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Budapest, 1981

MAGYAR
TUDOMÁNYOS AKADÉMIA
KÖNYVTÁRA

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CORRELATION OF THE HUNGARIAN TRIASSIC*

By

K. BALOGH

HUNGARIAN GEOLOGICAL INSTITUTE, BUDAPEST

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If the merely drilling-based Triassic in the substratum of the Great Hungarian Plain and southwestern Transdanubia and the still uncleared Triassic sequences of the Rudabánya Mts. be disregarded, so the Triassic of the mountains in Hungary belongs to four different facies areas (Central Mountains, Bükk, Aggtelek—Szilice and Mecsek—Villány). These are determined by the specifics and the mode of intertwining of the

* Manuscript finalized: May 15, 1980

lithological units of platform, basin, volcanic or fluviatile-lacustrine facies which followed the beach-, shallow-water marine or evaporitic sediments of the early Triassic Tethyan transgression. Facies shifts within the individual facies areas are particularly striking in the Transdanubian Central Mountains and in the Aggtelek Karst. The present writer has sought to place these lithostratigraphic units in a geochronological frame, which he judges to be best suited to this purpose of all the variants of the new international standard being still in the making. The principal statements can be gathered from the correlation tables are as follows:

The Lower Triassic of the *Central Mountains* shows a wider game of changes than was supposed heretofore. The reef-lagoonal facies of the Anisian (in contrast with the Balaton Highland and the Bakony) persist in the more eastern part of the Central Mountains even in the Ladinian. The Veszprém Marl Formation of Carnian age can be traced from Keszthely up to the Buda Mts., being intertongued here with the chert-rich Sashegy Dolomite Formation. The deposition of the Hauptdolomit Formation is not even in the southwestern Central Mountains confined in all occurrences to the Norian, while farther northeast it begins — with some fluctuations though — in ever deeper levels of the Carnian. This "descent" of the Hauptdolomit Formation is followed also by the onset of the Dachstein Limestone Formation: still definitely Rhaetian in age in the SW, the Dachstein Limestone is observed farther northwest to have begun to be formed in Alaunian and, in the horsts on the Danube's left side, as early as the Tuvalian. Interposed between the Dachstein Limestone and the Hauptdolomit Formation in the southwest, the "Kössen beds" will completely disappear in the northern part of the northern Bakony Mts. and to the east of it, to re-appear maybe only in the Feketehegy Formation spanning the greater part of the Norian (or reaching up perhaps into the Rhaetian, too) in the Pilis Mts. In connection with the disappearance of the "Kössen beds" the Dachstein Limestone Formation of the northern Bakony Mts. and Buda Mts. is partly or entirely intertongued with the Hauptdolomit Formation. It is important to know that, unlike believed earlier, the cherty Csővár Limestone Formation adjacent to the reef- and reef-lagoon facies Dachstein Limestone of the Triassic horsts on the Danube's left side is not restricted to the Carnian Stage, but it replaces the Dachstein Limestone Formation being here of Tuvalian-Sevatian age.

As a result of some conodont and foraminiferal finds the age of the lithostratigraphic units of the *Bükk area* has been pushed higher up the stratigraphic column than was earlier believed by the author.

The intricate stratigraphy of the *Aggtelek-Szilice area* has changed in that certain limestones of Alsóhegy earlier believed to belong to the Schreyeralm (i.e. Illyrian) Limestone, now, on the basis of conodonts, have proved to represent the Ladinian, Carnian and Norian members of the Hallstatt Limestone Formation. The borehole Szőlősardó-1 on the southern margin of this area has uncovered, in turn, under the Tuvalian-Alaunian Pötschen Limestone Formation, the Cordevolian—Lower Tuvalian Szőlősardó Marl Formation, hitherto unknown in the Szilice nappe. This formation, in turn, was found to be underlain by the Middle Illyrian—Longobardian Nádaska Limestone Formation representing an underwater slope facies. Below this latter — with the absence of the Lower Illyrian and Pelsonian Substages, owing to tectonic (?) causes — there follows a cherty dolomarl- and siltstone sequence with a volcanic ash content which, on account of its Bithynian (?) age, pushes, at least locally, even the age of the underlying Steinalm Formation considerably deeper than it was earlier supposed.

Palynological investigations in the *Mecsek—Villány area* have made evident the Lower Triassic age of the Jakabhegy Sandstone. The position of the known lithological units of the Middle and Upper Triassic has been somewhat modified by the revaluation of their fauna and flora.

Although the *lower boundary* of the Hungarian Triassic sequences is generally sharp but it cannot everywhere be evidenced biostratigraphically. Some of the sharp boundaries above the Permian clastic sequences are accentuated by erosional unconformity and basal conglomerates (Balaton Highland, Mecsek, Villány). The Lower Triassic of the southern foreland of the Vértes and of the Aggtelek Karst rests on Upper Permian evaporites. The Lower Triassic oölites of the Bükk, however, lie on Upper Permian marine limestones.

The upper boundary of the sequences in question varies from one mountain to the other. The Dachstein Limestone Formation of the Balaton Highland and the southern and northern Bakony and the Karolinavölgy Sandstone Formation of the

Mecsek pass uninterrupted into similar facies of the Hettangian Stage. The hiatus between the Triassic and the Jurassic in the Vértes, the Gerecse, the Pilis and the Aggtelek-Szilice mountains is relatively small. In the horsts on the Danube's left side and the Buda Mountains (where the Jurassic and the Lower Cretaceous is totally absent), however, for the moment there is no evidence enabling a judgement: how much of this hiatus may be owed to a break in sedimentation within the Mesozoic and how much to a Tertiary denudation.

Preface

The author should like to insert the presently distinguishable lithostratigraphic units of the Hungarian Triassic in such tables which not only illustrate their probable interrelations, but also determine their links to the international ammonoid zones. He believes that these tables show the now-fixable position of the units within the chronostratigraphic scale in a more up-to-date form compared to the earlier tentatives (K. Balogh—S. Végh—E. Végh-Neubrandt, 1963; K. Balogh, 1973, 1974; K. Balogh in "*Hongrie*", 1978).

Upon initiatives by Tozer, E. T.; H. Zapfe; L. Krystyn; R. Assereto, H. Kozur; J. D. Zakharov; Yu. V. Arkhipov and others, supported by international cooperation under IGCP Project No. 4, the viewpoints concerning the biostratigraphic and chronostratigraphic (or geochronologic) subdivisions of the Tethyan Triassic have considerably changed during the last decade. Several conventional chronostratigraphic units from the lowermost Triassic up to the middle of the Anisian have been proposed to be replaced by new ones. The Carnian Stage appears to be widened — at the expense of the Ladinian — by the Cordevolian Substage. Since 1973 it has seemed that the Rhaetian Stage cannot avoid its elimination from the chronostratigraphic (geochronologic) scale, unless degraded to some extent, i.e. united with the Sevatian Substage of the Norian Stage. Of the ammonoid zones, those — suitable for a long-distance, interregional correlation — have come to the fore. With a rational application of priority rules, stratigraphers endeavour to rely on the ammonoid zonation in redefining the boundaries of stages and substages: efforts presupposing an up-to-date approach to systematics. Beside adjustment to the new chronostratigraphic frame now being developed, the latest results of investigations of the Hungarian Triassic have also stimulated the writer to modify the earlier tables and to refine them in further details.

In the course of this work he paid particular attention to some parts of the Transdanubian Central Mountains. Namely, the areas along the shoreline of Lake Balaton, with their Triassic sequences containing the richest ammonoid fauna in Hungary, have since J. Böckh (1872, 1873); D. Laczkó (1909, 1911) and L. Lóczy Sr. (1913, 1916) served as an important basis of reference both in this country and abroad. According to I. Szabó (*in* F. Szentes et al. 1972; I. Szabó 1978), however, the development of the Lower and Middle Triassic does not only in the surroundings of Lake Balaton fluctuate more largely than was supposed by Lóczy, but farther northeast (Iszkahegy, Vértes and Buda Mts.), the reef-lagoonal facies of the Anisian Stage continue, in contrast with the case of the Balaton Highland and the Bakony, even in the Ladinian. In the horsts on the Danube's left side, however, even

cherty basin facies occur in immediate vicinity of Upper Triassic platform facies. In contrast with these well-motivated improvements of detail, the author has had to refrain from representing the Triassic of the Rudabánya Mts., because this area is built up by several tectonic units showing such a diversity of facies that their stratigraphy cannot be unambiguously established even today.

While compiling the tables the writer has had to take into consideration the following circumstances:

1. The process of transformation of the international chronostratigraphic scale has not yet been completed: the controversies concerning the beginning and the end of the Triassic and the stratigraphic range of almost all of its stages are still going on. *One of the causes of this fact consists in the limitations of the direct applicability of the ammonoid-based orthostratigraphic system.* Notably, the majority of the ammonoids are frequent only in some deep-water carbonate facies, but even in these it is not always possible to trace them layer by layer. In hosts of other carbonate facies (e.g. Hauptdolomit, Dachstein Limestone, Gutenstein Limestone), in turn, the ammonoids are scant, in fact, they may be totally absent. For this reason, to recognize the boundaries of ammonoid zones in the practice becomes usually subjective. *Consequently, it is necessary to develop, in addition to the orthostratigraphic system, also parastratigraphic ones most perfectly harmonized with the former.* Most suitable for this purpose are such facies-enduring (and therefore widespread) microfossils, characterized, at the same time, by a rapid rate of evolution, which occur in all (or most) members of the studied sequences and thus enable the establishment of a fine zonal scale with sharp and distinct zone boundaries. Most suited to this end within the Triassic are the conodonts, though good results can be expected from investigating the holothurioids, radiolarians, ostracods and sporomorphs with the same aim as well. *The most important and current task in Triassic stratigraphy consists in a precise correlation of parastratigraphic systems with the standard ammonoid zones, that are to be worked out coupled with eliminating the systematic and nomenclatural confusions within the afore-mentioned (and other) fossil groups.* This would be the only means enabling us to find the common language and to achieve that, at least the majority of the scientists should judge uniformly the spans of the chronostratigraphic units being used. Because of the different rates of evolution of the various fossil groups it is though not easy to pave this road, however, we have no other choice. And until this is not achieved, we must remain tolerant with some inaccuracies committed with good faith.

2. Since the Hungarian Triassic comprises a lot of lithological units identical in facies and age with the Alpine and/or Carpathian counterparts, it is advisable, in accordance with the conventional practice, *to apply to these the Alpine* (or Carpathian) *terms* (Werfen Formation, Gutenstein Formation, Steinalm Formation, Wetterstein Formation, Hallstatt Limestone Formation, Hauptdolomit Formation, Dachstein Limestone Formation, etc.). New, "Hungarian" names are proposed to use only for units different from the Alpine ones, difficult to identify with these.[1] The unit-names inserted in the tables usually agree with those proposed by K. BALOGH (1973, 1974) on the basis of the standpoint of the Triassic Subcommission of the Hungarian Stratigraphic Committee. Relying on the relevant literature, the members of the Subcommission already gave a brief characterization of these units in the 2nd edition of the volume "*Hongrie*" (1978) of the Lexique Stratigraphique International. The fact is, however, that these characterizations do not meet yet the requirements stipulated in the Guide issued by the International Subcommission on Stratigraphic Classification in 1972 and its Hungarian version (J. FÜLÖP—G. CSÁSZÁR—J. HAAS—E. EDELÉNYI, 1975). To eliminate these shortcomings will cost us still a very hard work to do, for up-to-date descriptions unambiguously fixing even the unit-boundaries are available only in the monographs devoted to the Triassic of the Bükk, the Mecsek and mainly the Villány Mts. and also to the Kössen Beds of the southern Bakony (K. BALOGH, 1964; E. NAGY, 1968; E. NAGY—I. NAGY, 1976; S. VÉGH, 1964b), though some statements even in these need to be revised. Particularly much to be done has still remained for us as far as the definitions of the lithostratigraphic units of the Aggtelek Karst and, especially, those of the Transdanubian Central Mountains are concerned. Field observations, drilling research and analytical works which have recently undergone an upswing here have already yielded hosts of new results which considerably modify the content of the local stratigraphic units outlined 3 to 6 decades ago, and *sometimes require the*

[1] The names of our units that are still to be clarified and that can by no means be regarded as official terms are given, for distinction, in quotation-marks.

introduction of new lithostratigraphic units (e.g. Nádaska Limestone Formation, Szőlősardó Marl Formation, Csővár Limestone Formation, etc.). *In other cases, units* already correctly outlined and defined, but *unproperly named in terms of present-day judgements, have had to be given proper* (geographic) *names.* This is how the "annulata-bearing" (earlier: "diploporal") dolomite of the Buda Mts. has become — by virtue of the principle of priority — the Budaörs Dolomite Formation, the "Tirolites marl" of the Balaton Highland—Bakony—Iszkahegy zone the Csopak Marl Formation (which is all the more acceptable, as it enables to redetermine the content of this unit, complemented with the members, hitherto unknown, observed by I. Szabó *in* F. Szentes et al., 1972), and the "Tridentinus limestone" thus has been renamed the Nemesvámos Limestone Formation. *Again in other cases, we have changed names adopted also in the 2nd edition of "Hongrie" (1978) for ones more suited to the state of exposure or the principle of priority;* this is why we are recommending the term Feketehegy Formation as a substitute for the "Cserepesvölgy Limestone" of the Pilis and the Sashegy Dolomite Formation for the "Mátyáshegy Dolomite" of the Buda Mts. New fossil finds by H. Kozur and R. Mock (1977) have enabled us to combine the "Nagyeged Limestone" of the Bükk with the Felsőtárkány Limestone Formation. The present insufficiency of evidence for identification justifies to keep the "Ederics Limestone" of the Keszthely Mts. separate from the "Sándorhegy Limestone" of the Balaton Highland and the Bakony (may be just provisionally). *For sake of simplification* we are inclined to unite the "Subalyuk (Hórvölgy)" and "Kisfennsík Limestone" formations of the Bükk with the Bervavölgy Limestone Formation owing to their very slight differences from one another. On the other hand, given the particular importance of the calcareous algal white limestones, distinguished, in the wake of L. Lóczy Sr. (1913, 1916), also by I. Szabó (1978) and intertonguing with the "Megyehegy Dolomite" or the "Reifling and Recoaro Limestone" of the Balaton Highland and the southern Bakony, it is necessary, however, to name it, for the moment informally, "Tagyon limestone". *Similarly important* is to keep within the Hámor Dolomite Formation of the Bükk the lower member under separate name, the Vadászvölgy Dolomite Member, as a distinction from the Nyavalyástető Limestone Member forming the upper part. Nevertheless, we seek, as a rule, to reduce the number of distinguished units and to reserve the right and possibility of further detailing for future contributors.

3. Although not wishing to find final solutions in the tables, *we consider it advisable to indicate the rank we propose to reserve for each unit* adopted within the lithostratigraphic hierarchy. In spite of the lack of exact definition of lithostratigraphic categories in the 1972 Guide of ISSC we believe the difficulties to be avoidable by a far-looking examination of the need for any possible further division. To treat by all means as independent formations, say, relatively little different parts of a continuous sequence being in many cases of restricted extension or still very little studied would be, however, unjustified, according to our opinion. Accordingly, we consider the six lithostratigraphic units hitherto distinguished within the Lower Triassic in the Balaton Highland, the southern Bakony Mts. and the Iszkahegy just members of the Werfen Formation. Within the Werfen Formation of the northern Bakony, the Vértes and the Aggtelek Mts., however, we speak already only of "Seisian" and "Campilian" beds, as, to subdivide in further detail the relevant sequences — known for the moment only in rough lines for the most part merely by drilling and in a heavy tectonic disintegration — belongs to the urgent tasks of the near future. *In order to emphasize the differences in facies* we do not identify the Ablakoskővölgy Formation corresponding to the Bükk Mts. Lower Triassic with the Werfen Formation, retaining the power to further subdividing it to a later time. We propose, however, to unite the four or, respectively, two lithostratigraphic units (E. Nagy, 1968; E. Nagy—I. Nagy, 1976) hitherto distinguished within the Mecsek and Villány Lower Triassic as members within the frame of the Werfen Formation, because such a separate assignment of these strata of allegedly "Upper Campilian" nature is motivated by their divergencies from the underlying Jakabhegy Sandstone Formation. (That this latter belongs to the Lower Triassic was convincingly proved by Á. Barabás-Stuhl (*in* A. Barabás—Á. Barabás-Stuhl—Gy. Majoros—K. Balogh, 1978). As far as the altogether 9 units of the Mecsek Middle Triassic recognized by E. Nagy (1968) are concerned, we propose to treat them henceforth as parts of the Misina Formation.

4. The lower boundary of Hungary's Triassic sequences being generally distinct lithologically, it cannot everywhere be proved very well biostratigraphically. Some of the striking boundaries are emphasized by erosional unconformity and the presence of basal conglomerates above the detrital sequences of the Permian (Balaton Highland, Mecsek, Villány). In those parts of the Transdanubian Central Mountains, where we have to rely merely on the very small surfaces of core samples in our judgement about the mode of contact between the Upper Permian and the Lower Triassic, particular precaution is required in stating the presence of continuous transition between the two systems even if the sedimentation trends may coincide.

The continuity of sedimentation here is only probable even in places, where the shallow-water to beach-rock sediments of the initial Lower Triassic rest on carbonate—shaly—evaporitic lagoonal formations of the latest Permian (GY. MAJOROS *in* A. BARABÁS et al., 1978). In contrast with K. BALOGH's definitively positive statement concerning the question of continuity of sedimentation from the Bükk's predominantly carbonate-built, though heavily bituminous, marine Upper Permian to the more agitated-water carbonate-and-shaly Lower Triassic (1964), the present-day standpoint is more inclined to admit such an opinion only as a probability owing to the tectonic deformation of the boundary formations. The contact of the purple to red "Seisian" sandstone of the Aggtelek Mts. with the anhydrite-bearing Upper (?) Permian underneath (exposed best by the borehole Szín-1 put down quite recently) does not seem to be undisturbed either.

5. The Hungarian Triassic sequences vary in time range and upper boundary from one mountain to the other. The Dachstein Limestone Formation of the Balaton Highland and the southern and northern Bakony Mts. and the Karolinavölgy Sandstone Formation in the Mecsek pass without any break to the Jurassic. In these places the upper boundary of the Triassic could be defined only artificially. For instance, the Liassic continuation of the Dachstein Limestone Formation, the Kardosrét Member, is distinguished from its footwall, beside the disappearance of the large-sized *Megalodontaceae*, mainly only by its being variegated-mottled (J. NOSZKY Jr., 1961). In the Mecsek the Liassic Hettangian Stage was counted for many years from the appearance of the first coal seam, called Seam Alpha, and it was quite recently that, after sporomorph studies by J. BÓNA, this boundary was proposed, by Z. NÉMEDI-VARGA (1971), to be placed by about 85 m higher.

In many other places there is a break in sedimentation between the Triassic and the Jurassic. Consequently, the differently-thick higher part of the Triassic is also absent and, as a rule, the lowermost horizons of the Liassic are not present either. However, this break is not accentuated by angular unconformity even there, where, like in the Villány Mts., it has resulted in the absence of the Upper Triassic as a whole. In the Gerecse, the Vértes, the Pilis, the horsts on the Danube's left side and the Aggtelek Mts., however, only the uppermost levels of the Triassic are lacking, in varying measure, and the hiatus under various horizons of the Liassic is minimal. In the Keszthely Mts., the Iszkahegy and the Buda Mts., in turn, the whole Jurassic—Lower Cretaceous, together with the very top of the Triassic, is absent. In such cases it is difficult to find out how much of this hiatus may be ascribed to a break in sedimentation within the Mesozoic and how much to the Tertiary denudation. The role of the latter is, wantonly and incorrectly, completely ignored by many authors even at present time.

6. The chronostratigraphic correlation of our lithostratigraphic units is rendered difficult by the fact that *most of the Hungarian Triassic faunae derive from samplings that were not made layer by layer.* In addition to this, the processing of new collections and the revision of old ones have been considerably delayed.

An additional impedance consists in the poverty in ammonoids of much of the Hungarian Triassic (E. VÉGH-NEUBRANDT, 1974). For this reason we have to rely increasingly on the use of parastratigraphic methods. On the other hand, the development of conodont stratigraphy, capable of filling the gap due to the shortage of ammonoids, however, is still far from being completed. In addition to this, an even distribution of diggings and boreholes and the analysis of the material for microfacies are the means from which we can expect further progress in stratigraphy.

Consequently, the new tables are far from being considered "finalized". They rather represent a kind of "en route" account, coupled with a number of recommendations and proposals, which has as its principal aim to point out — through a summarization of present-day knowledge — the major problems to be solved in Hungary's Triassic stratigraphy. To achieve this, we give an overview of the key problems first of the chronostratigraphic frame and then of the Hungarian correlation, *looking forward to hearing the opinions of colleagues* especially on the latter.

The tables and the explanations to them have been compiled by considerable collective efforts. Therefore the author is profiting of the opportunity to

express his gratitude to all members of the Triassic Subcommission of the Hungarian Stratigraphic Committee. Particular thanks are due to chief-geologist I. SZABÓ; E. VÉGH-NEUBRANDT, full professor; J. ORAVECZ, principal assistant; F. GÓCZÁN, section head in the Hungarian Geological Institute; Á. BARABÁS-STUHL, palaeontologist on the staff of MÉV and GY. RAINCSÁK and S. KOVÁCS, research associates of MÁFI, for their advice and their disinterested supporting the author with valuable information they have given to him.

Problems of the Chronostratigraphic Frame

Within the three series of the Triassic system in Hungary, developed in Tethyan facies, 7 stages and 15 substages may be proposed to be distinguished.

The LOWER TRIASSIC (Scythian) Series could so far be subdivided in greatest detail, in terms of ammonoids, in Canada (Table 1, 1), though Soviet authors have proposed hosts of other stage names now in circulation (Table 1, 2—5). In general, these can be well correlated with one another and with TOZER's stages. H. KOZUR (1973a, b, 1975) attempted to get several old terms (Gangetian, Brahmanian, Gandarian and Yakutian) revived in a sense somewhat different from their original definition by WAAGEN and DIENER (1895) and to associate them with the Olenekian proposed by KIPARISOVA and POPOV; and he recommended to replace TOZER's Upper Griesbachian by the term Ellesmerian (Table 1, 6). By introducing the name Nammalian, J. GUEX (1977, 1978) and E. T. TOZER (1978a) have simplified the hitherto four-member division to a three-member one (Table 1, 7).

By virtue of their fossil content, the Hungarian occurrences of the Lower Triassic, however, might let be shared by only two stages whose lithostratigraphic equivalents are usually referred to as "Seisian" and "Campilian beds" according to conventions, now seemingly out of date, adopted for the Alpine realm. To stress the time value attributed to these, occasionally, we made attempts at promoting our "Seisian" and "Campilian beds" to the rank of stages. That this approach is incorrect is evident from the fact that, recently, beside the Seis (= Siusi) and Campil Members, additional 4 to 6 members are also distinguished within the Lower Triassic of northern Italy (A. BOSELLINI, 1968; D. ROSSI, 1969; E. FARABEGOLI et al., 1977). For this reason, the author considers, for the moment, the two-member division recommended by Soviet authors to apply best to Hungarian conditions, reserving the right of eventual modification for future contributors.

Namely, out of the minimally 9 ammonite zones, spanning the Lower Triassic in the more eastern parts of Tethys, it is only the *Tirolites carniolicus* Zone proposed by L. KRYSTYN (1974b) and recognized by J. BYSTRICKÝ et al. (1978) also in the Western Carpathians that

Table 1

Stages and substages of the Lower Triassic

1			2	3	4	5	6			7		
Griesbachian		Spathian	Olenekian	Olenekian	Olenekian	Russian	Olenekian			Spathien = Russien		Scythien
		Smithian			Verchojanian	Ussurian	Jakutian			Nammalien	Smithien (Verchojanien, Ussurien)	
		Dienerian	Induan				Brahmanian	Gandarian			Dienerien = Gandarien	
	Upper			Induan	Induan	Induan		Ellesmerian		Griesbachien	Ellesmerien	
	Lower						Dzhulfian p. p.	Gangetian			Gangetien	

1. Tozer (1965, 1967); Silberling—Tozer (1968)
2. Kiparisova—Popov (1956, 1961)
3. Kiparisova—Popov (1964); Arkhipov (1974)
4. Vavilov—Lozovskij (1970)
5. Zakharov (1973, 1974)
6. Kozur (1973, 1975)
7. Guex (1977, 1978); Tozer (1978a, b)

could be identified in Hungary (of course, in addition to the *Tirolites cassianus* Zone surely present). The time spans of the other ammonite zones can, at present, be assessed still only on the basis of the presence or absence of the bivalve *(Unionites, Claraia, Gervilleia and Costatoria)* and gastropod faunae *(Naticella, Turbo)* characteristic of the Werfenian in Hungary. Namely, micropalaeontological studies that might enable a more detailed stratigraphy have so far been conducted mainly by palynological methods and solely in the Villány and Mecsek Mts. For this reason, the suggestion to lift, by virtue of examples picked out of the Germanic facies realm, our *Triadispora crassa-* and *Costatoria costata*-beds to the Anisian Stage would be, according to the author's opinion, premature to be accepted.

Properly speaking we are for the moment unable to contribute much to the question of the Permian-Triassic boundary itself. At present, this boundary is not evidenced palaeontologically even in those regions of Hungary, where a marine transition, seemingly continuous, from limestone to limestone is suspected (Bükk Mts.). Notwithstanding this, the boundary must be got figured within the time scale. If we accept the *Otoceras* Zone as initial member of the Lower Triassic — in support of E. T. TOZER (1967, 1974, 1978) and WANG YI-KANG and HE GUO-XIONG (1976, p. 233) and in spite of objections taken by H. KOZUR (1974, 1975) —, we back this way a convention now appearing to have been enrooted. The key to finding final solution to the problem, however, is not in Hungary.

In Hungary equivalents corresponding to the 3—4 ammonite zones of the Aegean and Bithynian (combined: Anatolian) Substages proposed and defined by R. ASSERETO as substitutes for the "Hydaspian" substage got invalide, forming the deeper parts of the Middle Triassic ANISIAN STAGE are fossil-poor dolomite and limestone complexes. Although these are lithologically rather different from the youngest beds of the Lower Triassic, the palaeontological definition of their age boundaries is not accurate.

This is reflected in L. KRYSTYN's proposal (1974b, p. 45) that the time equivalent of the *Keyserlingites subrobustus* Zone, overlying the North American counterparts of the *Tirolites carniolicus* Zone (the "*Subcolumbites Beds*") and thus still present there, but, in the Alps and Carpathians, not represented by ammonoids, should be sought in the lowermost, fossil-poor part of the limestone and dolomite complexes always considered to be of Anisian age. This opinion concerning the age of the Gutenstein Formation in the Aggtelek Mts. seems to be corroborated by the results of the borehole Szőlősardó-1 (K. BALOGH—S. KOVÁCS, 1981). The opinion that the *Keyserlingites subrobustus* Zone might belong already to the Aegean Substage (H. KOZUR, 1973d, pp. 9—10) cannot be considered proved even on the basis of the profile at Chios. The boundary between the Lower and Middle Triassic can be drawn, even on the basis of conodonts, between the *Subrobustus* and *Anodosum* Zones.

Much better possibilities exist for the separation of the Pelsonian and Illyrian Substages (respectively, of the *Balatonites balatonicus* and *Paraceratites trinodosus* Zones corresponding to them), in Hungary, though the absence of index ammonoids, stated in many places, cannot be perfectly compensated either by dasycladaceans or brachiopods. Namely, the time-range of most of the Anisian dasycladaceans may span the stage as a whole, and a considerable part of the brachiopods, indeed, pass well into the Ladinian.

That the author has substituted the *Balatonites balatonicus* Zone, as a biostratigraphic equivalent of the Pelsonian Substage, for the "*Paraceratites binodosus Zone*" reintroduced by R. ASSERETO (1971) (and originally proposed by MOJSISOVICS) is a reflection of the aptness of H. KOZUR's remarks on this matter (1974, pp. 170—171). Namely, *Paraceratites binodosus* is known only from the upper one of the three ammonite horizons of the zone named after it, and only from the top of it; consequently, the s. str. *Binodosus* Horizon can — as a subzone — be united with the *Paraceratites trinodosus* Zone following next above it; the more so, as this will create, compared to the earlier circumstance, a more favourable situation from the viewpoint of tracing the Pelsonian-Illyrian boundary microfaunistically. To use the *Balatonites shoshonensis* Zone (E. T. TOZER, 1974) instead of the "*Binodosus* Zone" is, in Hungary, not

justified, for this species has been known so far from the western USA only. The *Balatonicus* Zone, however, may certainly be regarded as an equivalent of the great part of the three ammonite horizons of ASSERETO's "*Binodosus* Zone" (H. KOZUR, 1974, Table 2). By the way, even E. T. TOZER (1978) has accepted the *Balatonicus* Zone as the biozone of the Pelsonian Substage in the Tethys realm.

Above the *Trinodosus* Zone — usually at the top of the Illyrian (e.g. H. ZAPFE, 1974, p. 246) in KOZUR, 1974, p. 172, however, at the base of the Fassanian Substage — the *Aplococeras avisianum* Zone, got revived by R. ASSERETO (1969), is usually indicated, too. Since *Aplococeras avisianum* does not occur in Hungary and the "associate" *Ptychites* and *Flexoptychites* fauna is not diagnostic of age (I. SZABÓ, 1978; I. SZABÓ—GY. MAJOROS, 1978), the author cannot help omitting the *Avisianum* Zone from its tables. Let us note in this connection that TOZER (1978a) considers the location of the *Avisianum* Zone at the base of the Fassanian Substage to be an undesired introduction of a convention deviating from the traditions. Involved in this may have been the consideration of that obvious error that E. MOJSISOVICS (1895) placed the *Avisianum* Zone still between the *Curionii* and *Archelaus* Zones.

Because of the doubtfullness of the lower and upper boundaries the splitting of the Middle Triassic LADINIAN STAGE into "Reitzi, Wengen and Cassian horizons" has recently got in a crisis. Namely, some authors assign the lower horizon of the Fassanian Substage to the Anisian, others place the old "Cassian Horizon" (= Cordevolian Substage), partly or completely, in the Carnian.

To place the Lower Fassanian *Protrachyceras reitzi* Zone in the Upper Anisian would be motivated best of all by the fact that, associated with "*Ceratites* of the Reitzi group" and with *Kellnerites*, one member of the *Nevadites* genus, *N. humboldtensis*, enabling a finer subdivision of the North American *Gymnotoceras occidentalis* Zone, believed to belong still to the Upper Anisian (and thus being very frequent there!), has recently been discovered even in Greece. Moreover, according to E. T. TOZER (1978), even the *Protrachyceras* sampled from Beds 98 to 112 of Switzerland's Grenzbitumenzone would be a *Nevadites*. L. KRYSTYN—I. MARIOLAKOS (1975, p. 186) and E. T. TOZER (1978) refer — in the wake of N. J. SILBERLING and E. T. TOZER (1968) —, in addition to the above, also to the fact that the lower boundary of the Ladinian Stage was defined already by E. MOJSISOVICS—W. WAGEN—C. DIENER (1895, p. 1279) with the *Protrachyceras curionii* Zone overlying the *Reitzi* Zone. The presence in the Tethys of *Nevadites* (forms absent in Hungary or not yet discovered) is really an argument in favour of the correlability of the *Reitzi* Zone with the *Gymnotoceras occidentalis* Zone.

The possible belonging of the *Reitzi* Zone to the Ladinian, however, would be precluded by such a situation which would arise, when the "*Ceratites*" *reitzi* from Hungary would turn out not to belong to the *Protrachyceras* genus. Thus, for the time being, nothing can hinder us in sticking to the convention of many decades defining the base of the Ladinian Stage with the *Reitzi* Zone — an approach more applicable to the conditions existing in the Balaton Highland.

As regards the Fassanian and Longobardian age of the *Curionii* and *Archelaus* Zones, respectively, no doubt has been awaked even in the recent

times. It is worth mentioning that some representatives of the North American counterparts of the *Archelaus* Zone, (namely ones of the *Meginoceras meginae* and *Maclearniceras maclearni* Zones, respectively) have been identified even in the Alps (E. T. TOZER, 1978b). The upper boundary of the *Archelaus* Zone, however, is better indicated by the last occurrence of *Daonella lommeli* than by the zonal index species which, as shown by some authors (see below), seems to reach up into the CARNIAN STAGE. The controversy about defining the base of the Carnian Stage has, notwithstanding this, stemmed not from this circumstance, but from the false interpretations taken by MOJSISOVICS.

Namely, in course of the gradually developing the stratigraphy of the Carnian Stage, MOJSISOVICS (Table 2) splitted up its lower part, in 1879/1882, into *Trachyceras aon* and *Trachyceras aonoides* Zones. In 1892/1893, however, he transferred the *Aon* Zone to the Lower Carnian, the *Aonoides* Zone, in turn (together with the *Trachyceras austriacum*-bearing "lens") to the Middle Carnian, noting that he had based the first zone on the *Aon* fauna of the Cassian Beds, the second on the *Aonoides* fauna of the Feuerkogel's Hallstatt Limestone. In 1895, however, when instituting the Cordevolian Substage corresponding to the Lower Carnian and the Julian Substage corresponding to the Middle Carnian, he designated the stratotype of the former to be the Cassian Beds, and that of the latter to be the Raibl Beds (overlying the former). This is surprising, because the Raibl section comprises ammonoids only at its base and only ones belonging to the *Aon* Zone, so that thus cannot be decided whether its higher part may extend upwards into the *Aonoides* Zone as well. After this final standpoint of MOJSISOVICS thus such a funny situation has arisen that the Julian Substage — whose key horizon, the *Aonoides* Zone, has a priority over the Cordevolian key horizon (i.e. the *Aon* Zone) — may now be regarded as a *later* synonym of the Cordevolian Substage owing to the *subsequent* designation of the Raibl type locality.

It is quite possible that the condensed and block-faulted nature of the Feuerkogel section, a fact confirmed by L. KRYSTYN (1978, p. 44) as well, was the main reason for MOJSISOVICS' proposing to select the Raibl section as type locality for the Julian Substage in the hope that later even fossils of the *Aonoides* Zone might be found in its higher part. This, hope, however, has not come true; but the correctness of the *Aon—Aonoides—Austriacum* succession has been approved in a number of other Alpine sections, moreover, partly even in the Balaton Highland. Consequently, the type locality for the Julian Stage would be designated in accordance with MOJSISOVICS' earlier views. We should accept, however, also L. KRYSTYN's statement (1978) that the *Aon* and *Aonoides* Zones cannot be assigned to two separate substages, as their index species are members of one and the same evolutionary line; consequently,

Table 2

Development of MOJSISOVICS' *views on the division of the deeper part of the Carnian Stage*

1869	1879/1882	1892/1893		1895	
				Tuvalian	
Aonoides Zone *C.*		Austriacum Z. *M.C.* Aonoides Zone	Feuerkogel	Julian	Raibl
	Aonoides Z. *L.C.* Aon Zone	*L.C.* Aon Zone	St. Cassian	Cordevolian	St. Cassian

C. = Carnian *L.C.* = Lower Carnian *M.C.* = Middle Carnian

such a separation is conceivable only between the *Aonoides* and *Austriacum* Zones. To place the *Aon* Zone in the Lower Carnian would mean to return to MOJSISOVICS' 1879/1882 statement and — with conservation of the term Julian — to discard the Cordevolian name.

Nevertheless, the solution is not so simple. It is true, that the deeper part of the Julian Stage constituted in accordance with the above is overlapped by the *Upper Part* of the Cassian Beds, the stratotype of the Cordevolian Substage, and this upper part comprises the *Aon* Zone as a whole and the base of the *Aonoides* Zone. The *Lower Part* of the Cassian Beds, however, on account of its containing *Frankites regoledanus* and *F. apertus* (URLICHS, M. 1974, 1977), belongs to the *Sutherlandi* Zone known, beside North America, from the Feuerkogel, from Romania, Greece and Turkey as well. H. KOZUR (1976), however, noted by right that the *Sutherlandi* Zone should be assigned — in contrast with the opinion heretofore adopted — not to the top of the Ladinian, but to the base of the Carnian, as an abrupte change of both the macro- and microfauna could be observed — as pointed out also by L. KRYSTYN and B. GRUBER (1976, p. 281—282) — only at the base of this zone. The *Sutherlandi* Zone is connected, in turn, with the *Aon* Zone by several common elements.

Therefore — although the index species of the Longobardian Substage, *Protrachyceras archelaus*, occurs, in the Tethys realm in the *Sutherlandi* Zone, moreover, according to M. URLICHS (1977) even in the Raibl Beds of the Schlern plateau, it would not be right to consider the Lower Cassian Beds to be of Longobardian age, but they must be placed, together with the Upper Cassian Beds, in the lower part of the Carnian Stage. This enables, at the same time, to conserve the validity of the Cordevolian Substage in the sense of MOJSISOVICS. Since, from the viewpoint of the stratigraphic subdivision of the platform facies common in the eastern Transdanubian Central Mountains and in North Hungary, this solution seems, on account of the enrooted use of the term, to be, for the moment, more favourable, the Carnian Stage on the time scale of our tables begins with the three-zoned Cordevolian and only the *Austriacum* Zone and the "*Sirenites*" Zone detachable from the former up to the appearance of *Tropites dilleri* remain to be included in the Julian Substage.

On the zonal and substage scale of the interval spanning the Upper Carnian to the NORIAN the majority of the authors, apart from very small differences, appear to agree. The Alpine zones of this interval are generally well correlable with the North American ones, and even the so-called "*Anatropites* domain" separated by H. ZAPFE (1974) and L. KRYSTYN (1974) is distinguished only by the absence of the index species from the North American *Klamathites macrolobatus* Zone (H. KOZUR, 1973b; E. T. TOZER, 1978a).

All the more disputed is the zonation of the youngest part of the Triassic. First of all, there is a controversy about the position of the "*Halorites* Horizon" placed at the top of the Alaunian Substage by H. ZAPFE (1974) and L. KRY-

STYN (1973, 1974). Some Alpine genera — restricted to the *Himavatites columbianus* Zone established in North America — were earlier taken to be characteristic forms of the *Cyrtopleurites bicrenatus* Zone. On the other hand, it has been verified that the "*Sirenites*" *argonautae* Zone distinguished by E. MOJSISOVICS (1893) within the Pötschen Limestone belongs to deeper parts of the *Columbianus* Zone (F. TATZREITER, 1978). The *Argonautae* Zone, however, was assigned, by MOJSISOVICS, still to the Sevatian, a substage created by him, and it was only upon initiatives of North American palaeontologists that it was placed at the top of the Alaunian. Therefore, according to H. KOZUR (1973, 1979), the *Columbianus* Zone should be regarded already as Sevatian. According to F. TATZREITER (1978), the genus *Himavatites* fills only in the boreal province the entire *Columbianus* Zone, being present only in its deeper horizons in the eastern Tethys and is known only in obscure traces in the Alpine profiles (e.g. it occurs even at the type locality of *Argosirenites argonautae*). The fact is that MOJSISOVICS' second *(Pinacoceras metternichi)* Zone of the Sevatian Substage has had to be rejected and E. T. TOZER (1967, 1971) has replaced it by the *Rhabdoceras suessi* Zone of global distribution, the lower part of which he named, in 1979, the *Gnomohalorites cordilleranus* Zone. Presence of *Monotis* species characteristic of the *Cordilleranus* Zone enables the recognition of the latter even in absence of ammonoids. According to H. KOZUR (1973, 1979), this zone in the Alps is correlable with the *Sagenites giebeli* Zone, known, unfortunately enough, from fissure infillings only (and this correlation can be extended, at least in part, also to a portion of the topmost, *Halorites* Subzone of the *Columbianus* Zone, though this subzone still lacks *Rhabdoceras*, but includes *Brouwerites*).

No matter how important its role in defining the base of the Sevatian, *Rhabdoceras suessi* is not suitable for a detailed zonation owing to its reaching up into higher horizons of the Alpine Kössen Beds. For this reason, the use of the "*Suessi* Zone" was given up, in 1979, even by its founder, TOZER, who proposed to use, instead of its lower part, the *Cordilleranus* Zone referred to above and, instead of its upper part, the *Cochloceras amoenum* Zone. Since this latter is correlable with the *Cochloceras suessi* Zone of the Alps, the problematics of the upper part of the Norian is reduced to decide whether the RHAETIAN can be retained as an independent stage or substage, either in its foregoing extent or in a volume including the quondam *Rhabdoceras suessi* Zone, too (extended up to the base of the *Sagenites giebeli* Zone), or it is to be regarded simply as the uppermost zone of the Upper Norian (= Sevatian). The present writer believes that, owing to the obvious priority and the enrootedness of the interpretation adopted heretofore, the efforts enabling to conserve the Rhaetian Stage in the conventional sense (i.e. encompassing the Alpine Kössen Beds and their biostratigraphically correlable equivalents of Zlambach and Dachstein facies) should be given preference. He is afraid that

any new interpretations deviating from the above would render extremely difficult the use of information of earlier literature. For this reason, the author joins in H. KOZUR's opinion, expounded several times between 1973 and 1979, that the Rhaetian can be saved as an independent stage, when split up into two ammonoid zones. The possibility for this exists, even if its lower — *Choristoceras haueri* — Zone is indicated (like in the Weissloferbach's section: H. MOSTLER et al., 1978), *only* by the presence of *Rhaetavicula contorta* and *Rhabdoceras suessi*. The danger that *Choristoceras marshi* may eventually appear already in the *Haueri* Zone (H. KOZUR, 1979, p. 76, footnote), can be easily eliminated, if the *Haueri* Zone is considered as an assemblage zone and if its upper boundary is defined as coinciding with the disappearance of *Rhabdoceras suessi*. Accordingly, only that, rather thin, part of the Alpine Kössen Beds which already contains only *Choristoceras marshi* would belong to the *Marshi* Zone.

