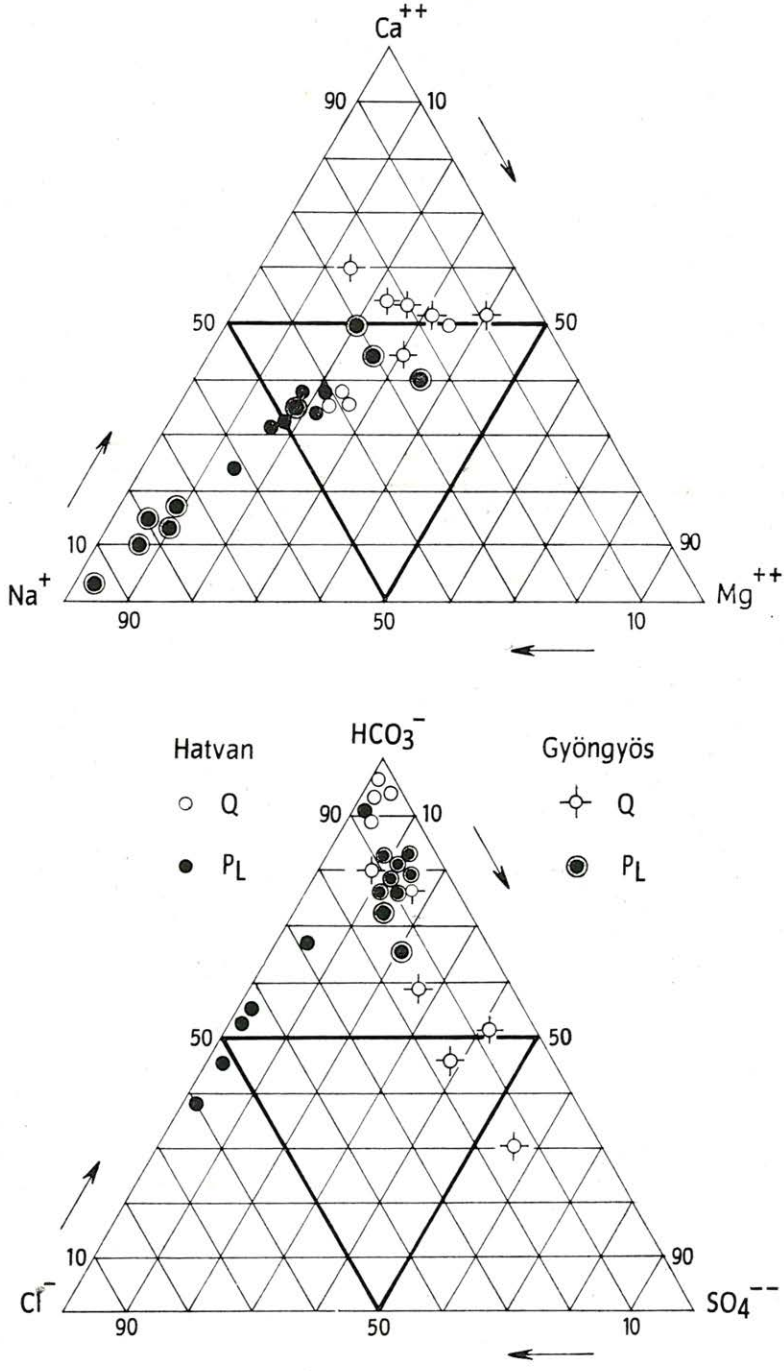

Fig. 9. The type of waters in confined aquifers near Gyöngyös and Hatvan

b) At Hatvan the Ca-Mg quantity as well as the total hardness of the water is increasing with the depth, whereas at Gyöngyös the value of both is decreasing gradually (Fig. 8).

c) In both areas an increasing sodium content can be observed, which at Gyöngyös is becoming meanwhile prevailing.

d) The chloride-ions are increasing in several waters of the Hatvan area with the depth into such measure that these waters must be ranged among the chloride-containing type of waters. At Gyöngyös no important chloride-ion increase can be observed with the increasing depth, bicarbonate being the prevailing anion independently of the depth.

e) Both at Gyöngyös and at Hatvan the upper confined waters—aquifers occurring at a smaller depth—contain equally mostly Ca-Mg bicarbonate. The waters in deeper layers begin at Gyöngyös to become a sodium bicarbonate type. At Hatvan too sodium prevails but Ca and Mg, further Cl content will remain important, therefore here mixed waters of sodium-Ca-Mg-bicarbonate type were formed (Fig. 9).

Physical characteristics

We are able to conclude on the recharge characteristics of confined aquifers, on the conditions developed between those aquifers and their variations also from other circumstances than hydrochemical characteristics.

Taking into consideration the position of the piezometric surface of the different strata in the Gyöngyös area it is a characteristic fact that the increase of piezometric levels with depth deviate from linear (Fig. 10). In the water supply area to be found in appr. north—south direction the piezometric slope of each aquifer can be easily explored as well as the spatial variation of the piezometric level, which follows approximately the slope of the water bearing strata. The lately accelerated general increase of water production had, of course, an important effect on the development of the piezometric levels. Any further breaking point in the decrease of piezometric surfaces is connected with the simultaneous increase of water production (Fig. 11). In the Hatvan area the connection between piezometric surface and depth of a layer is positive and linear. It is of interest, that the relationship is developing differently with Pleistocene layers (Fig. 12).

The water temperature has essentially the same tendency with depth as the piezometric surface and also in the Hatvan area the relationship is similar (Fig. 13).

Both circumstances refer to different characteristics concerning water recharge and water flow. In the case of the deeper aquifers at Gyöngyös the trend of the piezometric surfaces shows an important decrease following the production as well as the distribution of the temperature. Such unfavorable characteristics of the water-recharge are not proved still in the Hatvan area.

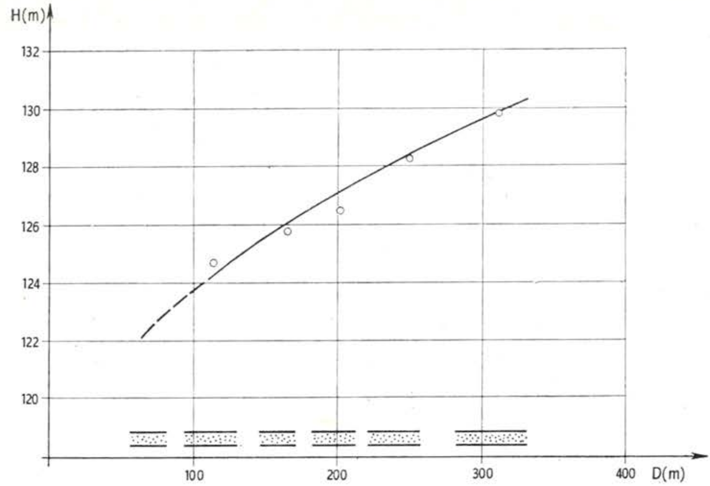

Fig. 10. The piezometric surfaces of confined aquifers at Gyöngyös

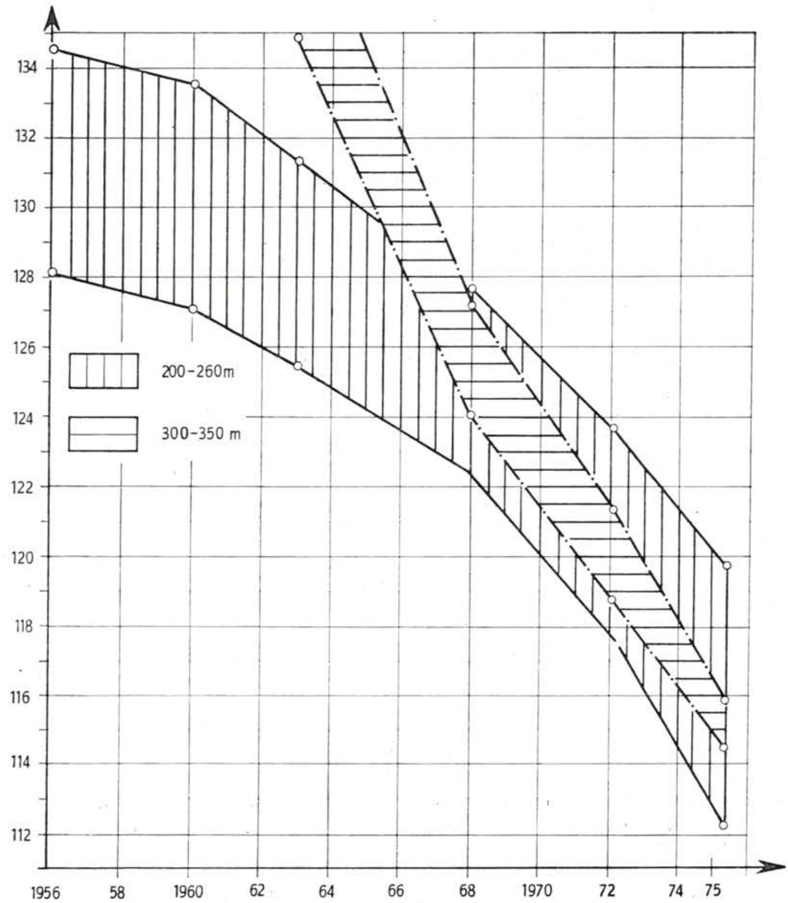

Fig. 11. Changes with time in the piezometric surfaces of confined aquifers at Gyöngyös

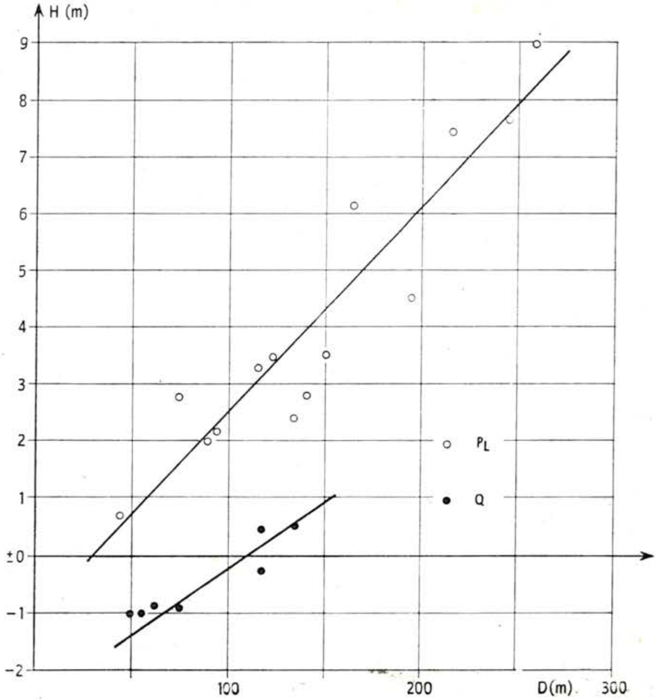

Fig. 12. The piezometric surfaces of confined aquifers near Hatvan

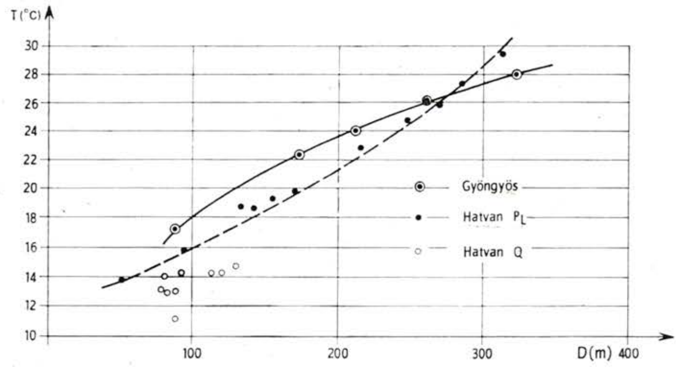

Fig. 13. The temperature of confined aquifers

Conclusions

When examining simultaneously the geological, hydrological and hydrochemical characteristics we become able to draw local and general inferences.

a) In the surroundings of Hatvan and Gyöngyös explorations proved unequivocally that the water-bearing gravel strata situated near the ground surface were settled on several spots directly—without intercalation of an impermeable clay layer—on the upper Pannonian water-bearing sand layers and a hydrological connection developed between them. In consequence, all preliminary geological and hydrological conditions of the water recharge of the deeper confined aquifers can be demonstrated on the southern side of the Mátra Mountain.

b) On ground of hydrochemical examinations it can be stated in the surroundings of Gyöngyös that the upper water yielding layers of the upper Pannonian sediment series are recharged out of the ground water of the aquifers situated near the ground surface. This is proved by the higher amount of Ca and Mg ions as well as by the hardness of the waters. The deeper water-bearing layers of this series are recharged from waters transported by the cracks and faults of the Mátra Mountain's andesite.

c) These statements are supported by the C 14 age determination of the waters contained in the confined aquifers in the surroundings (Nagyfüged and Boconád). The age of the water contained by these wells is supposed to be $19{,}580 \pm 560$, resp. $31{,}890 \pm 925$ years. By these age determinations and by other explorations it is proved that waters left over from the Pliocene ages are out of question because water exchange was already ensured. Just in consequence of the characteristics of water recharge and water production in the Gyöngyös area the contemporaneous waters of the upper Pannonian strata were already exchanged unto a depth of 350—400 m into waters with a reduced salt concentration of Ca-Mg-bicarbonate type resp. sodium bicarbonate type. Results concerning piezometric surfaces and water temperature distribution indicate the same circumstances.

d) Beyond the comprehensive characterization of local properties it can be stated as a general principle that a certain area can be evaluated reliably only by taking account simultaneously of all circumstances. During a characterization of this kind it is possible to draw reliable inferences and to eliminate over-hasty decisions by the reinforcing effect of elaborations carried on in different directions but completing each other at the same time.

PROBLÈMES HYDROCHIMIQUES DANS L'EXPLOITATION DU SOUFRE DE LA RÉGION DE TARNOBRZEG – EN POLOGNE

HYDROCHEMICAL PROBLEMS IN THE POLISH SULPHUR MINES OF THE TARNOBRZEG REGION

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ABSTRACT

Hydrogeological investigations carried out regularly on sulphur deposits since 1953 have given lot of information about chemical water composition.

Further, dewatering works in open pit mines made possible the realization of more detailed research to recognize water chemism in this region.

Sulphur deposits occur in the Miocene formation of the Forecarpathian Depression.

In the deposit area two water-bearing horizons have been identified:

I Quaternary horizon.

II Tertiary horizon.

The first, unconfined, water-bearing horizon with a free surface of water occurs in sands and Pleistocene gravel of 5 to 11 m thickness. Chemically, the Quaternary waters are not mineralized. The second water-bearing horizon (Tertiary) is composed of sulphur-bearing limestones and Baranów sand layers of 50 to 120 m thickness.

It is a confined horizon isolated from the Quaternary horizon by Krakowiec clays with a thickness of 40 to 110 m.

The Tertiary waters are mineral waters of high hydrosulphide content, enormously aggressive.

So, in the process of deposit exploitation problems have been faced in dewatering open pit mines and especially in the choice of filters, pumps and other dewatering instruments.

Data obtained from drainage wells in open pit mines and from exploitation wells in bore-hole mines are the main information source about water chemism.

Generally, 1500 chemical analyses have been used for the present report.

On this basis, maps and hydrochemical cross-sections have been prepared. The occurrence of a chemical composition zoning of the water has been stated. This differentiation consists in the increase of general mineralization and of Na^+ and Cl^- ion content together with the deposit occurrence.

The value of Ba^{2+} ion varies irrespective of depth. It is possible to show regions with an increased Ba^{2+} content, what is probably connected with the observed baryte content in sulphur-bearing limestones.

While dewatering the deposit an intensive precipitation of baryte on the pumps causing their quick wear, was observed.

Recognition of water chemism is also of great importance in the bore-hole sulphur mines.

On the basis of its differentiation it is possible to trace directions of water migration within the deposit and changes in chemism during sulphur exploitation and makes posible to observe migration of the heating water injected into the deposit in order to melt sulphur in it.

Introduction

Parmi les problèmes liés à l'exploitation des gisements de soufre dans la région de Tarnobrzeg, une place importante occupe le problème des eaux souterraines.

Les premières études hydrogéologiques des années 1953-55 ont déjà démontré, que les eaux liées aux zones soufrées ont une minéralisation élevée et une haute teneur en hydrogène sulfuré. A partir de cette constatation, il fût indispensable de bien reconnaître le chimisme des futures eaux d'exhaure, comme facteur indispensable aussi pour tous projets et réalisations visants l'évacuation de l'eau de la mine à ciel ouvert, ainsi que le système de son épuration — avant le transfert définitif dans les cours d'eaux superficielles. Les travaux de l'exhaure, débutés dans la région en 1967 par la mise en marche du système, qui fonctionne jusqu'à maintenant (barrières de puits d'épuisement des eaux) — ont permis de poursuivre l'étude détaillée du chimisme des eaux, de ses modifications et des difficultés de l'exhaure, causées par les propriétés spécifiques des eaux sulfurées.

L'auteur, tâche de présenter ici, en raccourci, les effets de cette étude.

L'étape suivante, des études de la variation du chimisme des eaux de l'exhaure dans la région, est réalisée durant l'exploitation par fusion directe du soufre dans le gisement — à l'aide de forages (FRASCH). Le grand nombre de forages de prospection, d'observation et d'exploitation directe a rendu possible la réalisation d'un échantillonnage important des eaux.

Il est à noter, que l'exploitation directe par forages présente des problèmes tout autres — aussi bien dans le domaine de la modification des conditions hydrauliques, que dans le caractère chimique des eaux de la zone exploitée, et, ce dont l'auteur fait part aussi, dans le rapport présent.

Grands traits de la structure géologique locale

La genèse des gisements de soufre de la région est étroitement liée à l'altération des dépôts gypseux, faisant partie des formations tertiaires, déposées directement sur le substratum cambrien.

La formation tertiaire est représentée ici par le Miocène, développé comme :

— argiles à lignite, considérées comme étant d'âge helvétique et constatées uniquement localement
— sables et grès de Baranów du Tortonien
— calcaires à soufre du Tortonien
— argiles cracoviennes du Tortonien et du Sarmatien.

A son tour, le Tertiaire est recouvert par le Quaternaire, développé sous forme de sables, glaises et cailloutis.

Conditions hydrogéologiques

La région des gisements de soufre se caractérise par deux nappes aquifères.

La première nappe, de caractère phréatique, englobe les sables, graviers et cailloutis du Quaternaire, dont la puissance varie de 5 à 15 m. Le niveau de la nappe se trouve, dans les conditions naturelles de la terrasse inférieure — en moyenne à la profondeur de 2 m environ du terrain. Sur la terrasse supérieure, à puissance de 1 à 5 m et composée en majeure partie de glaises, le niveau variable de la nappe est situé à 1 m environ, en moyenne, du terrain. La puissance du Quaternaire dépend de la configuration du substratum argileux du Tertiaire et de l'érosion superficielle.

La seconde nappe aquifère — captive, du Tertiaire, se trouve dans les formations de calcaires à soufre et les sables de Baranów sous-jacentes. La nappe tertiaire est isolée de celle quaternaire par la formation des argiles cracoviennes imperméables. Les calcaires, mentionnés plus haut, se caractéri-

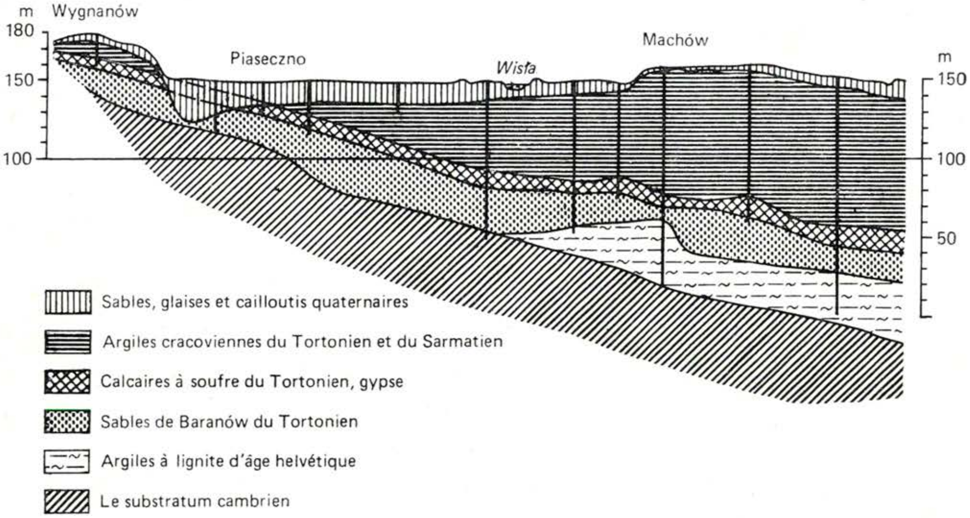

Fig. 1. Coupe géologique schématique de la région

sent par une fissuration abondante et des nombreuses crevasses, qui constituent un parcours aisé pour l'eau. La série de Baranów, sous-jacente les calcaires est constituée d'intercalations de sables fins et de grès. La puissance totale du complexe perméable de la seconde nappe varie de 50 à 120 m. *La nappe* captive *du Tertiaire* est sous pression hydrostatique, qui dépend de la différence de niveau de son toit imperméable envers le niveau d'alimentation. Dans la région considérée, cette différence varie de 60 à 120 m à partir du toit de la nappe. La valeur du coefficient de filtration, pour le complexe entier, est en moyenne 2.5 m/24 H ($\sim$0.003 cm/sec).

L'alimentation a lieu aux affleurements des calcaires et des sables — dans la zone de leurs contacts directs avec le recouvrement de grès et sables du Quaternaire.

Caractère chimique de la nappe quaternaire

Quoique le rapport concerne surtout la nappe tertiaire, il serait utile d'avoir, soit une notion générale de la nappe quaternaire dont le caractère chimique présente le tableau correspondant (Tableau 1). La nappe quaternaire constitue la base unique du ravitaillement en eau potable de la ville et pour

Tableau 1

Comparaison des compositions chimiques du Quaternaire

N° d'analyse	Minéralization mg/l	pH	Dureté totale	Cations (mg/l)					Anions (mg/l)			
				Na^+	K^+	Ca^{2+}	Mg^{2+}	Mn^{2+}	Cl^-	SO_4^{2-}	HCO_3^-	NO_3^-
1	454	6.8	13.9	31	34	83	10	.	48	178	98	13
2	430	7.9	11.7	30	20	66	10	0.3	31	96	18	150
3	376	7.0	9.7	21	52	42	16	0.3	45	74	61	73
4	318	7.1	8.1	18	6	40	11	0.7	37	81	79	65
5	429	7.1	11.7	41	32	64	12	.	57	138	73	20
6	700	7.0	20.1	48	72	50	57	.	170	201	92	0
7	482	7.0	19.4	30	7	81	35	0.3	147	114	79	17
8	289	7.0	8.8	12	15	52	6	0.4	28	154	24	2
9	162	6.9	6.7	4	7	32	10	.	15	62	37	4

les besoins de ménages et fermes de la région et, à ce titre, fait objet d'études et de contrôle.

Il existe quelques captages d'eau quaternaire, mais aussi une grande quantité de puits individuels de fermes et cultivateurs — dont les analyses chimiques donnent l'image de la qualité de l'eau de cette nappe. Dans le Tableau 2 sont présentées les teneurs minimum et maximum des composants respectifs. Du point de vue de la composition ionique, les eaux quaternaires représentent des eaux calco-bicarbonatées, avec une teneur en sulfates rehaussée, ou bien des eaux calco-bicarbonato-sulfatées. Elles contiennent beaucoup de fer et doivent être en principe améliorées pour la consommation.

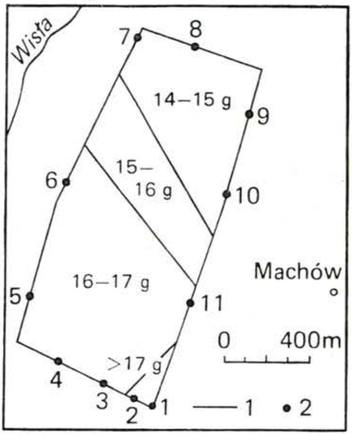

Fig. 2. Carte de la minéralisation de la nappe tertiaire

Caractère chimique de la nappe tertiaire

L'étude du caractère chimique de l'eau de la nappe est réalisée au moyen de:

1. analyses chimiques complètes avec détermination des propriétés physiques, pH, dureté, oxydabilité, basicité, teneur en hydrogène sulfuré et sulfures, minéralisation totale, ainsi que les anions et cations suivants: Na^+, K^+, Ca^{2+}, Mg^{2+}, Fe (tot.), Mn^{2+}, NH_4^{2+}, Ba^{2+}, Sr^{2+}, Cl^-, SO_4^{2-}, HCO_3^-, NO_2^-, NO_3^-, PO_4^{2-}, SiO_3^{2-};

2. analyses iodométriques de la teneur en H_2S.

Pour l'échantillonnage, visant les analyses chimiques complètes, on a désigné un certain nombre de puits d'exhaure, de sondages d'observation et de forages d'exploitation. Grâce au grand nombre de données analytiques, on peut caractériser les propriétés physiques et chimiques des eaux de la nappe de manière suivante :

1. *Propriétés physiques*: les eaux de la nappe sont de couleur verdâtre ou

Comparaison des paramètres hydrochimiques ainsi que des teneurs minimum et

N° d'analyse	pH	H_2S	Minéralisation	Teneur en mg/l.			
				Na^+	Ca^+	Mg^+	Sr^{2+}
1	6.8–7.0	382–476	17.535–18.478	6000–6450 261–280 87–88	369–414 18–21 6–7	184–222 15–18 5–6	11.1–22.7 0.2–0.5 0.1–0.2
2	6.6–7.3	335–505	17.974–18.918	6100–6600 265–283 86–88	407–473 20–24 6–8	175–251 14–21 4–6	5.7–15.9 0.1–0.4 0.0–0.1
3	6.4–7.4	316–467	14.459–17.638	4900–6000 213–261 85–88	396–442 20–22 5–8	116–209 10–17 3–6	9.0–16.8 0.2–0.3 0.1–0.2
4	6.5–7.4	316–494	16.360–17.508	5400–6000 235–261 85–88	403–494 20–25 7–8	119–184 10–15 3–5	6.9–21.3 0.1–0.4 0.1–0.2
5	6.4–7.9	316–416	16.490–17.690	5400–6000 239–261 82–87	397–477 20–24 7–8	161–198 13–16 4–6	7.1–24.8 0.1–0.5 0.0–0.2
6	6.4–7.3	224–456	15.930–17.230	5250–5950 228–259 84–87	305–609 15–30 7–11	126–278 10–23 3–8	6.7–17.0 0.1–0.4 0.1–0.1
7	6.4–7.2	282–341	13.959–14.645	4300–4500 185–196 79–81	593–692 30–35 12–14	131–210 11–17 5–7	7.3–35.5 0.1–0.8 0.1–0.9
8	6.5–7.2	263–350	14.045–14.708	4250–4400 185–191 78–80	669–726 33–36 14–15	122–170 10–14 4–6	8.7–29.1 0.2–0.6 0.1–0.2
9	6.8–7.1	247–380	14.166–14.861	4300–4600 187–200 79–80	674–772 34–36 14–15	133–150 11–12 4–5	10.5–13.9 0.2–0.3 0.1–0.1
10	6.7–6.8	259–297	14.296–14.480	4300–4425 187–192 77–79	707–756 35–37 14–15	148–158 12–13 5–6	10.3–14.7 0.2–0.3 0.1–0.1
11	6.3–7.2	352–424	15.292–17.136	5000–5850 217–254 83–86	462–565 23–28 8–10	150–192 12–16 4–6	6.9–16.9 0.1–0.3 0.1–0.1

noire, à odeur distincte d'hydrogène sulfuré et à température variante de 10° à 15 °C.

2. Quant à leur *composition chimique,* les eaux des respectives régions de gisements de soufre se distinguent entre eux par minéralisation et caractère chimique propres. Mais, comme le rapport présent concerne deux régions: de la mine à ciel ouvert et de l'exploitation directe par forages à distance de ca. 15 km – les eaux correspondantes seront caractérisées séparément.

La région de l'exploitation à ciel ouvert

La minéralisation de la nappe s'élève de 14 à 19 g/l en augmentant avec la profondeur du gisement. La disposition des zones à différentes minéralisations suggère, aussi, une relation entre les phénomènes hydrochimiques et les structures tectoniques.

Tableau 2

maximum des ions de la nappe tertiaire dans la région de la mine à ciel ouvert

m-val/l, % m-val					Paramètres hydrochimiques		
Ba^{2+}	NH_4^+	Cl^-	SO_4^{2-}	HCO_3^-	$\frac{rSO_4^{2-}}{rCl^-}$	$\frac{rCl^- - rNa^+}{rSO_4^{2-}}$	$\frac{rNa^+}{rCl^-}$
5.6–11.2 0.1–0.2 0.0–0.1	32.2–46.8 1.7–2.6 0.6–0.8	9927–10.460 280–295 91–92	167–270 3–5 1–2	1195–1232 19–20 6–7	0.015	3.87	0.94
1.9–16.8 0.0–0.2 0.0–0.1	18.0–69.4 1.0–3.8 0.3–1.2	10.211–10.796 288–304 91–92	152–259 2–5 1–2	1195–1259 19–20 6–7	0.013	4.13	0.93
1.7–10.7 0.0–0.1 0.0–0.1	30.6–63.1 1.0–3.5 0.3–1.1	7871–9998 222–282 88–91	164–411 3–8 1–3	1171–1281 19–21 6–7	0.025	2.57	0.93
5.1–14.0 0.1–0.2 0.0–0.1	27.0–61.3 1.5–3.4 0.5–1.2	9209–9857 260–278 89–91	302–424 6–8 2–3	1195–1299 19–21 6–7	0.029	2.04	0.93
3.5–31.0 0.1–0.4 0.0–0.2	2.8–90.2 0.1–4.5 0.4–1.5	8793–10.070 248–284 88–91	125–620 2–12 1–4	1220–1372 20–22 6–7	0.022	1.99	0.95
3.3–17.3 0.0–0.2 0.0–0.1	2.1–64.9 0.1–3.6 0.0–1.2	8545–9644 241–282 87–91	231–744 4–15 1–5	1195–1269 19–20 6–7	0.039	1.23	0.95
2.3–10.2 0.0–0.4 0.0–0.2	3.7–55.9 0.2–3.1 0.1–1.3	6701–7233 189–204 79–81	1432–1655 29–34 12–14	793–939 13–15 5–6	0.16	0.98	–
2.9–11.8 0.1–0.2 0.0–0.1	16.2–41.4 0.9–2.3 0.4–0.9	6701–6949 189–194 78–80	1602–1781 33–37 13–15	707–886 11–14 4–6	0.18	0.092	0.98
4.4–9.2 0.1–0.1 0.0–0.1	36.0–50.5 2.0–2.8 0.8–1.1	6772–7233 194–204 79–80	1544–1703 32–35 12–14	854–927 14–15 5–6	0.17	0.077	0.98
1.5–7.4 0.0–0.1 0.0–0.1	19.8–43.3 1.1–2.4 0.4–0.9	6736–6843 192–193 78–79	1715–1853 35–38 14–15	793–890 13–14 5–6	0.22	0.059	1.10
3.5–23.3 0.0–0.3 0.0–0.1	18.0–63.1 1.0–3.4 0.3–1.3	8084–9360 228–262 84–89	471–1126 9–23 3–8	976–1226 16–20 5–6	0.068	1.206	0.91

La teneur en hydrogène sulfuré dans l'eau de la nappe arrive à 550 mg/l. La plus grande teneur en H_2S est liée ici avec la plus haute minéralisation. Pour la nappe de cette région sont bien caractéristiques les deux principaux ions: de Cl^- et de Na^+ dont la teneur du premier arrive de 78 à 92 % et du second 76 à 88 % milligramme-équivalents.

Selon la classification de SZCZUKARIEW, ce sont des eaux du groupe chloro-sodique. A cause de cela, le pourcentage de la participation d'autres ions est fort réduite (Tableau 2). Pareillement aux zones signalées de différentes minéralisations, la nappe tertiaire se caractérise par la modification graduelle de la composition chimique.

Au fur et à mesure de l'augmentation de la minéralisation, monte, graduellement la teneur en ions de chlore et de sodium. Augmente aussi la teneur en ions Mg^{2+} et HCO_3^-, ainsi que de H_2S. Diminue, par contre, la teneur en ions Ca^{2+} et SO_4^{2-}. Dans l'étude de la nappe tertiaire, se font remarquer

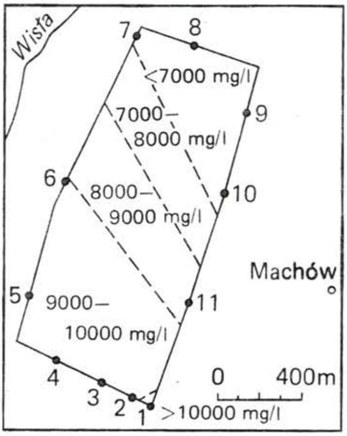

Fig. 3. Carte de la teneur en chlore

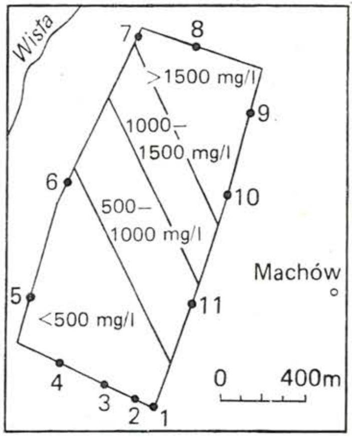

Fig. 4. Carte de la teneur en sulfures

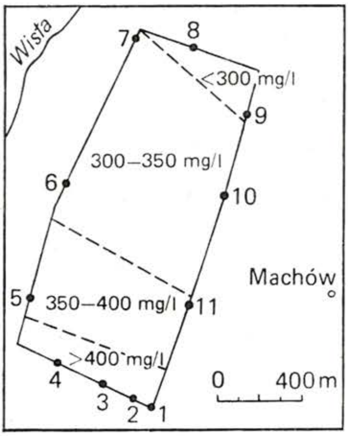

Fig. 5. Carte de la teneur en hydrogène sulfuré

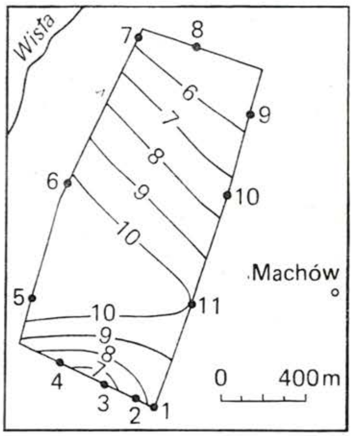

Fig. 6. Carte de la teneur en baryum

aussi — comme composants caractéristiques — le baryum et le strontium. Dans la répartition du ion de Ba^{2+}, on ne constate aucune régularité ou relation, ni avec le rang de la minéralisation générale, ni avec la profondeur. Quant aux réhaussements inexplicables de la teneur en Ba^{2+}, on pourrait, par exemple, les voir liés au fait — constaté dans la région — d'une augmentation locale et irrégulière de baryum dans les calcaires à soufre.

Les zones de minéralisation de la nappe tertiaire et la disposition des ions principaux présentent les cartes — constituant des graphiques correspondants — sous forme de pièces jointes: 2, 3, 4, 5, 6.

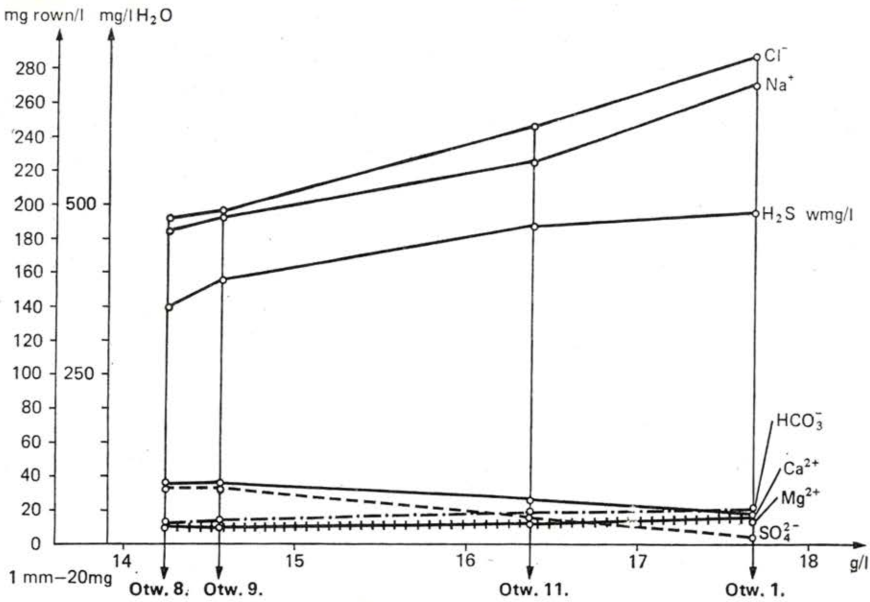

Fig. 7. Diagramme de la composition chimique de la nappe tertiaire

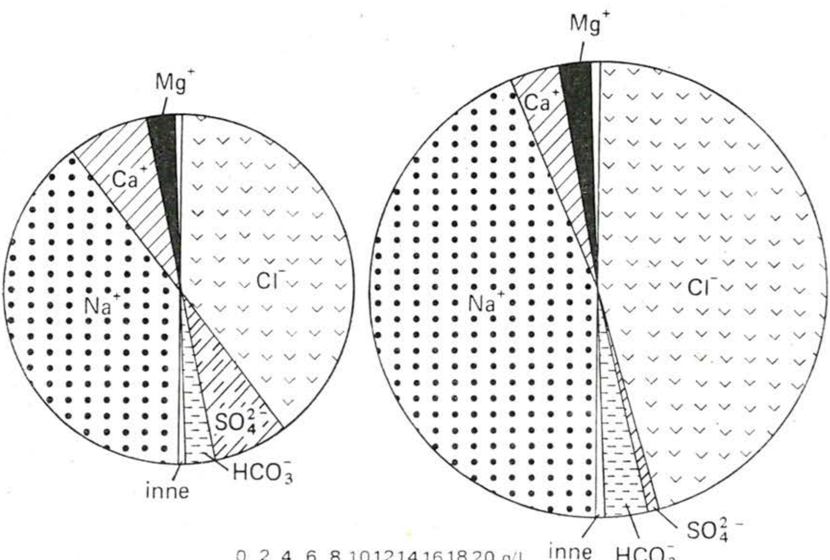

Fig. 8. Graphique de la composition ionique envers
la minéralisation moyenne

La région de l'exploitation par forages

La nappe tertiaire de cette région n'est pas reconnue à ce point, que dans celle de l'exploitation à ciel ouvert, car cela est rendu impossible par la technologie appliquée pour l'extraction du soufre, durant laquelle l'eau de la nappe subit de grandes modifications.

Initialement, la nappe démontrait une minéralisation de 4 à 7 g/l, une dureté générale de 60 à 120°n, ainsi que la teneur en hydrogène sulfuré de 120 à 250 mg/l. Selon leur composition ionique, ces eaux sont en majeure partie des quadri-ioniques eaux sulfuro-chloro-calcio-sodiques.

La composition détaillée est présentée dans le tableau 3.

Tableau 3

Comparaison des compositions chimiques de la nappe tertiaire dans la région de l'exploitation du soufre par forages

N° d'analyse	Minéralisation mg/l	pH	Dureté totale	Absorption de CO_2 mg	H_2S mg/l	CO_2 mg/l	Cations mg/mval			Anions mg/mval		
							Na^+	Ca^{2+}	Mg^{2+}	Cl^-	SO_4^{2-}	HCO_3^-
1	4310	6.7	56.0	520	204	292	950	360	23	1100	1184	550
							41	18	1	30	24	9
2	4802	7.4	88.5	546	122	371	925	589	26	950	1888	369
							40	29	2	26	39	6
3	5002	7.0	67.5	456	171	255	1140	407	45	1330	1434	532
							49	20	3	37	29	8
4	5500	6.9	83.0	400	244	173	1225	539	31	1430	1734	466
							53	26	2	40	36	7
5	5838	7.0	100.0	202	199	165	1240	636	47	1575	1871	424
							53	31	3	44	39	7
6	6300	7.4	129.0	143	219	39	1150	793	77	1750	2100	378
							50	39	6	49	43	6
7	7030	7.3	120.5	184	106	110	1925	721	84	2425	2317	457
							83	36	6	68	48	7
8	7948	6.8	107.5	201	167	127	1932	800	88	2475	2314	426
							84	40	7	69	48	6
9	8320	7.3	132.5	288	110	103	1975	800	88	2525	2368	484
							85	40	7	71	49	7
10	8695	6.9	118.7	184	180	132	2207	716	80	3025	2191	424
							96	35	6	85	45	7

Problèmes physico-chimiques défavorables dans le procédé de l'exhaure de la mine à ciel ouvert

L'existence de deux nappes aquifères de caractères chimiques différents et séparées par une puissante couche imperméable, a conditionné l'application de deux systèmes indépendants de l'exhaure : à part pour la nappe quaternaire et à part pour celle tertiaire.

L'exhaure de la nappe quaternaire n'a pas présenté de problèmes, et pour cela ne fera pas partie du rapport présent. Quant à la nappe tertiaire, son exhaure consiste en principe sur l'application des barrières de puits forés autour du champ d'abattage, au début à partir de la surface du terrain, en suite, à des niveaux inférieurs.

Les difficultés principales dans la construction et l'exploitation de ce système de l'exhaure, découlent des propriétés hydrochimiques de la nappe tertiaire, riche surtout en hydrogène sulfuré — ce qui a demandé l'application de filtres spéciaux: de résistance nécessaire et anticorrosifs, en même temps. Un problème à part et pas de moindre importance présentent les dépôts, dûs à la coprécipitation causée par la modification de l'équilibre physico-chimique de la nappe par les effets de l'exhaure. Ce facteur s'est montré primordial, tandis que la corrosion n'entrait pratiquement pas en jeu, car dès les débuts de l'exhaure de la nappe tertiaire, les dépôts ont été constatés aussitôt et en grande échelle. Déposés dans les pompes, ils réduisent vite leur rendement et les mettent hors d'usage. Le même problème concerne les collecteurs d'évacuation de l'eau, dont la section subit une réduction progressive à relativement vive allure. En dehors du côté technique, l'accroissement continuel des dépôts dans les pompes et collecteurs influence fort défavorablement l'économie et l'organisation des travaux de l'exhaure.

Suite de grippages, causés par les dépôts bloquant vite les pompes, intervient la nécessité de bien laborieux échanges de pompes et collecteurs ainsi que de nettoyages. En conséquence, cela cause de grandes fluctuations dans le nombre des pompes en marche du potentiel installé de l'exhaure et prolonge automatiquement sa durée.

Les études visant le discernement complet de ces phénomènes et leur élimination définitive, n'ont pas donné jusqu'alors des résultats totalement satisfaisants. Les eaux fournies par l'exhaure de la nappe tertiaire sont épurées dans un atelier spécial. Le procédé de l'épuration se compose de deux opérations essentielles:

— désorption de l'hydrogène sulfuré de l'eau dans l'air,

— absorption de l'hydrogène sulfuré par l'hydroxyde de fer avec la récupération du soufre.

L'eau, totalement épurée, est évacuée en suite pour utilisation industrielle ou directement aux cours d'eaux superficielles.

Problèmes hydrochimiques dans l'exploitation du soufre par forages

L'exploitation du soufre par forages — dans la méthode de sa fusion souterraine au moyen de l'eau chaude (Frasch) — cause automatiquement des effets caractéristiques dans le milieu naturel. En premier lieu, cela s'exprime par la modification des conditions hydrauliques de la nappe tertiaire et de sa composition chimique, à la suite de l'introduction par forages de l'eau chaude sous pression, provenante des cours d'eau de surface.

Pour l'étude des modifications de la nappe tertiaire, on échantillonne les forages d'exploitation, les sondages de décompression et les sondages d'observation. Sur la base des analyses des eaux tertiaires, sont dressées périodiquement des cartes de la minéralisation de toute cette région. On y constate des modifications, qui démontrent parfois une réduction de la minéralisation à un tiers (1/3) ou même à un quart (1/4) la minéralisation initiale dans les enceintes des champs d'exploitation.

La vitesse, le caractère et l'étendue des modifications dépendent des directions privilégiées pour l'écoulement des eaux de la nappe à travers le gisement, ainsi que de l'intensité de la décompression appliquée pour l'exploitation.

L'étude des modifications du chimisme, températures de la nappe et des pressions, facilite la constatation des changements de conditions physico-chimiques de la nappe tertiaire et quaternaire au cours de l'exploitation et permet d'appliquer, le cas échéant, les interventions qui s'imposent.

Conclusions

L'auteur voudrait souligner dans son rapport l'importance des recherches de chimisme des eaux pour résoudre les problèmes techniques et les résultats des recherches qui durent jusqu'à présent, donnant la caractéristique des eaux de région d'exploitation des mines.

Tous les problèmes ne sont pas encore résolus. L'exploitation de soufre sur une grande échelle permet aux hydrogéologues polonais de continuer de nouveau les recherches très intéressants sur le chimisme des eaux.

A CONTRIBUTION OF TRITIUM ANALYSES TO THE STUDY OF GROUNDWATER FLOW IN THE SEDIMENTS OF ŽITNÝ OSTROV

UNE CONTRIBUTION DE L'ANALYSE DE TRITIUM À L'ÉTUDE DU MOUVEMENT DES EAUX SOUTERRAINES DES SÉDIMENTS DE ŽITNÝ OSTROV

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RÉSUMÉ

On a étudié l'activité du tritium des eaux souterraines dans la région de Žitný ostrov. Les échantillons furent pris lors du forage de neuf puits. La détection d'une activité élevée (300 – 400 T. U.) dans le puits P_1/XIV avait permis de préciser la vitesse d'écoulement réelle dans l'intervalle de 60 m. Se basant sur l'interprétation des résultats des analyses, la région entière fut divisée en trois zones suivant la vitesse et la façon d'écoulement des eaux souterraines. La première en est caractérisée par l'influence de l'infiltration intense à partir du Danube, la seconde par la diminution de l'influence du Danube avec la prédominance de l'écoulement horizontal supposé. Dans la troisième zone l'eau est barrée par l'influence de la structure géologique et il y a l'expulsion des eaux plus âgées à la surface. On à aussi précisé le secteur du Danube avec l'infiltration la plus intense.

Introduction

Research concerning the groundwater movement in the sediments of Žitný ostrov is based on the observation of groundwater level in shallow observation wells. The knowledge thus acquired is used in solving several problems, for instance, detection of groundwater resources, rate of flow, etc.

It was found by the analysis of groundwater tritium activity that the notion acquired by up to date methods on many problems pertaining to groundwater movement is considerably simplified and frequently inaccurate. Inaccuracies occur mainly when, apart from the study of groundwater regime, also other natural phenomena and factors are not taken into account. Difficulties arise also from the fact that the observation network is not ideally distributed. Frequently it is necessary to consider the analogy with adjoining areas, or extrapolate the data to wider areas.

The tritium analysis of groundwaters also helps to solve some of these problems. The water samples taken by the author of the present contribution were analysed at the Department of Nuclear Physics, Faculty of Natural Sciences of the Comenius University, Bratislava (S. Saró).

Review of present-day opinions

The groundwater movement in the sediment of Žitny ostrov was investigated mainly by the personnel of the Groundwater Management Research Institute (BARTOLCIC, DUBA, GYALOKAY, LEHKY, SUPEK), then by PORUBSKY, KLAGO and others. GAZOVIC summed up practically all present opinions in his work (1972).

Generally accepted is the view-point that the groundwater conditions of the sediments in the upper and central part of Žitný ostrov are mainly governed by the bank-infiltration from the Danube. In proportion to this, the infiltration of atmospheric waters is of little quantitative importance. The most intense recharge from the Danube is in the sector reaching from Bratislava to Hamuliakovo (Fig. 1). The recharge in this sector forms the main groundwater flow, in SE direction, thus, almost parallel with the Danube. The second flow proceeds nearly parallel with the Little Danube, mainly along its northern bank. In the Bratislava region, it is supplied by affluents from the Little Carpathians and according to certain authors also by the Danube. The third groundwater flow, affects mainly the terrain north of Žitný ostrov, proceeds in N–S direction from the Váh river sediments. The main groundwater flow is slowed down by fine-grained sediments up to the high Neogene blocks near Kližská Nemá, built predominantly by clayey sediments. The dammed water is discharged by drainage channels (Čalovo-Okoč area) – JAKUBEC–PORUBSKY (1962, Fig. 3).

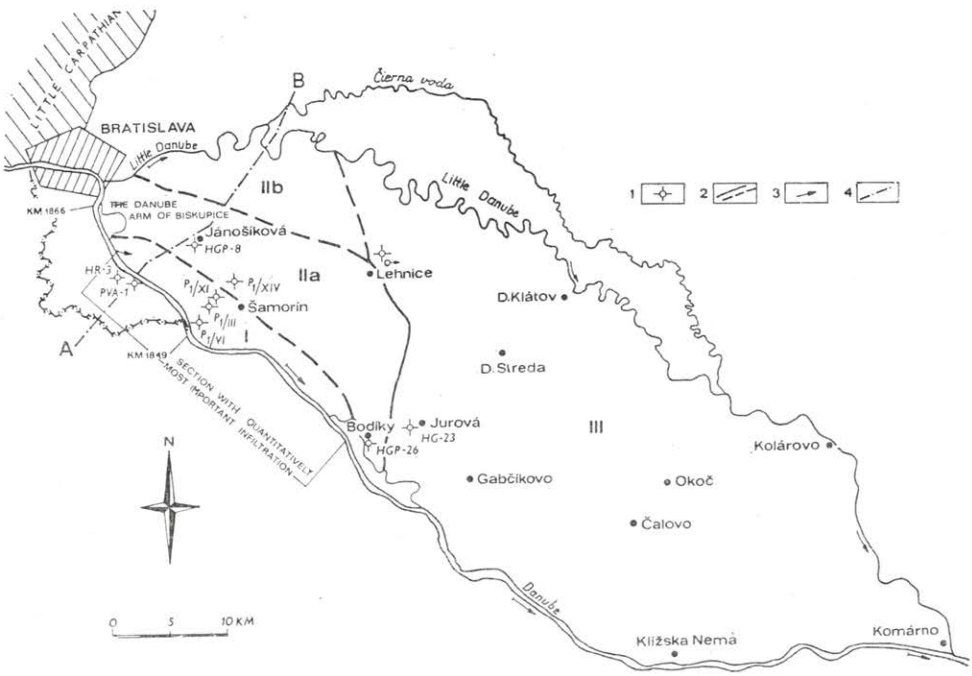

Fig. 1. Site of the studied area

1. Hydrogeological borehole for tritium analyses, *2.* boundaries of hydrogeological zones, *3.* place of infiltration of Danube water, flowing towards the borehole P_1/XIV, *4.* geological cross-section

The Žitný ostrov area is divided into three zones according to the influence of the Danube. The continuous infiltration under all flow conditions is characteristic for the near-river narrower zone. The Danube bed is formed on the top of an alluvial cone, so even during minimum flow, the water level in the river remains above the groundwater level. However, the amount of infiltrated water is changing. According to PORUBSKY et al. (1971) this zone extends almost to Gabčíkovo, according to DUBA up to Kližská Nemá. The direction of groundwater flow in this area is oblique to the Danube bed. The wider near-river zone can be characterized by short-term supply and long-term discharge periods of groundwaters. The groundwater level is affected only by significant river level fluctuations or by flow conditions of longer duration. According to PORUBSKY et al. (1970) also the drainage channels play important role in the second zone. Hydraulic relations between the groundwaters and the recipient change seasonally. The groundwaters are discharged and recharged alternatively in time.

The near-river outer zone has a groundwater level continuously lower than the minimum river water level. PORUBSKY et al. (1971) delineates this area in the region where affluents occur from the left-bank side of the Little Danube of Považie, Trnavská (hilly country) pahorkatina, Čierna voda and Subcarpathian area.

The problem of groundwater flow-velocity was dealt with by GYALOKAY, M. (1960). He states that the real flow-rate in the western part of Žitný ostrov (island formed by the Great and Little Danube) is 4 – 6 m/day. Based on this rate of movement he calculates the discharge rate of groundwater in the cross-section. Meanwhile he assumes that the water flow is homogeneous in the total depth of the aquifer.

Groundwater flow in the light of the tritium-activity studies

From the opinions mentioned in the previous chapter it can be concluded that the Danube is the main source of the groundwater. Atmospheric waters contribute only in a small part to the direct supply of groundwaters. However, it is necessary to consider the fact that the Danube-water originates predominantly from precipitation. Only in the winter season is the discharge supplied by groundwaters to significant degree. For the analysis and correct interpretation of the tritium activity of groundwaters, it is necessary to know the activity of Danube water and that of the precipitation in a longer time interval. These data were taken from the works of SÁRO, S. (1975), Figs. 2 and 3, as base for the judgement and classification of the groundwater in certain time periods. From Fig. 2 it can be seen that for the detailed observation period from 1972 – 1975 the activity fluctuated between extreme values from 50 to 250 T.U. The most part of the analysed samples showed an average activity from 120 to 140 T.U.

In the predominant part of the period the Danube-waters had higher activities than the atmospheric ones with an average of about 160 to 180 T.U. This phenomenon is caused probably by the nuclear power plant existing on the upper reach of the Danube. For the whole period interesting from the tritium-activity point of view (from 1950 to present) there is no precise activity measurement of the Danube water, only that of the atmospheric one

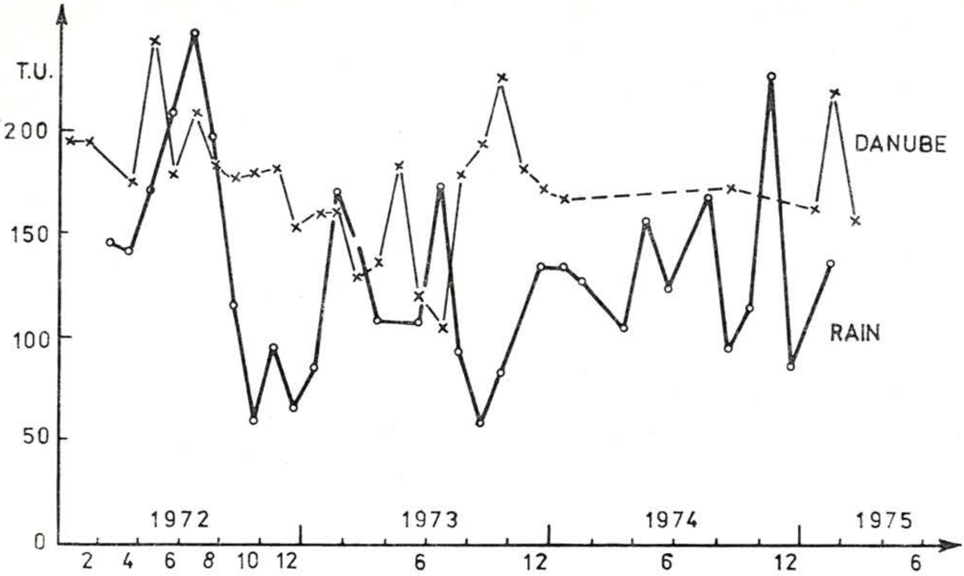

Fig. 2. Tritium concentration of rains in Bratislava (according to Sáro, S.)

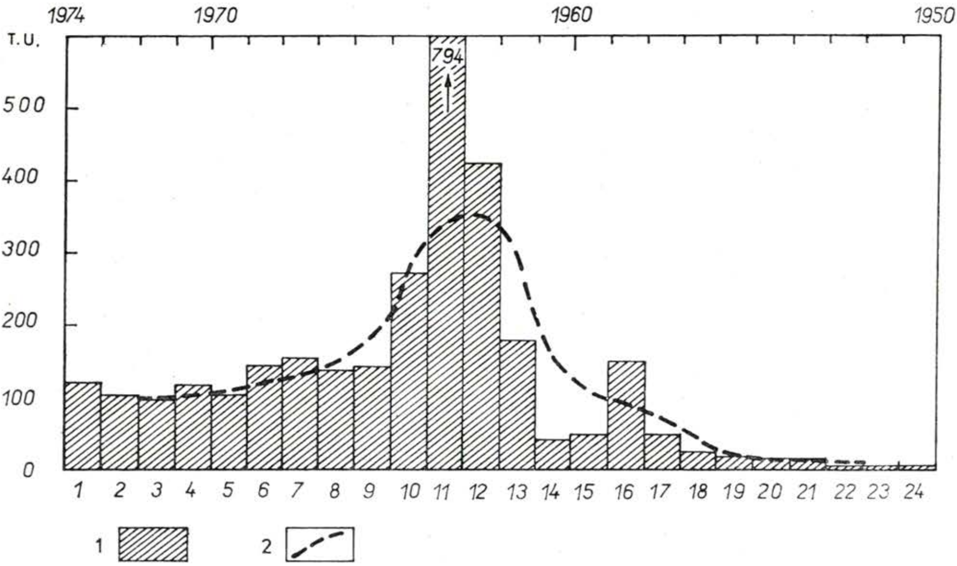

Fig. 3. Tritium concentration of rains for the period 1950—1974, yearly averages (according to Sáro, S.)

1. Residual activity of precipitation, *2.* probable distribution of present-day tritium activity of groundwaters

(Fig. 3). However, these data are sufficient for the purposes of relative dating. An eventual higher groundwater activity (300 to 400 T.U.) will help to place this water to the period of 1962—1964 with a relative certainty. The course of the residual activity from earlier and later periods shows that attempts made for a more precise absolute dating of waters with another activity could lead in several cases to incorrect data. For this reason it was ceased from a more precise determining the absolute age in most cases. The acquired results are used only for distinguishing groundwaters from each other, without determining their age.

What regards groundwater analyses from nine wells available, their distribution is shown in Fig. 4. Samples for the analysis were taken by pumping when sinking the wells. Water inflow to the well was possible only through the bottom. At each sampling only a small quantity of water was drawn—up to 2 l/s. As the aquifer consists of very permeable materials—$k = 10^{-2}$ to 10^{-3} m/s, no major disturbance was caused by pumping. Therefore, the obtained water can be unmistakably related to the depth to which the well was cased.

When comparing the character of tritium-activity changes with the depth in the well on the right side of the Danube (PVA-1), (Fig. 5) with the wells on the left side, their different character must be ascertained. It is important to note that though PVA-1 well is nearer to Danube, the variation of the tritium activity with the depth differs sharply from that in the P_1/VI well. The practically uniform activity in P_1/VI well enables to assume an intense movement of water along the depth, or its intense intermixing, along the left side of the Danube. Therefore, it can be assumed that there exists a very intense immediate infiltration from the Danube in this area, whose depth reach is obviously above 100 m. The infiltration intensity and the rate of water movement is so high that the decrease of activity as a result of radioactive decay did not appear in the analysed waters. In PVA-1 well it is im-

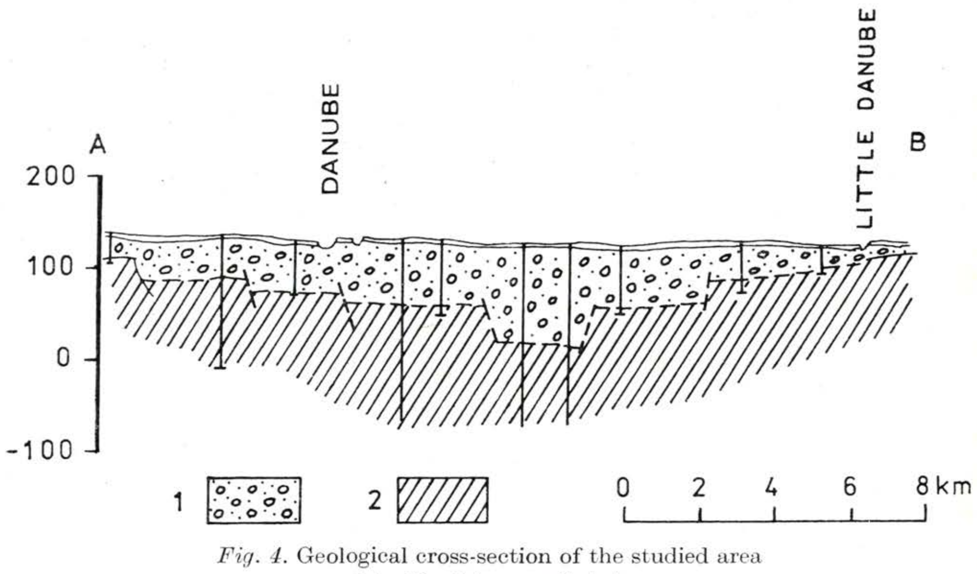

Fig. 4. Geological cross-section of the studied area
1. Aquifer, *2.* impermeable beds

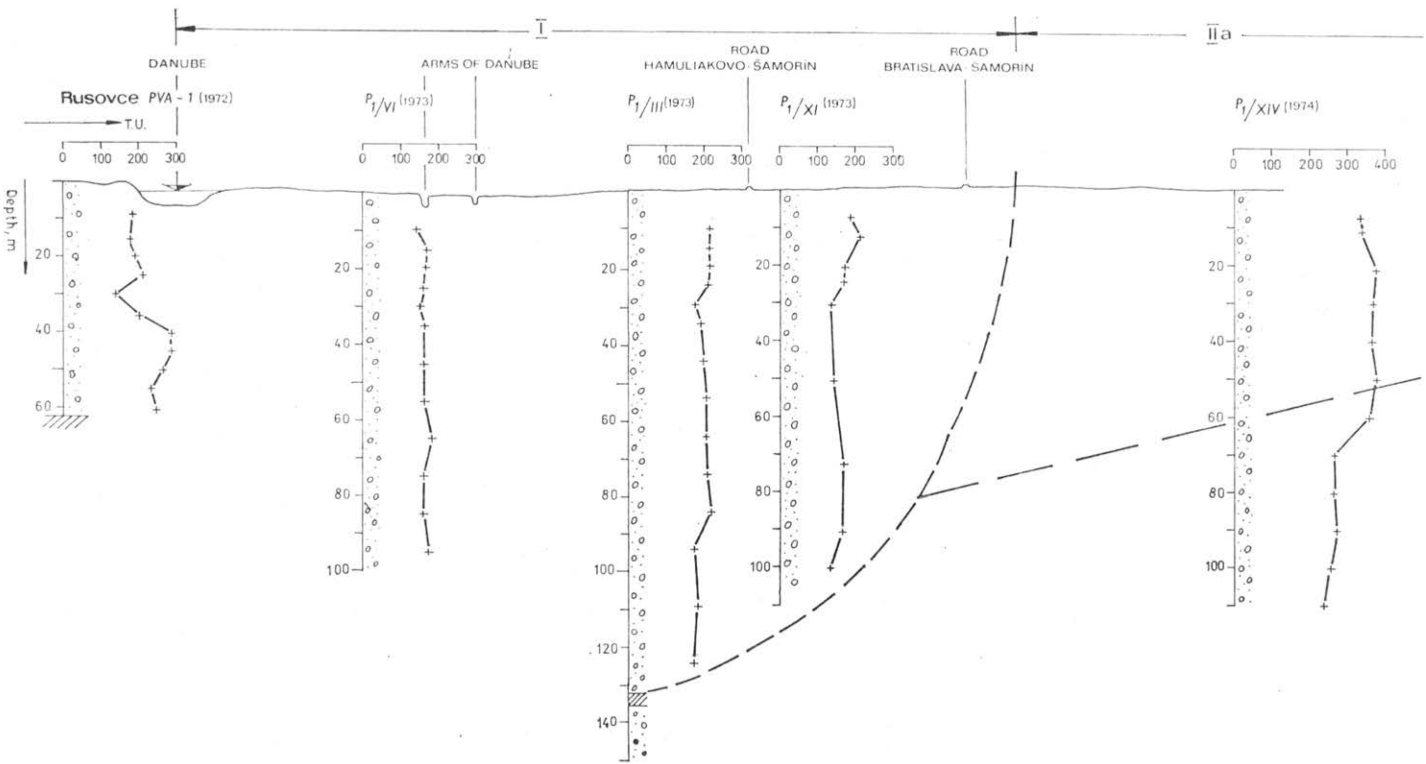

Fig. 5. Tritium concentration of groundwater in function of the depth, along the cross-section

portant to note that water activity to 25 m depth has the same level as in P_1/VI well, meanwhile at 35 m depth an increased activity occurs up to a value about 300 T.U. Taking into account the activity of atmospheric waters, the date of origin of these waters, i.e. from 35 m depth down to the gravel base, should be placed to the period of 1961–1965. Since the water activity in the Danube ranges in the recent years from 160 to 180 T.U., it is obvious that these waters in strata near the base cannot originate in this latter period. Based on data in Fig. 3 they can be considered older than 10 years. It can also be assumed that these waters flow into the region investigated from a greater distance, i.e. the place of infiltration from the Danube is not directly at the PVA-1 well level. In the studied area they occur below the young waters. A precise dating of these young waters, however, is not possible. From the activity levels in P_1/VI well as compared with PVA-1 well, it is possible to judge that the influence of infiltration from the Danube at the level of both wells was more intense along the left-bank side of the river. Attention to this problem was paid even earlier (Gazda–Pospísil 1974). The opinion was formed that water infiltrated from the Danube is on both sides directed towards the Žitný ostrov area as a result of the geological structure. The excentric position of the Danube related to the river-depression structure – causes that the infiltrated waters along the right side of the river have not sufficient space for movement. Therefore, they flow mainly into the Žitný ostrov area. Here the characteristic feature of the geological structure are the deepening of the depression towards south-east, and a considerable accumulation of gravels with sands. The fact creates suitable conditions for subsurface flow of considerable quantities of water.

The changes of tritium activity in Fig. 5 shows the influence of water flow along the left side of the Danube. The results from P_1/VI well give support to the assumption of an intense infiltration from the Danube in the area of this well. Down to a depth of 96 m the value of activity is practically uniform. This fact enables to place the investigated water to a similar period. In the other wells the variation of activity with the depth is more complicated. In P_1/III well the water activity down to 87 m is uniform, differing but very little from the activity of P_1/VI well. It decreases further with the depth. Very significant is the decrease in samples from depths of 125 and 150 m. Samples taken at a maximum depth of 99 m from P_1/XI well are also available. Insignificant decrease of activity in the interval of 88 to 99 m does not allow to draw more important conclusions. The drilling was finished at 100 m depth, i.e. further samples could not be taken. On the other side it cannot be excluded that the decrease of activity at this level indicates a similar course as was found in P_1/III well. In P_1/XIV well important is the fact that the activity of water down to 60 m is high – well above 370 T.U. It is possible to assume that the water in this well infiltrated in the years 1962–1965. Solely in this period can originate the high residual activity of 300 to 400 T.U. The next depth reach (70 to 110 m) shows again an increased activity compared with other waters in this profile of wells. The most probable explanation of this fact is that the water has mixed with waters of another activities. The dominant factor in it must have been water with an increased activity. In comparing the variations of activities in P_1/XI and P_1/XIV wells it is possible to note the change in the water movement in the area between these wells. In P_1/XI as well as P_1/III well it can be remarked an intense direct water influence from the Danube to 100 m or 130 m depth. In P_1/XIV well a change of activity in the 60 to 70 m reach is visible (357 to 264 T.U.) which excludes a uniform horizontal movement in the entire investigated depth-range. At 60 to 70 m depth there probably exists a change in the movement. This causes a contact of two water types of varied age which affect each other but to a small degree. With regard to P_1/II, P_1/III and P_1/XI well, it can be assumed further that the limit of the flow intensity shifts to the level around 65 m.

An intense change in water activity was detected also in HGP-8 well (Jánošíková – Fig. 5). Practically it is in a similarly distance from the Danube as P_1/XIV well. Then it is possible to assume approximately similar conditions of groundwater flow in the vicinity of both wells to about 65 m depth. The gradual but significant decrease of activity with the depth from this level testifies the fact that the age of water increases with the depth in the vicinity of Jánošíková. It can then be assumed that here the influence of the infiltration-intensity from the Danube is very rapidly ascertainable with the depth. This finding supports the view-point on the groundwater concentration in the central part of the depression. It has to be taken into account, namely, that westward from the reach of the Danube marked in Fig. 1 by arrow, the thickness of sediment strata decreases gradually to about 15 m beneath Bratislava. Groundwaters flowing in these sediments of inferior thickness cannot cope equally intensely with supplying the whole area, mainly as the depth increases. The intense groundwater flow is then limited to the deepest part of the depression. Greater depths (about 150 m), however, result here also in slowed groundwater flow and in the decrease of its activity.

The relatively precise determination of the water's "age" in P_1/XIV well from 0 to 65 m depth helps to acquire a certain notion on the rate of its flow. When the determination of infiltration places is based on hydroisohyps (the hydroisohyps were taken from the report of Porubsky et al. 1971) along the reach indicated by an arrow in Fig. 1 on the Danube, the distance to P_1/XIV well is 12.8 km. In case of 10 to 12 years age, the average flow-rate of water in the first shallow zone is 1.06 to 1.28 km/year. In average it is possible to consider it as 1.15 km/year i.e. 3.1 m/day. It is probable that in the vicinity of the river the velocity of groundwater flow is higher and slower at a greater distance from the river and in greater depths. The data on the rate of groundwater flow is near to a lower value of 4 m/day indicated by Gyalokay, M. (1960) for the Žitný ostrov (upper and middle part). The method for detecting the rate of 3.1 m/day – in fact tracing tests of considerable dimensions – allows to consider it for representative in a delineated strike, sector and depth interval.

Interesting subjects to consider are given also by the results of analyses from further wells – Fig. 6. In HGP-26 well at Bodiky it is important to note that although it was sunk in the immediate vicinity of the Danube branch, no similar variation of activity with the depth was detected as in P_1/VI well. The high water activity (> 300 T.U.) in samples from 60 and 70 m depths indicates a possibility to date this water within the period of 1962 – 1964. Water down to 30 m, its activity being about 200 T.U. corresponds to young water as detected in P_1/VI well down to 95 m. Such a change of activity with the depth can be interpreted as limit of the intense infiltration sector starting from the Danube. Signs of groundwater damming are to be assumed in the adjoining area in the close vicinity of the river.

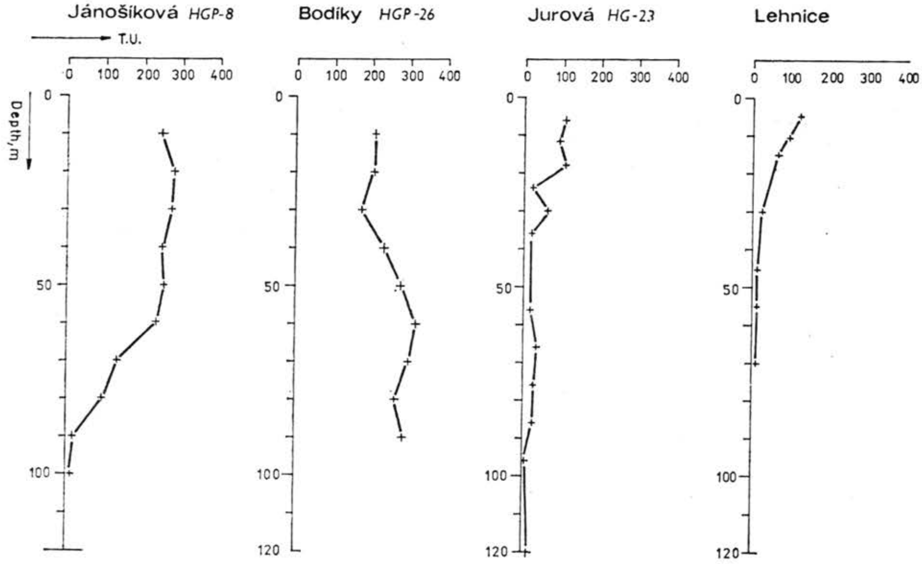

Fig. 6. Tritium concentration in various boreholes of the Žitný ostrov

The well in Jurová has a very interesting feature of activity. Although sunk at an equal distance from the Danube as are HGP-8 and P_1/XIV wells, the activity even in the shallowest positions exceeded but very little the value of 100 T.U. A very low activity down to 36 m confirms the assumption expressed in the work of Gazda–Pospísil (1974) that the influence of Gabčíkovo faults appears in damming the groundwaters well outside the area, where they occur. A very low activity near the subsurface can be explained by the damming of older groundwaters with a low activity. A low activity of about 100 T.U. down to approximately 20 m tells of the cessation of "young" waters, or of their considerable mixing with waters from great depths. Similar results were found also by activity measurement in the area of Lehnice, Fig. 6.

Based on the analysis of tritium activity in the whole studied area the narrowest zone adjoining the Danube can be delineated. The depth of intense exchange of water as a result of infiltration from the Danube exceeds 100 m in it. In the work of Gazda–Pospísil (1974) this zone was given the name of the groundwater forming place (Fig. 1, area I). In the adjoining area a zone of transport and mineralization is delineated. Here, approximately in the middle of the strata-series, a change in the age of the water can be observed. It divides the aquifer into the upper part with a relatively higher flow velocity (approx. 1.15 km/year) to a depth of 60 to 70 m and the lower part with a slower flow. The dividing surface between them is dipping. It cuts the surface of the area in the neighbourhood of Lehnice. It makes possible to divide this zone into two areas (IIa and IIb). In the outer one (IIb) a relatively uniform horizontal flow can already be assumed. Finally a zone of discharge (area III) was delineated in the eastern part of Žitný ostrov in which the groundwaters are squeezed out by the influence of the Gabčíkovo faults and refined sediments. Contrasted with the present knowledge, an influence of faults is therefore assumed as well as their damming effect in the area around Bodíky, Jurová and Lehnice. In the case of Lehnice, intense water outcrops of annual appearance in the draining channels are also considered as evidences of damming.