Correlation of Lithostratigraphic Units

A) THE CENTRAL MOUNTAINS FACIES AREA

1. The Keszthely Mts. (Table 3). The oldest known member of the geologic column of the Keszthely Mts. belongs to the higher-situated marly beds of the *Veszprém Marl Formation* extending from the Balaton Highland and the Bakony into the study area. These beds, being known from outcrops, reduced in extent, but rich in fossils (L. LÓCZY Sr., 1913, 1916; P. BOHN, 1975, 1979), have recently been uncovered, in a drilling thickness of about 100 m, without its footwall being reached. By virtue of foraminifera determinations by A. ORAVECZ-SCHEFFER *(Frondicularia bryzaeformis, Schmidtia inflata, Oberhauserella* cf. *carinthiaca)*, the "Ederics limestone" considered as back-reef formation by P. BOHN (1975) can be correlated — in contrast with G. KOLOSVÁRY's assigning it to the Norian (*in* SZENTES et al., 1972) — with the "Sándorhegy limestone" of the Balaton Highland and, in accordance with L. LÓCZY Sr.'s opinion (1913, 1916), it may be regarded as the top-most member of the Veszprém Marl Formation. However, whether this limestone was originally present between the Hauptdolomit and the marly members of the Veszprém Marl and whether it has been rolled out elsewhere upon tectonic effects is not clear even today.

The *Hauptdolomit Formation* overlying either the "Ederics limestone" or the Veszprém Marl is, as far as its higher part is concerned, certainly of Lacian —Alaunian age; that its deeper part reaches down to the Upper Tuvalian, however, can be only supposed by relying on Balaton Highland and South Bakony analogies.

On the basis of *Rhaetavicula contorta* the presence and Rhaetian age of the sediments rich in cherty dolomites and called conventionally the "*Kössen beds*" can be taken for sure (the afore-mentioned species seems to occur more frequently here than in the Bakony: L. Lóczy Sr., 1913, 1916; F. Szentes, 1943; S. Végh, 1964b; F. Szentes et al., 1972; P. Bohn, 1975, 1979). "Kössen beds" represented by hundreds of metres of sediment of rather Carpathian facies, consisting of alternating clay-marls to clays, limestones and dolomites, have been uncovered by drillings in the westward, deep-subsided extension of the Transdanubian Central Mountains. Farther northeast, however, they gradually pinch out, so that no typical Kössen facies is known from the Vértes —Gerecse range onwards. (Only in the Pilis' Feketehegy Formation they are supposed to reappear.) The "Kössen Beds" of the Balaton Highland and the Bakony always occur between the Hauptdolomit and the Dachstein Limestone Formation. In the Keszthely Mts., however, on account of the post-genetic denudation of their hanging wall, just their resting on the Hauptdolomit can be stated.

2. The Balaton Highland and the southern Bakony Mts. (Table 3). The succession and age of the units distinguished here and spanning the whole Triassic system have been indicated, essentially, in the sense of L. Lóczy Sr. (1913, 1916). But wherever it was possible or necessary, the author has replaced Lóczy's terms by geographic names. In the light of new discoveries made in recent time, their stratigraphic content must, of course, be redefined.

The initial beds of the Triassic rest, in the southwest, unconformably, in the northeast, peneconformably on the final beds of the Balaton Redsandstone Formation, dated by sporomorphs convincingly as of Upper Permian age (Á. Stuhl, 1961), and interbedded, locally, with greenish-grey sandstone and dolomite banks as well as coaly and anhydritic stringers. The biostratigraphic definition of the boundary between the two systems is, nevertheless, pregnant with some uncertainty, because the last Permian sporomorphs are found about 100 m below the lithologic and settling boundary; thus the last 100 m of the Permian are devoid of sporomorphs.

Of the six Lower Triassic members to be unified as Werfen Formation the lower two, the Nádaskút Dolomite and the Arács Marl, agree in their features with Lóczy's characterization over most of the area in question. The composition of the *Nádaskút Dolomite Member*, however, is variegated, in the borehole Szigliget-1, by gypsum-bearing and anhydritic and, in some places, heavily glauconitic dolomite layers. In other boreholes put down in the southwest part of the Balaton Highland the composition of the Arács Marl Member is modified as its base is constituted — in contrast with the greenish-grey colour of the bulk — by red marls and siltstones, its top, in turn, by thicker, locally anhydritic and dolomitic limestone layers. Consequently, in the southwest part of the Balaton Highland the initial members of the Triassic

are more strongly sulphatic and dolomitic than elsewhere (F. SZENTES et al., 1972).

The *Hidegkút Sandstone Member* — when normally developed — is dominated by alternation of red, shaly and laminated sandstones and siltstones, in which, from about the middle onwards, rust-mottled, grey, often gastropod-oölitic limestone laminae are intercalated. In other cases, in turn, the limestone laminae are completely missing and the member ends with grey, porous dolomite beds (I. SZABÓ *in* F. SZENTES et al., 1972, p. 34).

Nevertheless, the greatest change compared to LÓCZY's description has been observed within the *Csopak Marl Member*. Namely, above the grey initial member of the "*Tirolites* marl" including also limestone laminae and roughly corresponding to the Csopak Marl Member L. LÓCZY Sr. (1913, 1916) — for lack of suitable exposures — could still separate only the "*Tirolites* marl" proper, comprising also some grey or reddish limestone laminae and, above it, the yellow, crinoidal limestone laminae. According to I. SZABÓ (*in* F. SZENTES et al., 1972, p. 34), however, the thickness of these beds (about 120 m) is much greater than given by LÓCZY (75 m), because between the s. str. "*Tirolites* marl" and the overlying Aszófő Dolomite Member there is first a considerable thickness of red clay-marls and siltstones and then follows the red argillaceous sandstone alternating with crinoidal limestone laminae which, in a weathered condition though, LÓCZY could also observe. According to the friendly information obtained from I. SZABÓ, the appearance of *Dinarites (D. nudus, D. dalmatinus)* within the s. str. "*Tirolites* marl" precedes that of *Tirolites cassianus*. As he has stated, the lower, grey clay-marl sequence is characterized by *Naticella costata* and its relatives and the red marl sequence overlying the *Tirolites* beds is characterized by *Gervilleia* as most frequent fossils.

There is no need for changing LÓCZY's interpretation of the *Aszófő Dolomite Member* and of the *Iszkahegy Limestone Member*. Although I. SZABÓ (*in* F. SZENTES et al., 1972) — partly following J. BÖCKH's opinion (1872, 1873), partly by Alpine analogies — is inclined to attach it already to the older Anisian, there is no convincing fossil evidence that might corroborate such an approach. Notably, the hypersaline Aszófő Dolomite is completely unfossiliferous and the, rather poor, fossils of the Iszkahegy Limestone *(Costatoria costata, Gervilleia modiola)* are still of "Campilian" character. In spite of this fact, there would be some reason for attributing the Iszkahegy Limestone to the Anisian — from purely lithostratigraphic viewpoint —, yet a definitive solution to the problem might be achieved only by conodont studies.

The notion of the *Megyehegy Dolomite Formation* within the MIDDLE TRIASSIC, formulated first by J. BÖCKH (1872) and then by L. LÓCZY Sr. (1913, 1916), still holds firm even at present. As confirmed by new observations, this formation, not particularly rich in fossils, is not restricted to the

Aegean and Bithynian Substages, but locally extends well into the Pelsonian and, in the vicinity of Hajmáskér and Öskü, even into the Illyrian Substage, being overlain here immediately by tuffitic layers with *Trachyceras reitzi* (i.e. Lower Ladinian) (I. Szabó, 1978, Fig. 1).

That the Middle or Upper Anisian limestone bodies, now interbedded with marls, siltstones and tuffites, now cherty or dolomitic, are still referred to respectively as "*Recoaro*" or "*Reifling limestone*" is due to our simply having no other choice, for, on the one hand, their petrologic features do not correspond exactly to those of the Alpine units of the same name; on the other hand, on account of their faciological variability, related in part, to their remarkable changes in thickness, they show a kind of uniformity, so that they cannot by separated by a sharp boundary. At L. Lóczy Sr., this manifested itself in his attempting to refer to them jointly as "Alpine Muschelkalk". This contradictory term, however, is no longer used even in the Alps. Since before a profound elaboration of the outcrops settled by I. Szabó (1978, Fig. 1) and of other recoveries still to be done hereinafter it would not be quite right to coin hastily a new name encompassing the afore-mentioned two "members", an interrogation mark has been entered in its place in the Table 3.

It remains for further studies, however, to assess the stratigraphic value of the reef-lagoonal (Dasycladacea-bearing) "*Tagyon limestone*" replacing the "Recoaro" and the "Reifling limestones" in the vicinity of Vászoly and Szentantalfa and uncovered even toward the depth in the borehole Szentantalfa Szaf-1. Inasmuch as can be judged at present, this will be equivalent to the "Recoaro" and "Reifling limestones" as a whole. However, if the "Tagyon limestone" is totally devoid of ammonoids and the Megyehegy Dolomite Formation contains one specimen of *Balatonites balatonicus* (characteristic of the Pelsonian; J. Böckh, 1872, 1873) only in the topmost layers of its type locality, then the deeper part of the "Alpine Muschelkalk" does contain, beside brachiopods of the *Decurtella decurtata* Zone, the species *Balatonites balatonicus*, while its higher part contains, in addition to some "surviving" brachiopodal species and *Paraceratites trinodosus*, also *Semiornites* sp., *Flexoptychites*, etc. and, above these, even *Daonella sturi* (L. Lóczy Sr., 1913, 1916; I. Szabó *in* I. Szabó—Gy. Majoros, 1978). Thus, in the Anisian of the Lake Balaton region only two ammonoid zones can be identified in the present state of knowledge.

Although tuff and tuffite interlayers do occur even in these fossil-evidenced Middle to Upper Anisian sediments (I. Szabó, 1978; I. Szabó *in* F. Szentes et al., 1972, Fig. 11—12), the main eruption of the volcanic tuffs of potash-trachytic composition was associated, as suggested by I. Szabó—Cs. Ravasz (1970) and Cs. Ravasz (1973), nevertheless with the sedimentation of the "*Buchenstein Formation*" traceable, in varying thickness, in the hanging wall of all the Anisian facies over a large area (as proved by drilling). Dotted with

frequent limestone lenses, these peculiar, green and light purple tuffite beds of several m thickness serve as a sharp lithostratigraphic boundary for the Ladinian. The Buchenstein Formation of the Lake Balaton Region, however, includes, in addition to this (often heavily bentonitized) "*pietra verde*", even thicker tuffaceous limestone and cherty limestone beds and, at the top, also laminated cherty shale and radiolarite layers. The formation consists, however, of tectonically very mobile rocks, therefore to determine its composition and virtual thickness is often faced with considerable difficulty. It is overlain, in every case, by the red or light grey, cherty, nodular limestones of the Nemesvámos Limestone Formation, sometimes similarly with tuffite and marl interbeds.

On the basis of faunae deriving from new diggings and boreholes, I. Szabó (1978) placed the age of the "Buchenstein Formation" convincingly in the *Protrachyceras reitzi* Zone. These faunal assemblages, containing, beside *Protrachyceras reitzi*, the representatives of *Kellnerites*, *Hungarites* and *Flexoptychites*, can be readily identified with one another, in spite of the fact that they are associated with pietra verde of rather dissimilar sequences. The *Trachyceras reitzi*-bearing "pietra verde" of the Forráshegy at Felsőörs, here overlying the limestones of the *Paraceratites trinodosus* Zone, is, for instance, overlain by relatively thin, cherty limestones, followed, in turn, already by typical red layers of the lower part of the Nemesvámos Limestone Formation. The very thin, *Tr. reitzi*-bearing "pietra verde" of the Hajmáskér section, rests, on the contrary, on Megyehegy Dolomite which is followed, in turn, with intercalation of the calcareous-algal reef-lagoonal facies of the Budaörs Dolomite Formation pinching out in this region, by the Nemesvámos Limestone. The similarly thin, but fossiliferous, "pietra verde" of the Vászoly and Tagyon sections, respectively, however, is underlain already by a calcareous-algal "Tagyon limestone", while its immediate overlier is constituted by cherty limestones of "Buchenstein" type, interrupted by newer tuff and marl interlayers and much thicker than its Felsőörs counterpart. In the vicinity of Vászoly and Pécsely, radiolarite beds are added to this and the *Nemesvámos Limestone* lies only above these.

Lóczy's characterization of this formation overlying the "Buchenstein Formation" and referred to since I. Böckh (1872, 1873) as "Tridentinus limestone" has not changed much even in the light of latest results. Among its cherty-nodular limestone beds, red at the base and light grey higher up the profile, light grey tuff or marl interlayers occur in some places, as well. The deeper part of its basin-facies sequence of Hallstatt type represents, according to F. Frech (1911) — by virtue of the presence of *Protrachyceras villanovae* and *Arpadites* —, the *Protrachyceras curionii* Zone, the higher part the *P. archelaus* Zone. *Daonella lommeli* also occurs in the marly intercalations of its higher part.

UPPER TRIASSIC. The Nemesvámos Limestone is followed upwards immediately by the lowermost member of the *Veszprém Marl Formation* — the "*Füred limestone*". The similarity of the neighbouring beds of the two formations had become a source of various misunderstandings, though L. LÓCZY Sr. (1913, 1916) cleared the majority of these. However, it has remained an unanswered question, how to asses that thin formation called the "Sóly marl" which D. LACZKÓ (1909, 1911) wished to use for drawing a boundary between the Nemesvámos Formation and the "Füred limestone". I. SZABÓ (in "*Hongrie*" 1978), however, found the content of different horizons to have got mixed up in the notion of Laczkó's "Sóly marl", a phenomenon due to the poor quality of the exposures; therefore, he proposed to discard the term from among the stratigraphic units.

Disagreeing with E. VADÁSZ (1953, 1960) and I. SZABÓ (*in* F. SZENTES et al., 1972) who placed — obviously on the basis of *Daonella lommeli*, allegedly found J. BÖCKH in the "Füred limestone", but refuted by L. LÓCZY Sr. (1913, 1916) — this limestone still at the top of the Ladinian, the present writer is more inclined to regard it as part of the Veszprém Marl Formation (referred to as "Upper Marl Group" in earlier works). In fact, despite his refutal just quoted, LÓCZY himself did assign this chert-poor, but a little dolomitic, compact, light grey limestone, always unfossiliferous when typically developed, to the topmost Ladinian, because he found realiably Carnian fossils *(Trachyceras aon)* only in the overlying "Chondrites limestone" already undoubtedly belonging to the "Upper Marl Group". On the other hand, LÓCZY himself did emphasize that limestone interlayers very similar to the "Füred limestone" occur in higher horizons of the "Upper Marl Group" as well. A. ORAVECZ-SCHEFFER (1967), in turn, managed to collect *Carnites floridus* from the marl interlayer separating the Nemesvámos ("Tridentinus") Limestone and the "Füred limestone". Thus this species obtained from the footwall of the allegedly typical (unfossiliferous) "Füred limestone" and thus far known only from deeper parts of the "Upper Marl Group" bears a convincing witness to the Lower Carnian age of the "Füred limestone" and to its belonging to the "Upper Marl Group" (J. ORAVECZ, 1963).

The facies of the less typical occurrences of the "Füred limestone" differs only by their containing ostracods and/or tiny, thin-shelled bivalves (F. SZENTES et al., 1972) in the thin clay layers interbedded with the limestones. These beds are certainly fore-runners of the lowermost *Chondrites*- and *Daonella*-bearing limestone beds of the "Upper Marl Group" taken in LÓCZY's sense. Higher up the profile, these limestone beds — containing first *Daonella reticulata*, *D. latecostata*, *D. cassiana*, *D. esinensis*, *D. aperta* and *D. lóczyi* and then *Trachyceras aon* — are increasingly interbedded with shaly clay layers and their sequence grades — with complete continuity — into the so-called "*Estheria* marl" characterized mainly by *Halobia rugosa* and *Carnites floridus* (remark-

ably enough, *Trachyceras aonoides* is totally absent!). Only 1/10 of the total thickness, 150 m or so, of the deeper beds just enumerated is made up of the purple-grey, poorly cherty, thin-bedded "*Trachyceras austriacum* limestone" which contains, in addition, *Koninckina leonhardi* as well. Although *Tr. austriacum*, *Sirenites subbetulinus* and *Halobia rugosa* are found even at the base of the several hundred m thick "*Nucula* and *Lima* marl", following next higher up the profile, the higher horizons of the Triassic in the Balaton Highland, and what is more, throughout the whole Bakony area are already devoid of ammonoids. The "Upper Marl Group" of the Balaton Highland ends with the light grey "Sándorhegy limestone" interbedded with dark marls and/or bituminous limestones and dolomites and containing — beside *Julithyris julica* and *Alectryonia montiscaprilis* — *Cornucardia hornigi* and *Neomegalodon carinthiacus*, as well.

As a new lithostratigraphic name for the "Upper Marl Group" including the "Füred limestone", as well, the term *Veszprém Marl Formation* can be recommended; a name used by D. Laczkó (1909, 1911) in a more restricted sense, but reflecting very well the coherence of the constituent strata groups. This formation has now — as a result of deep drilling activities — become known from Keszthely up to Mány and Zsámbék. It seems, however, that its limestone members distinguished by Lóczy are restricted rather to the southern belt of the Central Mountains. Therefore we should like to handle them for the moment only as informal units until their detailed reinvestigation is completed. Such a new elaboration may then decide also the choice of the new names (based on geographic location of the occurrences) that are suitable to substitution for the Lóczy's terms accentuating the paleontological content.

That the near-Veszprém occurrences of the *Hauptdolomit Formation*, assignable after E. Kutassy (1933, 1940, 1945) for the most part to the Lacian-Alaunian, reach as far down as the Tuvalian was satisfactorily proved by S. Végh (1964a). The fact that the Sevatian age of the "*Kössen beds*" between the Hauptdolomit and Dachstein Limestone Formation has been evidenced by a rich bivalve fauna corroborates also the Rhaetian age of the *Dachstein Limestone* occurrences in this region (S. Végh, 1964b). That the Dachstein limestone facies extends well into the Liassic was earlier taken to be undoubted; but, lately, some authors have coined a new, separate name, the Kardosrét Member which they have begun to use for the so-called "Dachstein Liassic" which, as put by J. Noszky Jr. (1961), is distinguishable only on the basis of its more variegated colour and the scant brachiopod content replacing the representatives of *Megalodontaceae* vanishing here.

3. The northern Bakony Mts. (Table 3). The Triassic sequence of this mountains, though being considerably completed recently, is not known in full even today.

The borehole Alsószalmavár-1 has uncovered a *Werfen Formation*

evolving without any break (?) from the Upper Permian Balaton Redsandstone Formation and reaching roughly to the middle of the Olenekian (Cs. DETRE—E. NAGY, 1971). Represented by sandstones—evaporites—shales—dolomarls at the base and getting ever more calcareous higher up the profile, these beds were defined, by their describers relying partly on lithological analogies, partly on appearence of *Claraia* sp., *Costatoria costata* and *Sporomorpha* being of "Seisian" or "Campilian" character. These terms, however, cannot be attributed the content of formal units.

Neither the uppermost members of the Lower Triassic, nor much of the lower part of the MIDDLE TRIASSIC Anisian Stage could so far be explored. That the grey to dark grey dolomites in the 466.6 to 480.0 m interval of the borehole Bakonygyirót Bszl-5 may yet belong to the top of the Bithynian part of the Megyehegy Dolomite Formation is, again, just a hypothesis based on lithological analogies. Namely, in the drilling documentation, this dolomite figures as Ladinian, just as it is the case with the complex pierced by the drill in 80 m thickness between 387.3 and 466.6 m which consists of alternation of dark grey clay-marls, calcareous marls and dolomarls and contains the first traces of volcanic tuff eruptions. Despite this fact, we guess to have to do in these latter beds already with representatives of the Pelsonian and Illyrian Substages. Notably, with a view to the thick interbeds of potash-trachyte tuffs and tuffites, the 76-m-drilled complex consisting of cherty limestones, cherty dolomites, clay- and dolomarls within the 311.3—387.3 m interval above the afore-mentioned complex seems to correspond, already, to the "*Buchenstein Formation*" of the Balaton Highland (Cs. RAVASZ, 1973). Since, however, the Nemesvámos Limestone in the Balaton Highland also contains some interbedded tuff layers, it is for the moment impossible to decide whether the higher members of the cherty-tuffaceous sequence of the Bakonygyirót borehole may belong, at least partly, to this latter formation already. Because of the existing uncertainties, the author had to refrain from giving the Bakonygyirót sequences formal names.

UPPER TRIASSIC. Sometimes containing some thin limestone interbeddings, the dark grey marls, calcareous marls and carbonaceous siltstones of the *Veszprém Marl Formation* are known from boreholes only. The boreholes Pápateszér-1 (= Fenyőfő-2) and Bakonyszűcs Bűt-10 have intersected about 30 m of this sequence, the borehole Bakonyszűcs Bszű-1, however, could not get out of it even when reaching down to a depth of 1150 m. According to A. ORAVECZ-SCHEFFER (1971), it contains a foraminiferal assemblage of Cassian—Raibl type. Whereas the foraminiferal limestone outcropping at Bakonykoppány (A. ORAVECZ-SCHEFFER, 1967) and the brachiopodal limestone observable at Ugod (Cs. DETRE, 1972) are usually identified with the "Sándorhegy limestone" of the Balaton Highland, to subdivide the formation in a detail similar to the case of the Balaton Highland occurrences is not feasible

here, and, on top of that, it has not so far been possible to clear its relation to its footwall either.

Similarly to the case of the Balaton Highland and southern Bakony occurrences, the deposition of the *Hauptdolomit* began, in some places, as early as the Tuvalian, but the bulk of the formation is of Lacian-Alaunian age. It is connected with the *Dachstein Limestone Formation*, as a rule, by a transition with marked changes in lithology, but this transition is assigned, already, to the Dachstein Limestone.

Known only in the southern part of the northern Bakony Mts., the "*Kössen beds*" have recently been identified here in the borehole Csőszpuszta Tt-28, between the Hauptdolomit and the Dachstein Limestone, being represented by an alternation of pyritiferous clay-marls, calcareous marls, laminated, marly or dolomitic limestones and dolomarls.

The *Dachstein Limestone Formation* of the northern Bakony Mts. was considered in the literature up to E. VADÁSZ (1960) unanimously as of Rhaetian age. E. VÉGH-NEUBRANDT (1963) was the first to point out that its deeper part contains already such Norian *Megalodontaceae* as are known from the higher parts of the underlying Hauptdolomit as well. As judged today, this deeper part of the Dachstein Limestone Formation belongs to the Alaunian Substage. Consequently, the deposition of the Dachstein Limestone facies must have begun over the greater, "Kössen"-less part of the northern Bakony Mts. as early as the Alaunian and, spanning also the Rhaetian Substage, it must continue in the Liassic Kardosrét Limestone of Dachstein type.

4. The Iszkahegy (Table 3). The sequence resting on the Balaton Redsandstone Formation here agrees, from the basal Triassic up to the topmost Anisian, almost completely with that of the Balaton Highland (J. ORAVECZ—E. VÉGH-NEUBRANDT, 1961, 1962; J. ORAVECZ—J. PUSKÁS, 1966; A. BÉRCZI-MAKK, 1970). Above the "Buchenstein Formation" represented by cherty and marly limestones and dolomites, but reduced in thickness, however, it is not the basin-facies Nemesvámos Limestone, but the *Budaörs Dolomite Formation* of reef-lagoonal facies with *Diplopora annulata*, extending from the Buda Mts. up to here, that follows. It is overlain by clay-marls, marly dolomites, cherty and marly limestones in alternation, of which a more complete sequence is known from boreholes (J. ORAVECZ—J. PUSKÁS, 1966), its Cordevolian-Julian age being only suggested by the "smaller-megalodontid-bearing" initial beds of the overlying Hauptdolomit, at least for the moment so. On the basis of its position, this formation, known in a thickness of about 80 m, seems to be a more limestone-rich, but relatively thin variety of the *Veszprém Marl*, in which to distinguish between "members" similar to the case of the Lake Balaton region is presently impossible. It is not improbable, however, that the lack here of younger members of the formation may be due to an overthrust of rigid Hauptdolomit masses.

Predominant species in the deepest known beds of the *Hauptdolomit* is *Neomegalodon triqueter pannonicus* whose acme begins with the Tuvalian; the Tuvalian is suggested, however, to have been the time when the species *Neomegalodon hoernesi*, associated with the former, made its first appearance, too (G. TICHY, 1974).

The Hauptdolomit of the Iszkahegy and the *Dachstein Limestone Formation* overlying it roughly correspond in time span to their counterparts in the northern Bakony Mts. The Dachstein limestone facies evolves here too with an alternation from the Hauptdolomit, but its Rhaetian and Lower Liassic parts have been lost to postdepositional erosion.

5. *The Vértes Mts.* (Table 3). The Seisian and, then, Campilian-type sediments of the LOWER TRIASSIC are known only from boreholes put down in the southern mountain foreland (S. VITÁLIS, 1939; borehole *Tabajd-5*). They develop without any break in sedimentation from the Upper Permian shallow-water, marine, carbonaceous to marly Tabajd Formation (I. SZABÓ, 1964).

MIDDLE and UPPER TRIASSIC. After a large gap of knowledge, the next to follow is a *Diplopora annulata*-bearing dolomite sequence identificable with the *Budaörs Dolomite Formation* of the Buda Mts. overlain by the cherty limestone, marl as well as calcareous and dolomitic marl sequence of the *Veszprém Marl Formation*, thinner even than on the Iszkahegy (30—40 m), with *Zygpleura hybrida* and *Cornucardia julii*. The thinness of the formation appears to be due — like in the case of the Iszkahegy occurrence — to its being overthrust by the Hauptdolomit masses following next up the profile.

The fossil content *(Cornucardia hornigi, Neomegalodon triqueter pannonicus* and *N. carinthiacus)* of that part of the lowermost horizon of the *Hauptdolomit Formation* coinciding with the Julian/Tuvalian boundary confirms this hypothesis. Otherwise, the Hauptdolomit reaches up into the Alaunian Substage as well. The *Dachstein Limestone* evolving from it with alternation is of Upper Alaunian-Sevatian age. It is separated, by an erosional unconformity, from the Liassic Hierlatz Limestone.

6. *The Gerecse Mts.* (Table 3). For the moment, only the *Hauptdolomit* and the *Dachstein Limestone Formations* are known from here. Judging by data published by E. JAKUCS (1954), E. VÉGH-NEUBRANDT (1960), E. VÉGH-NEUBRANDT—J. ORAVECZ (1961) and E. VÉGH-NEUBRANDT—J. ORAVECZ (1960—1961), these seem to be of Tuvalian-Lacian or Alaunian-Rhaetian age, respectively. The Hauptdolomit and the Dachstein Limestone Formation are interconnected by a passage characterized by alternation in lithologic content. The Dachstein Limestone and the Liassic are separated by erosional unconformity.

7. *Mány—Zsámbék—Dorog* (Table 3). The known part of the Triassic begins with the *Budaörs Dolomite Formation* with *Diplopora annulata* of Ladinian age which was observed, in addition to its outcrops at Zsámbék,

Szár and Szomor (J. ORAVECZ, 1961), in the 766.3—(1200.5) m interval of the borehole Budajenő Bő-2 as well, where its observed thickness was 434.3 m (Á. JÁMBOR, 1977).

This formation appears to be overlain by the alternating sequence of calcareous marls, marls and cherty dolomite and limestone beds intersected in the 285.0—337.1 m interval of the borehole Zsámbék Zsá-6. The boundary beds of the "*Annulata* dolomite" and the cherty limestone uncovered by the Zsámbék drill, however are for the moment unknown to us. The boreholes Mány-94, -102, -116, -165, -174 and -179 intersected this limestone in a thickness of about 240 m without having got out of it. This can be correlated with the deeper part of the *Veszprém Marl Formation* because above it, in the borehole Zsámbék Zsá-6, 42 m, in the afore-mentioned Mány boreholes 80—90 m of such a uniform and monotonous, dark grey clay and marl sequence was intersected which seems, with highest probability, to be equivalent to the higher part of the Veszprém Marl Formation. Consequently, the Veszprém Marl Formation of the Balaton and Bakony area appears to be intertonguing in E direction through the Iszkahegy and the Vértes and particularly through the Zsámbék and Mány basins, with the chert-rich, though entirely calcareous-dolomitic facies of the Buda Mts. That this intertonguing, supposed to exist for long, but, for lack of intermediate exposures, not proved heretofore, could now be traced is one of the most eloquent results of research devoted to the Central Mountains Triassic.

Accessible to study in outcrops already, the *Hauptdolomit* spans, in addition to the Tuvalian Substage, the greatest part of the Lacian Substage as well, inasmuch as can be judged from J. ORAVECZ's data (1961). The *Dachstein Limestone* connected with it by an alternation extends, in turn, from the top of the Lacian up to the Sevatian. The unconformable superposition of the *Arietites proaries*-containing Liassic limestone to the *Rhaetomegalodon incisus cornutus*-bearing Dachstein Limestone can be studied in the eastern part of the Nagykőszikla at Dorog (GY. VIGH, 1913; J. ORAVECZ, 1961).

8. The Buda Mts. (Table 4). The Lower Triassic sediments are hidden at great depths under the dolomite and limestone masses exposed to the surface and so far they could not be explored by drilling. The oldest known Triassic sediment is the mostly light, though sometimes a little darker grey, rock of the *Budaörs Dolomite Formation* with *Diplopora annulata* and *Macroporella beneckei*, though at the top rather with *Foraminifera* only, the Ladinian age of which was first recognized by E. KUTASSY (1926, 1927). On account of its estimated thickness of 1000—1200 m, it may span the Ladinian Stage as a whole. (The borehole Budaörs-1 has intersected — with exception of the "andesite" dike cut in the 775.0—831.4 m interval — this formation down to 910 m depth; but underneath, after a fault zone of 25 m vertical amplitude, it has penetrated into such a dark grey dolomite which E. NAGY—G. NAGY—

F. SZÉKY (1967) identified as Upper Carnian Hauptdolomit, though they had not found any index fossil in it.)

Because of block-faulting and the uneven distribution of fossils the overlying, higher-Triassic sequence is very difficult to decipher stratigraphically. The stripe by stripe repetition of its cherty and chertless members were interpreted by their explorers very differently with such striking extremes as, on the one hand, the alternation of tectonic windows and nappes (F. HORUSITZKY, 1943, 1949, 1956, 1958, 1959, 1961), on the other hand, as stripe-patterned elevations and subsidences of the bottom of the one-time sedimentary basin (GY. WEIN, 1976, 1977a, b). Both workers sought to arrange the facies of the higher parts of the Triassic so that they should build up two different sequences a "cherty" and a "chertless" one. The cherty sequence according to both of them would end with a peculiar facies of the mountains, the *Monotis—Halorella—Halorelloides*-bearing "*Újlakihegy dolomite*", while the chertless series would terminate with the *Dachstein Limestone* containing the richest Norian fauna of the Alpine facies realm. In contrast with this, all the earlier geologists, furthermore, also E. VADÁSZ (1953, 1960) and E. VÉGH-NEUBRANDT (*in* T. BÁLDI et al., 1973; 1974), believe that all the Triassic facies are constituents of one and the same sequence. The only difference in opinion between these authors consists in the order of succession of the members of this sequence. From among these authors, indeed, only E. VÉGH-NEUBRANDT (*in* T. BÁLDI et al., 1973; 1974) did, by gathering all facts in respect to setting of the strata, produce a stratigraphic sequence agreeing very well with other parts of the Transdanubian Central Mountains. With some corrections though, her stratigraphy is that which has been presented in the Table 4.

Accordingly, the member of the Buda Triassic following above the Budaörs Dolomite Formation may be the strata group called "*Mátyáshegy limestone*". As the highest bed of this can be regarded that chert-nodular limestone of a little pinky grey colour exposed in the northern escarpment of the Szépvölgy over a small patch. Its deeper part is represented by the poorly fossiliferous, finely stratified dolomites, dark clayey dolomites and marly limestones standing vertically in the 13.3—150.3 m interval of the borehole Pálvölgy Pvbg-1. The "Mátyáshegy limestone" seems to represent a constituent of the *Veszprém Marl Formation* still requiring further clarification, a formation whose total thickness and whose relation to its supposed footwall are for the moment unknown. As a kind of probable relationship is mentioned, by E. VÉGH-NEUBRANDT (1974), that the cherty dolomite pebbles frequent among the pebbles of the Eocene basal conglomerate transgressing over the *Annulata* dolomite of the Csíki-hegyek must derive from a sequence overlying the Budaörs Dolomite. Her argumentation, however, makes probable the superposition to the Budaörs Dolomite *not* of the Mátyáshegy Limestone, but only of its hanging wall, the cherty Sashegy Dolomite to be discussed hereinafter.

That the Szépvölgy outcrop of the "Mátyáshegy limestone" is overlain by a dolomite sequence grading from the former through dolomites containing large grey chert nodules at the bottom to finely bedded, siliceous dolomites at the top (I. LŐRENTHEY, 1907; K. BALOGH, 1961b) has been approved of by several authors since K. HOFMANN (1871). This rock is proposed by the present writer to be named, for its most beautiful locality, the *Sashegy Dolomite Formation*. From the deeper beds of this formation the representatives of *Lingula* were mentioned from the road-cut near the Ördögorom-csárda (I. LŐRENTHEY, 1907), *Daonella* cf. *aperta* from the Melocco-quarry on the north-east side of the Mátyáshegy (Z. SCHRÉTER et al., 1958), while from a pebble, got into an Eocene basal conglomerate, of its laminated-siliceous upper part *Koninckina telleri* was quoted (GY. VIGH, 1933). (As for the even otherwise very doubtful determination of *Alectryonia montiscaprilis* recorded from the road-cut of the Ördögorom by F. SCHAFARZIK (1902), F. SCHAFARZIK—A. VENDL (1929) and later also by F. HORUSITZKY (1943, 1961), it cannot be checked anymore owing to the fact that the specimen(s) have since got lost.)

Even though the occurrence of *Koninckina telleri* in the upper part of the Sashegy Dolomite Formation may bear witness to that the latter extends up into the Tuvalian Substage, the initial complex of the varicoloured and variably textured *Hauptdolomit Formation* probably directly overlying it cannot be older than Tuvalian and its top cannot be older than Upper Alaunian. The Hauptdolomit of the Buda Mts. contains, namely, three index faunae whose correlations are difficult to judge, because they were found in different blocks separated by faults. Of these the fauna of the APÁTHY-szikla (APÁTHY's rock) appears to be the oldest, with the following species — determined by K. HOFMANN (1871, 1873) and J. ORAVECZ (1968): *Cruratula stachei*, *Amphiclinodonta suessi*, *Cardita* aff. *crenata*, *Placites* cf. *placodes*, *Anatomites dulcis*, *Arcestes* sp. div., *Orthoceras campanile*. *A. dulcis* being certainly of Middle Tuvalian age (E. T. TOZER, 1971), F. HORUSITZKY's attempt (1943) at regarding the rock of APÁTHY-szikla as Ladinian can be considered frustrated.

Almost synchronous with the above fauna is the smaller *Megalodontids* and *Cornucardia* bearing fossil assemblage of the Kisgellérthegy (E. KUTASSY, 1925, 1927) and the eastern side of the Sashegy (E. VÉGH-NEUBRANDT, 1974): *Cornucardia hornigi*, *Cuspidaria gladius*, *Neomegalodon carinthiacus*, *N. triqueter pannonicus*, *N. boeckhi* n. ssp.

Youngest is the rich Upper Alaunian fauna of the "*Újlakihegy dolomite*" in which, according to GY. VIGH (1933), *Monotis salinaria*, *Halorelloidea rectifrons* and *Halorella amphitoma* are also present. This can by no means be restricted anymore to the Carnian Stage, as suggested by CS. DETRE (1971).

That the Norian Hauptdolomit is replaced by the lower part of the *Dachstein Limestone Formation*, a phenomen thus far observed only in the northern

Bakony Mts. within the Central Mountains Triassic, is manifested with particular vigour in the Buda Mts. Namely, if this replacement in the northern Bakony Mts. is restricted to the upper part of the Alaunian, in the Buda Mts. — as evidenced by the rich faunae on the Fazekas- and Remete-hegy (M. PÁLFY, 1921; E. VADÁSZ, 1920; E. KUTASSY, 1926, 1927, 1936; F. GÓCZÁN, 1961a, b) —, it extends to the Lacian Substage as well. That the Fazekas-hegy fauna was believed even by PÁLFY and KUTASSY to belong to the *Tropites subbullatus* Zone may have been due only to the poorer knowledge of the stratigraphic range of the ammonoids. During her revision of the fauna on the basis of 120 specimens, A. BÉRCZI-MAKK (1969) has determined from here the following ammonoids: *Dieneroceras* sp., *Clionitites pseudonodosus*, *Styrites collegialis*, *Arcestes* cf. *antonii*, *A. decipiens*, *A. tomostomus*, *Cladiscites pusillus*, *Cl.* cf. *striatissimus*, *Megaphyllites applanatus*, *M. jarbas*, *Placites myophorus*, *Pl. placodes*, *Monophyllites* sp., *Rhacophyllites neojurensis*. On the basis of these forms — in harmony with F. GÓCZÁN's opinion based on gastropods — the lower part of the Fazekas-hegy's Dachstein Limestone should be assigned to the Lacian. On the contrary, the Dachstein Limestone of the Remete-hegy belongs to Alaunian—Sevatian; moreover, its beds with *Triasina hantkeni* can reach as high as the Rhaetian. Some ammonoids (*Alloclionites* sp., *Helictites* sp., *Epi-* or *Hyposulcites*) the age of wich can be placed after L. KRYSTYN on the Alaunian/Sevatian boundary are listed by F. TATZREITER (1979. p. 119) from a fissure-filling of the Dachstein Limestone of Remete-hegy, too.

The alternation of Dachstein limestone with Hauptdolomit is known for the moment only from the Várhegy at Pesthidegkút (E. KUTASSY, 1927; E. VÉGH-NEUBRANDT, 1974). This, however, cannot be studied anymore in the present state of exposure.

With this the Triassic sequence of the Buda Mts. is interrupted and, as a result of postgenetic denudation, the Jurassic and Cretaceous sediments are also unknown.

9. The Pilis Mts. (Table 4). The oldest known member of the Triassic is the grey cherty limestone with interbedded clay layers cut in 44.7 and 73.6 m thickness, respectively, at the bottom of the coal-exploratory boreholes Pv-46 and Pv-48 at Pilisvörösvár. This limestone contains, in the borehole Pv-48, even small *Posidonomya* and *Avicula* specimens (K. BALOGH, 1961b). This, together with the overlying grey clay-marls, observed in 59.1 m and 58.4 m drilling thickness under the Eocene terrestrica, in which thin, reddish, brownish or grey, sometimes cherty, limestone layers may also be interbedded, can be correlated with the *Veszprém Marl Formation*, because the cherty Sas-hegy Dolomite Formation in the Pilis has not been found (at least up to the present).

The relation of the afore-mentioned marls and the Hauptdolomit Formation is not cleared yet. From the deepest part of the outcrops of the *Haupt-*

dolomit, however, *Schafhäutlia mellingi* and, from its somewhat higher part, *Cornucardia hornigi* and *Neomegalodon* cf. *carinthiacus* are known (K. BALOGH, 1961b). On this basis, the author believes that the age of the Hauptdolomit in the Pilis Mts. corresponds to the Upper Julian or Tuvalian. Beside surface outcrops, the Hauptdolomit passes even in the profile of the karstwater-well Pilisszentlélek-7 with an alternation into the Dachstein Limestone.

The Lacian-Alaunian age of the *Dachstein Limestone Formation* is proved by *Dicerocardium* sp. found by K. FERENCZ (1953) and *Halobia* cf. *norica* lumachelle described from the Feketehegy at Pilisszentlélek (K. BALOGH, 1961b; G. NAGY, 1961, 1964). The crinoidal Liassic limestone following, after an obvious break in sedimentation, higher up the profile fills the fissures observable in the saddle between the Öreg- and Fehér-szirt.

Beside the Dachstein Limestone, however, the Norian in the Pilis Mts. has also another representative of different facies which evolved at the northwestern tip of the Feketehegy at Pilisszentlélek with gradual transition from a little brownish variety of the Hauptdolomit. This facies, the two most important members of which are the "dark, heavily bituminous, well-stratified dolomite" and the about 240 m thick "*Avicula* limestone" overlying the former or alternating with it, is present in the Cserepes-valley adjacent in the north. The marly limestones of this locality, however, made the impression in their explorers that they might belong to the "Raibl" marls of the Carnian (E. VADÁSZ, 1953, 1960; Á. JÁMBOR et al., 1966). Its Feketehegy occurrence, however, first described by G. STACHE (1866) and F. SCHAFARZIK (1884, 1902), was declared by GY. VIGH (1928) and F. SCHAFARZIK —A. VENDL (1929) to contain *Rhaetavicula contorta* and *Modiola minuta* (*?*) and assigned to the "*Kössen Beds*" of the Rhaetian. Only J. ORAVECZ (1961), K. BALOGH (1961) and G. NAGY (1964) did recognize the identity of the formations of the two localities. Thus they managed to eliminate a rather grave stratigraphic confusion the traces of which still exist in the stratigraphic terms figured in the second edition of "*Hongrie*" (1978).

While elaborating the fossils collected from the two localities by F. SCHAFARZIK, GY. VIGH, Z. SCHRÉTER, E. VADÁSZ and himself, J. ORAVECZ found among them, beside a lot of specimens of *Aviculae* and other forms being transient from the Carnian Stage, several explicitely Norian species *(Arcestes bicornis, Worthenia contabulata, W. escheri)*; moreover, in the upper beds, somewhat different, lighter in colour compared to the lower ones of the Cserepesvölgy locality, he could identify even *Rhabdoceras suessi*. Despite the similarity of these *Rhabdoceras*-bearing sediments to GY. VIGH's *Contorta*-beds, he brought to question the presence of *Rhaetavicula contorta* (and consequently, of the "Kössen Beds") in the Pilis area, as he himself had not found among his self-collected fossils any form that might be identified with it. Since the elaborated fauna contains — along with the predominantly "Carnian"

forms — already Norian ones as well, he placed all these fossiliferous beds in the upper part of the Carnian and, respectively, evaluated them as transient with continuous sedimentation, into the Norian.

This dating by ORAVECZ should, at any rate, be modified. Not until the fauna is re-sampled layer by layer, it is possible, however, to review the distribution of the faunal elements. Therefore, the position of the fossiliferous sediments within the Norian Stage is not clear either. In the present state of knowledge the presence of *Rhabdoceras suessi* does not at all exclude the presence of *Rhaetavicula contorta*, and the concurrence of the two species may bear witness to extension of this peculiar series from the Upper Norian up to the Rhaetian. The new fossil sampling should be coupled with studies aimed at clarification of diverging opinions still existing, on account of the intricate tectonic setting, between K. BALOGH (1961b) and G. NAGY (1961, 1964) as to the composition and relations to footwall and hanging wall of the Feketehegy Formation.