If the infiltration from the Danube and its areal effect are to be explained in the light of these data, one can express the assumption that quantitatively it is more important from the Biskupice branch flow to the Danube downstream to Bodíky. The upper boundary is confirmed also by the fact that the influence of polluted Danube water originating from the waste of the Slovnaft oil refinery and chemical industry can be followed in groundwater only close downstream of their flow into the river. At a greater distance the influence of pollution is not observed, therefore, an intense infiltration cannot take place in the vicinity of the waste inflow or downstream of it.

Conclusions

The tritium analyses and the knowledge acquired in Žitný ostrov enabled to clear up the characteristics of the groundwater flow in the sediment strata of the area. According to these facts the flow-rate of groundwater masses is not uniform in the entire thickness of the aquifer. The effect of hydraulic factors and the geological structure results in considerable anomalies in the movement of the groundwater, mainly in changes of flow-rates depending on the depth.

The determination of age of water taken from P_1/XIV well in 60 m depth as 10 to 12 years enabled to evaluate the real rate of water flow at this depth quite accurately as 3.1 m/day. This value holds mainly for zone IIa. In zone I a higher rate of water flow can be assumed and slower in zone IIb. In zone III with regard to water damming and outcrop of older waters to the surface the slowest flow can be assumed. The suggestion for dividing the water-bearing sediments in space is supported mainly by the analyses of tritium-activities. It will obviously be completed in the future. From practical point of view, a delineation of the section with the most considerable infiltration is significant. It is of immediate importance for the utilization of groundwaters in this area, for their protection and preservation their recharge conditions in the future.

Concerning the tritium analyses, it should be noted that circumstances will not always allow a precise water dating in complicated conditions. This is possible mainly in thick sedimentary complexes but only in exceptional cases. The tritium analyses, however, could frequently help to clear up facts which cannot be revealed by other means. In this way it is possible to distinguish waters of various age from each other, acquiring a more detailed knowledge of water flow even in spatial sense.

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GEOTHERMAL MAPPING IN CENTRAL AND EASTERN EUROPE

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Introduction

Within the Planetary Geophysical Commission of the Academies of socialist countries (KAPG) a Geothermal Working Group is active (member countries: Bulgaria, Czechoslovakia, German Democratic Republic, Hungary, Poland, Romania, Soviet Union). In the last ten years this working group concentrated its work on the geothermal mapping of the member countries. Two kinds of maps have been drafted: heat flow maps and geotemperature maps.

Heat flow maps

In the above mentioned countries the first data on heat flow were published by Stenz (1954) for Poland, Boldizsár (1956) for Hungary, Schlosser and Schwarzlose (1959) for the GDR, Lubimova et al. (1961) for the USSR and Cermák (1967) for Czechoslovakia. Up to the present time, 812 measurements have been published or are in the final stage of completion.

Fig. 1 shows the present state of heat flow measurements in Europe (Cermák, 1975).

National works

The number and density of heat flow measurements in the individual KAPG countries varies greatly. As an example, Fig. 2 shows Czechoslovakia, a thoroughly surveyed country. Here the total number of heat flow determinations is 81 and the density comes to 1 measurement per 1500 km^2. The determinations were carried out by the Geothermal Laboratory of the Geophysical Institute of Czecho-Slovak Academy of Sciences, using an automated divided bar apparatus and computerized evaluation.

International works

The geothermal instruments in the laboratories of various countries were calibrated by using crystalline and fused silica etalons and jointly measured rock samples.

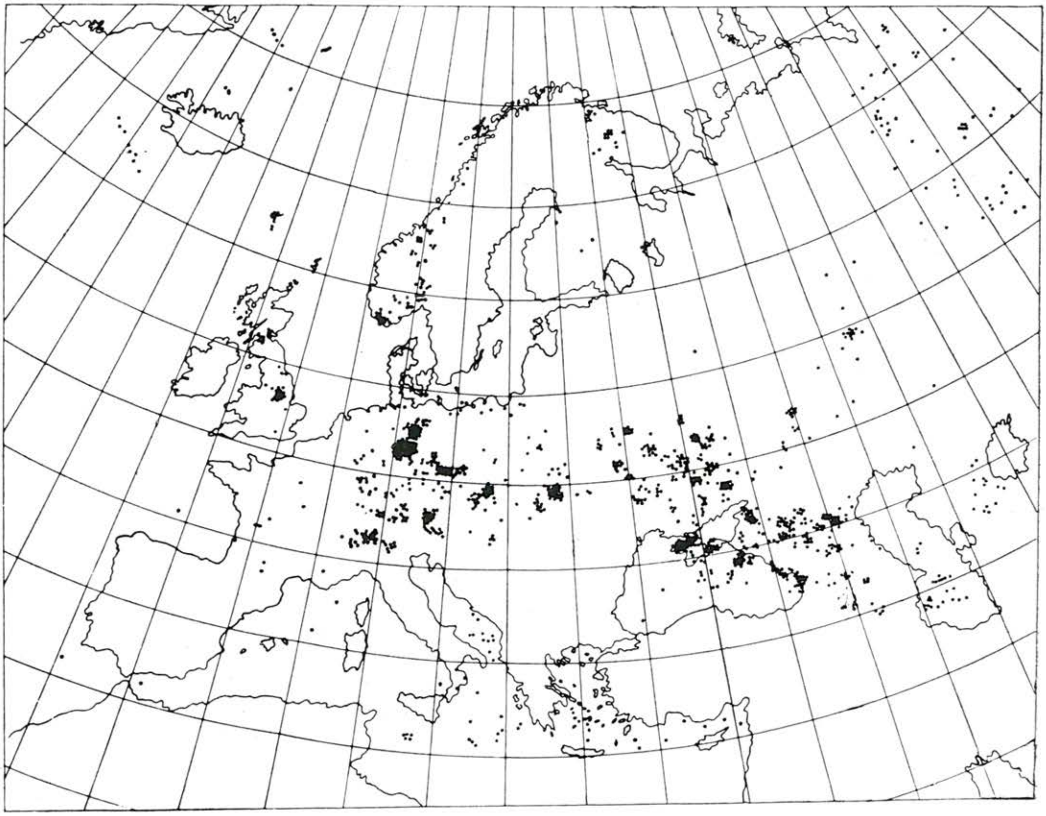


Fig. 1. Present state of heat flow determination in Europe, 1975. Jan.

Generally, the density of heat flow stations is of about 5–6 points in a $5° \times 5°$ square in stable tectonic areas of the Precambrian shields and of ancient platforms, and 3–4 points in a $1° \times 1°$ square in tectonically active areas of Alpine folding. The arithmetic mean of 812 values is 57.1 ± 27.1 mWm^{-2} ($= 1.36$ early HFU) which is lower than the mean value 1.5 ± 0.15 HFU proposed by LEE and UYEDA (1965) (Table 1).

The international heat flow map of the KAPG countries has a scale of 1:10,000,000. Multicoloured, with an isolines' interval of 15 mWm^{-2}. Fig. 3 shows a generalized black-and-white version.

Corrections. There may exist a more or less pronounced influence of such geological phenomena on observed heat flow as sedimentation, erosion, uplift, as well as the effects of past climatic changes. At present, no sufficient information is available for the application of reasonable corrections. Accordingly, in the present state of map construction preference was given to uncorrected values rather than introduce something what might be a subjective standpoint. This criterion has nothing to do with the application of technical corrections such as for local effects, conductivity contrast, borehole inclination, and with omitting some data obviously disturbed by underground water movements (e.g. from the hot spring areas of Teplice, CSSR and of Hévíz, Hungary).

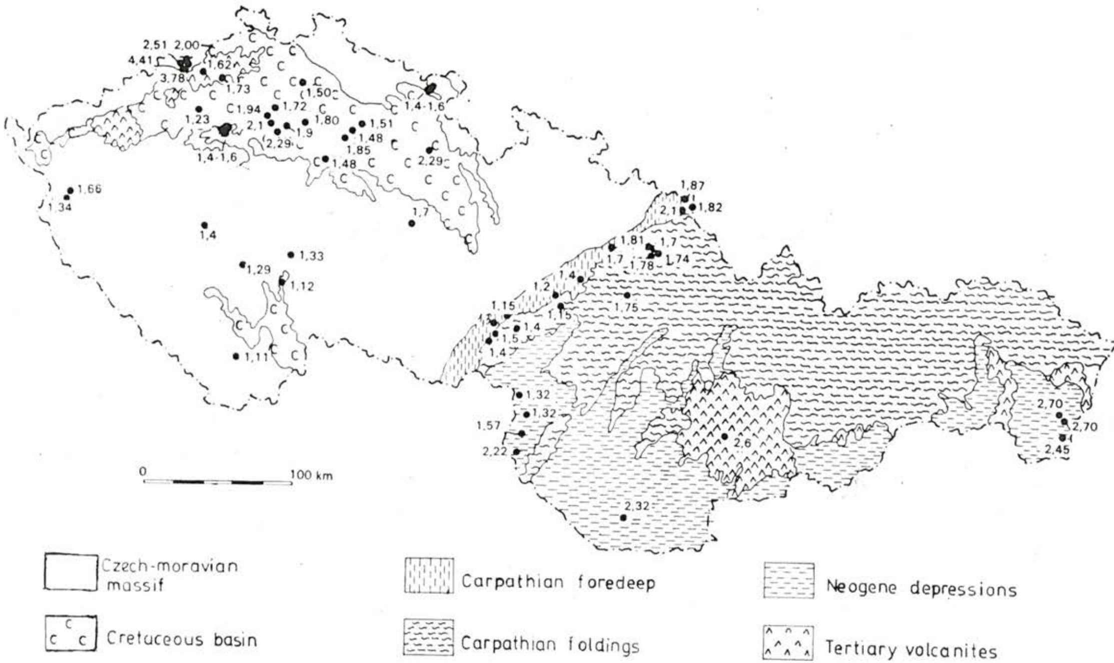


Fig. 2. Heat flow stations and generalized tectonics in Czechoslovakia. Heat flows are given in μcal/cm^2 sec

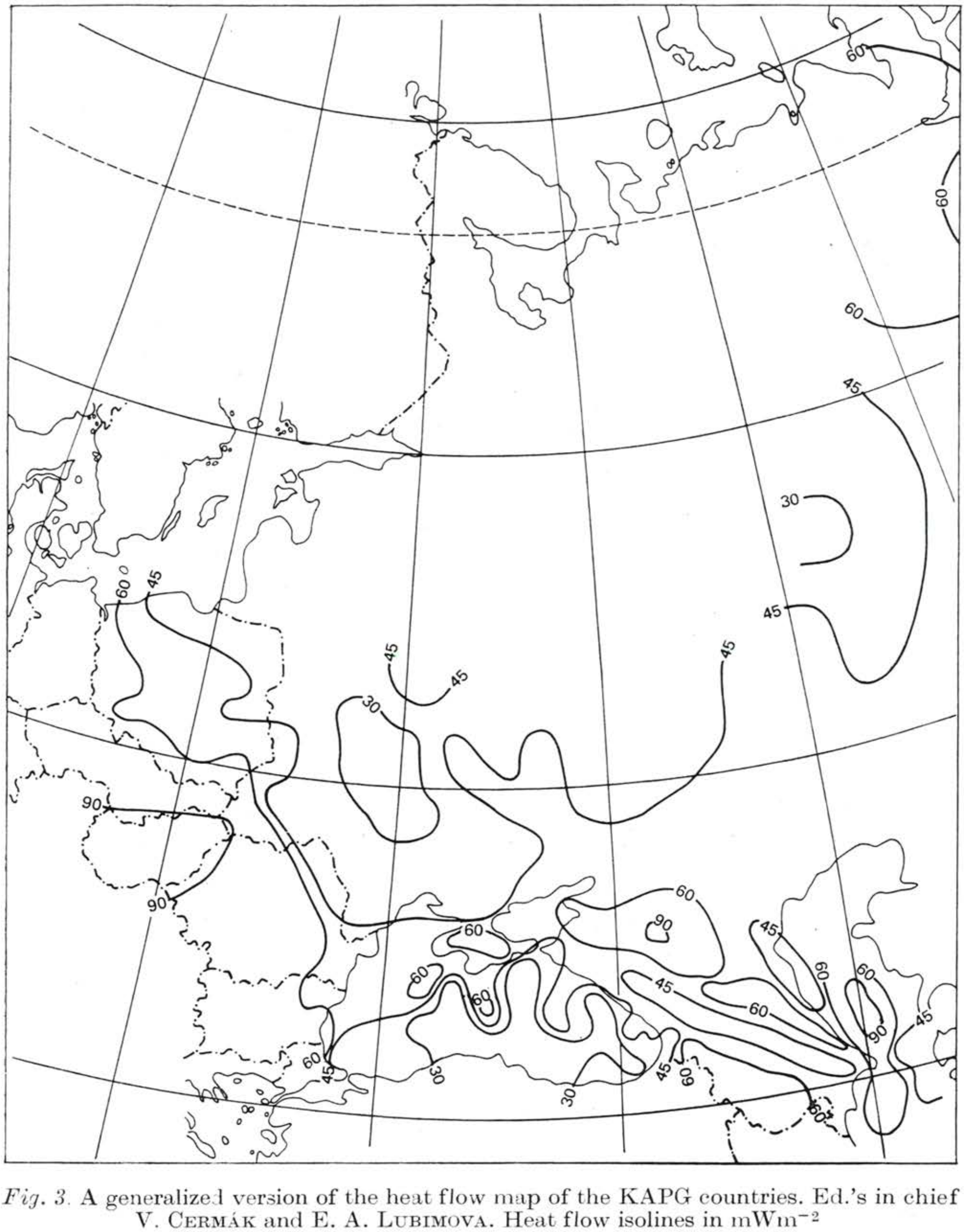


Fig. 3. A generalized version of the heat flow map of the KAPG countries. Ed.'s in chief V. Cermák and E. A. Lubimova. Heat flow isolines in mWm^{-2}

Table 1

Heat flow statistics for major geological units in Central and Eastern Europe

Geological units	Number of values N	Means in old HFU	mWm^{-2}	Standard deviation mWm^{-2}
1. Precambrian shields	83	0.86	35.8	6.9
Baltic shield	22	0.83	34.6	4.9
Ukrainian shield	61	0.87	36.2	7.5
2. Post-Precambrian platforms	325	1.33	55.5	17.1
Ancient platforms (Russian)	182	1.09	45.4	8.7
Epi-Paleozoic platforms				
Scythian plate	127	1.67	69.4	17.1
North-German Lowland	16	1.41	59.0	8.7
3. Post-Precambrian orogenic areas	321	1.51	63.0	19.8
Paleozoic folded areas	144	1.61	67.3	14.6
Central Europe	118	1.64	68.6	15.6
Ural Mountains	26	1.47	61.3	6.3
Mesozoic-Cenozoic areas	177	1.42	59.5	22.5
Frontal foredeeps	86	1.25	52.1	12.5
Intermountain basins	48	1.55	65.0	31.4
Faulted systems	43	1.63	68.3	21.6
4. Seas	77	1.12	46.9	21.6
Black Sea	56	0.97	40.4	16.1
Caspian Sea	21	1.54	64.4	24.7

Construction of isolines. For the construction of isolines, mathematical methods (e.g. linear interpolation, higher polynomes, filter theory) or smoothing to geological structures can be used. In this case, the last one was used, geothermics often being connected with geology. In addition, the map of geoisotherms for the depth of 1 km was used to help the construction of heat flow isolines.

Geothermal temperature maps

In the KAPG countries geotemperatures have been measured in many thousand boreholes. Consequently, the construction of a geotemperature map is better backed up than that of heat flow maps, although temperature data are sometimes affected with errors.

National maps

Each of the KAPG countries has prepared its own geotemperature map, for one or more depth levels. In the well-surveyed Hungary, based on 1225 boreholes, geotemperature maps have been drafted for several depths (0.5, 1, 1.5, 2, 3, and 4 km), in the scales of 1:500,000 and 1:1,500,00 (Fig. 4).

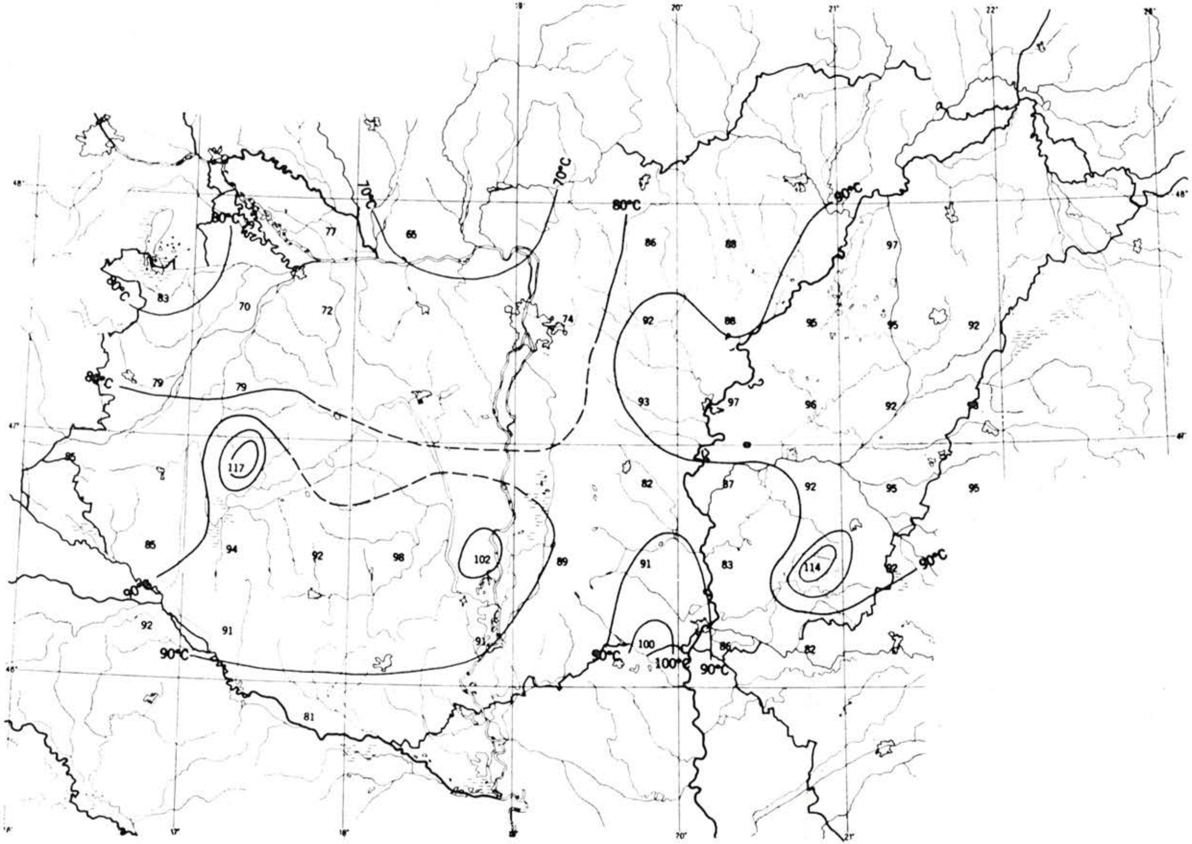


Fig. 4. A generalized version of the geotemperature map of Hungary, for the depth of 1.5 km. Ed. L. STEGENA

Reliability of data. Temperature data measured in boreholes often have unknown errors. To eliminate the incorrect data, the comparison of measurements at various depth and in neighbouring boreholes and consideration of the geological setting was used.

Corrections. Apart from technical (e.g. inequilibrium) and local corrections (neglecting of anomalous fault- or local superthermal-zones), no general correction (topographic, paleo-climatic) was used.

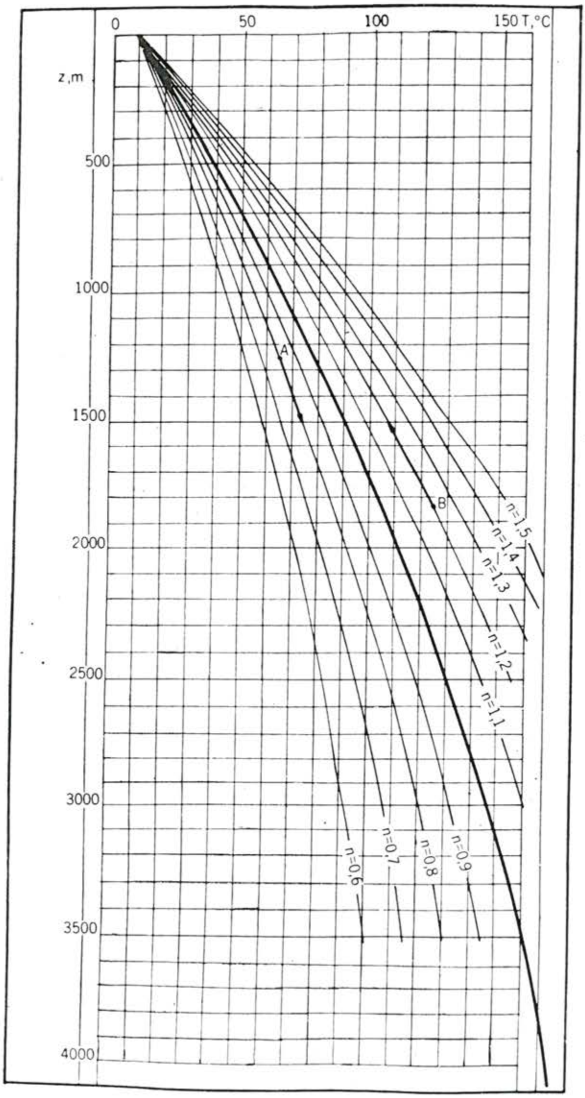


Fig. 5. Mean depth function of geotemperatures in Hungary, $T(z)$ and master curves used for the reduction of temperatures to depth ($n = 0.6 - 1.5$)

Reduction to depth. For the reduction of observed temperatures according to the depth levels of map construction, one has to use some hypothesis for the heat flow or the thermal resistivity of rocks. In Hungary, a so-called method of master curves has been used for the depth reduction (Fig. 5). After constructing reliable temperature-depth functions for the whole of Hungary and for its larger areal units as well as for the sheets of the 100,000 scale maps, it was observed that these functions are similar with a satisfactory approximation i.e. we have

$$T_1(z) = nT_2(z)$$

where n is a varying number with: $0.6 < n < 1.5$ and the surface value of temperature is given by the yearly average (10 °C). Using this method, individual geotemperature data could be reduced to the depths chosen for the construction of isotherms, without any particular assumption.

The international map

The international map of the KAPG countries is based on the national maps and/or on data banks. The multicoloured map has a scale of 1:10,000,000, the interval of isotherms is 5 °C. The map refers to the depth of 1 km from the surface. As regards the corrections, only technical and local corrections have been made, just as in the case of the heat flow map.

The geotemperature map, especially in the generalized version of Fig. 6, reflects the same essential features, as the heat flow map.

Geological implications

The above geothermal maps, constructed for a big area using common principles of construction allow to draw some not unessential geological conclusions.

Heat flow and tectonics

Fig 7. shows the correlation between mean heat flows and tectonics for the European part of the USSR and for Czechoslovakia. Fig. 8 delineates the connection between geological structure and heat flow for the whole of KAPG countries. From this formal statistics, as well as from the map (Fig. 3) one can see that the heat flow pattern can well be correlated with the major geological structures. Areas of approximately similar heat flow can be easily distinguished. Precambrian shields are characterized by a relatively low heat flow of 35.8 ± 6.9 mWm^{-2} with almost no regional variations. Uniform heat flow field is also typical for the old platforms: East European or Russian platform 45.4 ± 8.7 mWm^{-2}, North German platform 59.0 ± 8.7 mWm^{-2}. Considerable scattering of observed heat flow and increased geothermal activity is obvious for the areas of Alpine structures, quite prominently in intermountain depressions filled by Neogene sediments, such as the Pannonian basin (as more as 100 mWm^{-2}), or in areas affected by Tertiary volcanism

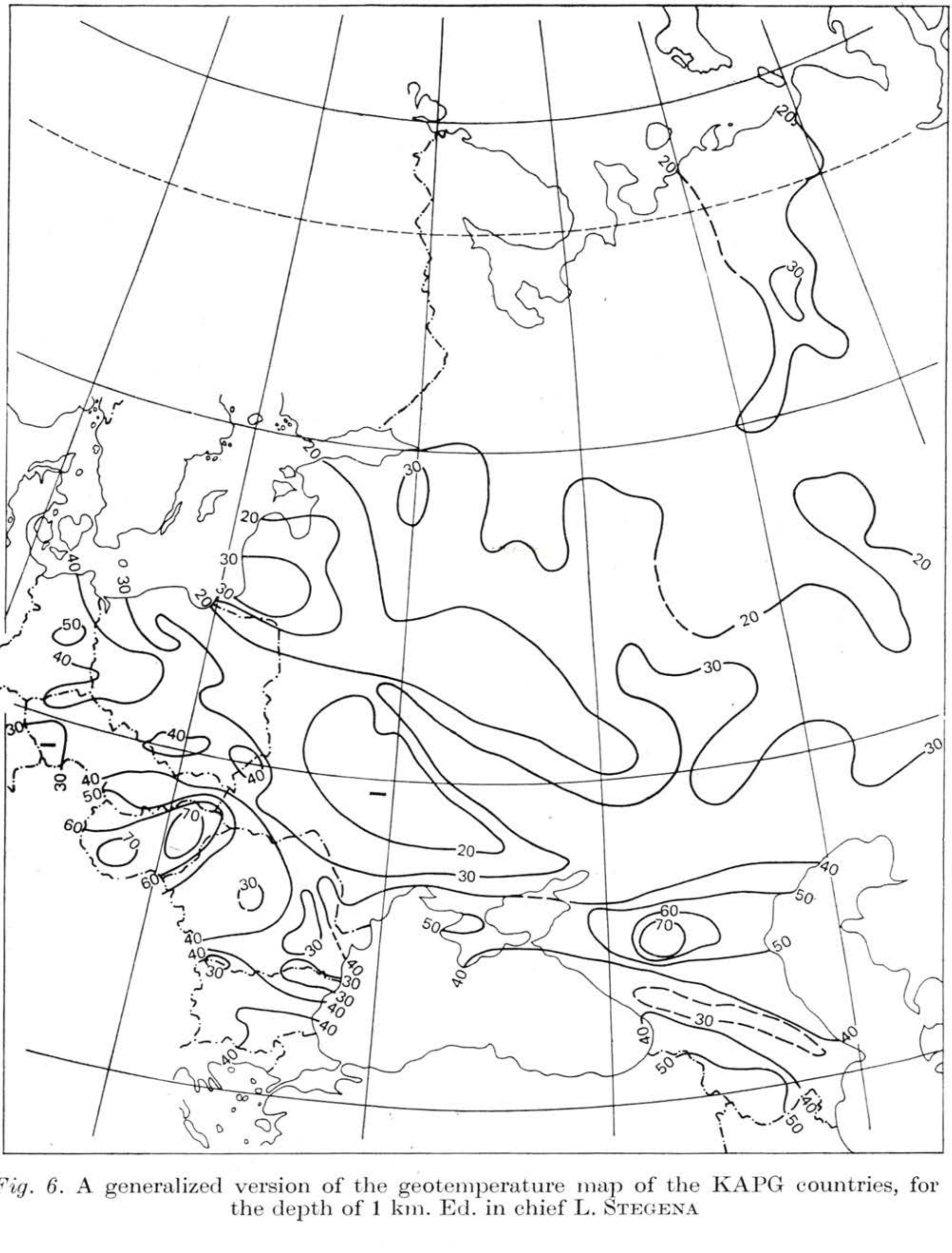


Fig. 6. A generalized version of the geotemperature map of the KAPG countries, for the depth of 1 km. Ed. in chief L. STEGENA

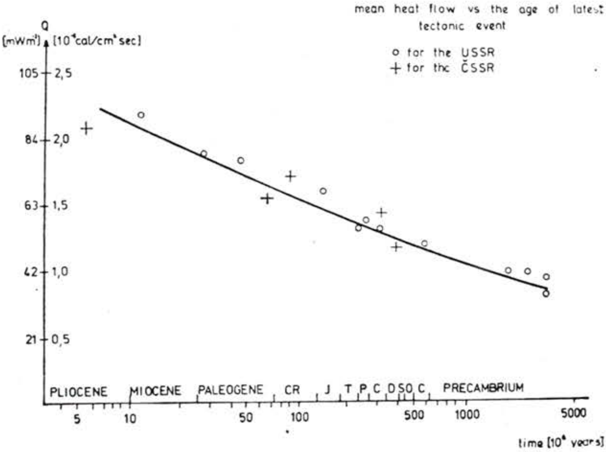


Fig. 7. Mean heat flow versus the age of latest essential tectonic event. Dots: for the European part of the USSR (after KUTAS, LUBIMOVA and SMIRNOV, 1975). Crosses: for Czechoslovakia (after ČERMÁK, 1975)

(Central Slovakia, Lesser Caucasus). — There is a general agreement between heat flow distribution of Fig. 8 and the results obtained by other authors correlating heat flow and geological features (LEE and UYEDA, 1965; LUBIMOVA and POLYAK, 1969).

Geothermics of the Alpine orogenic areas

The geothermal maps on Fig. 3 and Fig. 6 show clearly that—generally speaking—the areas affected by Alpine tectonics, marked on Fig. 9, coincide with a belt of geothermal high.

Recent crustal movements. Fig. 9 shows a generalized version of the map of recent vertical crustal movements in Central and Eastern Europe (MESCHERIKOV, 1972). Uplifting areas essentially coincide with the belt of geothermal high.

Seismicity. Fig. 10, a generalized version of the seismotectonic map (BELOUSOV, SORSKY and BUNE, 1966) shows that the northern boundary of earthquake activity also coincides with the northern border of the geothermally active zone (as far as the southern part, geothermal data are missing).

A detailed study of comparison between geothermics and seismicity is indicated on Fig. 11, for Hungary. Having a lot of geotemperature data, it was possible to construct a map of horizontal variation of geotemperatures. On the other hand, the seismicity of the Pannonian basin consists of sporadic crustal-quakes, with magnitudes less than 5.8. Having worked up all the earthquakes for the last century (1859—1958) (CSOMOR, 1974), the total seismic energy for each knot of a $10' \times 15'$ grid could be determined. Comparison of horizontal geothermal gradients and seismic energies showed that 94% of the seismic energy was released in that part of the country where the horizontal geothermal gradients are stronger than 1.3 °C/10 km.

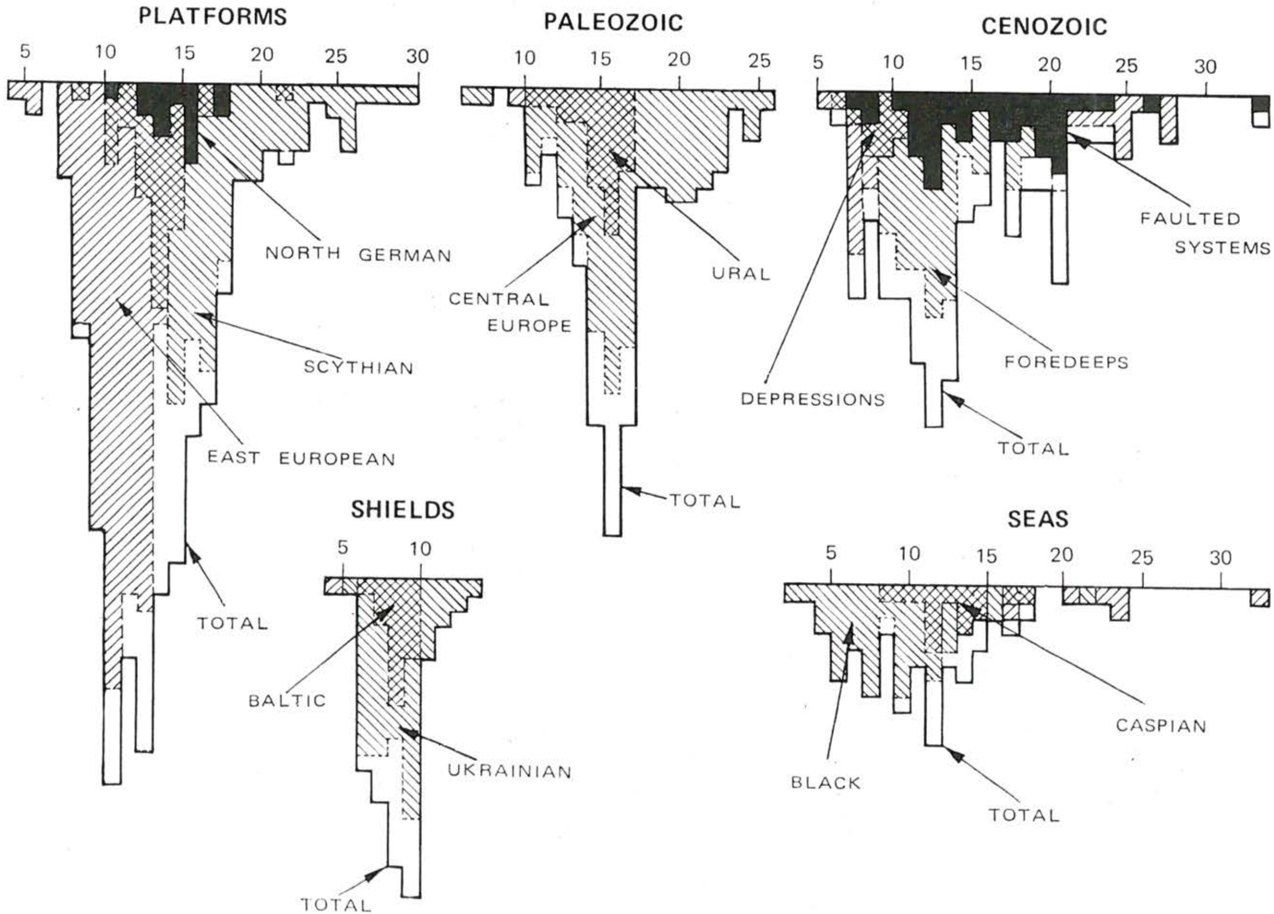


Fig. 8. Histograms of heat flow values from Central and Eastern Europe. Heat flow in $\mu cal/cm^2$ sec

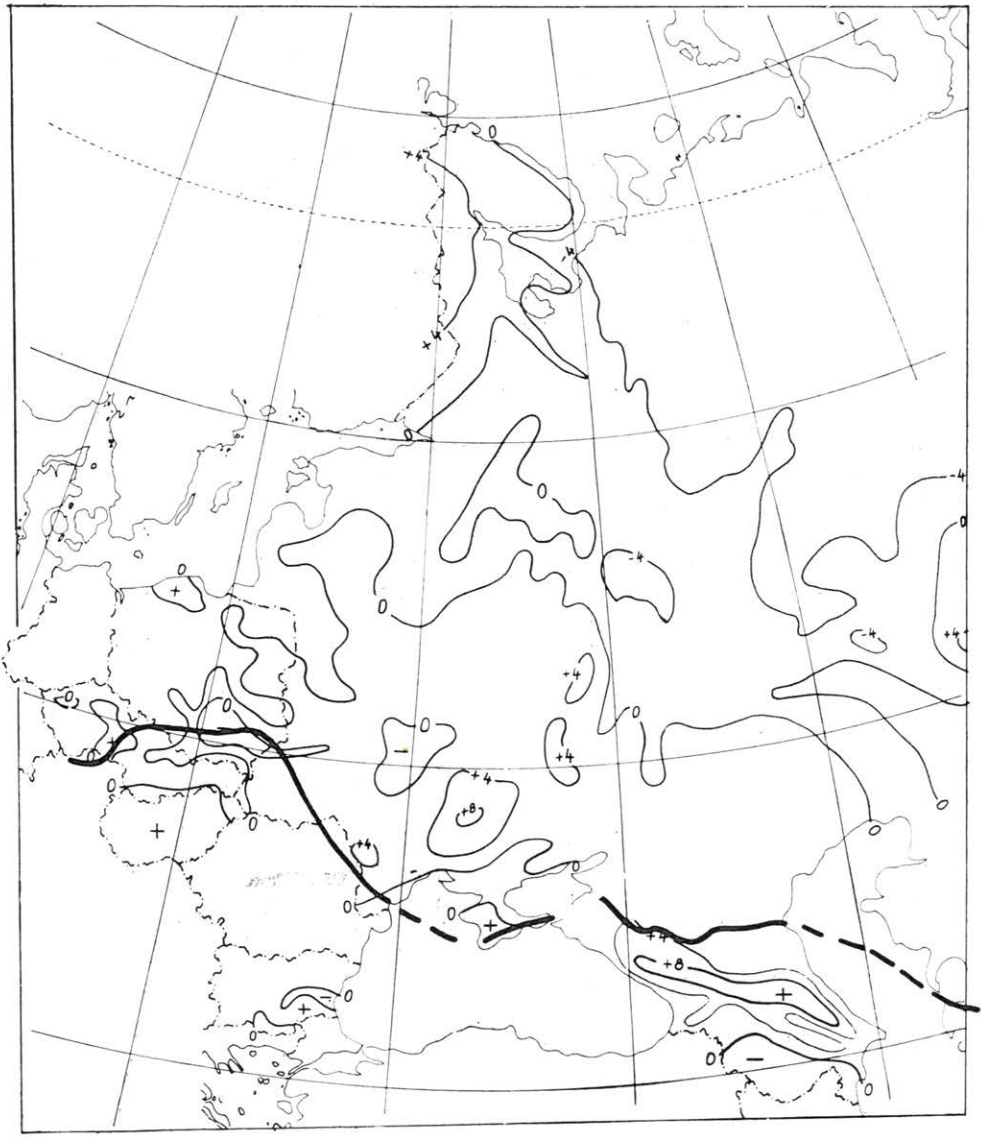
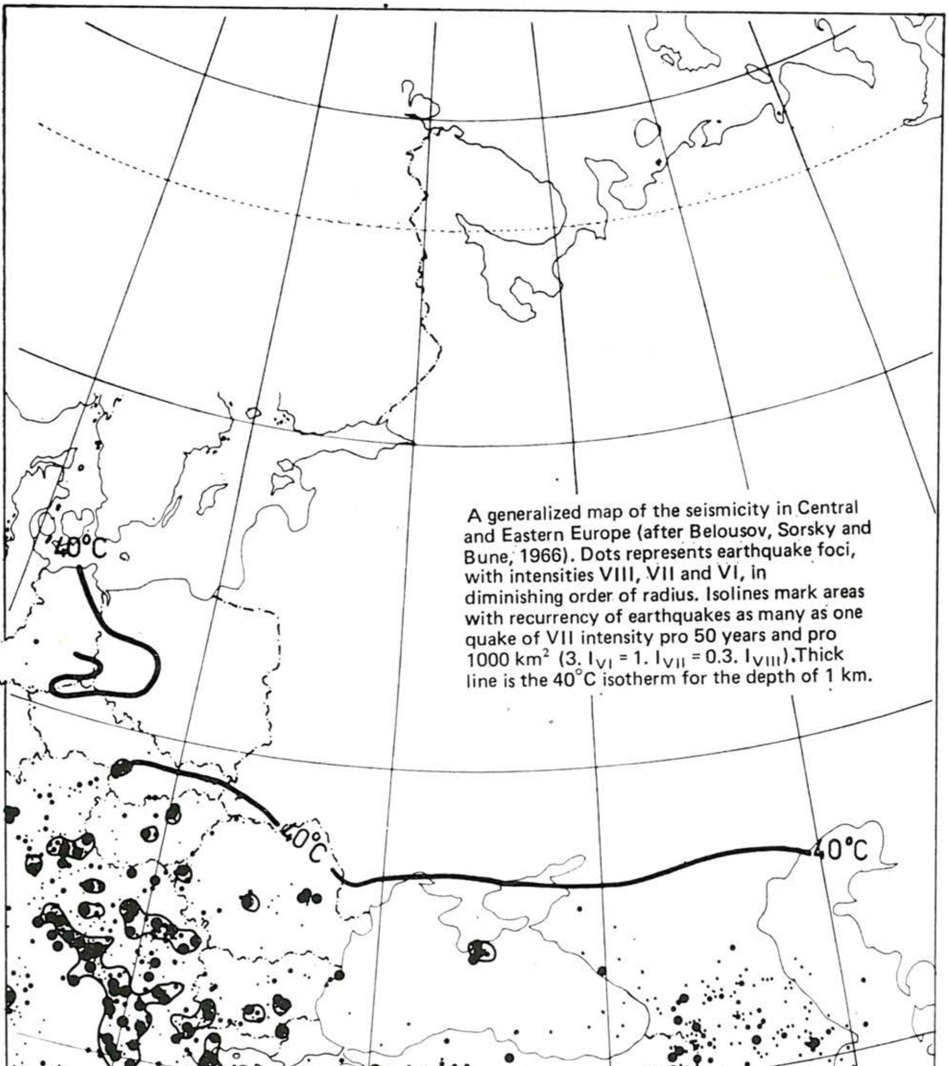


Fig. 9. A generalized version of the map of recent vertical crustal movements (after MESCHERIKOV, 1972) with the northern boundary (heavy line) of the areas affected by Alpine orogenesis. Uplifts (+) and subsidences (−) in mm/year



A generalized map of the seismicity in Central and Eastern Europe (after Belousov, Sorsky and Bune, 1966). Dots represents earthquake foci, with intensities VIII, VII and VI, in diminishing order of radius. Isolines mark areas with recurrency of earthquakes as many as one quake of VII intensity pro 50 years and pro 1000 km^2 ($3.\ I_{VI} = 1.\ I_{VII} = 0.3.\ I_{VIII}$). Thick line is the 40°C isotherm for the depth of 1 km.

Fig. 10

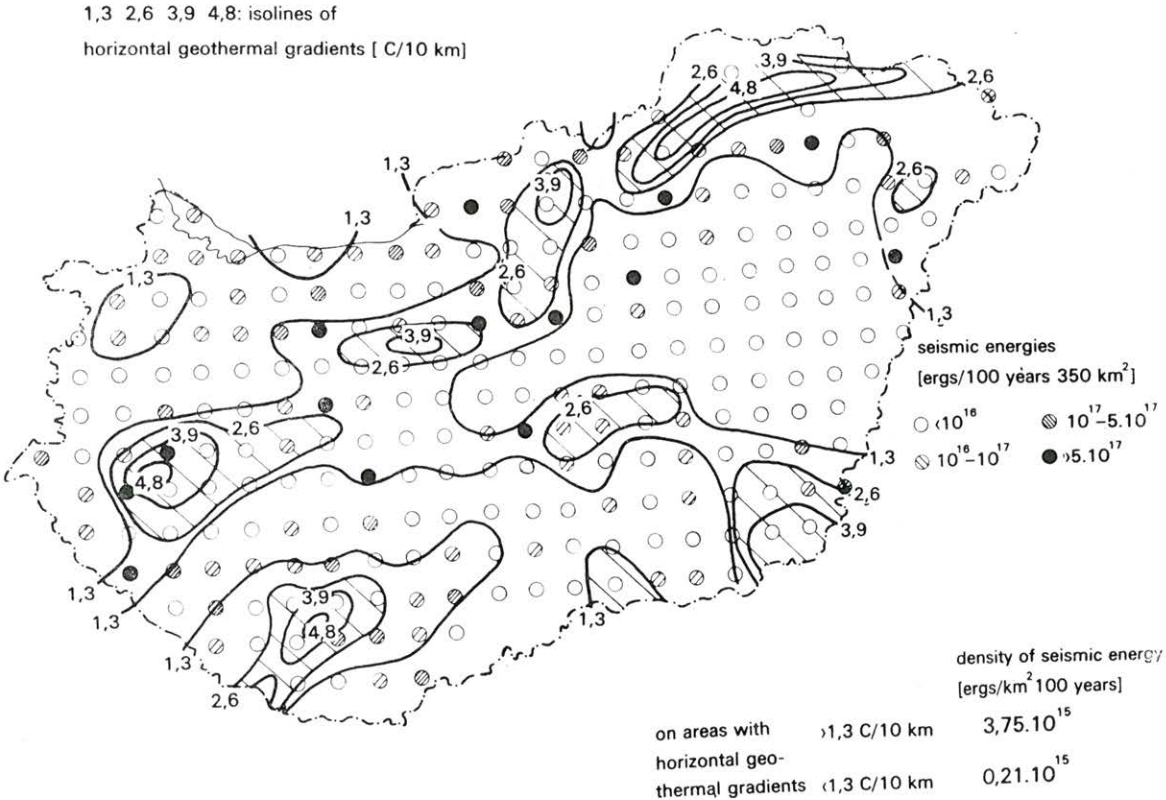


Fig. 11. Horizontal geothermal gradient and seismicity in Hungary. Isolines are in °C/10 km. Dots represent the total seismic energy observed during the last century (1859—1958), for a grid of 10′×15′. Seismicity after CSOMOR (1974)

It is not possible to prove the general validity of this correlation between crustal seismicity and horizontal geothermal gradient, having not anywhere a network of geotemperatures as dense as in Hungary. Nevertheless, many experiments have been published recently intending to delimit geothermally active underground zones by means of areas of increased seismicity (WARD, 1972; DOUZE and SORRELS, 1972; LYER and HITCHCOCK, 1974).

Explanatory remarks. Generally, on the areas affected by Alpine tectonics, geothermal high, crustal uplift and seismic activity appear. It is difficult to decide which of the three phenomena is of primary character. As it is probable, the geothermal high is developed in the upper mantle, too. This relates to the primary character of the geothermal high developed by means of deep currents. At present, crustal uplift and seismicity are consequences of the geothermal anomaly in the Central and Eastern European part of the Alpine orogenesis.

Geothermics and tectogenesis of the Pannonian interarc basin

Fig. 12. shows in detail that the Pannonian basin coincides with a big geothermal high. The Pannonian basin is the warmest part of the Alpine geothermal high belt.

Magnetotelluric soundings (Fig. 13) and anomalous high heat flows calculated for the upper mantle (Fig. 14a – b) show clearly that beneath the Pannonian basin the geothermal high is developed down to great depth. This extra heat could not be the result of heat conduction only, because of the young age (about 10 m.y.) of the basin, but convectionally transported heat is to be supposed as well (STEGENA, GÉCZY and HORVÁTH, 1975). The upward

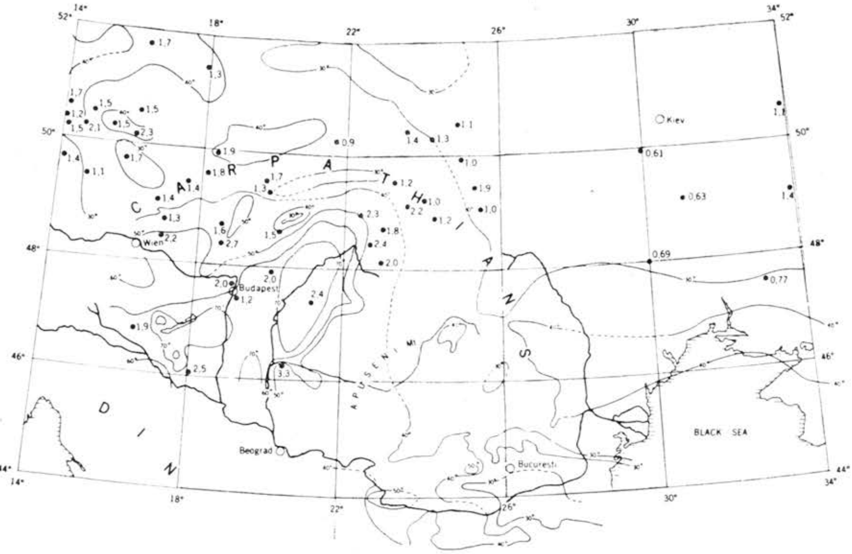


Fig. 12. Geoisotherms (°C) in the depth of 1 km and several typical surface heat flow values in the surroundings of the Pannonian basin. After STEGENA, GÉCZY and HORVÁTH (1975)

moving mantle material, "active mantle diapir", was generated by the subduction associated with the formation of the Carpathians. The evidences and consequences of the mantle diapir are: strong Mio–Pliocene volcanism, geothermal high, anomalistic upper mantle and finally, the thinning out of the crust and isostatic subsidence of the area (Fig. 15).

The study of other intermountain (interarc) basins suggests that the above hypothesis based on KARIG's (1971) tectogenetic model can generally be applied (HORVÁTH, STEGENA and GÉCZY, 1975).

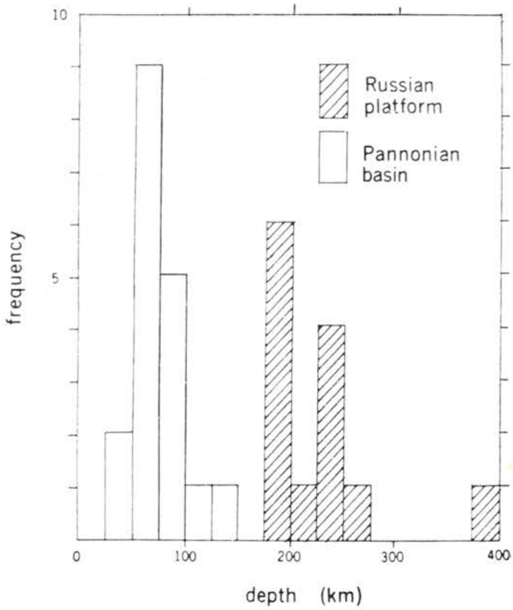


Fig. 13. Histogram of the depth of the highly conducting layer (HCL) for the Pannonian basin and the Russian platform, after magnetotelluric sounding data

General comments

The density of heat flow energy is very small, about 50 pro mill of that of the solar energy. A direct utilization of heat flow is illusory. Sources of geothermal energy to be considered are usually connected with upward moving material, heating up the near-surface rocks.

Recently, LUBIMOVA, LJUBOSHIZ and NIKITINA (1975) gave a correct mathematical solution for the calculation of the thermal field of upward moving materials. Applying this method to the mid-oceanic ridges (Fig. 16), the theoretical results agree with the experimental measurements (SCLATER, 1972). Besides, this combined interpretation showed that the heat loss from all ridges is much higher than believed before (7–17%); it is estimated to 30 % of the total heat loss of the Earth; this latter equals $7.7 \cdot 10^{12} \pm 10\%$ cal/sec.

The sediments of the Pannonian basin, heated up by the upward migrating mantle diapir, contain thermal energy as much as enough for the energy supply of the world for one thousand years. This is a speculation only, the total exploitation being impossible. In practice, the hot thermal waters of the sedimentary layers are exploited fairly systematically in Hungary, used for the heating of settlements, of glasshouses in the agriculture, for thermal baths and finally—if the water-chemism allows—for irrigation. This kind of use of geothermal energy is very practical; the quantity of this geothermal energy, however, is not considerable in the Hungarian energy household. Superheated steams connected with volcanic activity have not been encountered till now. The third possibility of using geothermal energy would be the utilisation of hot dry areas, splitting the rocks by hydropressuring (BLACKWELL, 1973). In the Pannonian basin there are some geological possibilities for this method, because the active mantle diapir brought about an enormous Neogene volcanism, partially covered by sediments. Even a fourth possibility,

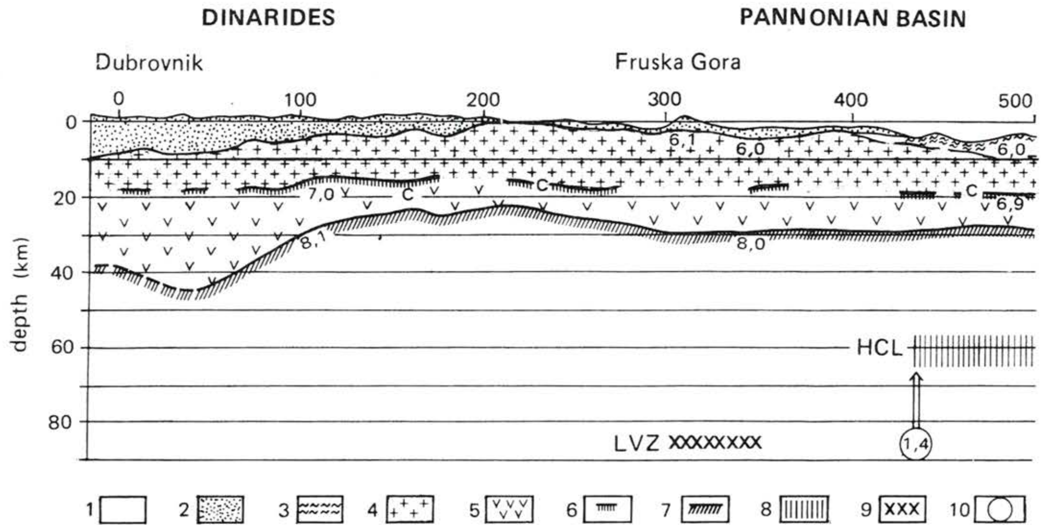


Fig. 14a Seismic crustal profile, heat flow values calculated for the upper mantle (BUNTEBARTH, 1975), position of the HCL and of the LVZ, across the Pannonian basin

1 = young sediments; *2* = sedimentary complex; *3* = Mesozoic basement; *4* = granitic layer, *5* = basaltic layer; *6* = Conrad discontinuity; *7* = Moho discontinuity; *8* = highly conducting layer; *9* = low velocity zone; *10* = heat flows in the upper mantle, in HFU (after STEGENA et al., 1975)

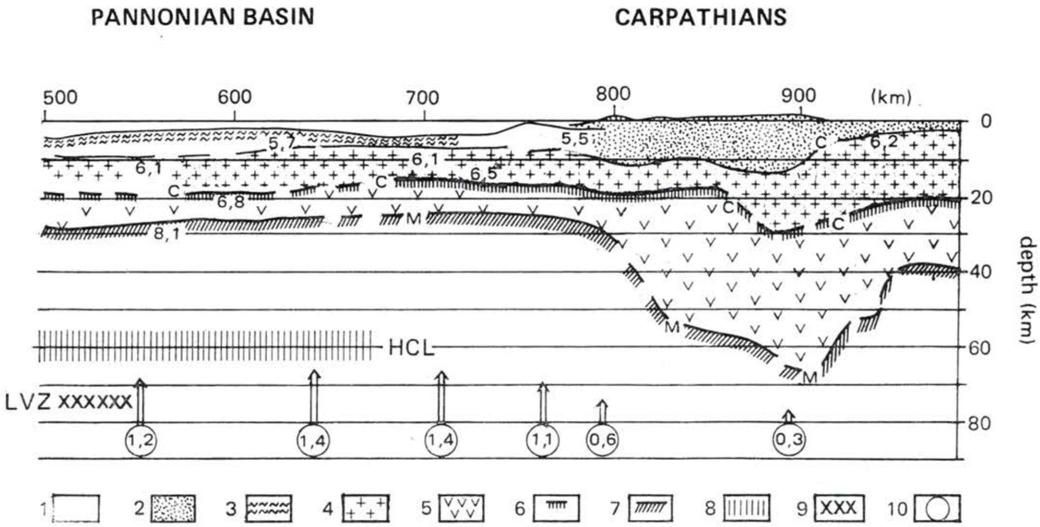


Fig. 14b For explanation see Fig. 14a

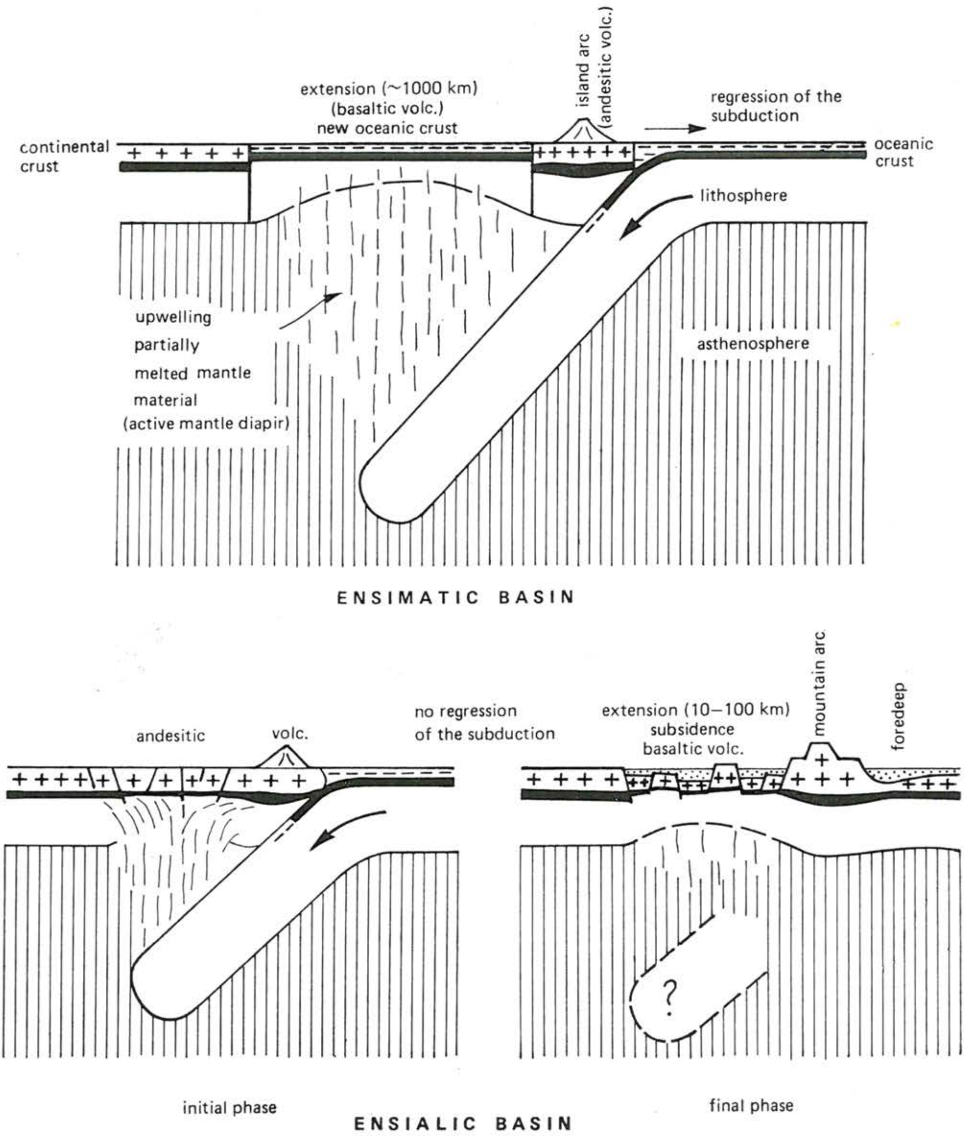


Fig. 15. Plate tectonical evolution model of the ensimatic and ensialic basins, as proposed by HORVÁTH, STEGENA and GÉCZY (1975)

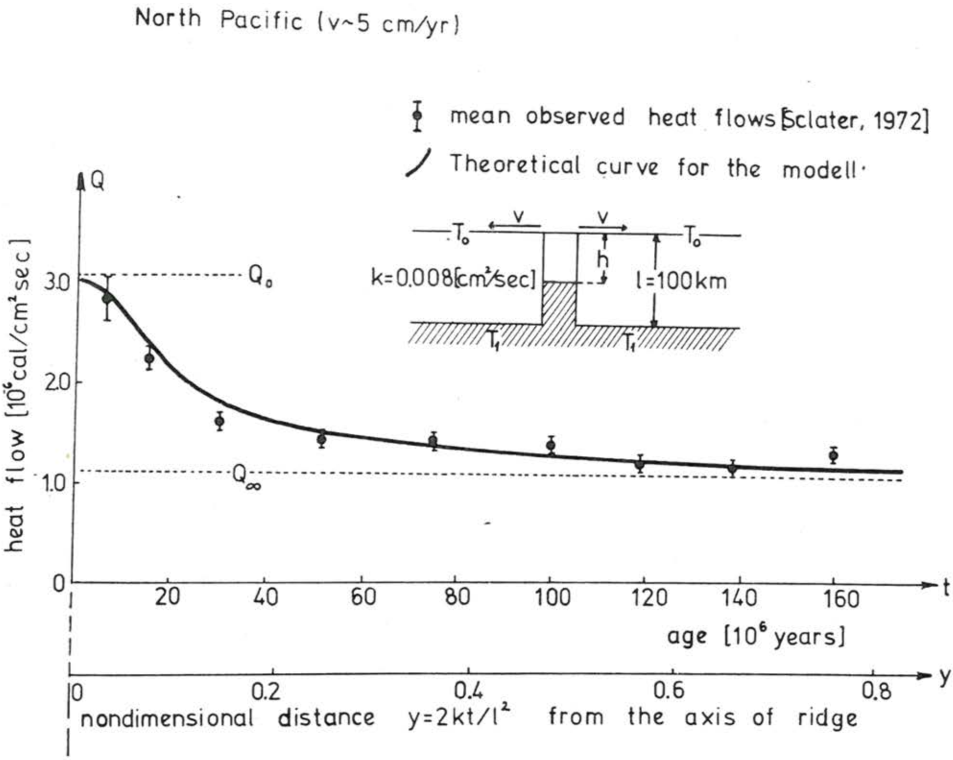


Fig. 16. Observed mean heat flows (after SCLATER, 1972) versus age of the ocean floor (or distance from the mid oceanic ridge). Solid line: theoretical curve of LUBIMOVA et al., 1975

the rather phantastic idea of setting up a magma-power station is not a totally utopian one in the Pannonian basin. Magnetotelluric soundings find the high conductivity layer generally in elevated position in the upper mantle beneath Hungary (ÁDÁM, 1965). At certain places (ÁDÁM, 1973), the high conducting layer approaches the surface up to a depth of 6–8 km.

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PREPARATION OF MAPS OF REGIONAL HYDROGEOLOGICAL UNITS

RÉDACTION DE CARTES HYDROGÉOLOGIQUES D'UNITÉS RÉGIONALES DISTINCTES

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RÉSUMÉ

Un groupe d'agents de l'Entreprise Nationale d'Exploration et Forage Géologiques a établi une carte hydrogéologique, au 1: 100 000[e], présentant une vue d'ensemble d'un territoire de 8000 km^2 et comprenant de nombreuses unités géomorphologiques et hydrogéologiques différentes. Le présent travail de caractère expérimental a été fait aux fins d'orientation rapide, comme un guide informatif envisagé pour les agents de gestion supérieurs.

Le mode de représentation graphique se base sur les caractères gîtologiques des nappes aquifères respectives.

Les quatre séries aquifères principales de la région ont été spécialement distinguées sur la carte. Les voici:

1. *Aquifère d'eau karstiques* — réservoir comprenant des nappes aquifères carbonatées du Paléo-Mésozoïque, avec des lentilles de calcaire éocènes y superposées. Utilisable pour l'implantation d'une usine hydraulique de distribution d'eau d'une capacité de 10 à 100 milles m^3 par jour, pour l'implantation de puits productifs d'eau de bain de l'ordre 1000 m^3 par jour.

2. *Aquifère d'eaux de fissures* — zones tectonisées de grès oligocènes et de roches éruptives miocènes. Utilisable pour l'implantation d'une usine hydraulique locale ayant un rendement journalier de 100 à 1000 m^3, pour la production d'eaux de bain, resp. minérales.

3. *Aquifère d'eaux interstitielles artésiennes* — complexe du Miocène moyen et du Pannonien supérieur. Permet d'alimenter une usine hydraulique d'un rendement journalier de 100 à 1000 m^3, les niveaux plus profonds sont utilisables pour le captage d'eaux de bain.

4. *Aquifère d'eaux phréatiques*, resp. *interstitielles artésiennes* — complexe de gravier et de sables grossiers fluviogènes pléistocènes dans les vallées des fleuves et dans les dépressions marginales de la Grande Plaine Hongroise. Permet d'alimenter des usines hydrauliques pour la distribution de 10 à 100 milles m^3 d'eaux potables et industrielles par jour.

On envisage de présenter à la Conférence les feuilles de carte de documentation, la collection de données fondamentales et les notices explicatives ronéotypées, déposées à l'Institut Géologique de Hongrie, ainsi que la feuille de carte combinée, rédigée par les agents de l'Entreprise Nationale de Recherches et Forage Géologiques et présentant les données de documentation et celles d'interprétation ensemble. Dans les transactions de la Conférence, on présentera des détails, en noir et blanc, des cartes en question, en échelle réduite par rapport à l'original.

Under the direction of the Central Office of Geology in Hungary, layout maps of regional units showing the various mineral deposits of Hungary on the scale of 1:100,000, are being prepared.

The first map of a regional hydrogeological unit has been made for showing the very diversified NE part of the country (Fig. 1). This diversity is constituted by the co-existence of the following features and facts: karstic mountains, volcanic rock masses, hilly landscapes constituted by Upper Tertiary sediments, river valleys incised in a Pleistocene gravel sequence, marginal depressions of the Great Hungarian Plain. Moreover an other motive was the high need for water to be consumed by the swiftly developing industry and the settlements expanding at an unprecedented high rate.

The plotting of maps of regional units have been aimed to serve as a handy, informative guide to be used mainly by state executives of regional and industrial planning. Therefore the following water-bearing sequences, most significant for water recovery purposes (Fig. 2) and representing, at the same time, the sequences of distinct geological time units, have been distinguished on the maps.

1. The geological substratum of the territory under consideration is the *Paleo-Mesozoic* basement. Its limestone and dolomite masses are considerable karst water aquifers. From a hydrogeological viewpoint, the *Eocene* limestone lenses and basal detrital materials overlying locally the aforementioned sequence should be assigned to the basement, too.

The outcrops of the basement, represented by the Bükk and Aggtelek Mountains, are *cold* karst water aquifers very important from the viewpoint of potable water recovery. The buried limestone and dolomite masses of 100 m

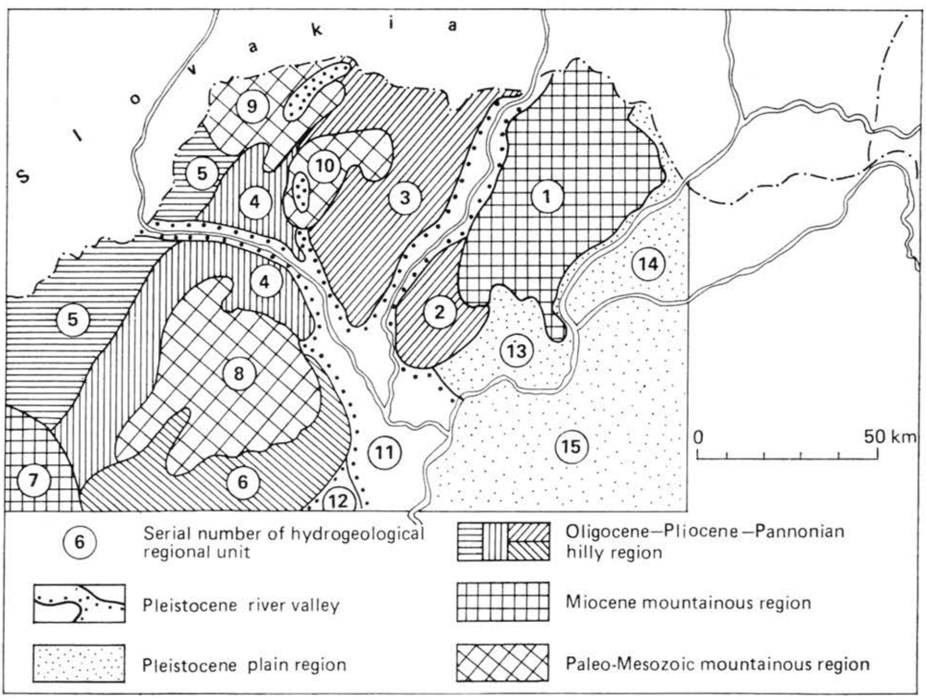


Fig. 1. Layout map of the hydrogeological regional units of Borsod

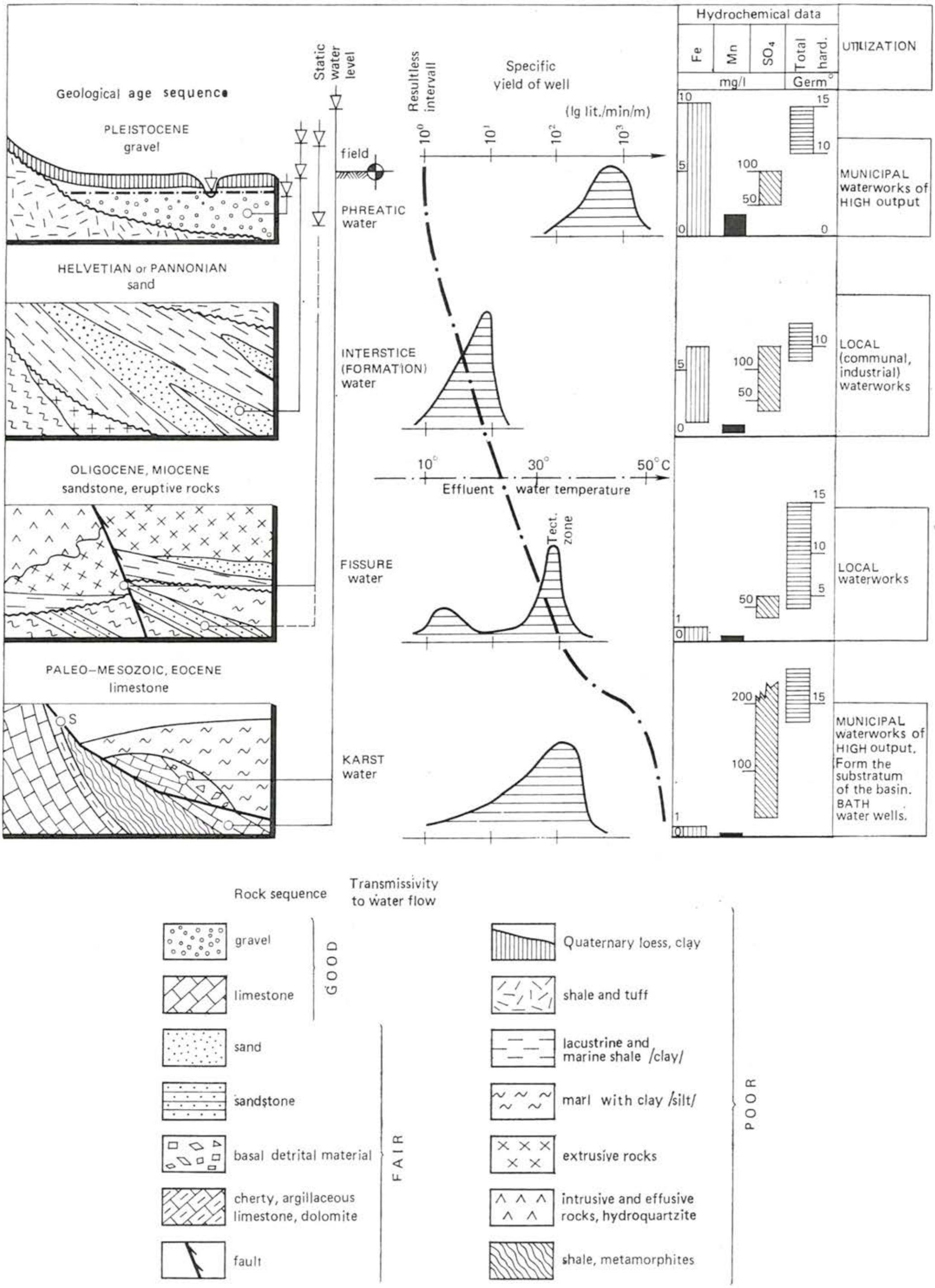


Fig. 2. Idealized columnar section of the major water-bearing sequences

depth of the sequence are thermal aquifers (the reciprocal value of the virtual geothermal gradient at the margins is about 10 metres per degree Centigrade).

2. Along fracture lines, the *Oligocene sandstone* and *Miocene* tuff and lava formations are significant thermal and subthermal (mineral, medicinal) aquifers holding fissure water.

3. The interstice (formation) water aquifers, occurring far away from the karst and river valley areas of high water-yielding capacity, are of local importance. In the northern basins the aquifers involved are *Helvetian* sands, and in the central and southern parts (close to the Great Hungarian Plain), they are *Pannonian* sands.

4. Very important for urban and industrial settlements is the *Pleistocene* gravel sequence of the rivers Sajó and Hernád, containing ground-water (phreatic water) down to a depth to which meteoric and river water soak; farther to the south, at the basal horizons of the sequence varying from 50 to 150 m in thickness, interstice (formation) water is held. At the margin of the Great Hungarian Plain this aquifer of good water-yielding capacity locally becomes sandy.

The marks and symbols used on the map have been determined by the four aquifers listed above.