10. The horsts on the Danube's left side (Table 4). Because of block-faulting, the stratigrapher is faced with a similarly difficult task, as in case of the Buda and the Pilis Mts. Since E. VADÁSZ (1910, 1911) up to most recent times almost everybody regarded the alternation of cherty limestones and grey marls in the Pokolvölgy quarry at Csővár as the Archimedean point of local Triassic stratigraphy. And since the afore-mentioned sediments were even quite lately sought to be pushed down to the Carnian Stage (CS. DETRE, 1969, 1970b), it was a great surprise to see H. KOZUR and H. MOSTLER (1973) conclude that the sediments in the afore-mentioned quarry are not of Carnian, but of Upper Sevatian age, as they contain, in addition to *Clionitites nicetae*, *Misikella hernsteini* as well. However, with a view to the fact that the top of the *Dachstein Limestone* in the immediate neighbourhood is — on the basis of its *Megalodontaceae (Dicerocardium hungaricum*, *Triadomegalodon tofanae*, *Neomegalodon complanatus*, *N. complanatus segestanus*, *N. hoernesi)* — also of (upper?) Norian age, the presence of parallel successions of two different (basin- and platform-) facies must be supposed here.

The sequence of *platform facies* probably begins with the "*Vashegy Hauptdolomit*" of unknown footwall; in absence of fossils, this sediment may be assigned to the Cordevolian-Julian. That it cannot be extended higher up to the Tuvalian is due to the fact that it is the *Dachstein Limestone* of reef facies of the Nézsa block with *Cornucardia hornigi*, *Sageceras haidingeri* and *Paralobites pisum* that has to be placed there. The Norian "normal" development of the Dachstein Limestone Formation would follow only after that.

The constitution of the *basin-facies* sequence ought to be given on the basis of the borehole Csővár-1. The section of this borehole, however, can be evaluated biostratigraphically only as regards the *Csővár Limestone Formation* extending below the surface to 522 m depth. Namely, as shown by H.

KOZUR and H. MOSTLER (1973), that part of this formation lying at 352.4 m depth and containing *Metapolygnathus diebeli*, *Metapolygnathus mostleri* and specimens (unfortunately totally lost in the meantime) of a new species, *Epigondolella pseudobidentata*, is still Cordevolian. Consequently, the basis of the formation situated round 170 m deeper may extend downwards as far as the upper boundary of the Ladinian. As far as the age of the 678 m of dolomite, uncovered by drilling and allegedly containing even cherts in its upper part (522.0—613.2 m) and called "*Pokolvölgy dolomite*" for the location of the borehole, is concerned, nothing can be said with certainty about its stratigraphic position owing to the insufficiency of description and to uncertainties of identification of *Daonella* specimens obtained in 564.3 m depth from cherty dolomites (CS. DETRE *in* H. KOZUR—H. MOSTLER, 1973). In case of normal position of the beds the "Pokolvölgy dolomite" may be of Ladinian age; if, however, its upper boundary is a tectonic one, its cherty part — modelled on the Buda Mountains — may reach up into the Julian-Tuvalian.

In the horsts on the Danube's left side the pre-Triassic Mesozoic was lost to erosion still before the Eocene, so that the Rhaetian/Liassic boundary cannot be studied here either.

B) THE BÜKKIAN FACIES AREA

11. The Bükk Mts. (Table 4). Beginning with oöidic limestones above the marine limestone beds of the Upper Permian, the Triassic sequence of almost 3.5 km thickness of the Bükk Mts., with its slight dynamothermal metamorphism and Carnic Alpine-Dinaric character, differs from all the other Hungarian contemporaries. In spite of the fifteen years that have elapsed since the publication of K. BALOGH (1964), only a few modifications on the succession of the described units are needed, merely the age of some members has shifted a little higher up the stratigraphic column, and only the isolated occurrences of limestone and diabase at Nekézseny, placed formerly to the Ladinian, proved to be — after findings of *Conodont* by S. KOVÁCS — as old as Lower Devonian. Notably, the *Nyavalyástető Limestone Member* of the *Hámor Dolomite Formation* evolving with alternation from the Lower Triassic *Ablakoskővölgy Formation*, in the wake of determinations done by courtesy of S. PANTIĆ-PRODANOVIĆ *(Glomospira articulosa, Gl. tenuifistula, Gl. sigmoidalis)* has turned out to belong — in contrast with its earlier assignment to the Lower Anisian — to the Illyrian already. The same can be suggested by the *Triadophyllum* finds known from this limestone for a long time, as corals from sediments older than this are unknown in the Alpine Triassic.

Because of the sharp distinction of the Nyavalyástető Limestone Member, the author proposes to distinguish the deeper, purely grey dolomitic part of the Hámor Dolomite Formation as *Vadászvölgy Dolomite Member*.

In new road-cuts on the east side of the Sebesvíz valley dolomite-conglomerates have been exposed to observation, sediments which — by virtue of their red-variegated, terrestrial clay interbeddings — may be interpreted as traces of a short, Late Illyrian break in sedimentation (H. KOZUR —R. MOCK, 1977).

As a consequence of extension of the Hámor Dolomite to the Anisian as a whole, all the sediments thus far found unfossiliferous have had to be made younger. This is how the *Szent-Istvánhegy Porphyrite Formation* overlying the Hámor Dolomite Formation as well as the *Fehérkő Limestone Formation* evolving with alternation from it have been assigned to the Ladinian, while a part of the *Kisgyőr Sericitic Shale Formation* (between Diósgyőr and Bélapátfalva) to the Carnian. This is confirmed by conodont finds that could be obtained from the pale-red, nodular interlayers of the Fehérkő Limestone and from some grey limestone beds in the Kisgyőr Sericitic Shale Formation as well as from limestone sequences overlying the latter and intertonguing both with one another and the *Óhuta Diabase Formation*. The composition of these has been presented in Table 3.

Macrofossils, the Norian age of which only recently proved *(Laballa suessi, Gemerithyris* sp., *Polygyrina elegans, Naticopsis hoernesi, Halobia* cf. *styriaca)* were even earlier known from the fossil-richest *Bervavölgy Limestone Formation* of the Bükk's Upper Triassic. To evaluate them properly was, however, handicapped by the presence of species being transient from the Carnian and by misidentifications of corals by G. KOLOSVÁRY. Recognition of the Norian age of the *Felsőtárkány Limestone Formation* was hindered by the flattened shape of the sparse fossils recovered from it and their uncertain determination *("Posidonia* cf. *wengensis"*, "*P.* cf. *pannonica*", "*Daonella pichleri*", "*Proarcestes* aff. *subtridentinus*", "*Trachyceras* cf. *roderici")*. But *Metapolygnatus posterus* and *M. abneptis* obtained by H. KOZUR and R. MOCK (1977) from this formation do confirm not only the Norian age; moreover they stimulate us to include also the "Nagyeged Limestone" with *Monotis salinaria haueri* in the notion of the Felsőtárkány Limestone Formation, even though it was earlier separated from this (K. BALOGH, 1964). It is all the more necessary because S. KOVÁCS has obtained *Metapolygnatus bidentatus* and *Gondolella steinbergensis* from it. Latterly the hypothesis arose that some representatives of the Kisgyőr Sericitic Shale in the SW Bükk would belong — at least partly — to the Jurassic. The prove of this hypothesis requires, however, further *detailed* investigations. In that work the *systematical* microfauna- and facies-determinations will play a decisive role.

C) THE AGGTELEK–SZILICE FACIES AREA

12. The Aggtelek Mts. (Table 5). Extending well to Slovakian territory and belonging to the Silice Nappe of the Western Carpathians, this karst plateau has a Triassic providing a fruitful subject for study owing to preservation of connections between its several ranges and in spite of the intricate distribution of the platform, reef-, reef-lagoonal and various basinal facies. For this reason, in the light of works of Hungarian and Slovak geologists (K. BALOGH, 1961a, 1964, 1976 and in "*Hongrie*", 1978; K. BALOGH—S. KOVÁCS, 1976; S. KOVÁCS, 1977, 1978a, b and 1979; J. BYSTRICKÝ, 1955, 1960, 1964, 1967, 1972, 1973; M. KOCHANOVÁ—J. MELLO—M. SIBLÍK, 1975; V. KOLLÁROVÁ-ANDRUSOVOVÁ—J. BYSTRICKÝ, 1974; V. KOLLÁROVÁ-ANDRUSOVOVÁ—M. KOCHANOVÁ, 1973; H. KOZUR—R. MOCK, 1973, 1974; J. MELLO, 1974, 1975, 1976, 1977; M. MIŠÍK—K. BORZA, 1976; R. MOCK, 1975; A. GAŻDZICKI—H. KOZUR—R. MOCK, 1979; M. SIBLÍK, 1967, 1971a, b, 1972, 1977; etc.), the constitution of its sequences is comparatively well known.

Further subdivision of the "*Seisian*" and "*Campilian beds*" of the Lower Triassic *Werfen Formation* and exploration of its relation to the Perkupa Evaporite Formation of Upper Permian (?) age need to be corroborated by additional evidence. Overlying the Werfen Formation, the *Gutenstein* and then the *Steinalm Formations* are present throughout the mountains and they continue, in many places, directly in the *Wetterstein Formation*, equally of reef- or reef-lagoon facies and of Ladinian-Julian age within which the larger dolomite bodies may even be considered to represent independent members. (Within the Steinalm Formation the dolomite bodies show a much more irregilar distribution, so that an independent "Steinalm Dolomite Member" can hardly be spoken of.) Since the Gutenstein sequence is entirely unfossiliferous, its reaching down into the topmost Lower Triassic is for the moment — after L. KRYSTYN, 1974b — merely supposed.

Much more information, based partly on *dasycladaceans*, partly on *brachiopods* and, less frequently, on *conodonts*, is available to the Upper Bithynian-Illyrian age of the Steinalm Formation (K. BALOGH—S. KOVÁCS, 1981).

The Ladinian age of the older part of the Wetterstein Formation is evidenced by *Diplopora annulata*, *Teutloporella nodosa*, and *Daonella lommeli*, its younger parts belonging to the Cordevolian-Julian being proved by finds of *Poikiloporella duplicata*, *Teutloporella herculea*, *Physoporella heraki* and *Uragiella supratriassica*.

The only known non-basin-facies formation of the Norian Stage is the "*Furmance Limestone*" (= "Dachstein-Riffkalk") with larger *Megalodontaceae* and *Heteroporella* cf. *carpathica* of the Somhegy at Drnava, Slovakia. It is overlain by the grey, crinoidal-coquinoidal *Drnava Limestone* of the

Bleskový prameň region, a formation of "Kössen type" containing numerous apparently Rhaetian brachiopods associated with ammonoids which were earlier believed to be Upper Norian. More than 19.6% of the bivalves and 47% of the gastropods could be identified with varying accuracy with known species; their stratigraphic value is not clear, however. No conodonts. On the basis of the ammonoids, V. KOLLÁROVÁ-ANDRUSOVOVÁ and M. KOCHANOVÁ (1973) regard the Drnava Limestone as of Sevatian age. H. KOZUR (1973) and A. GAŻDZICKI—H. KOZUR—R. MOCK (1979), however, have pointed out, with good reason, that these beds may belong, indeed, to the Lower Rhaetian already. This suggestion appears to be disapproved only by the absence of some characteristic Rhaetian forms *(Choristoceras marshi, Rhaetina gregaria, Austrirhynchia cornigera)* and the presence of the "Norian" *Halorella amphitoma.* On the other hand, it is corroborated by the following:

(1) five of the eight ammonoid genera extend from the *Columbianus* Zone up to the end of the Triassic, but *Cycloceltites, Vandaites (= Peripleurites)* and *Eopsiloceras* are already suggestive of the Rhaetian (E. T. TOZER, 1979);

(2) although the majority of brachiopods are equally transient forms, *Rhaetina piriformis, Triadithyris gregariaeformis, Zugmayerella koessenensis* and *Diplopora* cf. *phanerospora* also indicate a Rhaetian age.

Between the Drnava Limestone and the lowermost Liassic beds there is certainly an erosional unconformity.

In the more western parts of the mountains only the Illyrian portion of the Steinalm Formation and only the Ladinian (and maybe the Cordevolian) one of the Wetterstein Formation are replaced by basin facies represented by the *Schreyeralm Limestone Member* with *Flexoptychites flexuosus, Gondolella excelsa, G. mombergensis* and *Gladigondolella tethydis* and by the "Reifling Limestone Formation" containing *Gondolella polygnathiformis* and *Neocavitella tatrica* in its upper (chertless) part (S. KOVÁCS, 1979). The Upper Triassic, however, is dominated already in the west by basin facies; at Silická Brezova, Slovakia, village by the frontier, these appear from the top of the Julian onwards. Such a facies is the *Tisovec Limestone* overlying the Wetterstein Limestone, but still of transitional character in lithological sense, with a rich *Ammonoidea, Brachiopoda* and *Halobia* fauna of Upper Julian to Lower Lacian age; the white and red Lacian-Sevatian *Hallstatt Limestone Formation* with *conodonts* and *Monotis salinaria salinaria* (K. BALOGH, 1976) resting on the former; finally, the grey marls with *Misikella posthernsteini* representing the Zlambach Formation. According to J. MELLO and R. MOCK *(in* J. MELLO, 1975a), the Zlambach Formation at Bohuňovo, Slovakia, is overlain by "Gresten-type" (?) Lower Liassic sandstones and sandy limestones and, then by red brachiopodal-crinoidal Jurassic limestones. The exact age of these and their relation to the Zlambach Marl, however, still require further clarification.

Thus, for the moment, it is not known whether the break in sedimentation between the Rhaetian and the Liassic, well evidenced at Drnava, has extended here, too.

On the contrary, in the eastern part of the Aggtelek Mts. adjacent to the Rudabánya Mts., according to S. Kovács (1979), red facies of the *Hallstatt Limestone Formation* appears as early as the Ladinian. The *Steinalm Formation* with *Physoporella pauciforata* and *Neohindeodella aequiramosa* at the bottom of the borehole Szőlősardó-1 on the southern margin of the range, is followed, after some thin dolomarls, for the moment not given a special name owing to their transitional character, by the red-, grey- and green-mottled *Nádaska Limestone Formation*, observed also at the eastern tip of the Alsóhegy, a formation that may be considered, on account of its sedimentary structures, to represent submarine slope-deposited sediments. Judging by its conodont fauna, it may extend, from the Illyrian onwards, a little beyond the upper boundary of the Ladinian. Above it follow — in a way rather unusual for the Silica nappe — the Cordevolian-Lower Tuvalian beds with interbedded cherty limestone layers of the *Szőlősardó Marl Formation* which will then grade into the *Pötschen Limestone Formation* with *Halobia styriaca* and *Metapolygnathus nodosus* exposed to the surface. Although the time range of the Szőlősardó occurrence of this latter is not known exactly, it is taken to be identical with that of the Hallstatt red limestones. In the eastern part of the mountains — because of erosion of the Mesozoic younger than Upper Triassic — the Triassic/Jurassic boundary cannot be studied.

D) THE MECSEK—VILLÁNY FACIES AREA

13. The Mecsek Mts. (Table 5). The "germanic" Triassic of the Mecsek Mts. begins with the alluvial to lacustrine *Jakabhegy Sandstone Formation* overlying, after an erosional unconformity, with coarse basal conglomerates, the topmost "purple pebble conglomerates" of the Upper Permian Kővágószőlős Sandstone Formation. Although no fossil has ever been found in the bulk of the formation, it must nevertheless belong to the Lower Triassic, as, according to Á. Barabás-Stuhl (*in* A. Barabás et al., 1978), its immediate footwall contains, along with many typical Zechstein sporomorphs, still only very few Buntsandstein-type spores or pollen grains, while its topmost part belongs already to the Middle Buntsandstein, its hanging wall, in turn, connected with it through sediments of transitional character (the *Patacs Member* of the *Werfen Formation*), carries already Upper Buntsandstein palynomorphs. The conclusion that the Jakabhegy Sandstone Formation correlates in time with the Lower to Middle Buntsandstein, as was presumed already by J. Böckh (1876, 1881), has settled an important stratigraphic problem disputed for decades. After E. Vadász (1912a, b; 1917, 1935, 1953, 1960),

E. NAGY (1961) and E. NAGY and I. NAGY (1976) assigned the Jakabhegy Sandstone Formation, on the basis of lithological analogies, still to the Upper Permian and they commenced the Triassic with the ("Seisian") Patacs Member gradually evolving from it, though this was contradicted by both its lithological features and its *Costatoria costata* content. However, the *Sporomorpha* studies of Á. BARABÁS-STUHL, which involved both the Patacs Member and the *Magyarürög Member* overlying it, have borne a convincing witness to the respectively Upper Buntsandstein and "Lower Campilian" character of these members. The predominantly carbonate-facies Lower Triassic sequences overlying them, sequences now called, after E. NAGY (in "*Hongrie*" 1978), the *Hetvehely Dolomite* and the *Víganvár Limestone Member*, respectively, are undoubtedly of "Upper Campilian" character. By virtue of their geological features, the four Mecsek Lower Triassic members above the Jakabhegy Sandstone Formation are regarded as local facies of "Campilian character" of the Werfen Formation.

The *Middle Triassic* is spanned by the *Misina Formation* instituted by picking out the common features of the nine, predominantly carbonaceous stratigraphic units distinguished by E. NAGY (in "*Hongrie*", 1978). It would be needless to reproduce here their lithological description (E. NAGY, 1968), as that description has since been added to only by an account of a new variety (containing also coaly clay streaks) of the Mánfa Beds (B. WÉBER, 1978). However, if E. NAGY (1968) considered the whole Misina Formation, together with the *Trigodonus* and *Mánfa Beds*, to be Anisian, it appears now to span beside the Anisian the whole Ladinian as well. There are two reasons that account for this:

(1) The *Bertalanhegy Limestone Member*, containing, beside "Recoaro" brachiopods (*Decurtella decurtata*, *Coenothyris vulgaris*, etc.), i.a. *Paraceratites binodosus* and *Ptychites evolvens* and *Gondolella bulgarica*, as well, cannot be declared to be exclusively Pelsonian, as it (at least its upper part), extends into the *Paraceratites trinodosus* Zone (i.e. the Illyrian Substage) as well.

(2) The *Trigonodus Beds* at the top of the formation correlate best with the "Trigonodus Dolomite" at the base of the Lettenkeuper in the Germanic facies. The Lettenkeuper, in turn, corresponds, according to H. KOZUR (1975), to the upper part of the Longobardian Substage.

Thus, also the age determination of the two formations above the *Mánfa Beds* will change, of course:

(1) The transitional Mánfa Beds are overlain by 120 m of UPPER TRIASSIC sediment called the *Kantavár Formation* constituted by marls and limestones at its base and by a succession of shales and clay-marls and, finally, sandstones, higher up the profile. Nota bene, though all that which can be said about the vegetal remains obtained from the formation is that they are older than the Rhaeto-Liassic flora, its molluscs *(Anodontophora recta*, *A. lettica*, *Nati-*

copsis uhligi), however, indicate, even according to their age data published by E. NAGY (1968, p. 85), the presence of deeper part (Cordevolian-Julian) of the Carnian Stage.

(2) Evolving with a transition in lithology from the Kantavár Formation, the 330 to 520 m of sediment of varying colour and grain size called the *Karolinavölgy Sandstone Formation* (i.e. "the sandstone without coal-seems" of J. BÖCKH, 1876) can be divided into a lower (predominantly lagoonal and lacustrine), a middle (predominantly lacustrine) and an upper (fluviatile, deltaic and lacustrine) part. As evidenced by the borehole Pécs-28 chosen to be a standard key section, the lower part (constituted for the most part by greenish-grey or greyish-red sediment of 100 to 140 m thickness) contains, in its lower 1/3, *phyllopods*, in its upper 2/3, in turn, *fish remnants* (mainly *Semionotus)*. The *Isaura minuta*- and *I. ovata*-bearing "Phyllopoda group" corresponds, by and large, to the "Estheria Beds" of the Germanic Triassic, and these are declared by H. KOZUR, with conviction, to be of Julian age. Consequently, the *Semionotus*-bearing group would be Tuvalian, a hypothesis corroborated by *sporomorph* determinations by J. BÓNA (1972). Above the middle (Lacian—Alaunian) part of the Karolinavölgy Sandstone the beginning of the Rhaetian is counted from the fluviatile sediments deposited on the first surface of emersion; palaeontologically, it is reckoned with from the appearance of Rhaeto-Liassic macro- and microflora. Furthermore, the recognition of the Rhaetian members is facilitated by the occurrence of thin chamosite and clay-ironstone layers interbedded with their higher part.

The Rhaetian-Liassic transition is so gradual that the boundary between the two was up to recent times drawn with the first coal seam (E. VADÁSZ, 1935) or the first moory sediment. Nevertheless, Z. NÉMEDI VARGA (1971), relying on palynological results by J. BÓNA, the change in lithology and the occurrence of *Isaura hungarica* above the first coal seam, (this fossil was found heretofore only below that seam), proposed to draw the Triassic-Jurassic boundary 80 to 100 m above the first, so-called Alpha Seam. According to J. BÓNA (1973, 1979), some species of the characteristic spine-ornamented genera of the Upper Triassic *(Apiculatisporites, Anapiculatisporites, Anemiidites, Anaplanisporites)* do really survive — along with predominance of the smooth-skinned *Cyclina* genus — the horizon of the Alpha Seam, though the species *Ephedripites tortuosus*, known in Germany only from the Liassic, also appears already, associated with them. Consequently, the definition of the boundary has invariably remained a question of viewpoint.

14. The Villány Mts. (Table 5). The deeper part of the LOWER TRIASSIC is filled, here too, by the unfossiliferous *Jakabhegy Sandstone Formation* which, in absence of the Upper Permian Kővágószőlős Sandstone Formation known only from the Mecsek, directly transgresses onto the Middle (?) Permian Vokány Quartz-Porphyry Formation.

The higher part of the Lower Triassic is represented by local facies of the upper part of the *Werfen Formation*. Discovered not long ago in the northern foreland of the mountains and uncovered in the boreholes Turony-1, Bisse-1, Vokány-2, Peterd-1 and Rádfalva-1, these sediments, unfossiliferous at their base, but containing, in addition to *sporomorphs*, also *Costatoria costata* higher up the profile, were described as independent "*Turony*" and "*Vokány Formation*", respectively, by E. NAGY and I. NAGY (1976). The former begins with an alternation of red and green, fine-grained sandstones, but passes soon into an anhydrite-gypsum-bearing sequence. This consists of an alternation of dolomites, dolomarls, dolomitic limestones, anhydrites and gypsum and various sorts of siltsone, so that the evaporite-rich beds are replaced by sediments of increasing carbonate content higher up the profile. *By virtue of the above, the two "formations" cannot be considered but local members of the Werfen Formation.*

Excelling with the uniformly "Germanic" character of its members and representing an alternation of purely carbonate sequences (dolomite and limestone), the MIDDLE TRIASSIC can be split up into formations distinguished by E. NAGY and I. NAGY (1976). As far as the age of these formations is concerned, the author proposes merely a slight change of it. In agreement with J. BÓNA (1976), it is only a part of the *Brachiopoda* fauna, sampled *not* layer by layer, of the *Zuhánya Limestone Member* that can be regarded as Pelsonian. A considerable part of the "Recoaro" brachiopods did persist, as is generally known, throughout the Illyrian, moreover, at the base of the Fassanian as well. Thus the Illyrian age of the higher part of the Zuhánya Limestone Member and the Fassanian age of the *Csukma Dolomite Member* as a whole seem to be more plausible than their earlier dating (Pelsonian and Illyrian, respectively). Of course, this implies that the *Templomhegy Dolomite Member*, considered Lower Ladinian, by E. NAGY and I. NAGY (1976), ought to be transferred to the Longobardian.

Post-Longobardian Triassic formations are absent in the Villány Mts. and the Pliensbachian transgresses immediately on the Templomhegy Dolomite.

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Beside their outcrops in the mountain ranges the Triassic sequences discussed in the foregoing can be traced in the substratum of the intermontane basins as well. The Transdanubian Central Mountains Triassic is known to us as far as Tura to the east of Budapest and also from the bottom of the Zala Basin. The southern and southwestern continuation of the Bükkian Triassic has been drilled in the southern foreland of the Bükk Mts., at Jászberény and Tóalmás, while some members of the Mecsek Triassic have been uncovered by other boreholes even in the Danube—Tisza Interfluve area. However,

because of insufficiencies in core recovery and material testing and the great distances from the surrounding mountains, it is at present impossible to incorporate the Triassic sequences found in the substratum of the southern Great Hungarian Plain into a convenient stratigraphic system. Therefore, the author has had to refrain from introducing them, though a faciological and stratigraphic analysis of these formations could largely promote the understanding of the evolution of the substratum of the Hungarian basins, very important as it would be for assessment of the development of the Carpathian arc. This cannot be achieved, however, by simply listing the Triassic formations according to their age (Á. SZALAY—K. SZENTGYÖRGYI—A. SZŐTS, 1978), but it would require a comparative approach to formations in different areas and tectonic units, respectively.

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G.H.S.G. = Geologica Hungarica Series Geologica, Budapest
G.P.Sp. = Geologický Práce, Správy, Bratislava
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КОРРЕЛЯЦИЯ ТРИАСОВЫХ ОТЛОЖЕНИЙ НА ТЕРРИТОРИИ ВНР

К. БАЛОГ

Резюме

Не считая триасовых толщ Венгерской низменности и юго-запада Задунайского края, известных только по буровым скважинам, а также еще не выясненных толщ триаса в горах Рудабанья, триасовые отложения венгерских горных районов относятся к четырем различным областям развития (Среднегорной, Бюккской, Аггтелек-Силицкой и Мечек-Вилланьской). Эти фациальные области своей идентичностью обязаны своеобразию и способу переплетения литологических единиц платформенной, бассейновой, вулканической или речно-озерной фаций, последовавших за пляжевыми, мелководно-морскими или эвапоритными осадками раннетриасовой Тетисской трансгрессии. Фациальные сдвиги внутри отдельных областей развития триасовых отложений особенно бросаются в глаза на территории Задунайского среднегорья и Аггтелекского карста. Автор настоящей работы попытается включить эти литостратиграфические единицы в такую геохронологическую схему, которую он считает наиболее подходящей из всех вариантов нового международного стандарта, находящегося в стадии становления. Основные положения работы сводятся к следующему:

Развитие среднетриасовых отложений на территории среднегорий варьирует в большей степени, чем предполагали раньше. Рифогенные, лагунные фации анизийского яруса (в отличие от Балатонского нагорья и гор Баконь) продолжаются и в ладинском ярусе в более восточных частях среднегорья. Карнийская по возрасту Веспремская Мергельная Формация прослеживается от г. Кестхей вплоть до Будайских гор и там она зубовидно сочленяется с богатой кремнями Шашхедьской Доломитовой формацией. Образование Главнодоломитовой Формации даже в юго-западной части среднегорья не ограничивается везде норийским ярусом, а дальше на северо-восток оно — с некоторым колебанием — начинается в все более древних горизонтах карнийского яруса. Этому «опусканию ниже по стратиграфической колонке» Главнодоломитовой Формации следует и начало Дахштейнской Известняковой Формации. Дело в том, что образование Дахштейнских Инствестняков, имеющих на юго-западе еще выраженно рэтский возраст, дальше на северо-восток начинается в алаунском ярусе, а в глыбах левобережья Дуная уже в тувальском ярусе. В северной части Северной Бакони и восточнее ее так называемые «Кёссенские слои», включающиеся между Дахштейнской и Главнодоломитовой Формациями, полностью выпадают и они снова будут появляться только в Фекетехедьской Формации, охватывающей большую часть норийского яруса в горах Пилиш. В связи с выпадением из толщи «Кёссенских слоев» Дахштейнская Известняковая Формация Северной Бакони и Будайских гор об-

наруживает частичное или полное зубовидное сочленение с Главнодоломитовой Формацией. Очень важно, что кремнистая Чёварская Известняковая Формация, находящаяся в непосредственном соседстве рифогенно-лагунных дахштейнских Известняков глыб левобережья Дуная, не ограничивается карнийским ярусом, как полагали раньше, а здесь она замещена Дахштейнской Известняковой Формацией тувальско-севатского возраста.

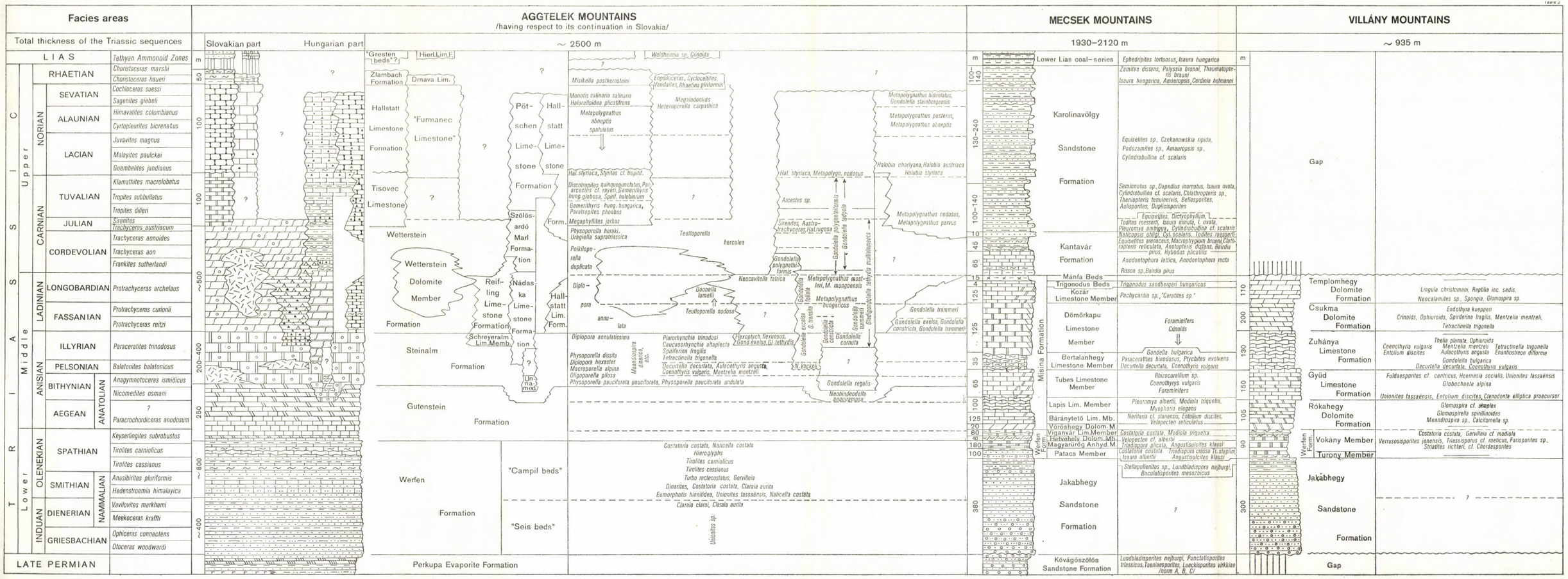
Возраст литостратиграфических единиц Бюкского района, судя по некоторым находкам конодонт и фораминифер, оказывается несколько сдвинутым вверх по сравнению с принятой раньше точкой зрения автора.

Сложная стратиграфия Аггтелек-Силицкого района изменилась всего лишь столько, что по данным исследований конодонт Альшохедьские известняки, ранее считавшиеся шрейеральмскими (то есть иллирийскими), теперь оказались ладинским, карнийским и норийским подразделениями Галлштаттской Известняковой Формации. В скважине Сёлёшардо-I, пробуренной на южном краю района, под Пётченской Известняковой Формацией-тувало-алаунского возраст вскрыли Сёлёшардоскую Мергельную Формацию кордеволо-нижнетувальского возраста, до сих пор неизвестную на территории Силицкого покрова, а в постели последней была вскрыта Надашкайская Известняковая Формация среднеиллирийско-лангобардского возраста, представленная фацией континентального склона. Под этой же формацией, в связи с обусловленным тектоническими причинами отсутствием (?) нижнеиллирийского и пелсонского подъярусов, следует толща кремнистых доломергелей и алевролитов, содержащих вулканические туфы, которая, из-за своего битинийского возраста (?) заставляет нас отнести — по крайней мере локально — и подстилающую ее Штейнальмскую Формацию значительно ниже по стратиграфической колонке, чем предполагалось раньше.

Палинологические исследования, проведенные в Мечек-Вилланьской области, доказали нижнетриасовый возраст Якабхедьского Песчаника. Положение известных литологических единиц среднего и верхнего триаса несколько изменилось в результате переоценки элементов фауны и флоры.

Нижняя граница триасовых толщ Венгрии является в общем литологически четкой, но биостратиграфически она не всюду доказывается убедительно. Часть резких границ над обломочными толщами перми (Балатонское нагорье, Мечек, и Виллань) подчеркивается эрозионным несогласием и наличием базальных конгломератов. Нижний триас южной передовой зоны гор Вертеш и Аггтелекского карста залегает на эвапоритах верхней перми. Нижнетриасовые оолиты гор Бюкк налегают, наоборот, на морские известняки верхней перми.

Верхняя граница толщ меняется из одного горного массива на другой. Дахштейнская Известняковая Формация Балатонского нагорья, Южной и Северной Бакони, а также Каролинавёльдская Песчаниковая Формация гор Мечек переходят без перерыва в аналогичную фацию геттанжского яруса. Перерыв между триасом и юрой в горах Вертеш, Герече и Пилиш, а также в Аггтелек-Силицкой области относительно невелик. Однако, в глыбах левобережья Дуная, а также в Будайских горах (где отсутствуют целая юра и нижний мел) у нас пока нет данных, по которым могли бы судить о том, сколько из этого проблема обусловлено внутримезозойским перерывом в осадконакоплении и сколько третичным размывом.


Facies areas
Total thickness of the Triassic sequences
LIAS
Tethyan Ammonoid Zones
TRIASSIC
Upper
Middle
Lower
RHAETIAN
SEVATIAN
ALAUNIAN
LACIAN
TUVALIAN
JULIAN
CORDEVOLIAN
LONGOBARDIAN
FASSANIAN
ILLYRIAN
PELSONIAN
BITHYNIAN
AEGEAN
SPATHIAN
SMITHIAN
DIENERIAN
GRIESBACHIAN
NORIAN
CARNIAN
LADINIAN
ANISIAN
OLENEKIAN
INDUAN
ANATOLIAN
NAMMALIAN
Choristoceras marshi
Choristoceras haueri
Cochloceras suessi
Sagenites giebeli
Himavatites columbianus
Cyrtopleurites bicrenatus
Juvavites magnus
Malayites paulckei
Guembelites jandianus
Klamathites macrolobatus
Tropites subbullatus
Tropites dilleri
Sirenites
Trachyceras austriacum
Trachyceras aonoides
Trachyceras aon
Frankites sutherlandi
Protrachyceras archelaus
Protrachyceras curionii
Protrachyceras reitzi
Paraceratites trinodosus
Balatonites balatonicus
Anagymnotoceras ismidicus
Nicomedites osmani
?
Paracrochordiceras anodosum
Keyserlingites subrobustus
Tirolites carniolicus
Tirolites cassianus
Anasibirites pluriformis
Hedenstroemia himalayica
Vavilovites markhami
Meekoceras kraffti
Ophiceras connectens
Otoceras woodwardi
LATE PERMIAN
AGGTELEK MOUNTAINS
/having respect to its continuation in Slovakia/
Slovakian part
Hungarian part
~ 2500 m
m
50
100
100
~500
200–400
250
~800
~400
Zlambach Formation
Dnava Lim.
Hallstatt Limestone Formation
"Furmanec Limestone"
Tisovec Limestone
Wetterstein
Wetterstein Dolomite Member
Formation
Steinalm Formation
Gutenstein Formation
Werfen Formation
Pötschen Limestone Formation
Szölősardó Marl Formation
Hallstatt Limestone Form.
Reifling Limestone Formation
Nádaska Limestone Formation
Schreyeralm Lim.Memb.
Hallstatt Lim. Form.
"Campil beds"
"Seis beds"
Perkupa Evaporite Formation
Misikella posthernsteini
Monotis salinaria salinaria
Halorellioidea plicatifrons
Metapolygnathus abneptis spatulatus
Hal. styriaca, Styriites cf. trupiti
Discotropites, quinqueunctatus, Parcarcestes cf. tayeri, Gemerithyris hung. globosa, Spirif. halobiarum
Gemerithyris hung. hungarica, Paratropites phoebus
Megaphyllites jarbas
Physoporella heraki, Uragiella supratriassica
Poikiloporella duplicata
Diplopora annulatissima
Physoporella dissita
Diplopora hexaster
Macroporella alpina
Diplopora pilosa
Physoporella pauciforata pauciforata, Physoporella pauciforata undulata
Meandrospira dinarica, etc.
Piarorhynchia trinodosi
Caucasorhynchia altaplecta
Spiriferina fragilis
Tetractinella trigonella
Decurtella decurtata, Aulacothyris angusta, Coenothyris vulgaris, Mentzelia mentzeli
Costatoria costata, Naticella costata
Hieroglyphs
Tirolites carniolicus
Tirolites cassianus
Turbo rectecostatus, Gervilleia
Dinarites, Costatoria costata, Claraia aurita
Eumorphotis hinnitidea, Unionites fassaënsis, Naticella costata
Claraia clarai, Claraia aurita
Unionites sp.
Gresten beds"?
Hierl.Lim.F.
Waldheimia sp. Crinoids
Eopsiloceras, Cycloceltites, Vandaites, Rhaetina piriformis
Megalodontids
Heteroporella carpathica
Teutloporella herculea
Teutloporella nodosa
Neocavitella tatrica
Daonella lomelli
Halobia charlyana, Halobia austriaca
Halobia styriaca
Metapolygnathus nodosus, Metapolygnathus parvus
Metapolygnathus bidentatus, Gondolella steinbergensis
Metapolygnathus posterus, Metapolygnathus abneptis
Hal. styriaca, Metapolygn. nodosus
Arcestes sp.
Sirenites; Austrotrachyceras, Hal.rugosa
Gondolella polygnathiformis
Gondolella tadpole
Gondolella polygnathiformis
Gondolella multidentata
Gondolella tethydis
Gondolella trammeri
Gondolella exelsa, Gondolella constricta, Gondolella trammeri
Metapolygnathus nevadites, M. mungenensis
Metapolygnathus hungaricus
Gondolella exelsa
G. transita
G. folana
Gondolella constricta
Gondolella cornuta
N. kockeli
Gondolella regalis
Neohindeodella aequiramosa
Hexoptych. flexuosus, Gond. exelsa, G. tethydis
MECSEK MOUNTAINS
1930–2120 m
Lower Lias coal-series
Karolinavölgy Sandstone Formation
Kantavár Formation
Mánfa Beds
Trigonodus Beds
Kozár Limestone Member
Dömörkapu Limestone Member
Bertalanhegy Limestone Member
Tubes Limestone Member
Lapis Lim. Member
Báránytető Lim. Mb.
Vöröshegy Dolom. M.
Viganvár Lim.Member
Hetvehely Dolom.Mb.
Magyarürög Anhyd.M.
Patacs Member
Misina Formation
Werfen Form.
Jakabhegy Sandstone Formation
Kővágószőlős Sandstone Formation
Ephedripites torfuosus, Isaura hungarica
Zamites distans, Palyssia bronni, Thaumatopteris brauni
Isaura hungarica, Amauropsis, Cardinia hofmanni
Equisetites sp., Czekanowskia ripida, Podozamites sp., Amauropsis sp., Cylindrobullina cf. scalaris
Semionotus sp., Dapedius inornatus, Isaura ovata, Cylindrobullina cf. scalaris, Chlathropteris sp., Theniopteris tenuinervis, Bellisporites, Aulisporites, Duplicisporites
Equisetites, Dictyophyllum,
Todites roesserti, Isaura minuta, I. ovata, Pleuromya ambigua, Cylindrobullina cf. scalaris
Naticopsis uhligi, Cyl. scalaris, Todites roesserti
Equisetites arenaceus, Macrotyppum brauni, Clathropteris reticulata, Anotopteris distans, Bairdia pirus, Hybodus plicatilis
Anodontophora lettica, Anodontophora recta
Rissoa sp., Bairdia pirus
Trigonodus sandbergeri hungaricus
Pachycardia sp., "Ceratites sp."
Foraminifers
Crinoids
Gondella bulgarica
Paraceratites binodosus, Ptychites evolvens
Decurtella decurtata, Coenothyris vulgaris
Rhizocorallium sp.
Coenothyrys vulgaris
Foraminifers
Pleuromya albertii, Modiola triquetra, Myophoria elegans
Neritaria cf. stanensis, Entolium discites, Velopecten reticulatus
Costatoria costata, Modiola triquetra
Velopecten cf. albertii
Triadispora plicata, Angustisulcites klausi
Costatoria costata Triadispora crassa Tr. staplini, Isaura albertii Angustisulcites klausi
Stellapollenites sp., Lundbladispora nejburgi, Baculatisporites mesozoicus
Lundbladisporites nejburgi, Punctatisporites triassicus, Taeniaesporites, Lueckisporites virkkiae /form A, B, C/
VILLÁNY MOUNTAINS
~ 935 m
m
110
200
130
150
105
90
300
Gap
Templomhegy Dolomite Formation
Csukma Dolomite Formation
Zuhánya Limestone Formation
Gyüd Limestone Formation
Rókahegy Dolomite Formation
Werfen Form.
Vokány Member
Turony Member
Jakabhegy Sandstone Formation
Gap
Lingula christomani, Reptilia inc. sedis, Neocalamites sp., Spongia, Glomospira sp.
Endothyra kuepperi
Crinoids, Ophiuroids, Spiriferina fragilis, Mentzelia mentzeli, Tetractinella trigonella
Thelia planata, Ophiuroids
Coenothyris vulgaris Mentzelia mentzeli Tetractinella trigonella
Entolium discites Aulacothyris angusta Enantiostreon difforme
Gondolella bulgarica
Decurtella decurtata, Coenothyris vulgaris
Fuldaesporites cf. centricus, Hoernesia socialis, Unionites fassaënsis
Globochaete alpina
Unionites fassaënsis, Entolium discites, Ctenodonta elliptica praecursor
Glomospira cf. simplex
Glomospirella spirillinoides
Meandrospira sp., Calcitornella sp.
Costatoria costata, Gervilleia cf. modiola
Verrusosisporites jenensis, Triassisporus cf. roeticus, Farispores sp., Striatites richteri, cf. Chordasporites
?