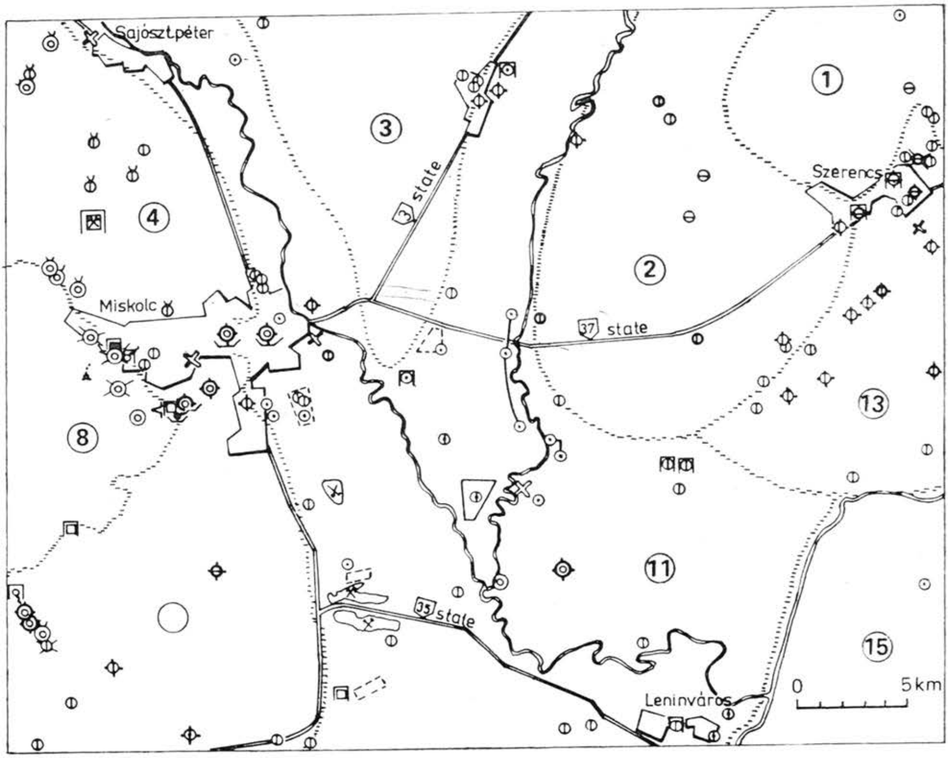
The maps and their supplements are specified as follows:

1. Documentation map

The documentation map contains data on various boreholes yielding *valuable information* for an all-round collecting of data. Let us list the various sources of information recorded on the map:

a) Drilled wells, hydrogeological exploratory boreholes, other boreholes furnishing hydrogeological information (e.g. coal exploratory boreholes, geological survey boreholes, etc.). After reviewing the data files of about 1000 drilled wells and of additional 2000 boreholes for different purposes, 270 drilling data have been recorded on the map.
b) Water-producing or exploratory shafts and adits.
c) Springs having a yield of more than 500 m^3 per day on the average.
d) Active water-transmitting cave galleries.
e) Lakes, storage basins, mine waste water ponds.
f) Major waterworks supplied with subsurface water.
g) Spas.
h) Places at which ground-water is polluted by sewage of industrial and municipal origin (refuse deposits, slag heaps).
i) Places where mine waste waters are lifted to the surface, with indication of an eventual utilization.
j) Areas of water exploration, successful or resultless.

Waters colder than 20 °C are marked in blue, those of higher temperature, in red. Among the symbols of boreholes and wells, four different types of wells corresponding to the 4 aquifers listed above, namely phreatic ground-water, interstice (formation) water, fissure water and karst water aquifers, have been distinguished. The symbols have been chosen as be able to represent



Water temperature

< 20C° > 20C° Boreholes

- Phreatic ground water (Quaternary)
- Interstice (formation) water (Neogene)
- Fissure water (Eocene–Miocene)
- Karstic water of the basement (Paleo–Mesozoic)
- Well with effluent water
- Point of water recovery and exploitation
- Without hydrogeological data
- Water producing shaft or well
- Spring with a yield of 500 m^3/day or more
- Water producing or exploratory adit
- Water producing series of wells (gallery)
- Hydrogeologically important cave or underground water course
- Water works
- Bath
- Water prospecting area
- Mine drainage water lifting point
- 11 Boundary and serial number of hydrogeological regional unit
- Mining lake
- Important pollution of ground water

Fig. 3. Documentation map

boreholes that have penetrated two or more horizons or two or more different types of aquifer. Both the wells yielding flowing water and the unsuccessful drills have been indicated on the map.

In Fig. 3 an extract from the documentation map, reduced in scale as compared to the original, is shown with black-line symbols of the above-listed grouping, instead of the original marks in colour. The scale of this map allows to indicate on it the boundaries of the 15 regional hydrogeological units.

2. *Collection of fundamental data*

The collection of fundamental data contains, in form of tabulations under a proper book cover, the most essential data on hydrogeological recording points shown on the afore-mentioned map, the symbols being the same as those used on the map, i.e. both letters and numerals. As being adequately prepared and fastened into a hard cover, these data collections have enable a quick retrieval of the data being looked for.

3. *Interpretation map*

This map in colours showing the whole surface of the surveyed area, is named interpretation map that contains the results and conclusions deduced from the fundamental data and other documents (on subareas). In the part-maps reduced in size shown in Fig. 4a and 4b the colour symbols have been replaced by black-and-white ones, the colour patches showing the transmissivity to water flow by hachure.

The four major water-bearing sequences distinguished have been represented, according to the legends of Fig. 4a and 4b, in the following way:

a) Paleo-Mesozoic basement and Eocene limestone lenses atop:

- their surface is indicated by black contour lines referred to sea level and paced at 0.5 km,
- their general water-bearing capacity (good, fair, poor, unknown) is given by graphic marking, at the edges of the various aquifer belts,
- in outcrops of karsted basement rocks, the transmissivity is already marked by colouring, in harmony with the colour symbols given in the following paragraph.

b) The various water-bearing basin sequences overlying the Paleozoic to Mesozoic and Tertiary basement, have been indicated by colour symbols (good, fair, poor aquifers) and sequences of uncertainly-known water-bearing capacity by hachure. Two or more water-bearing sequences in vertical succession are cross-hachured.

c) The contour lines of the Tertiary interstice (formation) water-bearing sequences represent a vertical spacing of 200 m (or of 100 m in some places), and are referred to the sea level. In areas of near-footwall interstice (formation) water aquifers with a temperature lower than 20 °C, the contour lines are blue; farther away in direction of the

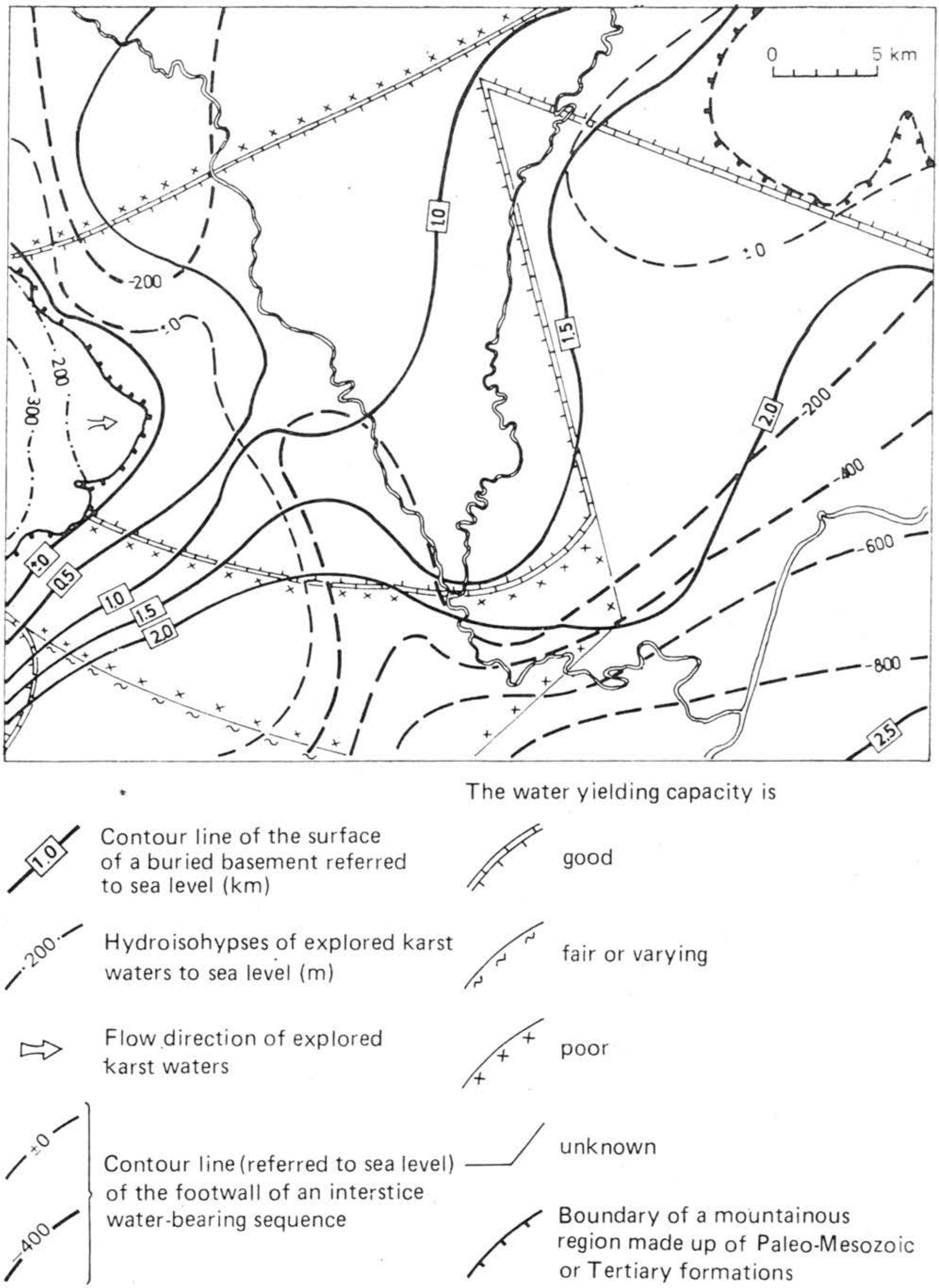


Fig. 4a Detail from the interpretation map

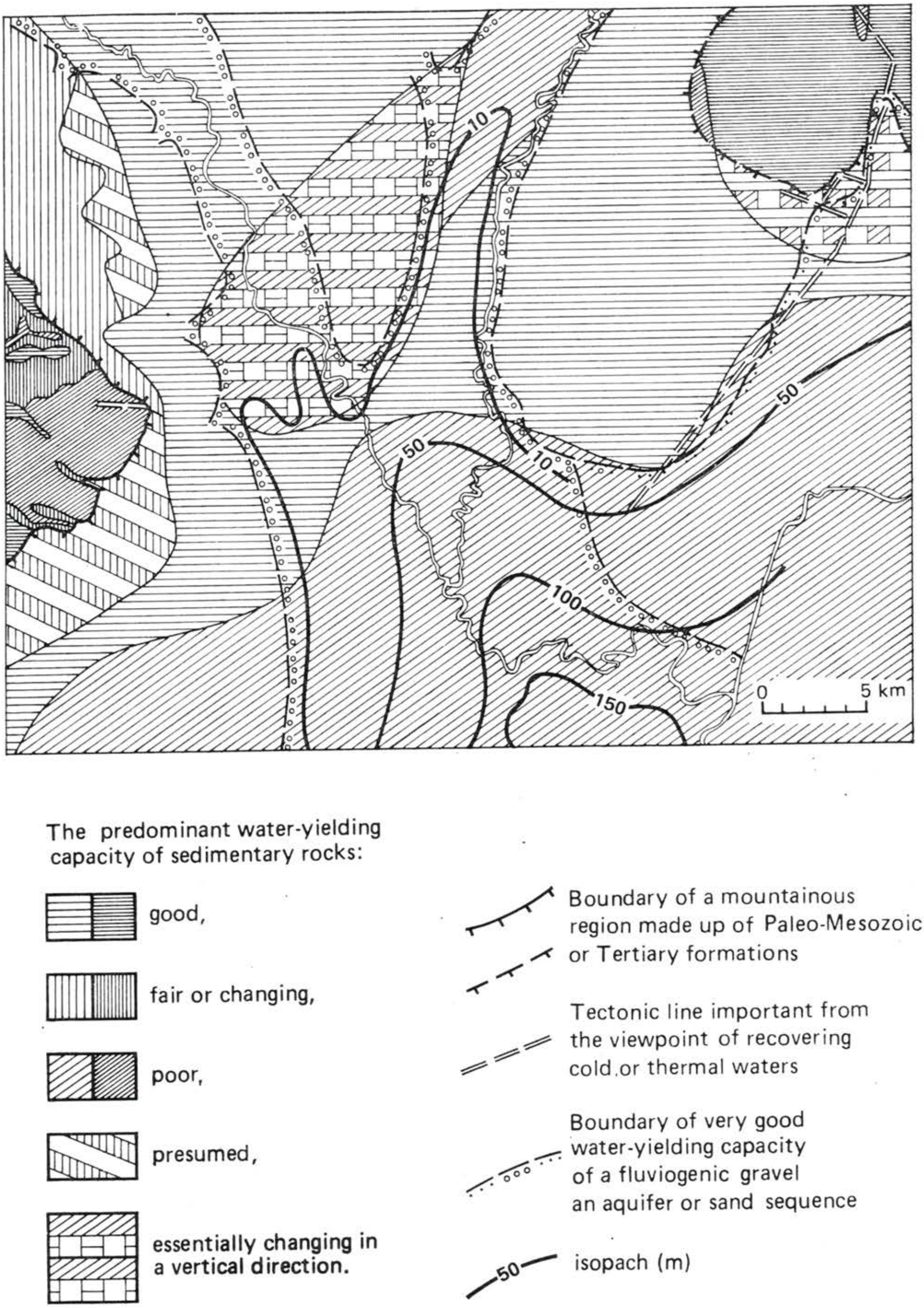


Fig. 4b Detail from the interpretation map

Great Hungarian Plain, where the temperature of the formation waters is already higher than 20 °C in the same aquifers, the contour lines are red on the original map.

d) The boundaries of the Pleistocene, fluviogenetic gravel sequence and its 10-, 50-, 100- and 150-m isopachs (where known!) are originally shown in orange.

Beside the four major water-bearing sequences, the following informations are recorded on the interpretation map:

a) Boundaries of mountainous regions of different age are traced because the mountains are independent hydrogeological units. (Another important hydrogeological boundary, the margin of river valleys, coincides with the outline of the Pleistocene gravel sequence.)

b) Reconnoitred fracture lines of hydrogeological significance as bearing/transmitting cold and thermal waters, are also shown.

Relations between the most significant hydrogeological formations and the legend are presented by an idealized profile in Fig. 5.

4. Explanatory Notes

As requested by the Central Office of Geology, Explanatory Notes for a better understanding of the maps have been prepared so as to include a short text and 68 figures (illustrations).

Contents of the descriptions

a) Principles for the preparation of the two map variants with Explanatory Notes to them.

b) A summarizing commentary on the documentation and interpretation maps.

c) A summarizing explanation to the 15 hydrogeological regional units distinguished within the quadrangles under consideration.

Types of figures

The colours showing transmissivity in the figures are the same as those used on the maps, but the symbols are substituted by hachure here. In the Explanatory Notes the following types of figures, grouped according to regional units, have been used:

a) Idealized hydrogeological columnar section for each particular regional unit.

b) Hydrogeological profiles in which, as essential, a hydrogeological evaluation of the respective sequences of chronostratigraphic units is given. Separately, at a larger scale, the static water levels in wells, i.e. boreholes as well as the temperature of effluent water and specific water yield data are indicated.

c) Local, hydrogeological sections on a larger scale.

d) Local, more detailed hydrogeological maps showing typical problems of single regional units.

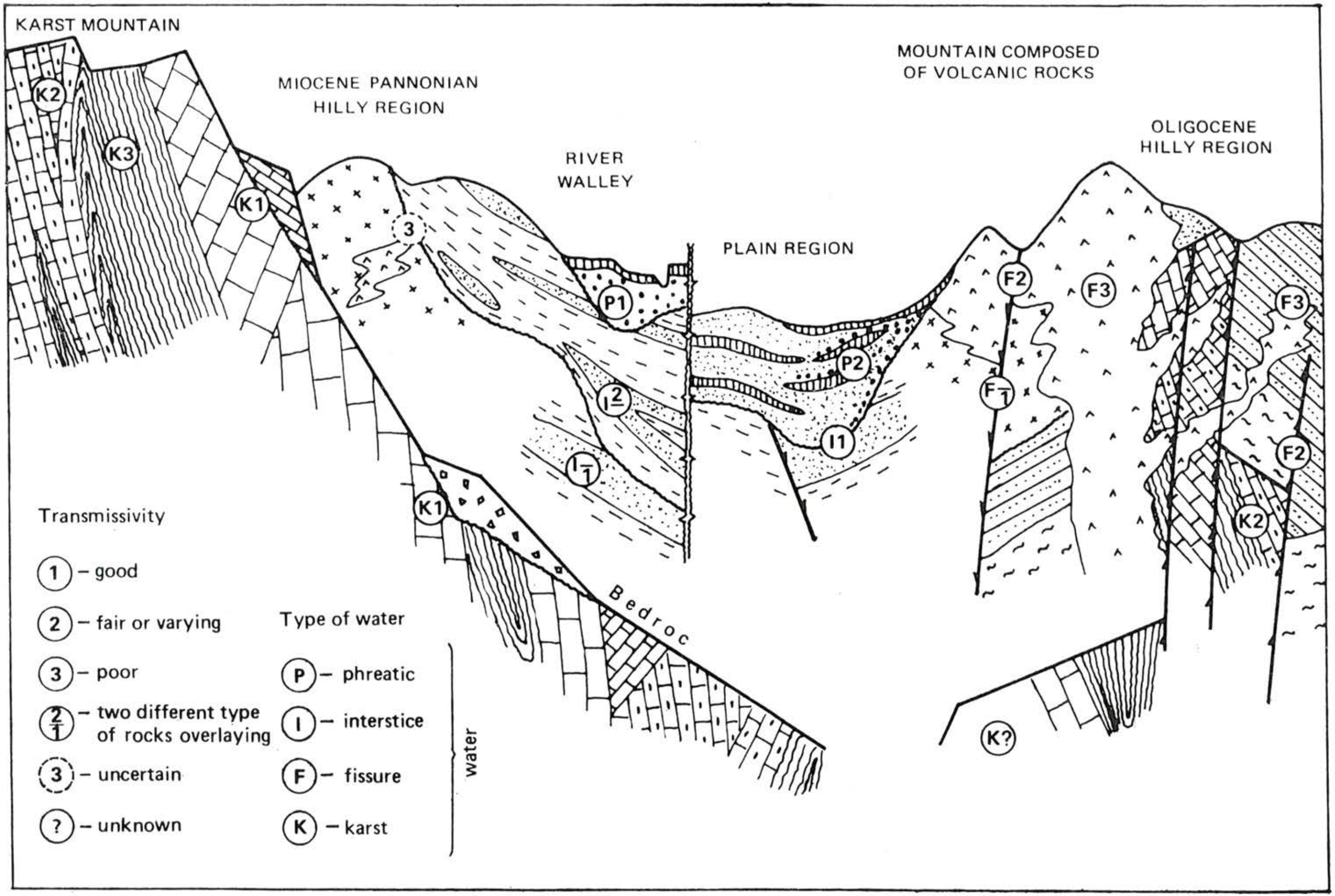


Fig. 5. Idealized section illustrating the representation of transmissivity to water flow

e) Detailed stratigraphic column and other data from hydrogeological boreholes yielding data on relevant features of each particular regional unit.
f) Characteristic water quality diagrams for each regional unit.
g) Block diagrams.

Synoptic figures enabling the reader to have an overlook on the whole material. Thus, the "$Q-H$" curves of wells and shafts tapping the typical aquifers within the above-listed four major water-bearing sequences are presented in one figure, plotted on an identical scale.

All illustrations—with symbols and marks corresponding to each particular type—can be found on a single outline map. The figure captions are listed in tabulation, according to both regional units and types, as given in the Explanatory Notes.

GEOCHEMICAL REGULARITIES IN THE DISTRIBUTION AND CONCENTRATION OF ALKALINE ELEMENTS OF THE CARBONATE WATERS IN GEORGIA, U.S.S.R.

RÉGULARITÉS GÉOCHIMIQUES DE RÉPARTITION ET DE CONCENTRATION D'ÉLÉMENTS ALCALINS RARES DANS LES EAUX CARBONIQUES DE LA GEORGIE

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RÉSUMÉ

Les eaux carboniques des dépôts schisteux du Jurassique du flanc méridional du Grand Caucase sont les plus enrichies en éléments alcalins rares. On peut constater que les éléments alcalins rares ont une tendance générale de s'accumuler avec l'accroissement de la minéralisation générale, suivie d'une modification de la composition chimique en séquence suivante:

$$HCO_3—Ca \rightarrow HCO_3—Na \rightarrow HCO_3—Cl—Na \rightarrow Cl—HCO_3—Na$$

Les eaux carboniques ayant la haute teneur en éléments alcalins rares sont caractérisées par des valeurs basses des rapports de:

$$Na/K; K/Li; K/Rb; K/Cs; Rb/Cs.$$

La série de migration des éléments alcalins rares, contenus dans ces eaux, a l'aspect suivant:

$$Li \rightarrow Cs \rightarrow Na \rightarrow Rb \rightarrow K.$$

L'accroissement de la teneur en éléments alcalins rares dans l'eau carbonique s'effectue principalement par suite de la désalcalinisation des roches, accompagnée par la superposition des processus endogènes.

Carbonate waters of Georgia with rare alkaline elements should be regarded among the most interesting occurrences of carbonate waters in the Alpine folded system. In Georgia carbonate waters are mainly confined to folded systems of the Great and Minor Caucasus and to a part of the Artivino – Bolnisi block of Minor Caucasus. In the above regions, which are mainly built up of eruptive rocks and sedimentary deposits of the Jurassic, Cretaceous and Paleogene, carbonate waters of soda and brine-alkaline composition occur.

Representative contents and relative concentrations of the rare alkaline elements in primary carbonate water regions of Georgia are listed in Table 1. As it is shown by this table, contents of rare alkaline elements in carbonate waters are not uniform. The highest contents are in carbonate waters of the folded system of the Great Caucasus Southern Slope. Here such an index as the coefficient of water migration related to the rare alkaline elements, is very high. Thus, the migration order of alkaline elements differs from that known for carbonate waters,

$$Na > Li > Cs > K > Rb.$$

Relative concentrations of rare alkaline

Geotectonic zones, water-bearing complexes	Lithium		
	content mg/l	coefficient of	
		concentration	water migration
Folded system of the Great Caucasus Southern Slope			
1. Schist	$1.89-29.0$	$0.05-0.9$	$14.28-218.9$
2. Carbonate fumarole	$1.3-3.3$	$0.04-0.1$	$6.6-16.8$
3. Terrigenous and carbonate fumarole	$5.26-20.0$	$0.164-0.62$	$15.79-60.0$
Minor Caucasus			
Carbonate and volcano-fumarole deposits	1.02	0.031	5.77
Artivino – Bolnisi Block			
Carbonate deposits	0.603	0.018	2.08

On the whole, the migration order of alkaline elements seems to be quite original here. It should especially be noted that in approaching zones with active tectonics (the Major Thrust Zone of the Great Caucasus Southern Slope and the axial part of the Adjara-Trialety system) lithium and caesium of carbonate water forestall sodium and potassium. Thus, here the migration order is as follows:

$$\mathrm{Li} > \mathrm{Cs} > \mathrm{Na} > \mathrm{K} > \mathrm{Rb}.$$

With the distance growing from the Major Thrust, within the range of carbonate deposits of the Great Caucasus Southern Slope, the rare alkaline elements content (particularly that of the caesium) decreases in waters stored by the aquiferous complex of Upper Jurassic to Lower Cretaceous age, showing the following order of migration:

$$\mathrm{Li} > \mathrm{Na} > \mathrm{Cs} > \mathrm{K} > \mathrm{Rb}.$$

As to the carbonate waters of the so-called epicontinental regions, such as the Artivino—Bolnisi Block, the migration order is of a normal pattern. Such an unusual nature of distribution outlined above, is caused by local conditions, under which rare alkaline elements in carbonate waters forestall sodium and potassium, in relation with their accumulation and geochemical mobility.

As a divergency from the usual chemical composition of carbonate waters, one may observe a trend of the rare alkaline elements to increase in concentration parallel with the increase of the total mineralization. The chemical composition is changing in a sequence as follows:

$$\mathrm{HCO_3{-}Ca} \rightarrow \mathrm{HCO_3{-}Na} \rightarrow \mathrm{HCO_3{-}Cl{-}Na} \rightarrow \mathrm{Cl{-}HCO_3{-}Na}.$$

The observed zoning in carbonate waters of the Caucasus is rather peculiar and is well defined in both vertical and horizontal profiles. Fig. 1 and Table 2 indicate that rare alkaline elements content increases with the increasing

Table 1

elements in carbonate waters of Georgia

Rubidium			Caesium		
content mg/l	coefficient of		content mg/l	coefficient of	
	concentration	water migration		concentration	water migration
0.14 – 1.2	0.009 – 0.037	0.13 – 1.06	0.26 – **3.3**	0.07 – 0.9	12.6 – 160.7
0.043 – 0.1	0.003 – 0.06	0.046 – 1.08	0.066 – 0.14	0.018 – 0.038	2.9 – 6.15
0.139 – 0.46	0.0092 – 0.031	0.089 – 0.29	0.07 – **0.38**	0.019 – 0.102	1.82 – 9.86
0.024	0.00016	0.028	0.015	0.004	**0.73**
0.23	0.0015	0.169	0.019	0.005	0.567

concentration of Cl^-, Na^+, K^+, HCO_3^--ions, while accumulation of caesium progresses much quicker than that of rubidium and even lithium.

Correlation Tables 3 and 4, presented below, confirm that mentioned above. It is readily seen that rare alkaline elements are related to each other, to Cl^-, $(Na^+ + K^+)$, HCO_3^--ions and to water mineralization (M), bands being strong and positive. Furthermore, rare alkaline elements are strongly correlated with boron, especially in carbonate waters of the Great Caucasus Southern Slope, this in its turn favouring their possible paragenesis. Besides that, the correlation of rare alkaline elements with potassium and sodium (individually) in carbonate waters has shown to exist a correlation band of the formers to be much stronger with potassium than with sodium.

The analysis of particular coefficients of correlation permitted to determine supplementary peculiarities of rare alkaline elements' hydrochemistry (Tables 5 and 6). Thus the relation between rare alkaline elements is so strong that even after neglecting the role in mineralization of stroncium, boron hydrocarbonate, sodium, potassium, and even that of the

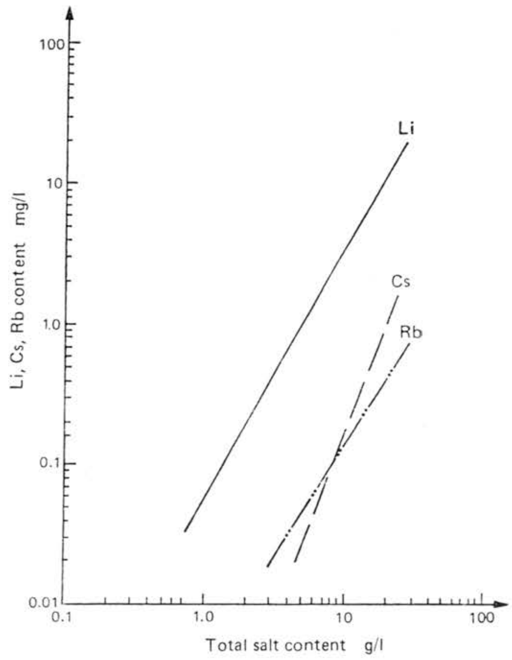

Fig. 1. Dependence of Li, Rb and Cs contents on the mineralization of carbonate waters of Georgia

Table 2

Rare alkaline elements distribution in carbonate waters of Georgia

Goetectonic zone	Chemical structure	Mineralization	Lithium mg/l	Rubidium mg/l	Caesium mg/l
Great Caucasus Southern Slope	HCO_3-Ca	$\frac{0.8-3.7}{2.12}$	$\frac{0.003-1.08}{0.45}$	$\frac{0-0.07}{0.045}$	$\frac{0-0.1}{0.03}$
	HCO_3-Na	$\frac{1.1-22.3}{7.15}$	$\frac{0.06-8.1}{2.22}$	$\frac{0.03-0.34}{0.125}$	$\frac{0-0.38}{0.115}$
	$HCO_3-Cl-Na$	$\frac{2.0-26.8}{9.92}$	$\frac{0.7-19.0}{5.38}$	$\frac{0.03-1.2}{0.138}$	$\frac{0-3.3}{0.46}$
	$Cl-HCO_3-Na$	$\frac{3.1-24.7}{11.16}$	$\frac{0.9-29.0}{8.49}$	$\frac{0.03-1.1}{0.383}$	$\frac{0-3.0}{0.896}$
Adjara-Trialety folded system	HCO_3-Na	$\frac{1.5-7.2}{6.1}$	$\frac{0.4-1.66}{1.08}$	$\frac{0.03-0.05}{0.0375}$	$\frac{0-0.035}{0.0203}$
	$HCO_3-Cl-Na$	$\frac{3.7-11.8}{6.6}$	$\frac{0.06-1.5}{0.606}$	0.03	0.01
	$Cl-Na$	10.7	1.5		
Artivino-Bolnisi Block	$HCO_3-Cl-Na$	$\frac{6.6-13.9}{9.05}$	$\frac{0.2-0.8}{0.603}$	$\frac{0.002-0.8}{0.23}$	$\frac{0-0.05}{0.019}$

rare alkaline elements, it frequently remains too strong and positive. Also, the relation between lithium and chlorine in carbonate waters both of the Great Caucasus Southern Slope and of the Adjara-Trialety folded system remains significant, even if the role of primary chemical compositional ions and that of the mineralization are neglected. Then, in carbonate waters of Adjara – Trialety there is a marked correlation of rubidium and caesium with potassium (Table 6).

In some cases the neglecting of a single component strengthens the correlation between the others. Thus, the mutual relationship will be stronger between caesium and lithium when neglecting the role of potassium and chlorine. It is similarly observed between lithium and chlorine. Likewise, the correlation gets to be closer between lithium and chlorine if one neglects the potassium and sulphate ions (Table 6). It is important to note the appearance of a strong negative band between lithium and hydrocarbonate if one excludes the role of potassium, chlorine and magnesium (Table 6). Besides that, in many cases one may observe a weakening in the correlation bond between hydrocarbonate and rare alkaline elements, when other components are not taken into account. This bond in carbonate waters indicates the essential role of heterogeneous processes, favouring the migration in solution of these components.

A strong bond of rare alkaline elements with each other and with other ions of the carbonate waters' chemical composition is also supported by regression equations quoted below (Table 7). These equations present a valid

Table 3

Correlation bands of chemical elements in carbonate waters of the Great Caucasus Southern Slope

	pH	M	Na+K	Ca	Mg	Cl	SO_4	HCO_3	Rb	Cs	Sr	Li	B
M	+0.60												
Na+K	+0.56	+0.76											
Ca	−0.15	+0.05	−0.09										
Mg	+0.34	+0.56	+0.48	−0.21									
Cl	+0.48	+0.69	+0.77	+0.06	+0.42								
SO_4	−0.10	−0.29	−0.24	−0.11	−0.12	−0.24							
HCO_3	+0.56	+0.86	+0.76	+0.81	+0.54	+0.62	−0.36						
Rb	+0.38	+0.42	+0.46	−0.11	+0.18	+0.50	−0.22	+0.44					
Cs	+0.32	+0.39	+0.40	−0.07	+0.21	+0.28	−0.25	+0.52	+0.58				
Sr	+0.10	+0.25	+0.14	+0.15	+0.06	+0.22	−0.36	+0.21	+0.21	+0.03			
Li	+0.43	+0.70	+0.74	+0.03	+0.44	+0.78	−0.33	+0.70	+0.62	+0.43	+0.24		
B	+0.44	+0.66	+0.74	+0.06	+0.43	+0.68	−0.36	+0.72	+0.52	+0.47	+0.06	+0.73	

Table 4

Correlation bands of chemical elements in carbonate waters of the Adjara-Trialety folded system

	pH	M	Na+K	Ca	Mg	Cl	SO_4	HCO_3	Rb	Cs	I	Br	B	Li
M	+0.66													
Na+K	+0.76	+0.86												
Ca	−0.03	+0.13	−0.03											
Mg	+0.07	+0.08	−0.18	−0.21										
Cl	+0.75	+0.84	+0.80	−0.02	−0.10									
SO_4	−0.29	+0.06	−0.20	−0.06	+0.51	+0.10								
HCO_3	+0.63	+0.94	+0.77	−0.16	+0.14	+0.71	−0.02							
Rb	+0.65	+0.65	+0.80	−0.04	−0.34	+0.69	−0.54	+0.58						
Cs	+0.68	+0.65	+0.86	−0.05	−0.38	+0.63	−0.57	+0.59	+0.89					
I	−0.17	−0.03	+0.14	−0.11	−0.44	−0.14	−0.60	+0.001	+0.01	+0.05				
Br	+0.38	+0.14	+0.25	−0.33	+0.32	+0.22	+0.21	+0.12	+0.09	+0.04	+0.12			
B	+0.55	+0.45	+0.48	−0.12	+0.40	+0.55	+0.17	+0.39	+0.27	+0.22	−0.10	+0.71		
Li	+0.66	+0.72	+0.89	−0.05	−0.30	+0.81	−0.27	+0.55	+0.87	+0.85	+0.04	+0.26	+0.47	

Table 5

Particular correlation bands of rare alkaline elements in carbonate waters of the Great Caucasus Southern Slope

1. $\tau_{RbCs} = 0.58$;	$\tau_{RbCs.Sr} = 0.58$;	$\tau_{RbCs.SrLi} = 0.45$;	$\tau_{RbCs.SrLiB} = 0.43$
2. $\tau_{RbLi} = 0.62$;	$\tau_{RbLi.Sr} = 0.60$;	$\tau_{RbLi.SrCs} = 0.47$;	$\tau_{RbLi.SrCsB} = 0.41$
3. $\tau_{RbB} = 0.52$;	$\tau_{RbB.Sr} = 0.52$;	$\tau_{RbB.SrCs} = 0.34$;	$\tau_{RbB.SrCsLi} = 0.13$
4. $\tau_{CsLi} = 0.43$;	$\tau_{CsLi.B} = 0.14$;	$\tau_{CsLi.BSr} = 0.15$;	$\tau_{CsLi.BSrRb} = 0.02$
5. $\tau_{CsB} = 0.47$;	$\tau_{CsB.Li} = 0.25$;	$\tau_{CsB.LiSr} = 0.24$;	$\tau_{CsB.LiSrRb} = 0.20$
6. $\tau_{LiCl} = 0.78$;	$\tau_{LiCl.M} = 0.57$;	$\tau_{LiCl.MHCO_3} = 0.58$;	$\tau_{LiCl.MHCO_3(Na+K)} = 0.47$
7. $\tau_{Li(Na+K)} = 0.74$;	$\tau_{Li(Na+K).M} = 0.45$;	$\tau_{Li(Na+K).MHCO_3} = 0.39$;	$\tau_{Li(Na+K).MHCO_3Cl} = 0.13$
8. $\tau_{LiM} = 0.70$;	$\tau_{LiM.Cl} = 0.35$;	$\tau_{LiM.ClHCO_3} = 0.04$;	$\tau_{LiM.ClHCO_3(Na+K)} = 0.02$
9. $\tau_{LiHCO_3} = 0.70$;	$\tau_{LiHCO_3.(Na+K)} = 0.32$;	$\tau_{LiHCO_3.(Na+K)Cl} = 0.27$;	$\tau_{LiHCO_3.(Na+K)ClB} = 0.22$

Table 6

Particular correlation bands of rare alkaline elements in carbonate water of the Adjara—Trialety folded system

1. $\tau_{RbLi} = 0.87$;	$\tau_{RbLi.HCO_3} = 0.81$;	$\tau_{RbLi.HCO_3(Na+K)} = 0.60$;	$\tau_{RbLi.HCO_3(Na+K)Cl} = 0.62$
2. $\tau_{RbCs} = 0.89$;	$\tau_{RbCs.HCO_3} = 0.83$;	$\tau_{RbCs.HCO_3(Na+K)} = 0.66$;	$\tau_{RbCs.HCO_3(Na+K)Cl} = 0.65$
3. $\tau_{RbHCO_3} = 0.58$;	$\tau_{RbHCO_3.SO_4} = 0.67$;	$\tau_{RbHCO_3.SO_4(Na+K)} = 0.06$;	$\tau_{RbHCO_3.SO_4(Na+K)Cl} = 0.002$
4. $\tau_{Rb(Na+K)} = 0.80$;	$\tau_{Rb(Na+K).SO_4} = 0.83$;	$\tau_{Rb(Na+K).SO_4HCO_3} = 0.67$;	$\tau_{Rb(Na+K).SO_4HCO_3Cl} = 0.62$
5. $\tau_{RbCl} = 0.69$;	$\tau_{RbCl.SO_4} = 0.75$;	$\tau_{RbCl.SO_4(Na+K)} = 0.27$;	$\tau_{RbCl.SO_4(Na+K)HCO_3} = 0.26$
6. $\tau_{CsLi} = 0.85$;	$\tau_{CsLi.HCO_3} = 0.78$;	$\tau_{CsLi.HCO_3(Na+K)} = 0.30$;	$\tau_{CsLi.HCO_3(Na+K)Cl} = 0.46$
7. $\tau_{Cs(Na+K)} = 0.86$;	$\tau_{Cs(Na+K).SO_4} = 0.93$;	$\tau_{Cs(Na+K).SO_4HCO_3} = 0.85$;	$\tau_{Cs(Na+K).SO_4HCO_3Cl} = 0.82$
8. $\tau_{CsHCO_3} = 0.59$;	$\tau_{CsHCO_3.Li} = 0.28$;	$\tau_{CsHCO_3.Li(Na+K)} = 0.06$;	$\tau_{CsHCO_3.Li(Na+K)Cl} = 0.15$
9. $\tau_{LiCl} = 0.81$;	$\tau_{LiCl.(Na+K)} = 0.35$;	$\tau_{LiCl.(Na+K)HCO_3} = 0.55$;	$\tau_{LiCl.(Na+K)HCO_3SO_4} = 0.56$
10. $\tau_{LiHCO_3} = 0.55$;	$\tau_{LiHCO_3.(Na+K)} = -0.46$;	$\tau_{LiHCO_3.(Na+K)Cl} = -0.61$;	$\tau_{LiHCO_3.(Na+K)ClMg} = -0.59$

Table 7

Equations of rare alkaline elements regression in carbonate waters of the Great Caucasus Southern Slope

1. $\ln Rb = 0.19 \ \ln (Na+K) + 0.35 \ Cl - 3.17$; $(\delta = 3.13)$
2. $\ln Rb = 0.75 \ \ln Cs + 0.75 \ \ln Sr - 0.8$
3. $\ln Cs = 0.31 \ \ln (Na+K) - 3.17$
4. $\ln Cs = 0.48 \ \ln HCO_3 + 0.33 \ \ln Rb - 2.44$; $(\delta = 1.02)$
5. $\ln Li = 1.34 \ \ln M - 2.20$
6. $\ln Li = 0.92 \ \ln (Na+K) + 0.56 \ \ln Cl + 0.31$; $(\delta = 2.08)$
7. $\ln Li = 0.77 \ \ln Mg - 0.55 \ \ln SO_4 - 0.32$

quantitative characteristic of ionic bonds. This bond being strong is also confirmed by an insignificant dispersion of the above regression equations.

As far as the spatial distribution of rare alkaline elements in carbonate waters is concerned, we should note that carbonate waters with significant coefficients of water migration and high rare alkaline elements concentration are usually confined to schists of the Great Caucasus Southern Slope (Table 1). As a whole, high clarkes of rare alkaline elements within shales as well as the carbonate waters with increased rare alkaline elements concentrations being confined to shales prove the primary role of lithologic-facies factors in the formation of similar waters.

We would note here that in carbonate waters of the considered region the rare alkaline elements concentration is not in full conformity with the zonality noted above, and one may distinguish some areas with high minor elements content. These anomalies in carbonate waters outlines regions of conjugate structures and linear to transverse deep faults, with high tectonic and geothermal ($q > 3$ microcalories/sm^2 sec.) intensity, mobilizing migration capacity of minor elements and favouring infiltration of endogenetic solutions. As it has been noted above, in this regions rare alkaline elements get accumulated in water more readily than sodium and potassium. As a result, values of their ratios are somewhat decreased: Na/K (about 4); K/Li (about 0.5); K/Rb (as high as 83); K/Cs (about 32); Rb/Cs (about 0.3). All these values are smaller than those of the rocks rich in rare alkaline elements content.

According to many researchers, such an intensive accumulation of rare alkaline elements in carbonate waters is due to the leaching of rocks under high temperatures and pressures. However, as shown by recent data on the activization of rare alkaline elements at intensive manifestations of postmagmatic metasomatism; on the isotope composition of CO_2 in carbonated waters of Georgia, and on the discovery of avogadrite $(K, Cs)BF_4$ in volcanic fumarolic sublimation products, there is a sound reason to suppose that along with the leaching processes endogenetic solutions play a quite definite role in the genesis of rare alkaline elements in carbonate waters.

ASPECTS HYDROGÉOCHIMIQUES DES SOURCES LIMITROPHES DU BASSIN DE SÃO PAULO, BRÉSIL

HYDROGEOCHEMICAL ASPECTS OF SOURCES AROUND THE SÃO PAULO BASIN, BRAZIL

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ABSTRACT

In our plan to study the waters of sources which are sold under the name of bottled "drinking water" of São Paulo State, the first part of our work was to study samples collected within the area of the São Paulo sedimentary basin.

12 sources were studied, within a total of more than 50 known sources in the State of São Paulo, which covers a surface of 225,000 km².

The São Paulo basin has a surface of 1180 km² and consists of Cenozoic sediments of a medium thickness of 200 m, which consists essentially of clays intercalated with sand lenses.

Limiting the basin the rocks are predominantly of pre-Cambrian origin with complex structures: granites, gneisses and micaschists. The fractures, fissures and diaclases of these rocks form favourable structures for outcrops. The waters of these sources have the following physico-chemical characteristics: temperatures are not very different from the average temperature of the air; a low conductivity related to a low concentration of dissolved elements (between 28—98 mg/l); an acid pH (4.7—5.5); radioactivity between 8.95—57.56 maches; major elements with predominance of the alcaline earth elements; a variable concentration of the trace elements.

A study of the exchangeable bases showed that there is a relationship between the chemical composition of the waters and the lithology, and also the existence of rapid circulation of these waters within the rocks.

According to the classification of Schoeller—Berkaloff, all waters are magnesio-bicarbonated and they clearly separate on the Piper diagram according to the rocks in which they circulate.

However, a difference could be observed between the waters of sources which emerge from the crystalline rocks around the basin and those of the sediments: in SO_4^{2-}, B^{3+} and in radioactivity.

Introduction

L'objectif de notre étude consistait à établir la corrélation entre la composition chimique des eaux et la roche magasin. Étant donné le nombre très grand des sources, environ 50 connues, nous nous sommes limités, au cours de la première phase, à l'étude des émergences de la région des environs du bassin sédimentaire de São Paulo. En effet l'État de São Paulo couvre, dans son ensemble, une superficie de 252,000 km² dont pratiquement, les 3/4 sont recouverts par les sédiments appartenant au grand bassin du Paraná. Parmi les eaux étudiées, dont la majorité sont exploitées et vendues dans le commerce, sous l'appellation d'eau de table, nous en avons examiné douze. Les sources

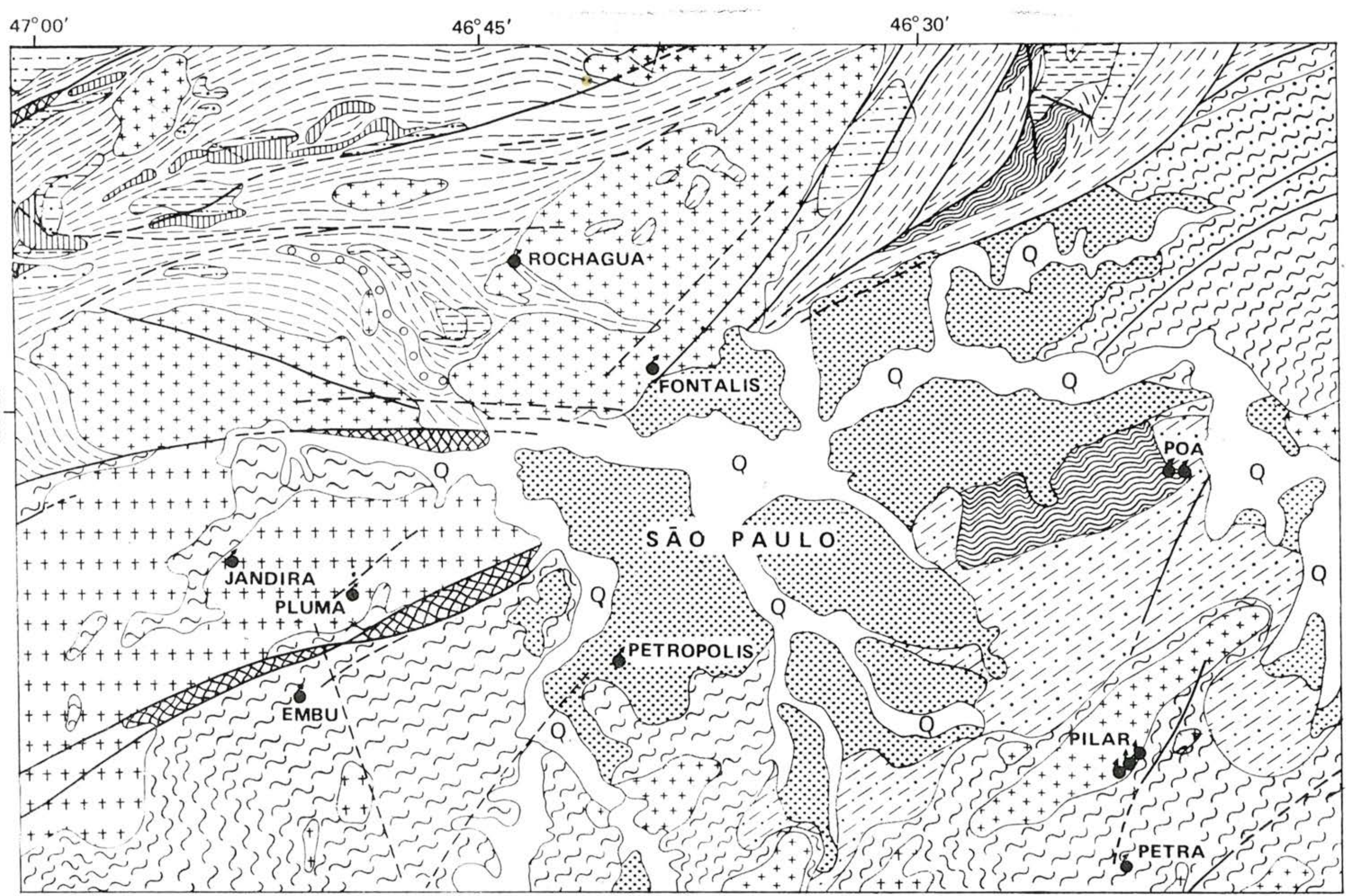

Fig. 1a Carte géologique de Sao Paulo (d'après le C.P.R.M. 1974)

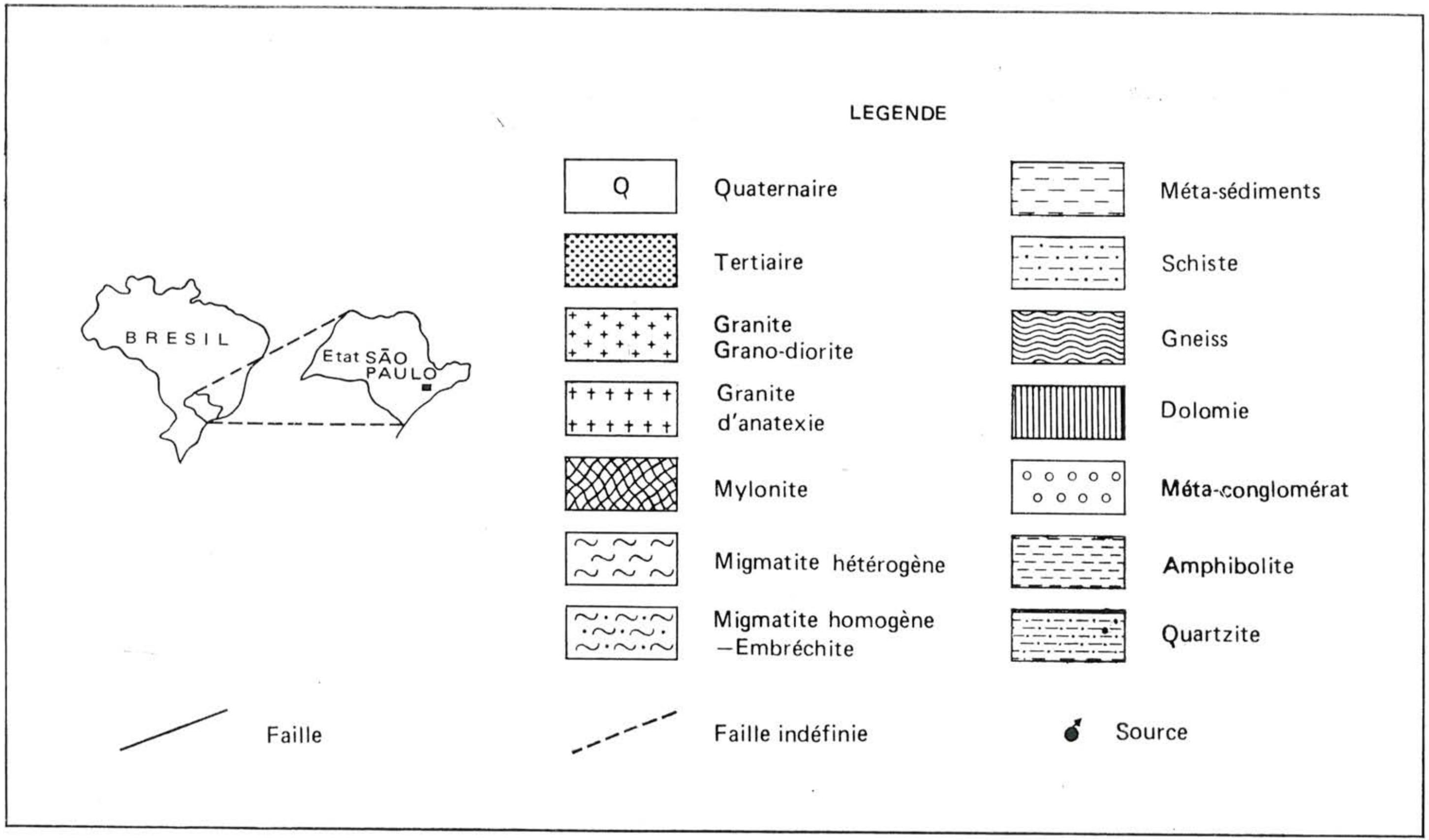
LEGENDE
BRESIL
Etat SÃO PAULO
Q
Quaternaire
Tertiaire
Granite
Grano-diorite
Granite
d'anatexie
Mylonite
Migmatite hétérogène
Migmatite homogène
—Embréchite
Méta-sédiments
Schiste
Gneiss
Dolomie
Méta-conglomérat
Amphibolite
Quartzite
Faille
Faille indéfinie
Source

Fig. 1b

sont situées sur les roches précambriennes dominant le petit bassin tertiaire de São Paulo de 1180 km². Avant de donner un aperçu des résultats chimiques, ainsi que la corrélation qui a pu être établie, il convient de faire un court rappel sur la géologie complexe de chaque émergence. Soulignons ici, que les analyses chimiques des éléments majeurs ont été réalisées sur le terrain, et les traces au laboratoire, avec le spectromètre portatif « HACH Water Analysis Kit ».

Cadre géologique

Comme nous l'avons déjà signalé, les sources se localisent sur les roches précambriennes qui appartiennent au Groupe São Roque et au Complexe Cristallin. La lithologie est essentiellement représentée par des migmatites, gneiss, micaschistes et granites ayant développé des métamorphismes de contact ou général (Fig. 1). Étant donnée la résistance de ces roches, en général, la surgie des eaux se fait à partir de fissures et de diaclases ouvertes selon la schistosité, leur permettant une circulation rapide ou bien au contact de roches différentes. Une description géologique sommaire de chaque source s'impose donc pour aboutir à une corrélation entre les différentes eaux. Les descriptions minérales des roches proviennent en partie des relevés cartographiques de Coutinho (1972), du C.P.R.M. (1974) et d'autre part d'une étude macroscopique des échantillons recueillis sur le terrain.

— Jandira: La géologie est représentée par un granite à deux micas, où l'eau sort par l'intermédiaire des diaclases qui recoupent la roche.

— Pluma: Fonte dos Bandeirantes : Il s'agit d'une source de contact entre deux types de granites ; d'une part un granite porphyroïde à phénocristaux de feldspath potassique et d'autre part un granite aplitique leucocrate.

— Embú : Fonte dos Jesuitas : La roche mésocrate, où apparaît la source est connue comme gneiss de Embú ; cependant elle présente une structure et une paragenèse qui rappelle un granite orienté et qui lui donne davantage l'apparence d'une migmatite plutôt que d'un gneiss. Sa composition minéralogique est essentiellement représentée par du quartz, microcline, plagioclase, biotite, muscovite et quelques minéraux secondaires tels que sphène, apatite, calcite et allanite. De plus, il n'est pas exclu, au niveau de la source, l'existence d'une faille de direction SW-NE, dans le sens de la schistosité, qui aurait disloqué verticalement les deux blocs, favorisant le jaillissement de l'eau emmagasinée en amont.

— Pilar : Fonte da Encosta No 1, Fonte da Montanha et Fonte Serrania. Bien que ces 3 sources soient très proches l'une de l'autre, une différence géologique les sépare nettement. En effet en ce qui concerne la Fonte da Encosta, elle se situe au contact, d'ailleurs peu visible puisque il est progressif, entre les gneiss et le granite à biotite orienté. Cette orientation devient nébuleuse au fur et à mesure que l'on s'éloigne du contact, de telle forme que ce granite mésocrate est très semblable à une migmatite. De plus, on note la présence de filons de pegmatites, à proximité de la Fonte Serrania, qui sont des pegmatites à quartz, feldspath potassique, muscovite et tourmaline. Les deux sources, Fonte da Montanha et Fonte Serrania sont dues aux diaclases et fissures suivant la fluidité de la roche.

— P e t r a : Fonte Dotta : Il s'agit de ce que l'on pourrait appelé une « fausse source », puisque elle a été engendrée par les galeries d'excavation réalisées pour exploiter les nodules de graphite contenus dans les gneiss. Dans ces gneiss, qui sont la continuation de ceux identifiés près de la Fonte da Encosta, apparaissent des filons de quartz, en forme de boudinage, où se trouve le graphite.

— P o á: Fonte Aurea et Fonte Primavera : Pour la Fonte Aurea, il nous a été seulement possible d'identifier la roche dans le fond du captage, car il n'existe aucun affleurement à proximité. La roche est un gneiss très diaclasé suivant la schistosité d'où sourd l'eau. Quant à la Fonte Primavera, éloignée 1 km de la précédente, elle naîtrait dans les micaschistes et peut-être même dans un schiste à séricite et chlorite difficile à identifier à cause du degré élevé d'altération.

— R o c h á g u a : Fonte Jaraguá : L'eau affleure au contact entre les schistes de direction NW-SE à pendage subvertical et des cornéennes vertes. Il s'agit d'une zone de métamorphisme de contact entre les schistes et le granite, où se trouvent également en filon des pegmatites à tourmaline, grenat et apatite, et des amphibolites à diopside et épidote.

— F o n t a l i s : Fonte São João : Cette source se situe en bordure du massif granitique porphyroïde de la Cantareira. Le granite où elle prend naissance est une variation de faciès au sein du massif. Il s'agit d'un granite mésocrate à texture équigranulaire présentant une composition minéralogique à quartz, feldspath potassique et plagioclase, et biotite lui donnant l'apparence d'une granodiorite. L'eau sourd au pied d'une petite falaise à partir des nombreuses diaclases ouvertes dans la roche.

— P e t r o p o l i s : Cette source non naturelle se localise au coeur du bassin de São Paulo et n'appartient pas directement au Complexe Cristallin. Son captage est fait grâce à un forage qui exploite l'eau des couches sableuses de l'ensemble sédimentaire. Ce forage de 140 m de profondeur traverse une succession de lentilles argileuses et sableuses avant d'atteindre le substratum

Tableau 1

Types d'émergence

Source	Roche	Type d'émergence
Jandira	granite	fissure et diaclase
Pluma — F. dos Bandeirantes	granite	contact entre deux granites
Embú — F. dos Jesuitas	migmatite	fissure et faille
Pilar — F. da Encosta No 1	gneiss — granite	contact gneiss — granite
Pilar — F. da Montanha	granite	fissure et diaclase
Pilar — Serrania	granite	fissure et diaclase
Petra — F. Dotta	gneiss	fissure — schistosité
Poá — F. Aurea	gneiss	fissure et diaclase — schistosité
Poá — F. Primavera	micaschiste ou schiste	schistosité et fissure
Rocháqua — F. Jaraguá	cornéenne — schiste	contact schiste — cornéenne
Fontalis — F. São João	grano-diorite	fissure et diaclase
Petropolis	sédiments tertiaires	forage
Mongaguá No 1 et 2	gneiss	contact gneiss — sol d'altération

granitique ou gneissique. Il est bien probable qu'il y ait un mélange de différents types d'eau, ainsi qu'un apport de celles du substratum.

— Mongaguá : Les sources émergent au contact entre le gneiss et le sol d'altération provenant de ceux-ci. Ces sources appartiennent à la région littorale et par conséquent reflètent l'effet des embruns marins très importants dans cette région.

La géologie des différentes sources ainsi que le type d'émergence sont synthétisés dans le Tableau 1.

Résultats des mesures physiques

L'ensemble des données physiques sont présentées dans le Tableau 2.

Les valeurs de température, pH et conductivité ont été mesurées sur place, dans la source même ou bien à la sortie du captage. Les valeurs de radioactivité sont tirées en partie des études effectuées par Longo (1967). Les débits ont été en général évalués à partir des volumes exploités, car il n'existe que très rarement une sortie de vidange du captage, et de plus, celui-ci peut fonctionner comme château-d'eau. Toutes les émergences se situent entre les côtes 750 et 800 m au-dessus du niveau de la mer.

Comme le montrent les résultats, la température des eaux des roches cristallophylliennes varie de 18.5 à 20 °C. Ces températures correspondent pratiquement aux températures moyennes annuelles de l'air dans la région. Dans le cas de la « source » Petrópolis, dont la température est légèrement supérieure (22 °C), cette augmentation serait due à l'influence du gradient

Tableau 2

Résultats des mesures physiques

Source	Date de l'analyse	T °C	pH	Conduct. µmho/cm	Radioact. nCi/l	Débit l/s
Jandira	13/01/76	20	5.3	38	0.7	
Pluma — F. dos Bandeirantes	7/11/75	19	5.3	47	8.0	0.47
Embú — F. dos Jesuitas	4/07/75	19	4.7	45	3.6	0.15
Pilar — F. da Encosta No 1	26/09/75	19	5.3	32	3.7	0.14*
Pilar — F. da Montanha	26/09/75	19	5.3	28	4.1	0.12*
Pilar — Serrania	26/09/75	18.5	5.3	35	3.3	0.06*
Petra — F. Dotta	10/10/75	18.5	5.0	28	3.9	0.23*
Poá — F. Aurea	10/07/75	20	5.0	30	15.3	0.19*
Poá — F. Primavera	10/07/75	19	5.0	24		0.10
Rochágua — F. Jaragua	24/10/75	20	5.5	107	20.9	0.37*
Fontalis — F. São João	19/11/75	19	5.3	58	4.8	0.05
Petropolis	19/11/75	22.5	5.3	95		16.6*
Mongaguá — F. No 1	08/75		5.3	75		
Mongaguá — F. No 2	08/75		5.3	87		

* Débit évalué d'après les volumes exploités

géothermique, si l'on tient compte de la profondeur du forage. D'après CASTANY (1967), ces eaux ayant des températures inférieures ou égales à 20 °C peuvent être considérées comme sources froides.

En fonction de la roche réservoir, comme on pouvait l'espérer, le pH varie de 4.7 à 5.5 et caractérise des eaux provenant du granite, des roches métamorphiques ou des sables siliceux du bassin sédimentaire. Toutes ces eaux sont acides à la source.

La conductivité s'échelonnant de 20 à 107 μmho/cm est en général basse et indique, par conséquent, une faible minéralisation, représentative des roches cristallines. Les concentrations en sels dissous sont faibles, même si l'on tient compte du « weathering » assez fort dans ces régions. On peut observer que la conductivité moyenne est proche de 35 $=\mu$mho/cm pour la grande majorité, à l'exception des eaux des Fontes Rochágua, Petropolis et Mongaguá 1 et 2. Pour Rochágua la dissolution des minéraux serait donc plus facile au contact de roches filoniennes, quant à Petropolis il y a certainement une accumulation en sels dissous dans les sédiments et pour Mongaguá l'apport des sels marins par les pluies orographiques très abondantes sur les contreforts de la Serra do Mar.

Les teneurs en radioactivité variant de 0.7 à 20.9 nanocuries/litre sont basses, et les eaux des sources Jandira, Embú, Pilar Serrania sont considérées peu radioactives, tandis que les autres sont radioactives à l'émergence d'après le Code des Mines du Brésil. Seulement la source Rochágua appartiendrait à la classe des eaux fortement radioactives. Cette radioactivité est certainement due à la présence de minéraux accessoires (apatite) identifiés dans les roches filoniennes et de plus, aux dépôts d'uranite dans les fractures des roches.

La plupart des débits ont été évalués d'après les volumes exploités ou bien d'après les débits antérieurs à la mise en exploitation. De manière générale, ils ne dépassent pas les 0.5 l/s ; ce sont par conséquent des sources à petit débit, sauf pour Petropolis, où l'eau est pompée. L'amplitude des variations annuelles des débits n'est pas connue, cependant, en fonction des précipitations abondantes et constantes dans le temps, on suppose qu'elle est très faible.

Composition chimique et interprétation

Les résultats des analyses chimiques des eaux, obtenus sur le terrain et au laboratoire, sont résumés dans les Tableaux 3 et 4.

Comme le montrent les valeurs du Tableau 3, et comme on pouvait l'espérer à partir des résultats de conductivité, les concentrations en STD sont basses ; ce qui indique une faible dissolution des minéraux constituant la structure des roches cristallophylliennes. De même, les éléments comme Ca, Mg, Na et K sont présents en petite quantité.

Le calcul de l'indice d'échange de base a donné des valeurs négatives pratiquement pour toutes les sources, montrant ainsi qu'il existe un passage rapide des eaux dans les milieux de circulation de ces roches. En conséquence, ces eaux n'ont pas le temps nécessaire de se charger en sels dissous, ce qui confirme les faibles teneurs des minéralisations. L'i.e.b. positif (+0.5) de la source Poá, F. Primavera, peut s'expliquer par la présence d'argiles provenant de l'altération très poussée des schistes à séricite et chlorite, ainsi que par l'apport de chlore provenant des chlorites. Quant aux sources de Mongaguá,

Tableau 3

Données chimiques des sources étudiées

Éléments majeurs (mg/l)

Nom	Date de l'analyse	Total sels dissous	O_2	CO_2	Ca	Mg	Na	K	Al	Fe	SiO_2	HCO_3	Cl	SO_4	PO_4	NH_4	NO_2	NO_3
Jandira	13/01/76	39	12	40	3.5	2.5	1.5	2.1	0.01	0.04	3.6	20	2	2	0.05	0.05	0.05	1.9
Pluma – F. dos Bandeirantes	7/11/75	42	16	48	1.9	2.3	3.3	1.5	0.01	0.03	3.7	20	2	1	0.18	0.3	0.1	6.0
Embú – F. dos Jesuitas	4/07/75	45	13	52	4.4	2.2	3.2	1.6	0.01	0.04	1.1	18	1	0	0.15	0	0.09	10.3
Pilar – F. da Encosta No 1	26/09/75	34	19	44	1.9	2.5	1.1	2.7	0	1.1	3.6	12	3.5	3	0	0	0.08	3.2
Pilar – F. da Montanha	26/09/75	32	17	48	1.8	2.6	3.2	1.1	0.08	0	3.7	18	1.5	0	0.06	0	0.2	4.2
Pilar – Serrania	26/09/75	38	17	32	1.2	3.5	2.1	0.6	0.018	0.05	3.0	18	2	1	0.1	0	0.06	3.6
Petra – F. Dotta	10/10/75	30	28	36	0.5	3.5	1.6	1.0	0.001	0.05	2.5	10	3	2	0.07	0.5	0.03	6.3
Poá – F. Aurea	10/07/75	32	13	60	2.3	2.0	3.2	1.1	0.009	0.08	2.1	12	4.5	3	0.05	0.25	0.06	4.6
Poá – F. Primavera	10/07/75	28	12	88	2.0	3.5	0.4	0.8	0.008	0.02	2.1	10	2.3	3	0.04	0	0.06	6.6
Rochágua – F. Jaraguá	24/10/75	98	17	44	5.7	7.0	4.3	3.3	0.005	0	3.2	50	8	3	0.11	0	0.05	6.0
Fontalis – F. São João	19/11/75	56	20	76	3.0	2.0	4.8	2.0	0.01	0	4.1	23	1.5	4	0.05	0.1	0.05	6.0
Petropolis	19/11/75	80	17	28	6.0	6.7	3.5	3.2	0.03	0.04	3.2	25	6	12	0.04	0	0.03	9.0
Monguaguá – F. No 1	8/75	65			6.0	10.0	8.3		0.012	0.02	2.2	15	17.5	5	0.08	0	0.05	8.8
Monguaguá – F. No 2	8/75	75			5.0	8.0	15.0		0.01	0.03	2.1	20	25	0	0.08	0	0.03	5.7

Analyses par: M. SZIKSZAY

Tableau 4

Données chimiques des sources étudiées
Éléments traces (mg/l)

Nom	I	Br	F	B	Ni	Cu	Cr total	Zn	Pb	Detergents (LAS)
Jandira	0	0.01	0.95	0.7	0	0.03	0.017	0.07	0.007	0.006
Pluma – F. dos Bandeirantes	0.3	0.01	0.95	1.0	0.04	0.04	0.015	0.09	0.028	0.005
Embú – F. dos Jesuitas		0.01	0.95	0.2	0.04	0.05	0.012	0.14		0.008
Pilar – F. da Encosta No 1	0.9	0.05	1.0	0.8	0.04	0.05	0.018	0.12	0.14	0.002
Pilar – F. da Montanha	0.5	0.07	0.8	1.1	0.04	0.03	0.08	0.13	0.13	0.004
Pilar – Serrania	0.3	0.02	0.95	0.9	0.03	0.03	0.018	0.09	0.14	0.002
Petra – F. Dotta	0.4	0.02	0.8	1.5	0.03	0.02	0.013	0.019	0.12	0.003
Poá – F. Aurea		0	0.95	0.8	0.03	0.03	0.02	0.09		0.005
Poá – F. Primavera		0	0.7	0.8	0.03	0.04	0.018	0.12		0.005
Rochágua – F. Jaraguá	0.4	0.01	1.1	1.0	0.04	0.04	0.018	0.09	0.11	0.006
Fontalis – F. São João	0.04	0.07	0.9	0.7	0.04	0.03	0.018	0.08	0.08	0.004
Petropolis	0.1	0.015	0.92	0	0.05	0.03	0.018	0.12	0.028	0.001
Mongaguá – F. No 1		0.012	0.9	1.3	0.03	0.04	0.0028	0.09		0.004
Mongaguá – F. No 2		0.02	0.7	0.4	0.03	0.03	0.02	0.11		0.012

Analyses par: M. SZIKSZAY

dont l'i.e.b. est de +0.07 et +0.26, il y a certainement un enrichissement en chlore par l'apport des embruns marins.

Le rapport rCa/rMg<1 implique que toutes ces eaux contiennent plus de Mg que de Ca sauf pour la source d'Embú, où rCa/rMg=1.22. Dans ce cas, le Ca plus élevé que le Mg, provient de la dissolution de la calcite identifiée dans la roche. La grande quantité de Mg par rapport aux autres éléments peut s'expliquer par la présence de biotite dans toutes ces roches.

En ce qui concerne les anions, suivant le procédé mentionné dans les lignes antérieures, le chlore se trouve en plus grande quantité dans les eaux des sources Mongaguá. Pour ce qui est du bicarbonate, il est représentatif des eaux de roches cristallines et caractérise une couverture végétale abondante, ainsi que les teneurs en PO_4. Ce qui peut être vérifié également par les teneurs en NH_4, NO_2 et NO_3, ainsi que le CO_2 provenant de la décomposition de la matière organique recouvrant le sol. Les concentrations en sulfate sont insignifiantes sauf dans le cas de Petropolis, où celui-ci provient des bancs argileux intercalés dans les sables.

L'analyse des éléments traces a montré que l'iode, le brome sont présents en petites quantités et trouvent leur origine dans la matière organique. Le

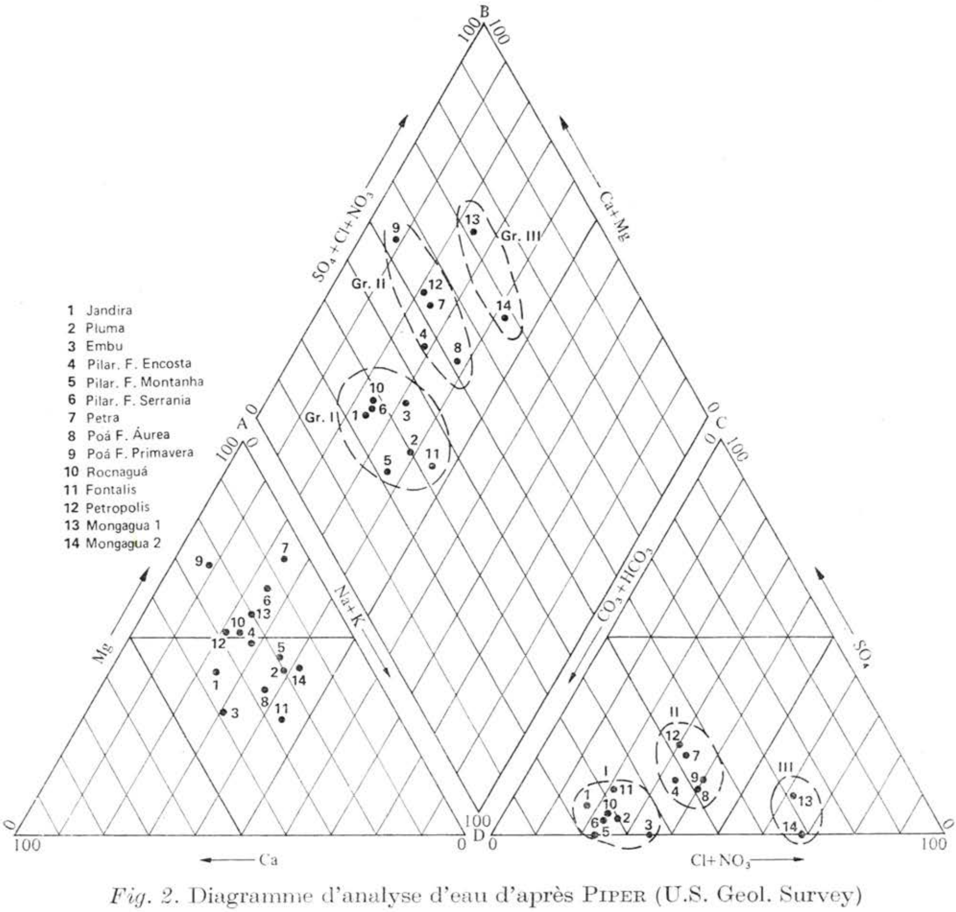

Fig. 2. Diagramme d'analyse d'eau d'après Piper (U.S. Geol. Survey)

fluor qui varie de 0.7 à 1.1 mg/l proviendrait, d'après certains auteurs, Hem (1959), Castany (1967), d'une origine juvénile, mais comme l'a prouvé l'i.e.b. négatif, il est préférable de supposer son origine dans l'apatite présente dans presque toutes les roches. Le bore en concentrations assez élevées dans la plupart des eaux, notamment Pluma, Pilar, F. da Montanha, Petra et Rochágua, correspond aux zones, où ont été recensés des filons de pegmatite à tourmaline noire. Le nickel et le cuivre sont toujours présents dans les eaux. Le zinc et le plomb proviennent des minéraux secondaires des roches granitiques. Les détergents (LAS), rencontrés dans les émergences, sont dûs aux lavages hebdomadaires des captages.

D'après ces résultats, une classification des eaux a pu être établie sur le diagramme de Schoeller—Berkaloff, comme suit : Les sources Jandira, Pilar — Fonte da Encosta No 1, Fonte da Montanha et Serrania, Petra, Poá — Fonte Primavera, Rochágua et Petropolis sont bicarbonatées magnésiennes ; les sources Pluma et Fontalis sont bicarbonatées sodiques ; la source de Embú est bicarbonatée calco-sodique ; la source de Poá — Fonte Aurea est bicarbonatée sodio-magnésienne ; et les sources de Mongaguá sont chlorurées-bicarbonatées magnésiennes.

Le diagramme de Piper a permis d'établir une séparation de ces eaux en trois groupes (Fig. 2) ; le premier groupe qui inclut les eaux des sources 1, 2, 3, 5, 6, 10, 11, le deuxième groupe les eaux des sources 4, 7, 8, 9, 12 et le troisième groupe les eaux des sources 13 et 14. Les sources du groupe I correspondent aux roches granitiques et migmatites, et celles du groupe II et III appartiennent aux gneiss, micaschistes, schistes ou roches sédimentaires. Dans le groupe II on peut distinguer la source 4 qui fait transition entre les deux et qui se situe au contact gneiss-granite ; la source 9, légèrement différente, trouve son émergence dans les schistes très altérés. Les sources 13 et 14 du groupe III sont dans des gneiss en bordure de mer. Cette séparation apparaît nettement sur le triangle des anions.

Conclusions

L'interprétation des résultats a permis de mettre en évidence d'une part les eaux provenant des granites et migmatites et d'autre part les eaux de gneiss et micaschistes.

Les eaux des sources des environs du bassin de São Paulo se caractérisent par une conductivité basse due à une faible minéralisation, conséquence d'une circulation rapide dans les fissures que confirme un indice d'échange de base négatif. Les pH acides et la prédominance des bicarbonates sont représentatifs des eaux des roches cristallines sous un climat tropical à subtropical. On note également dans ces eaux la présence de Br, I, NH_4, NO_2 et NO_3 provenant de la matière organique abondante sous ces climats. L'enrichissement en fluor est dû à l'apatite et le bore est toujours associé aux filons de pegmatite à tourmaline.