Facies areas			KESZTHELY MOUNTAINS	BALATON HIGHLAND & SOUTH BAKONY	NORTH BAKONY	ISZKA HILL	VÉRTES MOUNTAINS	GERECSE MOUNTAINS	MÁNY–ZSÁMBÉK–DOROG
Total thickness of the Triassic sequences			>600 m	~2730 m	>2160 m	~3600 m	~3200 m	>1800 m	>2700–3100 m
LIAS		Tethyan Ammonoid Zones							Arietites proaries
TRIASSIC, Upper	RHAETIAN	Choristoceras marshi		Kardosrét Limestone Member	Brachiopods		Hierlatz Lim. Form. / Waldheimia alpina	Hierlatz L. F. / Oxynoticeras oxynotum	
		Choristoceras haueri	"Kössen beds" / Rhaetavicula contorta	Dachstein Limestone Formation / "Kössen beds" / Rhaetavicula contorta, Cardita austriaca	Rhaetina gregariaeformis, Conchodon infrapunctatus, Rh. contorta, Tr. gryphoides; "Kössen beds" / Dachstein Limestone Formation; Dicerocardium pteriforme, Neomegalodon acutaminatus	Dachstein Limestone Formation	Dachstein Limestone Formation	Triasina hantkeni, Rhaetimegalodon bajotensis, Rh. incisus cornutus, Conch. infralasicus	Dachstein Limestone Formation / Rhaetimegalodon incisus cornutus
	NORIAN: SEVATIAN	Cochloceras suessi	Hauptdolomit Formation	Hauptdolomit Formation	Neomegalodon complanatus, N. boeckhi, Gemmelarodus seccoi	Hauptdolomit Formation	Hauptdolomit Formation / Triadomeg. cf. gryphoides	Neomegalodon complanatus, N. guembeli	?
	NORIAN: ALAUNIAN	Sagenites giebeli							Dicerocardium curioni, Neomegalodon complanatus, Neomegalodon boeckhi, N. mediofasciatus
	NORIAN: ALAUNIAN	Himavalites columbianus							
	NORIAN: ALAUNIAN	Cyrtopleurites bicrenatus							
	NORIAN: LACIAN	Juvavites magnus		Worthenia contabulata, Neomegalodon hoernesi	Neomegalodon hoernesi, Neomegalodon laczkói			Gemmelarodus paronai	Gemmelarodus seccoi
	NORIAN: LACIAN	Malayites paulckei							Isognomon exilis, Dicerocardium sp.
	NORIAN: LACIAN	Guembelites jandianus							
	CARNIAN: TUVALIAN	Klamathites macrolobatus	"Ederics limestone" / Veszprém Marl Formation / Frondicularia bryzaeformis, Corals / Spiriferina bittneri, Cornucardia hornigi, Alectryonia monticaprinis	"Sándorhegy limestone" / Veszprém Marl Formation	"Sándorhegy limestone" / Veszprém Marl Formation; Veghihynchia arpadica / Costatoria inaequicostata	Veszprém Marl Formation / Neomegalodon hoernesi, Neomegalodon triqueter pannonicus	Hauptdolomit Formation / Nuculids, Costatoria inaequicostata, Cornucardia hornigi, Neomegalodon triqueter pannonicus, N. carinthiacus	Hauptdolomit Formation / Costatoria cf. inaequicostata	Hauptdolomit Formation / Neomegalodon hoernesi, Neomegalodon triqueter pannonicus
	CARNIAN: TUVALIAN	Tropites subbullatus							
	CARNIAN: TUVALIAN	Tropites dilleri							
	CARNIAN: JULIAN	Sirenites		"Austriacum limestone" / Veszprém Marl Formation					
	CARNIAN: JULIAN	Trachyceras austriacum	? ? ?	"Lima & Nucula marl" / Veszprém Marl Formation; "Austriacum limestone" / Veszprém Marl Formation	Foraminifers				
	CARNIAN: CORDEVOLIAN	Trachyceras aonoides		"Estheria marl" / Veszprém Marl Formation; "Chondrites limestone" / Veszprém Marl Formation					
	CARNIAN: CORDEVOLIAN	Trachyceras aon		"Füred limestone" / Veszprém Marl Formation					
	CARNIAN: CORDEVOLIAN	Frankites sutherlandi							
TRIASSIC, Middle	LADINIAN: LONGOBARDIAN	Protrachyceras archelaus		Nemesvámos Limestone Formation					
	LADINIAN: FASSANIAN	Protrachyceras curionii							
	LADINIAN: FASSANIAN	Protrachyceras reitzi		"Buchenstein Formation" / Budaörs Formation; "Reifling limestone" / "Tagyon limestone" / "Recoaro limestone" / Megyehegy Dolomite Formation		Budaörs Dolomite Formation / Diplopora annulata; "Buchenstein Formation"	Budaörs Dolomite Formation / Diplopora annulata		Budaörs Dolomite Formation / Diplopora sp., Oberhauserella neotriassica, Trochammina alimtlensis
	ANISIAN: ILLYRIAN	Paraceratites trinodosus							
	ANISIAN: PELSONIAN	Balatonites balatonicus							
	ANISIAN: BITHYNIAN (ANATOLIAN)	Anagymnotoceras ismidicus							
	ANISIAN: BITHYNIAN (ANATOLIAN)	Nicomedites osmani							
	ANISIAN: AEGEAN (ANATOLIAN)	? Paracrochordiceras anodosum							
TRIASSIC, Lower	OLENEKIAN: SPATHIAN	Keyserlingites subrobustus		Iszkahegy Limestone Member		Iszkahegy Limestone Member / Costatoria costata			
	OLENEKIAN: SPATHIAN	Tirolites carniolicus		Aszófő Dolomite Member / Gervilleia modiola		Aszófő Dolomite Member			
	OLENEKIAN: SPATHIAN	Tirolites cassianus		Csopak Marl Member / Crinoids, Gervilleia, Tirolites cassianus, Dinarites, Naticella costata	"Campil beds" / Costatoria costata; Veryachium, Verrucosisporites morulae, Grandisp. ulrichi, Falcispor. enodys	Csopak Marl Member / Tirolites	"Campil beds"		
	OLENEKIAN: SMITHIAN (NAMMALIAN)	Anasibirites pluriformis		Hidegkút Sandstone Member / Eumorphotis hinnitidea, Pseudomonotis laczkói		Hidegkút Sandstone Member			
	OLENEKIAN: SMITHIAN (NAMMALIAN)	Hedenstroemia himalayica							
	OLENEKIAN: DIENERIAN (NAMMALIAN)	Vavilovites markhami		Arács Marl Member / Claraia aurita, Claraia clarai	"Séis beds" / Claraia sp.; Isaura alberti	Arács Marl Member / Claraia aurita, Claraia clarai	"Séis beds" / Claraia clarai		
	OLENEKIAN: DIENERIAN (NAMMALIAN)	Meekoceras kraffti							
	INDUAN: GRIESBACHIAN	Ophiceras connectens							
	INDUAN: GRIESBACHIAN	Otoceras woodwardi		Nádaskút Dolomite Member		Nádaskút Dolomite Member			
LATE PERMIAN				Balaton Red Sandstone Formation	Balaton Red Sandstone Formation	Balaton Red Sandstone Formation	Tabajd Formation / Gymnocodium bellerophontis		

Facies areas				BUDA MOUNTAINS				PILIS MOUNTAINS			HILLS ON THE DANUBE'S LEFT SIDE				BÜKK MOUNTAINS			
Total thickness of the Triassic sequences				> 2690–3200 m				> 1000–1520 m			> 1500 m				~ 3450 m			
LIAS			Tethyan Ammonoid Zones	m		m		m			m		m		m			
TRIASSIC Upper	RHAETIAN		Choristoceras marshi															
			Choristoceras haueri															
	NORIAN	SEVATIAN	Cochloceras suessi															
			Sagenites giebeli															
		ALAUNIAN	Himavatites columbianus															
			Cyrtopleurites bicrenatus															
		LACIAN	Juvavites magnus															
			Malayites paulckei															
			Guembelites jandianus															
	CARNIAN	TUVALIAN	Klamathites macrolobatus															
			Tropites subbullatus															
			Tropites dilleri															
		JULIAN	Sirenites															
			Trachyceras austriacum															
		CORDEVOLIAN	Trachyceras aonoides															
			Trachyceras aon															
			Frankites sutherlandi															
Middle	LADINIAN	LONGOBARDIAN	Protrachyceras archelaus															
		FASSANIAN	Protrachyceras curionii															
			Protrachyceras reitzi															
	ANISIAN	ILLYRIAN	Paraceratites trinodosus															
		PELSONIAN	Balatonites balatonicus															
		BITHYNIAN (ANATOLIAN)	Anagymnotoceras ismidicus															
			Nicomedites osmani															
		AEGEAN (ANATOLIAN)	? Paracrochordiceras anodosum															
Lower	OLENEKIAN	SPATHIAN	Keyserlingites subrobustus															
			Tirolites carniolicus															
			Tirolites cassianus															
		SMITHIAN (NAMMALIAN)	Anasibirites pluriformis															
			Hedenstroemia himalayica															
	INDUAN	DIENERIAN (NAMMALIAN)	Vavilovites markhami															
			Meekoceras kraffti															
		GRIESBACHIAN	Ophiceras connectens															
			Otoceras woodwardi															
LATE PERMIAN																		

BUDA MOUNTAINS

m: ~1000; 540; 1000–1200

m: 500–700; 600; 540; ? ; 1000–1200

Ujlaki hegy dolomite beds

Dachstein Limestone Formation

Haupt-dolomite Formation

Monotis salinaria salinaria, Halorilinoides recliffons, Halorella amphitoma

Triasina hantkeni; Rhabdoceras suessi, Triadomegalodon tofanae, Triadomegalodon damesi, Neomegalodon guembeli, Cladiscites neortus, Worthenia escheri, W. ornata, Purpurodiea excelsior, Hungarielta spinosa, Hungarielta pappi; Clionitites pseudonodosus, Styrites collegialis, Arcestes decipiens, A. tornostomus, Arc. tuhlabiatiformis, Cladiscites pusillus, Megaphyllites applanatus, Plucites myophorus, Pl. placodes, Rhacophyllites negiorensis

Sashegy Dolomite Formation

Cornucardia hornigi, Cuspidaria gladius, Neomegalodon boeckhi n. ssp., N. triqueter pannonicus, N. carinthiacus; Crurealula stachei, Amplicilinodonta suessi, Placites cf. placodes, Anatomites dulcis, Arcestes sp. div., Orthoceras campanile / Koninckina telleri

Lingulae; Daonella cf. aperta

"Mátyáshegy limestone"

Radiolarians, Foraminifers, Ostracods, Echinoderms

Budaörs Dolomite Formation

Foraminifers; Diplopora annulata & Macroporella beneckei

PILIS MOUNTAINS

m: ~280; ~600; >120

Hierlatz L.F.; Crinoids

Dachstein Limestone Formation

Fekete-hegy Formation

Dicerocardium; Helobia cf. norica; Rhabdoceras suessi; Aviculids; Worthenia escheri; Worthenia contabulata; Arcestes bicornis

Hauptdolomit Formation

Cornucardia hornigi, Neomegalodon cf. carinthiacus; Schafhäutlia mellingi

Veszprém Marl Formation

Ostracods; Aviculidae; Posidonomyae

HILLS ON THE DANUBE'S LEFT SIDE

m: ~1000; ~500

m: ~480; ~80; ~580

Dachstein Limestone Formation

Csővár Limestone Formation

"Dachstein Reef Limestone"

Hauptdolomit Form /Vashegy/

"Pokolvölgy dolomite beds"

Dicerocardium hungaricum, Triadomegalodon tophanae, Neomegalodon camptonotus, N. complanatus, N. segestanus, N. boemesi; Clionitites nicetae, Misikella hernsteini

Sageceras haidingeri, Paralobites pisum, Cornucardia hornigi

Metapolygnathus diebeli, Metapolygnathus mostleri

Daonella

BÜKK MOUNTAINS

m: 1500; 450; 400; 250; 50; 200; 600

Felsőtárkány Limestone Formation

Bükkfensik Limestone (Bagolyhegy Quartz-porphyry Mb.); Ohuta Diabase Formation; Formation

Répáshuta Limestone Formation

Bennetitlgy Limestone Formation and tyrrhenis

Monotis salinaria haueri, Met. bidentata, Metapolygnathus posterus, Met. abneptis abneptis; Neomegalodon sp.; C. Laballa suessi, o. Gemnithyris sp., r. Polygyrina aiagana, Naticopsis cf. boemesi, s. Halobia cf. styriaca

South-Bükk Cherty Shale Beds

Radiolarians

Kisgyör Shale Formation; Sericitic; Létrás Diabase M.

Metap. nodosus, M. pseudodiebeli, Gond. naviculis; Anotopteris sp.

Fehérkő Limestone Formation

Gladigondoella tethydis, Gondolella excelsa

Szent-Istvánhegy Porphyrite Formation

Hámor Dolomite Formation: Nyavalyás-tető Limestone Member; Savós Breccia Beds; Gap ?

Vadászvölgy Dolomite Member

Glomospira arcticarosa, Gl. tenuifistulata, Gl. sigmoidalis; Trigodphyllum; Meandrospira dinarica; Nerittaria stanensis

Ablakoskövölgy Formation

Naticella costata; Tirolites cassianus; Costatoria costata; Myophoria cf. laevigata, Unionitites canalensis; Costatoria costata; Claraia cf. aurita; Unionitites fassaënsis; Claraia sp.

Nagyvisnyó Limestone Formation

Gymnoc. bellerophontis, Anchignathodus minutus, Stepanovites dobruskinae; Leptodus nobilis, Tschernyschewia typica

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MICROFLORA OF THE PERMIAN AND LOWER TRIASSIC SEDIMENTS OF THE MECSEK MOUNTAINS (SOUTH HUNGARY)

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On the basis of the microflora-investigations of the 3.4 km thick Permian—Lower Triassic sequence of the Mecsek Mountains the Korpád Sandstone Formation (containing a *Potonieisporites* Association) has been assigned to the Lower Permian, the Boda Aleurolite Formation (including a *Lueckisporites* Association with the norm *Aa* of *Lueckisporites* palynodemes), however, with the underlying Cserdi Conglomerate Formation belonging to the same great cycle, but being unfossiliferous, to the Middle Permian. In the course of this classification the disconformity which separates the Cserdi Conglomerate Formation from the underlying Gyűrűfű Quartz-porphyry Formation, and the Boda Aleurolite Formation from the overlying Kővágószőlős Formation, respectively, has been also regarded. The Kővágószőlős Sandstone Formation including another *Lueckisporites* Association with the norms *Ab*, *Ac*, *B* and *C* of *Lueckisporites* palynodemes belongs to the Upper Permian. In the Palynological assemblage II_4 obtained from the upper part of this formation Lower Triass spores in a quantity of 25% also appear indicating the nearness of the Triassic. The Jakabhegy Sandstone Formation (containing the norm *D* of *Lueckisporites* palynodemes and Lower Triassic spores in its uppermost part) belongs to the lower part of the Lower Triassic. The Patacs Member and the Magyarürög Anhydrite Member (with a *Triadispora* and *Angusticulcites* Association) can be ranged to the upper part of the Lower Triassic.

The Permian/Triassic boundary has been drawn between the Kővágószőlős Sandstone and the Jakabhegy Sandstone Formation, at the diastrophical boundary to be found 40 m higher than the horizon of the Palynological Assemblage II_4.

1. History of Research, Aims

By the palynological investigations in the roughly 3400 m thick Permian and Lower Triassic sequence of the Mecsek Mountains (Fig. 1) consisting prevailingly of clastic rocks the chronostratigraphic classification of the distinguished lithological units and within this also the question of location of the Permian/Triassic boundary discussed for a long time has been largely promoted.

Fig. 1. Geographical position of the Mecsek Mountains

Although a deal of the investigated beds crops out in the perianticline of the western part of the mountains their entire recognition is the result of the drilling researches carried out in the last two decades (Fig. 2). Estimation of the age of their lithostratigraphic units because of poorness of the red sediments, which build up the bigger part of the sequence, in fossils, was subjected to great changes within hundred years (Table 1). For the most part the debates were gone about the stratigraphic position of the 300 m thick Jakabhegy Sandstone Formation. This was placed namely either into the Upper Permian or the Lower Triassic, but its classification, for lack of direct biostratigraphical evidences, was based upon its sedimentological characters or the consideration of its relation to the underlying beds which contain an Upper Permian macroflora. Separated from the underlying Kővágószőlős Sandstone Formation by a low-angle disconformity and coarse grained conglomerate banks, but continued in the overlying Patacs Member without a break, the Jakabhegy Sandstone was regarded already by J. Böckh (1876) as Lower Triassic. On the basis of its petrographic characters, however, it was considered by other scientists as Upper Permian (E. Vadász, 1912, 1917, 1935, 1957; A. Barabás, 1955; A. Barabás—J. Kiss, 1958; A. Barabás—O. Tőzsér, 1961; Á. Barabás-Stuhl, 1962, 1967, 1975; E. Nagy, 1961, 1968). By the drilling research beginning in the sixties it was evidenced the transgression of the Jakabhegy Sandstone had widely gone beyond its normal footwall. This fact and newer petrographical, sedimentological and facial investigations, by which latters the marine character of the formation was made probable (M. Kassai, 1969, 1971, 1976a), lead to connect the beginning of the Triassic transgression (on diastrophical ground) with the base of the Jakabhegy Sandstone. This is how was thought in the last 20 years by the

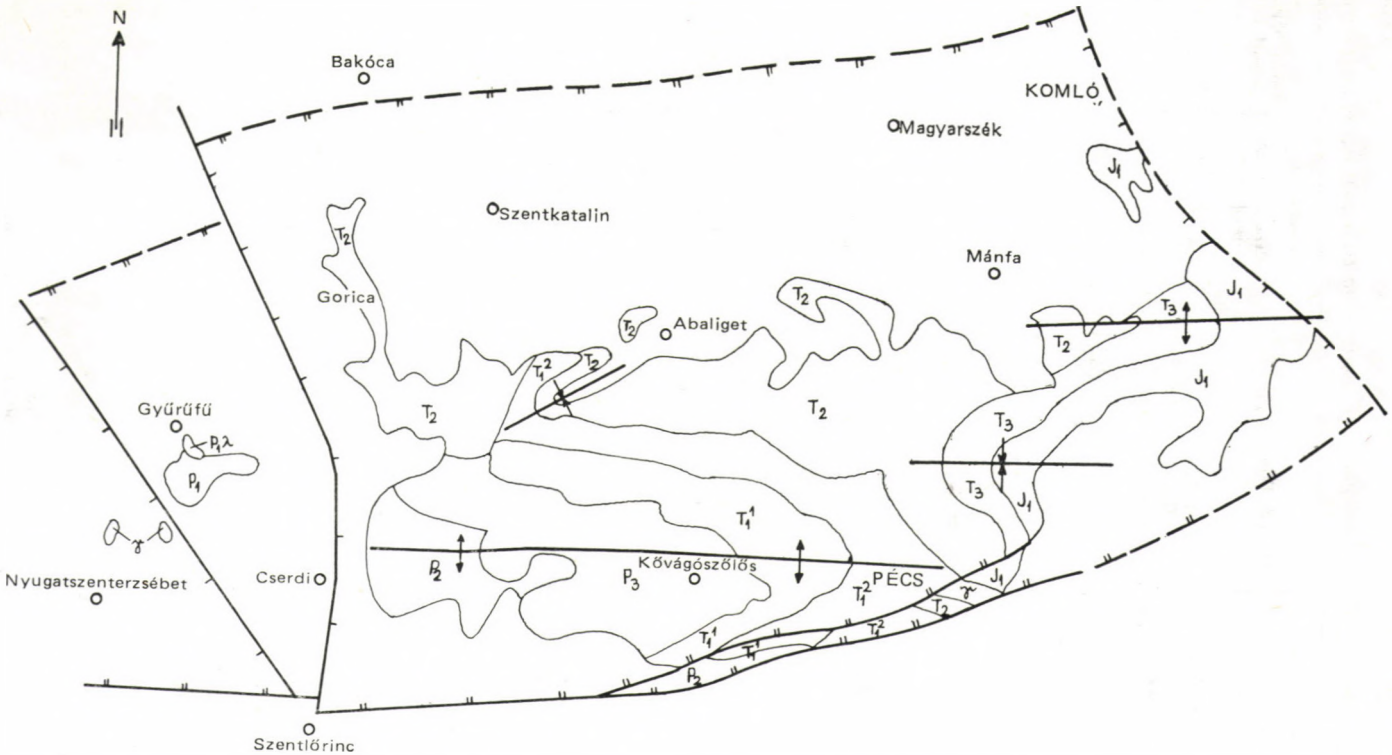

Fig. 2. Geological carte of the western part of the Mecsek Mountains. γ = Granite (Pre-cambrian); P_1 = Korpád Sandstone Formation (Lower Permian); $P_1\lambda$ = Gyűrűfű Quartzporphyry Formation (Lower Permian); P_2 = Cserdi Conglomerate and Boda Aleurolite Formation (Middle Permian); P_3 = Kővágószőlős Sandstone Formation (Upper Permian); T_1^1 = Jakabhegy Sandstone Formation (Lower part of Lower Triassic); T_1^2 = Patacs Member; Magyarürög Anhydrite, Hetvehely Dolomite, Víganvár Limestone Members (Upper part of Lower Triassic); T_2 = Middle Triassic; T_3 = Upper Triassic; J_1 = Lower Jura

Table 1

Development of opinions concerning the age of the Permian

Lithostratigraphy	Böckh, J. 1876	Vadász, E. 1935, 1953	Barabás, A. 1955 Barabás, A.—Kiss, J. 1958	Tőzsér, O. 1961; Kiss, J. 1961
Báránytető Limestone Mb.	Muschelkalk	Anisian		
Vöröshegy Dolomite Member				
Víganvár Limestone Mb.	Early Triassic	Werfenian: Campil beds		
Hetvehely Dolomite Member				
Magyarürög Anhydrite Mb.		Werfenian: Seis beds		
Patacs Member				
Jakabhegy Sandstone Formation		Late Permian		Early Triassic
Kővágószőlős Sandstone Formation	Late Permian	Middle Permian	Middle Permian	Late Permian
Boda Aleurolite Formation	Unknown		Early Permian	Middle Permian
Cserdi Conglomerate Formation				Early Permian
Gyűrűfű Quartz-porphyry Formation			Early Permian	
Korpád Sandstone Formation				
Granite	?	Praecarboniferous	Carboniferous	Early

and Lower Triassic lithostratigraphical units of the Mecsek Mountains

Barabás, A.—Tőzsér, O. 1961; Jámbor, Á. 1964; Nagy, E. 1961, 1968	B. Stuhl, Á. 1962, 1967, 1969, 1975	Kassai, M. 1969, 1971, 1975, 1976a	Balogh, K.—Barabás, A. 1972; Balogh, K.—Barabás, A. — Majoros, Gy. 1973	Kassai, M. 1976b, 1978	Barabás, A. 1977	B. Stuhl, Á. 1979
Anisian				Anisian		Middle Triassic
Campil beds				Campil beds	Lower Triassic	Upper part / Lower Triassic / Upper part
Seis beds		Lower Trias	"Seis beds"	Seis beds		
Late Permian	Late Permian			"Permo-skyth"		Lower p. / Lower part ?
Middle Permian		Upper Permian	"Zechstein"	Upper Permian	Upper Permian	Upper Permian
Early Permian: Upper Rotliegend		Lower Permian	"Rotliegend"		Middle Permian	Middle Permian
Early Permian: Lower Rotliegend		?	? ?		? ? Lower Permian	Lower Permian
		Carboniferous				
Carboniferous [?]		Early Palaeozoic	Early Carboniferous		Praecambrian	

majority of geologists working in this area (O. TŐZSÉR, 1961; J. KISS, 1961; M. KASSAI, 1969; K. BALOGH —A. BARABÁS, 1972; K. BALOGH —A. BARABÁS — GY. MAJOROS, 1977). A further confirmation of this was found by the discovery of *Costatoria costata* of undoubtedly "Campilian" character in the upper part of the type section of Patacs Member that was formerly regarded as of "Seisian age" (Á. JÁMBOR, 1966). Moreover, palynomorphs referring to Triassic were found in the uppermost part of the Jakabhegy Sandstone Formation accompanied by microflora elements being transient from Upper Permian (Á. BARABÁS-STUHL *in* K. BALOGH —A. BARABÁS, 1972; Á. BARABÁS-STUHL, 1975). But the question if the Permian/Triassic boundary should be drawn *within* or *below* the Jakabhegy Sandstone complex, was not decided by this as yet. This uncertainity is reflected in the fact, too, that M. KASSAI (1976b, 1978) considers the Jakabhegy Sandstone Formation in scope of a diastrophical conception, on the basis of comparison with the red sediments of the surrounding areas, as a "Permoscythian" sequence appearing above a regional unconformity and going beyond the underlying formations.

The present writer makes an effort to prove the belonging of the Jakabhegy Sandstone Formation to the Early Triassic by means of palynological methods. In order to a wider supporting of the argumentation, however, all the palynological and biostratigraphical results will be enumerated, which have been obtained in connection with the investigation of the whole Lower Permian —Lower Triassic sequence of Mecsek since the beginning of the sixties.

2. Geological Constitution

The oldest member of the investigated sequences is the Korpád Sandstone Formation resting disconformably on Precambrian granites (Fig. 3). This is built up by clastic rocks of predominantly red colour, but green and grey sandstone beds, too, occur frequently in it. It belongs to the continental fluviale facies (Á. JÁMBOR, 1964) and represents a single great sedimentary cycle. Frequency of burrowing organisms (bioturbation by worms) is a characteristic feature of the reddish brown, siltic, fine grained sandstones which close the cycle.

On the uneven surface of the Korpád Sandstone the Gyűrűfű Quartz-porphyry Formation rests with a low angle disconformity (Á. JÁMBOR, 1964). The quartz-porphyry is a predominantly purple-coloured rock produced by a volcanism of regional extension belonging probably to the Saalian orogenic phase. Above it, likewise disconformably (Á. JÁMBOR, 1964), the coarse grained, red, fluviale sandy sediments of the Cserdi Conglomerate Formation lie enclosing frequent pebbles derived from the underlying quartz-porphyry.

The highest beds of the Cserdi Conglomerate are followed by the variously coloured sandstones of the lower part of the Boda Aleurolite Formation

successed higher up the profile by reddish-brown siltstones (aleurolites). Among the varied sandy beds are present sandy limestone interbeds, too; but in the upper part of the aleurolitic series dolomite beds frequently occur. By virtue of its sedimentological characteristics this formation can be qualified as product of a standing water of beach facies (J. SOMOGYI, 1961), but it is rather of lacustrine than marine origin (Á. JÁMBOR, 1964). The Cserdi Conglomerate and the Boda Aleurolite Formation belong to a single great sedimentary cycle.

In the south side of the mountains varied, coarse grained sandy beds occur in the upper part of the Boda Aleurolite Formation, passing over to the concordantly layered, variegated clastica of the Kővágószőlős Sandstone Formation through progressive coarsening of grains. This latter formation begins with conglomerates.

The sequence of the Kővágószőlős Formation consists of a rhythmic alternation of coarser and finer grained beds. It represents for all that a single great sedimentary cycle beginning with conglomerates. Its middle part consists of relatively finer grained sandstone beds. Higher up the profile grains become coarser again and the uppermost part ends with a pebbled sandstone relating to regression. Within this great cycle four medium-size cycles *(A-B-C-D)*, within the latter, however, eighteen small cycles can be proved by means of statistical grain analysis of the sandstone (Á. BARABÁS-STUHL, 1969, 1973).

On the basis of the dominant colour of the beds the formation can be divided in three heteropical members: the Varied Sandstone, the Gray Sandstone and the Red Sandstone Member.

By virtue of their sedimentological characters the beds of the Kővágószőlős Formation are considered to be settled down in continental (fluvial and paludinal) facies (E. VADÁSZ, 1953; A. BARABÁS, 1964). But, with regard to the dolomite beds and the horizons with dolomitic concretions found in certain parts of the sequence as well as to the findings of *phyllopods* in the Grey Sandstone Member (K. VÁRSZEGI, 1961), M. KASSAI (1976) believes that they may belong to the marine facies.

The Jakabhegy Sandstone Formation overlaps the Kővágószőlős Formation with a low angle disconformity and, going beyond this far away, it can be found in the north foreland of the Mecsek Mountains above Silurian (or Carboniferous?) shales, in its south foreland, however, on Precambrian granites. Its basal part is built up of a characteristic conglomerate containing many quartz-porphyry and quartz pebbles (= Jakabhegy Main Conglomerate). Higher up the profile purple, pebbled sandstone, then purple equigranular, cross-bedded sandstone, finally reddish brown, siltic, fine grained sandstone beds follow. Above the latters the varied sediments of the so called "transitional sandstone" are settled on which the firstly red, then variegated, well bedded, fine grained sandstones, silt- and mudstones of the Patacs Member rest conformably.

In the upper part of the Patacs Member chemical sediments appear in form of thin dolomarl interbeds which become predominant higher up the profile (namely in the Magyarürög Anhydrite, the Hetvehely Dolomite and the Víganvár Limestone Member following each other conformably).

The fauna beginning with the Patacs Member shows unambigously the marine facies of the beds including it. Concerning the facies of the Jakabhegy Formation, as it no fauna includes, one can throw oneself only on sedimentary analyses. Recently it is believed to be of marine facies (M. KASSAI, 1976).

Among the mentioned lithostratigraphical units the beds of the Cserdi Conglomerate Formation, the Hetvehely Dolomite and the Víganvár Limestone Members contain no microflora. The Vöröshegy Dolomite and the Báránytető Limestone resting upon the Víganvár Limestone Member to be ranged on the basis of their fauna already with the Anisian stage of the Middle Triassic (E. NAGY, 1968) proved to be likewise steril palynologically.

3. Macroflora

The poor macroflora collected from the KORPÁD SANDSTONE FORMATION (Fig. 3, 1) has been determined by G. ANDREÁNSZKY (1961) as follows: *Voltzia heterophylla* BRGT. and *Pecopteris* sp. The presence of *Voltzia heterophylla* in these beds is, however, ambigous as this species is characteristic for the higher beds of the Lower Triassic.

Macroflora of greatest quantity was obtained from the grey beds of the KŐVÁGÓSZŐLŐS SANDSTONE FORMATION (Fig. 3, 2). This was determined firstly by O. HEER (1877). Subsequently J. TUZSON (1911) revised this material: *Carpolites geinitzi* HEER, *Ullmannites rhodeanus* GOEPP, *Ullmannites geinitzi* HEER, *Voltzites hungarica* HEER, *Voltzites boeckhiana* HEER. Some stems, found not long ago, were determined by P. SIMONCSICS (1956): *Dadoxylon schrollianum* GOEPP., *Dadoxylon transdanubicum* SIMONCSICS. Also *Baiera-* and *Araucarietites*-stems were described from this material by P. GREGUSS (1967). All this shows Upper Permian in age, moreover, *Baiera* and *Araucarietites* can indicate the end of the Upper Permian, as well, because they were more frequent only from the beginning of the Triassic.

From the varied sandstone beds of the upper part of the JAKABHEGY SANDSTONE *Voltzia heterophylla* BRGT. and *Equisetites* sp. were determined by G. ANDREÁNSZKY (Fig. 3, 3) (Á. JÁMBOR, 1964; K. VÁRSZEGI *in* M. KASSAI, 1976). In other places these forms are to find in the upper part of the Lower Triassic.

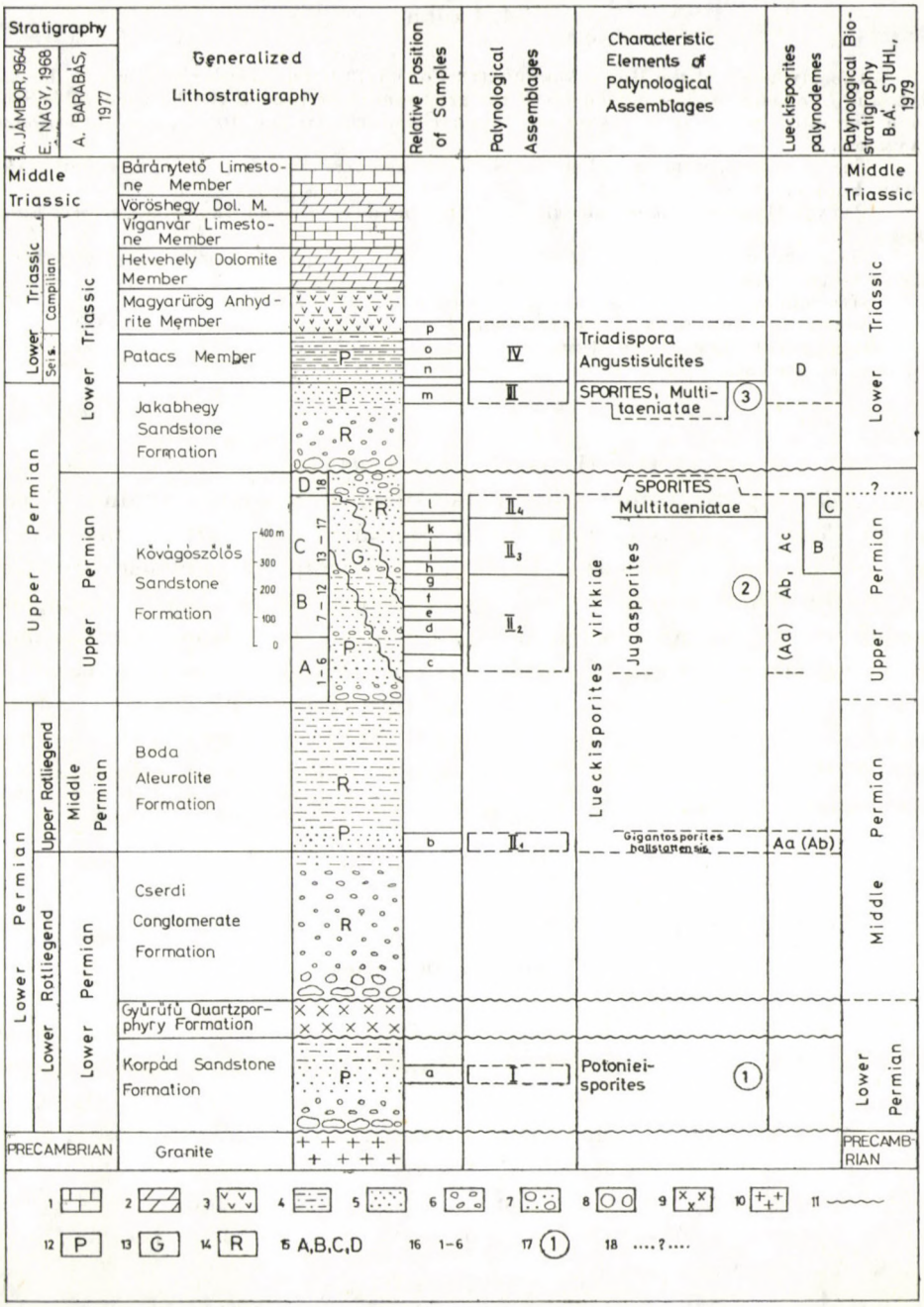

Fig. 3. Biostratigraphical classification of lithostratigraphic units of the Permian and Lower Triassic in Mecsek. 1. Limestone; 2. dolomite; 3. anhydrite and gypsum; 4. mudstone and aleurolite (siltstone); 5. finer grained sandstone; 6. coarser sandstone; 7. pebbled sandstone; 8. conglomerate; 9. quartz-porphyry; 10. granite; 11. disconformity. 12. variegated (red, green, grey) sediment; 13. grey sediments; 14. red sediments; 15. little cycles; 16. small cycles; 17. finding place of macroflora; 18. second possibility for drawing the Permian/Triassic boundary

4. Fauna

The *phyllopods* of the BODA ALEUROLITE FORMATION are: *Lioestheria lallyensis* DEP. et MAZ. and *Limnadia* sp. The *phyllopods* of the grey sandstone beds of the Kővágószőlős Sandstone Formation are after K. VÁRSZEGI (1961): *Euestheria dawsoni* JONES, *Eoleaia leaiformis* RAYMOND.

The most important faunal elements of the members overlying the JAKABHEGY SANDSTONE FORMATION are (E. NAGY, 1968):

PATACS MEMBER: *Isaura albertii* (VOLTZ), *Lingula tenuissima* BRONN, *Costatoria costata* ZENK.

MAGYARÜRÖG ANHYDRITE MEMBER: *Pseudomonotis* sp., *Costatoria* sp., *Velopecten* cf. *albertii* GOLDF., *Lingula tenuissima* BRONN.

HETVEHELY DOLOMITE MEMBER: *Velopecten* cf. *albertii* GOLDF.

VÍGANVÁR LIMESTONE MEMBER: *Modiola triquetra* SEEB., *Costatoria costata* ZENK.

BÁRÁNYTETŐ LIMESTONE MEMBER: *Velopecten reticulatus* SCHL., *Entolium discites* SCHL., *Neritaria* cf. *stanensis* PICHL.

It is noteworthy that the notice of PETERS (1862) that the hanging wall of the Jakabhegy Sandstone — the Patacs Member — would containe "*Posidonomya clarae*" (that is *Claraia clarai*) indicating the start of the Triassic appeared for some authors as an important stratigraphical evidence for rendering the Jakabhegy Sandstone as Upper Permian. But this is very ambiguous because nobody managed to refind this form as yet. In addition, PETERS spoke of embryonary specimens of *Posidonomya clarae* and so it is very reasonable to believe that it was really a *phyllopod*, namely a lot of *phyllopods* was found in this member ever since. *Costatoria costata*, however, to be found in the outcrop of Patacs (Á. JÁMBOR, 1964; E. NAGY, 1968), as well, relates already to the higher part of the Lower Triassic.

5. Palynological Characteristics of the Permian and Lower Triassic Formations in Europe

In Europe, the detritic sediments of the Permian of continental facies do not contain (or only sporadically) animal fossils being available for geological dating. Therefore their stratigraphical classification was taken by virtue of accentuation of the lithological differences and, occasionally, on the basis of their position compared to the fossiliferous beds to be evaluated biostratigraphically. Consequently the Permian/Triassic boundary was usually placed between different lithological units (e.g. Zechstein/Buntsandstein; Magnesian Limestone/Bunter, etc.), and a chronostratigraphic value was ascribed to it. Beginning of the palynological investigation of the Permian and Triassic falls in the fifthies. By the quantitative and qualitative analysis of the rich palynomorph assemblages described since then, the recognition of the Permian—Triassic floras has been largely promoted and the correctness

of marking the boundaries by reason of merely lithological differences has been made questionable in many places.

The main microfloristical features of the intervals within the Permian and Early Triassic, being currently acceptable, are as follows:

The Lower Permian can be characterized by *Potonieisporites* and *Vittatina* assemblages (DOUBINGER, 1974).

VISSCHER (1971, 1974) considers the disaccite pollen *Lueckisporites virkkiae* as the leading form of the Upper Permian. In his opinion, only Lower and Upper Permian exists in Europe, but the Middle Permian is missing. According to him, the deposits considered till now as Middle Permian ("Saxonian", "upper Rotliegend") are heterotopic equivalents of the Lower Permian ("Autunian") or else the Upper Permian ("Thüringian"). He has recognized the microevolutionary mutation of *Lueckisporites virkkiae*, being proceeded in time, determined the *Lueckisporites* palynodemes and distinguished the older and younger parts of the Upper Permian by the help of the presence of it norms *A (Aa, Ab, Ac)*, *B* and *C*.

KOZUR (1977), in turn, has different views as compared to VISSCHER. He came by means of literary data to the conclusion that the *Lueckisporites virkkiae* was not restricted to the Upper Permian in the Euramerican flora province at all, but it was frequent in the whole Middle and Upper Permian. In the Upper Permian the norms *Ab*, *Ac*, *B* and *C* of *Lueckisporites virkkiae* palynodemes, in the Middle Permian, however, the norm *Aa* were predominating. Beside the norm *Aa* of *Lueckisporites* palynodemes the presence of *Corisaccites* and *Hemiapollenites* was characteristic of the Middle Permian. Consequently, according to KOZUR, the Middle Permian is also represented in the Permian sequences in Europe. In his opinion, in case of dividing the Permian in two parts, the Middle and Upper Permian would be united into Upper Permian.

In VISSCHER's view (1971, 1974), the Permian/Triassic boundary can be traced by the sudden disappearence of the Upper Permian *Pinaceae* to be characterized by disaccite pollens *(Ullmannia bronni = Lueckisporites* palynodeme; *Ullmannites frumentaria = Jugasporites* palynodeme; *Pseudovoltzia liebana = Klausipollenites schaubergeri)* and by the prevalence of *Sporophyta*. In the lower part of the Lower Triassic these *Sporophyta* are predominating, but — beside them — the *Pteridospermopsida* to be characterized with multitaeniate pollens play a great role, as well. Their remnants are represented by the *Pleuromeia (Lycopsida)* being relatively frequent in the macroflora and by the rare *Glossopteridium (Pteridospermopsida)* found till now only in Poland.

The quick disappearence of the Upper Permian *Pinaceae* and the contemporary upswing of the *Sporophyta* indicate an important change in the floral evolution that can be explanated only by a great, regional change

in the development of the Earth. This is very well reflected in the composition of the microflora, too, and lends itself particularly well for tracing the boundary between Permian and Triassic.

According to VISSCHER the Lowermost Triassic can be characterized by the significant quantity of the cavate lycophytic microspore, *Lundbladispora nejburgii*, and by the norm *D* of *Lueckisporites* palynodemes *(Stellapollenites* sp.).

In the upper part of the Lower Triassic the *Pinaceae* prosper again; new genera having disaccite pollens *(Voltzia heterophylla = Triadispora, Yuccites vogesiacus = Angustisulcites)* appear. These pollens are very important elements of the palynological assemblages of the upper part of the Lower Triassic and of the Middle Triassic.