Dans la majorité, les eaux sont bicarbonatées magnésiennes et secondairement calcique ou sodique.

Toutes ces sources peuvent-être classifiées comme minérales seulement en fonction de la radioactivité à l'émergence et non d'après leur composition chimique qui les encadre dans la catégorie des eaux de table.

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THE ROLE OF NEOTECTONICS IN THE FORMATION OF THE THERMAL SPRINGS OF THE MONGOLIAN-BAIKAL OROGENIC BELT

LE ROLE DE LA NÉOTECTONIQUE DANS LA FORMATION DES HYDROTHERMES DE LA CEINTURE MONGOLO-BAIKALIENNE

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RÉSUMÉ

On examine la question du rapport entre l'activité hydrothermale et le volcanisme et le renouvellement de l'activité tectonique de la région dépuis le Mésozoïque et le Cénozoïque. Suivant les particularités du développement néotectonique et de la structure contemporaine on peut y distinguer trois zones:

1. La zone de rift du Baïkal est caractérisée par le grand développement des hydrothermes azotiques, métaniques et acidulés (carboniques). Elle subit la période Caïnozoïque (inachevée) du renouvellement de l'activité tectonique.

2. La zone néotectonique de Zabaïkalié des eaux acidulées froides qui a subie le renouvellement de l'activité tectonique depuis le Mésozoïque est marquée par le développement des thermes azotiques et carbonatées sur les terrains localisés dans des élévations voutées (Khentei-Dauria, Gaizimour).

3. L'élévation voutée de Khangai où des thermes azotiques sont très répandus.

Toutes les manifestations de l'activité hydrothermale sont en corrélation et forment un système spatial-temporel à deux branches du développement: *1* — thermes azotiques → thermes acidulés → eaux acidulées froides caractérisent des transformations réciproques (mutuelles) dans les régions du jeune volcanisme; *2* — thermes azotiques → thermes carbonatés → eaux acidulées froides sont typiques pour les dépressions intramontanes. Avec cela les différences entre l'intensification des mouvements tectoniques et de l'activité séismique actuelle dans les différents terrains de la région conditionnent l'échelle temporelle des transformations de chaque branche. Tout cela a prédéterminé, à l'époque actuelle, l'existence de tous les membres du système sur le territoire de la région.

The studied territory (about 2×10^6 km^2) is considered to be part of the vast Central Asian epicraton orogenic belt. The neotectonic structure comprizes two major regions of different types — *(1)* the southern, greatly activated part of the Siberian Cenozoic platform and *(2)* the epicraton orogenic belt of East Siberia and Mongolia. In these major structural regions the Cenozoic tectonic movements manifested themselves differently and conditioned the manifestation of hydrothermal processes both in time and space.

Cenozoic movements began in the Oligocene. The most intensive formation of relief and geomorphological differentiation took place in the Pliocene. The main features of the stage of recent activation are given by uplift (ridges and massifs) and subsidence (intramontane depressions).

Two types of recent mountain building (Baikal and Central Asiatic) were distinguished by N. A. FLORENSOV (1965, 1970) for this territory.

Neogene i.e. Quaternary movements sharply deformed the ancient consolidated substratum of Pribaikalia (= Baikalia) and North Mongolia. Mesozoic folded tectonic forms were greatly reworked (especially in the Baikal Rift Zone). However, the succession of the structural development of the main structures is expressed by the general uplift of the territory and the inherited development of the majority of large abyssal faults. Gentle fold-like flexures ("bulges") are the most common recent deformations. These flexures are complicated by the systems of renewed ancient and newly formed faults, which caused the formation of vast arch-block uplifts.

According to peculiarities of the neotectonic development and present structure, a series of neotectonic zones can be distinguished (Fig. 1). Among them three main zones are recognized—the Baikal rift zone, Zabaikalia (= Transbaikalia) and the Khangai arched uplift.

The Baikal Rift Zone is a first-order recent megastructure of a large arch-like-formed uplift. Almost all through this "arch" the axial part is complicated by a depression of Baikal type.

The development of the rift structures, proper in the Late Pliocene and Anthropogene, took place under the conditions of suppression of young plastic forms due to active rupture deformation. Abyssal processes associated with exceptionally high seismicity, intensive hydrothermal activity, geophysical anomalies, regularities of rupture formation etc., testify to the incompleteness of the rift-forming stage and its continuing development (FLORENSOV, 1960, 1970).

The Zabaikalian Zone is characterized by the alternation of low-and-mid-mountain-type uplifts with depressions of Zabaikalian and East Mongolian types. As a rule, large uplifts are weakly differentiated (Khentei-Dauria, Gazimur and others). The characteristic features of the neotectonic development are the prevalance of vertical movements and the inheritance of the Upper Mesozoic structural pattern. In connection with this, arched and arch-block uplifts prevail. Therefore, the basement of most depressions is situated at a higher hypsometric level than that in the rift structures.

The Khangai arched uplift is restricted by zones of large abyssal faults (Khangai, Orkhon, Dzavkhan, Bayankhongor etc.). The stage of its neotectonic activation was associated with intensive tectonic movements, and some volcanoes were active up to the 13th century (Khorgo, Togo, Togo-Shut). This testifies to the conditions of tension within the arch. Similar conditions were created in other smaller arches of Zabaikalia and Mongolia (Khentei-Dauria, Gazimur).

On the whole, in taking into account the leading role of fold deformation in the morphostructure of the territory studied, it is necessary to note that the neotectonics of zones mentioned above is mainly determined by the development of two systems of renewed faults—Baikal (submeridional and north-eastern) and Gobi or Khangai (submeridional). These "suture zones" have to be called the main "hydrothermal lineament". The surveyed territory of East Siberia and Mongolia is characterized by the abundance of different mineral waters (Fig. 1). From hydrogeological point of view, it is called "the Mongolian-Baikal hydromineral area". It is distinctly divided into two large provinces:

I. Thermal waters, bound to the Baikal Rift Zone;

II. Cold carbon dioxide waters and thermal waters of local development in Zabaikalia and North Mongolia. In this second province the hydrothermal springs are developed within large arched uplifts (Khangai, Khentei-Dauria and Gazimur).

It is not necessary to prove that mineral water manifestations testify to a tectonic activity in the region involved, during the Mesozoic and Cenozoic. It is more difficult to establish the genetic pattern of thermal and cold carbon dioxide waters, moreover their characteristics in time and space, determined by both evolutionary and revolutionary stages of tectonic development of the territory. The authors put forward the hypothesis that all manifestations of hydrothermal activity, being observed at the present epoch, are interrelated within the Mongolian-Baikal hydrothermal area and can be grouped into a genetic system with two main branches of development, depending on the caracter of tectono-magmatic activity. At the same time, the authors deliberately do not touch on the rather complicated interconversions of hydrothermae of an active period of volcanism, which do not occur in the region observed. Hydrothermal processes generated by active volcanism are thoroughly investigated within the island arcs (Kamchatka, Kuril Isls, New Zealand, etc.) and oceanic rifts, and are discussed in details in the literature.

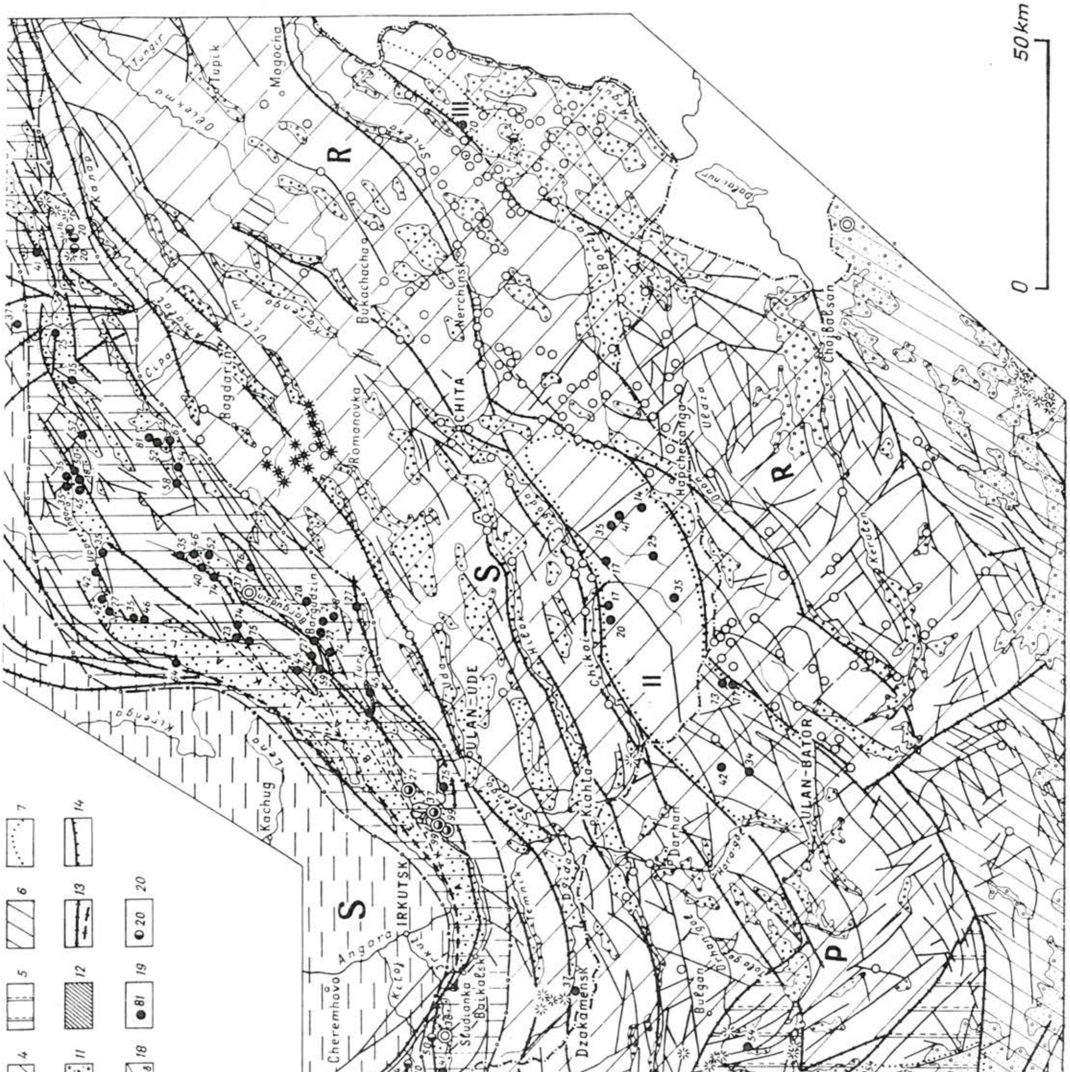
*I*st branch — nitric thermae, carbon dioxide-nitric thermae, nitric carbon dioxide thermae, carbon dioxide thermae, carbon dioxide cold waters or dry streams of carbon dioxide gas (regions with tectonic activity out of depressions).

*II*nd branch — nitric thermae, methane-nitric thermae, nitric-methane thermae, methane thermae, carbon dioxide—methane thermae, methane—carbon dioxide thermae, carbon dioxide thermae, carbon dioxide cold waters or dry streams of carbon dioxide gas (neotectonic depressions).

There is every reason to suggest that these two branches characterize in space and time the process of evolution of the hydrothermal activity. At the same time, the differences in the activity of tectonic movements in separate regions result in different scales of time of transformations in every branch. We shall consider the evolution of the hydrothermal activity based on the example given by the neotectonic zones mentioned above.

The Baikal Rift Zone

Nitric siliceous thermae are widely developed within this zone. The deposition of siliceous (geiserite-type) and sulphite sediments enriched in ore minerals is characteristic of them. In most large rift depressions (Baikal, Barguzin, Tunka, etc.) the nitric thermae rising onto the surface along faults of the basement, come across the sedimentary covering deposits. Favourable conditions for their metamorphism are created according to the direction of the *II*nd branch. Methane thermae do not appear on the surface: they are revealed at considerable depths (2500 – 3000 m) near the basement of depressions. The temperature of hydrothermae at these depths reaches 80 – 100 °C. Under such conditions an environment suitable for the formation of hydrocarbons of the methane sequence (including oil) is created. Oil concentrations are known in the Baikal depression. The intermediate members of the second branch (methane-nitric and nitric-methane thermae) are dispersed either with a smaller thickness of the overlapping sedimentary cover or in the marginal parts of depressions. The carbon dioxide – methane thermae and methane –


IRKUTSK
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CHITA
ULAN-BATOR
Cheremhovo
Angara
Kachug
Lena
Studianka
Bajkalsk
Dzakamensk
Kiahta
Darhan
Bulgan
Muren
Hatgal
Turtu
Bargudzin
Bagdarin
Romanovka
Bukachacha
Nerchinsk
Mogocha
Tupik
Tungir
Chojbalsan
Hapcheranga
Chikoj
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Fig. 1. A scheme of neotectonics and manifestations of hydrothermal activity of the Mongolian-Baikal orogenic belt

Main neotectonic structures:

1. Siberian Cenozoic platform;
2. the Baikal Rift Zone;
3. Zabaikalian and East-Mongolian neotectonic zones;
4. Sayano-Tuva neotectonic zone;
5. Khanghai neotectonic zone;
6. Dolinoozerskaya and Gobi neotectonic zones;
7. Contours of large arched uplifts with the development of nitric and carbon dioxide thermae (*I* – Khangai, *II* – Khentei-Dauria, *III* – Gazimur).

Main negative morphostructures:

8. mature depressions of Baikal type (rift);
9. minor depressions of Baikal type (embryonic and sub-rift);
10. depressions of Zabaikalian type;
11. depressions of Gobi type;
12. depressions of Khangai type.

Main dislocations with a break in continuity:

13. abyssal faults that were active in the Meso-Cenozoic;
14. rift-forming faults;
15. regional and local faults controlling morphology, giving rise (primarily before the Holocene) to the development of neotectonic structures;
16. areas of faults with the Holocene (seismogenic) renewal;
17. morphologically defined boundary of the Baikal rift zone.

Volcanicity and hydrothermal activity:

18. Cenozoic volcanoes, Quaternary *(a)* and Tertiary *(b)*;
19. springs of nitric thermae, figure—temperature of water in °C;
20. springs of carbon dioxide thermae, figure—the same;
21. springs of carbon dioxide cold waters;
22. methane thermae;
23. waters, recovered by a well

carbon dioxide thermae are confined to the intradepressional prominences of the basement (the Baikal depression). The presence of "windows" in the near-contact zone of the basement and sedimentary cover does not favour the formation of large oil and gas concentrations.

It is quite possible that in the course of time (in the descending branch of the Cenozoic stage of tectonic activity) the development of carbon dioxide thermae, followed by cold carbon dioxide waters could take place after the consolidation of the Neogene-Quaternary sediments.

N. A. LOGATCHEV has stated (1968) that within the Baikal Rift the sedimentation and volcanism are not directly associated, on the contrary, they are independent enough as being equally derivatives of abyssal processes. In connection with this, in the regions of intensive manifestation of the Cenozoic volcanism on the flanks of the rift zone (a group of the Tunka depressions, marginal zone of the East Sayan, the Udokan ridge), we can cross transformations of the first branch. Nitric thermae developed far from the volcanoes. Either carbon dioxide thermae (with a comparatively high temperature of 20 – 40 °C) or cold carbon dioxide waters and dry streams of carbon dioxide gas appear near the volcanoes. After V. P. SOLONENKO (1967), the authors are inclined to relate carbon dioxide thermae and cold waters with the mofette stage of volcanic activity. Many facts indicate that transformations of the *I*st branch are determined by changes in time of the volcanic process. Thus, the deposition of calcareous sediments (travertines) forming vast cones and fields, is observed everywhere. In cold carbon dioxide waters the sedimentation can be absent or very insignificant. Ferruginous travertines prevail in the composition of these sediments. Decrement in volcanic activity leads to transformations in the *I*st branch. Due to investigations carried out by LOMONOSOV (1974), the increased (often anomalous) metal concentrations are connected with the concerned hydrothermae; in some places the aureoles of ore mineralization are recognized.

Thus, the manifestations of metalliferous hydrothermal solutions are characteristic of the Baikal Rift Zone, which is on the stage of the Cenozoic tectonic activation; present-time ore formation is quite possible in the depth.

At present most geologists and hydrogeologists have no doubt that many ore deposits and zones of mineralization are not only connected in position but they are also united by ties of a "consanguinity" with mineral waters. A. V. LJVOV indicates that ore deposits and zones of mineralization are "petrified mineral water springs"; and springs acting at present are deposits forming nowadays. On such a ground we investigated the province of cold carbon dioxide waters and the local distribution of thermal waters (the tectonic activization took place in the Mesozoic). Within this province, such a tectonic activity has left numerous polymetallic and fluorite deposits, which are associated with abyssal faults. Ore deposits and zones of the hydrothermal alteration of basement rocks (zones of chloritization, epidotization, sericitization, kaolinization) evidence the paleohydrothermal activity to be intensive in the Early and Middle Mesozoic. According to the afore-mentioned hypothesis, it follows that the transformations of the first branch took place during a long period of its evolution. It predetermined a large development of cold carbon dioxide waters during the present epoch (Fig. 1). The process of transformations was not interrupted in time. This thesis can arouse doubt regarding the region studied, where the formation of weathering crusts in the Cretaceous to Paleogene is well known; this testifies to the "tectonic rest". However, the presence of volcanogenic formations in the sedimentary strata of Cretaceous-Paleogene age (the Tunka depression) evidences the continuing volcanic activity.

The Baikal rift-formation has influenced the adjacent territory of Zabaikalia and Mongolia where conditions suitable for the revival of the hydrothermal process were formed. Here the degree of permeability of the Earth's crust did not reach the values sufficient for the ascending migration of substance. In connection with this, the activation manifested itself only by emanation of carbon dioxide gas, enriching infiltration waters near the land surface.

At the same time, an exception will be constituted by regions of arched uplifts (Khangai, Khentei-Dauria, Gazimur) where the stage of the Cenozoic tectonic activity proceeded and is proceeding even now rather actively. Quaternary volcanism and present-day seismic activity with earthquake magnitude up to 7.75 and more (Khangai earthquake 1965, Mogodsky earthquake 1967) intensively manifested themselves in the Khangai and Khentei Dauria arches where the heights of mountain ridges reach 4000 m. Numerous springs of nitric thermae with high temperature (up to 92 °C), depositing geyserite, are characteristic of these arched uplifts. Thus, at present the conditions of hydrothermal activity are identical with the conditions in the Baikal Rift Zone.

The Gazimur arch (badly defined in the relief) is apparently on the descending stage of its development or the process of activation was weaker. As a result, hydrothermae are of carbon dioxide and nitric carbon dioxide type, with abundant travertines.

Special attention should be paid to the region of the Khalzanul deposit of cold carbon dioxide waters (PISARSKY, 1974). Obvious indications of the arched uplift are absent here. The Khalzanul spring is confined to the fault at the contact of the Mesozoic sedimentary deposits and intrusive rocks. Thick deposits of travertine evidence that the cold carbon dioxide waters had earlier been thermal. A hydrothermally changed zone and the presence of an extensive fluorite mineralization testify to the existence of nitric paleothermae during the period of the Mesozoic activation.

Within the *II*nd province we come across all the representatives of the *I*st branch of hydrothermal activity: beginning from nitric thermae to carbon dioxide waters and "dry" carbon dioxide streams.

We may state that the Cenozoic activation of the Mongolian-Baikal hydromineral area predetermined its intensive hydromineralization activity which manifested itself differently in time and space. The type of the mineral waters depends on a series of factors by various interactions and physicochemical reactions, taking place in the interior of the Earth. But the extent of neotectonic processes was the main factor in creating such conditions for the formation of waters.

Manifestation of a revolutionary tectonic activity is also reflected by the change of composition and temperature of hydrothermae. In this respect the regime of thermal springs in the period of heavy earthquakes attracts attention. According to V. P. SOLONENKO (1967), before the Muya earthquake on 27. 6. 1957, the temperature of the Hot Spring (Chara Depression) nitric thermal water was 42—43 °C. In the second day after earthquake the water temperature was measured as 47—48 °C, with sulphate and alkaline concentration raised.

During a week water was very turbid and only in 20 days the temperature decreased to 44–45 °C. When the Olekma earthquake occured (14. 9. 1958), water temperature of the spring raised again. On 5. 1. 1967, the Mogidsky earthquake with magnitude 10 took place in Khangai. In the Saikhan-Khuldzhi thermal spring (lying directly at the epicentre) a sharp increase in water temperature (12–14 °C) and discharge at some points (MARINOV–POPOV, 1963) were observed when investigating the spring in 1972 (five years after the earthquake). The increase of water temperature and discharge of wells were fixed on 5. 1. 1967 at a significant distance from the epicentre, after the earthquake in Pitatelevsky thermae near the town of Ulan-Ude.

The authors are sure that the evolution of hydrothermal processes in the Mongolian-Baikal orogenic belt (a region of neotectonics and present-day seismic activity) represents a regular process of the development of neotectonic zones of the planet.

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THE BOUNDARY BETWEEN FRESH- AND SALTWATER IN SALIFEROUS SEDIMENTARY BASINS

LIMITE ENTRE L'EAU DOUCE ET L'EAU SALÉE DANS LES BASSINS SÉDIMENTAIRES SALIFÈRES

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RÉSUMÉ

La région regardée appartient à la cuvette d'Europe centrale remplie des sédiments du Paléozoïque supérieur jusqu'au Quaternaire. Le fond atteint 5000 m au maximum. Intercalé aux dépôts pélitiques, psammitiques ou calcaires se trouvent des dépôts épais de sel gemme, appartenant au Zechstein (Thuringien), au Buntsandstein supérieur, au Calcaire coquiller et au Keuper. Les cavernes dedans sont remplies surtout par de l'eau minéralisée, dont la charge de sel va de quelques grammes jusqu'au plus de 200 g/l. Seulement près de la surface de la Terre on trouve de l'eau douce d'origine météorique. Les strates portant l'eau douce comptent normalement de 50 à 1000 mètres et ils n'atteignent 300 mètres qu'en cas extrêmes.

La limite entre l'eau douce et l'eau salée se détermine par l'extension régionale des strates imperméables, surtout des argiles de l'Oligocène moyen et aussi par l'hydraulique des relations de pression entre l'eau douce et l'eau salée. De nombreaux « endroits de sel » (c'est-à-dire des sources naturels à la surface de la terre) donnent une première impression concernant la dynamique et composition chimique des eaux minéralisées.

The area discussed is a part of the Central European sedimentary basin. It is filled with sediments from the Palaeozoic to the Quaternary. The hydrogeological conditions of the basin depend on the following geological factors:

— extensive salt layers in the Zechstein, Buntsandstein, Muschelkalk and Keuper sequences,

— great thickness of sediments (more than 10 km),

— long cycles of sedimentation interrupted only by short periods of regression in the central part of the basin,

— overall development of halokinetic processes from the Triassic upwards.

As shown by an analysis of the paleotectonic development, the lithological-faciological conditions and the dynamics and geochemistry of ground water, the following ground water complexes can be distinguished within the basin:

— Ground water complex of the pre-Zechstein which comprises the permeable and impermeable layers of the Carboniferous, Devonian and the Rotliegendes.

— Ground water complex of the Zechstein (Upper Permian) which is mainly represented by a regional impermeable layer (aquiclude) and the aquifers of Stassfurt serie and Plattendolomit.

– Triassic ground water complex (without Rhaetian) which combines the aquifers of the Middle Buntsandstein and Schilfsandstein with the joining relatively impermeable layers of the Lower and Upper Buntsandstein, Muschelkalk and Keuper (without Rhaetian).

– Rhaetian-Oligocene ground water complex, which contains the aquifers of the Rhaetian, the Jurassic, the Lower Cretaceous and the Paleogene, and the intercalated impermeable layers as well.

– The Neogene-Quaternary ground water complex, saturated mainly by freshwater, is separated from the former complex by the Middle Oligocene Rupelian clays.

The basic hydrodynamic regime of the mineral-water-bearing ground water complexes of the pre-Zechstein, Zechstein and the Triassic corresponds to the hydrodynamic model of an artesian slope. Fig. 1 shows a generalized hydrodynamic diagram.

The main catchment area of the ground water complexes can be found outside the area investigated, in the territory of Poland. Only the western tongue extends over the territory of GDR (Fläming). At this point, infiltration waters flow with different intensity into the various aquifers of the basin. Though infiltration is remarkable even in the pre-Zechstein, the lateral dimension of water is negligible. The bulk of the waters infiltrated is discharged, still in the vicinity of the catchment area, into the overlying aquifers or directly into surface water bodies. This model is corroborated by the frequency of near-surface mineral water anomalies south and west of Berlin. The quantity

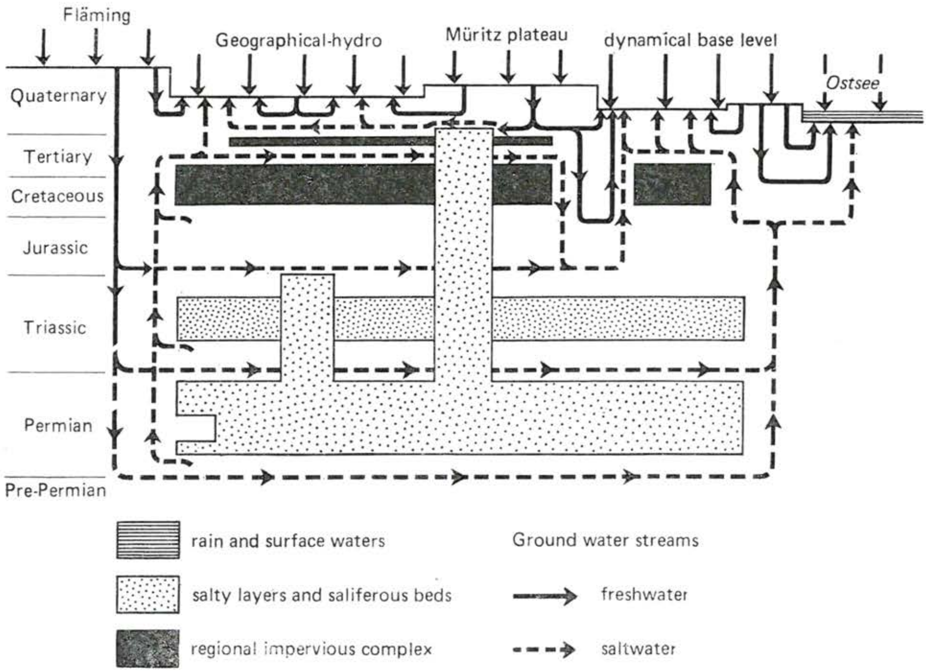

Fig. 1. Hydrochemical model of the Nord–East German Lowland (after ZIESCHANG, 1974, simplified and generalized)

of the infiltration water depends mainly on the storage capacity and on the lithological environment of the aquifer. This is why, in some cases, water have penetrated farther into the Rotliegendes than into the overlying aquifers of the Zechstein.

The recharge and discharge of salt solutions of the Zechstein and pre-Zechstein beds are heavily controlled by tectonic disturbances. In the ground water complex of the Triassic the zone of influence of infiltration is 50 – 60 km wide. In this ground water complex not the tectonic disturbances are the most important factors controlling infiltration and recharge, but the extension of the Rötsalinar (Upper Buntsandstein), the clayey-marly sediments of the Upper Cretaceous and the Rupelian clays. The lack of Rupelian clays in the SE areas as well as the partial transition of Upper Cretaceous clayey-marly formations into sandy ones constitute the proper conditions for freshwater supply to the Triassic ground water complex. This process is proved by the reduced overburden pressures calculated for the formation waters of the Triassic. The emerging saltwater is discharged into the valleys of the Spree, Oder and Havel rivers.

The northern part of the examined area represents another area of discharge of mineralized formation waters. The discharge of formation waters is promoted here mainly by the following factors: the outcrops of Zechstein saltbeds, the frequency of tectonic disturbances, the relatively shallow position of the aquifers (in the case of formation waters of Zechstein and pre-Zechstein ground water complexes), the lack of the Rötsalinar (in the aquifers of the Buntsandstein), the lack of impermeable layers in the Upper Cretaceous and Rupelian clays (in the Rhaetian-Oligocene ground water complex).

Contrasted with the dynamics of formation waters the dynamic conditions of freshwaters are determined by the morphology of the generally permeable Quaternary regions and the main drainage area. The basic principle is shown on Fig. 1. The young formations of Fläming, on the southern edge of the basin and of the so-called "nördlicher Höhenrücken" (in the model denoted with "Müritz plateau") were interrupted by the Pleistocene paleostream valleys. The northern discharge area is represented roughly by the slope sweeping down to the Baltic Sea. The altitude differences of ground water tables related to the base level of the Baltic (80 m in the main catchment areas, 30 m in the paleostream valleys) are determinant for the hydrodynamic conditions of both the freshwater sequence and the upper part of the saltwater complex. The dynamic relations of the two complexes constitute a criterion for the development of the fresh- and saltwater interface.

The presence or the lack of the uppermost regional ground water aquifuge, the Rupelian clay, constitute a further criterion for the depth down to which freshwaters penetrated and the height up to which mineral waters emerge. As evidenced unambiguously by several exposures the Rupelian clay, if intact, is that layer forming the boundary between the two hydrogeological complexes. However, this state is disturbed, mainly by isolated, local exarations in the south and contiguous areal exarations in the north (Fig. 2).

In this territory the variation of the fresh-saltwater interface is determined only by the hydrodynamics of the uppermost ground water complexes. The varieties of the form of such local boundaries are shown in Fig. 3. In the recharge area, at such "windows", freshwaters may penetrate 300 m down well below sea level. The saltwater swept away by the flowing freshwater

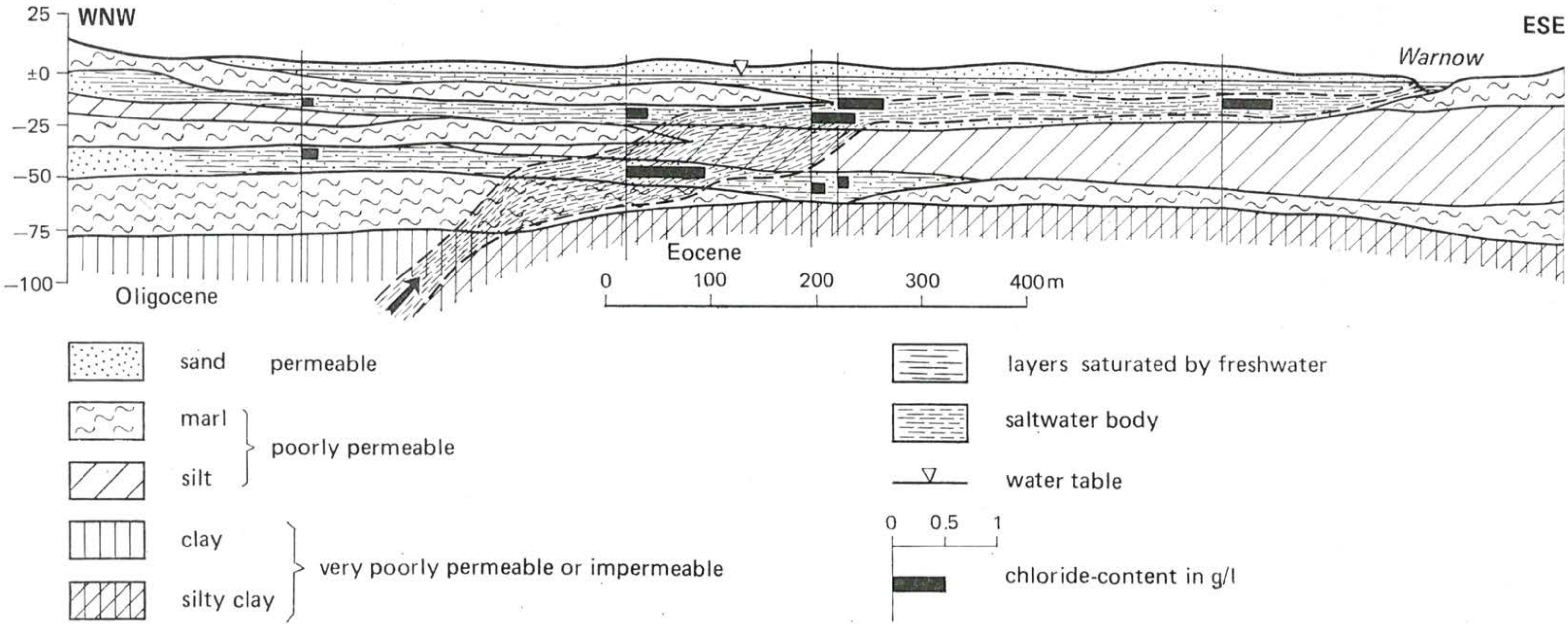

Fig. 2. Development of a saltwater zone in a freshwater complex by direct discharge into a natural catch-drain (example: Warnow-valley, north of Schwaan) (after ZEIDLER, 1976)

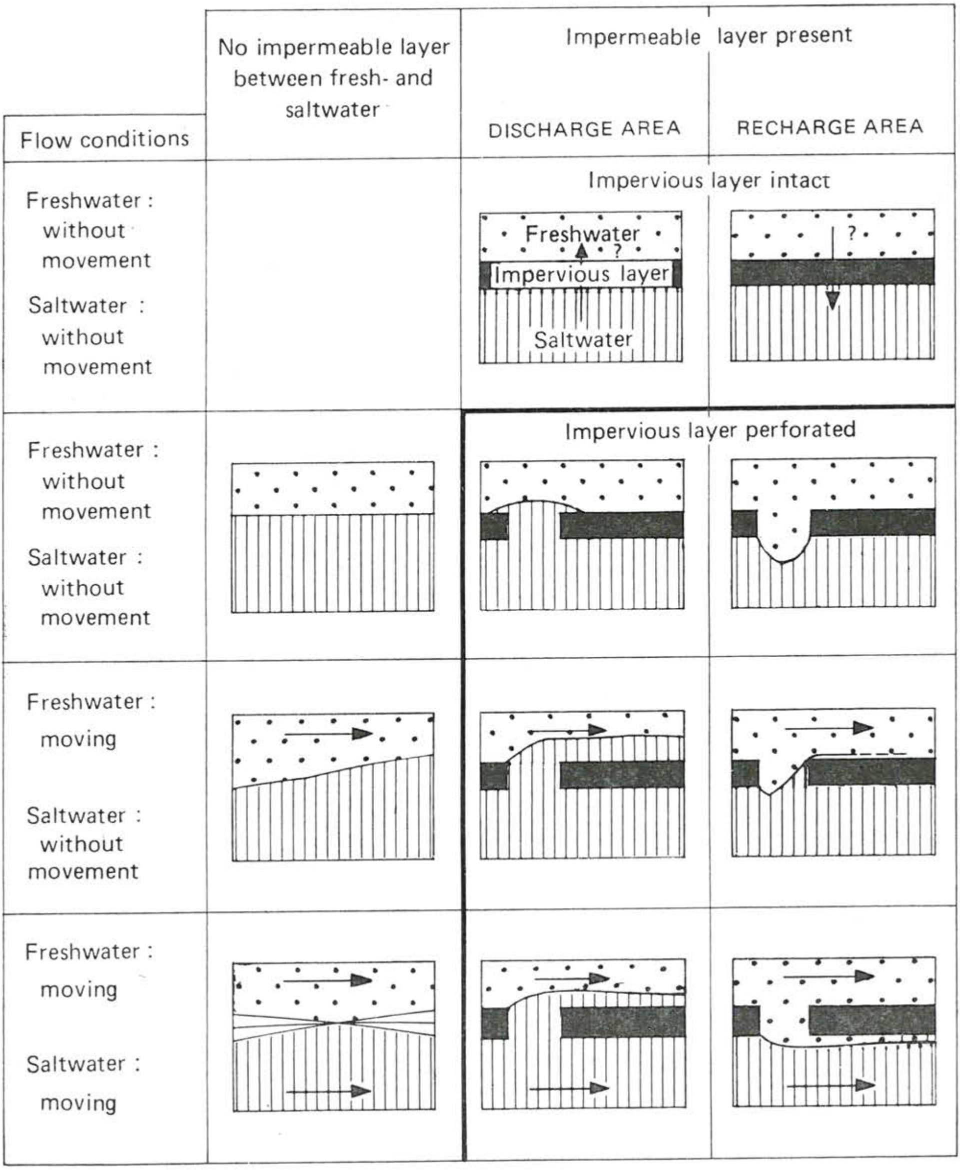

Fig. 3. Different types of boundaries between fresh- and saltwater zones

will be rather diluted and can be hardly detected. In the discharge area a great quantity of saltwater penetrates into the freshwater complex and in some cases it can reach even the water table, forming near-surface mineral water anomalies, so-called "Salzstellen". Most of these anomalies concerning—at least—their position, were described already 100 years ago.

Nevertheless, the hydrochemical-hydrogeological elaboration has been systematically completed only in recent years. As a result of the field investigations of the subsurface mineral water anomalies ("Salzstellen") the following observations could be made:

– the "Salzstellen" form distinct local patches which are definitely isolated from the surrounding freshwater;

– the distance from the peak of mineralization to the edge of "Salzstelle" is often not more than 100—150 m.

Accordingly, field investigations have allowed the author to confirm the results obtained by the so-called "Dreischichtsand" model (VOIGT—ZEIDLER, 1975).

According to the sources of the salt content of the waters 2 types of near-surface mineral water anomalies can be distinguished:

a) the high degree of mineralization is the result of leaching processes attacking salt domes;

b) the mineral waters represent discharging formation waters of the Rhaetian-Oligocene or Triassic ground water complexes.

In the "Salzstellen" occurring in the northern part of the basin the highest values of salt content (40 g/l or even more) could be determined. In most cases the mineral waters had a high Ca content, which feature shows their Triassic origin. Nevertheless, the matter in hand is that these are usually NaCl-waters. Inside the mineral water anomalies the changes in ionic ratios are insignificant. The sulphate content is also negligible.

The "Salzstellen" of the southern part are completely different from the northern ones. First of all the high sulphate content and the lower content of dissolved solids mark them out. Waters of major chloride content could be generally observed only in the centre of the anomalies. Towards the edges this water was gradually replaced by sulphatic water. In the area of the "Salzstellen" there are significant changes in the ionic ratios.

The scattered mineral water anomalies of the western part of the basin are generally connected with salt blocks and occur in a certain dispersion zone under the salt block.

In the regional position of fresh- and saltwater interface to be given hereafter these dot-like anomalies, i.e. the so-called "Salzstellen", have not been taken into consideration. Nevertheless, some connections do exist here.

The northern area encompasses the islands and peninsulas near the seashore and the mainland as far as the maximum extension of the Weichsel glacial stage. The fresh- and saltwater interface is here between 0 and −100 m m.s.l., in larger areas about −50 m m.s.l. In the deep valleys, especially beyond the Rupelian clay areas, there are several "Salzstellen".

The next area covers the territory of the maximum extension of the main (Saale) glacial stage. The freshand mineral water interface is below −100 m m.s.l. In extreme cases it can drop below −300 m m.s.l. No "Salzstellen" have been recorded here though some (80 – 90 m a.s.l.) halophyte-occurrences in hypsometric height have been described in which no near-surface mineral water anomaly is expected to occur. South of Frankfurt/Oder this area passes over into the southern edge of the sedimentary basin which is mainly aligned around the southern ridge, the Fläming. The features are the same as above. Between the two areas with a relatively thick freshwater zone along the rivers Spree, Havel and the lower reaches of the Elbe, the fresh- and saltwater interface rises again to the – 100 and – 50 m m.s.l. range.

The subsurface mineral water anomalies are mainly connected with local erosions of Rupelian clays and they occur along the lower reaches of the Elbe, on the "leeward" sides of high elevated salt domes, respectively.

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Discussion

H. SCHOELLER as a comment to the paper of CERMÁK – LUBIMOVA – STEGENA: "As you said, there are heat flow correlated with tectonics, and the mean heat flow diminished with the age. It is not only in this country, but also in other countries. You say that the belt of Alpine orogeny generally coincides with a geothermal high. The Pannonian Basin is a part of the area affected by Alpine tecto-orogeny. Here is not an uplift but, on the contrary, a subsidence. In the eastern part of the Pannonian Basin in the Rumanian territory there are many volcanoes with a great number of thermal springs."

T. BANDRABUR, comme réponse à la question posée par M. SCHOELLER concernant aux mesurages de la temperature dans le gisement: "Dans la dépression Pannonienne du territoire de la Roumanie on a effectué des mesurages de la température dans des forages seulement en ce qui concerne les roches; la temperature de l'eaux a été mesurée à l'orifice du sondage".

A la deuxième question de M. SCHOELLER quant à la situation similaire des eaux thermales en Roumanie à celle du Plateau Central de France, où le géothermisme est engendré par l'activité volcanique, la réponse de M. BANDRABUR est : « Les déterminations d'âge absolu effectuées sur les produits volcaniques de la chaîne de Călimani-Harghita, qui ont révélé des âges compris entre 7.37 et 3.92 m.a. nous empêchent d'admettre l'éxistence d'un géothermisme fossile. Les eaux thermales des régions à volcanisme néogène de Roumanie sont considérées des produits indirects des manifestations postvolcaniques. La distribution de l'énergie géothermique dans la lithosphère est due à un complexe de facteurs physico-géologiques rattachés à la structure majeure de l'écorce terrestre et aux types du méchanisme de transfert de la chaleur des zones internes vers l'extérieur de la planète. La contribution des roches volcaniques à l'apparition des anomalies géothermiques provient de leur conductivité plus élevée que celle des roches sédimentaires et metamorphiques, en favorisant un transfert conductif plus intense de chaleur des zones profondes de la lithosphère vers la surface. »

A. LORBERER: "According to Prof. STEGENA and his collaborators the Pannonian Basin is an interarc basin with a continental (ensialic) crust where the average depth of the crust is 26 km and the minimum is 23 km. Beside the smaller depth of crust its further important characteristics are: lower density of the upper mantle, smaller seismic velocity, raised state of LVZ and HCL, reduction of the lithosphere, beginning with an andesitic and ending with a basaltic volcanism, low crustal seismicity and positive Bouguer-anomaly. From the geothermal point of view it is characterized by high temperatures and great surface- and mantle heat flow.

In the light of these characteristics they explain the formation of the Pannonian Basin with the plate (global) tectonical theory and they compare it with the Great Basin in the U.S.A. According to them, the positive geothermal anomalies that may be observed in the Pannonian Basin can be traced back to the subduction of the neighbouring Carpathians surrounding the basin: above the subsiding lithospherical plate the partially molten material (active mantle diapir) generated within the Earth's mantle has laterally spread at the Moho-discontinuity and has reduced it from beneath, resulting in subcrustal erosion. High heat flow values can be explained with the heat convection of the ascending mantle-material in the upper mantle. So this heat convection does not occur in the basin sediments but in the upper mantle. Upon its effect, the surface heat flow has regionally increased which only in a small degree could be modified by the confined water circulation with its own heat convection in the basin sediments."

W. M. TURNER: "The work of our distinguished Hungarian colleagues in the Great Hungarian Plain has revealed significant geothermal anomalies within the ground-water system. The explanation for these geothermal anomalies seems to be based on either the shallow occurrence of the Mohorovićić discontinuity or shallow igneous activity. In my opinion these explanations are unlikely and unnecessary to account for the geothermal anomalies observed.

Ground-water in the Great Hungarian Plain is recharged around the fringes of the ground-water basin and moves in the subsurface to points of ground-water discharge. During its movement a considerable amount of the ground-water reaches great depths within the basin and consequently, because of the high heat capacity of water, the ground-water absorbs a considerable quantity of geothermal heat during its passage from its recharge to its discharge zone. The measure of the amount of heat in storage within the water, of course, is temperature. As has been pointed out by our Hungarian colleagues, the geothermal anomalies generally coincide with zones of geothermal discharge. This is to be explained by the fact that in zones of discharge deeply circulating ground-water moves upwards, thereby conductively conveying the geothermal heat towards the surface."

E. P. WRIGHT: Referred to DR. RÓNAI's paper on the Great Hungarian Plain with particular reference to the statement that seepage velocities could be evaluated only by the determination of the absolute age of the underground waters. He commented that there could well be errors in interpretation apart from uncertainties in the current techniques of dating. He reffered to studies carried out by the Institute of Geological Sciences in the large (200,000 sq.km plus) Sirte Basin in Lybia. Age studies showed considerable variations both in vertical and horizontal planes, the latter being related to periods of recharge mainly localized along fossil valleys. Ages ranged from 40,000 to 8000 years before present. With such complexities, interpretation of seepage velocities from relative ages could easily be in error. The speaker considered that age dating should be used in combination with other geochemical evidences and with the more standard hydraulic methods of measuring ground-water seepage velocity which are applicable as well.

P. LIEBE: "I would like to say a few words concerning the anomalies of the geothermal gradient of the Hungarian Basin.

Lectures and contributions here dealt with the movement of ground-water between the recharge belts and the permeable deposits of the discharge areas inside the basin. The question can be raised that the geothermal anomaly of the basin is not caused by ground-water warming up while moving from the recharge zones to the discharge areas of the basin.

In my estimation about 10 m³/sec. is the amount of water moving in the subsurface part of the hydrological cycle of the Great Plain. This quantity of water warming up to 15–20 °C decreases the heat-flow by about 20 percent in the recharge zones and increases it by about 10 per cent over the discharge areas. This phenomenon does not explain the entire basin's geothermal anomaly exceeding by some 60 per cent the global average, but has an effect on the temperature distribution inside the basin."

V. S. KOVALEVSKY: "I would like to say some words dealing with the paper presented here concerning the helium survey. Studies of helium distribution in the Earth's crust, carried out by I. N. YANITSKY and other investigators in the U.S.S.R., have shown a regular depthward increase in the helium content of the ground-water. Helium concentration increases most rapidly with depth in case of the occurrence of the crystalline basement at shallow depth. Helium is "squeezed out" from the earth's interior under overburden pressure owing to rock diagenesis and is transported by ground-water flow to the ground surface and then dissipates into the outer space. The most favorable paths of helium flow are faults, in which ground-water occurs, and hydrogeological windows where anomalously high concentrations of helium are recorded. Particularly intensive anomalies are observed in places of fault intersection. The regularities the observed migration of helium offer wide possibilities for the use of helium as a means allowing the solution of different practical tasks facing hydrogeologists. Among these tasks are:

1. Estimating the extent of isolation of an aquifer from adjacent aquifers for assessing the possibilities of industrial waste-water disposal and for other purposes.

2. Recognizing zones abounding in ground-water (fault zones, hydrogeological facial windows) for the construction of water-supply wells.

3. Precise determination of the origin of ground-water in general and that of mineral water in particular.

4. Studying the regime of confined ground-water in the zone of slow water circulation, including ground-water recharge estimation when common observation results are little constrasting and uniformative.

5. Estimation of the extent of man's economic impact upon ground-water (e.g., variation of the values of leakage or interaction of the aquifers and changes in the hydraulic interrelationship between surface and ground-water).

6. Prospecting for mineral deposits.

The above tasks may be most effectively solved in artesian basins."

A. RÓNAI: "I want to make some remarks upon the question raised by MR. SCHOELLER. I do not think that our temperature anomalies in the differently deep aquifers are caused by neovolcanism unlike the situation in Transylvania may be. We have huge anomalies over large areas covered by 3000–5000 m of rather unconsolidated Pliocene and Quaternary sediments which overlie rocks of very different types. There are anomalies over the Paleozoic and older cristalline rocks and metamorphites, over the Mesozoic sediments as well as in some places, though in smaller extent, over Mesozoic and Tertiary volcanics.

The general anomalies may be explained by the geophysical peculiarities of the Earth crust; some of the local differences may be caused by convective currents and by subsurface water circulation.

The great temperature differences observable close to the surface, in shallow aquifers over a distance of a few kilometers in different points cannot be explained by mechanisms other than subsurface water circulation. And this is what I want to answer to Mr. Wright's question as well. I agree perfectly with his comments made upon the value of interpreting seepage velocity from radiological dates."

H. Schoeller : « Je veux simplement indiquer que le système thermique fonctionnant comme un thermosiphon, ne peut exister avec une telle intensité que si le flux de chaleur terrestre a une intensité superieure à la normale. »

W. A. Visser referring to a statement in the General Report concerning the contribution by Čermák, Lubimova and Stegena, notably, that the geothermal high beneath the Pannonian Basin is due to upward-migrating mantle material (mantle diapir) generated by subduction processes provoked by the bordering mountains, asked to have this substantiated, e.g. by a section through the Earth's crust. According to other authors (Korim; Airinei and Pricajan), the results of seismic refraction and gravimetric surveys indicate the Moho discontinuity beneath the basin to be shallow, about 26 km, which, with Late Tertiary activities, would account for the geothermal anomaly.

Since these data involve the fundamental problem of the development of geosynclinal and orogenic belts, a further elaboration of the authors' views would be welcome, including their views on the relative movements of the European and African continents and their paleoboundary.

Theme 4

WATER EXPLORATION, WELLS, WATER REGIME, WATER BALANCE, SUBSURFACE THERMAL ENERGY, WATER RECHARGE. POLLUTION

Thème 4

PRODUCTION D'EAU, PUITS, RÉGIME HYDRAULIQUE, BILAN D'EAUX, ÉNERGIE GÉOTHERMIQUE SOUTERRAINE, RECHARGE. PROBLÈMES DE POLLUTION

GENERAL REPORT 1

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Fourteen papers have been presented in section 4 on the following topics:

– Hydrogeology and geothermics of the Pannonian Basin, geothermal anomalies and use of earth's heat	6 papers
– The phreatic aquifer, resources, fluctuations and recharge	4 papers
– Underground disposal of industrial waste and storage of warm water	2 papers
– Water resources in artesian basins	2 papers

These papers were prepared in the following countries (in alphabetical order):

France	1 paper	Romania	1 paper
W. Germany	1 paper	Russia	4 papers
Hungary	4 papers	Turkey	2 papers
Netherlands	1 paper		

For a large extent man is dependent on fresh ground water for his needs – household, agriculture, industry. Now in several parts of the world pollution of surface waters is becoming a serious threat, his dependence on this source of as yet not polluted waters may become even greater. In a certain sense, it being replenished from precipitation and surface waters, ground water is inexhaustible. However, good management and conservation of this natural resource is needed.

Another demand that is being made on subsurface waters is the extraction of geothermal energy. High-caloric energy (200° centigrade and more or natural steam) related to magmatic and volcanic activity in young orogenic belts has been developed in some countries (f.i. Iceland, Italy, New Zealand, California). It is found that in deep sedimentary basins the geothermal gradient is often considerably higher than the average value of 3 °C/100 m. Exploration and development of this source of low-caloric energy (up to about 150 °C) are still in their infancy; examples are the Pannonian Basin in Hungary, the Paris Basin in France and the foothills basin to the N. of the Caucasus in Russia. It is a matter of quality and capacity of the reservoir rock, whether here is a source of economically viable energy. The production and use of these warm and saline waters are not without environmental hazards.

Sedimentary basins contain still other natural resources: to mention oil and natural gas, coal and lignite, rock salt, some metallic ores. Since many scores of years brines produced with oil and gas have been disposed of underground in porous/permeable reservoir beds. In the USA from underground brine disposal the technique of disposal of industrial waste was developed that is now successfully applied in various countries. Caverns in rock salt are being used for the storage and recovery of economically valuable materials, including energy, but also for the disposal of, especially radioactive, waste.

It will be clear that, in order to be able to exploit the natural resources of sedimentary basins, a multidisciplinary exploratory effort, involving risk of money, is required. Management and supervision are needed to reconcile contradictory interests, for example production of oil versus disposal of waste, construction of a solution cavern in rock salt for extraction, or for disposal or storage, protection of useable ground water. In all cases harm to the environment must be avoided.

In any sedimentary sequence depth of burial and age bring about compaction and diagenetic changes in both sediment and interstitial waters. As a result porosity and permeability—and the yielding capacity of reservoir beds — tend to deteriorate, whereas the content of dissolved solids tends to increase with depth. In marine and even in terrestrial sediments the original (depositional) waters often reach salinities considerably higher than that of sea water and attain entirely different ion ratios. A number of chemical and physical reactions occur, in which clay beds play an important role. Clay, whose particles are electrically charged, may present a barrier for ions, whereas water molecules may be able to pass through. Indeed, one often sees that clay beds form boundaries between different water qualities. Similarly the compaction of clays is a very slow process, in which porosities diminish greatly (from some 60—90% to may be some 10% in a well consolidated and indurated claystone). Due to compaction water must be expulsed from the clays, and it can move only in an upward direction at an extremely slow rate. This rate determines the speed of compaction and diagenesis, both processes involving a geological span of time. In this process the, for human perception, stagnant waters reach chemical and physical equilibrium with the sediment.

These natural processes are so slow that as a rule (there are some notable exceptions) pore pressures in both, permeable and impervious or semipervious beds, remain hydrostatic: the weight of a water column, depending on the amount of dissolved solids, as an average 0.107 atm per meter depth. Quick human action (production of oil, gas, thermal waters), however, may disturb this system, pressure transfer being too slow a process for re-establishment.

Of course, what has been said applies to the deeper parts of a sedimentary basin. To depths of perhaps a few hundred meters infiltrating surface waters may circulate more or less freely through the sediments, their movement being impaired by clay beds and clay lenses and as a rule slowing down toward depth. At a certain depth, generally below a consolidated, thick and extensive clay bed, no movement or meteoric influence is discernable any more. In some cases, however, along some fracture zones or faults, and in artesian basins in the classical sense meteoric waters penetrate to great depths; in artesian basins through extensive beds of reservoir rocks that outcrop in the basin's rim. So is in the centre of the Paris Basin the water to a depth of about 1200 m fresh in marine sediments and deeper down it increases only slightly in dissolved solid content. Under true artesian circumstances pressures are higher than hydrostatic causing the waters to rise above surface when penetrated by a drill hole.

Nevertheless, in one basin artesian and non-artesian circumstances can go together, as it is clear from the papers on the Pannonian Basin presented at this conference.

KORIM's paper "*Hydrogeologic factors governing thermal water occurrences and recovery in the Pannonian Basin*" contains a concise description of the subsurface geology and hydrogeology of the Hungarian plain. The Pannonian Basin (Hungarian, Carpathian Basin as I have seen it called in other papers) is an intramontane basin surrounded by Alpidic chains. Subsidence, which began during mid-Miocene, was not uniform; differential epeirogenic movements gave rise to many sub-basins and widely varying thicknesses of the Neogene formations (depth contour map, Fig. 1, in KORIM's paper and the section, Fig. 3, in that of ALFÖLDI et al.). The neritic Miocene, overlying the crystalline basement is some tens of meters to 3000 m thick; it contains some limestone aquifers and blanket sands. During the Pliocene (Lower and Upper Pannonian and Levantian sub-stages) facies changed from neritic to shallow marine and near-shore or lacustrine. The Pliocene is up to 5000 m thick and it contains, especially in its upper parts, the most prolific aquifers in the Neogene/Quaternary sequence. The Quaternary, consisting of stream and channel deposits, contains excellent aquifers.

Multiple reservoir systems consist of lenticular and coalescing sands, laterally and vertically confined by wedging out and changes in permeability; they are of highly variable dimensions and depths.

The quality of the aquifers tends to decrease with age and depth of burial, and there is an increase in the contents of dissolved solids in the interstitial waters: from fresh in Quaternary and Levantian via fresh to brackish in Upper Pannonian and brackish to saline in Lower Pannonian and Miocene. The occurrence of brackish waters in fully marine sediments indicates artesian conditions; saline waters and the depletion of several thermal water reservoirs, accompanied by pressure decline, indicate other parts of the basin to be isolated from meteoric influences.

The low-caloric underground waters are produced from Neogene especially Upper Pannonian reservoirs. KORIM mentions a bottom-hole temperature up to 145° centigrade at depths between 2000 and 2300 m (that is in the deepest part of the basin in the SE); PIKLER* states temperatures at an average of 50° at 1000 m depth, of 100° around 2000 m.

ALFÖLDI et al. quote an average gradient of 5.4 °C/100 m in the area between Danube and Tisza rivers. In the Pannonian Basin the geothermal gradient is well above and almost twice as great as the average world gradient of 33 °C/100 m. This, with the large yields (1100 to 1500 and even up to 2000 l/min. — up to 120 m^3/h), gives an enormous geothermal potential.

The other regional paper on the Pannonian Basin is that of AIRINEI and PRICAJAN, "*A conceptual pattern for the complex investigation of thermal phenomena in the underground waters of Romania's Western Plain*".

* PIKLER, F. (1975) — Utilisation des Eaux thermales en Hongrie. Journées sur l'Emploi de la Géothermie dans le Chauffage domestique et industriel, Paris 23 et 24 juin 1975.

The West Romanian plain is the E rim of the Pannonian Basin bordered by mountainous areas to N, E and S. A sequence of Mesozoic, Tertiary and Quaternary deposits, several hundreds to thousands of meters thick reflecting horst and graben structures, overlie the crystalline basement. One must assume that Mesozoic and lower Tertiary wedge out toward the main part of the basin in the W and that the heavy fracturing as marked out on the sketch map, accompanying the paper, is essentially a border feature of the basin.

The plains were investigated by drilling and geophysics. The regional Bouguer anomaly and several local anomalies reflect basement structure. The six extensive magnetometric anomalies are stated to represent the effects of (basic?) instrusive masses conected with the Neogene volcanism in the mountains to the E. It is not clear to me, what is meant with "electrometric data"? Is here referred to telluric or to geoelectric (resistivity) surveys? The latter provide, if controlled by borehole data, an insight to a depth of several hundred meters into the distribution of aquifers and clay beds, water quality and temperatures.

Thermal conductivities were determined from rock samples. Thermal surveys in boreholes indicated that the geothermal gradient of 5.4 °C/100 m from the main part of the basin extends on Romanian territory and decreases to the normal value of almost 3.3 °C/100 m in the bordering mountains. Several positive geothermal anomalies showing three types of regionality are superimposed on the general gradient. In two regions (Oradea and Calacea Spa) thermal springs were found to be related to high temperature waters in the Upper Pannonian.

According to the authors distribution and circulation of underground waters are determined by faults and fissures in the sedimentary cover with lateral closures being present. "Temperature inversions", an example of which is given, show the problem to be of a considerable complexity. It leaves, however, a few questions:

- are the quoted temperatures bottom-hole values?
- the quoted, downward-decreasing pressures are obviously taken at well-head, and if so, there is reason to assume that they are artesian over-pressures; hence are there any outcrops of the containing reservoir rocks in the mountains to the E?
- what are the salinities of the underground waters?

Unless true artesian conditions occur, it is difficult to follow the authors in their notion of large and deeply penetrating circulation systems of descending and ascending streams; in my opinion it is questionable whether fracture zones could sustain large flow systems.

A large artesian flow through permeable reservoir rocks is concluded to be present below the Hungarian plain. This is a well documented report by Deák, "*Study on the recharge of deep groundwaters and their connection with shallow groundwaters using environmental isotopes in the Nagykunság region, Hungary.*"

Age determinations by $^{13}C/^{12}C$ isotope ratios showed that water from a depth around 400 m is significantly younger than that from a depth around 80 m, hence indicating an upward movement of water. This indicated flow was further investigated in the Nagykunság region. Due to human activities the tritium content of the precipitation increased about 20 years ago, and it was found that no low values occur in the ground waters, less than 20 m deep. The tritium concentration is determined by the degree of permeability of the soil; and it is concluded that the shallow ground waters are recharged by precipitation.

The deuterium/hydrogen stable isotope ratio in the shallow ground waters is comparable to that of the precipitation, but deviates from that of the deep waters (50 to 1000 m). The carbon isotope ratios in the shallow ground water match those from other parts of the Hungarian plain, but deviate significantly from those from the deep waters.

It is concluded that the shallow ground water is recharged by precipitation and that in the levels below 60 m artesian waters move upward. The stratum between 20 and 60 m depth collects both waters that are discharged into Tisza River. This is corroborated by piezometric observations: the level in the bed between 20 and 60 m deep is lower than that either in the shallower or in the deeper strata.

It should be noted from the piezometric graphs, Figs. 9–11, that in the deep waters pressures are higher than hydrostatic. The authors mention that the opposite is the case in the piedmont areas and in the higher parts of the basin between Danube and Tisza Rivers and in the NE; thus a huge artesian flow system is sustained toward the low region E of Tisza River. It would be interesting to learn whether this system affects the distribution of the geothermal gradients and how a local geothermal anomaly like the one described by ALFÖLDI et al. fit onto this picture. Bearing KORIM's paper in mind, it follows that in the Pannonian Basin zones are present that are isolated from the artesian flow system.

Local geothermal anomalies are discussed in two papers. One, entitled "*A geothermal flow system in the Pannonian Basin, case history of a complex hydrogeological study at Tiszakécske*" is compiled by ALFÖLDI, ERDÉLYI, GÁLFI, KORIM and LIEBE. The anomaly is situated in an area where the base of the Upper Pannonian is somewhat over 1000 m deep. Its lower part consists mainly of impervious shales with tight sandstones. The sand sequence with some lenticular clay and marl beds of the Upper Pannonian and Levantian and the sands and gravels of the about 500 m thick Quaternary form a multilayered continuous aquifer system with excellent permeability and porosity. It contains bicarbonate waters with some 30 mg/l of chlorine in the Quaternary and up to 360 mg/l of this ion in the Pliocene, suggesting a meteoric rather than a deep origin of these waters.

The anomaly was intensively explored by thermal measurements at the surface and in boreholes. Isothermal maps were drawn and the raw data were corrected for the diffent heat conductivities of loose sand and clay, in order to arrive at the distribution of the natural heat-flow values. From residual temperature logs (the differences between measured temperatures and average geothermal gradient against depths), it was concluded, that a hydrothermal convection cell is present to a depth of about 800 m.

This was substantially confirmed by the analysis of subsurface water pressures. A cold descending flow to the S of the anomaly is heated up in the interval from surface to 800– 1000 m depth and ascends below the anomaly with a drainage toward Tisza River. The velocity was calculated at 10^{-8} m/sec.

It should be realized that such a convection cell, though with very small velocity, is possible only in an aquifer of extremely good quality.

The other example of a local geothermal anomaly is discussed by WERNER and BALKE, "*Geothermal model for the rise of deeper ground water along faults in unconsolidated rocks*". It is situated in the W part of the Federal Republic of Germany in the Erft region, some 15 km W of Cologne (Köln). The Erft region is a small part of a huge tensional rift and graben zone in W Europe and North Sea, in part of which the River Rhine found its course.

The Miocene of the Lower Rhine Embayment contains numerous lignite seams in a sand/clay sequence that is overlain by Pleistocene gravels. The lignite is produced by open-cast mining in the NE bordering fault zone. The lignite seams are about 100 to 150 m deep, and for the mining operations the ground water table had to be lowered by pumping from boreholes. It was found that in the graben the geothermal gradient is about normal, a 12–15° centigrade at 100 m depth; in the fault zone, however, temperatures are up to 25° at 100 m. The authors propose that the anomaly is related to rising warm waters along faults in unconsolidated sediments. A mathematical model was set up, using idealized geological conditions. Numerical calculations showed that the proposed mechanism is sufficiently realistic to explain the geothermal anomaly.

However, as the authors remark, why should the ground water be rising at all? They invoke hydrologists to draw up hydrodynamic models. This brings us to the problem of the hydraulic properties of faults. Accumulations of oil and gas in fault traps are well known. In the coal-mining area of Netherlands South-Limburg the compressional thrust faults of Hercynian age were found to be impervious, the tensional normal faults of Cenozoic age to be conductive. Some 40 km to the NW of the geothermal anomaly in the Erft region in a borehole, situated in the graben a short distance from the bordering fault zone, FABIAN* reported the occurrence of fresh water to a depth of almost 1200 m in marine sediments of the lower Tertiary. This indicates meteoric waters percolating to great depths, in which flow, besides the here rather sandy nature of the Tertiary sediments, the fault zone may play a role. In the Central Graben and adjoining horst in the S Netherlands, however, the fault zone separates two hydrochemical provinces. In the graben the fresh/brackish water interface is about 400 m deep near the top of a well in consolidated Oligocene clay formation. Sediments containing waters of concentrations higher than that of sea water occur below 650 m in lower Tertiary and older formations. In the adjoining, intensively faulted horst, however, in Upper Carboniferous at a depth of 1400 m concentrations smaller than sea water were encountered, indicating meteoric influences in marine sediments of lower Tertiary and upper Cretaceous age and in the uppermost part of the Carboniferous.

The hydraulic role of faults depends on local geological circumstances, and thus the hydraulic properties of the fault zone in the Erft region remain an open question. I wonder, however, whether in the Erft region

- firstly, subsurface communication exists between the Erft fault zone and the structurally high area to the E,
- or secondly, a geothermic model could be envisaged in which a blanketing action of the lignite beds is considered.

* FABIAN, H. J. (1958) – Die Aufschlussbohrung Straeten 1. F. schr. Geol. Rheinl. Westf., 1, pp. 11–28.

Organic sediments have very low thermic conductivities (about 1 to 1.5 $mcal \cdot cm^{-1} \cdot sec^{-1} \cdot deg$ against an average value of loose, water containing, sediments of about 4).

The solution of a technical inconvenience related to earth's heat is discussed by IVANOV and RAIKHMAN in their paper "*Utilisation des échangeurs de chaleur souterrains dans les puits d'eaux carboniques afin d'éviter la formation de travertin*". Due to release of pressure while producing hot underground waters carbonic acid goes out of solution, and as a result calcium carbonate precipitates clogging the tubing. The most appropriate way to maintain the contents of carbonic acid is to cool the ascending thermal water by means of tubular heat exchangers placed inside the production well. The cooling medium is fresh surface water of 6° centigrade injected either through the innermost tube or through the inner annulus (see figure in the paper). Under the circumstances prevailing in the important health resort of Djermouk in Russian Armenia cooling from around 60° to 45–50° proved to be the most advantageous: there was only little decrease of yield and no precipitation of travertine; the production by gas lift was not impaired.

This brings us to the end of the papers discussing hydrogeothermics. The next 4 papers treat problems related to the phreatic aquifer.

AGAOGLU and KARAASLAN discuss the results of pumping tests in their paper "*Mise en valeur des résultats d'étude faite dans la plaine de Bursa-Nilüfer*". The alluvial plain of Bursa, about 290 km² in area, is situated at the S coast of the Sea of Marmara in NW Turkey. Crystalline rocks of probably Precambrian age and Palaeozoic to Tertiary sediments outcrop in the mountains bordering the plain and protrude in erosional remnants through its alluvial cover. The latter is apparently thick and consists of an irregular and variable sequence of gravels, sands, silts, clays and their mixtures; the beds have widely varying hydraulic properties.

Hydrologic observations are made in 31 research installations up to a depth of 40 m. In 7 of these pumping tests were conducted. A detailed account is given of the relevant calculations in one of the installations. Depending on suppositions made, diverging figures on transmissivity, permeability and storage coefficient were obtained. It concerns a 6 m thick sandy gravel bed overlain by about 11 m of clays and silts. It appears that this cover is semipervious and allows waters from the surface to feed the gravelly aquifer.

In his paper "*Economy of irrigation water wells within unconsolidated sediments*" KARAASLAN derives a formula relating the economically optimum well distance in an irrigation network to the desired discharge, the transmissivity of the aquifer and the costs. Costs are a function of the hydraulic properties and geometry of the aquifer (which are considered constants in a particular area) and the interdependent variables discharge and well radius. Here a not uncommon, but highly complex problem of optimization is well presented.

Rate of discharge affects the radius of influence of a well and hence well spacing, i.e. the distance between wells and the number of wells to be drilled. This in turn affects drilling and transport costs and costs of pipes. The well radius is an important item on the drilling budget; it affects—with costs of pumping and energy—the rate of discharge.

A statistical treatment of the fluctuations of the ground water level has been carried out by RÉTHÁTI, "*Peculiarities in the fluctuation of the ground water level in the Great Hungarian Plain*". The plains are considered a hydrological unit, where a close, but not completely uniform, correlation is found between the fluctuations of surface and (phreatic) ground waters. In order to investigate whether a periodicity occurs, the statistical technique of autocorrelation was applied to 24 observation wells for ground water levels over a period of 34 years between 1938 and 1972. The results can have a predictive value.

In autocorrelation an entire time sequence of events is compared to itself in all possible positions by computing the autocorrelation function, the linear correlation between a time series and the same series at a later interval of time. This interval, the "lag", is the amount of offset between the two series being compared. By so doing, the degree of similarity is determined at positions of good correspondence and the amount of dissimilarity at other positions. After plotting the autocorrelation versus the lag (the autocorrelogram) the location of points of maximum correspondence may reveal an existing periodicity. The correlation may drop from a value of $+1$ to zero and even to negative.

The results obtained in the Hungarian plain indicate periodicities of 12 to 13 and of 15 years in peak values of ground water levels. It should be noted that the autocorrelation coefficients (Fig. 1) are on the low side, 0.56 and 0.6 resp., and that the time series is rather short relative to the maximum interval of 1 year. Nevertheless, the areal distribution of the periodicities suggest that they are significant, in N and S the periods being 15, in the middle part of the plain 12 to 13 years (Fig. 2).

Although there appears to be no direct correlation between fluctuations in ground water level and in rainfall, rise or fall occur in the former when for a considerable period average rainfall is higher or lower than normal. Climate influences on ground water are of relatively small extent; the high Carpathian chain has a predominant influence.

I have obtained the impression that topography, the characteristics of the streams—re- or discharging—and of the soil have an influence on the fluctuations of ground water levels and so do even conditions outside the Pannonian Basin, viz. levels in the Danube River.

The interrelations of a lower confined aquifer and the phreatic aquifer are discussed by Palausi and Polveche in their paper "*Sur un cas particulier d'alimentation d'une nappe superficielle par une nappe profonde*". The valley and coastal plain of the Gapeau river in S France near Hyères are covered by a heterogeneous sequence of young Quaternary to Recent sands, clays and loams that give rise to a complex hydrogeological picture.

The phreatic aquifer (at a depth between 0.7 and -2 m) is mainly recharged by rains and surface waters; evapotranspiration is large. Its water, produced by means of dug wells, is an important economic asset for agriculture. It is underlain by a clay bed, 8 to 10 m thick, that covers the confined aquifer. The latter is recharged by surface waters infiltrating in higher parts of the valley, i.e. in valley slopes and river bed; it has a varying but high permeability. Due to its higher pressure and due to the imperfect cover, in places it feeds the phreatic aquifer, lowering its solved solids contents.

This mechanism goes a long way explaining hydrogeological anomalies, such as variations in water table and water quality. It is envisaged that it will be possible to increase the capacity of the phreatic aquifer recharging it from the confined aquifer by means of boreholes some 12 m deep.

From the phreatic aquifer to those, deep in the sedimentary basin, that are suitable for the disposal of waste is quite a long step. Two papers are presented on this subject, namely underground discharge of industrial waste for permanent disposal and underground storage of warm water for later recovery. Both papers deal with liquids to be injected in porous/permeable reservoirs; the use of rock salt caverns is not considered here.

Disposal or storage should be done only in aquifers that are isolated by under- and overlying impervious beds. Contamination of useful ground waters and of other natural resources must be avoided. This means discharge into beds that are isolated from the present ground water flow, where the interstitial waters are stagnant and saline*.

Any introduction into an aquifer, whether a liquid toxic waste or warm and fresh water, is apt to have an effect on the quality of the aquifer. Swelling of clay particles and precipitation of solids may clog screen and aquifer; solution of other minerals may open undesirable ways of escape.

The injection of waste induces flow in the aquifer, and it depends on the differences in viscosity and specific gravity of interstitial waters and liquid waste what the ultimate shape of the body of waste and its migration distance will be.

In the first paper on underground waste disposal to be discussed, that by PITIEVA, KOVALEVSKAYA, ALSHINSKY, ORLOV and VITVITZKY entitled "*On the possibility of using artesian basin deep-seated horizon for industrial discharge dumping*", solution is given for a most important hydrogeochemical problem; the degree of compatibility between waste, interstitial water and sediment and its influence on the aquifer and hence on flow rate and ultimate distribution of the waste in the aquifer. Laboratory experiments under high pressures and temperatures (simulating reservoirs between 1000 and 2500 m deep) were carried out, a difficult but not impossible task. Results were compared with those of experiments under laboratory conditions, and it appeared that for different types of waste prognoses could be made of the interactions under varying reservoir conditions.

In order to arrive at successful and safe disposal operations, the authors propose a prognostic combination of hydrological data and chemical factors of the waters. Provinces should be established based on hydrogeochemical characteristics, within which province areas should be singled out that are characterized by ion composition, concentration and pH of the wastes. Within the areas subareas are to be established on lithologic and hydraulic factors.

This end is to be reached by regional investigations in order to find thickness and extent and depths of the aquifers, their reservoir properties, the composition of the interstitial waters and the quality of the impervious beds. In the second stage of investigation the compatibility of waste, water and rock and the changes in the aquifer affecting flow and distribution of waste in the aquifer should be studied.

It is emphasized that highly toxic wastes only, that cannot be purified technologically or at a profit, should be disposed of underground.

In underground disposal of toxic wastes economic considerations are of subordinate importance only; costs should not be a deterrent. This is different in underground storage and recovery of warm waste water from industrial

* VISSER, W. A. (1974) — Prevention of ground water pollution in subsurface waste disposal, I.A.H. Memoirs tome X-1, p. 153—158, Montpellier, France.

VISSER, W. A. (1974) — Waste disposal and underground waters, Geol. Mijnb., vol. 53 (6), p. 249—256.

and public utility plants. In his paper "*Hydrogeological considerations on underground storage of low-caloric energy*", VISSER states that in such a project the energy balance should be positive and that it should be economically viable. It may provide a contribution to the conservation of energy and to reduction of thermal pollution.

The fresh and warm waste water is introduced into a saline and relatively cool environment. As its viscosity and specific gravity are lower than those of the native waters, the waste water tends to move freely through the aquifer and in upward direction. This movement is greatly influenced by clay streaks and inhomogeneities in the reservoir bed. This will cause irregularities in the waste front and "fingering" and, in recovery, loss of water and temperature. Moreover, physical-chemical reactions, impairing permeability, will restrict the life of a storage and recovery project.

A short while ago the results of an experimental project in the Federal Republic of Germany (to the N of Krefeld) were brought to my attention.* Fresh water was extracted from a confined aquifer between 11 and 29 m deep, heated to 45 °C at the surface and infiltrated in the phreatic aquifer between 1 and 4 m depth. If waste heat would be provided free of cost, such a project would be economically feasible. In my opinion, however, storage in the phreatic aquifer (or in any fresh water-bearing stratum) must not be done. The photograph in the paper by KLEY and NIESKENS, showing molten snow around the infiltration well, indicates that considerable ecological harm could result.

In a confined aquifer space can be created only by injection under pressure. If, however, the pore pressure is increased, the cohesion between the rock particles will be impaired, and eventually cracks will be formed. This hydraulic fracturing is a useful stimulation technique in the production of oil, it should be avoided in underground waste disposal. An injection pressure not higher than about one third of the overburden pressure (weight of rock particles plus interstitial water, as an average 0.23 atm per meter depth) appears to be safe.

The injection pressure causes the interstitial waters and the waste to be compressed, as are the rock particles, whereas the formation will dilate. These are elastic effects, which are quite small. Water is compressible by a factor of $5 \cdot 10^{-5}$ per atmosphere pressure, the combined effects on the rock depend on depth: at 500 m about $8 \cdot 10^{-5}$, at 1000 m about $5 \cdot 10^{-5}$ per atmosphere. The restriction on the injection pressure and the smallness of the elastic effects severely limit the amount of liquid that can be disposed of underground.

Elastic effects cause the water drive in petroleum production, a very effective set of forces driving the oil towards and up the producing well. Their role in pumping water from subsurface aquifers was realized by MEISSNER about 50 years ago, water being forced out by compaction of aquifer and surrounding aquitards and by expansion of the water itself.

It appears that in ground water production, apart from elastic effects, in case of imperfectly impervious confining beds forced communication between aquifers could be established after prolonged pumping. This is to a large extent the theme of the paper by IASVIN, BOREVSKI and PLOTNIKOV, entitled "*Lois de formation des réserves exploitables des eaux souterraines des bassins artésiens du type de plate-forme et quelques particularités de leur évaluation*".

* KLEY, W. and NIESKENS, H. G. (1975) — Möglichkeiten der Wärmespeicherung in einem Porengrundwasserleiter und technische Probleme bei einer Rückgewinnung der Energie. Z. dt. geol. Ges., 126, p. 397—409, Hannover.

In large epicontinental basins the authors distinguish two types of aquifers. The first type does not extend to the edges of the basin, it is well isolated and deep (100 to 300 m). The exploitable reserves are due to elastic effects only. The cones of depression are large (radii of 100 km), as are draw-downs (60–80 m or more). The regime is nonpermanent: reserves are depletable. The yield is small relative to the large volume.

The second type of aquifer in large basins occurs as a rule at smaller depths; the cones of depression are smaller; the depletion is slower and the regime is nonpermanent too. The exploitable reserves are replenished to a large extent by leakage from other aquifers through the less impervious parts of the aquitards ("hydrogeological" or "lithological windows" – permeability windows); elastic effects are of subordinate importance only.

Finally a third type of aquifer is distinguished that occurs in small basins and that reaches the edges of the basin. The reserves are constantly being built-up by infiltration from the surface and by leakage from aquitards.

The communication between various aquifers can be demonstrated by radioactive carbon measurements, by hydrogeothermic and hydrochemical surveys and by artificial tracers. The interrelation between aquifers and aquitards in any basin is a complex problem. For its solution–that is the evaluation of the reserves–mathematical simulation methods are indispensable.

To be able, however, to apply mathematical simulation methods a physical model that accurately reflects hydrogeological conditions has to be prepared first. An example is provided by SEMENOVA, LEVI, VODOVATOVA, EFREMOV, PECHERIN, KLYUKIN and FREYDINA in their paper "*Methods of regional estimation of subsurface water resources of multilayered artesian basins*". The authors report on the investigations carried out in the Terek-Kuma basin, situated in the European part of the U.S.S.R. along the central Caspian Sea up to about 250 km inland. The basin, of an area of about 100,000 km^2, is filled with Mesozoic to Cenozoic sandy and clayey deposits.

In order to establish a satisfactory physical model, apart from a thorough knowledge of the geological development, a subdivision of the sedimentary sequence based on hydrogeological properties is needed. The basic information was obtained by drilling and geophysical borehole measurements, from which a regional, basinwide correlation was obtained.

As a negligeable part of the aquifers was tested hydrogeologically, geophysical methods were looked for to obtain the needed data on transmissibilities. A correlation between the coefficient of permeability and the relative clay content derived from radioactivity logs was found. The correlation coefficients are low: 0.6 to 0.8.

It should be remarked here that analysis of natural radioactivity in conjunction with borehole diameter is a powerful tool to derive the clay contents of sand beds and to establish the depths of the sand/clay interfaces. Permeability, however, depends also on other factors, such as grain size and sorting. The research on the practicability of combinations of various borehole logs to find permeability is by no means completed. As yet the only accurate tools are pumping tests and drilling-core analysis. As this may be, the method used is a good approximation to distinguish between relatively permeable and impervious beds and to find the in the paper described trends of distribution.

In the Neogene to Quaternary sequence, that dips and thickens toward the Caspian Sea, up to a depth of almost 700 m seven water bearing complexes, consisting predominantly of variously grained sands and sandstones intercalated by clays, are separated by relatively impermeable complexes of clayey sediments of regional extent (Fig. 1).

The distribution of the piezometric heads shows some pecularities In the W and S basin margins heads decrease toward depth ("direct relationship", see the author's map, Fig. 2); in the N, S (Terek River) and E parts and in the middle course of Kuma River heads increase toward depth ("reverse relationship"). In the main part of the basin heads decrease to a certain depth and increase further down ("complex relationship"). Analysis of the distribution of piezometric heads enabled the authors to distinguish between areas of recharge and discharge and to establish the general hydraulic relationships and flows within the basin. This was confirmed by thermometry.

A further study was made on the role of the rivers. It was found that they recharge the ground waters in their upper reaches only, where they leave the mountains to enter the piedmontane area (question: is this in the area of direct piezometric relationship, piezometric heads decreasing with depth?). The rivers discharge the ground water in their middle courses (areas of reverse and complex relationships?), whereas still lower down, in the delta, surface run-off is balanced by evapotranspiration.

It is, however, difficult from the presented information to visualize what is actually happening. There appears to be a general flow from intake areas in the high terrain toward the zone of low heads in the area of complex relationship and hence to a discharge area in the lower Terek River. The distribution of salinities, Fig. 3, suggests a replacement to have taken place of originally saline, native waters by fresh artesian water. The same figure shows a zone of high salinity at and near the surface, salinities decreasing to fresh at a depth of about 150 m. What is its origin? Is it entirely due to evapotranspiration or is it a remnant of infiltration by a Pleistocene Caspian Sea?

The hydrogeology of the Terek-Kuma basin is complex and presents many difficulties of interpretation, to whose solution a successful attempt, involving many disciplines, has been made.

A variety of natural resources that may all be present in one and the same sedimentary basin has been reviewed—water, geothermics and disposal of waste. Some reference has been made to hydrocarbons and other organic deposits, to rock salt and ores. To exploit these resources to their full extent and to reconcile often conflicting interests, the basin in its totality has to be considered. That is to say a thourough knowledge is required of its geology and its history—stratigraphy, lithology, structure—for a successful, safe and economic development of its resources. In view of threatening shortages, f.i. of water, energy, metals, there is a challenge to earth scientists of many disciplines: geology, geophysics, geochemistry, hydrology, mining- and petroleum engineerings. Collaboration and mutual understanding are required to accomplish an ultimately rewarding task.

GENERAL REPORT 2

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The topic to be discussed here is a very complex one already in its title. This is reflected also by the reports received. 14 reports have arrived in this topic and almost each of the items listed in the title of the topic is treated in the reports.

As the subjects of the reports spread over a large area and the subjects are being approached from several aspects, it can hardly be drawn any general conclusion from the reports.

In spite of the difficulties I have tried to comprise the reports into certain groups following the subjects listed in the title of the topic, as far as possible. Another endeavour has been to progress from reports giving some general considerations through reports dealing with the description of areas investigated to the reports that give suggestions for actual technical solutions or experiences in connection with them.

The first report to be discussed is that of IASVIN, BOREVSKI and PLOTNIKOV, which gives the most comprehensive outline of the nature of artesian basins. When the authors are classifying the water resources of such basins into certain main groups, the study itself points out that there are many transitions between them. It is emphasized that though it is possible to establish the main characteristics of an artesian basin or of a region by determining the properties of the single layers that are building up a structure, these data do not give a reliable picture of the behaviour of the structure as a whole.

The study reports some cases when data of the geological structure itself would justify that only the elastic resources of some single layers would play the main role in the development of the utilizable water resources; however, observations on the actual exploitation show that there is a close interaction between seemingly isolated aquifers through hydrogeological or petrografical windows.

The authors are strengthening the conviction of hydrogeologists, referring to that during the preparation of actual water utilizations the investigations must be extended over the whole structure and to a large distance; that instead of analytical methods only the integrated utilization of mathematical modelling of natural radioisotopes and of long-term operational observations can give a comprehensive picture of the processes to be expected.

One of the most important statement of the report is that even the long-term pumping tests do not give a reliable picture of the behaviour of the subsurface system: as the processes induced by the actual operational exploitation are developing in a longer time than that of a pumping test. This gives a valuable support to us when stating that the economic aspects of the preliminary studies should be dealt with in a cautious way as being complex and comprehensive; and that the observation of operating water works is an indispensable tool of making conclusions on the behaviour of artesian basins when exploiting their water resources.

After this in-principle discussion SEMENOVA and her companions present an actual, comprehensive and large-scale study of an artesian basin in the Soviet Union. The area of this basin is approximatively the same as that of Hungary and this is a reason why the methods and conclusions of this investigation will be interesting for the Hungarian hydrogeologists and for hydrogeologist from countries of similar hydrogeological conditions. The other reason why we Hungarian hydrogeologists are interested in this study is constituted by an entirely similar approach that was applied also in Hungary in the past decade when compiling the Hydrogeological Atlas of Hungary and, later, in investigating smaller areas of the country.

Another investigated indicator was the vertical and horizontal change in the pressure head measurable in water wells in the area. The authors are distinguishing areas where the piezometric head in wells is increasing as going downwards to greater depths; areas where this tendency is of adverse character and the areas showing a combined tendency. Vertically increasing head is explained as an indicator of vertical upward migration of water while the adverse change indicates downward migration. They are investigating in their report the ratio of this vertical gradient of piezometric head to the horizontal one. This may be in the authors' opinion a tool of discerning areas where deeper aquifers are supplying the upper ones from areas where the direction of movement is directed downwards, therefore these are the areas of recharge of artesian water-bearing complexes. These statements are confirmed by studies into the thermal and chemical conditions of the underground waters of the area.

Without wanting to provoke any discussion with the authors, it may be noted that a vertical gradient in a multilayered complex is not always an indication of actual flow or migration of water: it may indicate in some cases only the possibility of such movement. Another thing that I should like to mention is that both the horizontal and vertical piezometric gradient are components of forces effecting on water and the resultant force is that influences the water movement according to the ratio of horizontal and vertical permeability of the complex.

Returning to the report of SEMENOVA and others: it is mentioned there that above the heads and gradients they have taken into consideration also the results of investigations with stable radioisotopes when analyzing the hydrogeological conditions of a large artesian basin in the Soviet Union. It is J. DEÁK, Hungarian specialist, who discusses in details the kinds of investigations that were carried out in Hungary with environmental isotopes directed towards the large-scale flow conditions of the Great Hungarian Plain as an artesian basin.

Former studies conclude to a similar situation in Hungary as in the case of SEMENOVA's report. In certain areas a vertical upward gradient can be experienced in the Great Plain while in other areas this gradient has a reverse direction. DEÁK has investigated whether it is any indication of an existing water movement investigating the ^{14}C and tritium content of subsurface waters in these areas.

An area has been chosen where the gradient in wells has an upward direction and the question was if is there any migration or flow according to this head? If yes, then the deeper waters should be the younger.

Samples has been taken from wells perforated in the 390 – 405 m interval and in the interval of 74 – 87 m; the ^{14}C content, the $^{13}C/^{12}C$ ratio of the samples has been determined according to the formulae taken from the international literature and the age of the waters coming from the two horizons has been determined. The age of the deeper waters has been found $10{,}400 \pm 2000$ years, while the age of the shallow waters has been $14{,}500 \pm 1000$ years. This suggests that there is in fact a movement with upward direction.

If it is so, the uppermost groundwater must have a detectable ^{14}C content or this water body must be, at least in part, older than groundwaters that have been infiltrated in the past 20 years. By analyzing the tritium and deuterium content of these waters it can be established whether they have infiltrated since the first thermonuclear bomb-experiments or they are older than those. The results of the investigations have shown that the uppermost groundwater is the youngest one and no indication could be found which would support the assumption that in this water body the waters coming from the deeper complexes and the water infiltrated from precipitation are being mixed.

If there is an upward-movement from the deeper layers with no mixing in the uppermost groundwater then there should be a tapping layer between them. By investigating profoundly the heads it has been found that there is a sandy layer in the 20 – 60 m depth interval the head of which is lower than that of the layers under and above it. This may be a layer that is in a connection with the Tisza river and which leads the mixed waters of the deeper and higher layers into this river. This has been in part demonstrated by samples taken from the Tisza river at its low flow.

These investigations were started in 1974. The results reported have been based only on a few samples and thus the quantity of data does not permit to draw any quantitative conclusion. Anyhow, the results are interesting and serve as a good example when different ways of investigation are supporting the same assumption.

The next report to be discussed belongs already to a different theme: this is the dumping of liquid materials into underground complexes. The subject of PITIEVA and her companions is, however, not the technical solution but the considerations in connection with the problem in general and the viewpoints to be taken into account when surveying the possibility of dumping.

PITIEVA's subject is the dumping of industrial waste-waters. The authors distinguish different kinds of interacting systems: such as the industrial discharge-rock system, the industrial discharge-underground water and the industrial discharge-underground water-rock system as getting farther from the site of dumping.

Before giving a picture of the stages in which the dumping of such materials can be planned and prepared, they are making a few establishment which may be of general interest. These are:

— The mixing of given industrial discharges and groundwaters of different composition brings about processes that can be predicted;

— There is an additional factor: the composition of the water-bearing material;

— The high temperature and pressure prevailing in deep-seated layers influences the interaction of industrial discharge, subsurface water and rock so that certain processes are quite different from those under athmospheric conditions.

The consequence of these may be that in certain cases the permeability (and receptivity) of a layer can increase as the waste water leaches it; in another context the permeability decreases and a solid precipitation falls out of the solution as a result of the temperature and pressure given.

In spite of the many variations, a well-planned dumping of industrial water is feasible by making successively the necessary preparatory steps. It is possible also to forecast the possibilities for dumping such materials and so, stage by stage, arrive at the actual realization:

— So it is possible to make prognostic charts displaying the consequences of mixing various industrial waste waters with the given subsurface water and rock according to their composition;

— Being aware of the various possible interactions of industrial discharges, subsurface waters and rocks, the prognostic regioning of an area becomes possible.

The main conclusion to be drawn from this report is that in spite of the complex interactions the load capacity of even subsurface waters may be determined. The dumping of waters is perhaps not permittable everywhere because of certain severe preventive prescriptions, however, it can be believed that this way of thinking can be utilized when aiming at the solution of spreading the various pollutions in subsurface waters.

VISSER's report deals with an analogue problem: this is the heat storage in subsurface complexes. It calls the attention to the possibility of storing the hot used water of power plants in saturated loose sediments or in caverns with the purpose of recovering the stored heat to meet peak demands. It is discussed also here that the injection of any material into subsurface water will impair the balance between the composition of native interstitial water and the aquifer. The waste water is hot and has a composition quite different from that of the native water. Various interactions that may cause the decay of permeability, are provoked. Besides there is the problem of heat loss: several cycles of injection and retrieval must follow each other until the environment of the injection point is being warmed, according to approximative calculations the temperature of the injected water will decrease by some 25 – 30 °C during the first injection-retrieval cycle. This decrease will be reduced after several cycles, however, it must be kept in mind that during retrieval cold water always occupies the place of the recovered hot one.

It must be taken into consideration that the injected hot water will float above the native subsurface water; and even more it will by-pass the latter;

sometimes in the form of a palm. This causes that the contact surface will increase causing an increase in heat loss.

There are also economic aspects that must be taken into account: the storage of hot water at a larger depth is more economical as the larger the depth the smaller the difference between the temperature of the injected and the native water; on the other hand drilling is then more costly and also the energy demand of such a project is larger. There are also other aspects that must be taken into consideration: e.g. the selected aquifer must be well-isolated from aquifers of good-quality drinking-water; the injection-pressure must be chosen so as not to harm the context of the aquifer or even the over- and underlying layers: the recovered water must not be corrisive and so on.

The report has an impressive actuality for the hydrogeologists of certain countries. In Hungary, for example, the valuable geothermal energy is utilized in such a manner that the thermal water of 35 – 100 °C temperature is used in a maximum two-stage heat-exchanger system and after having exploited its heat content the water is discharged into the surface waters. The problems dealt with by Visser and his approach has a spirit quite differing from this manner which seems to be something luxurious in the light of such an energy-saving thinking.

With this report we have arrived at the field of the geothermal energy utilization.

K. Korim, Hungarian geologist, the author of the next report discusses the thermal water condition of Hungary. His report gives a general description of the Hungarian Middle Pliocene aquifers and thermal waters originating from them. The reader gets a general picture of the various loose formations lying at a depth of 2000 – 2500 m, having a temperature of about 100 °C or more. It can be learned from the study that in the Hungarian Basin the loose sediments of sands and sandstones are containing a considerable amount of thermal water in formations of coalescent, lens-like or sandwich-like shape. These layers lie upon the crystalline base-rocks. The pressure conditions of these waters are of hydrostatic nature. In spite of this fact the majority of the thermal water flows out of the wells without pumping, due to their temperature and gas content.

The yield of the more than 500 thermal wells is about 1500 – 2000 l/min/well. In several cases the wells are tapping more than one aquifer and the majority of them must be screened in a profound way.

There are no satisfactory data available on the origin and recharge of these waters and some phenomena are indicating that they are not of a renewable nature. This suggests to the Hungarian geologists that they be careful when exploiting these waters, in spite of the fact that these waters can be reached in a relatively simple way.

I would return again to the point mentioned in connection with Visser's report: it is a general temptation to explore as much thermal water as possible depending on the financial limits only. However, it involves the danger to withdraw more water than is recharged from any supply source. The other danger is that water and its heat content is not utilized entirely e.g. in several stages. It can be easily understood that this is not the most economical way to utilize geothermal energy; it is cheap only.

AIRINEI and PRICAJAN of Romania are reporting on the geothermic or better to say: hydrogeothermal conditions prevailing in the western territory of Romania. Both the territory discussed and the topic of the report is in a close connection with those of K. KORIM, the author of the report discussed just before. The area discussed, its geological structure is in many respect the same as that of Hungary since only the national border separates them. On the other hand this is the territory by which the Hungarian Basin is being finished eastwards, therefore this area has some characteristic features differing from the properties of the Hungarian Plain.

The structure of this territory may be characterized by the presence of the crystalline basement at the lower structural layer and the sedimentary formations at the upper ones. The crystalline basement represents the westward continuation of the mountains forming the border of the Pannonian Basin and they are occupying ever lower position getting westwardly towards the centre of the Hungarian Plain. The upper formations are of Mesozoic, Tertiary and Quaternary ages.

The authors are convinced that the geothermal phenomena of the territory can be explained only through the knowledge of the interdependencies of the various factors governing the geothermal processes. Their objective is therefore to give as a comprehensive picture of the total complexity of geothermic conditions prevailing in this part of the Earth's crust, as possible on the basis of data available in a not too large number.

The knowledge of the structural conditions is based on gravimetric and magnetometric investigations. According to this knowledge and in the author's opinion the fault system of the territory may be classified into three groups: crustal fracture fields, regional fractures and local fractures. These interdepending systems are determining the circulation of subsurface water and are giving explanation to their thermalization together with such additional phenomena as radioactive processes below the surface. The territory investigated is the fleck of the regional geothermal anomaly of Central Europe, where the geothermal gradient is 18.5 m/1 °C instead of the value of 35 m/1 °C in the territories surrounding it.

The petrographic inhomogeneity, the variations in the thickness of layers forming here the crust induce the distribution of geothermal irregularities. One of the characteristic features is e.g. the extensive system of faults that may give a certain confinement to water-bearing layers and creates connection between different horizons. This is an explanation to the phenomenon called "geothermal inversion" when one can found in the discussed territory situations where the temperature of subsurface water is decreasing when going into greater depths. It may be mentioned that similar situations can be found in such fractured zones as here, in Budapest.

The crustal fracture fields are giving the possibility to the circulating subsurface water to take up heat from masses of good heat conductivity and to bring over to other complexes. When geothermal anomalies of the structure are meeting possibilities of accumulation of subsurface water, the result is the hydrogeothermal anomaly. The process is complemented in certain cases by some postvolcanic activity that gives a significant carbon-dioxide content to water.

ALFÖLDI and his companions – Hungarian hydrogeologists – investigated a characteristic geothermal anomaly in Hungary. Their report summarizes the results of these investigations and it is a good example of how to utilize all the data available when wanting to study an area without too much new and costly exploration.

The subject of the investigations is an area of an extension of about 100 km² in the Great Hungarian Plain near Tiszakécske. The assumed crystalline basement is covered here by the various loose marine deposits of the Pliocene. The proportion of clayey layers in the sequences drilled by wells in the area varies from 20% to 60%. The maximum temperature of water flowing from drilled wells is 64 °C at a yield of 400 – 2400 l/min and from the depth of 830 – 2048 m. There are 13 thermal wells in this area.

Beyond the purpose of determining the general hydrogeological and geothermal conditions of the area, the aim of investigation was also to answer the question: what is the origin of geothermal heat and what is the flow system prevailing in the area?

With this purpose an exploratory work was carried out. The main means of exploration were shallow drilled wells with a depth of 53 m in which temperature was measured in every 5 m below the depth of 20 m. After these wells the utilizable data of the 13 deep wells were also evaluated. Also the chemical composition of thermal water and its distribution was taken into account, and all these data were plotted on maps, in sections and graphs.

The main results of the investigations are as follows:

– The geothermal gradient varies from the value of 0.3 to 1.2 °C/10 m; it means 30 m/1 °C – 8 m/1 °C;

– The geothermal gradient shows a remarkable correlation with the clayey-proportion of the sequence;

– The heat production cannot be attributed to a larger rockmass of higher elevation; the main heat-production happens in an intermediate depth of max. 1000 m below which depth the geothermical gradient displays a normal value;

– Therefore the cause of heat production is the rising of hot water from greater depths along structural lines; and areas where water descends and areas where it ascends warming up the various formations, can be distinguished;

– As in the case of radiocarbon and tritium investigations mentioned above, also in this territory appears the possibility that there would be a layer near the surface that collects waters and gives them the possibility of flowing into the Tisza River.

As in most problems of science, there are common problems to be answered, and it can be often experienced that an approach applied in the solution of a certain problem can be utilized in another. This is the case when one reads the report of ALFÖLDI and WERNER one after the other.

The impressive report of WERNER and BALKE describes an idealized and simplified model of a given geothermal anomaly near the town of Köln.

The authors of the former report are arriving at the conclusion that the geothermal anomaly at Tiszakécske is to be attributed to a regional flow system in which subsurface water has its descending and ascending zones, and that regional fault systems are playing a basic role in promoting such a flow system. WERNER and BALKE are assuming that the anomaly may be modelled by blocks of higher elevation and limited by faults. As they observed the higher temperatures to be connected with faults, they assume in their model that warm water flowing upwards in the conductive faults heats the blocks up to the temperature of the anomaly.

Purely through approximative thermic calculations they arrive at the result that in the assumed stationary system the time necessary to warm the blocks of 1 km wide up to their present-day temperature constitutes about 20,000 years i.e. a very short time as compared to their geological age. The width of the faults should be between 2 and 10 m and the flow velocity of the ascending water between 3.5 – 36 cm/day. The discharge of the faults may be 12 m^3/year at 1 m width of the fault. It must be assumed that the faults are filled up by unconsolidated material.

Furthermore the authors are assuming that there is no movement in the blocks but they are filled with water.

The authors did not want more than to investigate whether the results of such a model are in contradiction to the hydrogeological rules. They have found that considering the values listed before there is no such a contradiction, at least not from a theoretical point of view. It should be mentioned that there are only a few reliable measurements in the area and the whole rough calculation is based on the rules of thermal conductivity.

One question remains open: what is the reason that forces the water to flow upwards along faults? The authors do not strive with answering it. Whithout any justification, it arouses the idea of putting up the question whether a continuous consolidation of the underlying deeper loose layers could force the water to be released from the aquifers being consolidated and risen along faults of higher vertical conductivity.

Remembering the statements made by Airinei and Alföldi having pointed out the significance of regional and local fractures, the approach of Werner and Balke should be appreciated in spite of its simplifications.

Also in some of the former reports it has appeared the possibility that deeper artesian waters can recharge the upper water bodies when circumstances can promote this process. Such an example is treated by Palausi and Polveche. The investigated territory is a part of the lower plain of the River Gapeau in France. Here is an alluvial upper unconfined groundwater below which follows an impermeable layer and after it an aquifer of good conductivity containing water under pressure. The upper water body has a considerable salt content because of the intensive evapotranspiration, while water in the confined lower stratum is, in general, fresh water. There may be experienced some anomalies in the salt concentration of the upper groundwater which was attributed to the eventual connections with the deeper fresh water. Through drilling a well it has been proved that there is a possibility that this upper groundwater of high salinity be recharged from the deeper aquifer, and the authors see some economic way of doing so to develop the utilizable water resources.

Réthati, in the next report, aimes at giving a general picture of the groundwater fluctuations in the Great Hungarian Plain. The report points out a general period of 7 years (from the minimum to the maximum level) and some differences between the southern and the northern parts of the Plain. The report states that the groundwater fluctuation is influenced by precipitation and the regime of the neighbouring water courses, so the groundwater regime is not clearly of basin-nature.

An actual investigation is described in AGAOGLU's report. The task was to determine the parameters of the aquifers in the Bursa-Nilüfer Plain. In this area of 290 km², under a semipermeable clay, there are water-bearing formations. The recharge of these is coming both from the surface and from the deeper layers.

The area was investigated by 7 exploration wells. The report describes the wells to be investigated by the JACOB – THEIS method, by THEIS's standard curve, the HANTUSH formula and according to DE GLEE. By these methods the transmissivity, the storage coefficient and other parameters of flow were determined, compared and averaged, securing so the basis for further planning activity.

KARAASLAN in his short report deals with the economy of pumping water from below the surface by wells, with the purpose of irrigation. In his method the hydraulic characteristics of the pumping activity and the hydrogeological parameters of the aquifers are combined. His endeavour is to determine the exploitation cost taking into account the hydrogeological conditions, the hydraulic losses, the cost of maintenance, depreciation, the interest etc., and he optimized the function obtained. The result is the optimum distance of wells when taking into consideration also their mutual influence.

A technical solution applied to the utilization of thermal waters is reported by IVANOV and RAIKHMAN in the last report of our topics. The problem is well known: in a number of cases certain kinds of thermal water of high carbon-dioxide contents are inclined to cause considerable precipitation of calcium carbonate. The phenomenon is connected with the fact that as the pressure of the rising water in the well-tube is decreasing, a certain amount of CO_2 is being released and the equilibrium of CO_2 and $CaCO_3$ is upset. The preservation of the equilibrium might be given when thermal water is being cooled before precipitation occurs. The authors report on experimental results of heat-exchangers.

The principle of heat exchange is that thermal water is being cooled below the surface by cold fresh water. Two types of exchangers were tested, both of them built up according to the same principle. The principle is that more concentric tubes are located around the tube of the well and by this way thermal water flows as surrounded by flowing cold water. During this process thermal water is being cooled, while the temperature of cooling water increased. This causes that the CO_2 content of the thermal water does not decrease, so does not occur the precipitation of lime in the well tube.

The length of the exchangers tested was 37 m and the original temperature of thermal water 55 °C. The thermal water was cooled to a temperature of 35 °C and the original yield of the well decreased by 17%. There is a critical temperature at which the operation of the so-called gas-lift ceases. This temperature in the case study was 25 °C.

The result was fairly satisfactory: during an operational period of 2 years there was no necessity of removing "travertine" from the well, while in the neighbouring well, without heat exchange, the removal of "travertine" was necessary in every 3 – 4 months.

It should be noted that this solution is applicable only in cases of no great significance when the temperature of the thermal water remains unchanged. In certain cases temperature is an important factor and heat exchange is not an admissible process. For certain cases such processes were developed, e.g. in Hungary, with the addition of some acid to water flowing in the well. The results are satisfying, each case and each well needs however a uniform design and experimental operation.

I should like to hope that this short run could reflect the colourful picture that is formed by the reports received. I think, the conclusions of the reports corroborate our opinion: hydrogeology is a complex field of sciences, and when intending to solve an actual technical problem, investigating a given area or evaluating the utilizable water resources, one must do his work in a complex and integrated way. For determining a subsurface flow-system, we have to explore the structural conditions, to utilize the phenomena of radioactivity, to take into consideration the chemical properties of waters and rocks, and we are obliged to use mathematical or analogue models, computers, etc.

A rich store of means and tools is at the disposal of hydrogeologists. I think that to be up to date in our thinking and in our methods means that the hydrogeologist has to utilize these means and tools always in the most suitable proportion but always with a very wide spectrum of thinking. I hope that both the reports received and their present summary contributes to this up to date thinking and work.

A CONCEPTUAL PATTERN FOR THE COMPLEX INVESTIGATION OF THERMAL PHENOMENA IN THE UNDERGROUND WATERS OF ROMANIA'S WESTERN PLAIN

RÉSULTATS PRÉLIMINAIRES DES RECHERCHES CONCERNANT LE PROCESSUS DE LA THERMALITÉ DES EAUX SOUTERRAINES DE LA PLAINE OUEST DE LA ROUMANIE

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RÉSUMÉ

Les auteurs présentent les résultats préliminaires de leur recherche concernant le processus de la thermalité des eaux souterraines de la Plaine Ouest de la Roumanie.

A cette étape, on a élaboré un modèle conceptuel nécessaire pour poursuivre ce phénomène pendant les étapes futures tant sur le territoire étudié que sur d'autre régions du pays. On part de la postulation que l'énergie thermique de l'intérieur de la croûte terrestre est une somme composée de la chaleur interne de la planète migrée continuellement vers l'extérieur et de la chaleur résultée de la désintégration radioactive. Les facteurs physiques et géologiques qui intéressent l'installation de la distribution non uniforme de l'énergie thermique, dans l'intérieur des volumes crustaux situées au même niveau de profondeur, sont la propriété thermique des roches et leur contenu en substances radioactives. La distribution hétérogène de ces facteurs est en fonction de la composition et la structure majeure de l'écorce terrestre et du système des champs de fractures crustales y compris leurs ramifications arborescents vers la surface. Les éléments géologiques profonds sont reflétés par les anomalies gravimétriques et magnétiques ainsi que par les résultats du sondage sismique profond, tandis que ceux qui sont rapprochés de la surface sont reflétés par la sismique et l'électrométrie. La distribution du champs thermique aux différents niveaux de profondeur a été déchiffrée par la thermométrie des sondes existentes, tandis que la radioactivité de certaines eaux souterraines indique les premières informations sur l'existances radioactives des roches lavées par celles-ci. L'intégration des données mentionnées a conduit les auteurs au modèle conceptuel décrit qui ouvre de nombreuses voies vers de nouvelles études.

Introduction

The Western Plain of Romania, situated on the eastern flank of the Pannonian Depression, lies between the western frontier and the Volcanic, Apuseni and Banat Mountains to the north, east an south. The surface geology in the territory of the Western Plain is monotonous; the older geological formations are entirely covered by alluvial and Quaternary deposits. From a tectonical viewpoint, the geological formations belong to two structural layers: the upper structural layer of sedimentary rocks, and the lower structural layer constituted by the crystalline basement (I. DUMITRESCU et al., 1962). The sedimentary group is represented by unequally distributed and still incompletely known Mesozoic, Tertiary and Quaternary formations. The Pre-Mesozoic basement represents the westward continuation of the morphostructures in the Southern Carpathians and the Apuseni Mts, and even in lower positions. Partitioning of the basement into blocks by horsts and grabens had taken place to such an extent that thereby the thickness of the sedimentary cover ranges from several hundred meters to several thousand meters.

The authors are met with the difficulties of knowledge of the thermic processes taking place in the interior of the planet and even in the lithosphere, due to the generally interpretational character of the available data, which in their turn are not at all numerous regarding the distribution of heat values with the depth and the pertinent factors of differentiation from place to place, transfer mechanisms of the heat passing from deeper zones towards the surface, etc. The authors are convinced that an improvement in this respect can be achieved by investigating the geothermal phenomenon through the ensemble of interdependences of the terrestrial phenomena, those manifested in the planet's lithosphere, respectively. It is well known that the categories of terrestrial physical phenomena are indicative enough of the geological situation prevailing in the domain of the lithosphere or in another parts of the crust. The integration of the categories of the respective significant geophysical and geological signals, results in a complex picture. Thereupon the understanding of relationships of the geothermal phenomena, after due analyses, might put the matter in new light.

The physical-geological elements considered

The picture of the deep geological structure, as far as we know it with respect to our particular interest, has been achieved in a gravimetric, magnetometric and seismic way. As for the upper structural layer with the zone of passage to the lower one, many informations were brought in by electrometry, thermometry and radiometry.

1. *Gravity and magnetic anomalies.* Both types of anomalies bring qualitative information on the degree of lithological heterogeneity of the region's subsoil, on the basement faults and those deeper in the Earth's crust, as well as on structures suitable for underground water accumulation. We would emphasize the following relevant features:

The distribution of the gravity anomaly (Bouguer) shows a maximum area with values over 15 mgal, oriented N-S and centred on the State boundary, indicated by a deep structure of the Earth's crust, which is remarkably thinned here. Moreover, several local maxima and minima of ± 15 mgal have been resulted by the crystalline basement partition.

The magnetic anomaly surveyed is characterized by several local maxima (six out of them with larger extensions and intensities: Carei, Secuieni, Oradea, Chișineu-Criș, Arad, Timișoara), reflecting on the effects of intrusive masses aligned in the directions of Neogene eruptions in the Apuseni Mountains.

2. *Seismic and electrometric data.* Seismic informations are grouped into two categories. In the first one, there are data referring to the structure of the sedimentary cover and its relations with the crystalline basement. In the second category, informations on the structure and thickness of the Earth's crust have been considered (L. Constantinescu et al., 1972). With regard to the first category, we would mention the basement fault fields as having an important part in the underground circulation of cold or thermal waters and controlling

the deposit-like water accumulations. Out of the second category informations, though poorer, we mention the fact that the Earth crust's thickness in the northern part of the region ranges between 26 and 28 km, the crust is slowly thickening north- and eastwards, and the granitic layer is situated at an average depth of 10 km. There is a good correlation between the depth of Moho discontinuity determined seismically and the depth deduced from gravimetric data (M. SOCOLESCU et al., 1964).

The existing electrometric data are suitable for detecting water-bearing horizons, and to reveal relationships between the latter and the geological structure, including the role of faults in the underground circulation of cold, thermal or CO_2-gaseous waters.

3. *Geothermal data and geo-radioactivity.* The results of thermometry belong to four categories: determinations of thermic conductivity of the rocks, borehole thermometry, temperature of underground waters and thermometric prospection.

The thermic conductivity determinations are at the very beginning, they are few in number and restricted to some rocks from a wider scale of them found in the subsoil of the region concerned (CR. DEMETRESCU, 1973). The extreme values correspond to andesite (6.16 m cal/cm.s.°C) and tuffs (2.20 m cal/cm.s.°C). It has been established that the value of thermic conductivity varies even for the same rock-type (depending on the diagnostic physical and mineralogical factors). Volcanics and densely fissured sedimentary rocks show higher values.

The borehole thermometry has provided the most important data regarding the thermal conditions in the Western Plain. It has been established that the subcontinental geothermal anomaly in the Pannonian Depression extends to the western territory of Romania (V. SCHEFFER, 1963), and the most developed geothermal system of the country's whole territory is in the Western Plain (V. NEGOITA, 1970). It has also been stated that in this subcontinental anomalous geothermal system there are several distinguishable categories of geothermal anomalies with successively decreasing regionality.

The thermal features of the underground waters were monitored on a systematic hydrogeological ground, in two perimeters: to the north of Oradea and in the region of the Calacea Spa. In each of these perimeters a *hydrogeothermal anomaly* has been revealed. In the central zone of the first hydrogeothermal anomaly, the geothermal gradient is of 5.5 °C/100 m, while at the surface the water temperature ranges between 85 – 95°. The temperature of thermal waters shows a radial and gradual decrease down to 40 °C. Further away the thermal water temperature decreases gradually to the cold water temperature. In the hydrogeothermal anomaly of the Calacea Spa, thermal water temperature at the emergence points ranges between 34 and 42 °C. In both cases the thermal water accumulation is stored in the Upper Pannonian.

The only thermic prospecting operation performed in the Western Plain territory, covers a surface of about 20 km² in the Oradea region, including the perimeters of Felix and 1 Mai Spas. The temperature distribution map reveals the tendency of a regional increase with local hydrogeothermal anomalies of 3 – 4 °C.

Up to now, the radiometric investigations refer only to the radioactivity of mineral waters, of the ground waters in deeper aquifers and of some rocks crossed by drills. These have a low content of radon, radium and uranium.

However, when comparing the radioactive substance contents in waters and in the watered rocks, it appears that the formers are richer than the latters. The surplus of radioactivity in the waters, has to be interpreted as originating from deeper zones where rocks with higher content of radioactive substances occur.

4. *Deep geological structure elements.* According to the map of the deep geological structure of the territory of Romania (I. Gavat et al., 1965), it appears that the substratum of the Western Plain has a complex structure.

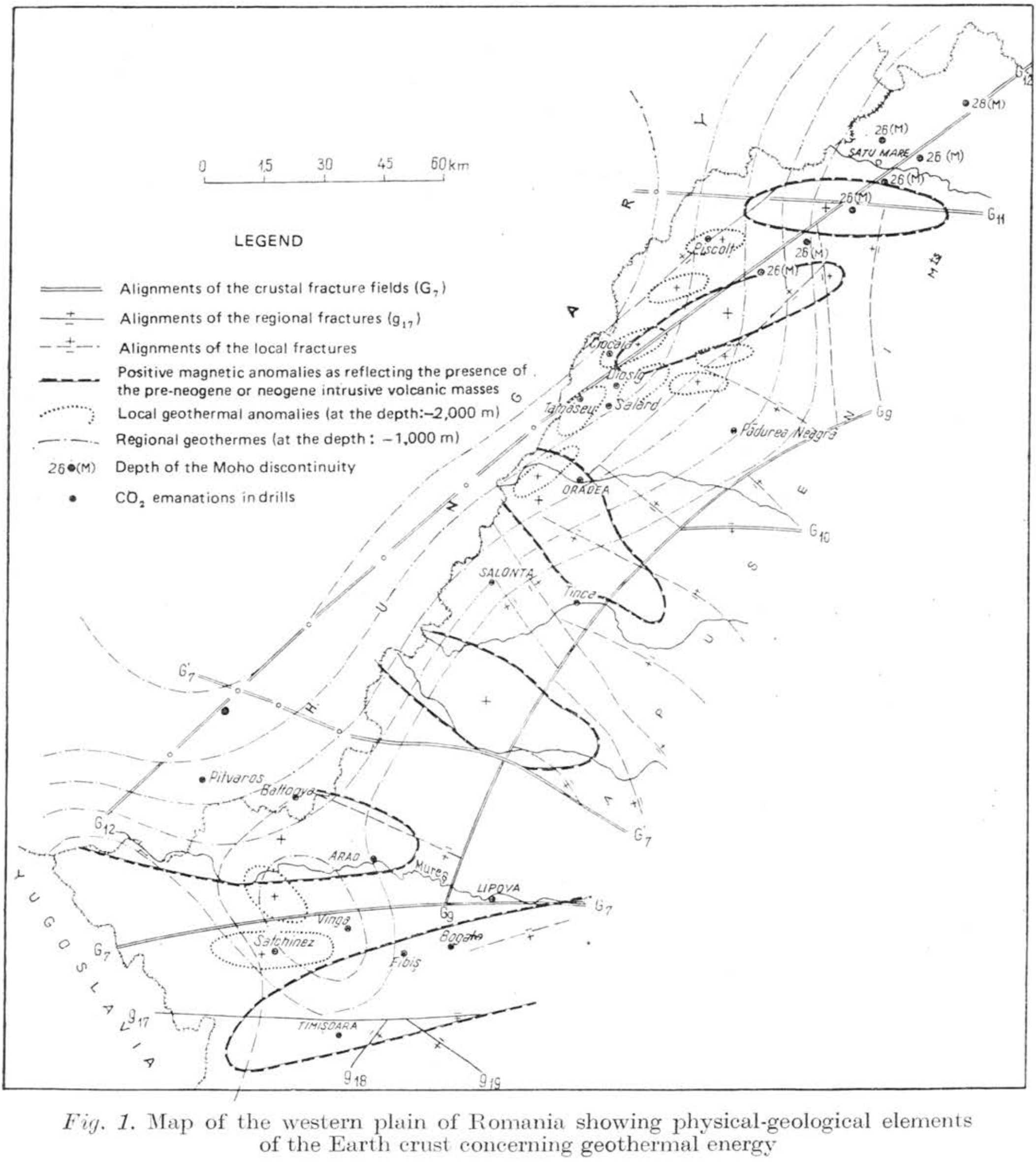

Fig. 1. Map of the western plain of Romania showing physical-geological elements of the Earth crust concerning geothermal energy

In the enclosed plate only a part of the geophysically determined fractures is shown, which belong to three classes: crustal fracture fields (1st order), regional fractures (2nd order) and local fractures (3rd order).

Five crustal fracture fields belong to the first category (G_7, G'_7, G_9, G_{10}, G_{11} and G_{12}). They intersect the whole Earth's crust, generally having continental extensions and are assumed to have a considerable role in the process of geological evolution of the region. At present they are overlain by sedimentary cover and, although being to a good extent welded, they represent the zones of the weakest lithospheric resistance.

Of the regional fractures there are partially reproduced three alignments (marked g_{16}, g_{17}, g_{18}), situated to the south of the Mureș Valley. They give the impression of being similar to the tectonic type in Banat, which continues to be Kraistidic to the south of the Danube.

By far more numerous are the local fractures, distributed in the spaces between crustal fracture fields and regional fractures. They are dividing into subordinated compartments, which as a result of the differentiated tectonical movements, have basement graben and horst positions. In addition to this, the detailed geophysical operations and drilling have shown that almost all the regional basement grabens and horsts are, in turn divided into local graben and horst compartments. The sedimentary cover is lying on this structural complex of the crystalline basement. An important role in the underground water circulation, besides the presence of the permeable horizons, is played by cover faults.

The geothermal phenomenon and the deep geological structure

To decipher the relationships between the geothermal phenomena and other physical phenomena manifesting themselves in the lithosphere of the Western Plain of Romania, the following considerations are to be kept in view:

1. *Geothermal process and the deep geological structure.* The distribution of the "geothermal field", as far as it is known now, reveals at least two categories of thermal effects of regional and local character.

The territory of the Western Plain is situated on the eastern flank of the regional geothermal anomaly of Central Europe, represented by the area of the Pannonian Depression. The characteristic values of this anomaly, regarded as deviations from the average value of the geothermal gradient, are: 18.5 m/1 °C in the central area of the Pannonian Depression and 35 m/1 °C at the exterior of the mountainous areas limiting it.

The lithological inhomogeneity and the variation of thickness of the layers making up the Earth's crust, are governing the distribution of geothermal values within the lithosphere of the Pannonian Depression to display areal irregularities. Such an irregularity, as subordinate regional geothermal anomaly, has been observed in the Western Plain of Romania. On the enclosed plate the sub-regional geothermal anomaly is partially sketched by isotherms drawn for a depth of 1000 m.

Actually, the lithological heterogeneity of the geological substratum, along with the areal distribution of the structural elements inside the frame of major tectonic units, establishes a heterogeneous system of geothermal properties being incidental to a profound modification of the temperature distribution and to the setting up of the local geothermal anomaly complexes of diverse areal extent.

In the area of the Western Plain there can be distinguished, according to their regionality degree, two local geothermal anomaly types. After the physico-geological causes that induce them, we may distinguish two categories: proper geothermal anomalies and hydrogeothermal anomalies.

The local geothermal anomalies of the first category are assigned, after their extension and intensity, to two sub-categories. The isotherm map for a depth of 1000 m contours two subregional anomalies of an intermediate regionality degree: one, between Oradea and Satu Mare, illustrated by a wide and strong inflexion of the isotherms from west to the east; the other, between Arad and Timişoara, also by an inflexion of the isotherms from NW to SE, and which, in addition, encloses a contour with temperatures of over 7 °C. Both subregional anomalies are situated in the intersection zones of two crustal fracture fields, each branched towards the surface by local cover fracture fields.

Over the described anomalies, there were detected groups of local geothermal anomalies by 8 contours in the anomaly area in the north and by 10 contours in the south (D. Paraschiv, M. Cristian, 1973, 1975). Their distribution is related to the directions of the fracture fields (I, II and III) constituted inclusively by cover faults.

The hydrogeothermal anomalies belong also to two sub-categories. To these we shell return in one of the following sections.

2. *Geothermal process and geo-radioactivity.* Regarding the thermic energy in the lithosphere of the Western Plain, we don't know almost anything about the proportion of heat contributions by the common internal heat of the Earth and that yielded by radioactive processes. The heat resulted from the decay of radioactive substances is added to that uprising from the inner zones of the Earth. Their distribution and movement in the lithosphere are in a function of the heterogeneity degree, thermic conductivity and radioactive substances content.

An intermediate geothermal anomaly is provided by the geological substratum between Arad and Timişoara. Numerous drills put down in the Şandra-Calacea structure show that subjacent to the relatively thin sedimentary cover, the crystalline basement is pierced by a granitic body, approaching the surface to as closely as 1100 m. The health-resort of Calacea, supplied with thermal water from wells, is situated on the apical zone of the granitic body. It is possible that this granitic body is part of the granitic layer of the crust, unblocked or bounded by faults towards the surface. These upward movements might have conditioned the general thinning out of the Earth's crust, as it has been suggested also by the maximum gravity anomaly of stretching in a N–S direction. Thus, the granitic body hit by wells, would be the physico-geological cause of the Arad–Timişoara subregional geothermal anomaly.

3. *Geothermal process and underground waters.* The knowledge of the hydrology of the Western Plain has been substantially improved upon a systematic investigation by drilling. The results obtained refer both to the horizons of cold water-bearing strata, phreatic and deeper, and to deeper or shallower levels of the thermal water horizons. Thermal waters occur in the Upper Pannonian, Lower Cretaceous and Triassic. These horizons are seldom issuing water through natural springs but mostly by wells. Water temperatures range from 25 to 100 °C.

The underground water temperature is variable in the zones of recharge to the water-bearing structures, where, obviously, waters are cold, but their temperature increases downwards to deeper levels i.e., it decreases when waters are involved in the ascending circulation. The heat lost by thermal waters is received by the rocks, involved in water circulation, giving them a surplus of temperature over their own, in proportion to their thermic conductivity.

Within the hydrogeological zonation of the Western Plain territory, the two hydrogeothermal anomalies, mentioned above, are well contoured: that to the north of Oradea, with the centre marked by the locality Galoșpetreu and the other to the south of Arad, at the Calacea Spa. Both hydrogeothermal anomalies are based on proper geothermal anomalies, somewhat more extended regionally, and whose caloric energy is responsible for the thermalization of the circulating waters in the area of the upper structural complex. Actually, the hydrogeological wells bored in the area of the northern hydrogeothermal anomaly, came across water-bearing complexes in three zones of the Mesozoic formations. At Borș, the wells yield waters of 90 °C. The resting zones are Oradea-Tabulia, in the SW of the hydrogeothermal anomaly and the Borod basin, eastwards from Oradea.

The circulation and distribution of the underground waters in the known hydrostructures is clearly predetermined by fault systems and zones of intense jointing in the sedimentary cover. Near the surface, their presence may be evidenced by thermometric prospecting, and on hydrogeothermal anomalies of low value and small extension. That is the case with the hydrogeothermal anomalies mapped in the zone of Felix and 1 Mai Spas. Moreover, the cover faults may form screen between the vertical differently moved sections with some lateral confinements of the water-bearing deposit, and even with lower water temperatures above and higher temperatures below. That is the case of the so-called "temperature inversions of the thermal waters" in the Oradea region, where the three water-bearing complexes show the following situation: the first complex, situated between 110—280 m, has a great flow capacity at a hydrostatic pressure of 1.5 atm, and a temperature of about 50 °C; the second, between 400—650 m, also with a great flow capacity, at a hydrostatic pressure of 2 atm and with temperatures of about 45 °C; the third, between 1000—1600 m, with a smaller flow capacity, at a hydrostatic pressure of 0.8 atm. and with a temperature of 25 °C. The facts are telling for themselves.

4. *Geothermal process and carbon dioxide.* The presence of carbon dioxide in the Western Plain has been proved by boring in three zones: the first to the north of Oradea; the second north-eastwards from Oradea, and the third, WSW of Lipova. The CO_2 gas originates from depths of more than 1000 m, and is flowing under pressure exceeding even 100 atm. Isolated CO_2 emergences are also known, as those at Pădurea Neagră, Tinca, Buziaș etc.

The northern zone is grouping wells on the area of the Ciocaia—Diosig—Sălard—Tămășău—Niudved localities, located on one side and the other of the crustal fracture field G_{12} and at the NW periphery of the magnetic anomaly which lies in the continuation of the Neogene eruptive rocks of Baia Mare.

The second zone is situated between the crustal fracture fields G_7 and G_7', on the northern flank of the magnetic anomaly of Arad. CO_2 gas is present in the wells of Sînana.

The third zone, situated WSW of Lipova and to the south of the crustal fracture field G_7, is characterized by surface manifestations of CO_2, either dissolved in the circulating underground waters (CO_2-gasiferous waters of Lipova, Bogata, etc.), or as CO_2 babblings (Vinga, Fibiș, a.o.). In wells, it was discovered only in the zone of Satchinez, at a depth below 1000 m.

We hardly know anything about the genesis of CO_2 gas, except for the mofettic aureola in the Eastern Carpathians (St. Airinei, A. Pricajan, 1975), as well as for a suggestion resulted from the attempt of involvement of the territory of Romania into the concept of "plate tectonics" (St. Airinei, 1975). The carbon dioxide occurrences of the Eastern Carpathians are considered, together with the geothermal phenomena, solfatara and the deposition of young carbonate rocks, to be postvolcanic manifestations of the Neogene volcanism of the Volcanic Mts. The same explanation was given also to the presence of CO_2 in the mofettic aureola of the Apuseni Mts. However, both mountainous segments lie on two subduction zones, responsible for the remobilization of the crustal fracture fields, for their surfaceward branching and, finally, for the putting in place of the Neogene eruptive rocks, as well as for the preservation of deep physico-geological phenomena connected with the remelting of the subduced forehead plates.

Instead of conclusions

The integration of geophysical and geological informations, specific to the concerned problem, has led to the formulation of a conceptual model regarding the geothermic investigation devoted to the interpretation of geothermal processes of the underground waters in the Western Plain of Romania. The obtained results, new though initial, open new perspectives for the future investigations.

The distribution of the geothermal energy in the substratum of the Western Plain gives a complex picture of phenomena consisting of a subcontinental thermal anomaly extending to the whole area of the Pannonian Depression, in which there are at least three types of local i.e. sub-regional geothermal anomalies superimposed. The complex distribution of geothermal energy is pre-determined by the material and structural distribution of rocks of the Earth's crust, with the role of their thermic conductivity and radioactive substances content being highly involved to. An important part is played by the crustal fracture fields along with all their branching towards the surface, as well as by the underground waters that take part in circulation. They borrow caloric energy from the crust, there after transporting and partly transmitting it into the rocks along their entire path, especially when ascending. Obviously, the degree of increase in temperature of the underground waters is proportional to the thermic energy of the medium of the water circulation. For one and the same depth level, the water temperature will be higher on account of local geothermal anomalies than it would be where the distribution of the thermal energy is quasinormal.

When hydrogeological structures display high storage capacity for thermal waters, such aquifers become sources of *hydrogeothermal anomalies*. The better case is when the hydrogeothermal anomalies are situated over the proper geothermal anomalies. From a practical viewpoint, hydrogeothermal anomalies are of an extraordinary significance, by the fact that waters bearing them are the very subject in utilizing thermal energies.

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A GEOTHERMAL FLOW SYSTEM IN THE PANNONIAN BASIN CASE HISTORY OF A COMPLEX HYDROGEOLOGICAL STUDY AT TISZAKÉCSKE

UN SYSTÈME D'ÉCOULEMENT DANS LE BASSIN PANNONIEN: HISTORIQUE D'UNE ÉTUDE HYDROGÉOLOGIQUE COMPLEXE A TISZAKÉCSKE

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RÉSUMÉ

Dans les eaux souterraines de Tiszakécske et Lakitelek il y a une concentration totale et une teneur en Cl relativement élevées, et elles sont caractérisées par une haute température. Les données géothermiques déterminées dans les puits indiquent une anomalie géothermique positive dans le territoire d'étude. Ainsi, le site semblait être favorable pour une production d'eau chaude.

Dans la première phase de cette étude géothermique un réseau de forage ayant un espacement d'un km a été établi et dans chaque puits un programme de thermométrie était accompli dans l'intervalle de profondeur de 30 à 50 m. L'exactitude des mesures était de ± 0,02 degré C.

Pour l'interprétation servent les gradients égalisés calculés pour chaque puits. La rédaction d'une carte réduite a été accomplie à la base de la répartition des materiaux les plus importants caractérisés par la conductivité de chaleur (c.d. sable — argile). La véritable base pour des données hydrogéologiques c'est la carte des gradients réduits qui représent simultanément la répartition relative du flux de chaleur.

Cette anomalie géothermique peut être expliquée surtout par une circulation des eaux souterraines, tandis qu'une position élevée du socle cristalline est moins plausible.

Dans la seconde phase, le développement de ce champ géothermique était exécuté par 10 forages de prospection et d'exploitation. Ces forages ont traversé des cuvettes sédimentaires quaternaires, pliocènes et miocènes. On peut distinguer deux nappes d'eau chaude souterraine: les couches aquifères supérieures quaternaires composées de sables et graviers et les couches aquifères inférieures renfermant des sables du Pliocène supérieur et moyen représentant des réservoirs multiples d'eaux thermales. A la base de l'étude détaillée des conditions géologiques, hydrogéologiques et des paramètres de réservoir on peut supposer la mise en communication des couches aquifères superposées.

L'anomalie géothermique locale peu profonde en question s'est produite par les eaux thermales ascendant de la nappe inférieure par suite de phénomènes stratigraphiques, hydrodynamiques et thermodynamiques particulières. En même temps, la partie méridionale du territoire d'étude est caractérisé par une anomalie négative produite par des eaux froides descendantes.

In the waters of the shallow wells of Tiszakécske (Fig. 1) and its environment a relatively high dissolved material and chlorine (Cl) content was found, and also from the drillings into medium depth warm water issued (M. Erdélyi, 1964.). The relatively great value of the apparent geothermal gradient computed from the data of these wells indicated a positive geothermal anomaly, the investigation of the area, hoping for heat energy production, came therefore to the front.

Fig. 1. Location map

1. Geology of the area

1.1. Stratigraphy

It is assumed that the basement at the site consists of paleozoic (or older) crystalline complex outcropping only outside the study area. The overlying Miocene formation which is separated by a major discordance has an average thickness of 300 meters and is composed of marls, sandstones and limestones of marine origin. The Pliocene sediments separated by a minor unconformity are developed in three groups: Lower Pannonian (=Lower Pliocene), Upper Pannonian (=Middle Pliocene) and Levantine (=Upper Pliocene) (Fig. 2). The sedimentary environment was developed progressively from marine to lacustrine character. The Lower Pannonian substage has a dominantly marl sequence of impervious character with intercalated sandstones. The thickness of this formation is about 450 m. The Upper Pannonian substage consists of two parts: the lower portion is a mainly shaly formation with tightly cemented sandstones, while the upper portion forms an alternating sand and clayey marl sequence (due to oscillating sedimentation) where sand beds are good aquifers. Total thickness of the entire Pannonian formation amounts approximately to 700 meters. The Levantine formation of about 250 m thickness has similar sedimentary pattern as the underlying part of the Upper Pannonian series. Some minor unconformities within the whole Pliocene sedimentary column may be presumed.

Average grain size of sands within the Pliocene aquifers is varying from 0.1 to 0.7 mm. In some places there are some intraformational gravel beds of 7 to 15 meters thickness.

There is no persistent, continuous, thick shale unit which could separate the underlaying multistoried aquifer system from the Quaternary formation. The junction is characterized by an apparent disconformity. The whole Quaternary sequence is of fluviatile origin and consists of thick gravel and sand deposits with interbedded clay formations. The total thickness of the Quaternary is about 500 meters. The afore-mentioned lack of an effective, impervious cover allows, in principle, a vertical communication among the different water-bearing rocks within the upper portion of the sedimentary column.

Particle size of the Quaternary gravels is varying from 2 to 8 mm while grain sizes of sands range from 0.1 to 2 mm, that is, forming a variety of fine sands through very coarse ones.

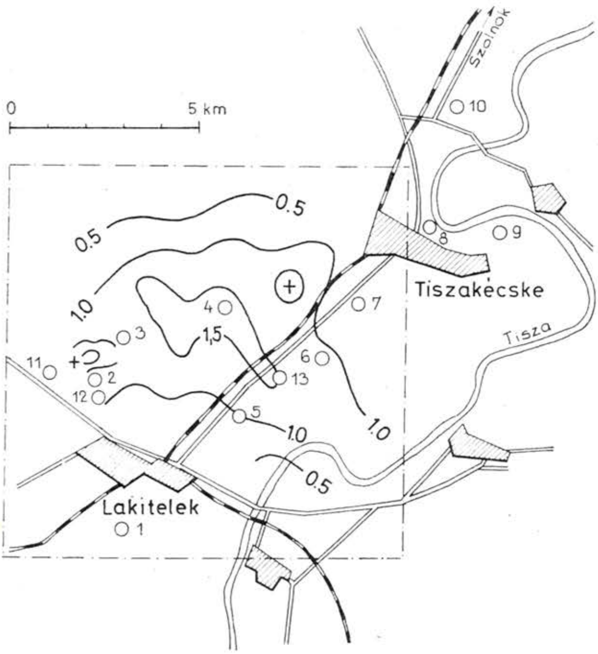

Fig. 2. Map of the study area. Well numbers from 1 to 10: deep thermal wells; numbers from 11 to 13: shallow wells. The thick isogradient lines, numbered in °C/10 m, represent the thermal anomaly

The surface of the study area is covered by thin, loose (blown) sand of high infiltration capacity.

The entire sedimentary column of Neogene age indicates a steady subsidence of the depositional environment only with slight unconformities at the Mio-Pliocene and Quaternary-Pliocene boundary.

1.2. Tectonics

The basin-architecture is a very simple one without any definite fracture or folding element. However, the existence of some local and limited atectonic fracture-forms must be taken into account. A diagrammatic geologic cross-section is shown in Fig. 3.

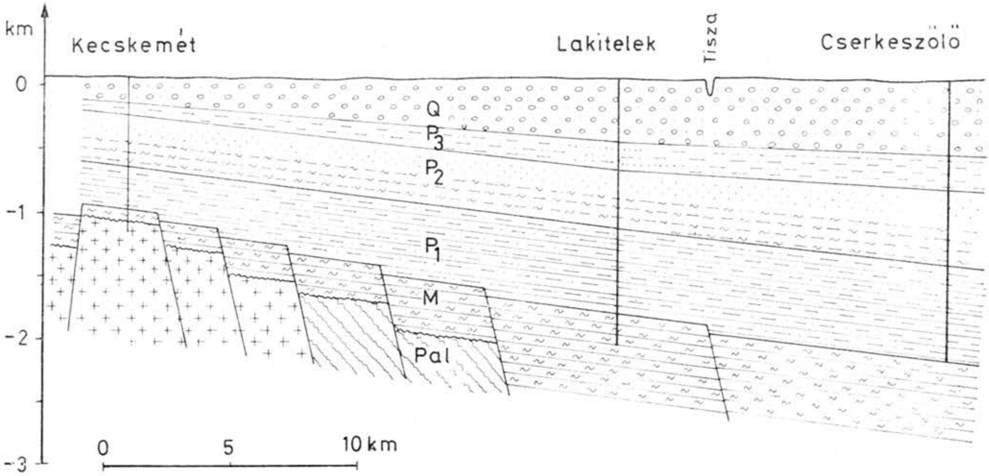

Fig. 3. Geological profile. Q: Quaternary; P_3: Levantine (= Upper Pliocene); P_2: Upper Pannonian (= Middle Pliocene); P_1: Lower Pannonian (= Lower Pliocene); M: Miocene; Pal: Paleozoic

1.3. Hydrostratigraphic units

Two distinct hydrostratigraphic units can be distinguished in the area. One is the Upper Pannonian and Levantine multiple sand reservoir system within 500 – 1000 m depth interval representing the main thermal water-bearing formation. The porous formations within the Upper Pannonian and Levantine sequence are sheet-like sand bodies which are often multi-layered, laterally coalescent and/or pinching and wedging out forming sand lenses and patches of limited lateral dimension.

The other is the Quaternary multistory aquifer system ranging from the near-surface to 500 m depth. Practically the lower section of the Upper Pannonian series below 1000 meters depth can be regarded as impervious barrier against any effective fluid flow. Based on the grain size distribution, the porosity of the sand bodies is thought to be as 30 – 42 per cent, due to the low degree of compaction. Assuming a relationship between porosity, permeability and grain-size distribution, the following range of permeability (k) and of hydraulic conductivity in case of water of 22 °C *(K)* can be estimated:
in the Pliocene aquifer system:

$$k = 0.5 - 5.0 \text{ darcy } (K = 0.5 \cdot 10^{-5} - 5.0 \cdot 10^{-5} \text{ m/s}),$$

in the Quaternary aquifer system:

$$k = 1.0 - 10.0 \text{ darcy } (K = 1.0 \cdot 10^{-5} - 10.0 \cdot 10^{-5} \text{ m/s})$$

or even more.

The multiple reservoir system represents an infinite reservoir complex without lateral boundary conditions as a whole. Due to the depositional-sedimentary development of the entire aquifer system the two hydrostratigraphic units mean only geologic classes because a certain form of communication between the upper and lower units is very likely. The entire aquifer system is partly open to the surface in the top part of the sedimentary column and precipitation can thus easily percolate downwards.

2. Surface temperature measurements

For the performing of temperature measurements some bases were already existing. Results of temperature measurements carried out elsewhere in 2–5 m deep bore-holes have proved that because of the propagation of daily temperature variations, on one hand, and of the irregular heat conductivity variations of the soil near the surface on the other, only large regional thermal anomalies can be revealed by this method (L. STEGENA, 1952). Measurements carried out in deep holes (of several hundred meters) are not subjected to these errors, the large costs are however, economically prohibitive. The drilling of several ten m deep holes carried out by a truck-mounted drilling rig can be realized quickly and at a low cost. The methodological task of the measurements at Tiszakécske was to clear up the question whether the geothermal anomaly could be detected with sufficient accuracy by drillings of such small depth.

2.1. Field procedure of thermic measurement

The mesurement area along the Tisza River is almost plane. On the area presenting a maximum level variation of 20 m the deepest point was selected as 0 level. The bore-holes, in which temperature measurements were made, had been driven in the corners of a quadratic network of km scale down to 53 m into the thick Pleistocene deposit. The temperature was measured in the lower section of the holes (from 20 to 50 m, at every 5 m). In this section the daily temperature variation is not perceptible any more, and the propagation of the slight and slow yearly variation is not disturbing (L. STEGENA, 1952).

The temperature measurement (M. HARTNER et al. 1964) was carried out with a resistance-thermometer connected to a Wheatstone-bridge according to the zerocompensation method. The expedient calibration and control methods and the minimum time needed to establish the thermal balance were determined (48 hours). The error of the geothermal gradient-measurement did not exceed 0.02 °C. Absolute temperature measurement (the comparison of the temperatures of different bore-holes) is less accurate, the error may attain the twofold of the error of the gradient.

The 53 thermal logs, comprising 7 temperature data each, which have been taken by direct temperature measurement in 53 holes drilled on the measurement area constitute the initial–raw–data for the processing of the geothermal measurements.

2.2. Utilization of thermal data

From the raw data the local temperature values *(t)* are utilized, when determining the depth of isothermal sites and the isothermal surfaces.

An illustration of a constant value can be obtained if isothermal surfaces are intersected by horizontal planes at a specified depth and in the plane

—as in the case of the horizontal plane of 30 m depth in Fig. 4—the intersecting curves of the isothermal surface are drawn. For the sake of comparison with geological or geophysical sections it is more advantageous to intersect the isothermal surfaces along a line by vertical planes and to illustrate the intersecting lines in that plane.

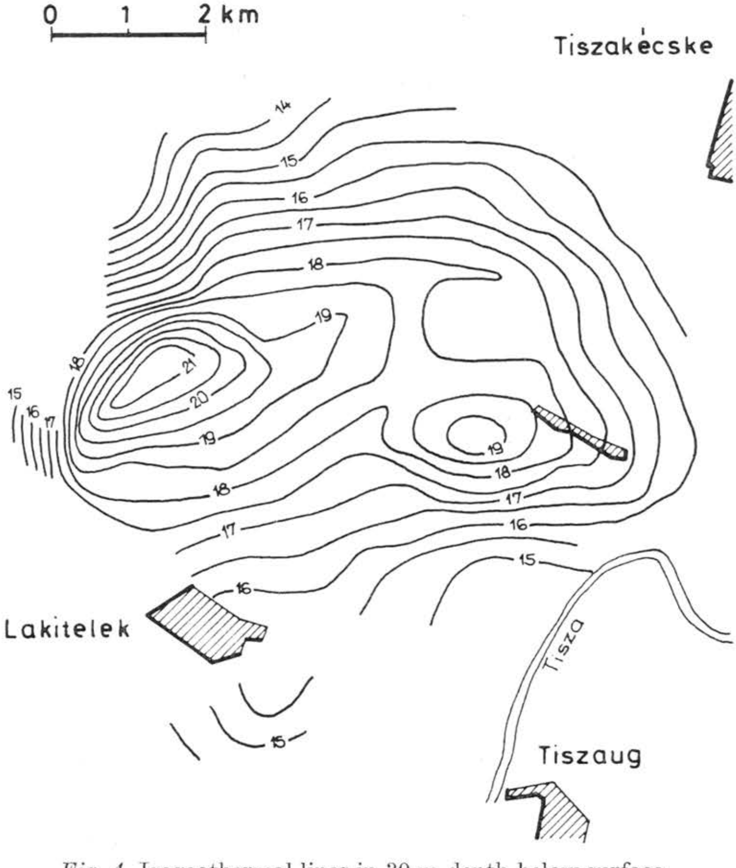

Fig. 4. Isogeothermal lines in 30 m depth below surface.
The value of isotherms is given in °C

For detailed investigation the *gg* data have been used, since these are the more accurate, primarily measured values. From the temperature log of each bore-hole, one gradient is obtained with the very plausible assumption that *gg* is constant at a great distance from the heat source, within a relatively short measuring range (30 m). In this case the best value of *gg* is obtained by determining the tangent *GG* of the straight line approximating best the temperature log of each bore-hole. The resulting *GG* values are given in °C/10 m in Fig. 5.

For studying the areal distribution of the *GG* values, their deviation from the areal mean gradient was calculated. The mean value between the Danube and Tisza River, as determined by BÉLTEKY, 1963 and SCHEFFER, 1964 equals to 0.54 °C/10 m. It is seen on Fig. 5, that the majority of the *GG* values is

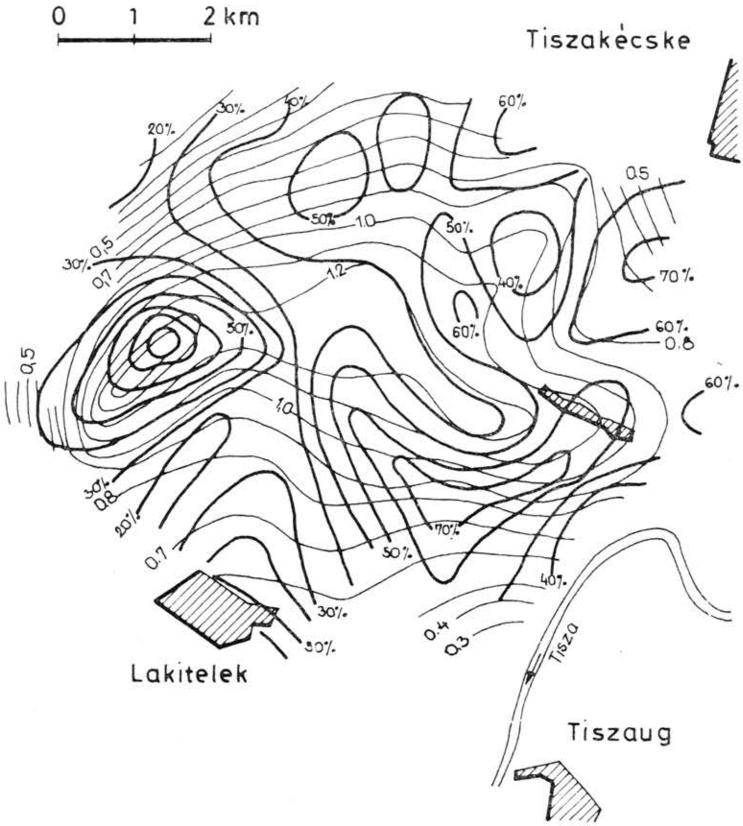

Fig. 5. Map of equalized thermal gradients. Value of isogradient contours (thin line) is given in °C/10 m. Thick lines represent places of equal clay content in per cent of the total quantity (clay and sand together). The isogradient and clay distribution curves show obvious similarity

above the *(c)* mean value – as to be expected – on the NW and S parts of the measurement area, however, *GG* values below the mean appear. Assuming identical heat conductivity (corresponding to the average of the Quaternary formations) over the entire area, – as would occur if *GG* values were transferred to the deeper layers unchanged – the heat flux would hardly exceed the Danube and Tisza Rivers, and at the margins an unjustified "negative" anomaly would appear. This first and more detailed analysis of the data focussed the attention on the heat conductivity-anomaly distorting the picture of the thermal anomaly.

The Pleistocene deposits in the area consist, of sandy-clayey layers, i.e. of the mixture of sands of better heat conductivity ($k_s \sim 6 \cdot 10^{-3}$ cgs) and of clays with worse one ($k_c \sim 3 \cdot 10^{-3}$ cgs). In areas where *GG* remains below the average, i.e. the decrease of temperature is small, a good heat conducting medium – sand – is to be assumed.

The contact of the sand and clay layers has been plotted from the samples taken during the drilling, allowing also the determination of the sand-clay ratio for each borehole. The average heat conductivity change can be concluded from the variation of the sand-clay ratio. On the sand-clay distribution referring to the lower 30 m of the drillings, the dotted curves of Fig. 5 provide information: they denote the clay content in per cent of the total quantity (clay and sand together). The *GG* curves and the clay distribution curves show a striking similarity: the gradient maximum at the W margin of the area is also the maximum of clay content, and the same coincidence can be found at the eastern *GG* terrace too. Fig. 5 refers to the fact that the *GG* maximum results – at least to some extent – performing the measurement in a bad heat conductive medium (i.e. in clay) and also to the *GG* values in consequence of the increasing sand content to the layers.

2.3. *Relative heat flux map*

These investigations referred to the relationship between the average gradient *(GG)* and the average clay content suggesting also the possibility of the reduction of thermal gradients.

After having investigated the possible correction methods the reduction of the *GG* data according to the percentage of the clay (respectively sand) content proved to be to best advantage. The reduction was made for a medium of $3.5 \cdot 10^{-3}$ cgs heat conductivity, i.e. for a sand-clay mixture containing appr. 70 per cent clay. The data of several hundred drillings, justify the assumption of a clay-sand mixture. Local pebble bodies, which are otherwise conspicuous enough, were not encountered on the area.

The $3.5 \cdot 10^{-3}$ cgs heat conductivity is somewhat higher than those determined for Quaternary layers in other areas by *in situ* measurements (J. GÁLFI et al., 1961). It has been taken into account, however, that in the area the quantity of bad heat conductive medium is relatively small. For *in situ* heat conductivity measurements there was no possibility – in routine work this is unfeasible – therefore the already mentioned average value was accepted. Laboratory heat conductivity measurements on core samples seems rather complicated since the samples taken from loose deposits do not remain in their original state.

Computing the heat conductivity (k_m) of the mixture composed of C percentage of clay and S percentage of sand by the $k_m^{-1} = C \cdot k_c^{-1}$ formula, assuming that $k_c = 3 \cdot 10^{-3}$ cgs and $k_s = 6 \cdot 10^{-3}$ cgs, we obtain the *GG* data reduced to the medium of $3.5 \cdot 10^{-3}$ cgs heat conductivity ($GG^{\times}$) as follows:

$$GG^{\times} = \frac{GG}{3.5} \cdot k_m = \frac{GG}{3.5} \cdot k_s \left(C \cdot \frac{k_s}{k_c} + S \right)^{-1} = GG \cdot \frac{1.7}{1+C} \, .$$

To the reduction formula a ratio of $k_s/k_c = 2$ has been used. Assuming very good heat conductivity for the sands – $k_s = 9 \cdot 10^{-3}$ cgs – the above ratio changes to 3, the reduction multiplicator being $2.16(2C+1)$. The two $GG^{\times}$ maps prepared by these two different reductions deviate only in unsignificant details, therefore as the last step of data processing only the $GG^{\times}$ map calculated by the first formula is given in Fig. 6.