KLAUS (1977) distinguished the following microflora associations being very similar to the above mentioned ones:

D: Triadispora Association (upper Bunter)
C: Association with *lycophytic spores* (middle part of the Lower Triassic)
B: Taeniaesporites Association (Lowermost Triassic)
A: Lueckisporites Association (Upper Permian)

Summarized:

- the Lower Permian is characterized by *Potonieisporites* and *Vittatina* associations,
- the Middle and Upper Permian by the *Lueckisporites* association; within this the Middle Permian by the norm *Aa* of *Lueckisporites* palynodemes, the Upper Permian, however, by the norms *(Aa)*, *Ab*, *Ac*, *B* and *C*,
- the lower part of the Lower Triassic by the increased frequency of the *spores* (mainly by that of *Lundbladispora nejburgii)* as well as the *multitaeniate pollens* and by the presence of the norm *D* of *Lueckisporites* palynodemes *(Stellapollenites* sp.),
- finally the upper part of the Lower Triassic by the *Triadispora* and *Angustisulcites* association.

6. The Permian and Lower Triassic Microflora of the Mecsek Mountains

The source of the microflora bearing sample groups is indicated in the Figs 3 and 4 by the letters *a—p*. The sample "*b*" was obtained from a trial trench, all the others derive from boreholes. Their exact position and the symbols of the individual samples are listed in Table 2. The samples of the Kővágószőlős Formation obtained of several boreholes are drawn together on the basis of their stratigraphical position within the small cycles.

One part of the palynological investigations (regarding the samples *c*, *d*, *e*, *f*, *g*, *h*, *i*, *j*, *k* and *m)* was finished already in 1962/63; their results were published in Á. BARABÁS-STUHL, 1975. The remaining samples were investigated only in the last years. By virtue of a particular qualitative and quantitative analysis of the spore and pollen bearing samples, several palynological assemblages were distinguished (Fig. 3). If the foundation of assem-

Table 2

List of the fossiliferous samples

Symbol of Samples	Borehole	Depth, m	Lithostratigraphical Units	Cycle	Relative Position of Samples in Fig. 3
Po 1692	M-13	941.0	Magyarürög Anhydrite Member		*p*
Po 1347	M- 4	647.0			*p*
Po 1687	M- 4	705.0	Patacs Member		*o*
Po 1688	M- 4	730.0			*o*
Po 328	M-14	292.0			*o*
Po 1691	M- 4	798.0			*n*
Po 1697	M-13	1065.0			*n*
Po 1362	M- 4	833.0	Jakabhegy Sandstone Formation		*m*
Po 1491	M-12	1013.0			*m*
Po 161	M-11	451.0			*m*
Po 218	M- 3	240.0			*m*
Po 726	M-10	610.0			*m*
Po 1599	M- 9	1108.0	Kővágószőlős Sandstone Formation	D/C 18/17	*l*
Po 1805	M- 8	1071.0		D/C 18/17	*l*
Po 1810	M- 8	1122.0		C 16	*k*
Po 33	M- 6	697.0		C 16	*k*
Po 21	M- 6	719.0		C 16	*k*
Po 319	M- 7	950.0		C 16	*k*
Po 11	M- 6	727.0		C 15	*j*
Po 6	M- 6	732.0		C 15	*j*
Po 2	M- 6	743.0		C 15	*j*
Po 233	M- 3	713.0		C 14	*i*
Po 235	M- 3	718.0		C 14	*i*
Po 1648	M- 5	1458.0		C 14	*i*
Po 1364	M- 4	1472.0		C 13	*h*
Po 1368	M- 4	1514.0		B 12	*g*
Po 1369	M- 4	1526.0		B 11	*f*
Po 1372	M- 4	1551.0		B 11	*f*
Po 245	M- 3	765.0		B 11	*f*
Po 255	M- 3	800.0		B 10	*e*
Po 257	M- 3	802.0		B 10	*e*
Po 259	M- 3	804.0		B 10	*e*
Po 1374	M- 4	1595.0		B 10	*e*
Po 1633	M- 4	1647.0		B 7	*d*
Po 295	M- 3	948.0		A 4	*c*
Po 300	M- 3	961.0		A 4	*c*
Po 1859	M- 2	trial trench	Boda Aleurolite Formation		*b*
Po 1860	M- 2				*b*
Po 1861	M- 2				*b*
Po 306	M- 1	86.0	Korpád Sandstone Formation		*a*
Po 1216	M- 1	87.0			*a*

Table 3

Percentual Distribution of the Main Palynomorph Groups in the Permian and Lower Triassic of Mecsek

Main groups	Palynological Assemblages						
	I	II_1	II_2	II_3	II_4	III	IV
	Samples						
	a	b	c—g	h—k	l	m	n—p
	%						
Triadispora+*Angustisulcites*	—	—	—	—	—	3	43
Norm *D* of *Lueckisporites*	—	—	—	—	—	8	6
Norm *C* of *Lueckisporites*	—	—	—	—	6	—	—
Norm *B* of *Lueckisporites*	—	—	—	6	12	—	—
Monocolpate alete pollens	—	—	—	+	7	2	2
Norm *A* of *Lueckiporistes*	—	51	30	24	7	—	—
Gigantosporites	—	9	+	+	—	—	—
Klausipollenites schaubergeri	—	2	5	13	4	—	—
Multitaeniate disaccite pollens	—	7	7	15	17	12	6
Jugasporites	3	10	34	24	3	—	—
Potonieisporites	53	—	—	—	—	—	—
Monosaccite pollens	18	3	5	3	10	—	—
Disaccite pollens	23	17	25	17	17	28	41
SPORITES	2	+	1	2	25	47	2
Vittatina	1	1	—	—	—	—	—
Lithostratigraphical units	Korpád Sandstone Formation	Boda Aleurolite Formation	Kővágószőlős Sandstone Formation			Jakabhegy Sandstone Formation	Magyarürög Anhydrite and Patacs Member

blages was based on investigation of a number of samples, those were characterized with the average frequency of the found forms. The microfloras of the distinguished assemblages were divided into groups according to the usual points of view of the palynological stratigraphy (Table 3). On the table the species, genera or groups proved to be of stratigraphical importance according to our own investigations, were also emphasized.

6.1. *Palynological Assemblage I* (Korpád Sandstone Formation; sample group "*a*". Plate I)

The microflora of the oldest beds of the Permian of Mecsek is very poor and bedly preserved. It is dominated by the following *monosaccite pollens: Cordaitina* sp., *Potonieisporites novicus* BHARADWAJ 1954, *Potonieisporites bhardwaji* REMY et REMY 1961, and cf. *Vesicaspora wilsoni* SCHEMEL 1951. These forms together come to 71%; the quota of the genus *Potonieisporites* is 53%. The quantity of the *disaccite pollens* is also of importance: *Pityosporites* sp., *Falcisporites* sp., *Jugasporites omai* HELBY 1966 and other indeterminable disaccite forms represent together 26%. The spores are represented by *Leiotriletes* sp. in a quantity not more than 2% (Table 3).

The Palynological assemblage I is mostly comparable with the assemblages described from the upper part of the Autunian deposits of France and from the Lower Rotliegend of Germany on the basis of the significant quantity of *Potonieisporites* and the relatively high procent of the disaccite pollens (HELBY 1966, DOUBINGER 1974). In opinion of DOUBINGER these assemblages are very similar to the microflora of the Artinskian—Lower Kungurian beds of the marine Permian in Soviet Union. On the other hand, the polyplicate pollen, *Vittatina*, appearing frequently in the comparable assemblages, is very rarely to find in the treated assemblage of Mecsek. This fact can be possibly interpreted by a local facies change.

6.2. *Palynological Assemblage* II_1 (Boda Aleurolite Formation; sample group "*b*". Plate II, III, IV

The very well preserved and rich microflora of the samples gained from the thin, greenish grey, fine grained sandstons intercalated to the red beds of the lower part of the Boda Aleurolite Formation follows in a great stratigraphical distance above the Palynological assemblage I (Figs 3 and 4).

It is dominated by *disaccite pollens* (Table 3). The grains of *Lueckisporites virkkiae* POTONIÉ et KLAUS 1954 found in a significant quantity prevaingly belong to the norm *Aa* of *Lueckisporites* palynodemes, a little quantity of them, however, to the norm *Ab*. The genus *Jugasporites* is represented by *Jugasporites omai* being transient from the Lower Permian, as well as by *Jugasporites delasaucei* (POT. et KLAUS) LESCHIK 1956 and *Jugasporites perspicuus* LESCHIK 1956 in a quantity of 10%; the latters being generally characteristic of VISSCHER's "Upper Permian". The quantity of *Gigantosporites hallstattensis* KLAUS 1963 (9%) is also remarkable. The entire quantity of the *multitaeniate disaccite pollens* (*Taeniasporites labdacus* KLAUS 1963 and *Striatites* sp.) comes to 7%.

Additional *disaccite pollens* are: *Limitisporites leschiki* KLAUS 1963, *Platysaccus papilionis* POT. et KLAUS 1954, *Gardenasporites heisseli* KLAUS 1963, *Falcisporites zapfei* (POT. et KLAUS) LESCHIK 1956, *Paravesicaspora splendens* (LESCHIK) KLAUS 1963, *Klausipollenites schaubergeri* (POT. et KLAUS)

JANSONIUS 1962, *Sulcatisporites* sp. Their proportion taken as a whole comes to 19%.

The *monosaccite pollens* (3%) are represented by *Striatimonosaccites* sp. being characteristic of the Lower Permian and *Nuskoisporites klausi* GREBE 1957 to be referred to the higher part of the Permian.

A representative of the *polyplicate pollens* is *Vittatina costabilis* WILSON 1962. The *spores* are represented by *Converrucosisporites* sp. Both of them were found only sporadically.

The microflora of the Palynological assemblage II_1 — particularly because of the relatively remarkable quantity of *Gigantosporites hallstattensis* — can be very well correlated with the microflora of the Middle Permian Gröden Sandstone of the Southern Alps, which was described by KLAUS, 1963. It is in a good agreement with the microflora determined by VISSCHER et al. (1974) from the beds of the "Faciès de Léouve" (Dom de Barrôt, France), as well. Formerly these ulterior beds were held, on the basis of their lithological aspect and position within the sequence, to be as old as Middle Permian (Saxonian) and only the above-named authors ranged them — because of the presence of the norm *Aa* of *Lueckisporites* — into the lower part of the Upper Permian.

6.3. *Palynological Assemblage* II_2 and II_3 (deeper part of the Kővágószőlős Sandstone Formation; II_2 = samples "*c*—*g*"; II_3 = samples "*h*—*k*". Plate V and VI)

The microflora containing samples of the Kővágószőlős Formation were found in a great stratigraphical distance above the Palynological assemblage II_1.

From among the three palynological assemblages of this Formation (Fig. 3) two ones — namely the Palynological assemblage II_2 and II_3, discussed here — have a very similar composition (Table 3). The main difference between them lies in that the norm *B* of *Lueckisporites* palynodemes was to find only in the younger assemblage (II_3) in a quantity of 6%. Further difference is that the quantity of the *multitaeniate disaccite pollens* is much greater (15%) than in the older assemblage (II_2) where they come to 7% only. Similar difference exists in the quantity of the disaccite pollen *Klausipollenites schaubergeri:* this comes in the assemblage II_2 only to 5%, in the II_3, in turn, to 13%).

A common characteristic of both assemblages is the great quantity of the norm *A* of the *Lueckisporites* palynodemes and the same of the *Jugasporites* genus (34 and 24%, respectively). The latter is represented by *Jugasporites delasaucei* appearing in the Palynological assemblage II_1, as well, and by *Jugasporites schaubergeroides* KLAUS 1963 to be found only in these assemblages.

Within the Palynological assemblage II_2 and II_3 the exact qualitative and quantitative distribution of forms inside the norm *A* of *Lueckisporites* palynodemes could not been with entire objectivity stated as preservation of the pollen grains is rather bad. By virtue of investigation of many samples, however, the subjective statement could be winned that the norms *Ab* and

Exclusively in this assemblage the *alete* pollen, *Pilasporites plurigens* BALME et HENN. 1955, was found in a quantity of 6%. The *monocolpate* pollen, *Cycadopites coxii* is present in 1%.

The pollen forms of this assemblage are known from several Upper Permian Formations (Bellerophon Limestone, Zechstein, etc.). The decided decrease of quantity of the norm *A* of *Lueckisporites* palynodemes, parallel with the increase of quantity of the norm *B* and the appearence of the norm *C* (Table 3) indicate the end of the Permian, the spores occurring simultaneously, however, refer to the beginning or nearness of the Triassic. The spores of this assemblage (without *Lueckisporites)* are significant to the Lower and Middle Buntsandstein in Poland (DÝBOVA—LASZKO, 1978); they are common in the Middle Buntsandstein in Germany (SCHULZ, 1964, 1966; REINHARDT et SCHÖN, 1967). The species of *Lundbladispora* and the cf. microspore of *Pleuromeia* are frequent in the Lower Triassic beds of the "New Red Sandstone" of Ireland (VISSCHER, 1971). In Hungary, the microflora of the uppermost beds of the Upper Permian in the Transdanubian Central Mountains (the so called Tabajd Formation) found in 40 m distance below the Lower Triassic beds containing *Claraia clarai* is also very similar to the Palynological assemblage II_4 as it is characterized with the coexistence of Upper Permian pollens (among which also *Lueckisporites* occurs) and Lower Triassic spores (BARABÁS-STUHL, 1979).

6.5. *Palynological Assemblage III* (Jakabhegy Sandstone Formation; sample group "*m*". Plate XII—XIV)

The microflora of the sample group "*m*" was obtained in a distance of about 300 m above the samples "*l*" of the Palynological assemblage II_4. The quantity of spores is in it still greater (47%) as compared with the assemblage II_4. Considering the qualitative and quantitative composition of the microfloras, there are some similarities, but differences, too, between the two assemblages. The mean difference consists in that the norms *A*, *B* and *C* of *Lueckisporites* palyodemes were not found in this assemblage, the norm *D* (= *Stellapollenites* sp.), in turn, occurred in a quantity of 8%.

Transient spores on the Palynological assemblage II_4 are as follows: **Azonotriletes:** *Calamospora* sp., *Punctatisporites triassicus*, *Leschikisporites aduncus*, *Baculatisporites mesozoicus*. **Zonotriletes:** *Lundbladispora nejburgii* as well as the cf. microspore of *Pleuromeia*. New spores of this assemblage are the azonotrilete *Apiculatisporites plicatus* VISSCHER 1966, *Guttatisporites guttatus* VISSCHER 1966 and *Guttatisporites microechinatus* VISSCHER 1966. From among the spores the genus *Guttatisporites* is of conspicous quantity (29%); the quantity of the other spore-forms ranges from 1 to 4%.

From among the *disaccite* pollens *Limitisporites moersensis* (2%) and cf. *Alisporites ovatus* (8%) were already present in the Palynological assemblage II_4, as well. New *disaccite* forms are: *Voltziaceaesporites heteromorpha* KLAUS 1964 (14%), *Paravesicaspora planderovae* VISSCHER 1966 (4%) and *Triadispora staplini* (JANSONIUS) KLAUS 1964 (only in traces). New representatives of the *multitaeniate disaccite* pollens are: *Taeniaepollenites noviaulensis* LESCHIK 1966 (12%), and *Angusticulcites klausii* (FREUDENTHAL) VISSCHER 1966 (3%). The *monocolpate* pollen grain, *Cycadopites coxii* (2%), was already present in the Upper Permian assemblages, too.

The *spores* of the Palynological assemblage III are known from the lower, middle and upper Buntsandstein of the Germanic facies (REINHARDT, 1964; SCHULZ, 1964, 1966; REINHARDT—SCHMITZ, 1965; VISSCHER, 1966, 1971; REINHARDT—SCHÖN, 1967; DÝBOVA—LASZKO, 1978).

Voltziaceaesporites heteromorpha is known from the lower Buntsandstein up to the Muschelkalk (DÝBOVA—LASZKO, 1978); it was listed by KLAUS (1964) from upper Werfenian beds, too. *Angustisulcites klausii* and *Triadispora staplini* is significant for the upper part of the Lower Triassic (VISSCHER, 1966, 1971; DÝBOVA—LASZKO, 1978; ANTONESCU, 1976). *Taeniaepollenites noviaulensis* was obtained from Upper Permian beds. *Paravesicaspora planderovae* was described by VISSCHER (1966) from the upper Bunter beds. *Stellapollenites* sp. (= *Nuskoisporites muelleri, Triadispora muelleri)* is known upwards from the middle Buntsandstein (REINHARDT, 1964; REINHARDT—SCHMITZ, 1965; SCHULZ, 1965, 1966; VISSCHER, 1966, 1971; ANTONESCU, 1976; DÝBOVA—LASZKO, 1978).

6.6. *Palynological Assemblage IV* (Patacs Member and Magyarürög Anhydrite Member; sample groups "*n—o*" and "*p*". Plate XV—XIX)

Samples "*n*" of this assemblage were obtained 40 m higher than the samples "*m*" of the assemblage III. Here a significant change in the composition of microflora can be experienced. This manifests itself mostly in decrease of quantity of spores and in increase of quantity of pollens. Among the spores and pollens many new forms occur.

From the *spores Calamospora* sp., *Leschikisporites aduncus*, *Guttatisporites guttatus* and *Lundbladispora nejburgii* were already present in the Palynological assemblage II_4, too. In this assemblage *Saturnisporites praevius* VISSCHER 1966, *Cyclotriletes* sp., *Retitriletes jenensis* (REINHARDT—SCHMITZ) SCHULZ 1965, *Concentricisporites pluriannulatus* ANTONESCU 1969 first occur. Their quantity is very low: that of the individual forms ranges from traces up to 4%.

The dominant *disaccite* pollen-form of this assemblage is the genus *Triadispora* (33%); this is represented by *Triadispora crassa* KLAUS 1964, *Triadispora falcata* KLAUS 1964 and the very rare *Triadispora epigona* KLAUS 1964. Among them *Triadispora crassa* is of the greatest quantity. *Voltziaceaesporites heteromorpha* having been present in the former assemblages, as well as *Alisporites grauvogeli* KLAUS 1964 the first appearence of which falls to this assemblage, occur in a major quantity (12% one by one).

The *multitaeniate disaccite pollens* are represented by new forms occurring firstly here *(Taeniaepollenites jonkeri* VISSCHER 1966, *Tubantiapollenites balmei* (KLAUS) VISSCHER 1966, *Striatobietites aytugii* VISSCHER 1966, *Angustisulcites gorpii* VISSCHER 1966) as well as by *Taeniaepollenites noviaulensis* and *Angustisulcites klausii* being present also in the Palynological assemblage III. Their total quantity comes to 16%; within this the genus *Angustisulcites* represents the greatest bulk (10%).

The *remaining disaccite pollens (Limitisporites moersensis*, cf. *Alisporites ovatus* and *Paravesicaspora planderovae)* known also from the former assem-

blage, as well as *Alisporites microreticulatus* REINHARDT 1964, *Colpectopollis ellipsoideus* VISSCHER 1966 and *Chordasporites* sp. (which latters first appear in this assemblage) participate in the composition of the microflora one by one with a quantity from 1 to 8%. The quantity of *Stellapollenites* sp. (= norm *D* of *Lueckisporites* palynodemes) comes to 6%.

This microflora is very well to correlate with the one of the upper part of the Alpine Werfen beds, the upper Buntsandstein, upper Bunter and Röt of the Germanic facies, the "Lower Anisian" of Roumania, the "Muschelkalk" of Southeastern France and the "Keuper" of England (KLAUS, 1964; MÄDLER, 1964, 1968; SCHULZ, 1965, 1966; VISSCHER, 1966, 1971; WARRINGTON, 1974; TAUGOURDEAU—LANTZ, 1974; ANTONESCU, 1976; DÝBOVA—LASZKO, 1978).

The spore species *Concentricisporites nevesi* and *Concentricisporites pluriannulatus* being present in the Palynological assemblage IV were found only in the Triassic of Roumania until now. This fact can relate to the presumable paleogeographical connections between the Mecsek Mountains and some Roumanian areas.

7. Stratigraphical Conclusions

The chronological change of the Permian—Lower Triassic microfloras of Mecsek entirely fits in with the ideas pictured on the page 61 which were established by the modern palynological stratigraphy considering the classification of the coeval sequences of Middle and Western Europe. This fact enables us to state the age of the lithostratigraphical units of the Permian—Lower Triassic of Mecsek as follows (Fig. 4):

(1) The Korpád Sandstone Formation should be assigned, by virtue of the *Potonieisporites* association dominating in the microflora of the Palynological assemblage I and of the great quantity of disaccite pollens, in spite of the very poor representation of *Vittatina*, to the Lower Permian.

(2) The Boda Aleurolite Formation should be considered Middle Permian by reason of the dominating of the norm *Aa* of *Lueckisporites* in the Palynological assemblage II_1 as well as of the presence of a few Lower Permian forms *(Striatimonosaccites, Jugasporites omai)*.

It should be mentioned again that — according to VISSCHER — the *Lueckisporites* association would be significant only for the Upper Permian and within this association the norm *Aa* of *Lueckisporites* would indicate the lowermost part of the Upper Permian. Considering the disconformity being generally present between the Lower Permian and the Permian beds overlying it VISSCHER supposes the Middle Permian deposits to be absent in the continental areas of Europe. A disconformity appears in the Mecsek Mts., too, at the lower and upper boundary of the Gyűrűfű Quartz-porphyry Formation. Though the dimensions of this break are unknown, the divisibility of the very thick clastic sequence, settled down after this gap, into three great cycles, makes probable also the presence of the Middle Permian in the Mecsek beside the Upper Permian.

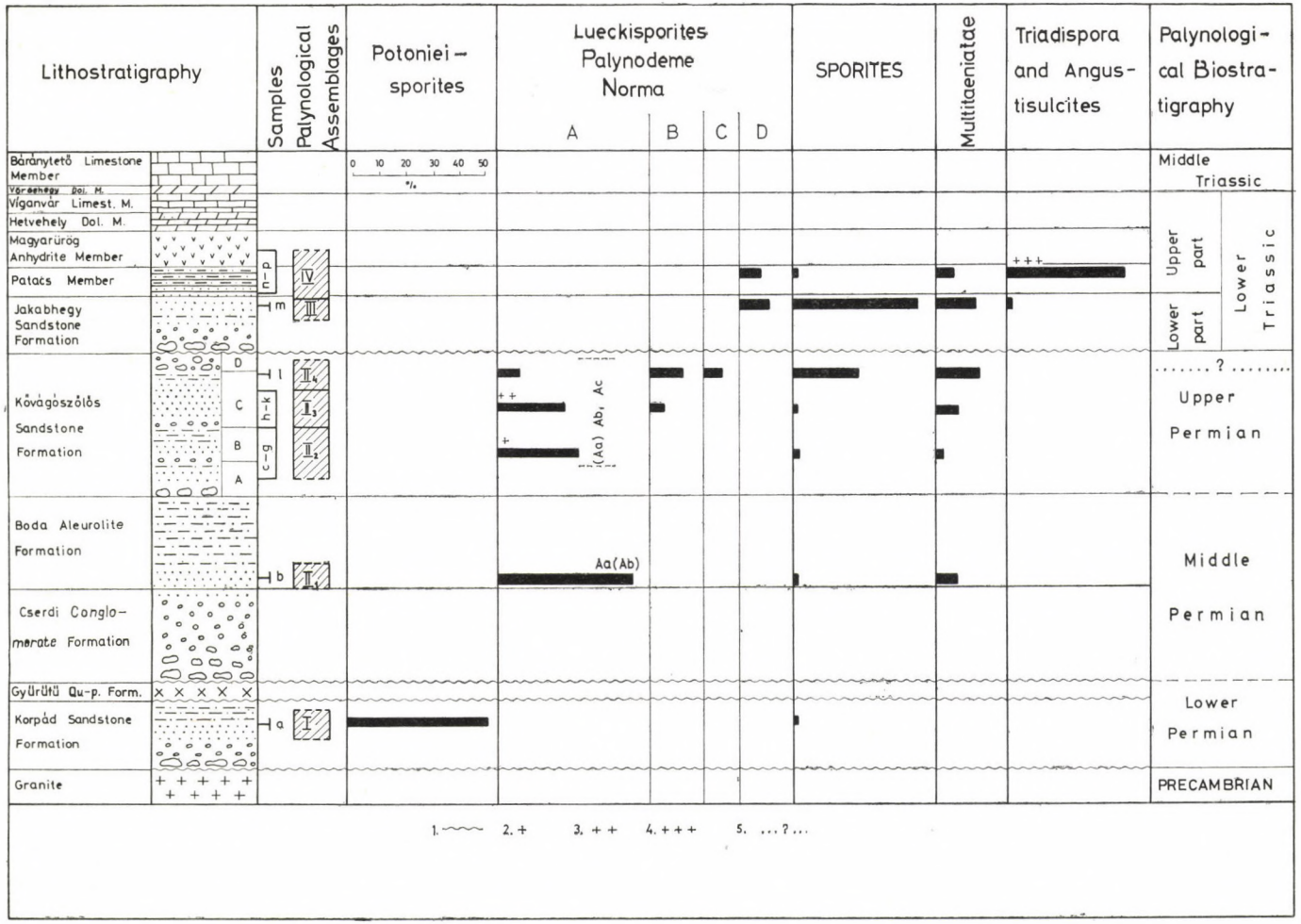

Fig. 4. Distribution of the characteristic forms of Permian and Lower Triassic microflora of Mecsek in the palynological assemblages. 1. unconformity; 2. average of five sample groups; 3. average of four sample groups; 4. average of three sample groups; 5. second possibility for drawing the Permian/Triassic boundary

Ac are more frequent, while the norm *Aa* of *Lueckisporites* palynodemes is very rare in these assemblages. In the Palynological assemblage II_2 the norm *Ab*, in the Palynological assemblage II_3, however, the norm *Ac* appears more frequently.

The *multitaeniate disaccite pollens* are represented by *Taeniaesporites ortosei* KLAUS 1963 and *Taeniaesporites labdacus* KLAUS 1963 as well as *Striatites jacobii* JANSONIUS 1962. But this latter is very rare in both assemblages.

Others disaccite pollen grains of the Palynological assemblage II_2 and II_3 are *Limitisporites leschiki*, *Platysaccus papilionis*, *Gardenasporites heisseli* and *Falcisporites zapfei* that were already present in the Palynological assemblage II_1, too. The quantity of the different species varies between 1 and 10%. The form cf. *Alisporites ovatus* JANSONIUS 1962 appears in the uppermost sample of the Palynological assemblage II_3 in quantity of 1%.

The *monosaccite pollens* are represented by *Nuskoisporites dulhuntyi* POT. et KLAUS 1954 (3—5%), the spores to be found very rarely, in turn, by *Converrucosisporites eggeri* KLAUS 1963 and *Calamospora* sp. In the uppermost sample of the Palynological assemblage II_3 occurs the *monocolpate pollen* grain, *Cycadopites coxii* VISSCHER 1966. The quantity of both the spores and the monocolpate pollen grains is very small (about 1%).

The composition of the Palynological assemblages II_2 and II_3 is in a good accordance with that of the microflora of the Upper Permian Formations of the Alpine and Germanic facies in Europe (Bellerophon Limestone, Magnesian Limestone, Kingscourt Gypsum Formation, Zechstein). In general, the whole Euramerican flora realm, well known in palynological respects, is characterized by similar assemblages. In Hungary, the Upper Permian beds of the Transdanubian Central Mountains (the Balaton Red Sandstone and the Tabajd Formation, respectively) contain similar microfloras, as well (KLAUS, 1953, 1954, 1963; POTONIÉ et KLAUS, 1954; LESCHIK, 1956; GREBE, 1957; SCHWEITZER, 1960; ČORNA—ILAVSKA, 1962; STUHL, 1962; CLARKE, 1965; KALIBOVA, 1967; VISSCHER, 1968, 1971, 1974; PLANDEROVA, 1973; DÝBOVA, 1974; CLEMENT, 1974; DÝBOVA—LASZKO, 1978).

6.4. *Palynological Assemblage II_4* (Kővágószőlős Sandstone Formation; sample group "*l*". Plate VII to XI)

The samples "*l*" derived from beds lying in a distance of 50—60 m above the uppermost sample of the Palynological assemblage II_3 show a sudden and strong change of the microflora. Though the disaccite pollens of the older assemblages being characteristic of the Upper Permian (and among these the norm *A* and *B* of *Lueckisporites* palynodemes, as well) appear still in them, but, beside the latters, many new forms (among them Lower Triassic spores, too) occur and become quantitatively significant. Whilst the amount of spores ranged only from 1 to 2% in the microflora of the older assemblages, their quantity increases in this assemblage as far as 25%.

The important change occurring in the microflora of the Palynological assemblage II_4 can be joined with a geological change, too. Namely the samples

"*l*" containing this palynological assemblage are derived from the highest, fine-grained beds of the medium-size cycle "*C*", and the coarse-grained, pebbly sandstone beds of the newer medium-size cycle "*D*" rest above them as regressive closing members of the Kővágószőlős Sandstone Formation (Fig. 3).

Among the *disaccite pollens* of the Palynological assemblage II_4 being characteristic to the Upper Permian, the quantity of the norm *A* of *Lueckisporites* palynodemes comes to 7%; this is far less as compared with the older assemblages; the same of the norm *B*, in turn, is increased to 12%. The norm *A* is represented by forms belonging to the norms *Ab* and *Ac;* between these the norm *Ac* dominates. The presence of *Guttulapollenites* sp. is also very important in this assemblage. This form is identical with the norm *C* of *Lueckisporites* palynodemes that indicates, in VISSCHER's (1971, 1974) opinion, the Latest Permian. Its quantity is 6%.

The total quantity of the *multitaeniate disaccite pollens* (17%) deserves attention. *Taeniaesporites ortosei*, *Taeniaesporites labdacus*, *Striatites jacobii* to be found also here were present already in the older assemblages, too; but *Strotersporites* sp. appears only in this assemblage.

Further *disaccite pollens* of the assemblage are: *Jugasporites schaubergeroides*, *Gardenasporites heisseli*, *Platysaccus papilionis*, *Falcisporites zapfei*, *Klausipollenites schaubergeri*, cf. *Alisporites ovatus* known also from the assemblages II_1, II_2 and II_3. The quantity of the individual forms comes, however, to not more than 1 to 4%.

Beside the "Upper Permian" *monosaccite pollen Nuskoisporites dulhuntyi* — which is present, however, also in the assemblage II_2 and II_3 — three new *monosaccite* forms *(Endosporites* sp., *Monosaccites I* and *Monosaccites II)* appear. The latters bear a resemblance mostly to certain Goudwana-forms. The total quantity of the monosaccite forms (10%) is much greater than the same in the earlier assemblages (II_1, II_2, II_3).

In the Palynological assemblage II_4 the quantity of *spores* increases abruptly as compared to the earlier assemblages. Beside *Calamospora* sp. and *Converrucosisporites* sp. known also from the older assemblages many new spore genera and species occur. The followings can be listed: **Azonotriletes:** *Punctatisporites triassicus* SCHULZ 1965, *Punctatisporites velatus* LESCHIK 1955, *Leschikisporites aduncus* (LESCHIK) POTONIÉ 1958, *Leiotriletes* sp., *Leiotriletes adrienneforma* NILSSON 1958, *Conosmundasporites* sp., *Laevigatisporites* sp., *Baculatisporites mesozoicus* SCHULZ 1964, *Verrucosisporites morulae* KLAUS 1960, *Verrucosisporites* sp. **Zonotriletes:** *Lundbladispora nejburgii* SCHULZ 1964, *Lundbladispora echinata* REINHARDT et SCHÖN 1967. **Monoletes:** *Laevigatosporites* sp., cf. microspore of *Pleuromeia*. But neither form plays a significant role; the quantity of the individual ones ranges from trace to 4%. The significant microspore of the lower part of the Lower Triassic, *Lundbladispora nejburgii* participates in the assemblage only with 1%.

As the presence of the Middle Permian was to be confirmed biostratigraphically only in a thin bed of the lower part of the Boda Aleurolite, one must lean at tracing its lower and upper boundary on diastrophical data.

As the Cserdi Conglomerate, lying above the disconformity connected with the Gyűrűfű Quartz-porphyry, represents with the Boda Aleurolite Formation together a single great sedimentary cycle, both formations should be assigned to the Middle Permian. The belonging of the Cserdi Conglomerate to the Middle Permian is supported by the fact that *Lueckisporites* is present in a great strength in the Palynological assemblage II_1 (51%); consequently its virency may be started already in time of sedimentation of the beds underlying it.

(3) The Kővágószőlős Sandstone Formation belongs by virtue of the norms *Ab*, *Ac*, *B* and *C* of *Lueckisporites* being dominant in the Palynological assemblage II_2, II_3 and II_4 to the Upper Permian.

The spore—pollen spectrum of the Palynological assemblage II_4 obtained from the upper part of the formation indicates the end or the nearness of the Triassic on the basis of appearence of *Guttulapollenites* sp. (= norm *C* of *Lueckisporites*) as well as the great quantity of Lower Triassic spores (among them *Lundbladispora nejburgii*, as well) and multitaeniate pollens.

(4) By lack of the norms *A*, *B*, *C* of *Lueckisporites*, by presence of the norm *D* and by the very high amount of spores, the belonging of the uppermost part of the Jakabhegy Sandstone Formation to the lower part of the Lower Triassic has been proved. The small quantity of *Triadispora* and *Angustisulcites*, however, indicates the nearness of the upper part of the Lower Triassic. This implies that on the top of the Jakabhegy Sandstone Formation younger beds of the lower part of the Lower Triassic are already present. The same results from the fact, that *Lundbladispora nejburgii*, dominating — in accordance with literary data — during the early Lower Triassic, appears in the Palynological assemblage III only in a very small quantity. Consequently the virency of this lycophytic microspore, observed in the early Lower Triassic of other areas, falls in the Mecsek Mountains onto the time-span that was filled between the creation of the Palynological assemblage II_4 and III by the palynologically sterile deposition of the 40 m thick, purple-pebbled conglomerate belonging to the cycle "*D*" of the Kővágószőlős Formation and of the 300 m thick Jakabhegy Sandstone.

(5) The Patacs and the Magyrürög Anhydrite Members belong by their *Triadispora* and *Angustisulcites* associations of the microflora of the Palynological assemblage IV unambigously to the upper part of the Lower Triassic. This classification is affirmed also by the appearence of *Costatoria costata* in these members.

(6) According to the above the Permian/Triassic boundary may be placed *either* on to the base of the coarse initial ("main conglomerate") beds

of the Jakabhegy Sandstone Formation *or* onto the sole of the "purple-pebbled conglomerate" filling the cycle "*D*" of the Kővágószőlős Sandstone Formation. Namely, a diastrophical change being suitable — also for lack of fossils — for separation of two geochronological periods appears only at these horizons within the whole sequence. The tracing such a boundary by virtue of diastrophism is supported by a sedimentation's break known through wide areas of the Tethys realm and attached in all places to the Permian/Triassic boundary. No doubt that the greater change is indicated by the "main conglomerate" of the Jakabhegy Sandstone; therefore the base of it is here proposed for fixing the Permian/Triassic boundary.

On the basis of composition of the Palynological assemblages II_4 and III, the possibility of an Upper Permian age of the Jakabhegy Sandstone is entirely eliminated, and further discussions may be emerged, if only, in respect of the classification of the cycle "*D*" of the Kővágószőlős Sandstone Formation.

(7) The boundary between the Lower and Middle Triassic can not be palynologically marked because no microflora was found in the beds above the Palynological assemblage IV as yet. In Figs 3 and 4, the Lower/Middle Triassic boundary has been drawn by virtue of faunistical data, in perfect harmony with E. NAGY (1968).

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МИКРОФЛОРА ВЕРХНЕПЕРМСКИХ И НИЖНЕТРИАСОВЫХ ОТЛОЖЕНИЙ ГОР МЕЧЕК (ЮЖНАЯ ВЕНГРИЯ)

А. БАРАБАШ-ШТУЛ

Резюме

На основании изучения микрофлоры пермско-нижнетриасовых отложений мощности 3,4 км в горах Мечек Корпадская Песчаниковая Формация (ассоциация с *Potonieisporites)* автором относится к нижней перми, Бодайская Алевролитовая Формация (ассоциация с *Lueckisporites*, *Lueckisporites* palynodeme norma *Aa)* вместе с подстилающей ее Чердской Конгломератовой формацией, относящейся к тому же самому мегациклу, но не содержащей ископаемых органических остатков — к средней перми. При этом она приняла во внимание и несогласие, отделяющее Чердскую Конгломератовую Формацию от нижележащей Дюрюфюской Кварцево-порфировой Формации и Бодайскую Алевролитовую Формацию от покрывающей ее Кёвагосёллёшской Песчаниковой Формации. Кёвагосёлёшская Песчаниковая Формация (ассоциация с *Lueckisporites*, *Lueckisporites* palynodeme norma *Ab*, *Ac*, *B* и *C*) относится к верхней перми. В палинологической ассоциации II_4, происходящей из верхней части формации, наряду с верхнепермскими пыльцевыми зернами появляются — 25%-овом количестве — уже и нижнетриасовые споры, указывая на близость триаса. Якабхедьская Формация *(Lueckisporites* palynodeme norma *D* и нижнетриасовые споры) относится к нижней части нижнего триаса. Патачское Подразделение и Мадьярюгёдское Ангидритовое. Подразделение (ассоциации с *Triadispora* и *Angustisulcites)* могут быть отнесены к верхней части нижнего триаса.

Граница между перью и триасом автор проводит на диастрофической границе, прослеживаемой в 40 м над сообществом II_4, то есть между Кёвагосёлёшской Песчаниковой и Якабхедьской Песчаниковой Формациями.

Plate I

Palynological assemblage I

Fig. 1—2. *Cordaitina* sp.
Fig. 3. *Potonieisporites novicus* BHARADWAJ, 1954
Fig. 4—5. *Potonieisporites bhardwaji* REMY et REMY, 1961
Fig. 6. cf. *Vesicaspora wilsoni* SCHEMEL, 1951
Fig. 7. *Jugasporites omai* HELBY, 1966
Fig. 8. *Pityosporites* sp.
Fig. 9. *Falcisporites* sp.
Fig. 10. *Disaccites indet.*
Fig. 11. *Leiotriletes* sp.

Plate II

Palynological assemblage II_1

Fig. 1. *Converrucosisporites* sp.
Fig. 2. *Nuskoisporites klausi* GREBE, 1957
Fig. 3. *Striatimonosaccites* sp.
Fig. 4. *Jugasporites omai* HELBY, 1966
Fig. 5. *Jugasporites delasaucei* (POT. et KLAUS) LESCHIK, 1956
Fig. 6. *Jugasporites perspicuus* LESCHIK, 1956
Fig. 7. *Limitisporites leschiki* KLAUS, 1963

Plate III

Palynological assemblage II_1

Fig. 1. *Platysaccus papilionis* POT. et KLAUS, 1954
Fig. 2. *Gigantosporites hallstattensis* KLAUS, 1963
Fig. 3. *Gardenasporites heisseli* KLAUS, 1963
Fig. 4—8. *Lueckisporites virkkiae* POT. et KLAUS, 1954
(Lueckisporites palynodeme norma A, VISSCHER, 1971)
Fig. 9—10. *Taeniaesporites labdacus* KLAUS, 1963

Plate IV

Palynological assemblage II_1

Fig. 1. *Striatites* sp.
Figs 2—3. *Paravesicaspora splendens* (LESCHIK) KLAUS, 1963
Fig. 4. *Sulcatisporites* sp.
Fig. 5. *Klausipollenites schaubergeri* (POT. et KLAUS) JANSONIUS, 1962
Figs 6—7. *Falcisporites zapfei* (POT. et KLAUS) LESCHIK, 1956
Fig. 8. *Vittatina costabilis* WILSON, 1962
Fig. 9. *Ephedripites primus* KLAUS, 1963

Plate V

Palynological assemblage II_2—II_3

Fig. 1. *Converrucosisporites eggeri* KLAUS, 1963
Figs 2—3. *Nuskoisporites dulhuntyi* POT. et KLAUS, 1954
Fig. 4. *Jugasporites delasaucei* (POT. et KLAUS) LESCHIK, 1956
Fig. 5. *Jugasporites schaubergeroides* KLAUS, 1963
Fig. 6. *Limitisporites leschiki* KLAUS, 1963
Fig. 7. *Platysaccus papilionis* POT. et KLAUS, 1954
Fig. 8. *Gardenasporites heisseli* KLAUS, 1963
Figs 9—11. *Lueckisporites virkkiae* POT. et KLAUS, 1954 *(Lueckisporites palynodeme* norma A)

Plate VI

Palynological assemblage II_2-II_3

Figs 1—3. *Lueckisporites virkkiae* POT. et KLAUS, 1954 *(Lueckisporites palynodeme norma A)*
Fig. 4. *Lueckisporites virkkiae* POT. et KLAUS, 1954 *(Lueckisporites palynodeme norma B)*
Fig. 5. *Taeniaesporites labdacus* KLAUS, 1963
Fig. 6. *Taeniaesporites ortosei* KLAUS, 1963
Fig. 7. *Striatites jacobii* JANSONIUS, 1962
Fig. 8. *Falcisporites zapfei* (POT. et KLAUS) LESCHIK, 1956
Fig. 9. *Klausipollenites schaubergeri* (POT. et KLAUS) JANSONIUS, 1962
Fig. 10. cf. *Alisporites ovatus* JANSONIUS, 1962
Fig. 11. *Cycadopites coxii* VISSCHER, 1966

Plate VII

Palynological assemblage II_4

Fig. 1. *Calamospora* sp.
Fig. 2. *Leiotriletes adrienneforma* NILSSON, 1958
Fig. 3. *Leiotriletes* sp.
Fig. 4. *Punctatisporites triassicus* SCHULZ, 1965
Fig. 5. *Punctatisporites velatus* LESCHIK, 1955
Fig. 6. *Converrucosisporites* sp.
Fig. 7. *Conosmundasporites* sp.
Fig. 8. *Laevigatisporites* sp.
Fig. 9. *Verrucosisporites morulae* KLAUS, 1960
Fig. 10. *Baculatisporites mesozoicus* SCHULZ, 1964

Plate VIII

Palynological assemblage II_4

Fig. 1. *Verrucosisporites* sp.
Fig. 2. *Leschikisporites aduncus* (LESCHIK) POTONIÉ, 1958
Fig. 3. *Lundbladispora echinata* REINHARDT und SCHÖN, 1967
Figs 4—5. *Lundbladispora nejburgii* SCHULZ, 1964
Fig. 6. *Laevigatosporites* sp.
Fig. 7. cf. *microspora* of *Pleuromeia*
Fig. 8. *Endosporites* sp.
Figs 9—10. *Nuskoisporites dulhuntyi* POT. et KLAUS, 1954

Plate IX

Palynological assemblage II_4

Fig. 1. MONOSACCITES I.
Fig. 2. MONOSACCITES II.
Fig. 3. *Endosporites* sp.
Fig. 4. *Jugasporites schaubergeroides* KLAUS, 1963
Fig. 5. *Limitisporites moersensis* (GREBE) KLAUS, 1963
Fig. 6. *Platysaccus papilionis* POT. et KLAUS, 1954
Fig. 7. *Gardenasporites heisseli* KLAUS, 1963
Figs 8—9. *Lueckisporites virkkiae* POT. et KLAUS, 1954 *(Lueckisporites palynodeme norma A)*

Plate X

Palynological assemblage II_4

Figs 1—3. *Lueckisporites virkkiae* POT. et KLAUS, 1954 *(Lueckisporites palynodeme norma A)*
Figs 4—6. *Lueckisporites virkkiae* POT. et KLAUS, 1954 *(Lueckisporites palynodeme norma B)*
Figs 7—9. *Guttulapollenites* sp. *(Lueckisporites palynodeme norma C)*
Fig. 10. *Taeniaesporites ortosei* KLAUS, 1963
Fig. 11. *Taeniaesporites labdacus* KLAUS, 1963
Fig. 12. *Strotersporites* sp.
Fig. 13. *Striatites* sp.