Taking into account the method of reduction given above the $GG^{\times}$ map represents also the areal distribution of relative heat flux data.

The anomaly — at least in brief outlines — could be limited, according to different points of view, by indirect geological methods (M. ERDÉLYI, 1964). As a result of the surface geothermal measurements, the *GG* map prepared in the course of the detailed elaboration furnishes information (Fig. 5) in some further details. In the W part of the area a great maximum can be seen, in the centre there is a terrace arrangement, which suggests a crossing of two structural lines. The reduced gradient map (Fig. 6) clears up the thermal

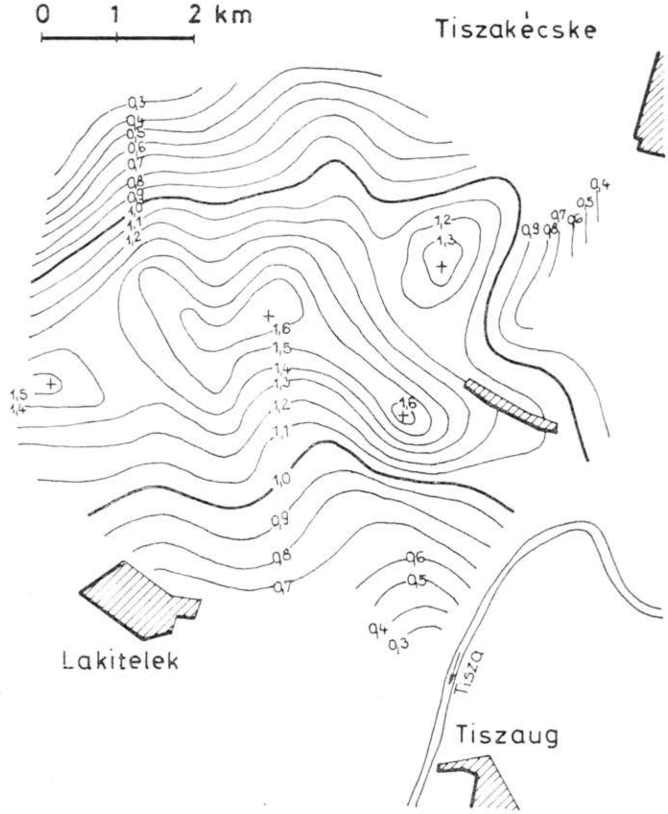

Fig. 6. Map of reduced thermal gradients. Values of equalized gradient are reduced to sand-clay mixture having $3.5 \cdot 10^{-3}$ cgs heat conductivity value. The isogradient curves are numbered in °C/10 m. This map represents also the relative distribution of natural heat flux

conditions also in details. The western peak value of the gradient is shifted towards E and the crossing of the two principal structural directions forms it decisively. The developing secondary maxima show a slight deviation in value from the middle maximum. The principal strike of the thermal anomaly follows the great W — E structural line. *By the systematic elaboration of the geothermal measurement data the number of possible geological models has been derived:* the anomaly can hardly be interpreted by an elevated basement, rather thermal water is to be assumed to rise up along some atectonic fracture mentioned in Part 1.

3. The study of the thermal anomaly by drilling

The preliminary research work was followed by an intense drilling exploration campaign. A total of 13 thermal water wells were located within the study area. The first exploratory (Well No. 4) was only 830 m deep. The second exploratory (Well No. 3) has a total depth of 2048 m and penetrated Quaternary, Pliocene and Miocene formations, while the basement was not reached. However, it is supposed to be crystalline basal complex of Paleozoic age. Both wells were located on the geothermal anomaly which was indicated by thermal data of shallow prospecting well-network. Both wells have mainly scientific objectives. The aim of drilling the other 11 wells was the thermal water production for agricultural and bathing purposes.

The initial *water-yielding capacity* of the single wells ranges from 400 to 2400 l/min while specific yield is from 58 to 194 l/min/m. The annual water production of single wells varies from 0.12 to 0.50 million cu.m. The cumulative water withdrawal by the end 1972 was 10.81 million cu.m, from which 6.52 million cu.m comes from the Quaternary aquifer system and 4.29 million cu.m from the Pliocene. It must be noted that beside 3 thermal water wells completed in the Quaternary water-bearing formation, the entire Quaternary

Table 1

Main water-yielding characteristics of production wells

No. of Well	Total depth m	Producing interval m	Initial water-yield l/min	Initial dynamic water level m	Specific yield l/min/m	Flowing water temperature °C
1	1200	995 – 1025	500	−8.0	58	38
2	904	756 – 830	440	+0.5	96	54
5	1201	767 – 984	1050	+1.5	66	57
6	1425	925 – 1003	2012	+2.0	109	64
7	1202	779 – 939	2160	+1.4	138	58
8	1200	771 – 962	2400	+3.9	194	53
9	1482	802 – 1302	1700	+3.0	89	55
10	1100	759 – 883	1600	+1.0	144	51
11	253	167 – 247	600	−6.0	115	38
12	212	181 – 206	500	−8.0	58	38
13	225	211 – 220	1500	+1.5	143	42

All the depth values are given in metres below surface. Positive and negative values of the dynamic water level denote levels above and below surface, respectively

aquifer system is highly developed by an extensive shallow well-network producing water of lower temperature (less than 35 °C) for drinking purposes.

The majority of these wells yields flowing water, the dynamic level ranges from zero to about 2.0 m above surface.

Table 1 shows the main water-yielding data, Fig. 2 the location of the thermal wells.

4. Thermal water chemistry

The two hydrostratigraphic units, namely the Pliocene and Quaternary aquifer systems show a marked difference in the composition of water.

In the Pliocene aquifers concentration increasing from SW to NE. Thus the southernmost occurrence shows a *total solid content* of 647 mg/l (Well No. 1) while the northernmost site has a value of 3618 mg/l (Well No. 10). Chlorid ion content ranges from 20 to 360 mg/l varying in the same manner as the total dissolved content. The character of these waters is defined by its prevalent alkaline hydrogene carbonate content (bicarbonates) which suggests that the main volume of this water has been originated directly or indirectly from the surface.

Waters in the Quaternary formations have usually lower concentration, although in the central part of the study area which coincides with the maximum value of the geothermal anomaly, total dissolved solid content is the same as in the underlying Pliocene waters (1350 mg/l in the Well No. 13). Chlorid ion content, however, is everywhere low (about 30 mg/l) in accordance with the previous hydrochemical observations made by M. ERDÉLYI (1964). The character of these waters are similarly of bicarbonate type.

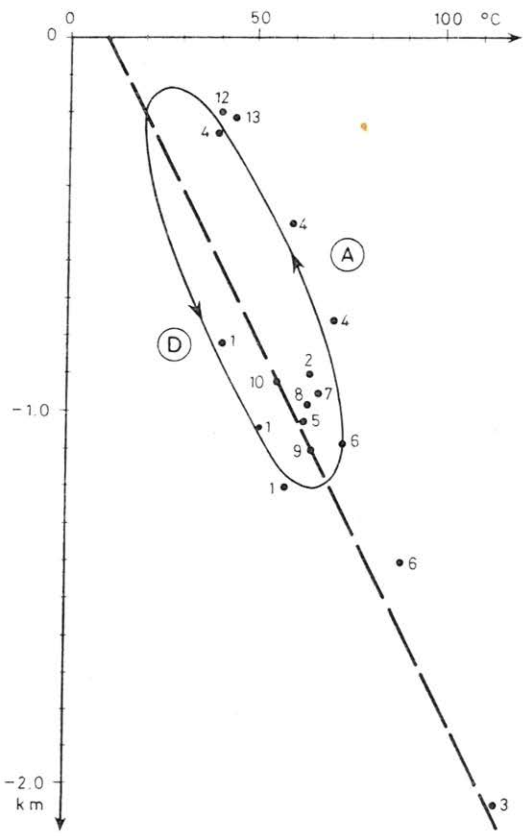

Fig. 7. Formation temperature versus depth. Numbers at measuring points refer to the numbering of wells given in Fig. 2. The average increase of temperature versus depth is shown by broken line. The sides *D* and *A* represent areas of descending and ascending groundwater, respectively

Natural *gas content* of the main thermal water reservoir system ranges from 100 to 1000 litres per cubic meters, in which methane makes up the greatest part (from 46 to 89 per cent). In addition, there is a variable but smaller amount of nitrogen and a minor amount of carbon dioxide. The gas-water ratio ranges from 0.1 to 1.0.

An increase of the dissolved gas content from SW to NE was observed.

5. Flowing water and formation temperature

5.1. Vertical distribution of temperature

Data of formation temperature measured in bore-holes as well as calculated from the temperature of the outflowing water provide the picture of the vertical structure of the thermal field.

The highest flowing water temperature in case of the two geologically determined reservoir systems was found within the previously described geothermal area (Part 2), that is 64 °C (Well No. 6) and 42 °C (Well No. 13) respectively. The measured and calculated temperature values are plotted versus depth in Fig. 7, where the average increase of geothermal heat in vertical direction (L. STEGENA, 1973) is indicated by the dotted line.

It is evident that temperature values on the right hand side of this line are indicating regions in which heat is being produced, those on the left hand side refer to areas of heat removal and values on it are marking "normal" regions i.e. regions without heat supply.

The anomalous area, encircled in Fig. 7, extends vertically to the upper 1000 meters of the sedimentary column. Below this level, formation temperature can be considered as normal.

5.2. Influence of convective thermal flow

To determine the theoretical form of the variation of stratum temperature (T) with depth (h) in cases where conductive and convective flow of heat are present at the same time the modified Fourier Equation should be used (CARSLAW, 1959).

In the area of thermal anomaly where the homogeneous flow of thermal water can be well represented by constant rate of heat production, a three layer model in which the thermal conductivity k has constant value and heat is produced only in the middle layer with a constant rate, gives fairly good approximation.

It follows from this consideration, that (L. ALFÖLDI, 1975) the curve T versus h, the thermal log, is concave/convex downward in a region where heat is being produced/removed, and (L. STEGENA, 1952) the zone of conductive heat flow is bounded by constant values of geothermal gradient (gg) i.e. by the linear segments of the thermal log. Thus, the convex shape of the thermal log of Well No. 1 in Fig. 7 which is situated in the infiltration area, indicates heat removal by the downward moving cold surface water and the thermal logs measured within the positive thermal anomaly display concave form in accordance with heat transport to this region by thermal water.

5.3. Residual temperature graphs

Though the thermal anomaly can be delineated by the method described here its shape and spatial extension will be better illustrated by the residual temperature logs. The residual log $r(h)$ is defined as the difference between the actually measured temperature values $T(h)$ and the values of average earth temperature pertaining to depth h:

$$r(h) = T(h) - \overline{gg} \cdot h - T_0$$

where $\overline{gg}$ and T_0 denote the average geothermal gradient and surface temperature, respectively. It should be noted that the main geothermal gradient

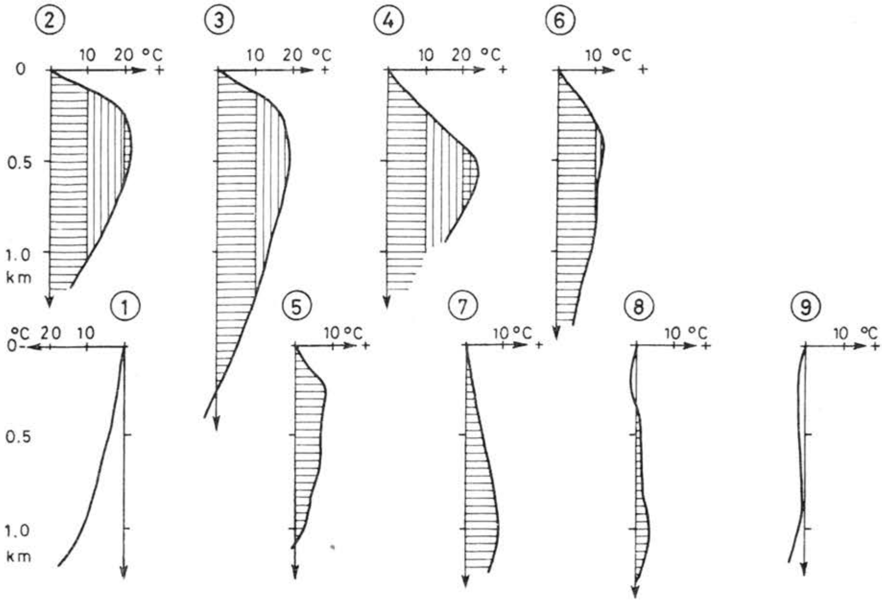

Fig. 8. Residual temperature graphs. The thick line represents the deviation of the measured temperature value from the average Earth temperature versus depth. Intervals of 10 °C are marked by hatching. The encircled numbers refer to the numbering of wells given in Fig. 2

calculated from data given in Fig. 7 for the upper 1000 meters of the study area equals to the gg averaged for the Hungarian Basin.

It is easy to see that the considerations given in 5.2 are valid for the residual graphs too: *(1)* the layer of convective thermal flow is limited by linear segments of residual curves and *(2)* $r/h=0$ (naturally within the error of measurement) is significant for non-anomalous areas.

From Fig. 8 in which the residual thermal logs are shown, can be concluded: *(1)* The layer of main convective thermal flow ranges from 200 m to 800 m depth beneath surface, rising up slowly to the Tisza River (from Well No. 4 to Well Nos. 13 and 6). *(2)* The thermal flow is decreasing from Well No. 6 in NE direction and is bounded by Well No. 7. Well Nos. 8 and 9 near Tiszakécske, penetrate non-anomalous layers. *(3)* The anomaly is bounded in SW direction between Well Nos. 5, 2 and 1 and, as it can be concluded from the shape of the corresponding logs, this boundary indicates the contact zone between areas of heat supply and removal.

5.4. Velocity of convective thermal flow

Flow velocity can be well estimated within areas of homogeneous flow of water with constant vertical velocity v. In calculating the rate at which heat crosses any horizontal plane a convective term must be added to the part due to conduction

$$F=-k\cdot(\partial T/\partial h)+\varrho\cdot c\cdot T\cdot v$$

provided heat generation owing to processes of compression and friction can be neglected. Here $\partial T/\partial h$ denotes the geothermal gradient in the absence of convection i.e. gg, c and ϱ the specific heat and density of water respectively. The study of thermal and electric logs measured in the zone of thermal anomaly as well as in normal area has not been shown any marked change in the value of heat conductivity within the upper 1000 meters. Thus, any finite difference ΔF of thermal flux F taken along the vertical axis can be written

$$\Delta F = \varrho \cdot c \cdot \Delta T \cdot v$$

where ΔT denotes the difference in temperature. This equation simply states that the difference of the flux of heat is converting to the warming up of the descending water in the inflow area and vice versa in the zone of ascension. If, instead of flux F the measured geothermal gradient gg is used, the above equation can be transformed in

$$k \cdot \Delta(gg) = \varrho c \cdot \Delta T \cdot v$$

or

$$v = k \cdot (gg_2 - gg_1) \cdot (\varrho \cdot c \cdot \Delta T)^{-1}$$

where gg_1 and gg_2 denoted the value of gg in depth h_2 and h_1 respectively. Estimation of the fluid velocity along a vertical flow-path can be carried out on this basis, if an approximate value of $k/\varrho \cdot c$ is determined. As a result of calculations, values of flow velocity within the order of $5 \cdot 10^{-9}$ m/s were obtained.

6. Reservoir and well-head pressures

6.1. Correction of well-head pressure data

Well-head pressures (measured in meters the elevation of static water level above surface) are influenced by temperature, gas content, dissolved solid content of water and by the elevation. Thus, static water levels of the deeper wells are higher than that of the shallow wells. In addition, the static water level in the southernmost part of the study area (Well No. 1) is lower than that of the other wells of similar depth. As a measure of formation pressure, the length of the water column in the well could be used if the density of water were the same, i.e. 1000 g/cm^3 in each well. Correcting the real densities according to the influencing factors mentioned before the corrected static water level will be obtained which represents the real formation pressure. Each correction was done in meters and the sea level was used as common reference level. Thus the corrected static level will be given by

$$H_{0\,\mathrm{corr}} = H_0 - H_T - H_g + H_s$$

in meters, where $H_{0\,\mathrm{corr}}$ = static water level a.s.l. corrected to density of 1000 g/cm^3, H_0 = measured static water level, H_T = correction due to temperature, H_g = correction due to gas content and H_s = correction due to dissolved solid content.

6.2. Vertical pressure distribution

The values of $H_{0\ \mathrm{corr}}$ data versus depth h, are shown in Fig. 9. It is seen and it is confirmed by a few measurements that the original static reservoir pressure within the Pliocene water-bearing system has a hydrostatic character with minor deviations. A definite minimal pressure exists within the geothermal anomaly at the Well No. 6.

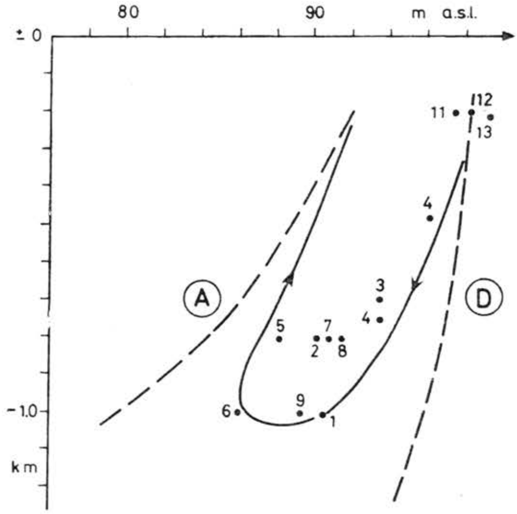

Fig. 9. Formation pressure versus depth. The dotted lines represent the hydrostatic pressure versus depth, the solid lines equalise the measured values in areas of descending (D) and ascending (A) character, respectively

6.3. Hydrodynamic gradient

Using the $H_{0\ \mathrm{corr}}$ data given in Fig. 9 the true pressure gradient and the difference of this gradient from the standard value of 1000 can be calculated in horizontal and vertical sense, respectively. Since the hydrostatic pressure P is defined as

$$P = \gamma \cdot h$$

the horizontal gradient in a hydrostatic field should be equal to zero, while vertical gradient has to be equal to the specific weight γ at any point, determined by temperature, dissolved salt content and pressure. The difference between the vertical hydrostatic gradient and the measured one have been indicated by the flow of water. The difference of these gradients will be referred as hydrodynamic gradient.

The calculation of the flow velocities needs the knowledge of the average horizontal and vertical values of hydraulic conductivity. The average horizontal permeability is assumed to be of $5 \cdot 10^{-5}$ m/s (Part 1.3) while the vertical permeability is thought only one tenth of that value, that is $5 \cdot 10^{-6}$ m/s.

6.4. Flow paths and flow velocity

Two major flowing paths or zones were investigated, notably *(1)* the assumed descending zone in the southern part of the study area at Well No. 1. and *(2)* the assumed ascending zone within the central part of the area near Well No. 6. These and other possible flow paths can be followed on Fig. 9, where the hydrostatic pressure versus depth is marked for these regions by dotted line.

All these data confirm the reality of the above assumptions and they indicate flow velocities of the order of 10^{-8} m/s. At the same time it might

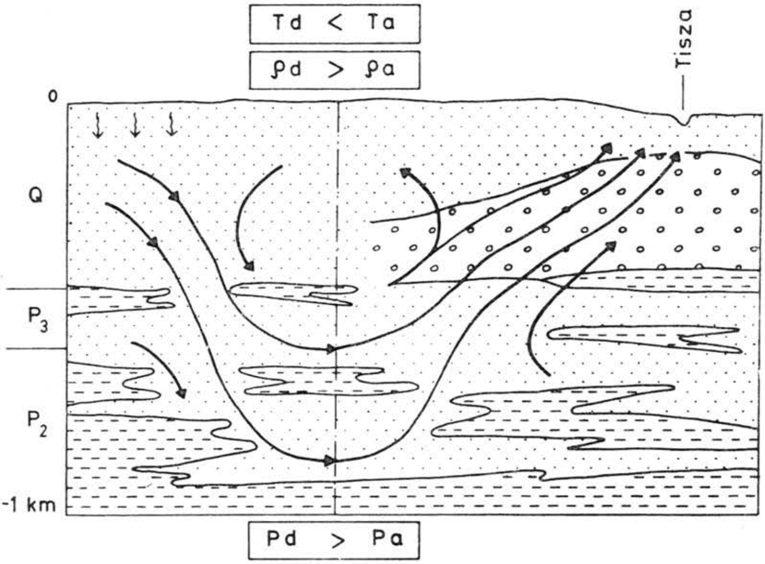

Fig. 10. Idealized diagram of the circulation system. *T:* temperature; ϱ: density; *P:* pressure; Q: Quaternary; P_2: Upper Pannonian (=Middle Pliocene); P_3: Levantine (=Upper Pliocene). Areas marked by dots and circles represent pervious sand and gravel layers, respectively, the dashed areas are impervious interbeddings and substratum. Letters *a* and *d* refer to the areas of ascending and descending groundwater. The supposed flow paths of groundwater are shown by thick lines and arrows

have been indicated that the main thermal water reservoir within the 800 to 1000 m depth interval, was drained by a shallow near-surface river-connected aquifer having a similar pressure to the average stage of the Tisza River which is the same.

The conformity between the flow velocity values calculated before and estimated through thermodynamical considerations (Part 5.4) suggests that the convective thermal flow is supported by thermal water movement. A sketch of the flow-system is given on Fig. 10.

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STUDY OF THE RECHARGE OF DEEP GROUNDWATERS AND THEIR CONNECTION WITH SHALLOW GROUNDWATERS USING ENVIRONMENTAL ISOTOPES IN THE NAGYKUNSÁG REGION, HUNGARY

ÉTUDE DE RAVITAILLEMENT DES EAUX DE PROFONDEURS A L'AIDE D'ISOTOPES NATURELS DANS LA RÉGION NAGYKUNSÁG, HONGRIE

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RÉSUMÉ

Par conséquence de nos analyses au tritium, au deuterium ainsi qu'au ^{13}C et ^{14}C nous avons pu établir ce qui suit:

— par datation des isotopes du carbon on a pu démontrer avec beaucoup de probabilité un courant d'eau montant en deux puits de l'eau artésienne, de différente profondeur, aux alentours de la ville Törökszentmiklós,

— se servant d'études au tritium, au deuterium et des isotopes stabils du carbone datation que la couche d'eau phréatique de la région Nagykunság est alimentée de la précipitation et nous ne pouvions y démontrer de l'eau de profondeur.

— le courant d'eau montant des eaux de profondeurs n'atteignent pas la couche des eaux souterraines phréatiques. Leur base de mise en perce se trouve à 20—60 m de la surface du sol,

— cette couche de 20 à 60 m draine non seulement les eaux souterraines profondes montantes, mais aussi les eaux phréatiques ruisselant à l'influence du gradient potentiel négatif des couches superficielles. Donc cette eau est une eau de mélange,

— cette eau de mélange écoule probablement dans la rivière Tisza ou quelque autre eau de surface,

— sur la partie de à peu près 300 km entre Polgár et Algyő on a pu constater l'accroissement du débit d'à peu près $17 \pm 5{,}5$ m^3/sec, sur la base de nos valeurs de tritium.

Cette étude n'est que la permière étape d'une investigation plus étendue, visant à étudier les systèmes d'écoulement intermédiers de la Grande Plaine Hongroise se servant d'isotopes environmentaux.

The most of the domestic and industrial water demands is met by using subsurface waters in the Nagykunság region. It is extremely important, therefore, to know the processes influencing the flow and replenishment of the groundwater. Based on hydrogeological methods, the following conclusions have been achieved previously [1, 2]:

— in the regions having a negative gradient of energy potential (the territory between the Danube and Tisza Rivers, the Nyírség, the Deliblát and the piedmont regions) the infiltrating water flows towards the central deep-lying areas of the Great Plain;

— in the regions having a positive gradient of energy potential (e.g. the Nagykunság), the deep groundwater moves upwards, and—after having been mixed with the shallow groundwater—it evaporates;

— the shallow groundwater in the Nagykunság area is hardly recharged by precipitation, its overwhelming part originates from the deep groundwater percolating upwards [2].

The purpose of our investigations using radioisotope techniques was to clear — the recharge conditions of deep groundwaters in the Nagykunság region; and — the connections between the shallow and deep groundwaters [3].

The main characteristics of the environmental isotopes and the possibilities of their application in the research of subsurface waters have been already dealt with in earlier papers [3, 4], therefore, only the conclusions drawn up from the data of isotope analyses are presented in this paper.

The investigation concerning the upward movement of deep groundwaters in the Nagykunság region using the radiocarbon waterdating method started in 1975. For the time being, age data for deep groundwater samples taken only from two wells of different depths in the Törökszentmiklós area are available. It is to be mentioned here that all radiocarbon age data presented in this paper have been corrected on the basis of the $^{13}C/^{12}C$ ratio (δ_{13})*, using the following formula:

$$t = \frac{5730}{\log 2} \log \frac{100\%}{A \cdot K} \text{[years]} \tag{1}$$

where A is the specific ^{14}C activity of sample [%], and

$$K = \frac{-25‰}{(\delta_{13})_{\text{sample}}}$$

In the Törökszentmiklós area, the water originating from the 390 — 405 m deep strata is younger (10,400 ± 2000 years) than the one originating from a shallow well perforated between the depths of 74 and 87 metres (14,500 ± 1000 years), consequently, the upward movement of deep groundwaters in the Törökszentmiklós area can be supposed to be probable.

The study of upward movement of deep groundwater is being continued in the Nagykunság region. It is expected that not only the existence of the upward flow will be successfully cleared up, but also the calculation of its velocity and discharge will be possible.

The method of $^{13}C/^{12}C$ stable isotope analysis was selected for the determination of the recharge area of deep groundwaters in the Nagykunság region. According to Münnich's model [5], the infiltrating waters receive their final HCO_3^- and CO_2 content in the upper 8 — 10 m deep soil layer and these parameters do not change significantly during the process of percolation. The $^{13}C/^{12}C$ isotope ratio remains constant similarly, thus this value is characteristic for the given area. For example, the stable carbon-isotope composition of deep groundwater percolating through the Upper Pannonian strata in the piedmont of the Mátra Mountain towards the Great Plain is in agreement with that of samples taken from shallow groundwaters recharged by the same outcropping

*

$$\delta_{13} = \frac{\left(\frac{^{13}C}{^{12}C}\right)_{\text{sample}} - \left(\frac{^{13}C}{^{12}C}\right)_{\text{PDB}}}{\left(\frac{^{13}C}{^{12}C}\right)_{\text{PDB}}} \cdot 1000 [‰]$$

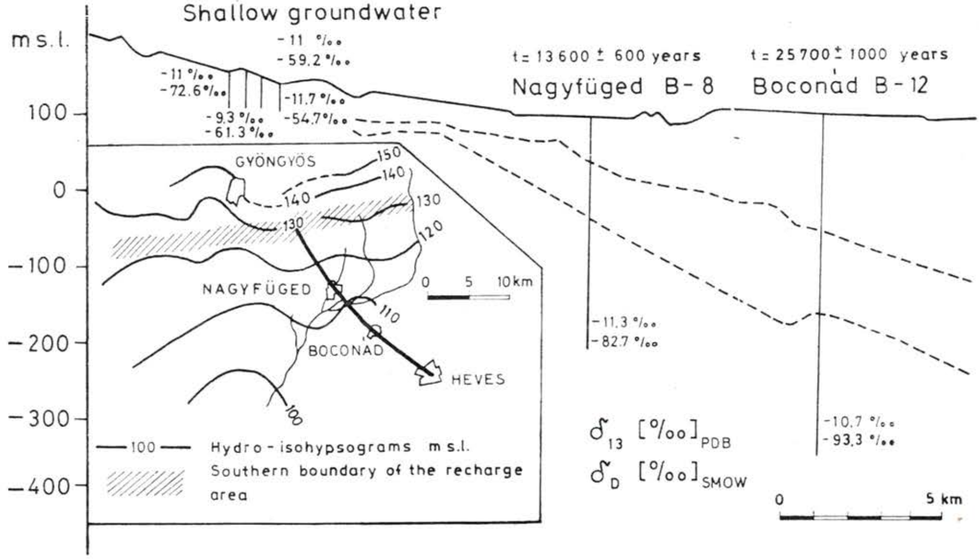

Fig. 1: Stable carbon- (δ_{13}) and hydrogen- (δ_D) isotope composition of shallow and deep groundwaters in the piedmont of the Mátra Mountain [8]

strata (Fig. 1), however, the deuterium (δ_D)* data of the same deep groundwaters—presumably due to a climatic change about 12,000 years ago—are significantly more negative than the average D/H isotope ratios of the shallow groundwaters. (The average deuterium concentration of precipitation is proportional to the local annual mean temperature [6].) This practical example demonstrates also, that the δ_{13} value of the percolating water does not change significantly.

In samples originating from groundwaters deeper than 80 m (Fig. 2) in the Nagykunság region and analysed so far, the characteristic δ_{13} value was found around $-20‰$, therefore, an area was sought, where the shallow groundwaters have got similar carbon-isotopic composition. The shallow groundwater samples collected in the assumed recharging areas in Hungary, however, did not show δ_{13} values of $-20‰$ (Fig. 3). The most negative values (-14 to $-15‰$) were found in samples taken in the Nyírlugos area as well as in Tarnazsadány and in Zaránk. Consequently, the further investigations concerning the determination of the recharging region are going to be concentrated on these areas and on the loess-ridge at Debrecen.

The discharge of deep groundwaters moving upwards in the Nagykunság region was investigated with the help of data obtained in the course of radioactive hydrogen and carbon-isotope analyses.

The basis of tritium (^{3}H) studies was the empirical fact that the ^{3}H concentration of precipitation has shown a sudden increase in the last 20 years,

*

$$\delta_D = \frac{\left(\frac{D}{H}\right)_{\text{sample}} - \left(\frac{D}{H}\right)_{\text{SMOW}}}{\left(\frac{D}{H}\right)_{\text{SMOW}}} \cdot 1000[‰]$$

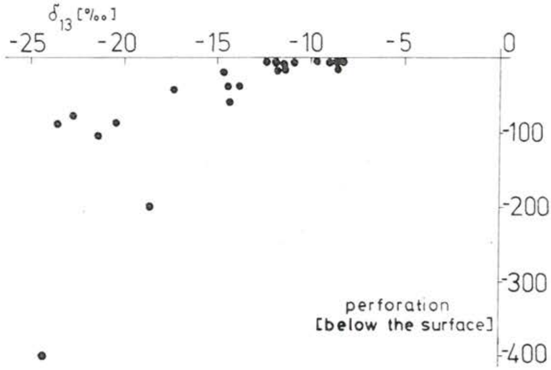

Fig. 2. Stable carbon-isotope composition (δ_{13}) of shallow and deep groundwaters in the Nagykunság region

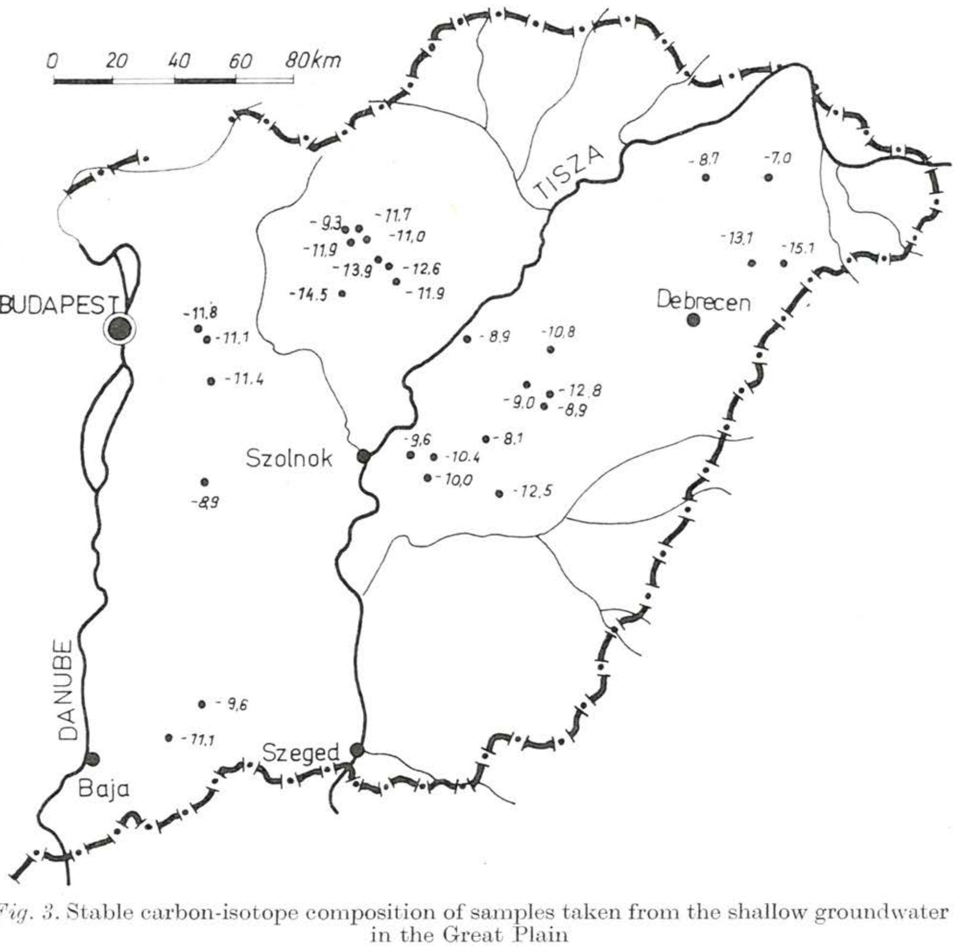

Fig. 3. Stable carbon-isotope composition of samples taken from the shallow groundwater in the Great Plain

due to thermonuclear bomb tests and due to operation of atomic reactors (Fig. 4). The tritium concentration of precipitation-waters is practically zero (background), therefore, they can be separated from waters originating from more recent precipitations. Areas having shallow groundwater with low ^{3}H concentration can not be found in the Nagykunság region, where the mixing of the water in the first unconfined aquifer with artesian waters of definite zero tritium content could be assumed (Fig. 5). The ^{3}H concentration of shallow groundwaters is influenced by the feature of the topmost layer (Fig. 6.) The high ^{3}H concentration in the case of sandy soil, and the low tritium concentration below clay unambiguously demonstrates the rapid recharge of shallow groundwaters from precipitation in the first case and the slowness of the process through impervious covering layer. The relatively low ^{3}H concentrations obtained in areas covered by loess or clayey loess top layer can be explained by the capillary storage theory [7]. All this considerations support the opinion that the primary source of recharge for the shallow groundwaters in the Nagykunság region is the precipitation.

The values of the D/H stable hydrogen isotope ratio, (δ_D), in the Nagykunság can be interpreted in a similar way (Fig. 7). The average deuterium concentration of the shallow groundwaters, (–63‰) is in good agreement with the multi-annual average of the deuterium concentrations found in the precipitation in Vienna (–68‰), and remarkably deviates from the average value of δ_D in the 50–100 m deep Upper Pleistocene artesian waters (–85‰). This fact indicates also, that the shallow groundwater in the Nagykunság region receives its recharge mostly from precipitation, and the recharge from artesian waters cannot be proved.

The interpretation of data obtained by the $^{13}C/^{12}C$ stable carbon isotope analyses leads to the same conclusion. The δ_{13} values of the shallow groundwaters of the Nagykunság (Fig. 2) are in good agreement with those measured in samples taken from other regions of the Great Plain, varying between –8 and –13‰ in the former and between –7 and –15‰ in the latter case (Fig. 3), however, they deviate significantly from the values between –18

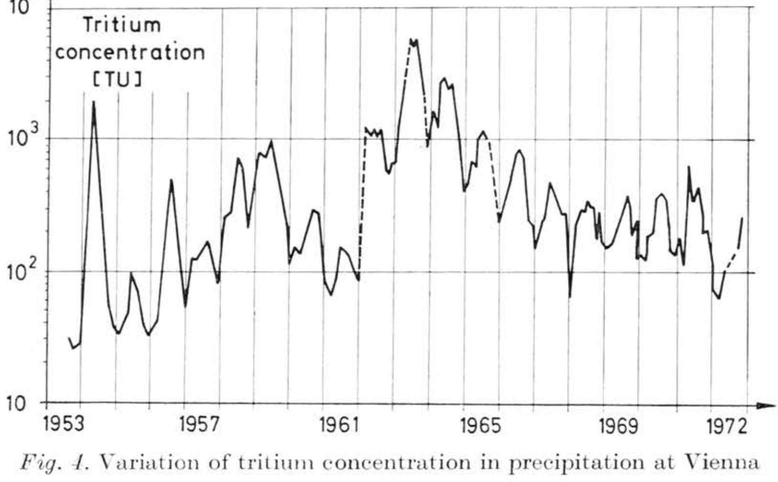

Fig. 4. Variation of tritium concentration in precipitation at Vienna

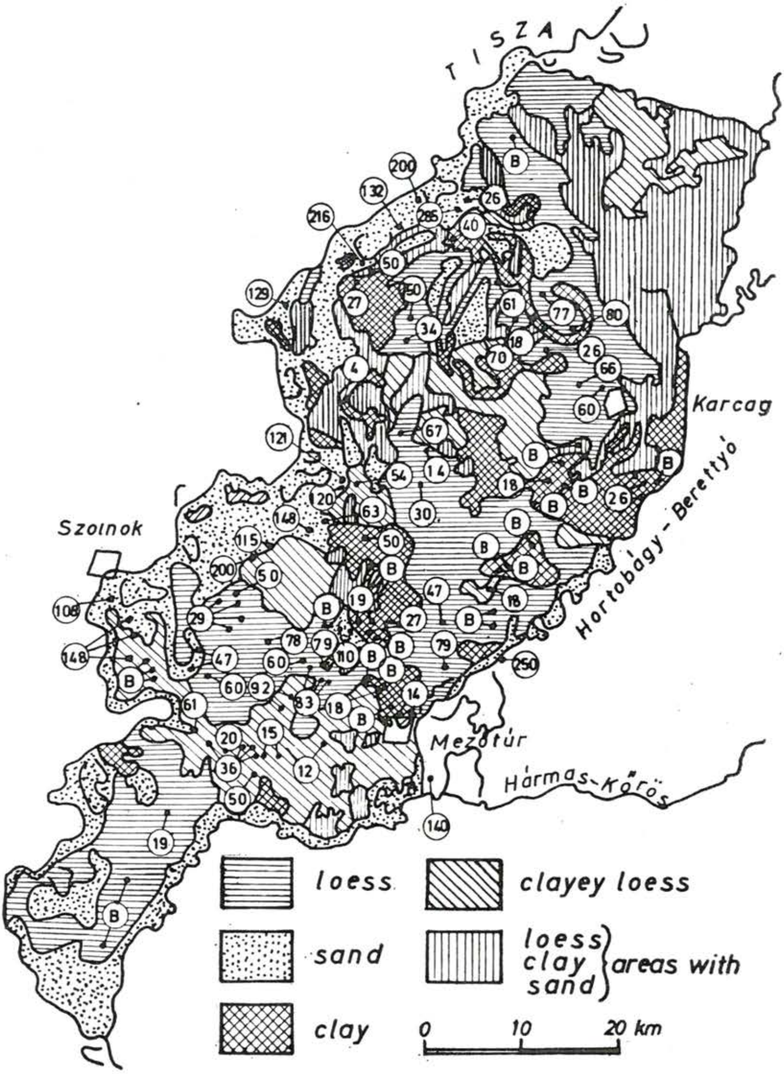

Fig. 5. Tritium concentration of shallow groundwaters in the Nagykunság region

and $-24‰$, which are characteristic for groundwaters deeper than 60 m. Thus the data of stable carbon isotopes demonstrate also, that the shallow groundwater in the Nagykunság region—like the ones in the piedmont of the Mátra mountains, in the Nyírség area and in the territory between the Danube and Tisza Rivers—gets their recharge from precipitation, and no component of artesian origin can be traced in them.

According to our investigations carried out so far, the deep groundwaters moving upwards in the Nagykunság region do not rise up to the shallow groundwater but are collected presumably by the sandy layer lying between 20 and 60 m depths and then are discharged into the Tisza River or into another stream. The δ_{13} data of 20–60 m deep groundwaters found to vary between -14 and -17‰ (Fig. 2) form a transition from the stable carbon isotope concetration of the shallow groundwaters to the one of the artesian waters. Since the $^{13}C/^{12}C$ ratio remains unchanged during the process of subsurface flow [5], these intermediate values can only be interpreted in such a way that the 20–60 m deep groundwater is a mixture of the descending shallow groundwaters and the artesian waters moving upwards. This assumption is also supported by the water-age data obtained by radiocarbon analyses carried out so far (Table 1).

The increase of average age of water in direct proportion to the depth can not be attributed to the slow (4 mm/year velocity, calculated on the base of age data) percolation of the shallow groundwater towards the 80–100 m deep groundwater. Such a deep percolation would be inconsistent partly with the change of the δ_{13} values with depth, partly with the fact that in all the three investigated subareas (Szolnok, Karcag and Kisújszállás) (Fig. 8), the minimum piezometer levels were observed in the 20–60 m deep horizon (Figs 9, 10 and

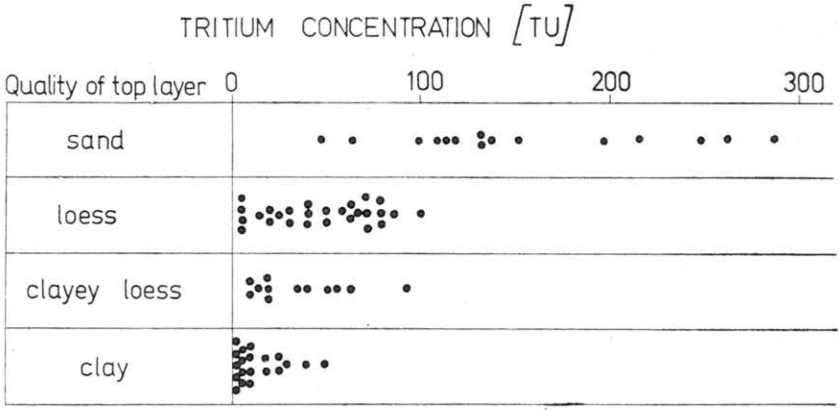

Fig. 6. The average tritium concentration of shallow groundwaters in the Nagykunság region as function of the soil type of the top layer [8]

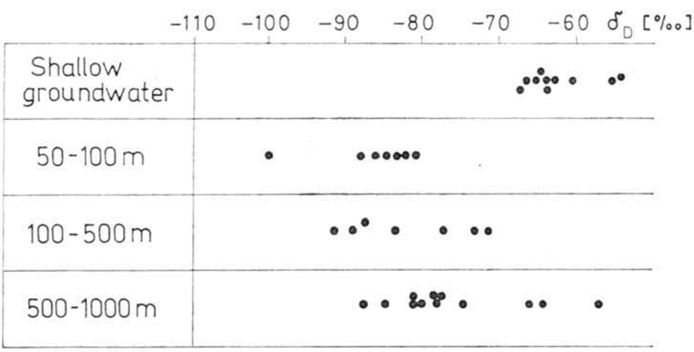

Fig. 7. Deuterium concentration of shallow and deep groundwaters in the Nagykunság region [8]

Table 1

Sampling site	Depth of perforated segment (under rim of well) m	δ_{13} $(‰)_{PDB}$	^{14}C $(\%)_{modern}$	Age (years) (corrected for ^{13}C)	Average age (years)
Karcag	< 10	−10.5	48.7	900 ± 500	
Karcagpuszta	< 10	−12.8	49.8	250^{+300}_{-250}	500 ± 500
Kenderes	< 10	−11.5	43.8	400^{+900}_{-400}	
Szolnok K−41	42−46	−17.2	28.9	7100 ± 900	
Karcag K−140	25−29	−14.8	30.1	5300 ± 500	5400 ± 700
Kisújszállás K−78	32−44	−13.9	34.2	4000 ± 800	
Törökszentmiklós B−42	74−87	−20.2	14.0	14.500 ± 1000	
Karcag K−129	75−86	−23.7	12.8	16.500 ± 1300	17.300 ± 1500
Kenderes K−26	60−130	−20.4	6.5	21.200^{+4000}_{-2700}	

11). The piezometric level of the 20−60 m deep groundwaters is lower than the levels of both the shallower and the deeper groundwaters, consequently, this stratum is draining both of the latter strata, and its water is a mixture of the shallower and deeper groundwaters. The age obtained for this mixed water, 5400 years, cannot be regarded as actual age, it is a weighted average of the 17,300 and 500 years age values.

It is assumed that the 20−60 m deep groundwaters are discharged into surface streams (in the first place, into the Tisza). In order to verify this assumption, the variation of tritium content was investigated in a reach of the Tisza River between Polgár and Algyő [8]. The tritium balance equation can be written as follows:

$$Q_A \cdot \mathrm{T}_A + Q_m \cdot \mathrm{T}_m + Q_t \cdot \mathrm{T}_t + \sum_{i=1}^{n} Q_i \cdot \mathrm{T}_i = Q_B \cdot \mathrm{T}_B \tag{2}$$

where Q = discharge [m³/sec]

T = tritium concentration [TU]

A, B = upstream and downstream sections of the investigated river reach

m = artesian water

t = shallow groundwater

$1,2 \dots n$ = tributaries

Considering that Székely [9] demonstrated that along the investigated reach of the Tisza River the contribution of shallow groundwaters to the streamflow is negligible ($Q_t = 0$), and that the tritium concentration of artesian waters recharging the river is zero ($\mathrm{T}_m = 0$), further that

$$Q_B = Q_A + Q_m + Q_t + \sum_{i=1}^{n} Q_i, \tag{3}$$

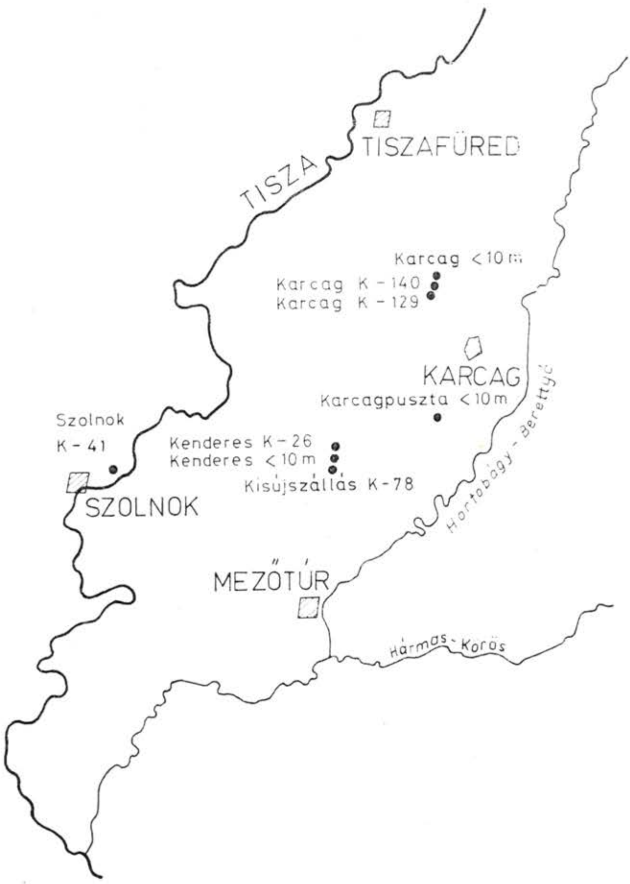

Fig. 8. Sampling sites for radiocarbon water-age determination

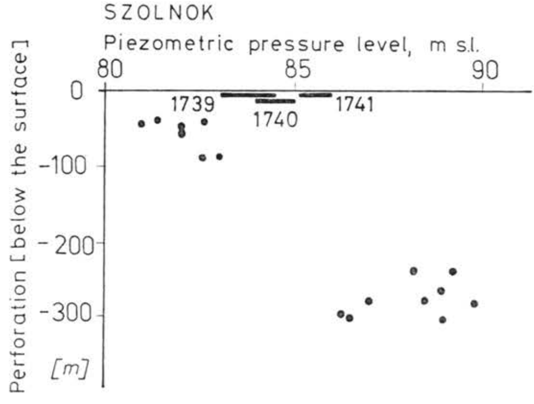

Fig. 9. Piezometric pressure levels of artesian waters at Szolnok in different depths and in groundwater observation wells No. 1739, 1740 and 1741

Equ. *(2)* takes the following form:

$$Q_m = Q_A \cdot \frac{T_A - T_B}{T_B} + \sum_{i=1}^{n} Q_i \cdot \frac{T_i - T_B}{T_B} \, [\text{m}^3/\text{sec}]. \tag{4}$$

Reasonably, the sampling procedure was carried out at flows of the Tisza River as low as possible, because during such periods the ratio of inflowing yield of artesian waters (Q_m) to the discharge Q_A is the highest. The tritium concentration of samples taken on 1 and 2 February 1973 between Polgár and Algyő from the Tisza showed a linear decrease (Fig. 12). Taking into

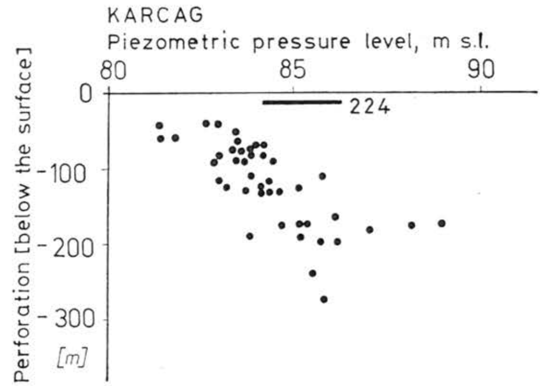

Fig. 10. Piezometric pressure levels of artesian waters at Karcag in different depths and in the groundwater observation well No. 224

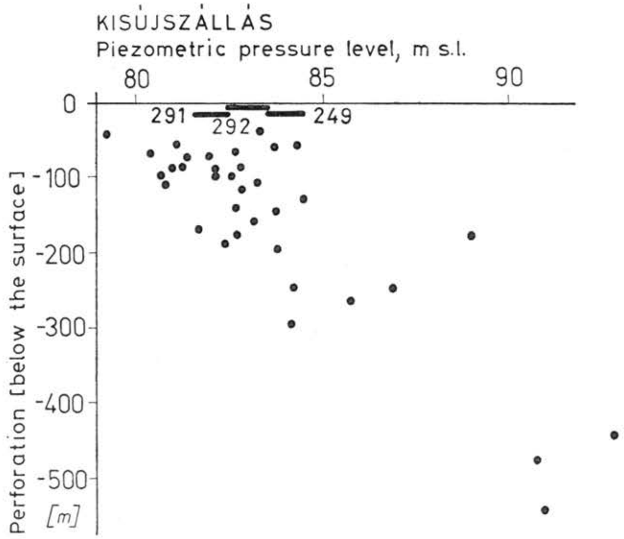

Fig. 11. Piezometric pressure levels in artesian waters at Kisújszállás in different depths and in groundwater observation wells No. 249, 291 and 292

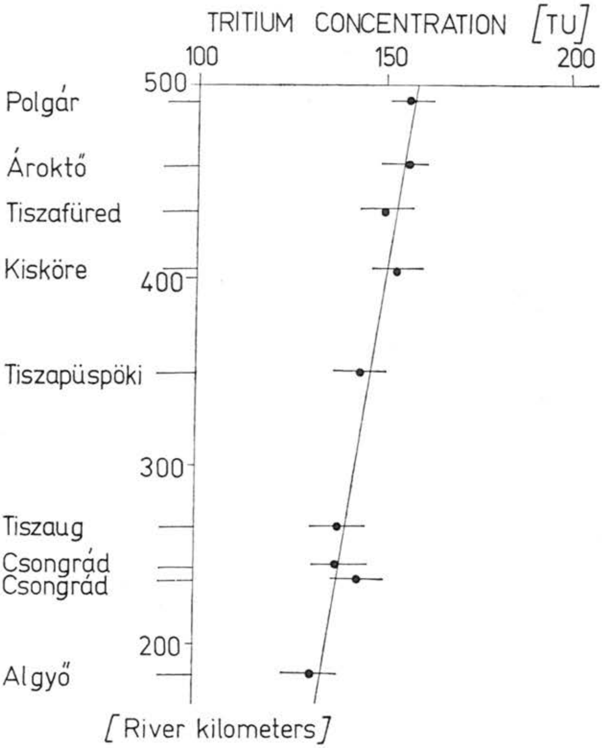

Fig. 12. Tritium concentration in the Tisza River between Polgár and Algyő

consideration the 110 m³/sec discharge at Polgár, the yield of deep groundwaters was calculated using this decrease:

$$Q_m = 17 \pm 5.5 \text{ m}^3/\text{sec} \tag{5}$$

(The error was evaluated by the least-squares method, applying the criterion of 1σ.) Due to the lack of extreme low flows in the Tisza River, the sampling could not be repeated.

Summarizing the results of all isotope analyses, the following conclusions can be drawn:

– the shallow groundwaters investigated in the Nagykunság region are not recharged by artesian waters, quite the contrary, they are percolating downwards, under the influence of the negative gradient of potential in the near-surface layers;

– the 20–60 m deep strata collect both the artesian waters moving upwards and the shallow groundwaters percolating downwards;

– this 20–60 m deep stratum discharges the collected mixed water into the Tisza River (or possibly into other surface streams).

It is planned to extend these isotope investigations to the entire area of the Great Plain. The main goal is to study the intermediate subsurface flow system of the Hungarian Basin with the help of environmental isotope analyses, as well as to clarify the subsurface flow conditions.

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LOIS DE FORMATION DES RÉSERVES EXPLOITABLES DES EAUX SOUTERRAINES DES BASSINS ARTÉSIENS DU TYPE DE PLATE-FORME ET QUELQUES PARTICULARITÉS DE LEUR ÉVALUATION

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Les eaux artésiennes des bassins du type de plate-forme jouent un rôle considérable comme des sources d'alimentation en eau potable des villes et des régions industrielles de l'U.R.S.S. (principalement de la partie européenne du pays). A l'heure actuelle dans les bassins artésiens des plates-formes sur le territoire de l'U.R.S.S. plus de 800 de gisements et de terrains isolés avec des réserves exploitables d'eau souterraine de l'ordre de 300 m^3/s sont explorés. Le plus grand nombre de gisements des eaux souterraines de ce type est mis au point sur le territoire des bassins de Moscou, de Dniepr-Donetz, d'Azovo-Kouban, dans les régions de la Mer Noire et des républiques baltiques, de la partie sud de la Sibérie Occidentale.

Les eaux souterraines de ces bassins artésiens sont largement utilisées pour l'alimentation en eau des villes des régions centrales de la Fédération de Russie, d'Estonie, de Lituanie, de Lettonie, de la majorité de l'Ukraine, de Biélorussie et de Moldavie.

Les sources principales de formation des réserves exploitables des eaux souterraines des bassins artésiens du type de plate-forme sont:

a) réserves élastiques de la nappe en exploitation;
b) réserves naturelles (y compris les réserves élastiques) des nappes contiguës;
c) réserves élastiques des couches intermédiaires peu perméables;
d) réserves naturelles de la nappe aquifère exploitée dans la zone de décharge où la nappe est libre;
e) ressources naturelles de la nappe aquifère;
f) les eaux superficielles*.

L'importance de toutes ces sources de formation des réserves n'est pas la même dans des conditions naturelles différentes, elle est conditionnée, principalement, par les particularités de la structure géologique et hydrogéologique du territoire étudié. D'après les conditions de formation des réserves exploitables des eaux souterraines les gisements des bassins artésiens du type de plate-forme peuvent

* Dans le rapport on comprend sous les « réserves naturelles » le volume d'eau dans la nappe et sous les « ressources naturelles » — l'afflux d'eau annuel dans la nappe en conditions naturelles.

être divisés en deux sous-types: *a)* gisements des eaux souterraines, éloignés considérablement des limites du bassin, pendant leur exploitation les cônes de dépression ne touchent pas les limites de la nappe; *b)* gisements des eaux souterraines en bordure des bassins artésiens.

L'analyse des résultats de l'exploitation des eaux souterraines des bassins artésiens des plates-formes montre que les gisements du premier sous-type peuvent être à leur tour divisés en deux groupes d'après le rôle prédominant de telles ou telles sources: *a)* gisements dont les réserves exploitables ne sont dues qu'aux réserves élastiques de la nappe aquifère, les autres sources d'alimentation étant absentes ou bien leur rôle étant négligeable; *b)* gisements dont les réserves exploitables sont dues à un fort degré à la drainance de l'eau des nappes contiguës à travers les « fenêtres hydrogéologiques » dans les couches imperméables ou peu perméables, ainsi qu'aux réserves élastiques ayant dans la plupart des cas une signification subordonnée.

Les gisements du premier groupe sont associés en règle générale aux nappes ou complexes relativement profondes (100—300 m et plus) bien isolés des nappes aquifères sus-jacentes par les couches imperméables ou peu perméables sur toute la surface d'extension de la nappe d'eau. L'exploitation de ces gisements s'effectue en régime non permanent. La prise d'eau est suivie de la formation des grandes cônes de dépression dont les rayons atteignent 100 km et plus, et l'abaissement des niveaux dans les parties centrales des dépressions est de 60—80 m et plus. A titre d'exemple on peut citer les gisements du Crétacé inférieur du bassin artésien de Dniepr-Donetz, des sédiments cimmériens du bassin Azovo-Kouban, des sédiments de l'étage de Gdov dans la région des républiques baltiques, etc.

Comme il est indiqué ci-dessus, les réserves élastiques constituent la source principale de la formation des réserves exploitables pour les gisements de ce groupe. En somme, les réserves élastiques des bassins artésiens ne sont pas importantes. Par exemple, pour le rabattement du niveau piézométrique de 100 m, grâce aux propriétés élastiques de l'eau et des roches et pour la valeur caractéristique du coefficient d'emmagasinement de l'ordre de 0.001 on peut extraire une tranche d'eau à l'épaisseur de 2—5 mm pendant 20—50 ans d'exploitation. Cependant, pour les ouvrages de captage isolés l'utilisation des réserves élastiques a une importance décisive (en rapport avec une surface immense du développement de la dépression dans le cadre de laquelle les réserves exploitables des eaux souterraines se forment). Comme exemple, on peut citer la nappe du Crétacé inférieur dans le bassin artésien de Dniepr-Donetz, où les réserves exploitables dont la formation est due aux réserves élastiques atteignent des centaines de milliers m^3/jour et sont utilisées pour l'alimentation en eau des grandes villes.

Pour les gisements entrant dans le deuxième groupe du premier sous-type les réserves et les ressources naturelles des nappes contiguës sont les sources principales de formation des réserves exploitables des eaux souterraines, les réserves élastiques de la nappe exploitée et des couches intermédiaires de faible perméabilité ayant en général une signification subordonnée.

La plupart des gisements se rapportant à ce groupe, sont caractérisés par la faible profondeur des nappes. L'exploitation de ces gisements s'effectue en régime non permanent, mais la vitesse du rabattement par rapport à celle des gisements du premier groupe se ralentit considérablement. Les dimensions des cônes de dépression se formant lors de l'exploitation sont aussi moins grandes.

Tout ceci est dû au rôle prédominant des processus de drainance des eaux souterraines des nappes contiguës dans la nappe en exploitation, soit à l'existence de l'alimentation supplémentaire des nappes au cours de l'exploitation.

Actuellement, l'examen des données expérimentales montre que la drainance à travers les couches intermédiaires peu perméables s'effectue par les « fenêtres lithologiques », ainsi que directement par les roches de faible perméabilité si elles sont fissurées car la perméabilité interstitielle de ces roches déterminée en laboratoire ne peut pas assurer la drainance en grand observée dans la nature. L'expérience d'exploitation des ouvrages de captage montre que les processus de drainance ont lieu comme dans les régions avec de grandes « fenêtres hydrogéologiques » (par exemple, la vallée de dénudation dans le bassin artésien de Moscou près de la ville de Moscou, dans le bassin artésien des républiques baltiques près de la ville de Riga), ainsi que sur les surfaces, où il n'y a pas de discontinuité visuelle des sédiments. De ce fait, l'étude spéciale des phénomènes de drainance des eaux souterraines en conditions naturelles est nécessaire pour l'évaluation des réserves exploitables.

Les gisements du deuxième sous-type des bassins artésiens des plates-formes situées en bordure de ceux-ci sont aussi caractérisés par les conditions complexes de formation des réserves exploitables des eaux souterraines. Ici, avec les réserves et les ressources naturelles des nappes contiguës pénétrant dans la couche aquifère exploitée lors de drainance, les ressources naturelles de celle-ci, les eaux superficielles et le drainage des roches aquifères aux endroits de leur affleurement jouent un rôle important dans la formation des réserves exploitables. On peut attribuer à ce sous-type des petits bassins artésiens où pendant l'exploitation des ouvrages de prise d'eau la dépression a envahi pratiquement toute la surface du bassin. Les gisements du deuxième sous-type se caractérisent par les conditions les plus favorables de formation des réserves exploitables des eaux souterraines.

A titre d'exemple, on peut examiner les lois de formation des réserves exploitables des eaux souterraines d'un grand gisement situé en bordure du bassin artésien du Dniepr-Donetz dans la vallée du fleuve Dniepr. Les nappes principales sont associées aux sédiments cénomaniens, calloviens et bajociens. Les sédiments de la nappe supérieure principale (cénomaniens, calloviens) sont représentés par les sables de différente granulométrie avec des intercalations des grès, aléurolithes, calcaires, argiles. L'épaisseur totale de la nappe est de l'ordre de 30 – 40 m, la profondeur du toit atteint 110 – 120 m. La plus grande partie de la surface d'extension de la nappe est recouverte par les marnes et les craies dont la puissance est de l'ordre de 40 – 50 m en certains endroits de l'emplacement des ouvrages de captage. Au-dessus de cette assise l'aquifère de l'étage de Boutchak se trouve et il est représenté par les sables fins épais de 30 – 40 m. Cette couche aquifère est séparée à son tour des couches alluviales quaternaires sus-jacentes (dans la vallée du Dniepr) et des couches aquifères des étages de Kharkov et de Poltava (sur les lignes de partage d'eau) par les sédiments marneux de l'étage de Kiev à l'épaisseur de 20 m. Dans la vallée du fleuve Dniepr les marnes de l'étage de Kiev sont délavées et les sables alluviaux reposent directement sur les sables de l'étage de Boutchak.

A la base des sédiments cénomaniens, calloviens l'assise des argiles bathoniennes, épaisse de 100 m, repose séparant les nappes principales. La nappe inférieure principale est associée aux sables bajociens à l'épaisseur de l'ordre de 20 – 30 m. Son toit se trouve à la profondeur de 200 – 220 m.

L'exploitation de ces deux nappes pendant plusieurs années a abouti à la formation des grandes dépressions. Le rayon de dépression dans la nappe inférieure est de l'ordre de 40–60 km (l'exploitation s'effectuait en régime permanent), dans la nappe supérieure la dépression a été localisée dans la région de Kiev (en rayon de 10–15 km) en régime pratiquement stationnaire de la filtration. Ce phénomène peut être expliqué par les conditions favorables de l'alimentation de la nappe supérieure par l'assise de faible perméabilité des marnes et des craies. Cette explication est confirmée par le fait de formation d'une petite dépression dans la nappe non exploitée de l'étage de Boutchak qui est en communication avec les eaux superficielles dans la vallée du fleuve Dniepr.

L'analyse des lois de formation des réserves exploitables des eaux souterraines associées aux sédiments cénomaniens, calloviens et bajociens, effectué sur le modèle analogique USM-1 a montré qu'après déjà quelques années d'exploitation le rôle du ruissellement de surface et des ressources naturelles en eaux souterraines est devenu prédominant dans le bilan total des réserves.

Si pour la nappe supérieure cette conclusion n'est pas innattendue, elle est au moins surprenante pour la nappe bajocienne où, d'après l'analyse, la formation des réserves exploitables pour 70–75% est aussi due à la drainance des nappes sus-jacentes.

Ce mécanisme de formation des réserves n'est pas logique pour les structures géologiques mentionnées car cette nappe est séparée de celle sus-jacente par l'assise des argiles bathoniennes d'une épaisse de 100 m. Cependant, l'analyse de l'exploitation a montré précisément l'importance dominante des processus de drainance. Si on ne les prend pas en compte, le rabattement des niveaux et leurs valeurs absolues doivent fortement dépasser les valeurs observées en conditions naturelles.

L'existence des processus de drainance a été démontrée aussi par les résultats de l'analyse de la teneur du carbone radioactif dans les eaux souterraines renfermées dans les deux nappes témoignant leur rajeunissement dans la vallée du Dniepr, région de l'exploitation la plus intense, ainsi que par les recherches hydrogéologiques.

L'exemple cité montre qu'il est indispensable de prendre en considération les processus de drainance des eaux souterraines dans les bassins artésiens des plates-formes et d'étudier la perméabilité des sédiments intermédiaires. Cependant l'étude de ces processus et la détermination des caractéristiques des couches intermédiaires pendant la prospection d'après les données des essais de pompages ne sont pas réelles dans la plupart des cas, car lors même des pompages intenses et de longue durée ces processus ne peuvent pas être fixés. Ceci a lieu parce que pour une faible perméabilité des couches intermédiaires les lois des variations des niveaux de long terme (parfois mesuré par les années) sont influencées par les processus de drainance. C'est pourquoi, pour l'évaluation des réserves exploitables des eaux souterraines des bassins artésiens des plates-formes on obtient les meilleurs résultats en faisant l'analyse des données de l'exploitation de plusieurs années et en utilisant la méthode d'analogie hydrogéologique. Aussi, une grande importance revient-elle dans l'utilisation des méthodes d'étude de la drainance comme du processus de transfert de masse et de chaleur pour l'argumentation scientifique des conditions de communication entre les nappes.

La méthode la plus connue dans ce groupe est la méthode hydro-géothermique basée sur l'analyse de la répartition de la température sur la coupe du système aquifère considéré. Elle permet de mettre au point les endroits à degré de communication des nappes différent et d'évaluer la vitesse et la direction du mouvement des eaux souterraines à travers les couches intermédiaires.

La méthode hydrochimique est moins répandue pour la résolution des problèmes. Elle consiste à l'analyse de la répartition sur la coupe de la couche peu perméable de différents composants chimiques en vue de l'étude du degré de lixiviation de l'assise isolée. Les premiers résultats des recherches montrent que cette méthode est très perspective.

Comme il a été mentionné plus haut, de bons résultats des recherches des conditions de communication entre les nappes peuvent être obtenus lors de l'étude des variations de la teneur dans les eaux souterraines des indicateurs isotopes stables et radiogènes en surface et en coupe qui peuvent être considérés en qualité d'indicateurs naturels de la drainance. Pour l'analyse des conditions de communication hydraulique on peut utiliser aussi des indicateurs artificiels (électrolytes, colorants et fluorescéines).

Or, la caractéristique des lois de formation des réserves exploitables des eaux souterraines des bassins artésiens du type de plate-forme montre que pour leur évaluation il est nécessaire de tenir compte soit de l'interaction d'un grand nombre des ouvrages de prise d'eau occupant une surface considérable (pour les gisements du premier groupe du premier sous-type), soit de l'interaction de plusieurs nappes (pour les gisements du deuxième groupe du premier et du deuxième sous-type). C'est pour cette raison que l'utilisation de la méthode de simulation mathématique pour l'évaluation des réserves d'eaux souterraines en qualité de méthode principale devient indispensable. Il est à souligner qu'en rapport avec les particularités caractéristiques citées ci-dessus l'évaluation des réserves exploitables des eaux souterraines des bassins artésiens est un problème régional pour la résolution duquel il faut tenir compte des variations des propriétés de filtration et capacitives sur un territoire considérable. Puisque dans la majorité des cas en conditions naturelles une hétérogénéité remarquable des paramètres hydrogéologiques a lieu, les méthodes de simulation mathématique sont préférables par rapport aux méthodes analytiques. Il en découle encore une particularité de l'évaluation des réserves exploitables des eaux souterraines des bassins artésiens, notamment la nécessité de l'étude des paramètres non seulement sur les terrains de prospection, mais aussi sur la surface plus étendue, où les réserves exploitables des eaux souterraines doivent se former. Pour la résolution de ce problème et pour l'établissement des cartes des paramètres toutes les données sur la structure géologique du territoire en question doivent être analysées, y compris les données sur la variation de la composition lithologique-faciologique des roches aquifères et les modalités de la structure tectonique et géomorphologique de la région étudiée.

UTILISATION DES ÉCHANGEURS DE CHALEUR SOUTERRAINS DANS LES PUITS D'EAUX CARBONIQUES AFIN D'ÉVITER LA FORMATION DE TRAVERTIN

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Assurer la stabilité de la composition chimique des eaux carboniques thermales précieuses en médecine pour le traitement balnéaire offre un problème important pour leur exploitation dans les stations balnéaires, car la stabilité de leur composition chimique les préserve contre les précipitations des carbonates de calcium (formation de travertin) qui, comme on le sait, peut complètement boucher les puits et couper l'éruption des eaux carboniques.

Ce problème est devenu surtout important ces dernières années à la suite d'augmentation de la profondeur du sondage aux eaux minérales et d'utilisation par endroits des puits aux eaux superthermales qui arrivent de grandes profondeurs.

Dans les grandes profondeurs sous haute pression hydrostatique les eaux carboniques peuvent contenir à l'état de solution et même à la température très élevée des quantités considérables de CO_2 (des dizaines de g/l) qui en vertu du principe d'équilibre carbonique maintiennent dans la solution beaucoup d'hydrocarbonates de calcium.

Pendant l'ascension des eaux carboniques thermales dans le tube du puits, sous pression naturelle une partie considérable de CO_2 solubilisée sous l'effet de tension superficielle diminuée se dégage de la solution pour former la phase gazeuse des eaux.

Cela aboutit au phénomène, connu sous le nom de gaz-lift qui augmente beaucoup le jaillissement des eaux carboniques (augmentation du débit). En même temps cela trouble l'équilibre de carbonates existant à cette profondeur ce qui favorise les précipitations des carbonates de calcium, possibles dans les parties supérieures du tube aussi bien que dans son couronnement et sa tuyauterie.

Ce processus se traduit par l'équations :

$$2\,HCO_3^- \rightleftarrows CO_3^{2-} + CO_2 + H_2O$$

La formation de travertin à la sortie des eaux carboniques superthermales à la teneur même insignifiante en calcium se produit (changement à droite de l'équilibre carbonique déjà mentionné) dans des thermes carboniques à composition ionique assez variée (voir le Tableau 1).

Tableau 1

Composition chimique des eaux carboniques thermales formant des travertins

Pays source sondage	Température des eaux à la sortie °C	M g/l	Composition chimique	Teneur en g/l			
				solution CO_2	le tout CO_2	HCO_3^-	Ca^{2+}
U.R.S.S. Archan, sond. 28	43	4.0	$\frac{HCO_3\ 69\ \ SO_4\ 27}{Ca\ 60\ \ Mg\ 21}$	1.040	10.000	2.220	0.639
Djermouk, sond. I–k	62	4.2	$\frac{HCO_3\ 60\ \ SO_4\ 24}{(Na+K)\ 77\ \ Ca\ 15}$	0.470	5.100	1.950	0.156
Isti-Sou sond. I – a	54	7.4	$\frac{HCO_3\ 50\ \ Cl\ 34}{(Na+K)\ 90\ \ Ca\ 8}$	0.600	–	3.000	0.156
Nagoute sond. 9	50	12.2	$\frac{HCO_3\ 70\ \ Cl\ 29}{(Na+K)\ 92\ Ca\ 4}$	0.640	10.300	6.870	0.14
Tch. Carlovy-Vary, source Vrjidlo	71	6.0	$\frac{HCO_3\ 43\ \ SO_4\ 36\ \ Cl\ 21}{(Na+K)\ 87\ \ Ca\ 8}$	0.375	–	2.105	0.12

Il convient de noter que plus la température est élevée, plus la teneur en calcium est grande et plus brusque, enfin, la différence de pressions au cours de l'ascension des eaux montant des profondeurs à la surface, plus forte et plus intense sera la formation des travertins dans les zones proches de la tête de puits.

Le moyen le plus rationnel pour empêcher la formation des travertins consiste à maintenir la concentration de CO_2 nécessaire à l'équilibre de carbonates.

On obtient cet équilibre en augmentant la pression dans le système ou bien en baissant la température de l'eau dans la zone supérieure du puits.

Étant donné que l'augmentation de la pression sur la tête des puits est pratiquement impossible puisqu'elle amène la diminution considérable de leur débit, le seul moyen c'est de refroidir les eaux à l'aide des échangeurs de chaleur souterrains qui utilisent comme agent frigorifique les eaux douces froides.

Pour la première fois un puits pareil avec échangeur de chaleur fut proposé en 1950 par I. I. Kobozev, V. N. Mikhailov et A. F. Radine au cours de reconnaissances des sources thermales dans la station balnéaire Isti-Sou de la République Socialiste Soviétique d'Azerbaïdjan.

Cependant, l'installation des puits équipés d'échangeurs de chaleur fut effectuée pour la première fois seulement au cours des travaux de prospection hydrogéologique dans la station balnéaire de Djermouk de la République Socialiste Soviétique d'Arménie en 1961 – 66 sous la direction d'un des auteurs de cet article.

La ville d'eau Djermouk est une grande station balnéaire qui utilise les eaux carboniques thermales pour le traitement de bains et aussi pour l'usage médical interne. Djermouk c'est aussi une usine produisant des eaux minérales mises en bouteilles.