Plate XI

Palynological assemblage II_4

Fig. 1. *Striatites jacobii* JANSONIUS, 1962
Fig. 2. *Sulcatisporites* sp.
Fig. 3. cf. *Alisporites ovatus* JANSONIUS, 1962
Fig. 4. *Falcisporites zapfei* (POT. et KLAUS) LESCHIK, 1956
Fig. 5. *Klausipollenites schaubergeri* (POT. et KLAUS) JANSONIUS, 1962
Fig. 6. *Pilasporites plurigens* BALME et HENN, 1955
Fig. 7. *Cycadopites coxii* VISSCHER, 1966
Figs 8—9. *Tracheide*

Plate XII

Palynological assemblage III

Fig. 1. *Calamospora* sp.
Fig. 2. *Punctatisporites triassicus* SCHULZ, 1965
Fig. 3. *Punctatisporites* sp.
Fig. 4. *Apiculatisporites plicatus* VISSCHER, 1966
Fig. 5. *Baculatisporites mesozoicus* SCHULZ, 1964
Figs 6—7. *Guttatisporites guttatus* VISSCHER, 1966

Plate XIII

Palynological assemblage III

Figs 1—2. *Guttatisporites microechinatus* VISSCHER, 1966
Fig. 3. *Punctatisporites* sp.
Fig. 4. *Leschikisporis aduncus* (LESCHIK) POTONIE, 1958
Fig. 5. *Lundbladispora nejburgii* SCHULZ, 1964
Fig. 6. *Triadispora staplini* KLAUS, 1964
Fig. 7. *Limitisporites moersensis* (GREBE) KLAUS, 1963
Fig. 8. *Stellapollenites* sp. *(Lueckisporites palynodeme norma D)*

Plate XIV

Palynological assemblage III

Fig. 1. *Taeniaepollenites noviaulensis* LESCHIK, 1966
Fig. 2. *Angustisulcites klausii* (FREUDENTHAL) VISSCHER, 1966
Fig. 3. *Paravesicaspora planderovae* VISSCHER, 1966
Fig. 4. *Voltziaceaesporites heteromorpha* KLAUS, 1964
Fig. 5. cf. *Alisporites ovatus* JANSONIUS, 1962
Fig. 6. *Cycadopites coxii* VISSCHER, 1966

Plate XV

Palynological assemblage IV

Fig. 1. *Calamospora* sp.
Fig. 2. *Cyclotriletes*
Fig. 3. *Guttatisporites guttatus* VISSCHER, 1966
Fig. 4. *Retitriletes jenensis* (REINH. et SCHMITZ) SCHULZ, 1965
Fig. 5. *Concentricisporites pluriannulatus* ANTONESCU, 1969
Fig. 6. *Concentricisporites nevesi* ANTONESCU, 1970

Plate XVI

Palynological assemblage IV

Fig. 1. *Lundbladispora nejburgii* SCHULZ, 1964
Fig. 2. *Saturnisporites praevius* VISSCHER, 1966
Fig. 3. *Leschikisporites aduncus* (LESCHIK) POTONIÉ, 1958
Fig. 4. *Triadispora staplini* KLAUS, 1964

Figs 5—8. *Triadispora crassa* KLAUS, 1964
Fig. 9. *Triadispora falcata* KLAUS, 1964
Fig. 10. *Triadispora epigona* KLAUS, 1964
Fig. 11. *Triadispora* tetrad

Plate XVII

Palynological assemblage IV

Figs 1—3. *Stellapollenites* sp. *(Lueckisporites palynodeme norma D)*
Fig. 4. *Taeniaepollenites noviaulensis* LESCHIK, 1966
Figs 5—6. *Taeniaepollenites jonkeri* VISSCHER, 1966

Plate XVIII

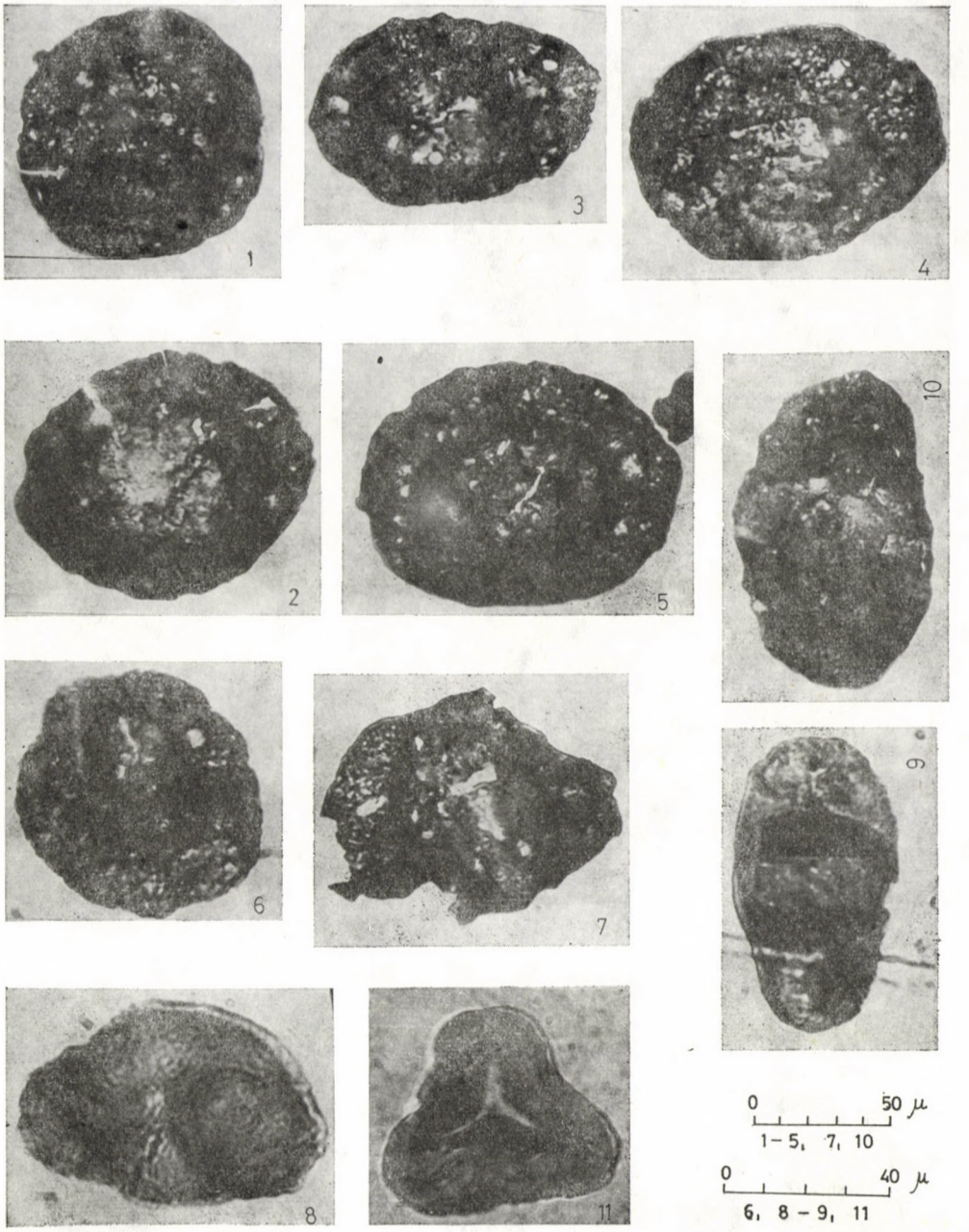
Palynological assemblage IV

Fig. 1. *Angustisulcites klausii* (FREUDENTHAL) VISSCHER, 1966
Figs 2—3. *Angustisulcites gorpii* VISSCHER, 1966
Fig. 4. *Striatoabietites aytugii* VISSCHER, 1966
Fig. 5. *Tubantiapollenites balmei* (KLAUS) VISSCHER, 1966
Fig. 6. cf. *Alisporites ovatus* JANSONIUS, 1962
Fig. 7. *Alisporites grauvogeli* KLAUS, 1964

Plate XIX

Palynological assemblage IV

Fig. 1. *Alisporites grauvogeli* KLAUS, 1964
Fig. 2. *Alisporites microechinatus* REIHHARDT, 1964
Figs 3—4. *Voltziaceaesporites heteromorpha* KLAUS, 1964
Figs 5—6. *Colpectopollis ellipsoideus* VISSCHER, 1966
Fig. 7. *Chordasporites* sp.
Fig. 8. *Cycadopites coxii* VISSCHER, 1966



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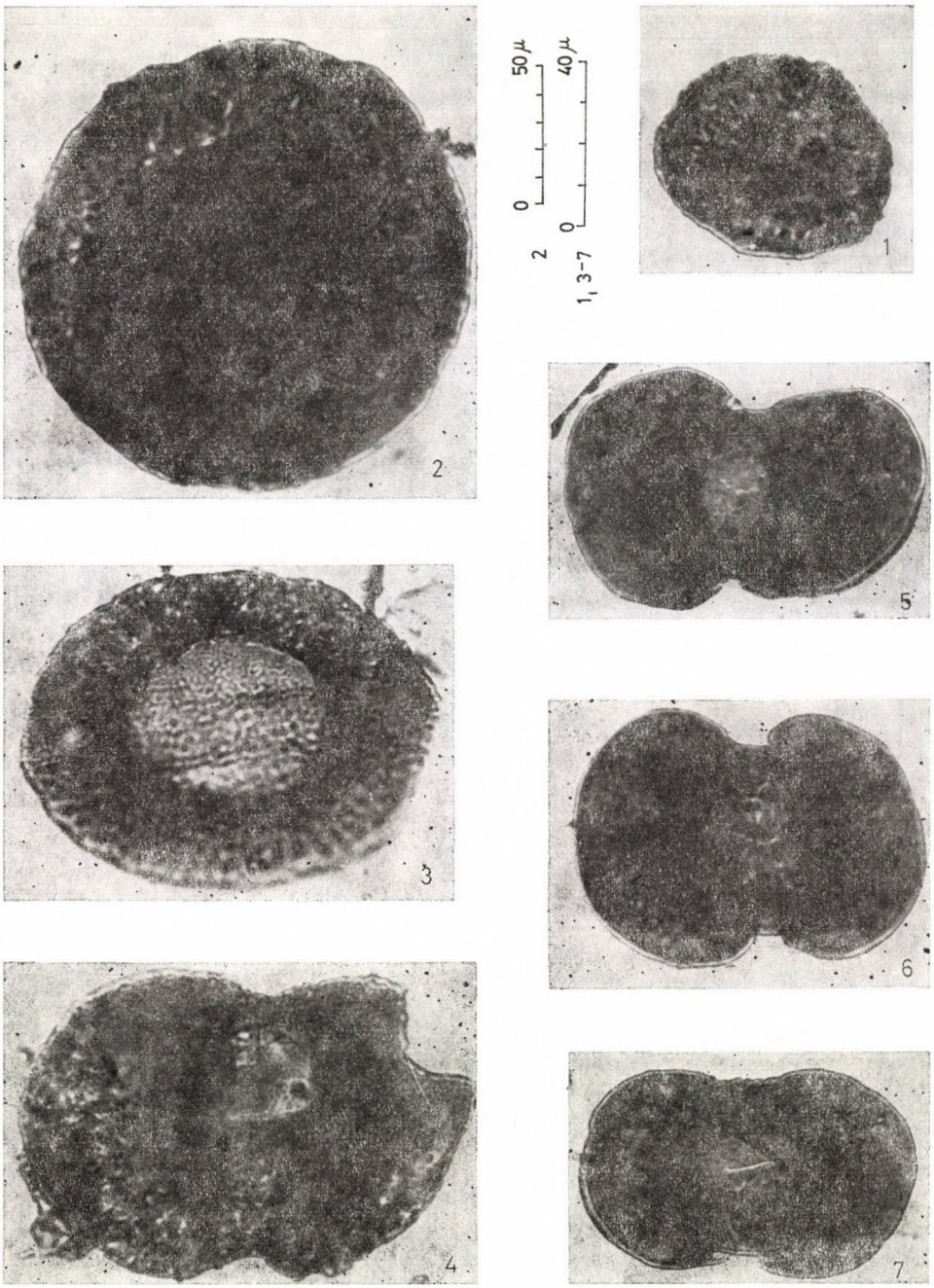

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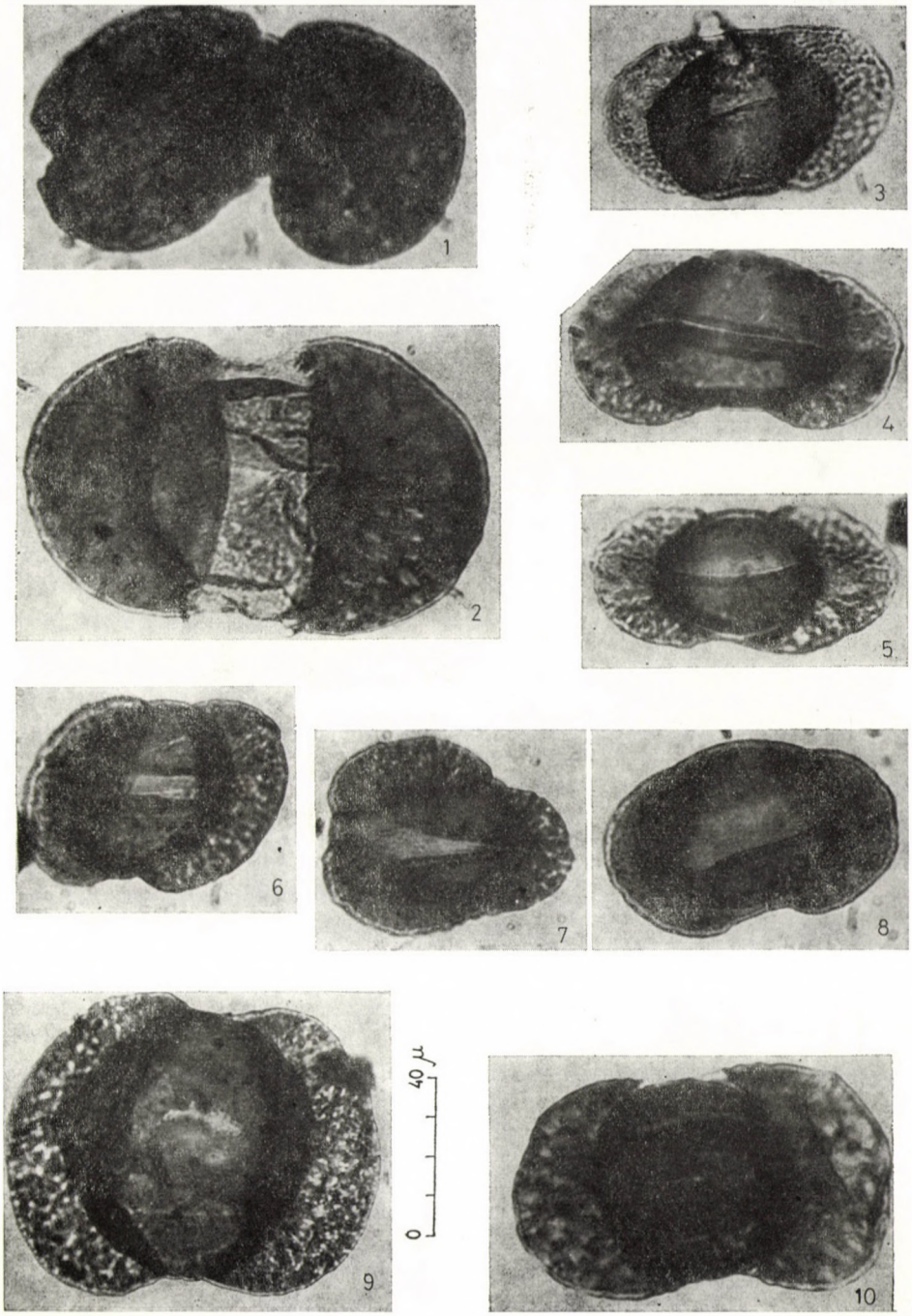

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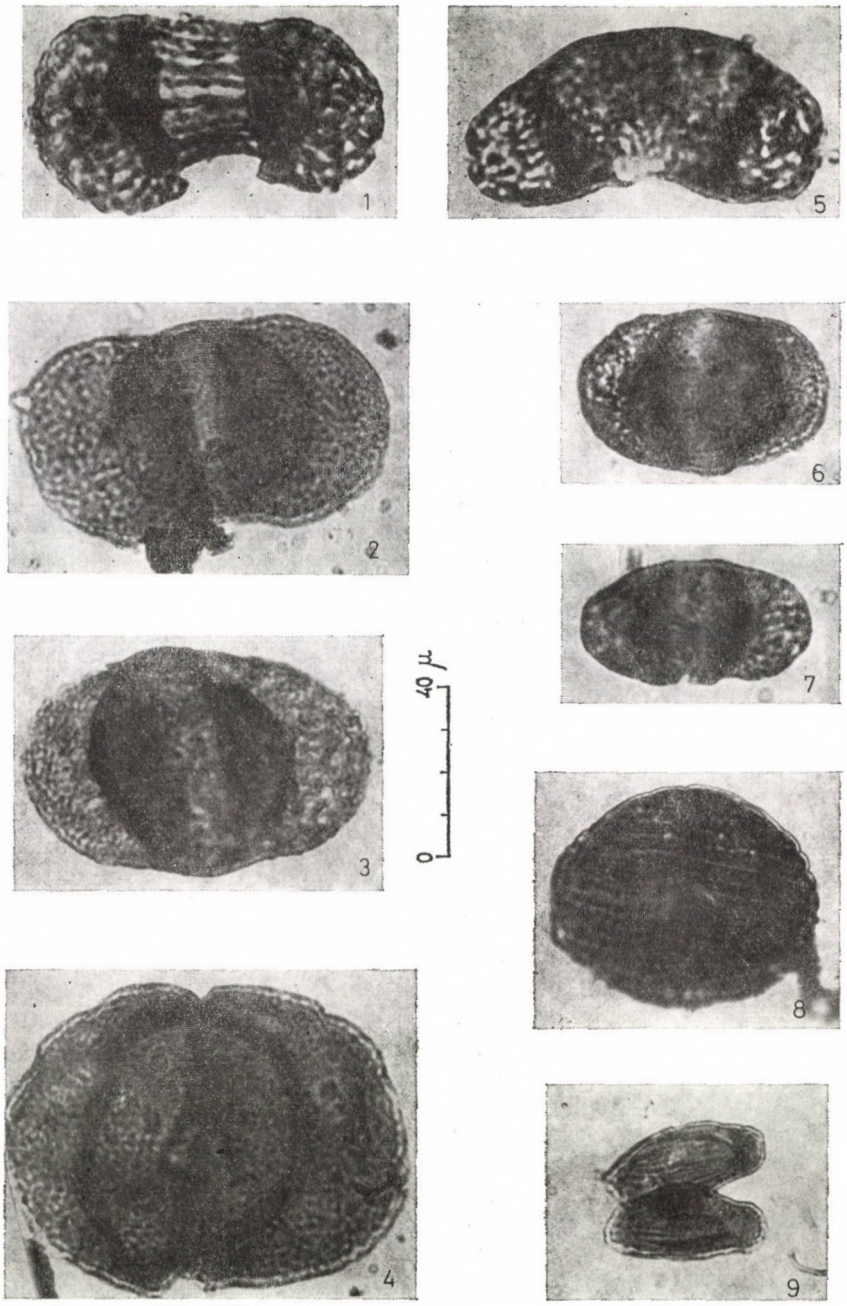

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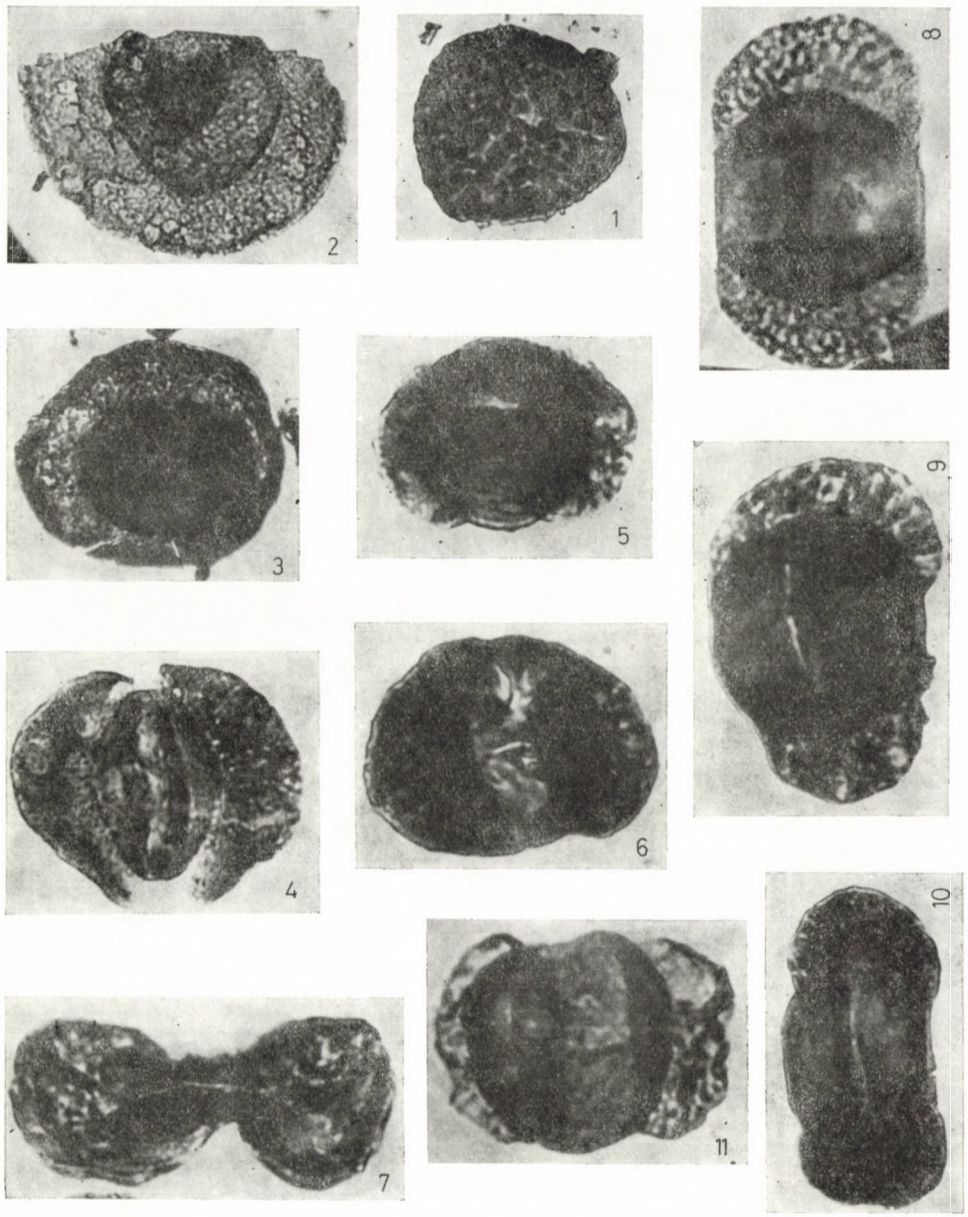
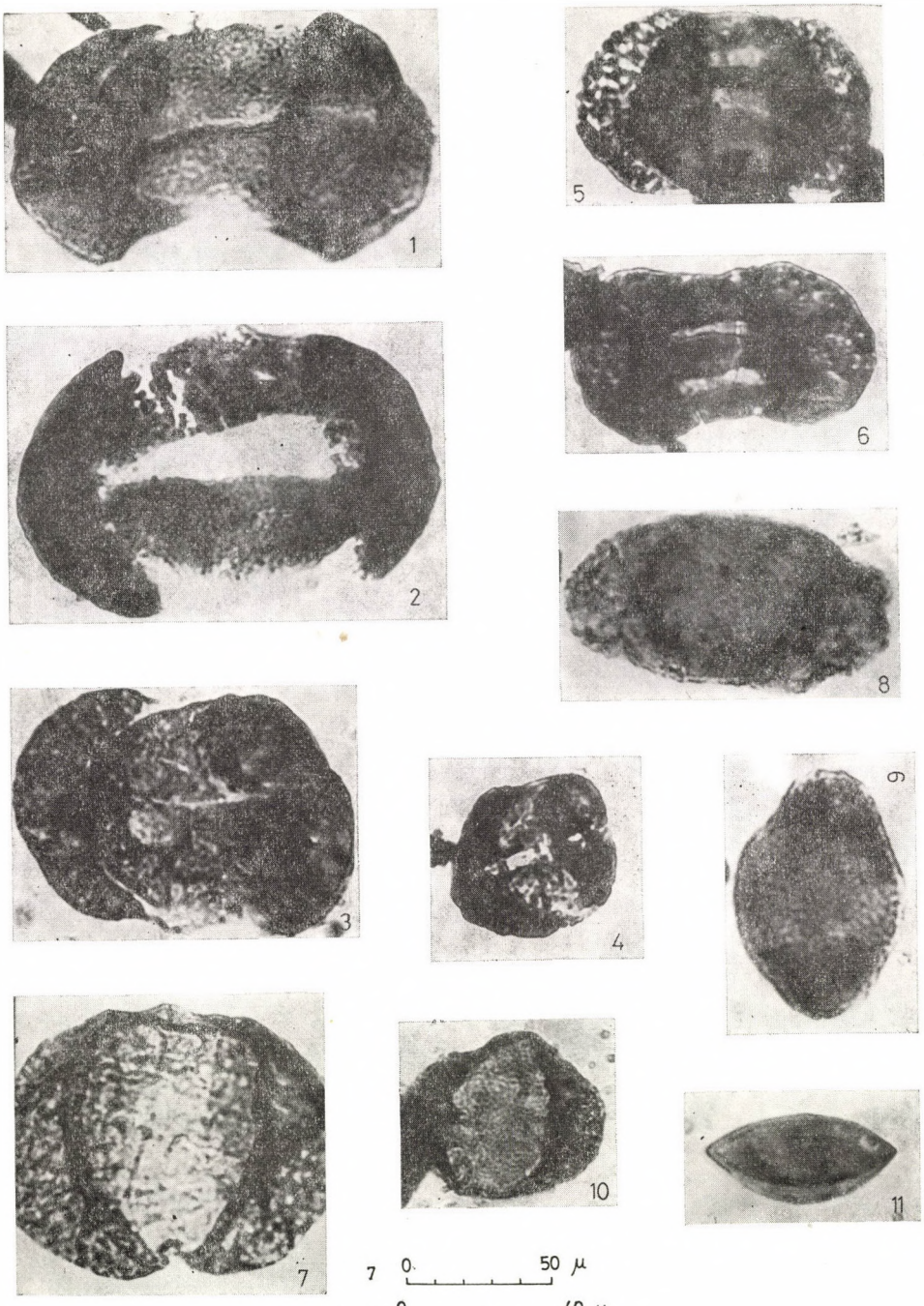


2–3 0 50 μ

1, 4–11 0 40 μ

Plate V



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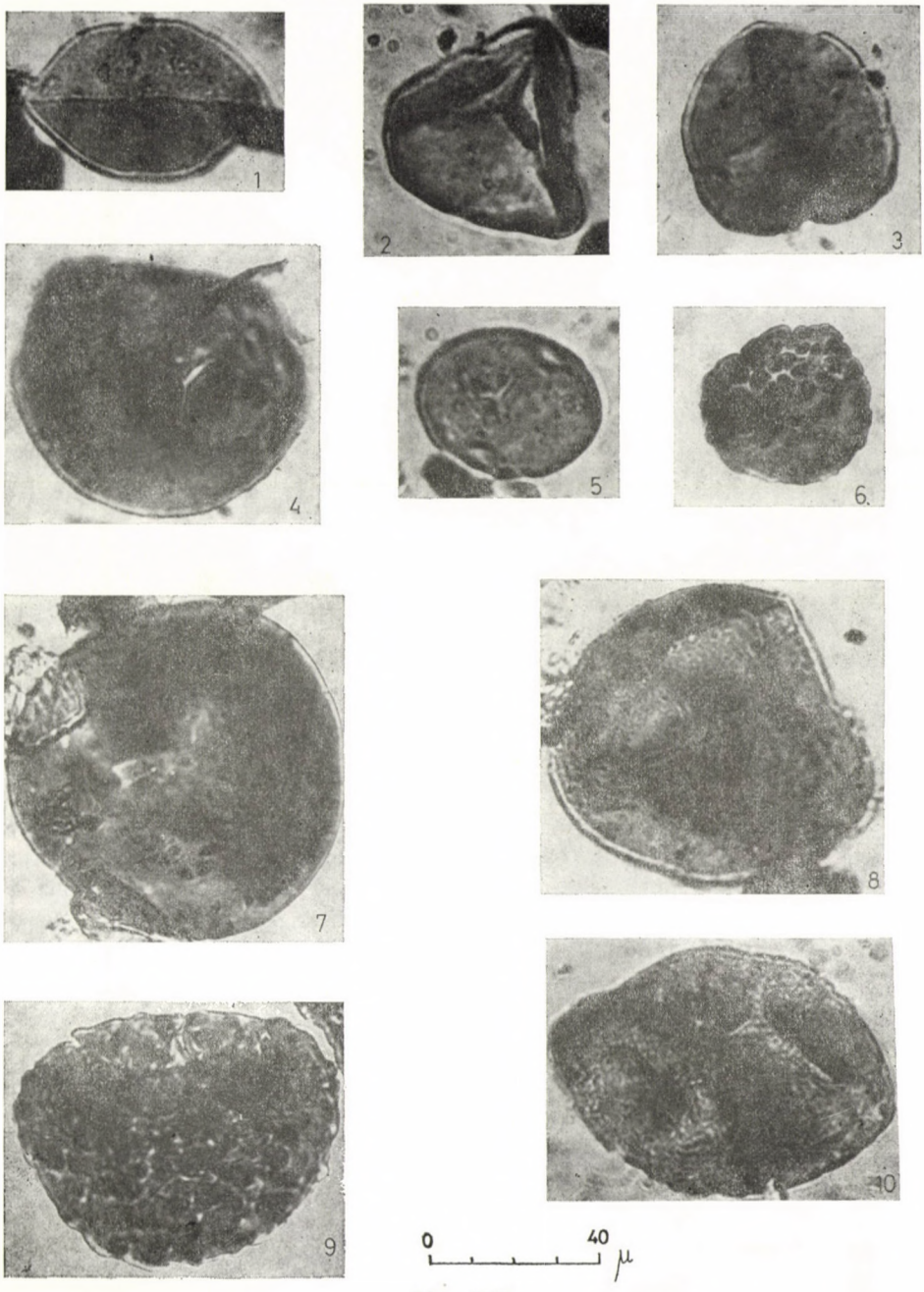

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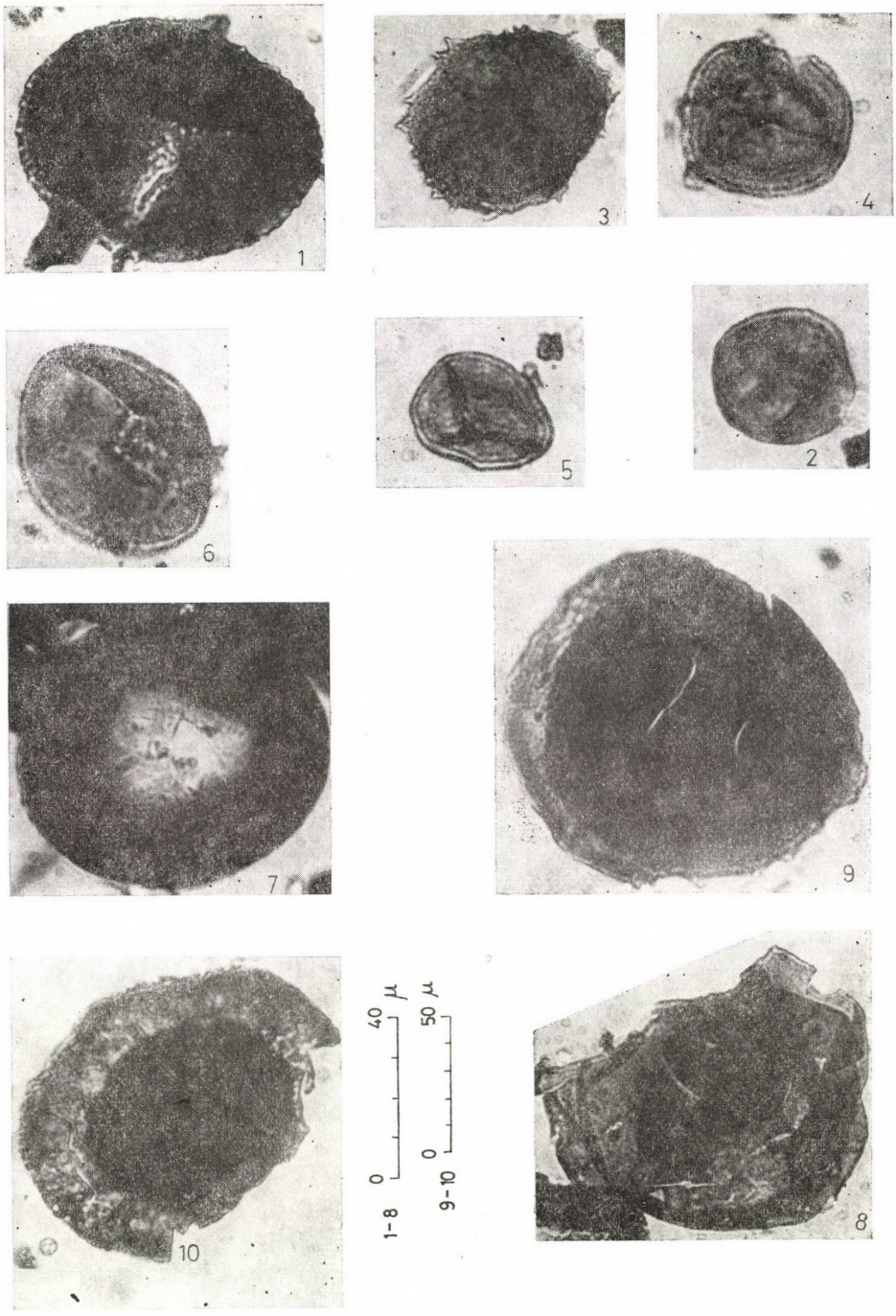

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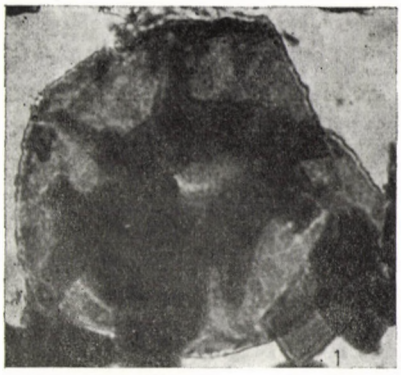

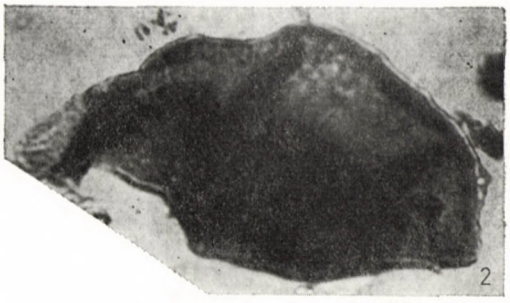

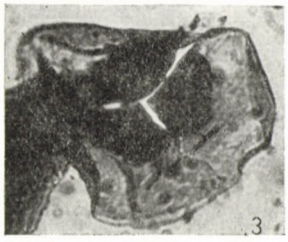

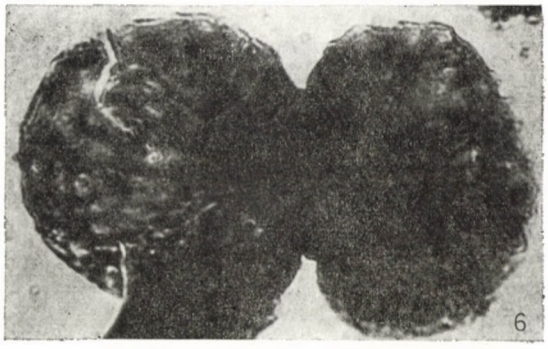

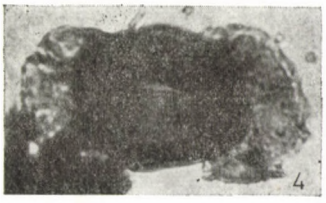

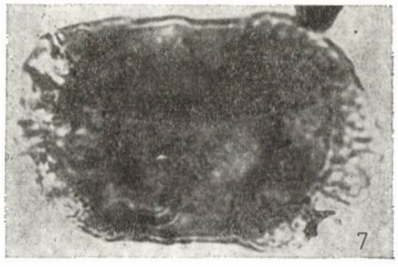

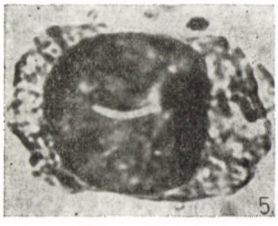

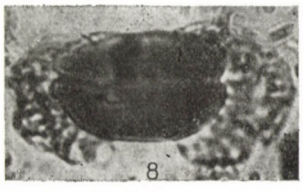

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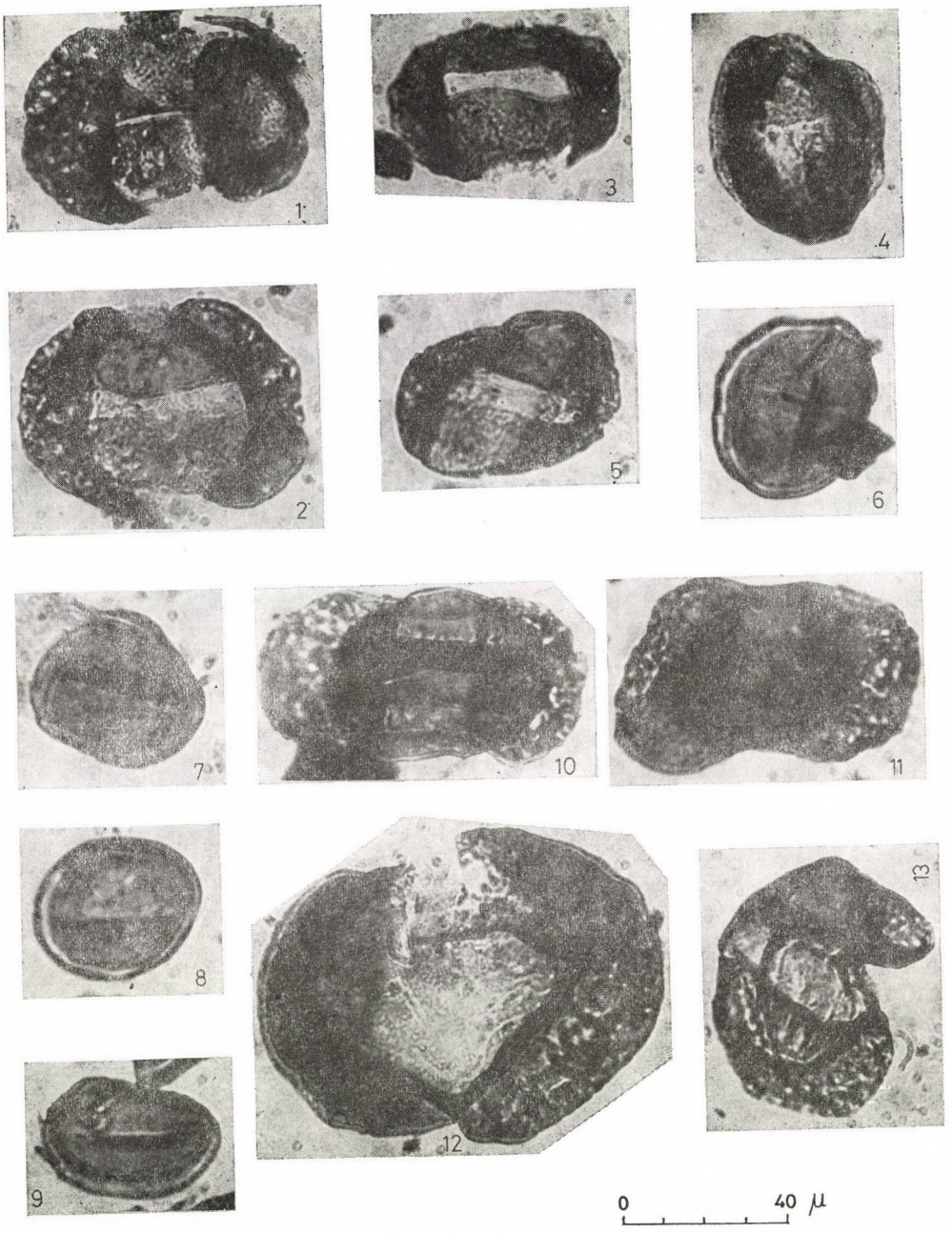

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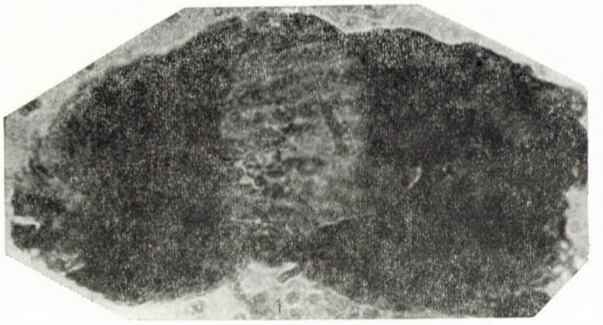
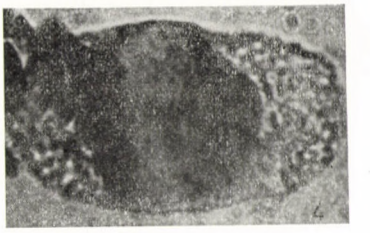
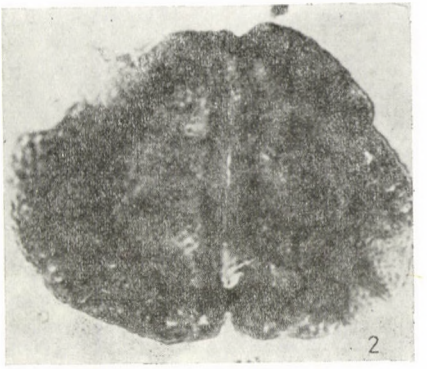
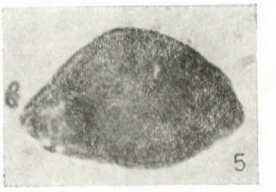
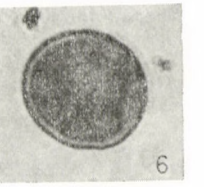
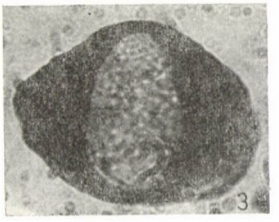






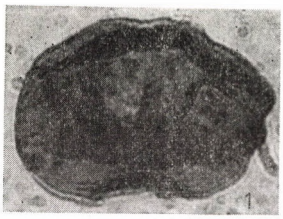


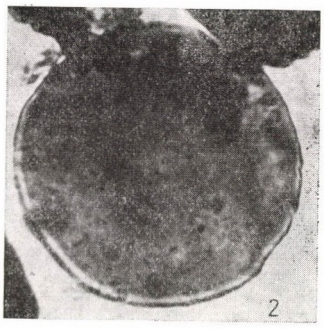


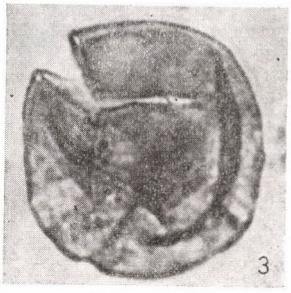


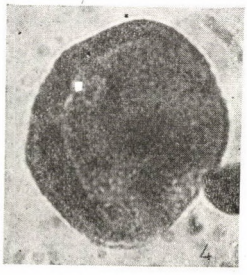
0 40 μ

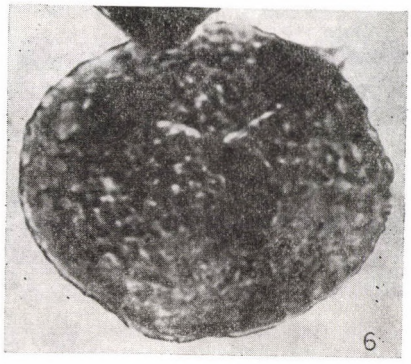
Plate XI











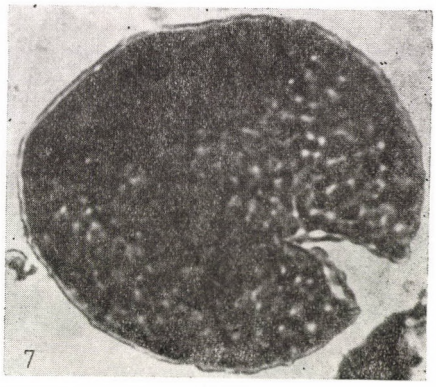


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Plate XII

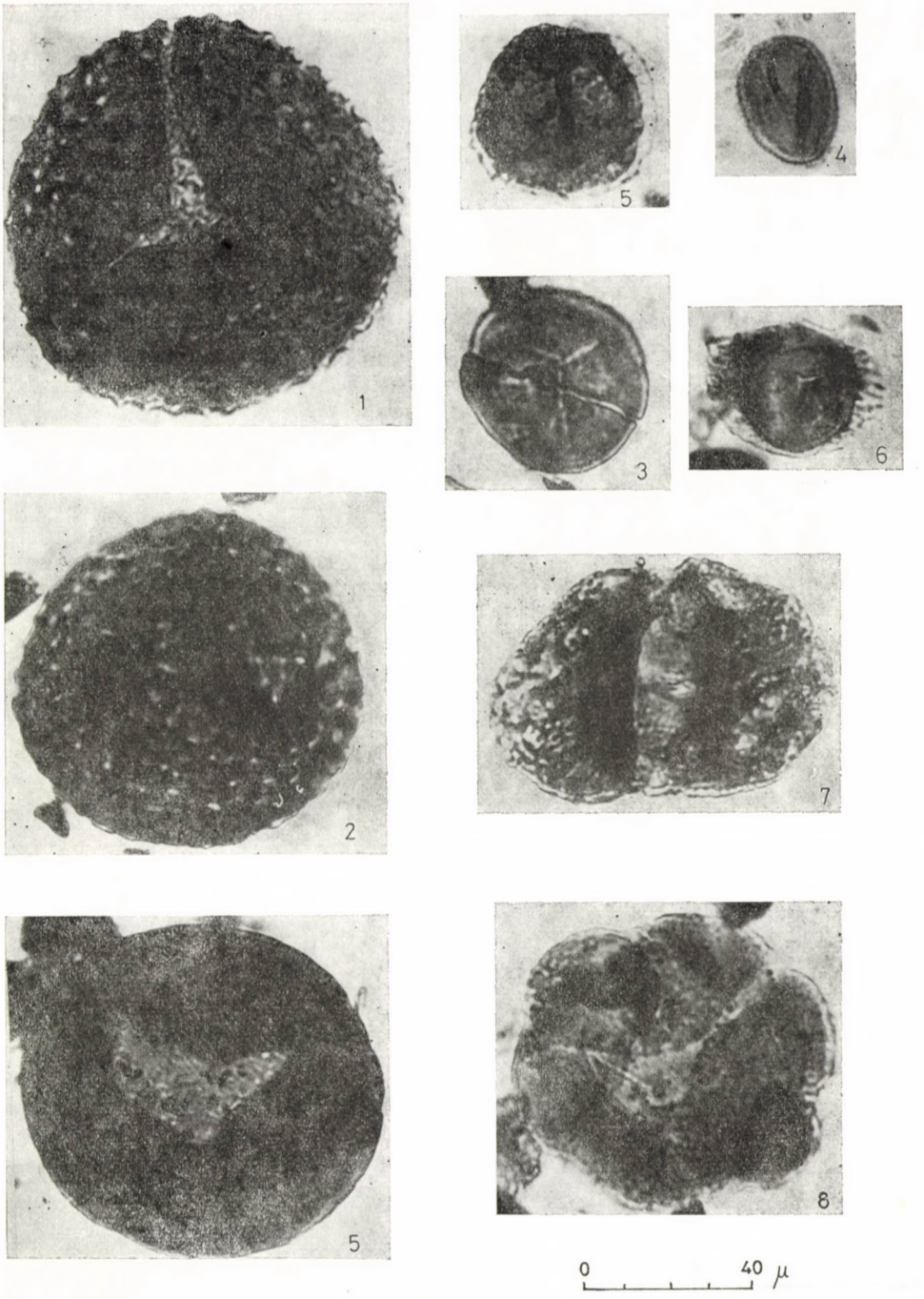


Plate XIII

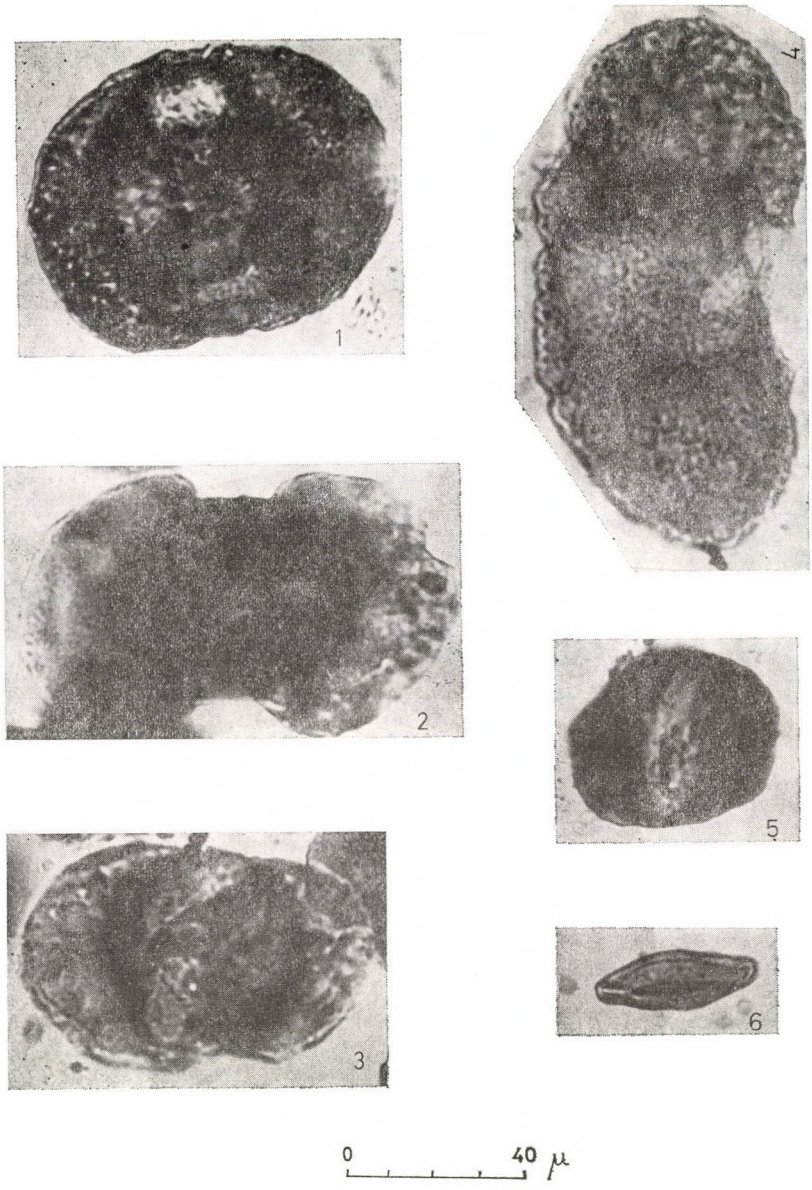
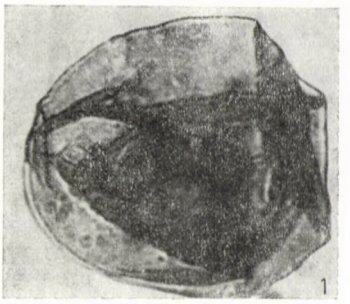
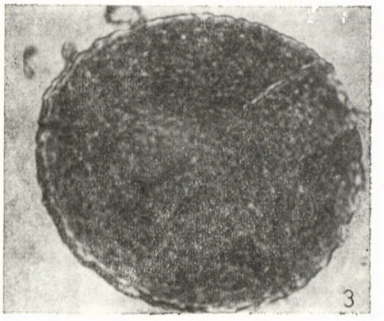
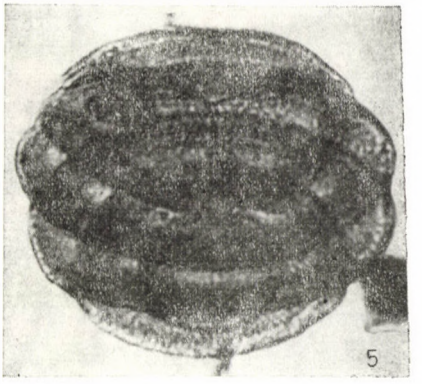


Plate XIV







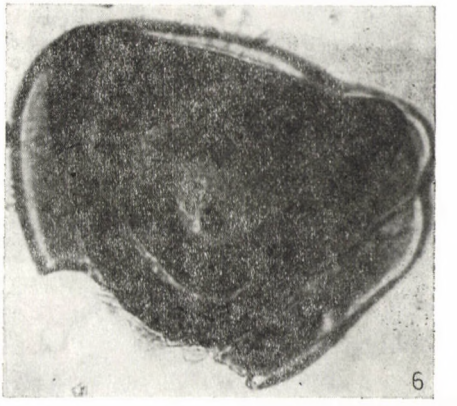


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Plate XV

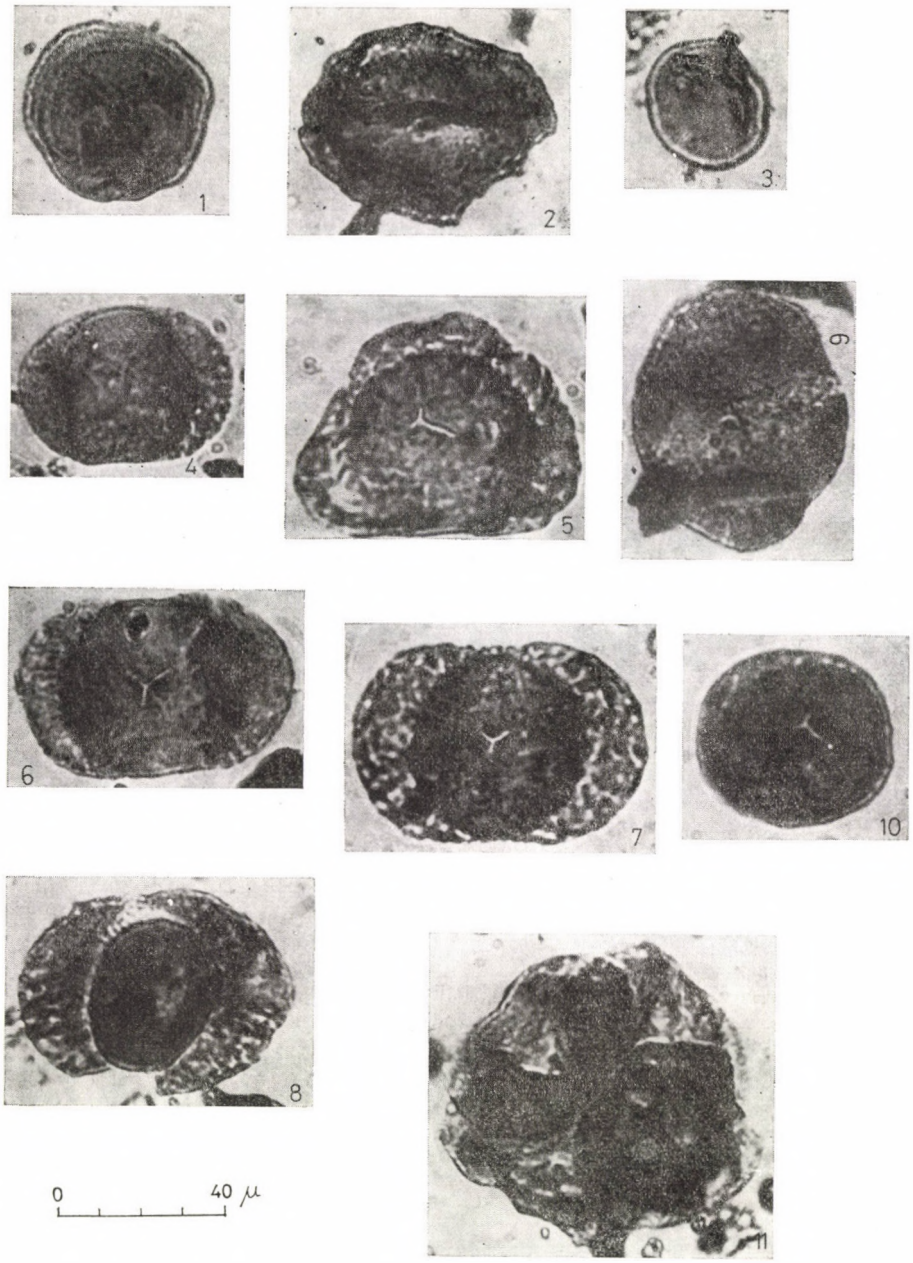
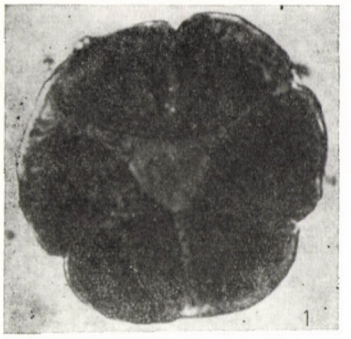
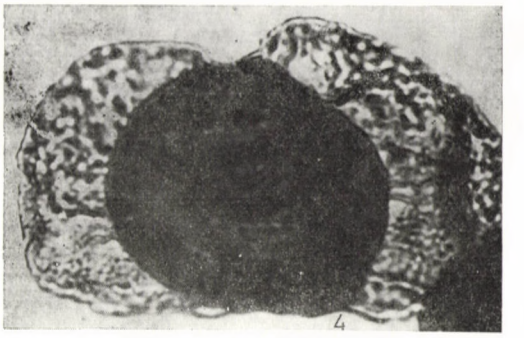
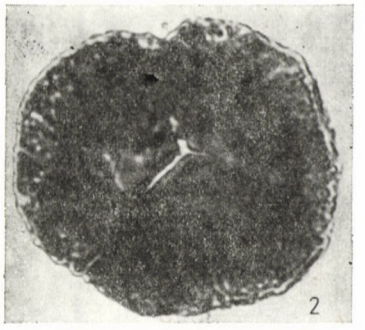
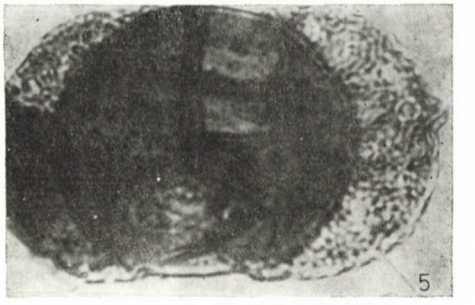


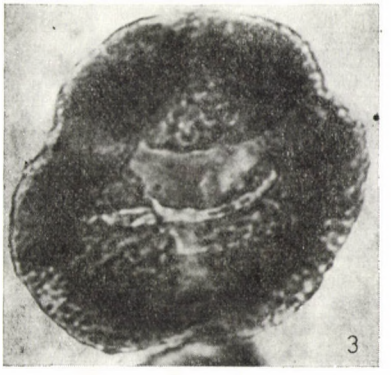
Plate XVI











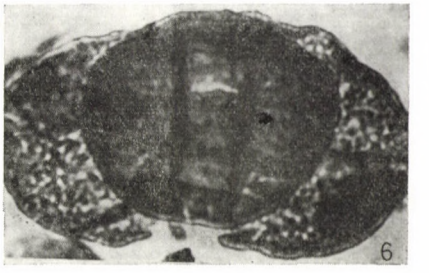


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Plate XVII

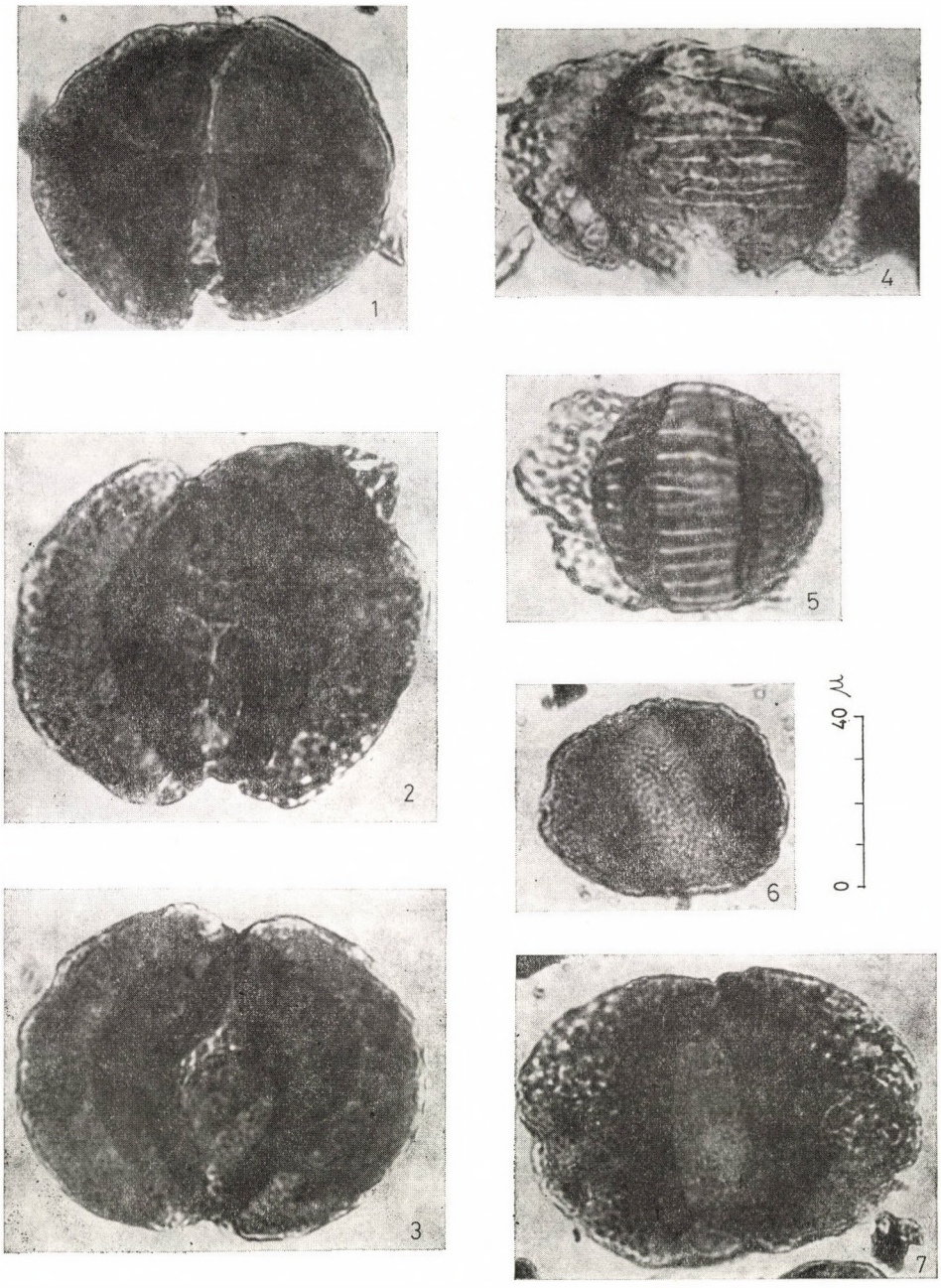
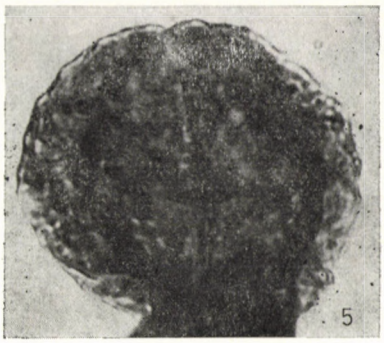
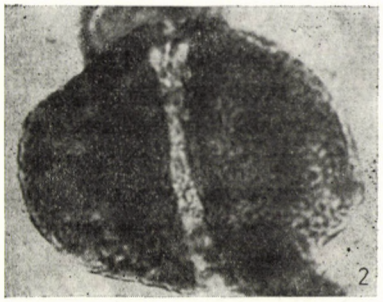
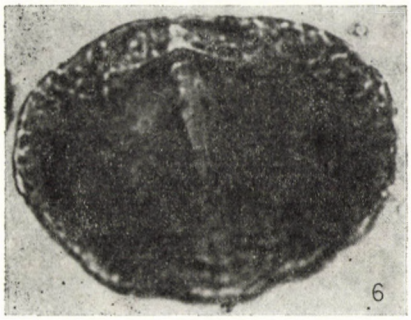
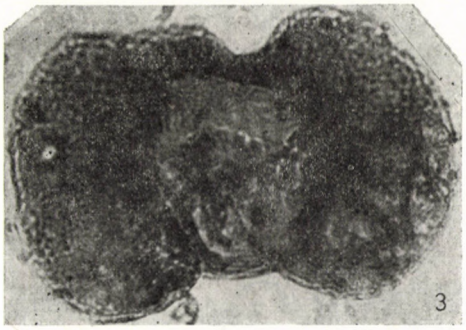


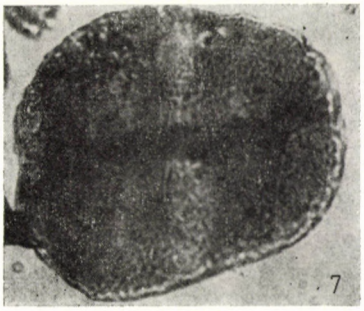
Plate XVIII

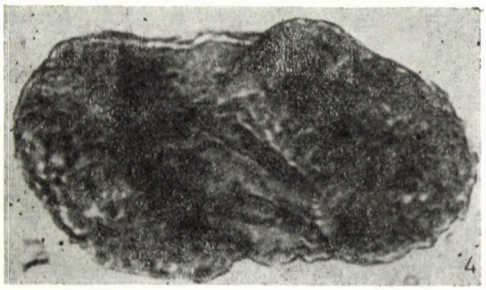












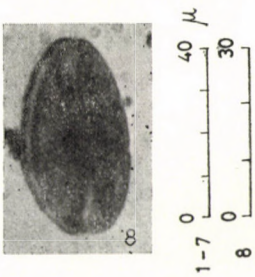


Plate XIX

Acta Geologica Academiae Scientiarum Hungaricae, Vol. 24 (1), pp. 99—115 (1981)

BIOSTRATIGRAPHIC AND SEDIMENTOLOGIC CHARACTERISTICS OF UPPER PERMIAN AND LOWER TRIASSIC SEDIMENTS IN THE GUČEVO MT. AREA, YUGOSLAVIA

By

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Authors present the investigations of Upper Permian and Lower Triassic microfossil association and the sedimentary environment within the time interval from Uppermost Paleozoic to Lowermost Mesozoic on the southeastern slopes of Gučevo Mountain (northwestern Serbia) at localities along the Vrelo Potok (Fig. 1) and Krivi Potok (Fig. 2) streams.

Introduction

Microfossil association of Upper Permian in northwestern Serbia was described first time by LÓCZY jun. (1919). Later, SIMIĆ, whose geological investigations of Late Palaeozoic and Triassic in western Serbia had the greatest significance, mentioned in his papers Upper Permian *Algae* and rare *Foraminifers* (1933 and 1938).

In this part of Serbia the fossil biocoenosis of Upper Permian was analyzed with noteworthy results: PANTIĆ (1959), KOCHANSKY-DEVIDÉ—HERAK (1960) and PANTIĆ (1969).

The Triassic microfossil association in the area of Gučevo Mt. was described by PANTIĆ—MOJSILOVIĆ (1968) after the investigations of facial characteristics of Podrinje and Valjevo areas.

Presenting in this paper two short geological sections from the Vrelo Potok (Fig. 1) and Krivi Potok (Fig. 2) localities, the authors intended to show some interesting results of detailed work within the interval Uppermost Permian—Lowermost Triassic, belonging to the part of Jadar area characterized by numerous outcrops, especially of the Palaeozoic rocks.

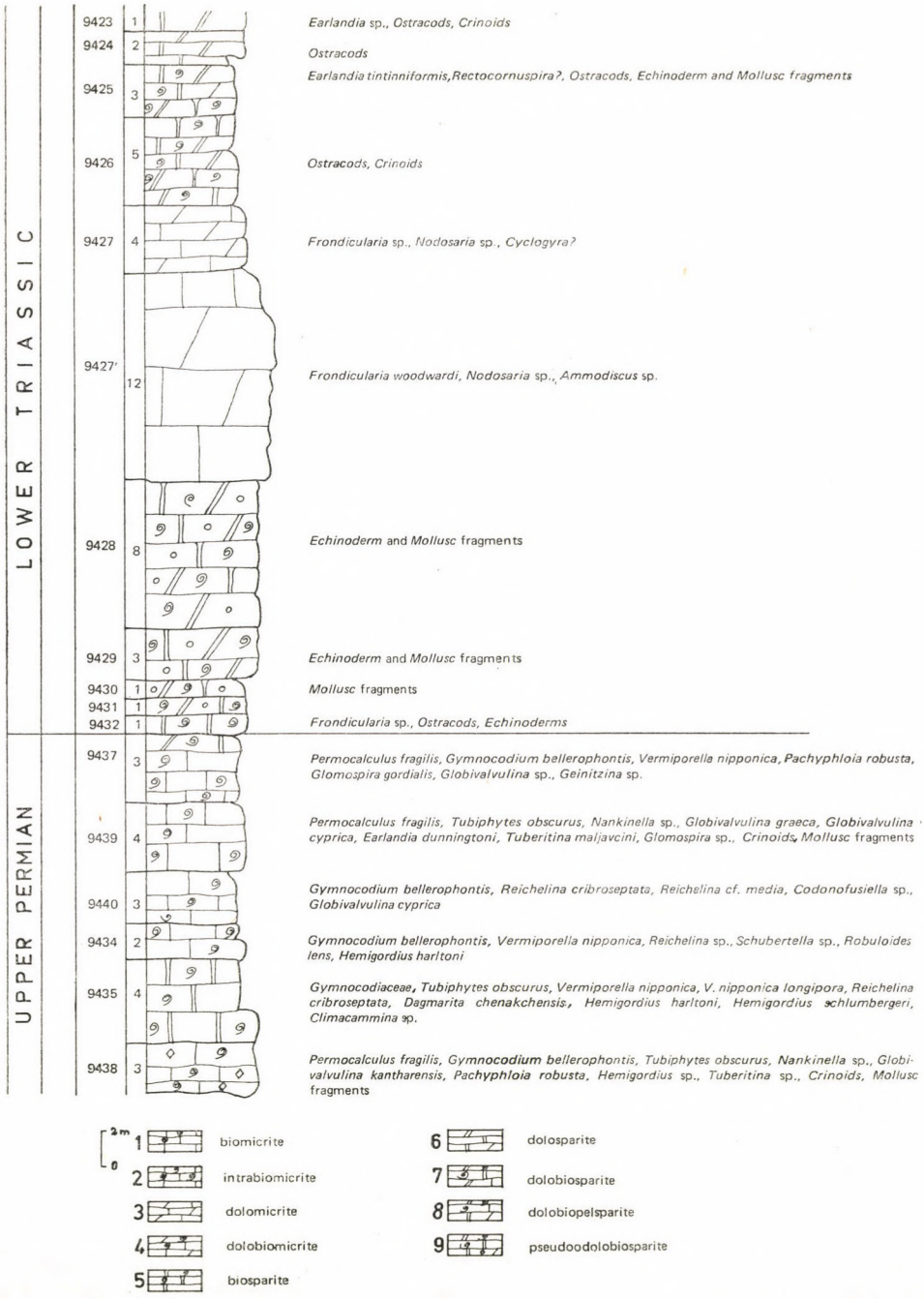

Fig. 1. Geological section of the Vrelo creek. 1. Biomicrite; 2. Intrabiomicrite; 3. Dolomicrite; 4. Dolobiomicrite; 5. Biosparite; 6. Dolosparite; 7. Dolobiosparite; 8. Dolobiopelsparite; 9. Pseudodolobiosparite

MAGYAR TUDOMÁNYOS AKADÉMIA KÖNYVTÁRA

Analyzed data showed very clearly the developments in the sedimentary basin containing variable biocoenosis and its evolution in the time interval from the end of Permian to the beginning of Triassic.

Micropaleontologic and Sedimentologic Characteristics of Geological Sections

1. *Geological Section at Potok Vrelo Stream (Fig. 1)*

In the exposed geological section at Vrelo Potok (Fig. 1), the UPPER PERMIAN sequence begins with cca 20 meters of bedded, dark grey, bituminous intrabiomicrite (specimen No 9438) with Algae: *Permocalculus fragilis* (PIA), *Gymnocodium bellerophontis* (ROTHPLETZ), *Tubiphytes obscurus* MASLOV; then Forams: *Nankinella* sp., *Pachyphloia robusta* K. M. MACLAY, *Globivalvulina kantharensis* REICHEL, *Hemigordius* sp., associated with Crinoids and Echinoderms.

Overlaying are thick-bedded, black biosparite (specimen No 9435) containing besides Gymnocodiums and Tubiphytes: *Vermiporella nipponica* ENDO, *Vermiporella nipponica longipora* PRATURLON; and Forams: *Reichelina cribroseptata* ERK, *Dagmarita chanakchensis* REITLINGER, *Hemigordius harltoni* (CUSHMAN et WATERS), *Frondicularia tumida* K. M. MACLAY, *Climacammina* sp.

At the beginning of the exposed section the sediments indicate the environment of somehow increased energy of the media, comparatively shallow, with the tendencies to pass into slightly deeper lagoonal environment.

Upwards dark grey, bedded/thick-bedded biomicrites (9434, 9440 and 9439) follow containing the same Algae *(Gymnocodiums, Permocalculuses* and *Tubiphytes)*, and slightly more variable Forams: *Globivalvulina graeca* REICHEL, *Globivalvulina cyprica* REICHEL, *Reichelina* cf. *media* K. M. MACLAY, *Earlandia dunningtoni* (ELLIOTT), *Tuberitina maljavkini* MIKHAILOV. This, the thickest interval within the Upper Permian sequence is represented by sediments deposited in a comparatively deep lagoon.

The final part of Upper Permian sequence consists of dark grey to black dolobiosparite (specimen No 9437), containing generally the same biocoenosis, only slightly enriched by several genera and species of Foraminifers: *Glomospira gordialis* (JONES et PARKER), *Pachyphloia robusta* K. M. MACLAY, *Geinitzina* sp., *Globivalvulina* sp., *Earlandia dunningtoni* (ELLIOTT). Sediments belonging to the final part of Upper Permian sequence at the Vrelo Potok locality, show gradual shallowing and closure of the lagoonal environment, followed by gradual increase in the energy of the media.

LOWER TRIASSIC begins with cca 40 meters of bedded, thick bedded and massive dolobio-oosparite to pseudo-oodolobiosparite, containing fragments

of macrofauna, often rounded by rolling action of certain agents. Examples of true oolites overnumber the pseudo-oolites. The above sediments were deposited in an area under the influence of submarine or surface marine currents. In this case, it could be interpreted, for example, as the consequence of the existence of high energy submarine channel linking the lagoon with the open sea. Microfossils are rare: only couple of *Frondicularia* sp., some *Ostracods* and detritus of *Echinoderms* (specimens Nos 9432—9428).

The interval of high energy circulations was substituted by the prolonged period of comparatively quiet lagoonal sedimentation, usually characteristic of this part of Lower Triassic (specimens Nos 9427—9423). In such environment, interchanges of grey to light grey, bedded to thick bedded dolobiomicrite and dolobiosparite could be quite appropriate. The sediments contain: *Frondicularia woodwardi* HOWCH., *Earlandia* MIŠÍK (ELLIOTT), *Nodosaria* sp., *Ammodiscus* sp., *Cyclogyra ?*, *Rectocornuspira ?*, *Ostracods* and *Crinoids*. The sediments as a whole are the product of deposition in the environment of rapid changes in the energy, with the marked periods of sea level oscillation, indicating, in a way, the decrease of the lagoon's size.

2. *Geological Section at Krivi Potok Stream (Fig. 2)*

At the Krivi Potok (Fig. 2) locality, the thickness of exposed Upper Permian sediments is approximately 30 ms., and the thickness of Lower Triassic sediments represents only lowermost parts of the Seisian substage.

The UPPER PERMIAN sequence begins with black bituminous biomicrite (specimen No. 9476) containing Algae: *Gymnocodium bellerophontis* (ROTHPLETZ), *Vermiporella nipponica* ENDO, *Vermiporella nipponica longipora* PRATURLON, together with Forams: *Codonofusiella* sp., *Hemigordius schlumbergeri* (HOW.), *Hemigordius harltoni* (CUSHMAN et WATERS), *Globivalvulina graeca* REICHEL, *Geinitzina ovata* LANGE, *Earlandia dunningtoni* (ELLIOTT).

Overlaying is the thick bedded, black, intrabiopelsparite with numerous calcite veins (specimen No. 9475) with abundant microfossils and pellets. The fossil biocoenosis consists of *Tubiphytes obscurus* MASLOV, fragments of *Gymnocodiaceae*, *Nankinella* sp., *Geinitzina cuneiformis* (JONES), *Nodosaria patula* K. M. MACLAY and *Pachyphloia* sp.

The next member is the comparatively thick interval of black, bedded biomicrite (specimen No. 9474) containing the same algal association. *Foraminifers*, besides listed above, are also: *Agathammina pusilla* (GEINITZ), *Hemigordius grozdilovae* IGONIN, *Geinitzina chapmani* CUSHMAN—WATERS, *Nodosaria sumatrensis* LANGE, *Climacamina* sp.

The column continues with bedded, black biosparite (specimens No. 9473 and 9472) with numerous microfossils, extremely abundant organogenic fragments, little pyrite and few small nests of organic (carbonaceous) matter

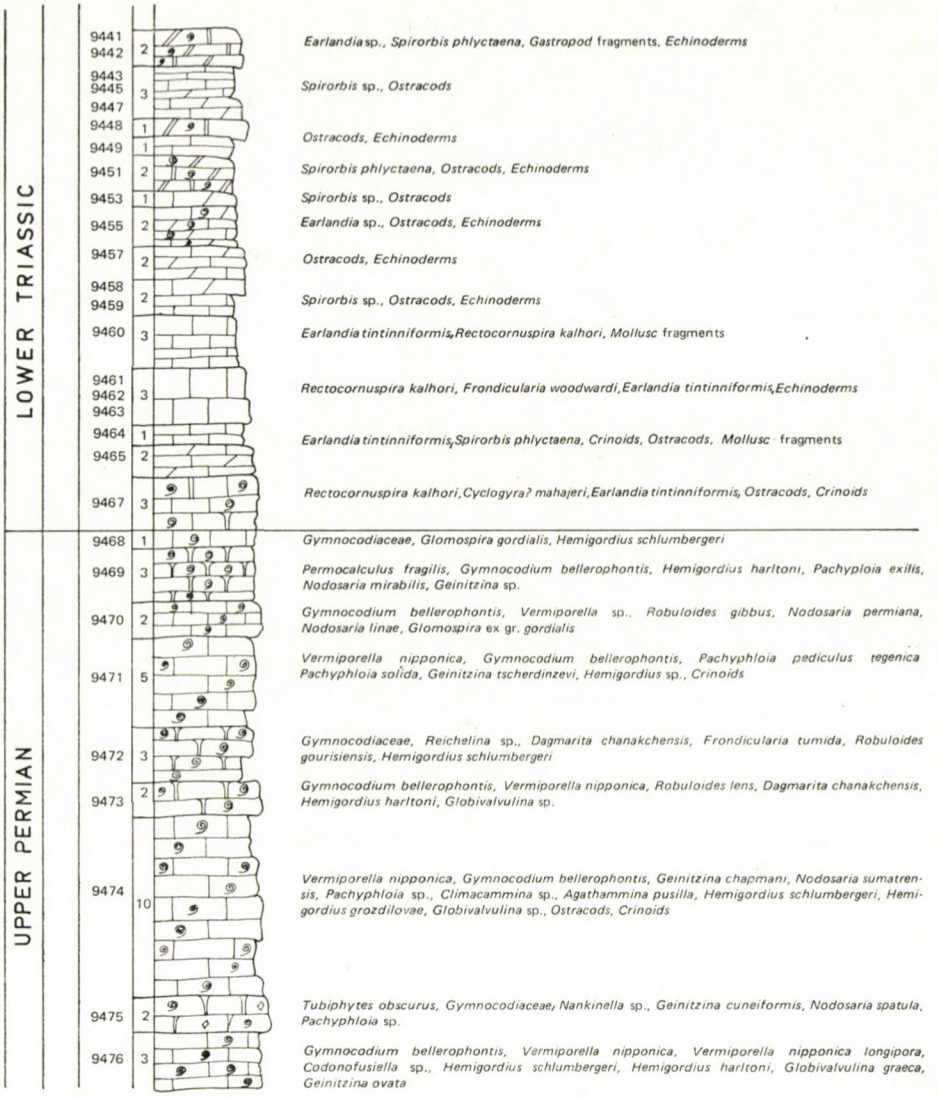

Fig. 2. Geological section of the Krivi creek (Yugoslavia). 1. Micrite; 2. Biomicrite; 3. Dolomicrite; 4. Dolobiomicrite; 5. Biosparite; 6. Dolobiosparite; 7. Intrabiopelsparite

in the sparry calcitic cementing matter. Thin calcitic veins are numerous. *Forams* are very frequent, represented by following genera and species: *Robuloides lens* REICHEL, *Robuloides gourisiensis* REICHEL, *Dagmarita chanakchensis* REITLINGER, *Frondicularia tumida* K. M. MACLAY and *Reichelina* sp. *Algae* are represented by *Gymnocodiums* and *Vermiporellae.*

Sediments described above and whose thickness amounts to cca 20 meters in the geological section (specimens Nos 9476 and 9472) indicate lagoonal depositional environment with infrequent changes in the energy of the media.

The last interval belonging to the Upper Permian at Krivi Potok stream (Fig. 2) is represented by interchange of biomicrite (specimens Nos 9471 and 9470) and biosparite (specimens Nos 9469 & 9468) containing *Algae* identical to those in the lower parts of the sequence with addition of numerous *Permocalculus fragilis* (PIA). Association is joined by *Lagenids: Pachyphloia pediculus tegenica* K. M. MACLAY, *Pachyphloia exilis* LUPETO, *Nodosaria permiana* (SPANDEL), *Nodosaria linai* K. M. MACLAY, *Nodosaria mirabilis* LIPINA, *Geinitzina tscherdinzevi* K. M. MACLAY, accompanied by: *Glomospira gordialis* (JONES et PARKER), *Robuloides gibbus* REICHEL, *Hemigordius umbilicatus* KIREEVA.