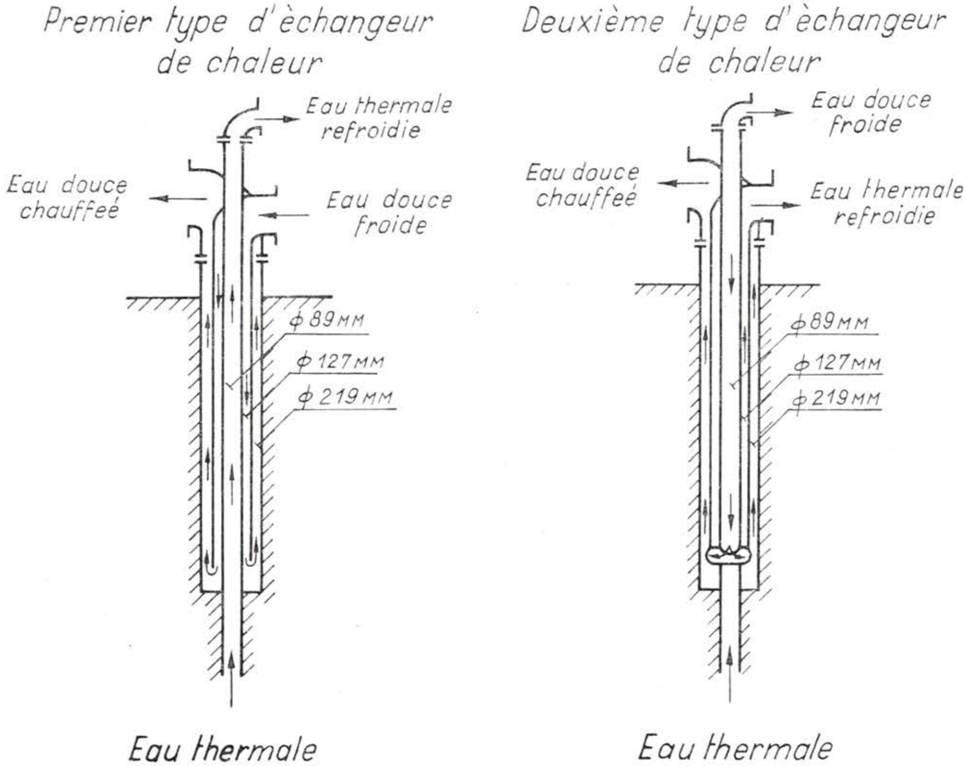

Fig. 1. Schéma des échangeurs de la chaleur utilisés dans les puits des eaux carboniques à la station balnéaire de Djermouk

Les eaux thermales de Djermouk sont des eaux carboniques siliceuses; elles contiennent des sulfates, des hydrocarbonates et du natrium avec la teneur en calcium élevée et une basse teneur en CO_2 solubilisé (450–550 mg/l) à la sortie.

Les travaux d'essai avec les échangeurs de chaleur ont été entrepris à Djermouk dans les puits I-OK et I-k aux eaux de composition presque identique qui ne se distinguent que par la température et la teneur en gaz totale (Tableau 2).

Dans les puits I-OK ont été essayés deux types d'échangeurs de chaleur-

Tableau 2

Composition chimique des eaux carboniques thermales des puits I—OK et I—k de la station balnéaire de Djermouk

N° du sondage	Température de l'eau °C	M g/l	H_2SiO_3 g/l	CO_2 solubilisé g/l	Teneur en gaz totale g/l
I–OK	55	3.8	0.080	0.540	1.8
I-k	62	4.2	0.090	0.470	5.1

refroidisseurs-souterrains ayant des systèmes de sabots et des schémas du mouvement des eaux thermales et des eaux douces et froides différents.

La structure des deux types d'échangeurs de chaleur est un ensemble de tubage composé de colonnes avec 219 mm, 127 mm, 89 mm de diamètres placées d'une manière concentrique dans les puits (selon le schéma « tube dans le tube »).

La partie active refroidissante de ses colonnes a la même longueur 37 m et un sabot spécial.

L'eau douce superficielle (à la température de 6°) destinée au refroidissement a été conduite vers le puits de la pente droite de la vallée de l'Arpe située à 25 m plus haut que la bouche du puits par les tuyaux à écoulement libre.

Dans le premier type d'échangeur de chaleur (dessin 1) l'eau thermale arrivait à la surface par la colonne centrale (89 mm); l'eau douce froide pénétrait du haut par l'espace annulaire entre les colonnes 89 mm et 127 mm pour arriver à la bouche du puits par le jeu périphérique entre les colonnes de 127 mm et 219 mm. Dans le second type d'échangeur de chaleur l'eau thermale montait par l'espace annulaire entre les colonnes de 89 et 127 mm en se refroidissant des deux côtés à l'aide de l'eau douce qui arrivait par la colonne centrale de 89 mm et rentrait à la bouche du puits par le jeu périphérique entre les tuyaux de 127 et 219 mm.

Une vanne réglait la quantité d'eau froide dont dépendait le degré de refroidissement des eaux carboniques.

Les paramètres initiaux des eaux thermales de Djermouk dans le puits I-OK étaient les suivants: température 56°, débit 3.07 l/sec, facteur de gaz 1.38. L'eau de charge au niveau dynamique négatif était amenée au jour sous l'action du gaz-lift naturel.

Dans le premier type d'échangeur de chaleur, l'eau thermale refroidissait de 55° à 35°; dans le deuxième type — de 55° à 40°, graduellement tous les 5° (55—50—45—40—35°).

Le refroidissement de l'eau thermale s'accompagnait par la diminution du facteur de gaz due à l'augmentation de la solubilité du gaz carbonique dans l'eau froide.

La chute de la température de 55° à 37° amenait la baisse du facteur de gaz de 55 % (de 1.38 à 0.62). La teneur en CO_2 solubilisé dans l'eau s'est accrue de 0.54 à 0.93 g/l (Tableau 3).

La saturation en gaz sommaire demeurait alors constante et oscillait dans les limites d'exactitude de mesurage (1.7—1.8 g/l). La baisse du facteur de gaz qui influe sur l'intensité du gaz-lift s'accompagnait de la diminution du débit d'eau (Tableau 3).

Comme il ressort du tableau 3, l'abaissement de la température de l'eau jusqu'à 45° s'accompagnait de la diminution graduelle et relativement insignifiante du débit (de 17 %) ; la baisse ultérieure de la température jusqu'à 40—37° a provoqué la chute brutale du débit presque d'un tiers (28—30%).

Au cours du refroidissement ultérieur (jusqu'à 35°) le jet était coupé et le jaillissement du puits cessait. Cela s'explique par le fait que la teneur en CO_2 spontané égale à la température de 35° à 0.8 g/l est pour ce puits la limite inférieure qui assure l'action du gaz-lift.

L'analyse comparative du fonctionnement des deux types d'échangeurs de chaleur a montré que si les autres conditions demeuraient les mêmes pour

Tableau 3

Changement de température de débit et de teneur en CO_2 dans l'eau carbonique thermale du puits I—OK au cours de refroidissement

Expérience conduite en conséquence directe					Expérience conduite en conséquence inverse		
Température de l'eau °C	Débit d'eau l/sec	Baisse du débit par rapport au débit initial %	Facteur de gaz	CO_2 solubilisé g/l	Température de l'eau °C	Débit d'eau l/sec	Augmentation du débit par rapport au débit initial %
55.7	3.07	–	1.38	0.54	40	2.6	–
50.2	2.89	6	0.95	0.61	45	2.83	9
45.5	2.54	17	0.89	0.69	50.5	2.91	12
40.0	2.20	28	0.82	0.79	55.5	2.94	13
37.0	2.16	30	0.62	0.93			

obtenir le même niveau de refroidissement de l'eau thermale, la dépense d'eau douce dans le deuxième type d'échangeur de chaleur diminue de 30—40%. Cependant la structure du premier type d'échangeur est plus simple, elle permet d'introduire aisément à l'intérieur du puits les outils, ce qui rend possibles le cas échéant des travaux de réparation sans démontage. Puisque la dépense de l'eau douce servant à refroidir les thermes de Djermouk n'était pas limitée, on a pu installer sur le puits de captage I-k l'échangeur de chaleur du premier type en vue de poursuivre les essais plus prolongés. Les travaux d'essai ont duré pendant 3 mois environ et ont abouti aux résultats analogues (voir le Tableau 4).

Avec le refroidissement jusqu'à 50°, le débit diminuait de 18.2%, à 30° de 45.6% du débit initial. La température ciritique était 25.7°, à laquelle le jaillissement du puits cessait.

Il découle des expériences faites que le refroidissement des eaux jusqu'à la température 45—50° est le meilleur pour l'exploitation des eaux thermales de Djermouk. A cette condition le débit des eaux diminuait relativement très peu et la formation des travertins n'avait pratiquement pas lieu.

Tableau 4

Changement de la température, du débit et de la teneur en CO_2 dans les eaux thermales carboniques du puits I—k au cours de refroidissement

Température des eaux °C	Débit des eaux l/sec	Diminution du débit à partir de l'initial %	Facteur de gaz	CO_2 solubilisé g/l
62.2	2.30	–	2.95	0.47
49.7	1.88	13.2	2.88	0.66
38.9	1.58	31.3	2.65	–
29.7	1.25	45.6	2.43	–
25.9	–	–	–	1.2

L'exploitation ultérieure au cours de deux ans du puits I-k équipé de l'échangeur de chaleur indiqué a prouvé que les précipitations des travertins quelque soit peu considérables n'eurent pas lieu dans le tube du puits. Tandis que le puits voisin 4/5 I devait être nettoyé des précipitations de travertin tous les 3–4 mois car il était complètement bouché.

Le gaz se dégageant du puits expérimental qui maintenait le gaz-lift était évacué dans un séparateur de gaz spécial; les eaux thermales plus riches en acide carbonique dilué, ce qui constitue un facteur médical important, étaient amenées au consommateur pour être utilisées à des fins médicaux.

Conclusions

1. L'utilisation de chaleur (refroidisseur) souterraine est le moyen le plus efficace contre la formation de travertin dans des puits d'eaux thermales carboniques.

2. Pour le refroidissement, les échangeurs de chaleur peuvent utiliser n'importe quelles eaux froides superficielles ou souterraines.

3. Au cas de quantité suffisante des eaux froides il est mieux de se servir d'échangeur de chaleur type I (système d'échangeur dans lequel les eaux carboniques arrivent par le tube central), sa structure plus simple permettant de faire des travaux de réparation dans le puits. Si les réserves d'eaux froides sont limitées, il convient d'utiliser l'échangeur type II (système où les eaux carboniques sortent par l'espace annulaire entre les tuyaux).

4. Pendant le jaillissement des eaux carboniques sous l'effet du gaz-lift, le degré de refroidissement des eaux dans l'échangeur de chaleur ne doit pas perturber le fonctionnement de ce dernier.

5. L'utilisation des échangeurs de chaleur souterrains permet d'éviter la formation de travertin, assure la teneur plus élevée en CO_2 et permet d'obtenir en plus des eaux douces chauffées.

HYDROGEOLOGIC FACTORS GOVERNING THERMAL WATER OCCURRENCES AND RECOVERY IN THE PANNONIAN BASIN

FACTEURS HYDROGÉOLOGIQUES CONDITIONNANT LES COUCHES PROFONDES DES EAUX THERMALES ET LEUR EXPLOITATION DANS LE BASSIN PANNONIEN

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RÉSUMÉ

La cuvette sédimentaire du Bassin Pannonien renferme les gisements d'eaux thermales les plus importantes en Hongrie. La plupart des puits thermaux (400 environ) ont drainé les sables aquifères du Pannonien supérieur. Ils constituent un système des réservoirs multiples contenant des nappes superposées et captives. Les dimensions des nappes aquifères sont très variés variant de petites lentilles jusqu'à des couches étendues. Les couches aquifères superposées ne communiquent guère entre elles. Une communication limitée se fait seulement exceptionellement dans certains endroits par des accidents stratigraphiques. (Le mouvement vertical des eaux souterraines ou « vertical leakage ».)

Les caractéristiques hydrogéologiques des couches aquifères comme la porosité efficace, coefficient d'emmagasinement, perméabilité, sont la fonction de la profondeur, diminuant progressivement dû aux processus diagénétiques. Dans les dépôts sédimentaires pliocènes les facteurs de l'écoulement souterrain et les paramètres de réservoir sont favorables ce que produit des conditions excellentes pour l'exploitation des ressources d'eaux thermales.

Introduction

Hungary is very rich in thermal water resources. The majority of thermal water occurrences is within the Neogene sedimentary sequence in the multipartite Pannonian Basin. More than a two-third part of the thermal water wells (about 400) are tapping Upper Pannonian (=Middle Pliocene) sand and sandstone formations. These confined and usually deap-seated aquifers have favourable specific yield and their flowing water temperature ranges up to the 100 Centigrades. These low-enthalpy thermal waters are utilizable not only for balneological but also for industrial and agricultural purposes. The balneological utilization has already an old tradition in Hungary and about 50 per cent of the thermal water wells supply different kinds of bath, swimming pool etc. The use of the geothermal energy—whose carriers are mainly the Neogene formation waters—has recently been developed to a relatively high level, especially in SE Hungary, where most favourable hydrogeologic conditions prevail. As a result, an extensive geothermal utilization is here available (district heating, greenhouses, PVC-tent and tunnel-, as well as soil-heating, drying, etc.). Moreover, there are some places with multi-purpose thermal water utilization.

Geologic framework of the thermal water provinces

The infrastructure of the Pannonian Basin is a rigid crystalline basement of Precambrian and Paleozoic age with superimposed Mesozoic carbonate rock complex. The development of the great Pannonian sedimentary basin began in the Tertiary, or to be specific in the Miocene period due to the Alpine orogenic movements. They resulted in a general subsidence of epeirogenic character, which was associated by significant crustal phenomena owing to magmatic i.e. volcanic processes. As a result, the crust was thinning, the mantle was situated relatively near the surface (the depth of Moho discontinuity is approx. 24–26 km). All this is assumed to be responsible for the regional positive geothermal anomaly existing in the Pannonian Basin.

Mechanics of the basin evolution

The Pannonian Basin as a representative intermontane or interarc basin was formed mainly during the Neogene by successive and continuous submergence. There was an interplay between depositional subsidence, uplift and erosion. Thus a close relation between sediments and tectonics can be recognized. The architecture of the overall depression indicates the existence of a combination of normal sedimentary basin and a primary dynamic one that endows it with a complex character.

The rate of the subsidence was different in space and time and, as a consequence, many sub-basins of different shapes and depths were formed. The maximum subsidence took place in the Pliocene.

Stratigraphic model

The sedimentary environment had developed progressively from early marine through lacustrine to late fluviatile character, and a combination of transgressive and regressive marine to lacustrine processes with intervening deltaic phenomena was operated, from Miocene to the Quaternary. A clear cyclic pattern of the Pliocene and Pleistocene sedimentation is indicative of tectonic, eustatic and provenance fluctuation. The closing member of the entire Neogene sedimentary prism is the fluviatile-deltaic Quaternary sequence, which has a maximum total thickness of 800 m.

The lower part of the overall sedimentary sequence consists of Miocene and/or Pliocene basal conglomerates and autoclastic rocks, then a subsequent marly to shaly formation with here-and-there layered organic limestone formations of Tortonian age. The total thickness of Miocene is varying from a few ten meters to 3000 m.

The Pliocene sequence is characterized exclusively by arenaceous to argillaceous sediments with some basal conglomerates. The total thickness of the entire Pliocene sequence may range up to 5000 m.

Thermal water reservoir systems

(Table 1)

Several hydrostratigraphic units and aquifer systems have been formed in the Pannonian Basin within the Neogene-Quaternary sedimentary sequence. The major units are as follows:

A) *Local occurrences*

1. Miocene and Pliocene basal conglomerates with limited extent. Their water-yielding capacity is poor to moderate. They usually represent salaquifers.

2. Miocene stratified calcareous deposits and sandstones. They are also of limited extent, generally of moderate water-yielding capacity and they are salaquifers, too.

B) *Regional occurrences*

1. Lower Pannonian (=Lower Pliocene) sandstones. They are characterized by low water-yielding property and unfavourable water quality (very often salaquifers).

2. Upper Pannonian sand and sandstone reservoir system as the main and most important thermal water-bearing-and-yielding complex due to its high water-yielding capacity and excellent water quality (low salt concentration) (Fig. 1).

3. Levantian (=Upper Pliocene) sand formations hydrogeologically similar to the Upper Pannonian aquifers, however, appearing in a limited regional extent and yielding water of lower temperature.

4. Thick gravel and sand deposits within Quaternary sequences along depressed areas, with excellent water discharge and quality (fresh waters).

Thermal water-bearing formations

System Series	Sub-stage	Aquifer	Confining beds
Quaternary Pleistocene		Gravel, sand	Clay
Neogene	Levantine	Sand	Clay, silty clay
Pliocene	Upper Pannonian	Sand, sandstone	Clayey marl, silty clay
	Lower Pannonian	Sand, sandstone, conglomerate	Clayey marl, marl, siltstone
Miocene	Sarmatian Tortonian Helvetian	Sandstone, conglomerate, limestone	Clayey marl, marl

Geometry of the thermal water reservoirs

Shape and dimension of sand (sandstone) bodies

Sand bodies represent major aquifers and thermal water reservoir systems within the Neogene to Quaternary sedimentary sequence. Their shape and dimension varies in wide ranges according to differences in the depositional and environmental conditions of the Pannonian Basin during this period.

Miocene sand and sandstone bodies are characterized by limited size occurring in form of lenticular bodies responding accordingly to any aquifer test or production, that is, showing a relatively quick depletion of the ground water contained.

Lower Pannonian sandstone: Due to the normal neritic sedimentation a well-differentiated, alternating sandy to shaly (marly) sequence was formed. The layered sandstone bodies are rather persistent, blanket or sheet-like, confined by thick, impervious pelitic rocks. Nevertheless, these sandstone bodies tend to be pinched out or coalescent. As a result, many lenticular forms of limited extent or, on the contrary, bundles attaining to a relatively great thickness and considerable extent occur. These aquifers, however, are of low hydrogeologic importance due to their unfavourable petrophysical conditions.

Upper Pannonian sand and sandstone bodies were formed in a gradually shallower, near-shore situated depositional environment involving a highly oscillating sedimentation that resulted in a densely alternating sandy to clayey sedimentary series. Well-differentiated, discrete and relatively thick sand bodies occur mainly in the lower part of the Upper Pannonian sub-stage, however, along with a lenticular and multilateral pattern more characteristic here than within the Lower Pannonian sequence. Lenticular rock bodies and frequent pinching-out as well as the ubiquitous occurrence of laterally coalescent sandbed groups (so-called bundles) display a great variety of thermal aquifer forms and patterns (Fig. 2).

Table 1

and their characteristics

Effective porosity %	Permeability in darcy	Water-yielding capacity	Quality of water
30 – 40	1 – 5	Excellent	Fresh water
25 – 30	1 – 2	Very good	Fresh water
20 – 30	0.5 – 1	Good	Fresh to brackish water
10 – 20	0.2 – 0.5	Poor	Brackish to salt water
5 – 15	0.2 – 0.5	Poor to moderate	Brackish to salt water

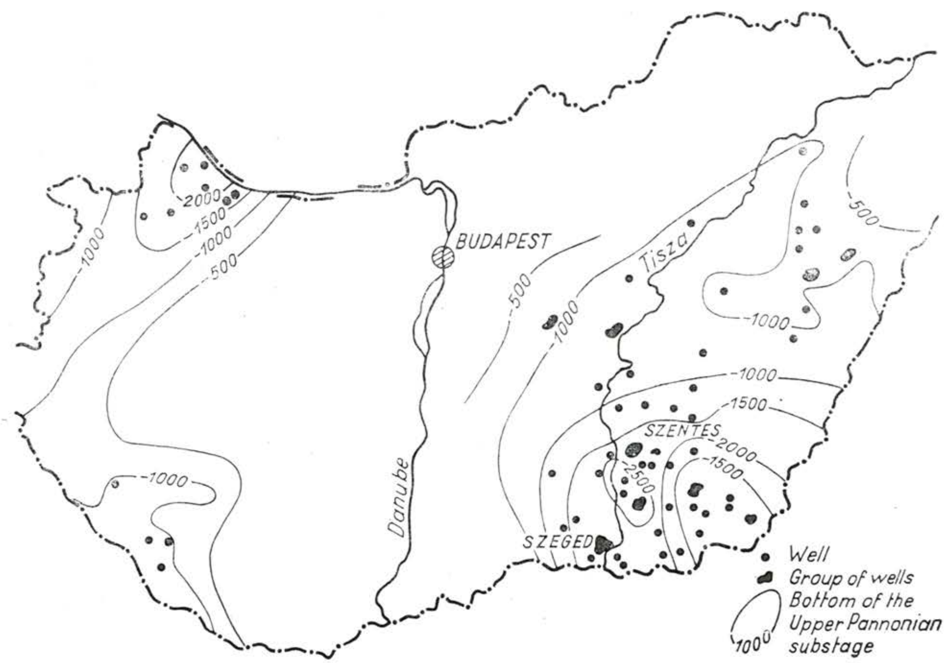

Fig. 1. Thermal water wells tapping Upper Pannonian aquifers (flowing water temperature > 60 °C)

Due to the upward-increasingly alternating sand, silt and clay beds, so-called sandwich-type sedimentary sequence were formed in the upper part of the Upper Pannonian sub-stage in many regions.

It should be emphasized that the overall geometry and gross hydrogeological properties of the Upper Pannonian sand (sandstone) bodies are primarily governed by the depositional characteristics of this peculiar sedimentary environment.

Somewhat similar sedimentary features and aquifer geometries can be found within the sandy-clayey-type *Levantian* series in different sub-basins.

Quaternary gravel and sand deposits frequently occur in form of "fossil" channel fillings and of a succession of superimposed stream deposits, which may be grouped to form widespread deposits or may show a sinuous, meandering pattern.

General features of reservoir geology

The locally developed, *stratified limestone aquifers* of Miocene age form thin, sheet-like reservoirs. Due to their limited lateral extent and moderate hydraulic characteristics, the water-yielding capacity of this kind of aquifer is low (about 200 – 300 litres/minute) with a lasting production time. These aquifers are interbedded with confining beds and contain no renewable water resources.

The principal *Upper Pannonian thermal water-bearing sequence* is, as a whole, a multiple reservoir system. This multistory, multiunit reservoir system consists mostly of superimposed and multilateral confined aquifers. The confining beds (aquicludes and aquitards) are present in form of clay, clayey marls and silts hydraulic conductivity of which may range from zero to some low value. The dimension of the individual aquifers is highly variable as ranging from small pods to widespread laterally persistent beds, which cover many tens of square kilometer. The interconnection of the different, single aquifer units is limited, at least in the lower portion of the Upper Pannonian sub-stage (Fig. 3).

Nevertheless, in certain parts of the sedimentary column with high sand percentage and due to the presence of stratigraphic windows, a vertical intercommunication of the subsurface waters ("vertical leakage") can be observed, which results in a high rate of infiltration of meteoric waters and the development of a descending and ascending water system. Such a convective ground water circulation was revealed in the surroundings of Tiszakécske, governed by temperature difference of the subsurface waters. Within these exceptional regions some renewable ground water resources may be taken into account. Elsewhere, especially in the deep parts of the sub-basins there are no renewable thermal water resources.

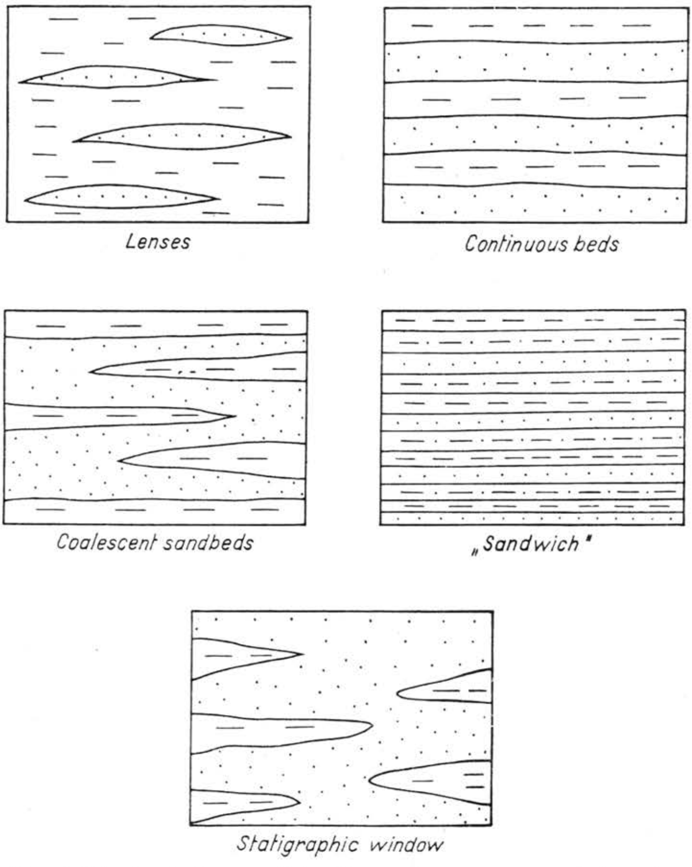

Fig 2. The basic forms of the Upper Pannonian sand deposits

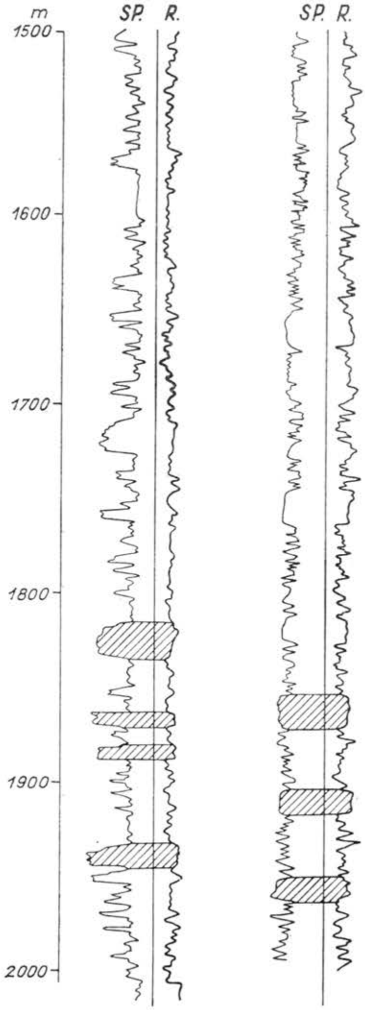

Fig. 3. Typical multiple-reservoir system in Szentes shown in electrical logs with perforated sections

The Upper Pannonian aquifers are characterized by two types of aquifer boundary: *1.* Sedimentary one, such as lens, wedging out, facies changes; *2.* Petrophysical one, such as permeability decrease or contrasts. The role of the tectonic boundaries (faults) are insignificant here.

The layered character pre-determines the anisotropy i.e. inhomogeneity of the overall aquifer system. In addition, from many hundreds of core analysis, a definite horizontal (lateral) anisotropy indicated by a remarkable variation in porosity, permeability and cement, has been found within the individual layers or bundles. It could be observed that mainly shale partings and silts are provoking the permeability contrasts and, therefore, many anisotropic flow packets occur within the sand bodies.

Regarding the gross properties of the Upper Pannonian thermal water reservoir system, the depth interval of this multiple reservoir system is ranging from about 500 to 2500 m. Its being highly differentiated is due to the densely changing boundary conditions, barriers, stratifications, subsurface divides between different sub-basins. The most favourable aquifers were developed in the lower part of the Upper Pannonian sub-stage. The individual sand beds are here relatively thick (from 5 to 25 m) and their lateral extent and persistence are good enough to form efficient thermal water reservoirs.

The sub-basins within the overall Pannonian Basin are generally well-separated multiple-reservoir systems without any interconnection because of the lack of a regional persistence of the water-bearing sand bodies or bundles. Occasionally, however, the existence of some communications may be assumed between certain parts of the reservoir system of the different, nearby subbasins.

Similar multiple-reservoir systems have been developed in the Upper Pliocene (Levantian) sedimentary sequence and deeply-buried Quaternary deposits. These aquifers can be recharged in part by vertical leakage due to the poorly consolidated character of a locally developed sedimentary column constituted mainly by sand and gravel.

Petrophysical properties

The practically most important Upper Pannonian thermal water-bearing formations consist of mature quartzose sand of orthoquartzite character with 70—90 per cent of quartz content. Cementing materials are mainly of calcite (rarely dolomite crystals) amounting to 15—20 per cent that leads to the heterogeneity of the reservoir. The predominant particle size of the sands is 0.1 to 0.2 mm. Representative effective porosity of the main Upper Pannonian thermal water-yielding horizon is 25 per cent on the average, whereas the representative permeability is about 500 millidarcies. Their values are varying, however, in wide ranges from 5 to 35 per cent and from 200 to 1500 millidarcies, respectively, according to the depositional and diagenetic pattern of the sand formations. The dominant clay mineral is montmorillonite.

The hydraulic characteristics of the upper section of the Pliocene and Quaternary sedimentary column are more favourable where the available pore space is ranged from 30 to 40 per cent, while the permeability equals the order of darcy.

It has been found that the water-yielding properties of the aquifers within the overall Neogene sedimentary column is deteriorating versus depth due to gravity compaction and diagenetic processes. In general, the hydraulic characteristics, such as the coefficient of storage, the permeability and transmissivity are a function of the depth. The excellent water-yielding capacity of the Upper Pannonian aquifers is due to the relatively early stage of lithification of this sedimentary sequence, which is generally the domain of mechanical compaction.

The loose character of these young deposits requires a screening technology during well completion. Perforated casing without screen structures is applied only below 1200—1500 m depth, whereas perforated casing sets without grouting are used below 2200 m in the deepest portion of the Upper Pannonian basin.

Performance and operation of the main reservoir system

The multiple sand reservoir represent a complex, multi-unit flow system. The individual aquifers as sand bodies of various dimension are characterized by a specific yield varying in a wide range (from tens to a few hundred liters/meter/minute), as it has been confirmed by well-tests (aquifer-tests) and by flow-meter surveys. Sand beds of great thickness (more than 10 m) and of considerable lateral extent are proved to be the most prolific and efficient. Their practical sustained yield (or safe-yield) is excellent. In practice, a multi-unit completion of well is made (3 to 15 single sand beds are screened or perforated) with a production ranging from a few hundreds to 2000 liters/minute.

The performance of the thermal water reservoirs is primarily determined by the reservoir energy stored within the system. The contributing sources of reservoir energy, manifesting themselves in the formation or reservoir pressure, are within the Pliocene confined aquifer system showing dissolved gas content, high reservoir temperature and compression of the confined water. The dissolved gas content of thermal water expressed by the gas-water ratio (GWR), is widely ranging up to 10 m^3/m^3 or even higher. The gas may be methane or carbon dioxide. Especially high gas content in thermal waters can be measured near the oil and gas fields, where these formation waters can be regarded as edge-water of the oil or gas pools. The reservoir or bottom-hole temperature is ranging up to 145 Centigrades within the Upper Pannonian sub-stage and, as a consequence, the flowing water temperature is reaching the boiling-point in the surroundings of the town of Szentes. Here thermal water is deriving from a depth interval of 2000 to 2300 m.

The formation or reservoir pressure of the aquifers comes, without exception, close to the hydrostatic value, i.e. normal pressures prevail. This hydrostatic depth equilibrium indicates that some kind of connection (submicroscopic permeability) was established between the aquifers and the surface during geological times. The depletion of several thermal water reservoirs accompanied by pressure decline proves, however, that this connection does not exist for the time being.

From a reservoir-engineering point of view, the Pliocene thermal water reservoirs are mostly infinite reservoirs where existence of an external boundary is not felt. There is generally a nonsteady-state flow of virtually single-phase fluid. This flow has a time-varying nature in these confined aquifers. The knowledge of the production history and the cumulative pressure drop is a prerequisite condition in order to judge and predict the performance and future behaviour of such a reservoir.

ON THE POSSIBILITY OF USING ARTESIAN BASIN DEEP-SEATED HORIZON FOR INDUSTRIAL DISCHARGE DUMPING

POSSIBILITÉ D'UTILISATION DES NIVEAUX PROFONDS DE BASSINS ARTÉSIENS POUR L'ENFOUISSEMENT DES EAUX D'ÉGOUT INDUSTRIELLES

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RÉSUMÉ

L'enfouissement des écoulements industriaux dans les nappes aquifères profondes des bassins artésiens exige des investigations hydrogéologiques spéciales. A l'Université de Moscou on a élaboré la méthode de ces investigations. Les buts principaux des investigations hydrochimiques sont: le choix de la nappe aquifère, qui peut être utile à l'enterrement des écoulements industriaux, changement de la possibilité de réception des écoulements industriaux et des nappes aquifères à cause des processus physico-chimiques entre les écoulements et les nappes aquifères ; l'établissement des processus physiques et chimiques pendant la répartition des écoulements industriaux dans la nappe aquifères dans le temps.

Pour résoudre ces problèmes il est nécessaire de faire:

a) des explorations régionales pour étudier la composition chimique des eaux souterraines, les propriétés de filtration de la couche-collecteur, les conditions hydrodynamiques et leurs liaison avec les autres nappes aquifères.

b) Un étude de l'intensification de l'action physico-chimique entre les écoulements industriels et des eaux souterraines et des roches, et l'influence sur cette action de différents facteurs (comme la composition des écoulements industriaux, celle des eaux souterraines et des roches ; la pression, la température, etc.).

c) Les changements quantitatifs de la possibilité de réception de la couche-collecteur à cause des processus physiques et chimiques.

L'influence des facteurs était étudié non seulement théorétiquement, mais aussi expérimentalement.

Le résultat des investigations est la division de la nappe aquifère en subdivisions régionales d'après les conditions hydrogéochimiques de l'enterrement des écoulements industriaux.

The utilization of artesian basin horizons, in a number of cases, demands considerable hydrogeochemical investigations. The justification of the industrial discharge dumping into deep water-bearing horizons is one of the most important hydrogeochemical problems.

Protective measures against damages by industrial discharges are being carried out on a large scale in many countries. Along with various techniques of making industrial discharges safe (mechanical purifying through filters and settlers; physico-chemical purifying by neutralization, freshening, ion exchange; biological cleaning etc.), it often becomes necessary to resort to the underground dumping of industrial discharges.

Highly toxic industrial wastes are those that technologically cannot be purified or their cleaning is uneconomic.

The major conditions to be necessary when dumping industrial discharges are:

1. The water-bearing horizon is to be rather capacious;

2. The aquifer has to be safely isolated from water-bearing horizons containing fresh-water or any kind of water of economic importance;

3. The water-bearing horizon itself must not contain any water used as natural resource;

4. The industrial discharge spreading through water-bearing horizon is to be lasting enough.

Besides the points of view of above, economy is imperative in every particular case.

Hydrogeological and hydrogeochemical investigations to solve the problem of dumping has to be performed stage by stage. The first-stage investigations are devoted to recommend water-bearing horizons suitable for storing industrial wastes. To choose such water-bearing horizons it is important to carry out regional investigations for studying rock filtration properties, underground water chemical composition, dumping depth and the thickness and spreading of the strata-collectors, besides the possibility of underground water movement from a stratum-collector into other horizons. Regions with one or more water-bearing horizons are characterized by different degree of usableness for industrial discharge dumping.

The second stage of investigations, directed to solve a problem of industrial discharge dumping into deep-seated water-bearing horizons involves a task of justification of the changes in receptivity to industrial discharges of the water-bearing horizons, regarding also the movements therein at the time of dumping and in the adjoining period. The changes in water-bearing horizon receptivity for industrial discharges are influenced by physico-chemical interactions of the wastes with underground waters and rocks.

In the areas of industrial discharge dumping several systems are formed. Immediately near the well where industrial discharges are dumped an industrial discharge—rock system is formed. Far from the well industrial discharge—underground water and industrial discharge—underground water—rock systems develop successively. The largest area is occupied by an industrial discharge—underground water—rock system; the industrial discharge—underground water system is an intermediate one between the above-mentioned first and the last systems.

In the systems indicated the study of physico-chemical processes is very difficult due to a number of natural and man-made factors, e.g. conditions like the industrial discharge composition, underground water composition, rock composition, pressure and temperature and rock filtration properties. Physico-chemical processes originating in artesian basins of platform-like areas when industrial discharges interact with underground waters and rocks, develop under the conditions of the geostatic pressure (200—500 atm.), hydrostatic pressure (100—250 atm.) and stratum temperature of 40—80 °C. The study of the natural factors' influence on the interaction of industrial discharges with underground waters and rocks under the conditions of increased pressures and temperature of above, is possible both theoretically and experimentally. Nowadays experimental investigations of industrial discharges—rock, industrial discharge—underground water and industrial discharge—underground water—rock systems are under development. The available literature on the problem of industrial discharge underground dumping deals mostly with the theoretical questions of hydrogeological and hydrodynamical justification of dumping. The authors, with the use of the UIPK installation of batch production, have carried out an experimental modelling of the interaction between industrial discharges and underground waters and rocks, making approaches to secure the conditions that exist in the strata. The installation working conditions allow to create geostatic pressure on a rock from 1 to 600 atm. and a hydrostatic pressure of 1 to 300 atm. along with a temperature of 20 – 80 °C. The interaction between industrial discharges and rock was being generated when filtering industrial waste and that mixed with water through the rock at a given pressure and temperature, characteristic for stratum conditions under which industrial discharges are really dumped. The changes in industrial discharges, underground waters and rocks, as a result of their interaction, were determined by the changes in the liquid phase.

The results of experiments under stratum conditions were compared with experimental investigations under the conditions of atmospheric pressure and of the temperature of 18 – 20 °C. Experiments have showed that under increased pressure and temperature, in industrial discharge – underground water, underground discharge – rock and industrial discharge – underground water – rock systems, respectively, complex physico-chemical processes develop. Their development is governed by factors such as rock composition, moreover peculiarities of the industrial discharges and underground waters. All these factors determine the direction of the ionic migration in the process of industrial discharge interaction with underground waters and rocks. Upon the character of the interaction with industrial discharges and underground waters, the rocks are subdivided into the following types:

a) carbonate, sulphate, terrigenous, feldspathic;
b) terrigenous, quartzose.

The first group of rocks, as a rule, can be leached out by industrial discharges and their mixtures with underground waters. With regard to the second group of rocks, industrial discharges and their mixtures with underground waters interact with it very slightly or do not interact at all. According to their ion composition, pH and degree of mineralization, the industrial wastes of the factories are different. Some of them have acid pH, others alkaline pH; they can be fresh, slightly or highly mineralized. Underground waters stored in deep-seated water-bearing horizons are mainly mineralized, the main ions being chloride and natrium. Solutions of different composition are formed when industrial discharges are mingling with underground waters, constituting the following types:

a) fresh and slightly mineralized with varied ion composition and pH;
b) mineralized and highly mineralized, most often natrium chloride and natrium-calcium-magnesium chloride of acid pH;
c) mineralized and highly mineralized natrium chloride and natrium-calcium-magnesium chloride waters of neutral or alkaline pH.

The first-group solutions interact actively with different rocks (except quartzose); the second-group solutions interact but slightly with all rocks; the third-group of rocks interact but very slightly with all rocks.

Geostatic pressure affects the physico-chemical processes that develop in the systems investigated. The increase of pressure on the industrial discharge – rock and industrial discharge – underground water – rock systems favour to the transition of substances from rock into solution, a fact seemingly connected with rock crystal lattice destruction. The increase of pressure in industrial discharges and their mixing with underground water (i.e. in a liquid phase) promoves the tightening of the solution structure, resulting in a possible removal of substances (as a rule of those rich in solutions: chlor and natrium from natrium chloride solutions, sulphate and calcium from calcium sulphate solutions etc.). Hence, an increased pressure in industrial discharges and industrial discharge-underground water system brings about a substance's falling out into sediment, while in industrial discharge – rock and industrial discharge – underground water – rock systems it brings about rock leaching. The processes of leaching give rise to increasing rock permeability, and, therefore, to an increasing receptivity to industrial discharge of a stratum-collector. The process of a substance fallout into sediment brings about a decrease in rock permeability and receptivity to industrial discharges of a stratum collector. The influence of temperature on ion behaviour in solutions was studied by I. G. KISSIN (1969) and others. Interesting results testify to a significant change in the behaviour under increased temperatures of solutions, e.g. of calcium sulphate (fallout into a sediment) and some others. The influence of temperature on industrial discharge interaction in rocks under increased pressure is very complex, and further experimental investigations are needed. In connection with the fallout of substance into carbonate and sulphate sediments, balance investigations are most important. The quantitative evaluation of the carbonate and sulphate balance involves serious difficulties regarding the complex routes of substance and ion migration. Apart from inner factors of migration, the complexity of a substance and ion migration is determined, to a significant extent, by outer factors of migration.

The hydrogeochemical character of the environment is especially important when considering outer factors of migration. These hydrogeochemical peculiarities affect significantly migration properties of the individual ions and substances. The evaluation of calcium carbonate and sulphate falling out from industrial discharge mixtures and underground waters as well, is performed on the basis of evaluating solution saturation degree with these substances.

According the substances indicated, the degree of a solution saturation is established by comparing active ionic concentration in solution with the solubility of a substance. If active ion concentration in solution exceeds that of the soluble substances, it means that underground water is oversaturated with these substances and hence their fallout into the sediment is quite possible. The determination of the degree of saturation with calcium carbonate and calcium sulphate in mineralized industrial discharge mixtures and underground waters is very difficult because of the complexity of determining ionic activity ratios. The ionic activity ratios depend mainly on the degree of the underground water mineralization and its ionic composition. In literature (GARRELS 1968; ZVEREV et al. 1969) values of the ionic activity ratios for solution of low-degree mineralization have been cited as showing slight variations in its ionic composition. It makes it difficult to use this ratio in qualifying natural solutions of particularly high mineralization. To calculate exactly the saturation degree of these solutions with hardly soluble substances it is necessary to determine experimentally the values of ionic activity factors. The experimental determination of ionic activity values-ratios for highly mineralized solutions is based on determining solubility of substances the compositional ions of which are dealt with and, further on, it is based on the solubility data, activity ratio calculations.

The authors have experimentally determined the activity ratios Ca^{2+}, HCO_3^-, $Ca(HCO_3)_2$, $CaSO_4$ in natrium chloride and natrium-calcium chloride solutions with mineralization of 100–250 g/l and more.

To calculate the mean ratio of calcium sulphate activity ($\gamma_{\pm CaSO_4}$) the following equation was used:

$$\gamma_{\pm CaSO_4} = \frac{(K)^{\frac{1}{2}}}{m_{CaSO_4}} = \sqrt{\frac{K}{m_{Ca^{2+}} \cdot m_{SO_4^{2-}}}} \tag{1}$$

where

K — is a balance constant,

$m_{Ca^{2+}}$ and $m_{SO_4^{2-}}$ — molar concentration of the calcium and sulphate ions in underground water under the condition of balance with calcium sulphate.

The calculation of the activity ratios for calcium bicarbonate Ca and HCO_3 ions in solution has been performed according to the following equations:

$$\gamma_{\pm Ca(HCO_3)_2} = \frac{10^{-5.86} \cdot P_{CO_2} \cdot a_{H_2O}}{m^2_{HCO_3^-} \cdot m_{Ca^{2+}}} \tag{2}$$

$$\gamma_{Ca^{2+}} = \frac{10^{9.8} \cdot a^2_{H^+}}{P_{CO_2} \cdot a_{H_2O} \cdot m_{Ca^{2+}}} \tag{3}$$

$$\gamma_{HCO_3^-} = \frac{10^{-7.83} \cdot P_{CO_2} \cdot a_{H_2O}}{a^2_{H^+} \cdot m^2_{HCO_3^-}} \tag{4}$$

where

γ — the activity ratio value,

m — the ion molar concentration,

P_{CO_2} — the partial pressure CO_2 (atm.),

a_{H^+} and a_{H_2O} — are hydrogen and water ion activities.

The calculation of underground water saturation with calcium sulphate is carried out as follows:

1. The calculation of component concentration in form of molar concentration:

$$m_i = \frac{x_i \cdot 1000}{(1000 - \Sigma_M) \cdot N} .$$

m_i — quantity of molar of component i in 1000 g H_2O,

x_i — concentration of component i in g/1000 g of solution,

Σ_M — solution general mineralization in g/1000 g of solution,

N — ion weight.

2. The calculation of ion solution power:

$$M = \frac{1}{2} \Sigma_{m_i z_i^2}$$

M – solution ion power,
m_i – component molar concentration,
z_i – component valency.

3. Finding of the calcium sulphate activity ratio value ($\gamma_{\pm CaSO_4}$) as ion power function [f(M)] by the graph made on the basis of experimental data by the solubility of gypsum in natural water.

4. Calcium sulphate ion activity product,

$$IAP_{CaSO_4} = \gamma_{\pm CaSO_4} \cdot m_{Ca^{2+}} \cdot m_{SO_4^{2-}}.$$

5. Solution saturation degree with calcium sulphate (α_{CaSO_4})

$$\alpha_{CaSO_4} = \frac{IAP_{CaSO_4}}{K^0_{CaSO_4}}$$

where

$K^0_{CaSO_4}$ – value of thermodynamic product of the solubility of sulphate of calcium by experimental data (for $P = 1$ atm. and $t = 25$ °C, $K^0_{CaSO_4} = 3.36 \cdot 10^{-5}$).

If

$\alpha_{CaSO_4} = 1$ – the solution saturated with gypsum,
$\alpha_{CaSO_4} < 1$ – the solution is not saturated with gypsum,
$\alpha_{CaSO_4} > 1$ – the solution is oversaturated with gypsum.

6. The amount of calcium sulphate (x) that can be solved in a solution given (i.e. at definite $m_{Ca^{2+}}$, $m_{SO_4^{2-}}$, M) is found by the quadratic equation of

$$(m_{Ca^{2+}} + x)(m_{SO_4^{2-}} + x)\,\gamma_{\pm CaSO_4} = K^0_{CaSO_4}$$

For solutions of high concentration i.e. with $M > 0.7$, one is allowed to neglect changes by $\gamma_{\pm CaSO_4}$ due to a slight increase in concentration on account of solution.

The calculation of the underground water saturation with calcium carbonate (α_{CaCO_3}) is performed by the formula

$$\frac{\gamma_{HCO_3^-} \cdot m_{HCO_3^-} \cdot \gamma_{Ca^{2+}} \cdot m_{Ca^{2+}}}{a_{H^+} \cdot 93.2} = \alpha_{CaCO_3}.$$

The activity ratio values γ_{HCO_3} and γ_{Ca} are also determined upon experimental data:

a_{H^+} – hydrogen ion activity in the solution, i.e. at pH = 7.0.

$$a_{H^+} = 10^{-7.0}.$$

The formula is obtained for $P = 1$ atm. and $t = 25$ °C.

As a result of finding relationships in the processes of leaching and falling out into sediments in industrial discharge-sediment, industrial discharge–rock, industrial discharge–underground water and industrial discharge–underground water–rock systems, prognosis is given for every particular area of dumping for the changes in receptivity to industrial discharges in a stratum-collector.

In characterizing changes in rock receptivity over larger territories under the influence of physico-chemical processes, a regional division of the individual water-bearing horizons is to be prevised for.

A forecast map for showing regions on the basis of hydrogeochemical conditions of the industrial discharge dumping is made upon the informations about the water-bearing horizon's natural hydrogeochemical setting; moreover on filtration properties of the strata-collectors, on physico-chemical processes of interaction between industrial discharges and underground waters and rocks, on natural factors influencing the industrial discharge-rock, industrial discharge—underground water and industrial discharge—underground water—rock systems.

A chart showing prognostic regions upon regarding the hydrogeochemical conditions of the industrial discharge underground dumping is purposeful as being aimed at the solution of concrete practical problems. Thus it represents a model for a regional division on the basis of taking into account natural facts combined with those resulting from man's economic activity, in particular by the chemistry of industrial discharges. Areal units where individual water-bearing horizons occur are as follows: province, area, subarea of the first order, subarea of the second order. Province is the largest unit that corresponds to a hydrochemical unit of the subsurface water's composition. As a hydrogeochemical principle for distinguishing a province, it should be used to show the extension of underground waters of different hydrogeochemical types, classes and groups. Within a province one can single out areas by the character of industrial discharge pumped. Industrial discharges are characterized by their ionic composition, degree of mineralization and pH. Within every area it is established an assumed composition of the industrial discharge mixing with underground water and, taking it as a principle, one can reveal industrial discharge compatibility to underground water by calcium carbonate and calcium sulphate.

The 1st-order subarea is singled out on the basis of geochemical characteristics of the watered rocks. For example, if a water-bearing horizon is composed of carbonate rocks, then areas built up of a combination of limestones and dolomites will be singled out. In the case of an aquifer composed of terrigenous rocks, areas constituted by rock-forming feldspathoids or quartz could be separated.

Therefore, a subarea of the first order is a part of a water-bearing horizon, characterized by a lithological i.e. geochemical homogeneity. In the first-order subareas peculiar physico-chemical processes between liquid and solid phases are to be developed depending on the combination of the composition of industrial discharge, industrial discharge mixture, underground water and rock as well. Depending on the characteristic processes, changes in the filtration and storage capacity of rocks can develop.

The second-order subarea is singled out upon inhomogeneity in the filtration capacity of rocks. A part of the water-bearing horizon characterized by definite properties of rock filtrants makes up this subarea. Rock filtration properties are influencing on the intensity of physico-chemical interaction between rocks and industrial discharges, industrial discharge mixtures and underground waters, respectively. Therefore on regional maps there are singled out water-bearing horizon areas with different changes in the filtration and storage capacity of rocks. Quantitative stages of rock permeability should be established for any particular area where industrial discharges would be dumped, on the basis of statistical techniques of separation.

During the second stage of investigation dealing with the problem of industrial discharge dumping into artesian basin deep-seated water-bearing horizons, special attention should be given to hydrogeological investigations devoted to establish arrangements in the capacity of filtration of water-bearing horizons and to set up hydrodynamic schemes. Hydrodynamic schemes and inhomogeneity of rock filtrants for water are of primary importance in calculating presumable movements provoked by industrial discharges along water-bearing horizons and in evaluating changes in the receptivity of the industrial discharge water-bearing horizon due to the physico-chemical interaction of wastes with underground waters and rocks.

Porosity, permeability, pumping receptivity and transmissivity are plotted on rock filtration inhomogeneity schemes. Porosity and permeability are determined upon well-logging and laboratory analyses. When making scheme of the rock filtration inhomogeneity one should take into account qualitative indices of the filtration properties of rocks, the presence of fractured zones, structural alignments, the extent of paleokarsts etc.

Hydrodynamic charts of the industrial discharge dumping are made by two steps:

1. When feeding wastes into a water-bearing horizon;
2. After the execution of industrial discharge pumping.

The spreading of industrial discharge pumped into the water-bearing horizon is first determined by technological peculiarities and pumping regime. More often, at the time of pumping industrial discharges can move a short way off but a few kilometres in some decades. As to the second step, after pumping, industrial discharges spread along water-bearing horizons under the conditions of a natural hydrodynamic regime and by underground streams. At this second stage, the duration of spreading through a water-bearing horizon of the wastes is determined by the underground water movement's velocity and upon the physico-chemical character of the processes exerted. It is a problem for hydrogeological prognosis to determine distance and time of the reduction of the harmful constituents to an acceptable rate in industrial discharges and underground waters.

In brief, for the solution of the problem how to dump properly industrial discharges into deep-seated horizons, the following hydrogeological investigations are to be carried out:

1. The classification of the industrial discharges on the basis of their chemical composition, along with the selection of the group of the biologically most harmful industrial discharges.

2. The elaboration of prognosis charts showing the presumable chemical composition of underground water, mixed with industrial wastes.

3. The appraisal of the saturation of the industrial discharge/groundwater mixture with calcium carbonate and calcium sulphate.

4. The compilation of charts of the spreading of industrial discharge mixed with underground water, characterized by a particular degree of oversaturation with calcium sulphate and calcium carbonate. It is necessary to single out on a chart water-bearing horizons unfavourable to industrial discharge dumping due to calcium carbonate and calcium sulphate sediments.

5. The examination of the physico-chemical processes an interaction between industrial discharges (and their mixture with groundwater) and rocks.

6. The investigation of pressure, temperature, composition of industrial discharge, underground water and rock and of the filtration properties of rocks influencing physico-chemical interaction with rocks of the industrial discharges.

7. The definition of the main kinds of the physico-chemical interaction between individual groups of industrial discharge and rocks. Main kinds of physico-chemical processes making influence on water-bearing complexes' receptivity to industrial discharges.

8. The division of areas into prognostic regions regarding an individual water-bearing horizon or a series of aquifers according to the hydrogeochemical conditions of the industrial discharge dumping.

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SUR UN CAS PARTICULIER D'ALIMENTATION D'UNE NAPPE SUPERFICIELLE PAR UNE NAPPE PROFONDE (VALLÉE DU GAPEAU — PLAINE D'HYÈRES, VAR, FRANCE)

SUPPLY OF PHREATIC AQUIFERS FROM DEEPER HORIZONS (GAPEAU VALLEY — HYÈRES PLAIN, VAR, FRANCE)

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ABSTRACT

Gapeau's low valley conceals two superimposed aquifers. A reduction in salinity of the upper aquifer becomes possible, being recharged by the underlying confined aquifer.

L'hydrogéologie de la basse plaine du Gapeau, près de Hyères (Var) a fait l'objet de nombreuses études, qui ont été synthétisées dans une publication du B.R.G.M.*

Rappelons brièvement que le comblement de la vallée comprend à la base, plaquées sur un socle imperméable, des alluvions datant de la transgression flandrienne (sables argileux et micacés, argiles bigarrées), puis des dépôts récents formés de cailloutis divers, avec des intercalations sporadiques de niveaux argileux imperméables. Au-dessus de ces dépôts, à structure souvent lenticulaire, viennent des formations de surface plus ou moins limoneuses.

L'hétérogénéité de ces formations, dans leur ensemble, et les passages latéraux fréquents que l'on observe entre les zones de perméabilités très différentes, se traduisent par un régime hydrogéologique complexe. Certaines anomalies, en effet, peuvent être observées :

— variations irrégulières de la valeur du gradient le long d'un profil rectiligne ;
— axes de drainage différents du tracé actuel du Gapeau ;
— équilibre précaire de la nappe, ou plus exactement de l'ensemble des nappes ;
— remontées anormales du biseau salé ;
— alternance d'alimentation et de drainage de la nappe (ou des nappes) par le Gapeau ;
— sinuosités inattendues des courbes de teneur en ClNa, indépendantes des indentations que l'on observe sur les courbes de niveau du toit de la nappe superficielle.

* G. Durozoy, Cl. Gouvernet, P. Jonquet, Ch. Olivo et L. Potie : « Données sur l'hydrogéologie de la basse vallée du Gapeau et de la plaine d'Hyères (Var) ». Bull. du B.R.G.M., 2e Série, Section III no 2, 1971, pp. 9—28.

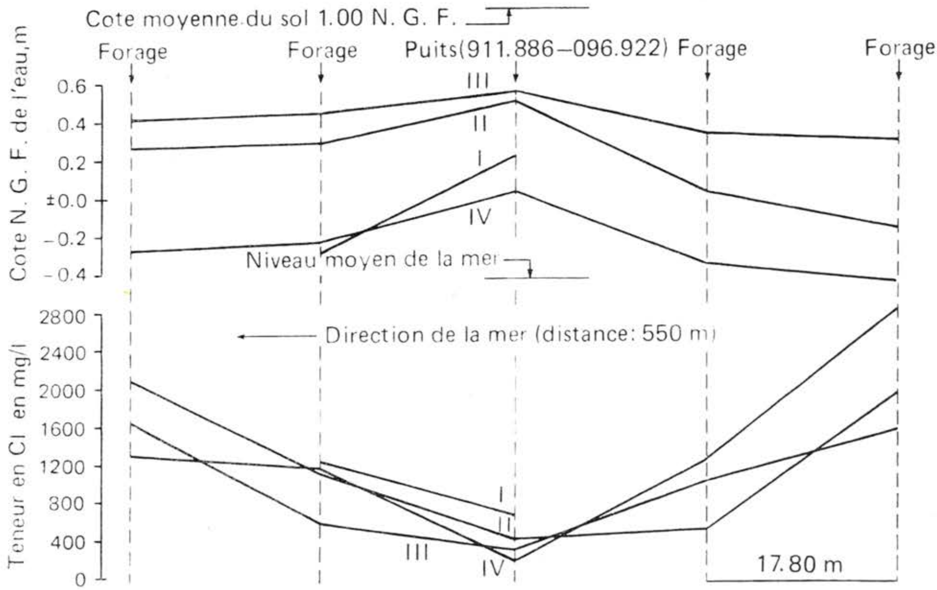

Fig. 1. Plaine du Gapeau. Cote NGF de l'eau et teneur en Cl dans un puits et dans les forages voisins

Dates des mesures: I: 31/08/72, II: 12/10/72, III: 25/04/73, IV: 20/09/73 (en raison de la différence d'échelles le gradient apparent vaut 4, 5 fois le gradient réel)

Cependant, de manière générale, la plus grande partie des puits et forages exécutés indiquent que l'on trouve d'abord la nappe phréatique superficielle, dont le toit est peu élevé au-dessus du niveau de la mer, puis une couche argileuse imperméable épaisse de 8 à 10 m, et enfin au-dessous une nappe presque toujours en charge, dans un terrain de perméabilité variable, mais en moyenne assez élevée ($K=10^{-2}$ m/sec.).

La nappe superficielle

Elle est alimentée :

— par les pluies (en moyenne 800 mm par an);

— par les éventuelles inondations du Gapeau, qui étaient autrefois plus fréquentes qu'actuellement;

— par des infiltrations latérales provenant du lit du Gapeau;

— par des arrivées d'eau provenant de la nappe profonde.

La nappe superficielle qui se situe entre $+0.7$ et -2 m est très affectée par l'évaporation et l'évapotranspiration; elle se vidange directement en mer au niveau du rivage. Son débit est faible, comme ses réserves. Elle joue néanmoins un rôle essentiel dans l'économie locale car de très nombreux puits captent cette eau.

La nappe profonde

Elle est alimentée :

— par des infiltrations de piémont;

— par des infiltrations latérales provenant de la partie haute du lit du Gapeau.

C'est une nappe en charge dont le débit est très important. Elle se vidange, soit dans le bief aval du Gapeau, soit en mer où l'on observe des résurgences d'eau douce.

Les rapports entre ces nappes

La nappe superficielle étant plus riche en sels que la nappe profonde, il était intéressant d'étudier les possibilités d'alimentation de la nappe superficielle par la nappe profonde en vue d'envisager une utilisation plus rationnelle des eaux de ce secteur.

Un puits, creusé jusqu'à la couche argileuse, avait été prolongé par un forage traversant cette couche et atteignant les formations inférieures perméables. L'extrémité supérieure du trou de forage étant obstruée, il était très spectaculaire de voir, lors, du débouchage, le niveau de l'eau dans le puits remonter de plus d'un mètre en quelques minutes.

La communication verticale entre les nappes était ainsi mise en évidence, de même que la mise en charge de la nappe inférieure, ce qui était d'ailleurs prévisible étant donné la cote d'alimentation de cette dernière.

Les deux nappes étant mises en communication permanente, on a procédé, à plusieurs mois d'intervalle, à des mesures de cotes de toit et de teneurs en chlorures, non seulement dans le puits, mais encore dans quatre sondages exécutés dans la nappe superficielle, à la tarrière à main, alignés dans le sens de la pente de la nappe (2 en amont, 2 en aval).

La figure jointe montre, pour 4 mesures faites au cours d'une année, la relation qui peut être établie à partir d'une nappe profonde dont la charge (donc le débit) variait, au bénéfice d'une nappe superficielle dont la cote augmentait en même temps que la salure diminuait. La salure du sol est essentiellement due à l'évaporation consécutive à la remontée capillaire dans les sols limoneux, et à l'évapotranspiration importante provoquée par des cultures maraîchères.

On peut tirer de ces mesures deux observations principales :

1) La mise en charge de la nappe inférieure se traduit, ponctuellement, par un débit très important modifiant les caractéristiques de la nappe superficielle. Comme il est certain que la couche argileuse, au toit de cette nappe profonde, est discontinue, il semble évident qu'il puisse y avoir localement des remontées d'eau profonde importantes à l'occasion de perméabilités plus grandes : cela pourrait expliquer certaines anomalies constatées dans le régime hydrogéologique.

2) L'influence d'un tel forage, dans ces conditions précises, se fait sentir sur une surface de terrain qui est certainement supérieure à un hectare. On dispose donc d'un moyen simple, un forage d'une dizaine de mètres, pour augmenter le volume utilisable de la nappe superficielle, lorsque les circonstances lithologiques s'y prêtent et procéder ainsi à une irrigation verticale, de bas en haut, dessalant les terrains superficiels. Bien entendu, une étude plus générale s'impose alors.

En conclusion

Cette étude ponctuelle sur une assez longue durée permet d'expliquer les anomalies des résultats d'un important travail exhaustif sur l'ensemble des nappes alluviales du Gapeau. L'analyse à grande échelle d'une telle nappe, utilisée de manière intensive, mérite donc d'être effectuée ; elle aurait l'avantage de montrer, de manière exemplaire, l'importance des études très détaillées des puits en zone alluviale lorsqu'il est possible de modifier facilement les relations naturelles entre les nappes. On voit là, tout l'intérêt économique que présentent de tels travaux.

PECULIARITIES IN THE FLUCTUATION OF THE GROUND WATER LEVEL IN THE GREAT HUNGARIAN PLAIN

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The hydrogeography considers the Great Hungarian Plain and—with a somewhat daring generalization—the whole Carpathian basin to be a *hydrological unit.* This is affirmed, among others, by its uniform morphology and hydrography, and by its limited extent: the distance between its northern and southern boundary amounts to not more than 250 kilometers. The consequence is that a rather close correlation can be found between the fluctuations of the surface- and the underground water levels. So during the flood of the Tisza and its tributaries in 1970 also the ground water level was very high.

In spite of these facts some phenomena may be found which point to an incomplete uniformity. The wheather in the Carpathian basin is being directed by three climatic systems: the oceanic, the continental and the mediterranean as well. The high mountain chain of the Carpathians will also determine, which of them will be preponderant in a given time. The year 1965 produced, in this respect, a characteristic hydrological example: there was a rather high and longlasting flood wave running down on the Danube and, at the same time, the ground water level was just slightly over the mean of many years. (An additional example will be mentioned later.)

The investigation of the fluctuations in the ground water level beneath the Great Hungarian Plain has been facilitated by the many observation wells established in the time interval between 1930 and 1940. According to previous studies (see RÉTHÁTI, 1974), there are 17 wells having neither natural nor artificial disturbing effects (as e.g. lake, irrigation, water pumping etc.). Besides, there are 7 further wells with a homogeneous time series, where, however, the maximum or the minimum is not realistic.

One of the most important tasks in the practice of hydrology is the prediction of the limiting values of ground water level. A number of methods, using even basically different approaches, has been concluded to the same result in the literature. In the recent years, several methods using the laws of the mathematical statistics have been published. One of these consists of analyzing autocorrelation viz. spectral functions. In the following, this approach will be used for investigating our wells. We take the number of intervals —considering the rather short time series—for max. 16. The characteristics

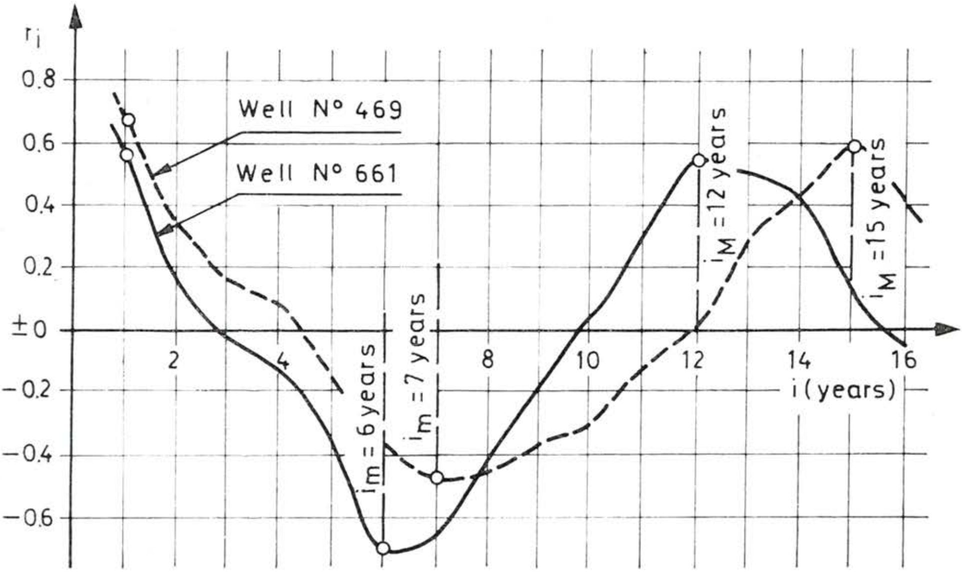

Fig. 1. Autocorrelograms for two observation wells derived from the mean September ground water level for the time series 1938—1972

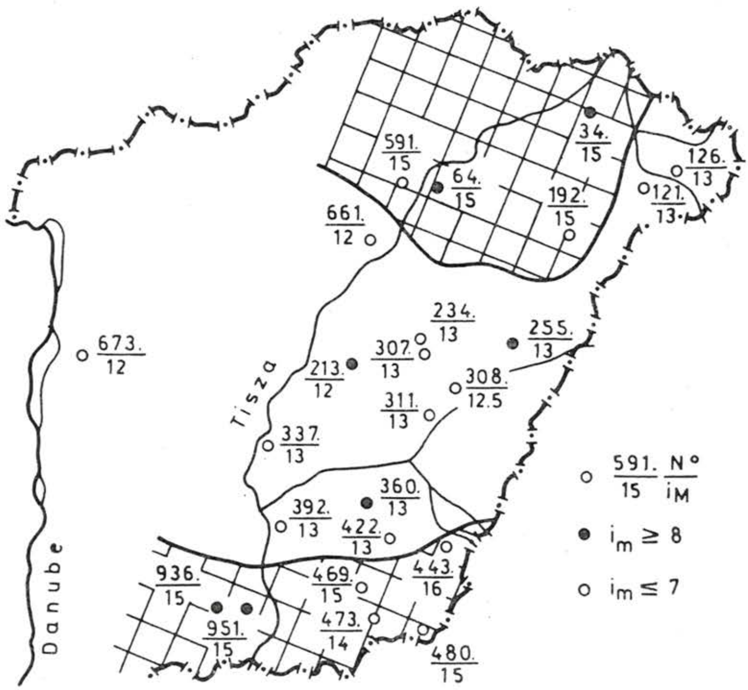

Fig. 2. Territorial distribution of the values i_m and i_M

Table 1

Some characteristic data in the 1938–72 time series of precipitation and of the ground water respectively

No.	Site	MW (cm)	Soil type	Trend cm/year	For water levels		Peak in periods		For precipitation	
of the well					i_m	i_M	1.	2.	i_m	i_M
34.	N	294	1.74	0.370	7.5	15				
64.	N	288	2.59	−2.071	8	15	41	71	7	13
121.	M	349	2.62	−2.819	6	13	41	66	7	14
126.	M	374	2.00	−2.092	7	13	41	70	7	11
192.	N	293	1.00	−0.043	7	15			7	14
213.	M	363	3.19	−6.159	8	12				
234.	M	336	2.84	−1.749	7	13	40	66	7	14
255.	M	295	3.00	0.173	8	13	40	66	7	12
307.	M	361	2.76	−2.749	7.5	13	42	66*	7	13.5
308.	M	363	3.00	−1.997	6.5	12.5			8	14
311.	M	338	3.59	0.513	7	13				
337.	M	268	2.44	−1.655	6	13	41	66	7.5	12
360.	M	393	1.68	−2.578	8	13	45	70	7	11
392.	M	429	2.66	−3.167	7	13				
422.	M	399	3.00	−0.949	8	13	42	70	7	11
443.	S	228	3.22	−1.320	7	16			7	11
469.	S	168	3.99	−0.649	7	15				
473.	S	478	3.84	−0.186	7	14	42	70	7	13.5
480.	S	392	3.00	−1.737	7	15	42	70	7	11
591.	N	176	1.00	0.514	8	15				
661.	M	245	2.69	−0.237	6	12	41	66*	7	13
673.	M	69	3.00	−0.134	8	12			7	11
936.	S	184	1.00	−0.102	8	15	42	70	7	14
951.	S	170	1.00	−0.891	8.5	15	40	66	7	11

* and 1971

sought for is the average September ground water level, the time interval considered 1938–1972.

Fig. 1 presents autocorrelograms for two wells. We may make the following statements:

a) the shape of the curves is relatively regular, the "memory" of the ground water can be traced well from year to year,

b) the limiting values of the autocorrelation coefficients are high, which is favourable for making predictions,

c) the length of the period (that of the wave) is 12–15 years.

An interesting picture will be obtained if plotting the time of the positive limiting value on the map of the Hungarian Plain (Fig. 2). In the middle part of the territory the values of i_M are consequently lower than those near the boundaries to the north and south, respectively. Table 1 gives a deeper insight. The meaning of the characteristics therein is the following.

MW (column 3) gives the mean level of the ground water, i.e. the arithmetic mean of all readings made between 1938 and 1972.

"Soil type" is an index, pointing to the permeability of the soil layers above MW. For sands it is *1*, for silty sands *2*, for silt and loess *3*, for lean clay *4*, for clay and sandstone *5*, respectively. For more than one layer a weighted average has to be calculated since, for a seepage which is perpendicular to the boundary, the permeability of the layer having finer grains is decisive. Therefore, in the case of two layers, the thickness of the finer one was multiplied by two, in the case of three layers, the finest by three, the intermediate by two.

The trend of the changes is given by the inclination ($\tan \alpha$) of the straight line averaging the ground water levels. (The negative sign means continuous rise since the depth beneath the surface was measured.)

The values i_m and i_M, obtained for the water levels have been used in the sense of Fig. 1; the same is true for the values i_m and i_M of precipitation.

The columns "peak" give the year when the water level in time intervals between 1940—1945 and 1966—1971, respectively was the highest.

Using the data collected in Table 1, let us calculate first the territorial averages of the caracteristics (Table 2).

The last raw of the Table 1 supplies the coefficient of correlation between i_M and the given characteristics, irrespective of the site of the well. i_M the smaller, the deeper the ground water, the more cohesive the soil and the greater the inclination of the trend line. According to the Table, all three characteristics act in a direction to make i_M the smallest in the middle part of the Plain. In reality, these three variables are not independent from each other, the cohesion of the soil increases the virtual depth of the ground water (it increases the seepage time), on the other hand, the trend is, according to the study quoted above, a function of the average depth of the ground water. In addition we have to keep in mind that the dynamic equilibrium level of ground water is located deeper in cohesive soils.

The average depth of ground water will define the date of the highest ground water level in a given rainy period. Calculating the average value of MWs in identical years, correlation is obtained in Table 3.

Table 2

	MW	Soil	Trend	i_M	
				Ground water	Precipitation
M	321	2.75	−1.724	12.75	12.4
N+S	267	2.24	−0.612	15.0	12.5

Table 3

	First period			Second period	
	1940	1941	1942	1966	1970/71
Average water depth (cm)	267	305	368	289	358

According to the i_M-values for precipitation and the coefficient of correlation, respectively, there is no correlation between the i_M-values for the precipitation and for the ground water. Fig. 3 seems to support this statement, by giving the average values of the precipitation-autocorrelograms for the different territories. In spite of this, there are phenomena, pointing toward the peculiarities in the distribution of the precipitation and through that in the ground water fluctuations caused by the geography of the Great Plain ("basin character").

We are going to investigate the meteorological characteristics of the two rainy periods, considering the northern and the southern part of the Plain separately.

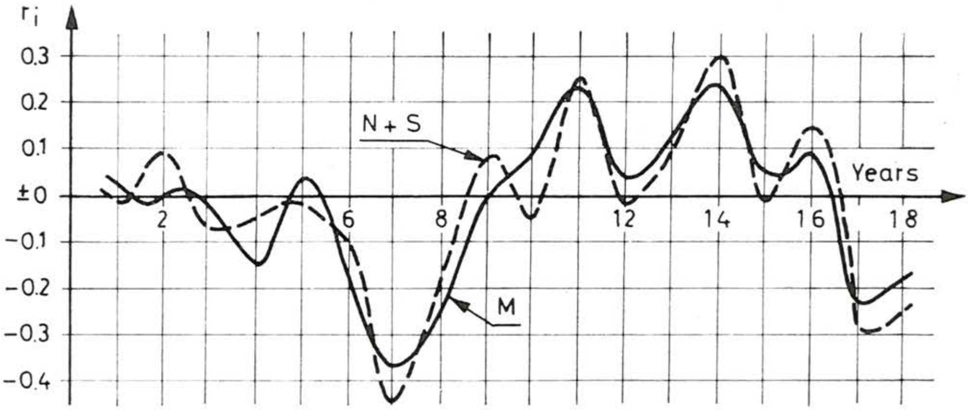

Fig. 3. Mean autocorrelograms for the precipitation of the middle part of the Hungarian Plain (M) and for the bordering territories in the North and in the South (N + S)

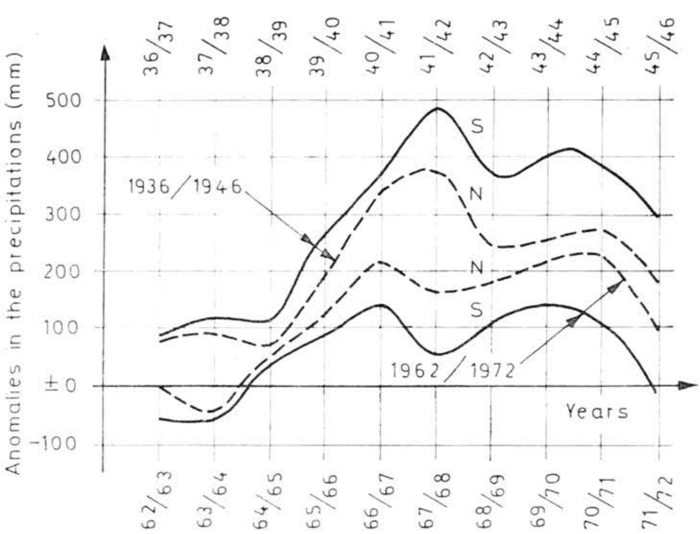

Fig. 4. Integral curves of anomalies in the precipitation in the North (N) and in the South (S)

It is a well-established fact that the average ground water level keeps balance with precipitation and evaporation of many years. A continuous rise will occur if the amount of precipitation is higher than the average for a series of months or years. The process can be described best, if determining the integral curve of the anomalies in the precipitations. According to Fig. 4 the curve for the period 1940–1945 runs higher for the southern part, for the period 1966–1971 the curve for the northern part runs higher.

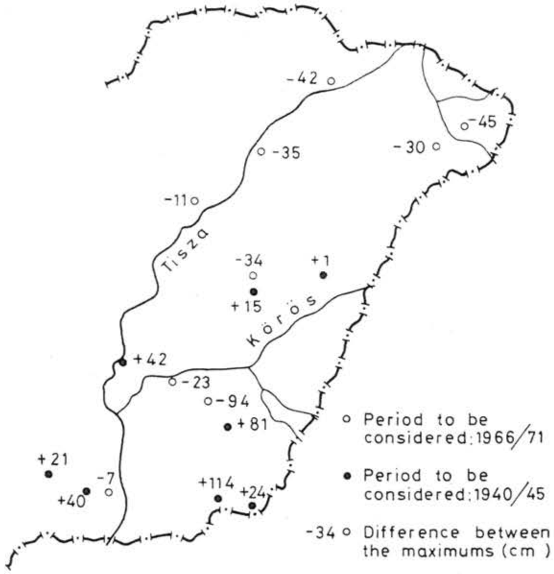

Fig. 5. Occurrence of the highest ground water level in the Hungarian Plain

The consequencies of the situation revealed on Fig. 4 are shown on Fig. 5. According to this, the highest water levels in the southern part of the Plain occurred (neglecting a relatively narrow strip) in 1940/42, in the northern part in 1966/71. This is in accordance with the integral curves of the anomalies in the precipitations, and also with the fact that the highest floods in 1970 occurred on the northern tributaries of the Tisza River.

The aim of this paper was to prove that the "boundary effect" and the "basin character" may influence the ground water fluctuation in plains of relatively small extent.

METHODS OF REGIONAL ESTIMATION OF SUBSURFACE WATER RESOURCES OF MULTILAYERED ARTESIAN BASINS

MÉTHODES D'ESTIMATION DES RESSOURCES D'EAUX SOUTERRAINES DANS DES BASSINS ARTÉSIENS À PLUSIEURS NAPPES AQUIFÈRES

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RÉSUMÉ

En appréciant régionalement les ressources d'eaux souterraines de systèmes captifs à plusieurs nappes il est difficile et exigeant beaucoup de travail de créer le schéma spatial de calcul.

Ce problème peut être résolu sur la base de recherches complexes: géologiques, hydrogéologiques, géophysiques et hydrologiques.

L'histoire du développement géologique de la région permet d'ouvrir des lois spatiales sur la répartition des nappes aquifères et des couches étanches et sur la variation de la filtration.

Les résultats géophysiques donnent une grande quantité d'informations complémentaires à propos de la construction du système de plusieurs nappes et leurs proprietés filtrantes.

Sur la base des investigations hydrodynamiques et thermiques on a distingué des régions de l'alimentation et celles de déchargement des eaux artésiennes.

L'analyse des matériaux hydrogéologiques et hydrologiques permet d'apprécier quantitativement le caractère et la dimension de l'alimentation et de déchargement des eaux souterraines dans les vallés des rivières.

Le modèle physique, qui est construit sur la base des investigations complexes. exprime avec la plus grande vérité et exactitude les conditions naturelles de la région,

The scientific background of the techniques used to estimate subsurface water resources of a region was developed mainly by F. M. Bochever, N. N. Bindeman, N. I. Plotnikov, L. S. Yazvin, and other researchers. The estimation consists principally in analytical calculation or analog modelling of either the network of evenly spaced wells or actual or potential interacting water intakes.

The physical model that would thoroughly and correctly reflect the natural conditions in the region is difficult to design. This complicates the matter and occasionally causes errors in the estimation of resources.

Multilayered water-bearing systems, composed of sandy-clayey sediments, are characterized by complex spatial relations among lithologically different rocks and by changes in permeability along horizontal and vertical direction and in relations among complexes.

The Terek-Kuma artesian basin (TKAB) in the south of the European U.S.S.R. is typical in this respect. It is a multilayered system of water-bearing complexes enclosed in Mesozoic-Cenozoic sandy-clayey sediments. Our investigations conducted in TKAB have made it possible to come to certain conclusions as to the methods of estimating subsurface water resources in similar basins.

To mathematically simulate a basin, it is necessary, above all, to make the basis for the recognition of major water-bearing complexes, trace their boundaries, and discern systematic changes in their transmissibility and the relations among water-bearing complexes and between these and surface waters. This permits to estimate the regional aspects of formations of subsurface water resources in the basin.

Several papers dealing with the hydrogeology of such basins distinguish water-bearing complexes in the section on a stratigraphic ground. This can, on one hand, be attributed to the monotonous lithology of stratigraphically different sediments and, on the other, to the poor knowledge of the hydrodynamic features of the various formations. It is obvious that as to multilayered formations the recognition on such grounds is extremely tentative and unacceptable for the regional estimation of subsurface water resources whereby an intensive water intake is envisaged for a number of water-bearing complexes regardless of their stratigraphic position. Moreover, the stratigraphic basis of recognition would vague and occasionally distort the hydrodynamic situation in the basin.

We have therefore distinguished water-bearing complexes on a hydrodynamic ground after layer-by-layer studying the transmissibility of reservoirs and changes in hydraulic head.

Sequences are extremely difficult to correlate exclusively on geologic ground because they are monotonous in lithological composition and lacking a distinct rhythm. Experience has shown that in this case water-bearing complexes can be distinguished only through the analysis of the geologic hystory of the basin and correlation of geologic evidence with hydrodynamic and geophysical one.

The subsurface flow in an artesian basin is being formed during, and so reflects all the characteristic features of, the long geologic history of the region. Therefore a careful analysis of the general geological situation at different stages of the development of the region allows us to discern, on a reliable basis the principal systematic features of the non-uniform transmissibility of reservoirs and hydrodynamic and hydrochemical features.

In this case the geological history is a major factor affecting the distribution of subsurface water resources, their resumption and prospects for their intake.

As a distinction from local estimation, the regional estimation of resources necessitates investigations into the geologic history of the multilayered formations of the artesian basin.

It was proven possible to subdivide the multilayered sandy-clayey formations and to distinguish the less water-permeable and more water-permeable complexes only through wide application of geophysical information which adds much to our understanding of the geologic structure of the basin and permits integration of the hydrodynamic characteristics of individual layers into a hydrodynamic system.

After purposeful reinterpretation of geophysical data recorded in different types of wells, above all from hydrogeological wells, which are consistantly logged on a wide scale, a detailed correlation of the sequence was made.

In studying the multilayered sequence of TKAB, the principal method was making stratigraphic-correlation charts on wells by using all the available geophysical logs (electrical, radioactivity, lateral, caliber, temperature, etc.).

The correlation was carried out both within localities and over the entire region. In the former case, the geologic structure and the lithology of the sequence were discerned, these being of decisive importance for further correlation. The regional correlation was conducted by using a system of profiles which are evenly spaced within the basin and directed so as to be supplemented by the greatest amount of geologic, geophysical, and hydrogeologic information. Principal markers are (from bottom to top): marly-sandy member of the Mamay bed of middle Sarmatian age (regional marker), clayey member of the Akchagylian (southern and central basin), and Lower Quaternary clay (steppes of Dagestan). Other characteristic intervals of the sequence were occasionally taken as markers to correlate in more details.

As a result of a basin-wide integration of the regional correlation charts a geologic-geophysical foundation has been laid down for compiling detailed structural maps.

Seven principal water-bearing complexes were thus distinguished within the basin in Neogene-Quaternary sediments, and their water in place and recoverable water reserves were determined. Fig. 1 shows a diagrammatic section along the dip of rocks as an example of subdividing a multilayered formation.

Study of changes in the permeability of water-enclosing sediments on which inadequate information is available is an important step toward the substantiation of the physical model of the basin. To estimate the transmissibility of water-bearing complexes, there were widely used geophysical techniques. A routine set of geophysical investigations was used to raise the accuracy of the transmissibility characteristics of the sequence over the entire depth interval under investigation. This is of special importance for TKAB, because only a negligible part of the water-bearing formation is generally tested by hydrogeologic wells.

We have found a correlation between the coefficient of transmissibility and the relative clayiness derived from radioactivity logs, and used it to estimate the transmissibility of subsurface water reservoirs. The correlation coefficients of the above values range between 0.6 and 0.8 for different water-bearing complexes.

The substantial geophysical information available and the existence of correlation between the clayiness of loose sediments and the coefficient of transmissibility have made it possible to reveal clearly the following trends observable in the multilayered formation along vertical and horizontal:

– relatively narrow ranges of average transmissibility coefficients both in horizontal and vertical direction;

– considerable dispersion of their particular values;

– varying vertical distribution pattern for average coefficients of transmissibility;

– the main direction of changes in transmissibility follows in principle that of the predominant removal of material;

– local areas characterized by higher average coefficients of transmissibility coincide with areas of local tectonic structures.

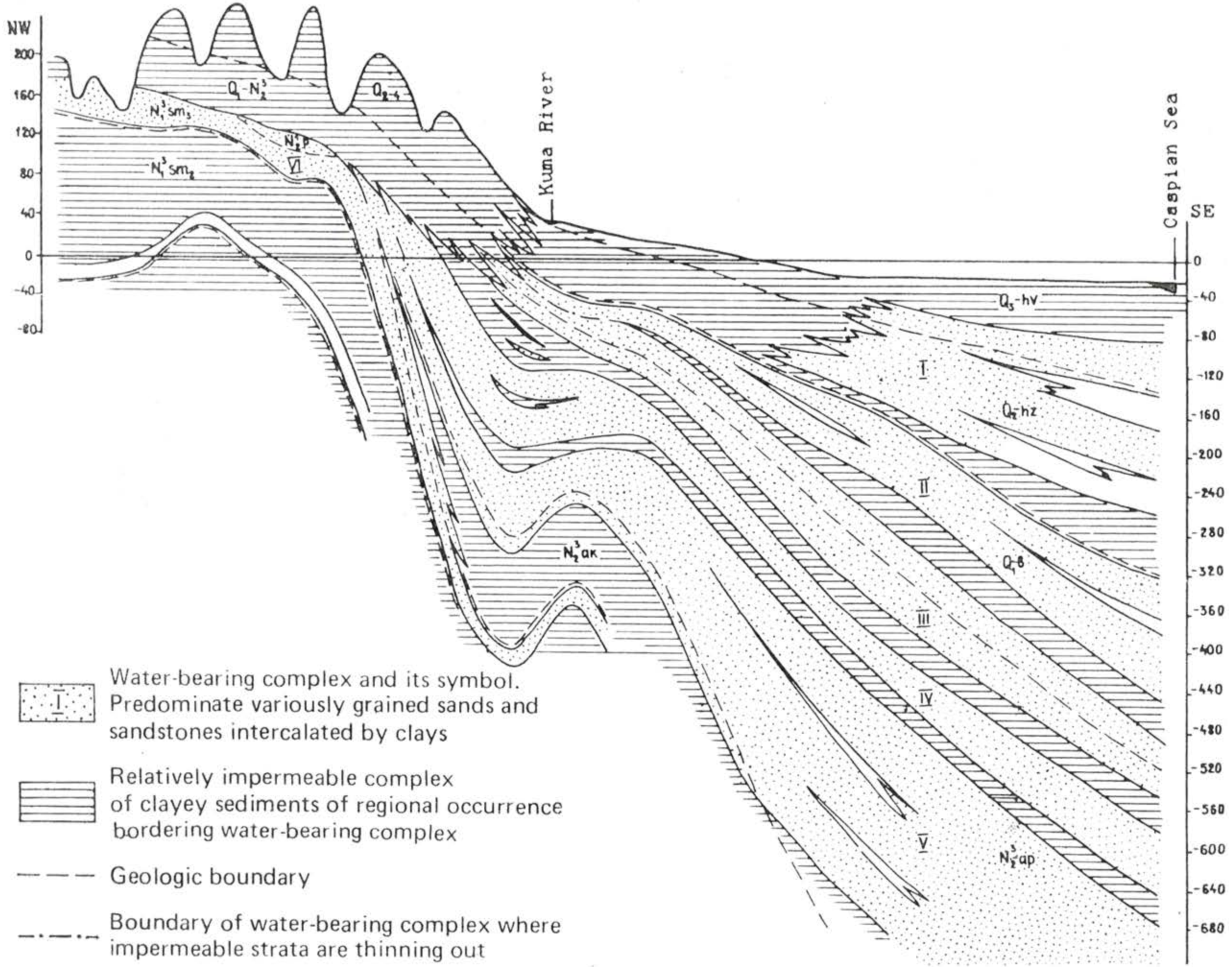

Fig. 1. Diagrammatic hydrogeologic section

Statistical treatment of the transmissibility coefficients has made it possible to elucidate a close relationship between their averages and the lithologic facies of the basin.

By geophysical techniques the occurrence of relatively impermeable clayey layers has been discerned, which is of interest in evaluating the relationship among water-bearing complexes. These techniques enabled us to detect large permeable patches associated with the lateral replacement of clayey sediments by more permeable lithologic varieties. Geologic and geomorphologic data collected from permeable patches indicate that these occur in areas of vigorous neotectonic activity in the eastern Terek-Kuma interfluve. Investigations have shown that it is within these patches where the subsurface waters of the artesian water-bearing complexes are mainly discharged.

Areal vertical (upward) migration of artesian waters in areas of neotectonic structures, which are local, is clearly shown by piezometric maps prepared for a number of water-bearing complexes and by maps of temperature contours made for certain depths.

The relationship among water-bearing complexes was studied by analyzing changes in piezometric levels and gradients of the head in vertical direction. Numerous head vs. depth plots, drawn for wells, were interpreted, and as a result three patterns of changes in the head were distinguished (Fig. 2). A relatively narrow zone follows the western and southern margins of the basin, where the piezometric levels progressively decrease with the depth of the interval tested. In the northern, southern, and eastern parts of the basin the head rises with the depth of the aquifer. The picture is complicated in the central part of the basin. Here, piezometric levels are descending from the ground water surface to a certain depth, but further downwards the heads are rising progressively. Levelling of piezometric surfaces takes place in a more permeable water-bearing complex generally overlain by poorly transmitting strata. This pattern of changes in the head is probably caused by the draining action of the water-bearing complex and also by a considerable isolation, as to their transmissibilities.

To acquire knowledge of the spatial hydrodynamic structure of the basin and to reveal general regularities, there was made a system of hydrodynamic profiles in characteristic directions and drawn piezometric contours. An example of such a hydrodynamic profile is shown in Figure 3. Profile directions were chosen with due account of the geologic structure of the basin and the main regional flow direction. To compile the profiles, wells were used where water-bearing rocks were tested interval by interval.

The head contours thus obtained reflect the general hydrodynamic structure of the basin in a given section. A combined analysis of the system of profiles makes it possible to reveal its spatial regularities and the character of changes in the horizontal and vertical gradients present in the area. With a close relationship between the hydrodynamic conditions in the basin and its hydrodynamics, the profiles show zones of the subsurface waters mineralized to varying degrees.

In an illuminating manner, the profiles reflect the principal elements of the structure of flow in vertical plane. In the southern part of the basin, percolation along bedding planes predominates. Here, the value of the longitudinal gradient of subsurface water flow is much greater than that along vertical, which indicates subsurface water recharge. The character of flow

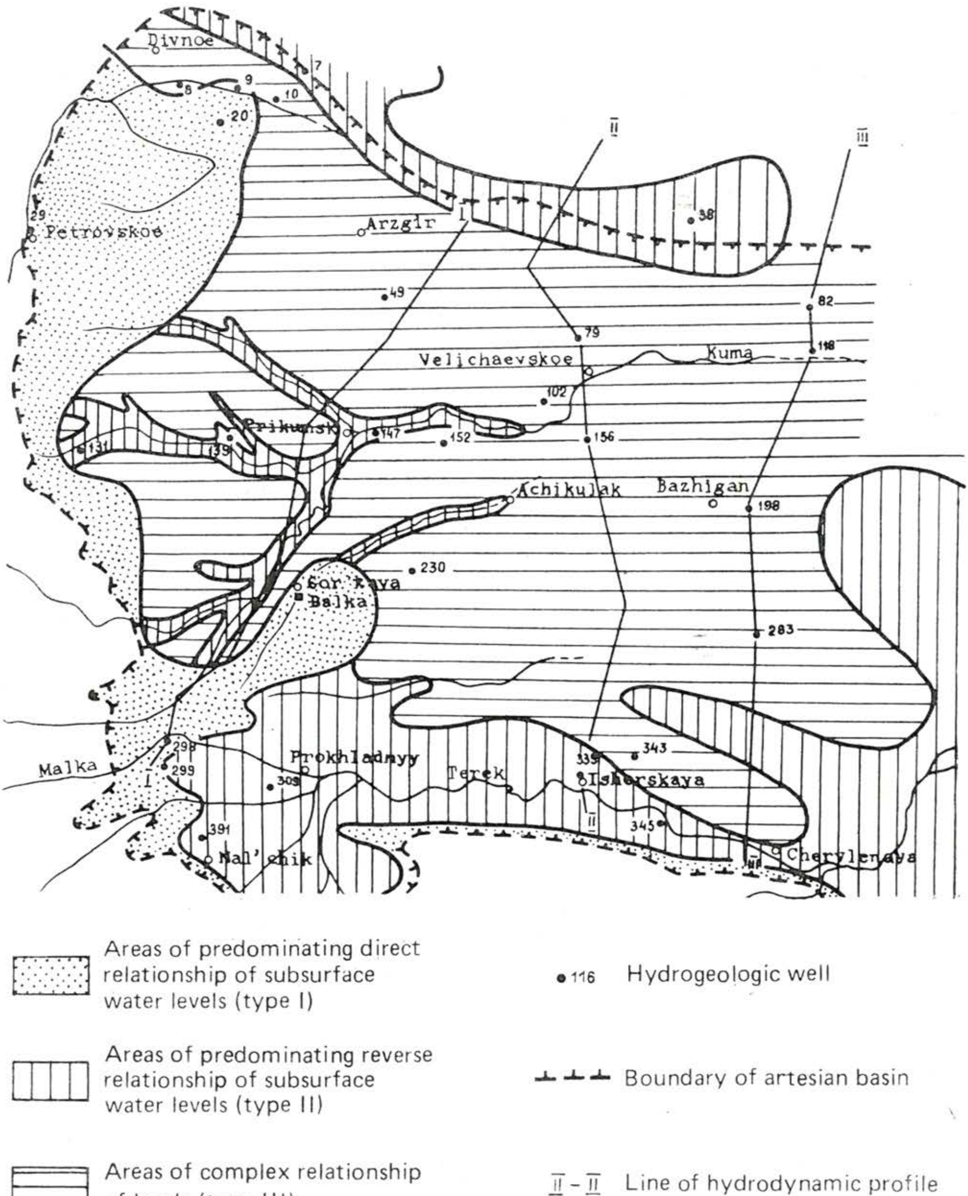

Fig. 2. A diagrammatic map showing hydrodynamic zones of the Terek-Kuma artesian basin

changes greatly northwards. Here, the gradient of flow is much greater transversely (upwards) than longitudinally, which suggests discharge of subsurface waters.

Analysis of relations among piezometric levels, the distribution of areas of the same type of level changes with depth, and the system of hydrodynamic

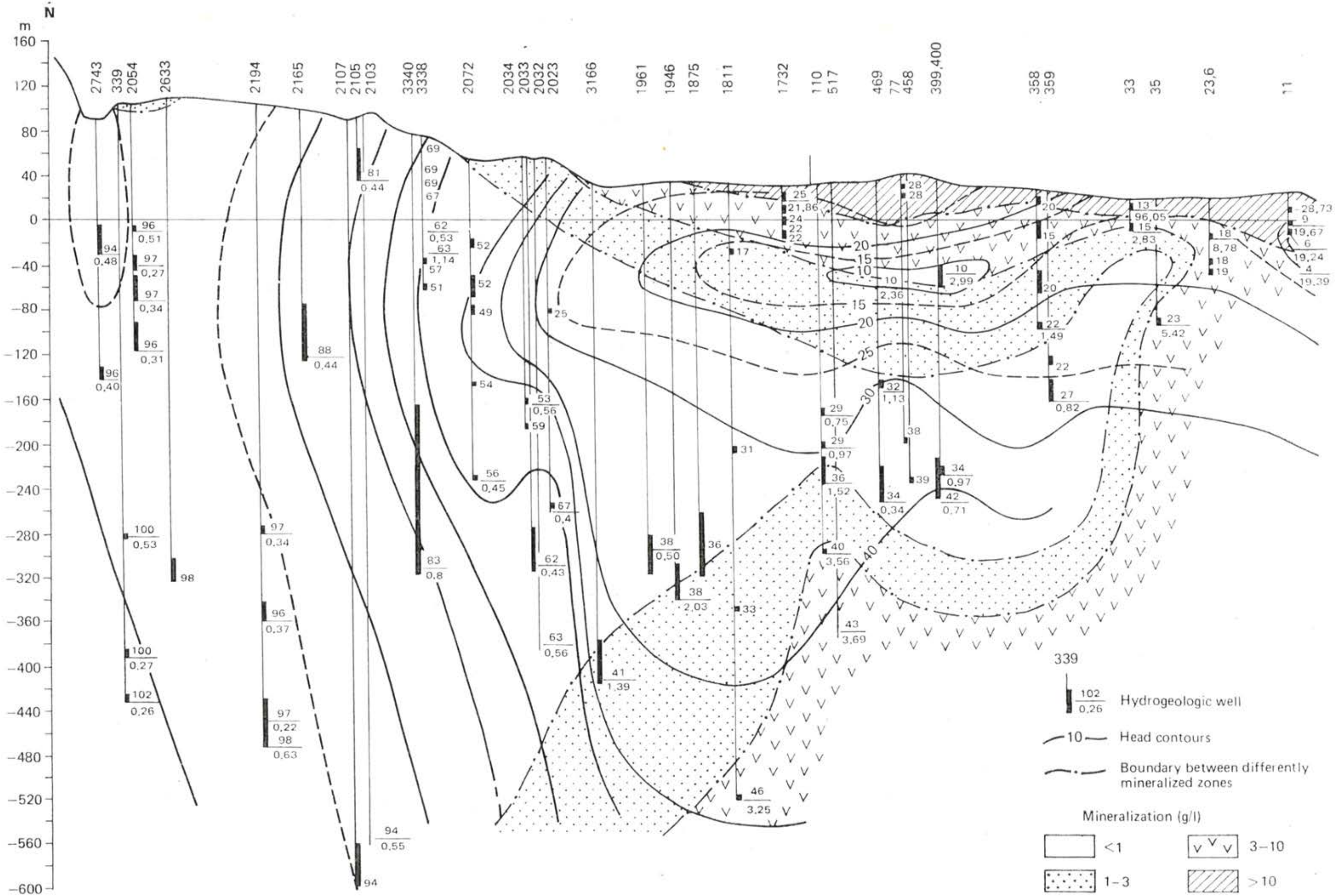

Fig. 3. Hydrodynamic profile II–II

profiles have thus made it possible to get an idea of the hydrodynamic features of subsurface water flow, distinguish areas of recharge, transit, and discharge, and elucidate relationships among water-bearing complexes.

The result was the preparation of diagrams of hydraulic relationship for every water-bearing complex. These diagrams show the occurrence of the complexes and their areas of recharge and discharge effected by overlying and underlying water-bearing complexes.

Additional data on relations among subsurface waters were obtained by compiling temperature maps for certain depths. We used not only natural temperature logs but also temperature logs recorded in flowing wells. It was suggested that the temperature at the well head equals that of the aquifer provided for a sufficiently large artesian discharge. The temperature anomalies revealed confirmed the existence of areal vertical migration of subsurface waters, as was suggested from the analysis of hydrodynamic test results.

In conducting the regional estimation of subsurface waters, studies should be made of the relations of resources with surface waters and of the role of rivers in the recharge of artesian basin resources. It should be taken into account that the history of a river valley may be complicated; hence, the sections of the valley whose geological ages are different may be incised to different depths and differ in lithological composition of alluvium and in trends of relations between river waters and subsurface waters. Therefore, to assess correctly the importance of a river valley for the water balance, the valley should be zoned by trend in interconnection process (river discharges or recharges of subsurface waters) and by the intensity of the process. During such zoning, there are taken into account the paleogeology and paleogeomorphology of the river valley, the lithologic-facial characteristics of the alluvium, and the relationship between the levels of subsurface and river waters.

Studies of the above problems have shown that, on the contrary to previous views, the Terek and Kuma Rivers do not recharge subsurface waters at the overdeepened sections of their middle course. It has been found that they recharge the subsurface waters only in their upper reaches where they leave the mountains to enter a piedmont plain. The role of the rivers as recharge sources decreases farther downstream, and they discharge subsurface waters in their middle courses. Water percolates from the river in its lower reaches characterized by a dendritic delta with the valley slightly incised and alluvium containing much clay. Here, the surface run-off is consumed by transpiration, rather than recharges surface water resources.

A quantitative estimation of the processes under discussion was made with the use of both hydrogeologic and hydrologic techniques. The former have made it possible to establish the value of recharge and discharge of subsurface (ground and artesian) waters according to observations of subsurface water regimes at hydrogeologic controls, and the latter, to prepare the channel balance determining changes in its recharging and discharging elements along the river valley. Experience has shown that the correct concept of the conditions and the intensity of processes of relationship between surface and subsurface waters on a regional scale is possible only after a combined analysis of the results obtained by the two methods. An example of the above is presented by the middle course of the Terek River. By hydrologic data, the section under consideration is characterized by a steady loss of the surface run-off, which was interpreted in previous investigations as recharge of the artesian waters of the basin. Analysis of the relationship of levels of the river and ground waters and heads of artesian aquifers indicates, however, that it is impossible for the subsurface run-off to be recharged at this section by the surface run-off.

Further investigations have shown that on the whole the valley in the middle course of the Terek River discharges the subsurface run-off. The major role is played in this process by the flood plain of the valley which is entered by ground waters, recharged by infiltration within vast river terraces, and by artesian waters, discharged as a result of flow from the aquifers below to those above, through slightly permeable clayey layers. The flood plain is also entered by river waters that are river run-off losses, according to hydrologic records. Within the flood plain, the entering waters are mainly consumed by evapotranspiration from the ground water surface. In the semiarid areas, evapotranspiration is considerable. This greatly affects the estimates of the recoverable subsurface waters of the basin, because the lowering of the ground water level will cause smaller evapotranspiration losses, which will contribute to the resources.

Therefore, as a result of comprehensive studies based on an analysis of geologic, hydrogeologic, geophysical, hydrologic, and other information on a complex multilayered system, there were distinguished major water-bearing complexes, established the character of relationship among them, and determined the transmissibility of water-enclosing rocks and the limiting parameters of the artesian system, which has made it possible to create a detailed spatial physical model of the Terek-Kuma artesian basin whose area is about 100,000 sq.km. This transmissibility model was used to estimate, on a regional scale, subsurface water resources by analog simulation.

Summary

In conducting the regional estimation of subsurface water resources of multilayered artesian systems, the spatial physical model is extremely difficult to design and it is labor consuming. This problem can be settled only on the basis of comprehensive geologic, hydrogeologic, geophysical and hydrologic investigations.

The geologic history of the region permits elucidating the patterns of spatial distribution of water-enclosing and water-separating rocks and of changes in their permeability.

Geophysical evidence contributes much to the information on the structure of multilayered formation and its transmissibility.

Areas of the recharge and discharge of artesian waters are distinguished by hydrodynamic and temperature investigations.

A combined analysis of hydrogeologic and hydrologic information makes it possible to estimate the character and the amount of recharge and discharge of subsurface waters in river valleys.

The physical model derived from such all-round investigations reflects most thoroughly and correctly the natural conditions of the region.

HYDROGEOLOGICAL CONSIDERATIONS ON UNDERGROUND STORAGE OF LOW-CALORIC ENERGY

CONSIDÉRATIONS HYDROGÉOLOGIQUES SUR L'EMMAGASINAGE SOUTERRAIN D'UNE ÉNERGIE DE VALEUR CALORIFIQUE INFÉRIEURE

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RÉSUMÉ

L'emmagasinage dans le sous-sol des déchets de l'eau chaude des industries et des services publiques et la récupération pendant des périodes de haute demande doivent contribuer à la conservation d'énergie et à la réduction de la pollution thermale. Les problèmes et les limitations de l'injection des eaux dans des roches réservoirs sont comparables à ceux du rejet souterrain des déchets, comme la nécessité de protéger des eaux utilisables, aussi que des restrictions hydrogéologiques: la pression maximum et la vitesse d'injection limitée, la capacité de la couche réservoire, la mobilité relative des eaux interstitielles et injectées, la compatibilité des eaux et les réactions du gisement. D'ailleurs, contrairement au rejet des déchets, une grande partie de l'eau emmagasinée doit être récupérable sans réduction excessive de la température, de sorte que le bilan énergétique soit positif et l'économie du projet soit viable. Des récherches et des expériments en terrain sont justifiés.

Introduction

Industrial and public utility plants dispose large amounts of low-quality heat by cooling through air or by discharge into surface waters. Apart from the hazards of thermal pollution, this energy, generally in the form of hot water, is irrevocably lost. Variations in diurnal and seasonal demand for electricity, for instance, cause a discrepancy between production capacity and consumption; it appears to be attractive to cover peak demand by stored energy. A component of such a "total-energy system" could be the underground storage of hot waste waters during summer to be recovered during winter. MEYER and TODD [11] consider this an economically favourable approach; HERRMANN [6] discussed the storage of compressed air in a solution cavern in rock-salt near Bremen (Western Germany). In The Netherlands under the auspices of the Netherlands Organization for Applied Scientific Research TNO the feasibility of underground storage is studied along with the possibility of the development of low-caloric geothermal energy.

Underground storage of energy, whether by means of drilled wells in porous and permeable formations or in rock-salt caverns, implies similar problems and encounters similar limitations as underground waste disposal. Underground waste displacement can be done if it is done with care and with the application of all science and technology we possess; if properly carried out it is a useful contribution to the solution of the waste problem of modern industrial society [1, 3, 5]. Thus underground storage of hot waste water can contribute to the conservation of energy. Unlike underground waste disposal, however, a considerable part of the stored water must be recoverable.

Techniques and limitations of underground storage

The techniques and limitations of underground waste disposal, viz. emplacement in reservoir rocks or in rock-salt caverns, are well known; there exists a considerable literature and much practical experience has been gathered [1, 3, 5]. Similar techniques are appropriate storing energy underground.

A liquid of a composition different from that of the interstitial native water is injected under pressure into a confined reservoir rock, for instance a sandstone or limestone bed of favourable porosity, permeability, thickness and extent, that is under- and overlain by impervious beds. In case of underground storage of energy, the injected liquid will be fresh water. Its temperature should be at least 70 °C, preferably 100° or more; its viscosity and density are less than 0.45 cP and 0.98 g/cm³ respectively. Leakage of this fluid toward beds that contain subsurface waters useful for human consumption or to the surface should be prevented, because apart from loss of energy, it possibly implies thermal pollution. Hence storage should be carried out only in confined aquifers well below those containing fresh or brackish waters, preferably in saline and stagnant aquifers that are isolated by extensive impervious beds from the present hydrologic cycle [12]. This means storage at rather great depths, say, 500 to 1000 m. Here temperatures vary from about 25 to 50 °C; the interstitial waters are saline with viscosities and densities greater than 0.6 cP and 1.03 g/cm³ respectively. In this depth range pore pressures increase from about 50 to 100 kg/cm², overburden pressures from about 100 to 230 kg/cm².

The differences in viscosity and density between stored and interstitial waters cause the former to move more freely through the aquifer than the latter and to float upward. Pore pressures should be sufficiently large to prevent boiling of the injected water.

It goes without saying that a cavern in rock salt is not a suitable receptacle for fresh hot water. Solution may cause an unwanted enlargement and there is the problem of disposal of the brine after recovery. A rock-salt cavern, as discussed by Herrmann [6], can be used to store compressed air. The technique of artificial recharge in an unconfined fresh water aquifer is well known and the method is widely practised. Storage in the non-saturated zone of water below boiling point at atmospheric pressure is feasible, but may meet thermal pollution.

In the following the storage will be considered of hot water in confined saline aquifers that are isolated from the hydrologic cycle.

a) *Maximum injection pressure and storage capacity*

In a confined aquifer space can be created only by injection under pressure. Water and rock matrix will be compressed, whereas the formation will dilate. These elastic effects are small; those for water amount to a compression of

$5 \cdot 10^{-5}$ per atmosphere pressure, for rocks the combined effect decreases toward depth: at 500 m about $8 \cdot 10^{-5}$, at 1000 m about $5 \cdot 10^{-5}$ per atmosphere. If, however, by injection pore pressures are increased above the so-called "break-down pressure", the cohesion between the rock particles will be impaired and cracks will appear in the rock, about horizontal at depths smaller than 300 to 500 m, vertical or near-vertical at greater depths (HOWARD and FAST [7], HUBBERT and WILLIS [8]). Fracturing is a useful stimulation technique in the production of oil; it should be avoided in underground disposal and storage. Continuous high pore pressures may cause the fissures to extend through the confining beds, and thus cause leakages. Break-down pressures vary from 0.6 to 1.2 times overburden pressure, and to remain on the safe side injection pressures should be well below these; at a depth of 500 m not above, say, 40, at 1000 m not higher than 100 kg/cm^2.

Radial flow from a well into a porous and permeable medium that is homogeneous and isotropic, of uniform thickness and of infinite extent is approximated by DARCY's formula that relates the various parameters in a linear function.

The force required to maintain the desired injection rate is mainly determined by permeability and thickness of the reservoir bed and the viscosity of the injected fluid. The quantity per unit volume that can be stored depends on porosity and thickness. A mathematical treatment is given by VAN EVERDINGEN [4], who also clearly defined the conditions under which the given solutions are valid; KOROTCHANSKY [10] discussed a case history.

In an ideal reservoir bed, for which DARCY's formula is valid and that has a permeability of 0.25 D and a thickness of 1000 cm, about $12 \cdot 10^3$ and $30 \cdot 10^3$ cm^3/sec of hot water with a viscosity of 0.45 cP can be injected

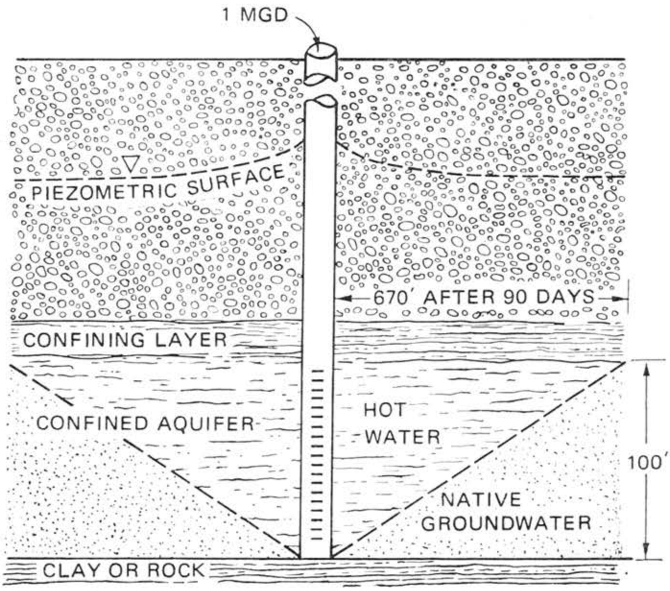

Fig. 1. Heat storage in a confined aquifer. From MEYER and TODD [11]

under forces (i.e. the differences in pressure between the fluids in the well bore and in the rock pores) of 40 and 100 kg/cm^3 respectively (about 40 and 100 m^3/h or in half a year's time, if injection is continuous, 170,000 and 400,000 m^3 respectively). In such a reservoir the hot water will form a body of an inverted conical shape around the injection well (Fig. 1). If total displacement takes place, after half a year of injection the quantities mentioned will occupy inverted cones with radii at the base of the upper confining bed of about 290 and 450 m respectively.

However, an ideal reservoir does not occur in nature; permeability and porosity vary within the aquifer, semi-impervious streaks are present, thickness changes and the reservoir is not infinite; whether total displacement occurs is doubtful. The figures therefore give no more than an indication of what really happens.

b) *Relative mobility of native and injected waters*

Due to its smaller density the injected hot water will tend to float on the cooler, more concentrated and denser native water; in other words it will tend to raise within the aquifer until its way is blocked by the overlying impervious bed. Due to the small rate of natural flow, subsequent up-dip migration of the injected fluid will, as a rule, be negligible. In some cases, for instance, the presence of steeply dipping beds and a highly permeable aquifer, structural closure will be advantageous to prevent loss of water.

Due to its smaller viscosity the injected water is able to move more freely through the aquifer than the original interstitial water. It will have a preference to flow through the more permeable parts, the "fingering" of the injected fluid. Fingering causes the front of the injected water to be irregular, while a considerable amount of the native water may be held around the rock grains by capillary forces.

The effect of viscosity on the relative mobility of two liquids is well known in petroleum production; water may by-pass and even block the oil. An example of the effect of inhomogeneities in the aquifer is given in Fig. 2. The results of a flow-meter survey in a producing groundwater well in The Netherlands show considerable variations in specific yield.

Brown and Silvey [2] discussed at length the results of experiments on storage and retrieval of fresh water in a brackish-water aquifer in Virginia (U.S.A.). Fig. 3 is taken from their publication. Although the screen is 90 feet long, 90% of the injected water is taken by 10 feet of aquifer.

c) *Compatibility of injected and native waters and the sediment*

The chemical and physical equilibrium that in the course of geological history is attained between sediment and interstitial water will be disturbed by the injection of another fluid. This is well known from underground waste disposal and as a rule it tends to impair permeability. In the experiments discussed by Brown and Silvey [2], who were quoted before, the change of electrolyte concentration caused in this particular case the dispersion of clay particles. This lead to an irreversible 40 to 50% reduction of the permeability,

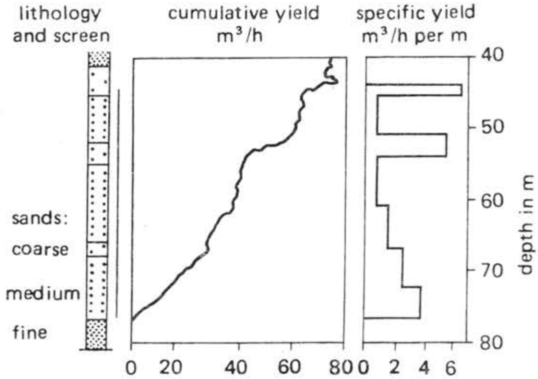

Fig. 2. Flow-meter survey in a producing water well in The Netherlands. Courtesy: Groundwater Survey TNO; Delft

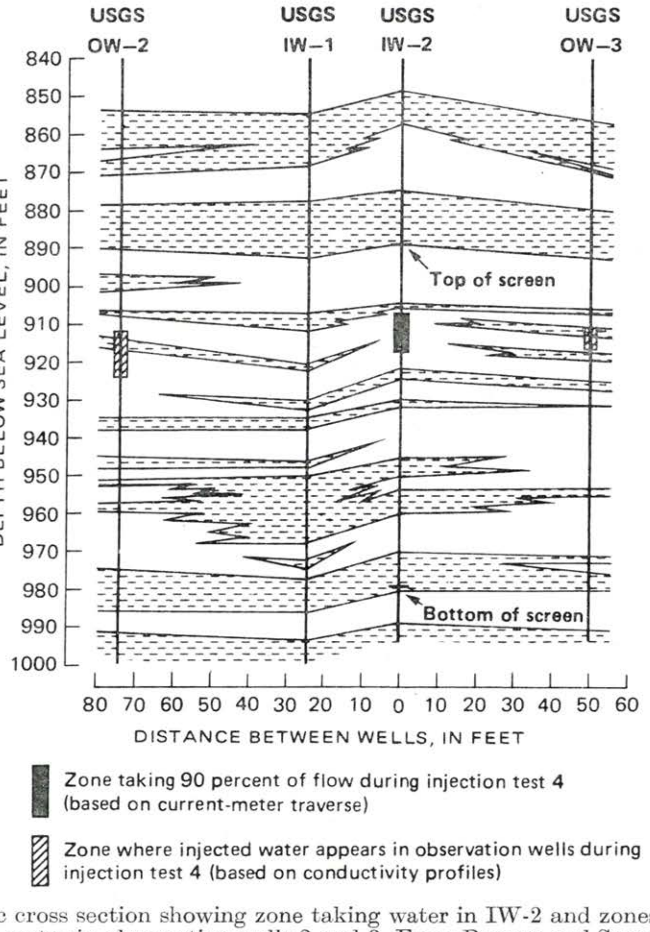

Fig. 3. Geologic cross section showing zone taking water in IW-2 and zones of detection of fresh water in observation wells 2 and 3. From BROWN and SILVEY [2]

causing an abnormally high pressure build-up around the well, changes in the flow pattern in the aquifer and a diminished rate of injection and capacity.

As a general experience in underground waste disposal, it was found that oily substances or clay particles in the injected fluid will clog the well screen and the reservoir. Microbial action may likewise have adverse effects. Pure liquids or solutions only are admissible for disposal. In spite of careful pretreatment of the waste, in the long run reservoir properties deteriorate, and the life of a waste disposal project is restricted to a few tens of years. The same is to be expected in storing hot water underground.

d) *Aquifer evaluation*

Before an underground storage project is undertaken, the reservoir properties of the proposed aquifer have to be evaluated carefully. As the stored fluid should remain around the well screen in a body as compact as possible, the degree of uniformity and homogeneity of the aquifer is of paramount importance. Further, the aquifer should have a thickness and an extent, a porosity and a permeability sufficient to store the desired quantity at the desired rate.

The sealing properties, especially of the overlying bed, along with the break-down pressure, should be investigated. The presence of structural closure may be advantageous in some cases.

Chemical and physical compatibility should be investigated in the laboratory; pretreatment of the hot water or even of the aquifer may prove necessary.

Conservation of heat and retrieval

Due to loss of heat to the immersed rock particles the injected water will cool. Assuming that all native water is displaced in a conical body around the well screen, then—at a porosity of 20%—a rock volume four times as large as that of the injected water has to be heated. As the specific heat of sand is about 0.45 per cm^3 per degree centigrade, this means that during the first storing/retrieval cycle the temperature of the water will decrease by some 25 to 30 °C. Due to the small heat conductivity of rocks—for water saturated clays and sands 5 to 10 $microcal.cm^{-2}.sec^{-1}$ per degree centigrade—loss of heat to the rock mass that surrounds the injected body is negligible. During retrieval, however, the pores will become filled again with native water of the ambient aquifer temperature. It is to be expected that after a few storing/retrieval cycles the temperature balance will improve.

Loss of heat will, however, be considerable, if by fingering and floating the body of injected water becomes irregular in shape. The area of contact will then be greatly enlarged, and heat losses will be determinable only by field experiments. Loss of heat in such a case may jeopardize the project.

From field- and model experiments, reported on by Brown and Silvey [2] and Kimbler et al. [9] respectively it appears that 75 to 90% of the stored water can be recovered; contamination with saline and cold native water will set an upper limit to the amount retrievable. Too great a percentage of native water has a disadvantageous effect on the temperature and, moreover, may cause difficulties in disposal. The recovered water must not be corrosive or otherwise harmful to the environment. After use it can only be disposed in surface waters; return to the subsurface is impracticable.

Economic aspects

The economic value of hot waste water stored underground is the cost of the conventional fuels that are saved after retrieval and use. Another, intangible, factor is the reduction of thermal pollution of atmosphere and surface waters. Underground storage can only be seen in relation to a total energy system in a production/consumption centre. The recovered water is to be used in district heating and agriculture to cover peak demand; if feasible its temperature could be increased by heat pumps to generate electricity. Another consideration in achieving an optimal result is the temperature at which the water to be stored is drawn off.

The economic factors to be considered can be expressed in terms of money and of energy; in the end both should be positive. Exploration for a suitable aquifer at not too great a depth and at not too great a distance involves risk; it may be proved that the looked-for aquifer is not present. Costs of drilling and aquifer evaluation, of well completion and maintenance increase with depth. Laboratory investigations and possibly necessary pretreatment of water and aquifer involve an increment of costs. On the one hand, the reservoir properties and temperature of the aquifer and the salinity of the original interstitial water determine the rate of injection and capacity, the quantity and temperature of recoverable water and the duration of the project. On the other hand, the draw-off rate and the temperature of the water to be stored and peak demand determine, together with the distance, the costs of construction and maintenance of the surface installation and of the transport- and distribution facilities. A further factor is the cost of disposal of spent water that, preferably, should possess a temperature and a composition comparable to those of the prevailing surface water.

It is clear that, as is the case with other forms of geothermal energy, investment costs are high, maintenance costs are low, such in contrast to conventional power plants. The rate of availability of waste water at the surface and of stored water in the subsurface should balance, and large amounts are advantageous to reduce investment costs per unit volume. A long life expectancy is important in amortization of the investment.

It follows that, before an underground storage project is launched, many factors should be carefully considered and analysed. Although theoretical considerations indicate that, if an ideal aquifer is assumed, storage and retrieval of hot water is economically feasible, a practical application should be proceeded by a pilot project. Such a project involves selection of an area and study of existing geological and hydrogeological data, field experiments through drilling, injection and retrieval.

Conclusion

Underground storage and retrieval of hot waste water from factories should result in a positive energy balance and should be economically viable. The reduction of thermal pollution of atmosphere and surface waters should be taken into consideration.

These requirements will be adversely affected by the reservoir properties of natural aquifers and their lithological composition and by the salinity of the native interstitial waters. These factors restrict the rate of injection, the capacity, the retrievable quantity and the duration of a storage/retrieval project. The search for a suitable aquifer involves risk money.

The possibilities of this method to conserve energy and to reduce thermal pollution are considered of sufficient importance to warrant further research and the execution of a pilot project.

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GEOTHERMAL MODEL FOR THE RISE OF DEEPER GROUNDWATER ALONG FAULTS IN UNCONSOLIDATED ROCKS

MODÈLE GÉOTHERMIQUE POUR LA MONTÉE DES EAUX SOUTERRAINES DE GRANDE PROFONDEUR LE LONG DES FAILLES DANS DES SÉDIMENTS NON CONSOLIDÉS

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RÉSUMÉ

Un simple modèle est présenté, lequel permet d'expliquer les anomalies de la température trouvées dans des sédiments non consolidés, perturbés par des failles.

Dans ce modèle, la montée d'eaux souterraines le long de failles est admise.

Pour la calculation on se servirait des mesures de la température effectuées dans le Niederrhein-Bucht. Une relation quantitative a été trouvée entre l'anomalie thermique et le flux de l'eau montante. Comme résultat principal, il est démontré que la quantité d'eau nécessaire à l'explication de ces anomalies est relativement très petite. Ce modèle ne mène pas à de contradictions hydrologiques.

Introduction and generalities

Pronounced geothermal anomalies are most commonly observed in tectonically disturbed zones, where hydrological anomalies generally occur in conjunction with the latter. It is thus appropriate to investigate models which may describe the relationship between geothermics and hydrogeology. The Niederrhein-Bucht is presented here as an elementary set-up for such models. To date there are but few areas in the Federal Republic of Germany which have received sufficient geothermal investigation as to be suitable for consideration. Among these few ones there exist the Erft region, which is dealt with here (Balke, 1973), the northern Oberrhein-Graben (Werner – Doebl, 1974) and the Schwäbische Alb (Carlé, 1974).

The Niederrhein-Bucht is a young area of depression which received its tectonic character during Tertiary and Quaternary times. The area is composed of several component blocks (Fig. 1), the result of tectonic dilation (Ahorner, 1962, Teichmueller, 1974). The greatest crustal stretching occurred in a SW – NE direction, in other words normal to the Erft fault and the fault system of the Rurrand.

The pre-tertiary basal complex consists of Devonian slates and graywackes, covered from place in angular unconformity by Mesozoic sediments. The Tertiary is represented by a succession of gravels, sands and clays with numerous lignitic seams (Fig. 2). These beds are overlain by Pleistocene terrace gravels.

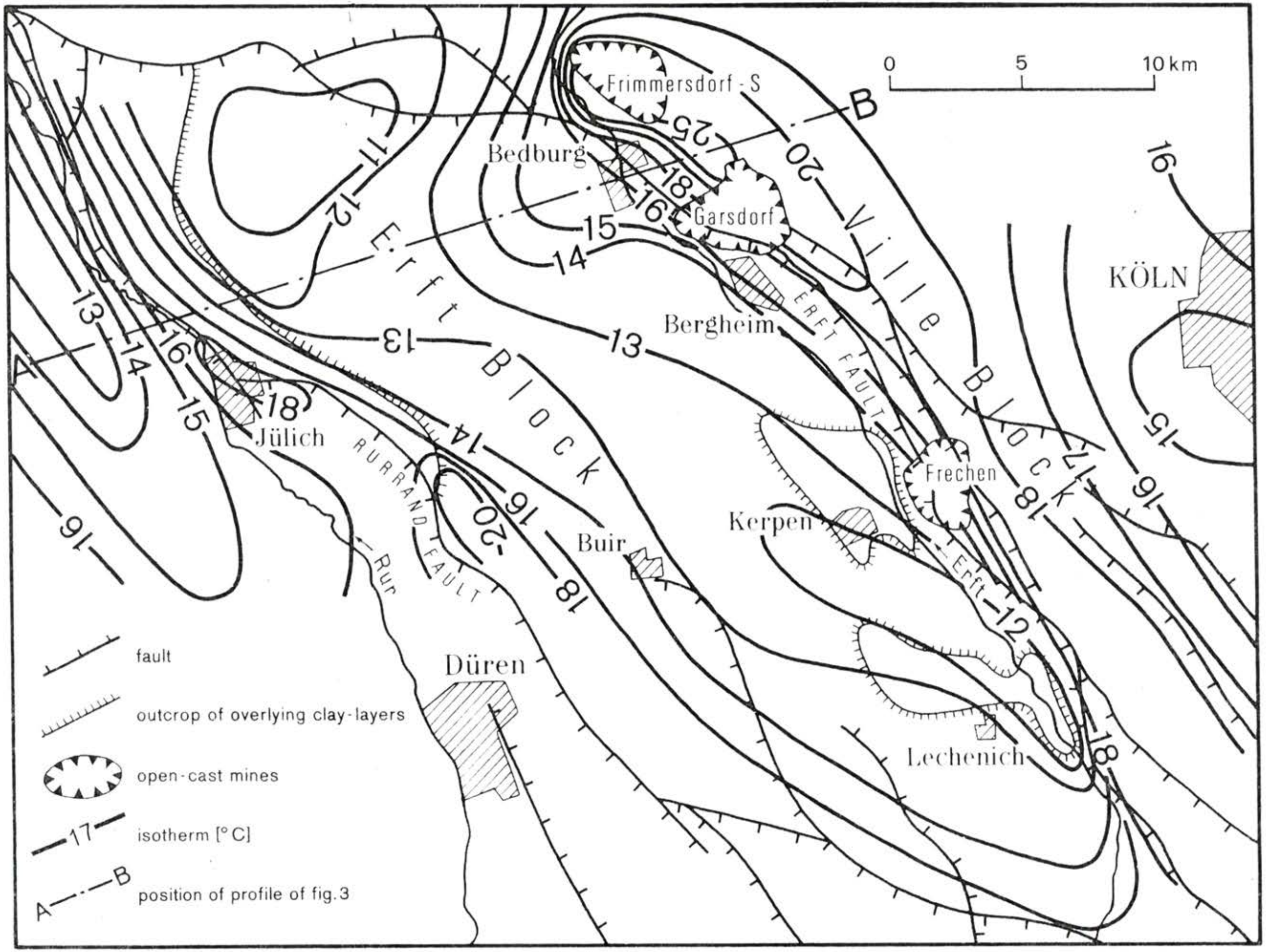

Fig. 1. Isothermal map of the Erft region (Niederrhein-Bucht). Temperatures relate to a depth of 100 m below sea level. The temperature anomaly of the Erft fault is striking

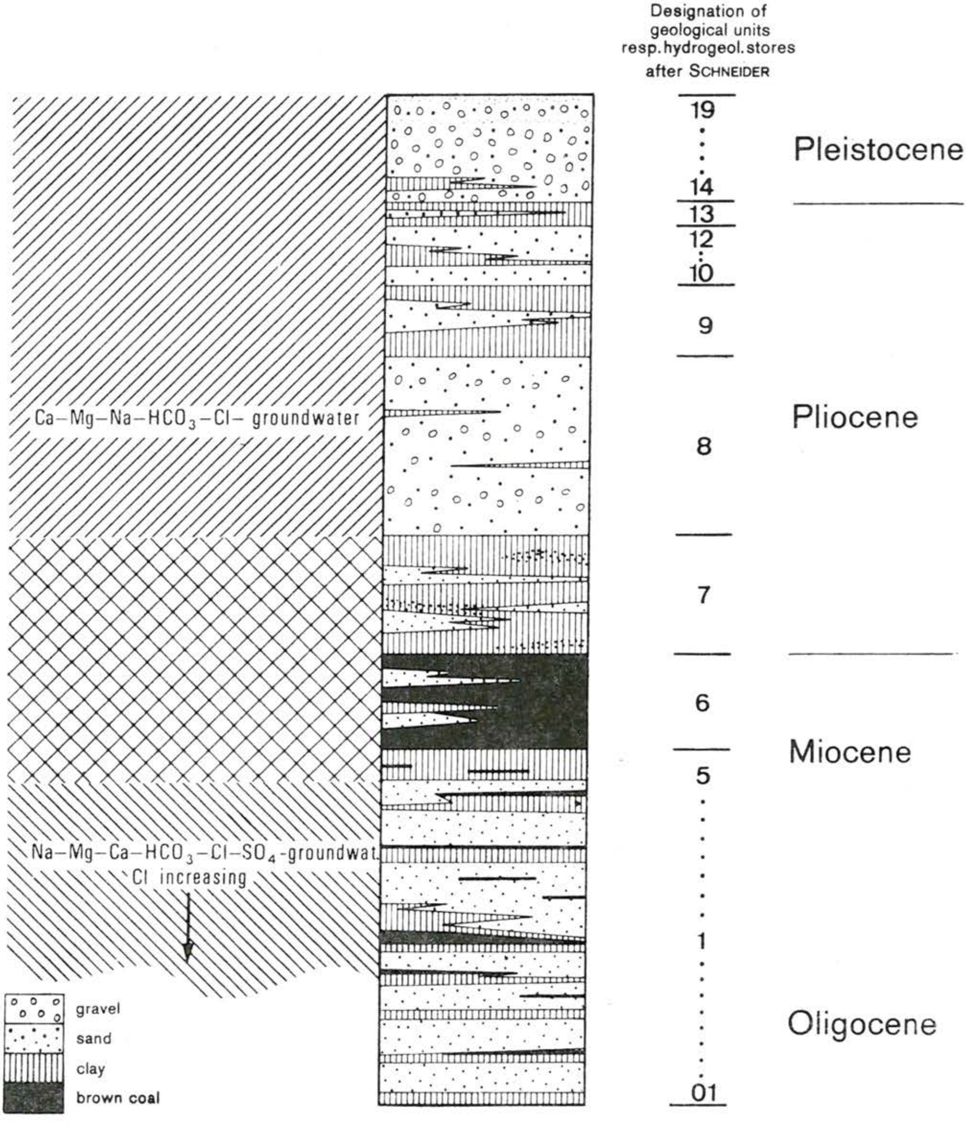

Fig. 2. Schematic geological sequence of the Niederrhein-Bucht

Since 1955 open-cast mining of the lignite from the Ville Block has been carried on, which necessitated lowering of the water table below mining level. It was noted that in several drainage shafts – especially in the case of the open-cast mine Fortuna-Garsdorf – water rose from a depth of 300 m with a temperature of about 40 °C.

The situation has been presented and analysed by BALKE (1973). Numerous temperature measurements in groundwater depth gauges have revealed normal geothermal conditions within the Erft Block. In the neighbourhood of the Erft fault system and, to a lesser extent, at the Rurrand fault too

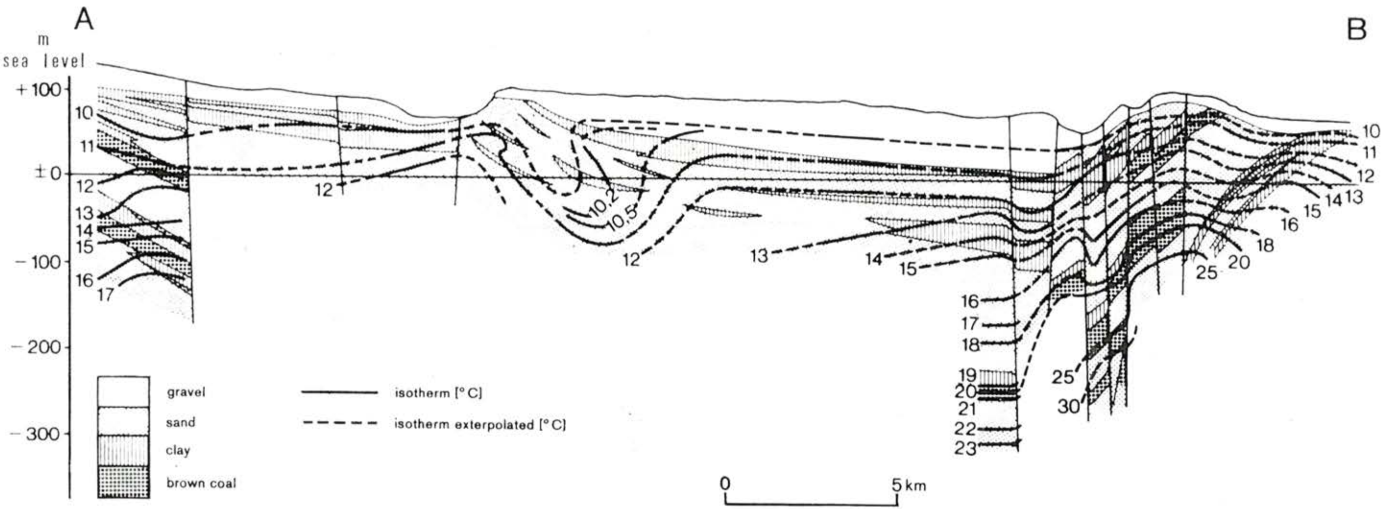

Fig. 3. Temperature profile A – B (cf. Fig. 1). The thermal anomaly along the Erft fault is explained as a hydrological consequence of the fault

considerable temperature increases were observed (Fig. 1 and 3). Fig. 1 represents temperature distribution in the form of isotherms at a depth of 100 m below sea level. The anomaly is especially pronounced in the Ville Block near Bedburg. The thermal profile (Fig. 3) and the isothermal map (Fig. 1) indicate clearly that the positive temperature anomalies are connected with tectonic disturbances. The cause of such temperature increases may be related to rising warm deeper groundwaters.

Little is known concerning the paths taken by rising water in disturbed unconsolidated rocks. But it is quite certain that open joints cannot play a role. It seems reasonable to attribute a certain vertical permeability to the zones of disturbance. This is the consequence of the dislocation of aquifers and impermeable beds. The case may be illustrated in a simplified and idealized form in Fig. 4. One can say, that the disturbances are "smeared" zones.

There are observations about the materials within the fault zones. In an exposure of the Erft fault (at the western edge of the former opencast mine between Türnich and Mödrath) clayey as well as sandy sediments were encountered which had been dragged into the fault zone (AHORNER, 1962).

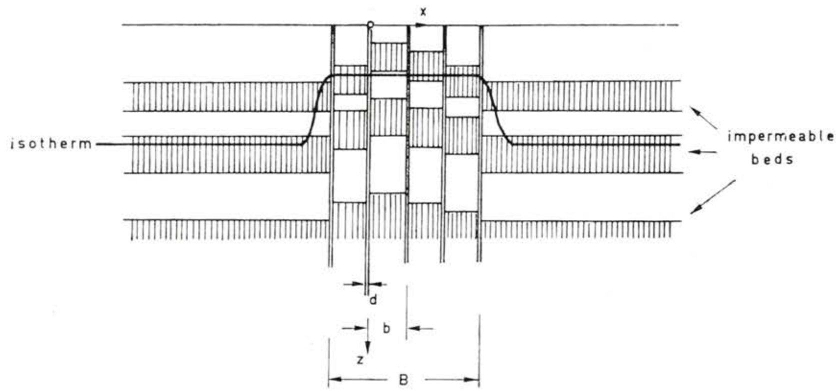

Fig. 4. Idealized model of the fault system. In the narrow faulting zones ($d \approx 3-10$ m) warm groundwater rises and warms the intervening blocks ($b \approx 1$ km)

More detailed investigations of flow conditions of deeper groundwaters in the Erft area do not exist. This paper relies solely on temperature data (BALKE, 1973). Using our model a quantitative relationship between temperature anomaly and flow of deeper groundwater can be established.

The question is: what extent of rising warm water is required in order to explain, within the context of the model, the measured temperature increases.

Model of calculation

In order to attain a practical mathematical formulation it is necessary to idealize the geological situation. A half space with uniform surface temperature (mean annual temperature T_0) and constant thermic material parameters (K=thermal conductivity, $\varkappa$=thermal diffusivity) are given. Within this undisturbed system the existence of an area of width B is presumed (Fig. 4), which is typified by two characteristics: *1.* an increased temperature, *2.* parallel vertical zones of faulting of width d. Between these fault zones d there exist thermoconductive (but without groundwater flow) blocks of width b. The blocks b are about 1 km across, whereas the width of the fault zones d lies between 3 and 10 m (AHORNER, 1962). Thus it follows $d \ll b$. In the model the fault zones d are treated as surfaces on which isothermal boundary conditions obtain.

This raises two problems: *1.* The calculation of the warming of blocks b under isothermal boundary conditions, *2.* the temperatures obtaining in the faults (zones d) must be quantitatively related to the rising groundwater flow.

Problem *1* serves to prove that the warming of the block b is a fairly quick process compared to the age of the fault. A component part b of the half space is considered which is surrounded by three isothermal surfaces; the surface and two fault planes. The calculation is two-dimensional and is based on a difference procedure corresponding to the equation of thermoconductivity:

$$\frac{\partial \Theta}{\partial t} = K\left(\frac{\partial^2 \Theta}{\partial x^2} + \frac{\partial^2 \Theta}{\partial z^2}\right) = 0. \qquad (1)$$

We distinguish a stationary undisturbed temperature field $\Theta(z)$ from a non-stationary anomaly component $\Theta(x, z, t)$. These two make up the total field:

$$T(x, z, t) = \Theta(z) + \Theta(x, z, t).$$

For the depth here considered $\Theta(z)$ is linear:

$$\Theta(z) = T_0 + \alpha z.$$

We have $\Theta(x, z, t) = 0$ at the starting point $t = 0$.

The boundary conditions are as follows:

$$\Theta(x, 0, t) = 0 \qquad \text{(surface)}$$

$$\Theta(0, z, t) = \Theta\ (b, z, t) = \Theta_0(z) \qquad \text{(fault planes).}$$

The solution of problem *2* of our model is of greater importance. The isothermal boundary conditions proposed for the fault planes require continual energy input. The energy is assumed to be supplied from water flow I. The following relationship (WERNER, 1975) between water flow I (cm²/s), temperature gradient $\mathrm{d}T_v/\mathrm{d}z$ in the fault plane, and the horizontal temperature gradient $\mathrm{d}\Theta/\mathrm{d}x$ in the immediate neighbourhood of the fault plane can be drawn up:

$$I c_w \varrho_w \frac{\mathrm{d}T_v}{\mathrm{d}z} = 2K\left(\frac{\partial \Theta}{\partial x}\right)_0 \qquad (2)$$

with c_w, ϱ_w — specific heat, density of water,
K — thermal conductivity.

Assuming an effective porosity p for the filling material of the fault zones, one can write:

$$I = pvd, \qquad (3)$$

where v represents the mean vertical flow velocity.

Our model is founded on the idealized assumption that within the thermal conductive blocks b there is no water movement. Only vertical water flows in the fault zones are considered.

Results

The solution of problem *(1)* with $\varkappa = 7 \times 10^{-3}$ cm²/s results in a temperature field as given in Fig. 5. On the right the model temperatures $T_v(z) = \Theta(z) + \Theta_0(z)$ are shown. Isotherms (continuous lines) represent the stationary temperature distribution which, assuming a block width $b = 1$ km, is attained after approx. 20,000 years. The marked sagging of the isotherms is conspicuous. This results from surface conditions which exert a cooling influence. Such behaviour cannot be demonstrated with existing temperature measurements. There are two separate reasons for this state of affairs. Firstly the observation positions were not close enough. Secondly the model is too simplified and too idealizing for such details.

In the context of the model, horizontal water movements between the fault planes are not allowed for. The model error thus caused can be estimated by supplementing equation *(1)* with a convective term:

$$\frac{\partial \Theta}{\partial t} - \varkappa \left(\frac{\partial^2 \Theta}{\partial x^2} + \frac{\partial^2 \Theta}{\partial z^2} \right) + p v_x \frac{\partial \Theta}{\partial x} = 0.$$

For this relationship one can also insert a corresponding difference procedure. Let us consider, by way of example, an aquifer at a depth of between 50 − 250 m with a horizontal flow velocity $v_x = 10$ cm/day and with $p_n = 20\%$. This results in the dashed isotherm diagram in Fig. 5. This secondary aspect serves to throw light on the problems involved in drawing up a realistic model for local hydrothermal phenomena.

The result of problem *(1)* which we anticipate is as follows. The minimum age for the thermal anomaly of about 20,000 years contrasts with that of the block faulting which began in the Oligocene, i.e. about 30 m.y. ago. Thus one can state that the time necessary for the warming up of the entire anomaly area B is small with respect to the age of the faulting.

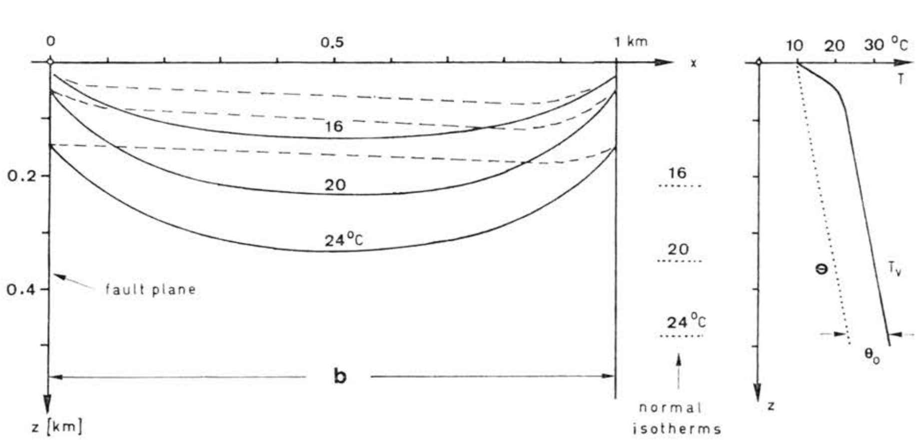

Fig. 5. Isothermal diagram of a model block $b \approx 1$ km (continuous lines). The temperature distribution is a result of isothermal boundary conditions, in our case: virtually stationary conditions after 20,000 years. $\Theta(z)$ represents the undisturbed field, $\Theta_0(z)$ the effect of increased water temperature within the fault

Problem *(2)*, the determination of water flow I from *(2)*, presume knowledge of the factor $\left(\frac{\partial\Theta}{\partial x}\right)_0$. A mean value is known from the solution of problem *(1)*. The gradient $\mathrm{d}T_v/\mathrm{d}z$ is already stipulated by $\Theta(z)$ and $\Theta_0(z)$ (Fig. 5), which were selected according to data observed in the field.

The flow of deeper groundwater can be estimated by use of the numerical values $\left(\frac{\partial\Theta}{\partial x}\right)_0 \approx 2\times10^{-4}$ deg/cm, $c_{w\varrho w}=1$ cal/cm³ deg and $K=4\times10^{-3}$ cal/cm s deg:

$$I=2\frac{K\left(\frac{\partial\Theta}{\partial x}\right)_0}{c_{w\varrho w}\frac{\mathrm{d}T_v}{\mathrm{d}z}}\approx 4\times10^{-3}\ \mathrm{cm^2/s}\approx 12\ \mathrm{m^2/annum}$$

This means: for a length of 1 m (normal to the plane of the drawing) of the fault plane an annual discharge of 12 m³ occurs. This is a very slight flow. The discharge of this water near to the surface leads to no further hydrological problems.

We give a clue as to the order of magnitude of flow velocity corresponding to flow I. It is founded on equation *(3)* where $p=10\%$, then

$$d=1,\quad 5,\ 10\ \mathrm{m},$$
$$v=35,\quad 7,\ 3.5\ \mathrm{cm/day}.$$

One can see that the factors p and v assume values which can also be found approximately in normal horizontal aquifers. This result agrees with the idea outlined at the beginning of this paper in which the faults were looked upon as "smeared" zones.

The whole problem under consideration here may also be looked upon from a simpler, generalized point of view. One limits oneself to a stationary heat flow balance. Once again we consider water flow I, which is necessary to maintain constantly the amount of heat contained in the anomaly. The sole supposition is that the entire block B is pervaded by watercourses so that the heat can be distributed. For this we consider both the mean heat flows on the surface q_A in the disturbed area B and q_N in the neighbouring undisturbed area. The difference $\Delta q=q_A-q_N$ must be made up by heat input from the deeper groundwater flow I. The thermal energy lost by the water to its surroundings in rising through the disturbed zone may be calculated from the temperature reduction ΔT of the water; however, this depends on the depth under consideration. Thus it follows:

$$I=\frac{\Delta q b}{c_w\varrho_w\Delta T}$$

b represents (as in the case of our model outlined above) a characteristic width of about 1 km.

The heat flow situation in the Erft area (BALKE, 1973) is known to an approximation. Taking $q_A\approx3.5$ HFU (10^{-6} cal/cm² s), $q_N\approx1.4$ HFU and $\Delta T\approx40$ deg one arrives at $I\approx5\times10^{-3}$ cm²/s. This forms another confirmation —we are dealing here only with approximate values—of the results already attained.

S u m m a r i z i n g one can say: the rising of deeper groundwater along faults in unconsolidated rocks proves to be a sufficiently realistic model to explain temperature anomaly. The simple model considerations presented in this paper lead neither to geothermal nor to hydrological contradictions. Unresolved, however, remain the hydrodynamic causes for the rise of deeper groundwater. This problem applies not only to the Niederrhein-Bucht, but also to the Oberrhein-Graben and the Schwäbische Alb as well as to structures which are geothermically and tectonically striking.

It would be exceedingly interesting and advantageous if hydrodynamic models, too, were to be drawn up. This would oblige geothermal research to orientate itself to rival models.

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Discussion

J. Deák comments again on the methods used in isotope measurements for the radiocarbon dating of the ground-water at different depths.

S. Krajewski, chairman of the Polish International Committee, announces that, in May 1978, Poland will organize an international conference on the hydrochemistry of mineralized waters. He invites heartily the participants in this Conference to attend the Polish meeting.

N. V. Rogovskaya expresses her regret that there is no time for further discussions. In many countries there are problems with water supply and distribution. In the Soviet Union the work is done already over large territories and the methods are also elaborated. They could be interesting for other countries. There are of cause further duties and tasks to solve, to what we may also use the results produced in other countries.

The Soviet delegation thanks Mr. Kovács and Mr. Dubertret for the invitation and congratulates them for the success of the meeting.

CLOSING OF THE CONFERENCE

CLÔTURE DE LA CONFÉRENCE

Professor S. BUCHAN president of IAH summarizes the results of the Conference:

"We are at the end of the Conference, this is the closing session. On this occasion I have to express my heartiest thanks to the organizers of this meeting. To the leaders of the opening session, to director KONDA and to DR. RÓNAI, the secretary general. I thank all the members of the Organizing Bureau and everybody who participated in the sessions not forgetting those operating the projector and those – we have not seen them – who helped to break down the barrier of the language, the translators.

I am deeply grateful to the representant of IAHS and of UNESCO, who have consponsored this meeting. It was the first occasion we have worked side by side which helped to make this meeting a great succes.

We had a wonderful welcome from our Hungarian colleagues, we were presented with a fist of excellent scientific results. We have got new ideas, we have made new friendships. It was a splendid meeting. It made a significant scientific contribution which will, I hope form a foundation for an important worldwide programme of study of the hydrogeology of great sedimentary basins.

I prefer not to see this session as an end, but a beginning of the collaboration of geologists, engineers and of all kind of other specialists to resolve our contemporain hydrogeological problems. I say my warmest thanks to all participants of this meeting and I wish further success in your scientific efforts in the future."

MR. J. KONDA Director of the Hungarian Geological Institute, chairman of the Conference closes the session and the Conference:

"The activity of the participants represents the result and this is which qualifies a scientific conference. I thank all the leaders of the two international organizations and of the collaborating Societies; the session chairmens, the general rapporteurs, the authors and those contributing to the discussion as well as all the members and participants of the meeting. I wish good success in the excursion and close the Conference."

LIST OF PARTICIPANTS OF THE INTERNATIONAL HYDROGEOLOGICAL CONFERENCE, BUDAPEST, 31 MAY TO 5 JUNE, 1976

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