According to the sedimentological characteristics, the sediments belonging to the final part of Upper Permian (specimens Nos 9471—9468) indicate the stagnation of lagoonal phase, possibly with the increased energy of the environment. Generally speaking, in the investigated locality the sediments of Upper Permian age are the product of (most likely) lagoonal deposition, where several changes in the energy of the media, as well as gradual changes from basically deeper lagoonal environment to the very shallow one could be observed.

In the geological section of Krivi Potok (Fig. 2), the LOWER TRIASSIC begins with black, bedded biomicrite and biosparite (specimens Nos 9467 and 9466), in the normal succession over the Permian sediments. After the interval of high energy environment in the Upper Permian, sedimentation at the beginning of Lower Triassic took place in a stable lagoonal environment, very likely isolated, as the faunal association appears to be very sparce. Microfossils are rare, represented by *Foraminifers: Rectocornuspira kalhori* BRÖNN. —ZANIN —BOZORGN., *Cyclogyra ? mahajeri* BRÖNN. —ZANIN. —BOZORG., *Earlandia tintinniformis* (MIŠÍK) in the association with *Ostracods* and *Crinoids.*

The lagoonal environment was the place of sedimentation of the next successive interval of Lower Triassic sediments (specimens 9465 —9448). For the difference, this interval is represented by the interbeddings of micrite, dolomicrite, dolobiomicrite and dolobiosparite, indicating the prolonged existence of the lagoon, but this time within the higher energy zone, with the frequent oscillation of the water depth and prominent influence of climatic and other exogenetic agents. The fauna is more numerous as compared to the

Fig. 3. Locality map

preceding interval, being represented by *Forams. Frondicularia woodwardi* HOWCH., *Rectocornuspira kalhori* BRÖNN.—ZANIN.—BOZORG., *Earlandia tintinniformis* (MIŠÍK) in the association with *Spirorbis phlyctaena* BRÖNN.—ZANIN., *Ostracods* and *Crinoids*.

Upwards in the isolated lagoon, quiet, short-lasting episode of sedimentation is represented by typical micrite (specimens Nos 9447—9443) with the minimal biogenic content — Spirorbis and Ostracods.

By the end of the exposed and investigated section, grey, bedded dolosparite (samples Nos 9442—9441) prevail again with *Spirorbis phlyctaena* BRÖNN.—ZANIN., *Earlandia* sp., Echinoderms and Gastropods. Lagoonal environment becomes shallower and energy of the media increases.

Conclusion

In the investigated area of northwestern Serbia, in both localities: Vrelo Potok (Fig. 1) and Krivi Potok (Fig. 2) creeks, Lower Triassic sediments overlay Upper Permian sequence, having normal contact difficult to observe without help of detail biolithofacial investigations.

The *Upper Permian* sediments at VRELO POTOK (Fig. 1) locality are represented by: a) *dark grey, intrabiomicrite* with Algae — *Permocalculus fragilis* (PIA), *Gymnocodium bellerophontis* (ROTHPLETZ), *Tubiphytes obscurus* MASLOV; Forams — *Nankinella* sp., *Globivalvulina kantharensis* REICHEL, *Pachyphloia robusta* K. M. MACLAY, *Hemigordius* sp., *Tuberitina* sp., Crinoids and Echinoderms; *black biosparite* containing Algae — *Vermiporella nipponica* ENDO, *Vermiporella nipponica longipora* PRATURLON and others listed above, then Forams: *Reichelina cribroseptata* ERK, *Dagmarita chanakchensis* REITLINGER, *Hemigordius harltoni* (CUSHMAN et WATERS), *Hemigordius schlumbergeri* (HOWCH.), *Frondicularia tumida* K. M. MACLAY, *Climacammina* sp., and c) *black biomicrite* containing the same Algae and Forams enriched by following genera and species: *Schubertella* sp., *Codonofusiella* sp., *Globivalvulina graeca* REICHEL, *Globivalvulina cyprica* REICHEL, *Robuloides lens* REICHEL, *Earlandia dunningtoni* (ELLIOTT), *Tuberitina maljavkini* MIKHAILOV, finally d) uppermost part of the sequence represented by *black, bedden dolobiosparite* containing almost the same fossil biocoenosis.

At KRIVI POTOK (Fig. 2) creek, *Upper Permian* sequence consists of: a) *black biomicrite* containing Algae: *Gymnocodium bellerophontis* (ROTHPLETZ), *Vermiporella nipponica* ENDO, *Vermiporella nipponica longipora* PRATURLON and Foraminifers: *Codonofusiella* sp., *Hemigordius schlumbergeri* (HOWCH.), *Hemigordius harltoni* (CUSHMAN et WATERS), *Globivalvulina graeca* REICHEL, *Geinitzina ovata* LANGE, *Earlandia dunningtoni* (ELLIOTT), then b) also *black intrabiopelsparite* with: *Tubiphytes obscurus* MASLOV, *Nankinella* sp., *Geinitzina cuneiformis* (JONES), *Nodosaria patula* K. M. MACLAY and *Pachyphloia* sp. Overlaying are c) *black biomicrite* containing the same species of *Gymnocodiums* and *Vermiporellae*, as at the beginning of Upper Permian exposed sequence, and Forams: *Geinitzina chapmani* CUSHMAN et WATERS, *Nodosaria sumatrensis* LANGE, *Agathammina pusilla* (GEINITZ), *Hemigordius grozdilovae* IGONIN, together with already listed *Hemigordiuses* from the preceding intervals. The Upper Permian sequence ends with d) *interbedded biosparite and biomicrite*, containing the same association of Algae, and variety of Forams: *Robuloides lens* REICHEL, *Robuloides gourisiensis* REICHEL, *Robuloides gibbus* REICHEL, *Dagmarita chanakchensis* REITLINGER, *Glomospira gordialis* (JONES—PARKER), *Geinitzina tscherdincevi* K. M. MACLAY, *Pachyphloia exilis* LUPETO, *Pachyphloia solida* K. M. MACLAY, *Pachyploia pediculus tegenica* K. M. MACLAY, *Nodosaria mirabilis* LIPINA, *Nodosaria permiana* SPANDEL, *Nodosaria linae* K. M. MACLAY, *Reichelina* sp.

All the mentioned sediments together with their biocoenosis were deposited in a large, deep lagoon with frequent changes in the energy of the environment, indicating most probably interchanges in shallowing and deepening episodes in the development and life of the lagoon.

The *Triassic* sediments at the VRELO POTOK (Fig. 1) locality have the

normal lower contact with the underlaying Permian sediments. The Triassic sequence begins with dolobio-oosparite and pseudo-oodolobiosparite containing numerous macrofossils and their fragments rounded by the rolling action; then scarce *Frondicularia* sp., *Nodosaria* sp., *Ostracods* and *Echinoderms*.

At KRIVI POTOK (Fig. 2) stream, slightly different, lithologically variable sediments are the equivalents of oolitic and pseudo-oolitic sediments, also conformably overlaying the Upper Permian sediments. In the first place, these are black biomicrite and biosparite with Forams: *Rectocornuspira kalhori* BRÖNN.—BOZORGN., *Cyclogyra mahajeri* BRÖNN.—ZANIN.—BOZORGN., *Earlandia tintinniformis* (MIŠÍK) in association with *Ostracods* and *Crinoids*. Upwards follows the part of the sequence represented by the interchanges of thin-bedded dolomicrite, dolobiomicrite, micrite and dolobiosparite containing generally the same microfossil association, and finally, micrite with *Spirorbis phlyctaena* BRÖNN.—ZANIN., *Earlandia* sp., *Ostracods*, *Echinoderms* and *Gastropods*.

In the both investigated localities, the sediments at the base of Lower Triassic sequence indicate the continuous existence of the Permian lagoon. At the beginning of Triassic, the lagoon has been stabilized, with very little evidence on the changes in sedimentary conditions. At that time the lagoon has been either affected by strong currents, or connected with the other part(s) of the shallow sea by channels. In either case the result is the presence of oolitic and pseudo-oolitic sediments.

In the geological sections of Krivi Potok (Fig. 2) and Vrelo Potok (Fig. 1) the final part of the Lower Triassic sequence is represented by dolomicrite and dolosparite containing *Earlandiae*, *Srirorbidae*, *Ostracods*, *Echinoderms* and fragments of *Molluscs*. These sediments are the product of deposition in somewhat restricted lagoon, and most likely indicate the expiry of the Triassic lagoon, giving the place to the development of Middle Triassic reef.

Though within the investigated interval: Upper Permian—Lower Triassic the sedimentary environment was lagoonal, as the existence of the more or less continuous lagoon has been proved by the characteristics of the biocoenosis and lithological composition of the sediments, it should be noted that great changes including ecological ones occured on the boundary between Upper Permian and Lower Triassic, changing the organic world. The rich and variable microfaunal and microfloral biocoenosis of the Upper Permian almost completely expired. Only sparce *Lagenids* and *Crinoids* passed into Triassic, followed by the occurence of sometimes frequent *Ostracods* and *Spirorbises*.

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БИОСТРАТИГРАФИЧЕСКИЕ И СЕДИМЕНТОЛОГИЧЕСКИЕ ХАРАКТЕРИСТИКИ ВЕРХНЕПЕРМСКИХ И НИЖНЕТРИАСОВЫХ ОТЛОЖЕНИЙ В РАЙОНЕ ГОР ГУЧЕВО В ЮГОСЛАВИИ

С. ПАНТИЧ–В. РАДОШЕВИЧ

Резюме

Авторы сообщают об исследованиях верхнепермских и нижнетриасовых ассоциациях ископаемых микроорганизмов и осадочной среды промежутка времени от самых верхов палеозоя до самого раннего мезозоя на юго-восточных склонах гор Гучево (СЗ Сербия), в речках Врело Поток (виг. I) и Криви Поток (фиг. 2).

Plate I

Vrelo Creek, Yugoslavia
Upper Permian

Fig. 1. *Geinitzina caucasica* K. M. MACLAY, thin section 9437, ×90
Fig. 2. *Robuloides lens* REICHEL, thin section 9434, ×90
Fig. 3. *Reichelina cribroseptata* ERK, thin section 9440, ×90
Figs 4—5. *Globivalvulina graeca* REICHEL, thin section 9439, ×70
Figs 6—7. *Hemigordius harltoni* (CUSHMAN et WATERS), thin section 9435, ×70

Plate II

Vrelo Greek, Yugoslavia
Lower Triassic

Figs 1—3. Dolobio-oosparite with *Echinoderm* and *Mollusc* fragments, and pellets, thin section 9429, ×30

Plate III

Krivi Creek, Yugoslavia
Upper Permian

Fig. 1. *Vermiporella nipponica* ENDO, thin section 9476, ×22
Fig. 2. *Gymnocodium bellerophontis* (ROTHPLETZ), thin section 9470, ×30
Fig. 3. *Glomospira gordialis* (JONES et PARKER), thin section 9468, ×70
Fig. 4. *Nodosaria linae* K. M. MACLAY, thin section 9470, ×70
Fig. 5. *Hemigordius schlumbergeri* (HOWCH.), thin section 9473, ×70
Fig. 6. *Reichelina* sp., thin section 9476, ×85
Fig. 7. *Geinitzina ovata* LANGE, thin section 9471, ×85
Fig. 8. *Frondicularia tumida* K. M. MACLAY, thin section 9472, ×85
Fig. 9. *Robuloides gibbus* REICHEL, thin section 9470, ×85

Plate IV

Krivi Creek, Yugoslavia
Lower Triassic

Figs 1—2. *Earlandia tintinniformis* (MIŠÍK), thin section 9463, ×85
Fig. 3. *Earlandia* sp., thin section 9460, ×85
Figs 4—6. *Rectocornuspira kalhori* BRÖN.—ZANIN.—BOZORG., thin section 9467 (4) and 9460 (5 and 6), ×85
Fig. 7. *Cyclogyra? mahajeri* BRÖN.—ZANIN.—BOZORG., thin section 9467, ×85
Figs 8—9. *Nodosaria* sp., thin section 9463, ×85
Fig. 10. *Spirorbis phlyctaena* BRÖN.—ZANIN., thin section 9463, ×22

Plate V

Krivi Creek, Yugoslavia
Lower Triassic

Fig. 1. Dolobio-oosparite with pellets, thin section 9454, ×30
Figs 2—3. Dolobio-oosparite with *Gastropods* and *Echinoderms*, thin section 9441, ×30

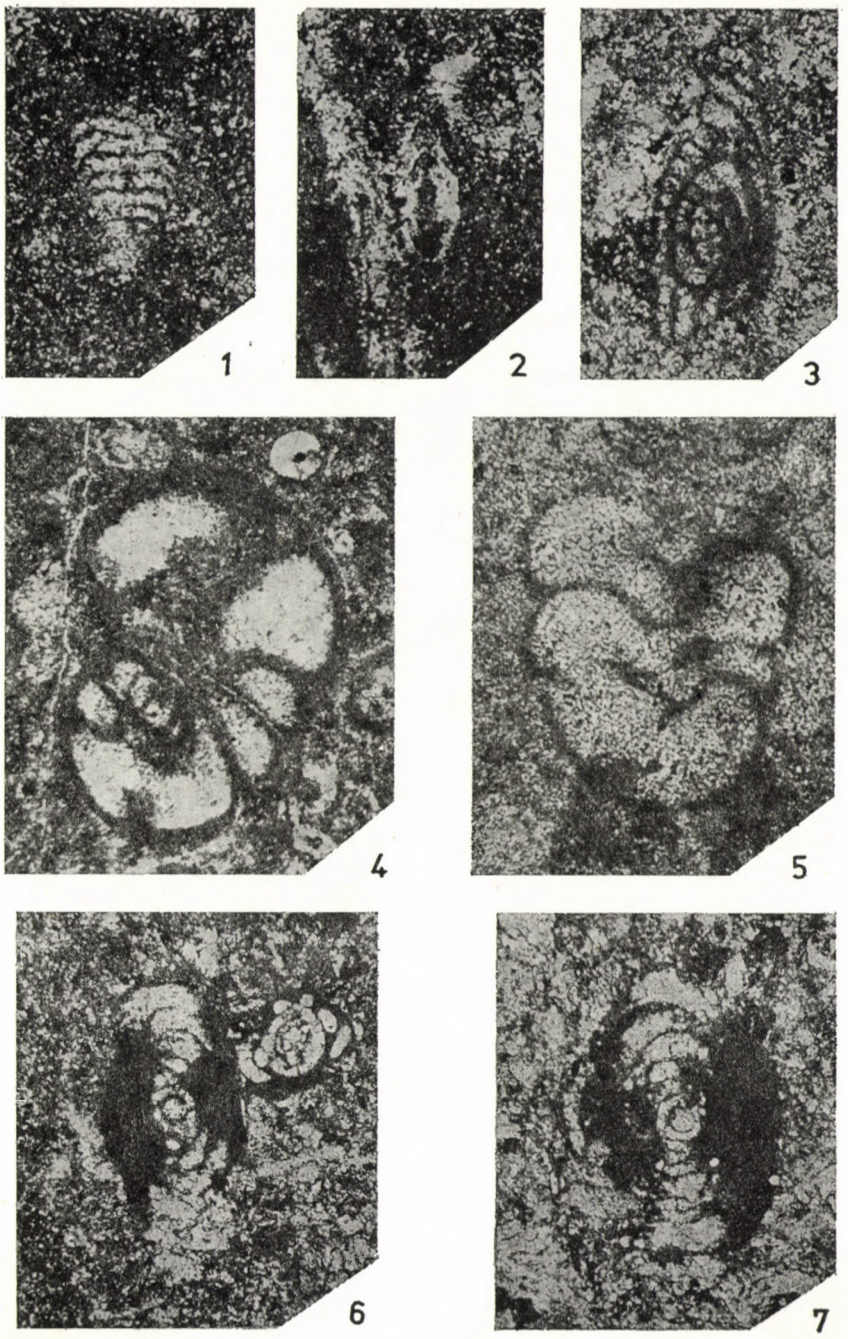

Plate I

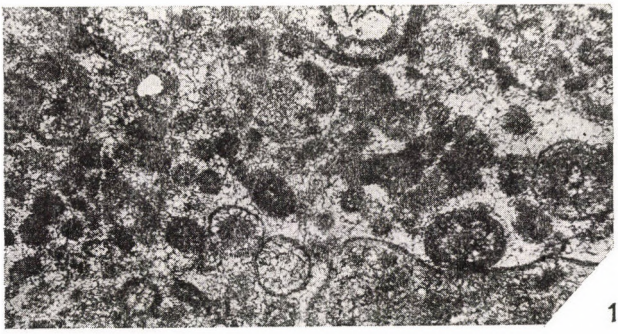

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Plate II

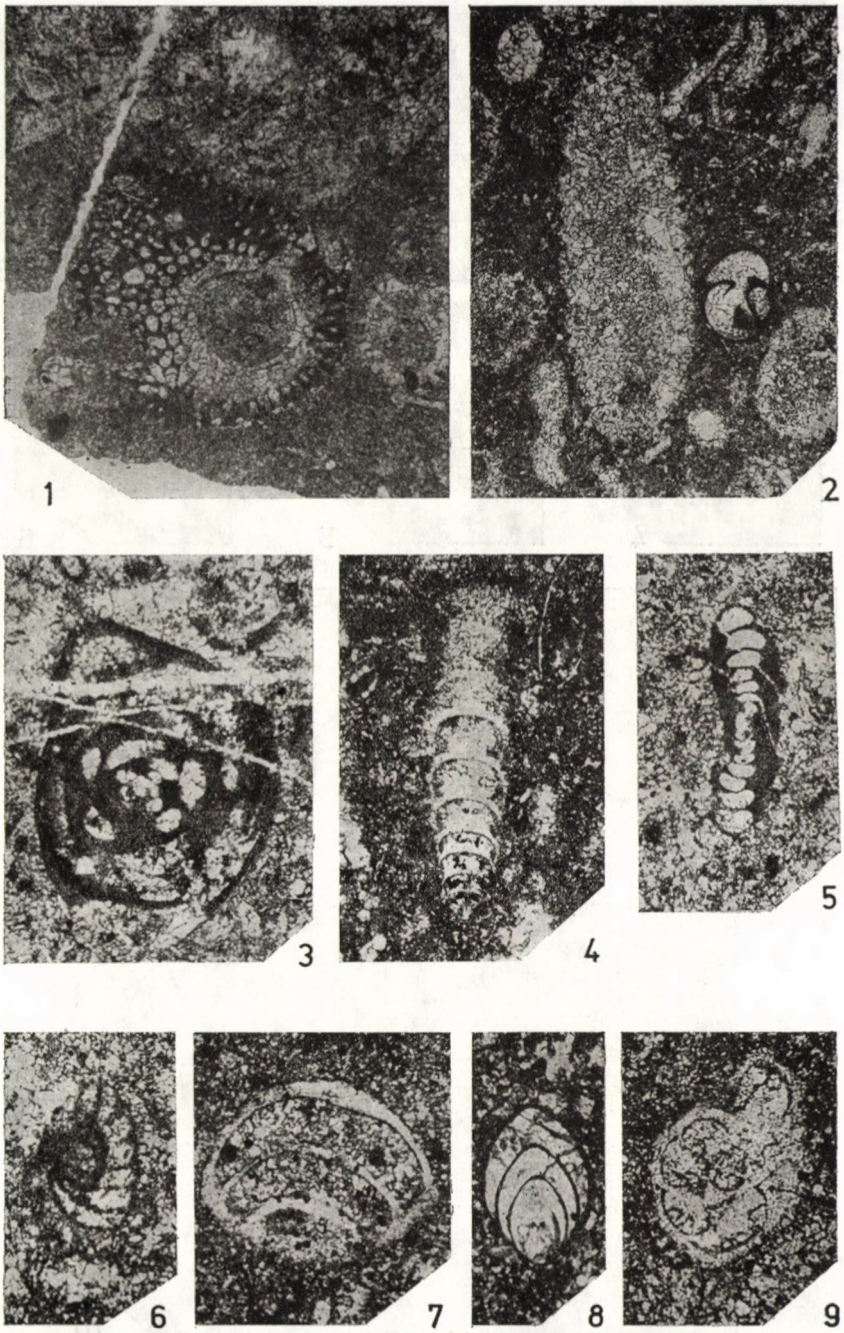

Plate III

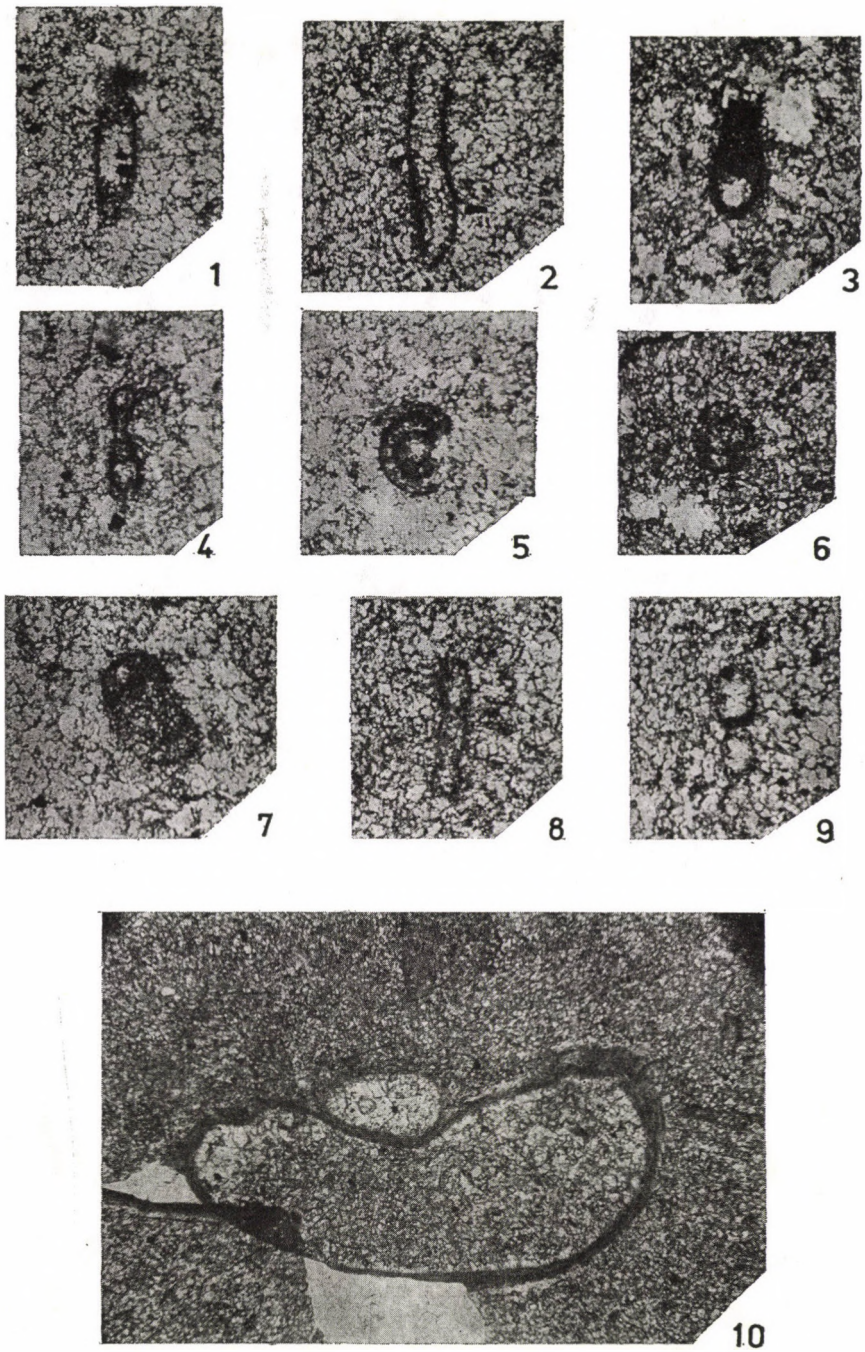

Plate IV

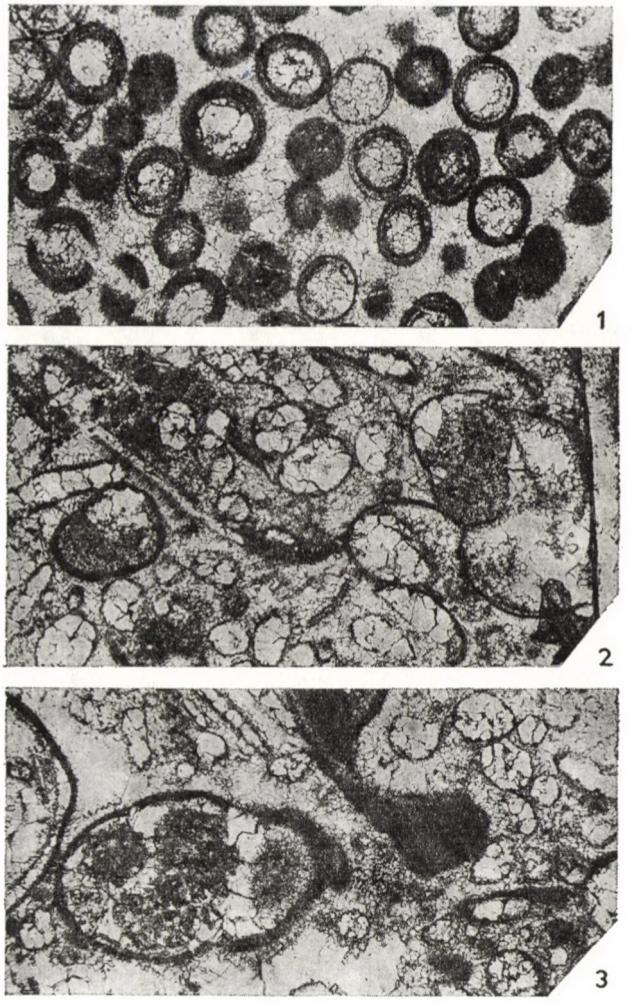

Plate V

Acta Geologica Academiae Scientiarum Hungaricae, Vol. 24 (1), pp. 117—128 (1981)

MARINE LOWER AND MIDDLE PERMIAN IN THE OIL EXPLORATORY WELL ÚJFALU-I (SW-HUNGARY)

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An apparently 1130 m thick Permian sequence has been intersected by the borehole Újfalu-I. The lower 788 m thick part of this sequence dominantly consists of fine-grained clastic rocks and contains fossiliferous dolomitic limestone beds and brecciated lenses of reefaceous character only in a subordinate proportion. On the basis of these, this sequence can be easily correlated with the clastic Košna-facies of the "Trogkofel stage" in Yugoslavia and can be assigned to the upper part of the Lower Permian. The 342 m thick dolomitic sequence overlying it may be the counterpart of the calcareous facies replacing the Middle Permian Gröden beds in the Velebit Mts. (Croatia). This new occurrence of the marine Lower and Middle Permian in Hungary corroborates the ideas formulated earlier concerning the continuation of the Late Palaeozoic sediments of Dinaric type in Hungary on the basis of occurrences of marine Uppermost Carboniferous layers at Karád and marine Upper Permian beds from Central Transdanubia to the Bükk Mts.

The oil exploratory well Újfalu-I spudded-in south of Lovászi village (Fig. 1) intersected a Permian sequence with an apparent thickness of 1130 m which is overlain by Middle Miocene beds of transgressive nature. The lower 788 m mostly consist of fine-grained clastic rocks (pale-grey fine-grained quartzose sandstone, silty shale, dark-grey to black argillaceous siltstone) and contain dark-grey dolomitic limestone beds or brecciated lenses of reefaceous character only in subordinate proportion. On the contrary, the upper 342 m part is a calcareous sequence, which is composed of dark-grey pure or argillaceous dolomite and dolomitic limestone.

Index fossils have been recovered only from the limestone interbeddings in about the middle part of the clastic sequence (Fig. 2). Namely, the brecciated limestone with dark-grey shale and fine-grained sandstone interbeddings of the core No. 5 taken from this part (3880.0—3886.5 m) contains a relatively populous microfossil assemblage which can be assigned to the upper part of the Lower Permian (Fig. 3):

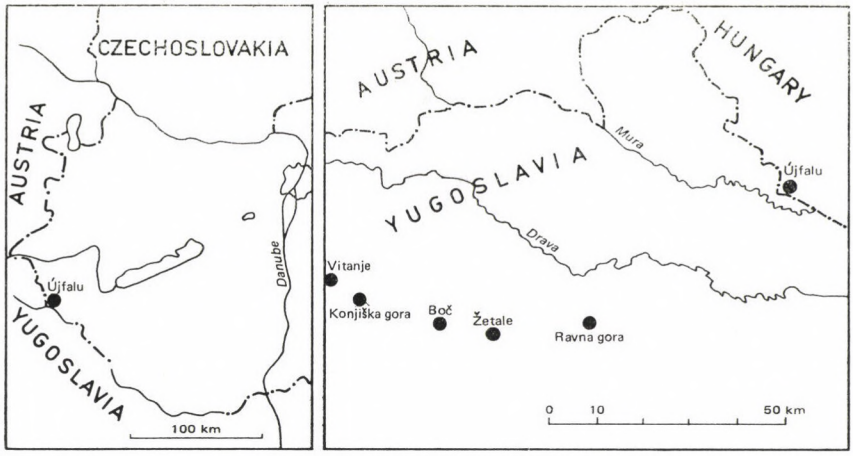

Fig. 1. Location of the well Újfalu-I and some localities of the Trogkofel beds in Yugoslavia

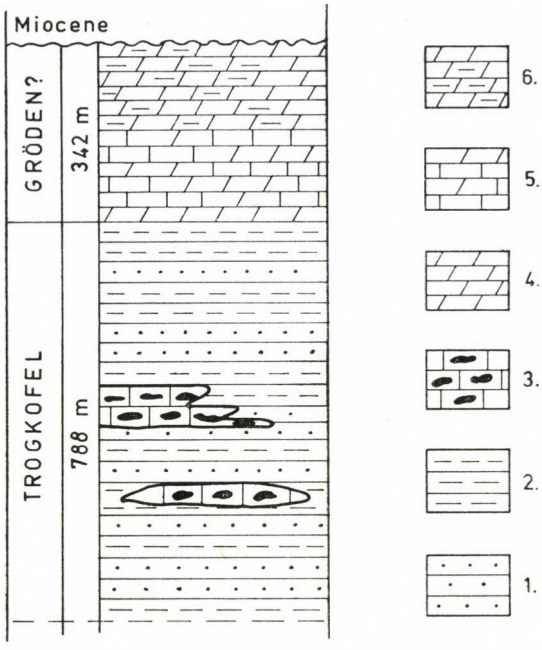

Fig. 2. Lithology of the Lower and Middle Permian beds in Újfalu. *1—3.* Lower Permian Trogkofel beds: *1.* pale-grey fine-grained sandstone; *2.* dark-grey shale; *3.* dark-grey brecciated limestone with Fusulinae. *4—6.* Middle Permian beds (Gröden equivalents): *4.* finely-crystalline dolomite; *5.* pale-grey dolomitic limestone; *6.* grey argillaceous dolomite

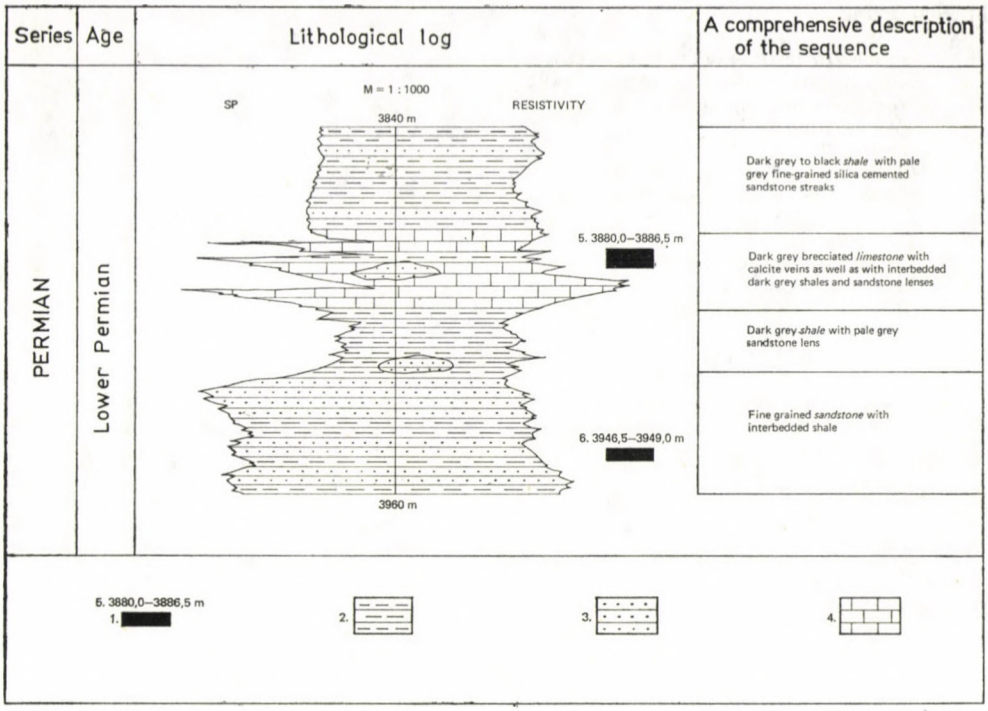

Fig. 3. The SP and resistivity logs of the marine Lower Permian in Újfalu between 3840.0—3960,0 m. 1. position, number and depth of the cores; 2. dark-grey shale; 3. fine-grained sandstone; 4. brecciated limestone

Algae

Tubiphytes obscurus MASLOV
Mizzia cornuta KOCHANSKY-DEVIDÉ et HERAK
Epimastopora sp.
Macroporella sp.
Gyroporella nipponica ENDO et HASHIMOTO

Fusulinidae

Nankinella sp.
Pseudoreichelina slovenica (KOCHANSKY-DEVIDÉ)
Schubertella australis THOMPSON et MILLER
Schubertella kingi DUNBAR et SKINNER
Schubertella cf. *paramelonica* SULEJMANOV
Schubertella sp.
Biwaella europaea KOCHANSKY-DEVIDÉ et MILANOVIĆ
Biwaella ? sp.
Paratriticites ? sp.
Darvasites contractus (SCHELLWIEN)
Darvasites sp.
Rugosochusenella sp.
Pseudofusulina ? sp.
Pseudoschwagerina sp.

Microforaminifera

Ammovertella inversa (SCHELLWIEN)
Neotuberitina reitlingerae (MIKL.-MAKL.)
Neotuberitina maljavkini (MIHAJLOV)
Neotuberitina sp.
Apterinella sp.
Climacammina elegans (MOELLER)
Climacammina cf. *rugosa* MOROZOVA
Climacammina sp.
Deckerella sp.
Tetrataxis nana MOROZOVA
Tetrataxis sp.
Globivalvulina parva CHERNUSEVA
Globivalvulina sp.
Glyphostomella cf. *triloculina* (CUSHMAN et WATERS)
Lasiodiscus minor REICHEL
Lasiodiscus irregularis MIKL.-MAKL.
Endothyridae gen. et sp. indet.

Accessorial elements

Bryozoa
Gastropoda embrional test
Echinoid fragments
Crinoid fragments
Baccinella sp.

The material of the core No. 5 can be well correlated both lithologically and paleontologically with the Darvasites Horizon of the "Trogkofel stage" in the Carnian Alps, Karavanke Mts., in Slovenia, Croatia and Montenegro (A. RAMOVŠ and V. KOCHANSKY-DEVIDÉ, 1965; E. FLÜGEL, 1966; W. HOMANN, 1972; V. KOCHANSKY-DEVIDÉ, 1969a, b, 1970, 1972; V. KOCHANSKY-DEVIDÉ et al., 1973). On this basis the clastic sequence can be correlated with the clastic Košna facies of the "Trogkofel stage" in the Lower Permian of the Dinarides, although in the shale and sandstone layers of the core No. 6, lying in a deeper position (3946.5—3649.0 m), was not found any fossil.

Contrarily, on its spatial position and lithological similarity, it is very probable, that the calcareous sequence intersected by the borehole Újfalu-I can be regarded as the counterpart of the Middle Permian dolomite series of the Velebit Mts. (Croatia), if it is not tectonically adjoined with its footwall. For the time being, however, the validity of this supposition can not be paleontologically evidenced owing to the lack of *Neoschwagerina*-finds characteristic of the Velebit sequence. In spite of this fact this dolomitic sequence may represent a calcareous facies replacing the Gröden sandstone of the Gröden stage in the Middle Permian.

In Hungary, Lower and Middle Permian formations of marine origin have not been known until recently. This first Hungarian occurrence associated to localities with sediments of similar age in Slovenia and Croatia (Fig. 1), however, corroborates the ideas about the continuation of the Late Palaeozoic

sediments of Dinaric type in Hungary. These ideas formulated earlier based on an occurrence of the marine uppermost Carboniferous at Karád, and on the occurrences of the marine Upper Permian from Central Transdanubia via Sári village up to the Bükk Mts. (North Hungary).

For a lenghty period, the term "Trogkofel limestone" was applied to the pale grey and pink reef formations of the Carnian Alps and Western Karavanke Mts. During the last 16 year period, however — starting from the initiative of RAMOVŠ (1963) — it has become evident (A. RAMOVŠ—V. KOCHANSKY-DEVIDÉ, 1965; V. KOCHANSKY-DEVIDÉ, 1973; A. RAMOVŠ, 1974) that in the Velebit Mts., and in the area from the Karavanke Mts. via Vitanje, Konjiška Gora, Boč, Žetale to Ravna Gora[1], this reef formation is in connection with a conglomerate—sandstone series having debris of older (Upper Carboniferous—Lower Permian) formations with fossils characteristic of the ages involved. The upper part of this series, however, has dark-grey to black, organic limestone beds and brecciated lenses of reefaceous origin, in form of intercalations. The microfauna and microflora of these intercalations are of the same age as that of the Trogkofel reef limestones. The Darvasites bearing clastic Permian sequence of the borehole Újfalu-I belongs to the group of occurrences of this type.

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[1] Zebanec is a bore-hole ENE from Ravna Gora, near the Drava river. Microfossil association has been found in the core 2472—2476 m in autumn 1980.

МОРСКИЕ НИЖНЕ- И СРЕДНЕПЕРМСКИЕ ОТЛОЖЕНИЯ В НЕФТЕПОИСКОВОЙ СКВАЖИНЕ УЙФАЛУ-I (ЮВ ВЕНГРИЯ)

А. БЕРЦИ-МАКК—В. КОХАНСКИ-ДЕВИДЕ

Резюме

При бурении скважины Уйфалу-I пройдена пермская толща мощности 1130 м. Нижняя 788-метровая часть толщи сложена преимущественно тонкозернистыми обломочными породами и она содержит доломитовые известняки с ископаемыми органическими остатками, а также рифогенные брекчиированные линзы только лишь в подчиненных пропорциях. На этом основании данная толща легко коррелируется с обломочной фацией Кошна «трогкофельского яруса» в Югославии и её можно отнести к верхней части нижней перми. Вышележащая доломитовая толща мощностью 342 м является, по-видимому, эквивалентом известковистой фации, замещающей среднепермские грёденские слои в Велебитах (Хорватия). Эта новая находка морской нижней и средней перми в Венгрии подтверждает высказанные раньше идеи о позднепалеозойском продолжении на территории Венгрии, выдвинутые исходя из присутствия морских отложений верхов каменноугольной систему у с. Карад и морских верхнепермских отложений в зоне, тянущейся от центральной части Задунайского края вплоть до гор Бюкк на северо-востоке страны.

Plate I

1. *Ammovertella inversa* (SCHELLWIEN) ×70
2. *Neotuberitina* sp. ×60
3. *Neotuberitina maljavkini* (MIHAJLOV) ×50
4. *a) Neotuberitina* cf. *maljavkini* (MIHAJLOV); b) *Apterinella* sp. — a-b: ×95
5. *Neotuberitina reitlingerae* (MIKL.-MAKL.) ×110
6. *Globivalvulina parva* CHERNUSEVA ×100
7. *Climacammina* cf. *rugosa* MOROZOVA ×70
8. *Deckerella* sp. ×54
9. *Globivalvulina ?* sp. ×100
10. *Climacammina elegans* (MOELLER) ×60

Plate II

1. *Lasiodiscus irregularis* MIKL.-MAKL. ×140
2. *Lasiodiscus minor* REICHEL ×100
3. *Tubiphytes obscurus* MASLOV ×45
4. a) *Tetrataxis ?* sp.; b) *Neotuberitina maljavkini* (MIHAJLOV); c) *Apterinella* sp. — a-b-c: ×45
5. *Apterinella* sp. ×45
6. *Tetrataxis* cf. *nana* MOROZOVA ×100
7. *Apterinella* sp. ×60

Plate III

1. *Pseudoreichelina slovenica* (KOCHANSKY-DEVIDÉ) ×53
2. *Schubertella* sp. ×100
3. *Endothyridae* gen. et sp. indet. ×150
4. a) *Schubertella kingi* DUNBAR et SKINNER; b) *Schubertella* sp.; c) *Apterinella* sp. — a-b-c: ×45
5. *Nankinella* sp. ×60
6. a) *Schubertella kingi* DUNBAR et SKINNER; b) *Schubertella australis* THOMPSON et MILLER; c) *Globivalvulina* sp.
7. *Gyroporella nipponica* ENDO et HASHIMOTO ×25
8. a) *Schubertella australis* THOMPSON et MILLER; b) *Apterinella* sp. — a-b: ×100

Plate IV

1. *Darvasites contractus* (SCHELLWIEN) ×10
2. *Darvasites ?* sp. ×40
3. a) *Darvasites* sp.; b) *Schubertella* sp. a-b: ×40
4. *Schubertella* cf. *paramelonica* SULEJMANOV ×45

Plate V

1. *Biwaella europaea* KOCHANSKY-DEVIDÉ et MILANOVIĆ ×50
2. *Biwaella europaea* KOCHANSKY-DEVIDÉ et MILANOVIĆ ×45
3. *Glyphostomella* cf. *triloculina* (CUSHMAN et WATERS) ×55
4. *Rugosochusenella* sp. ×43
5. *Paratriticites ?* sp. ×40
6. *Rugosochusenella* sp. ×35

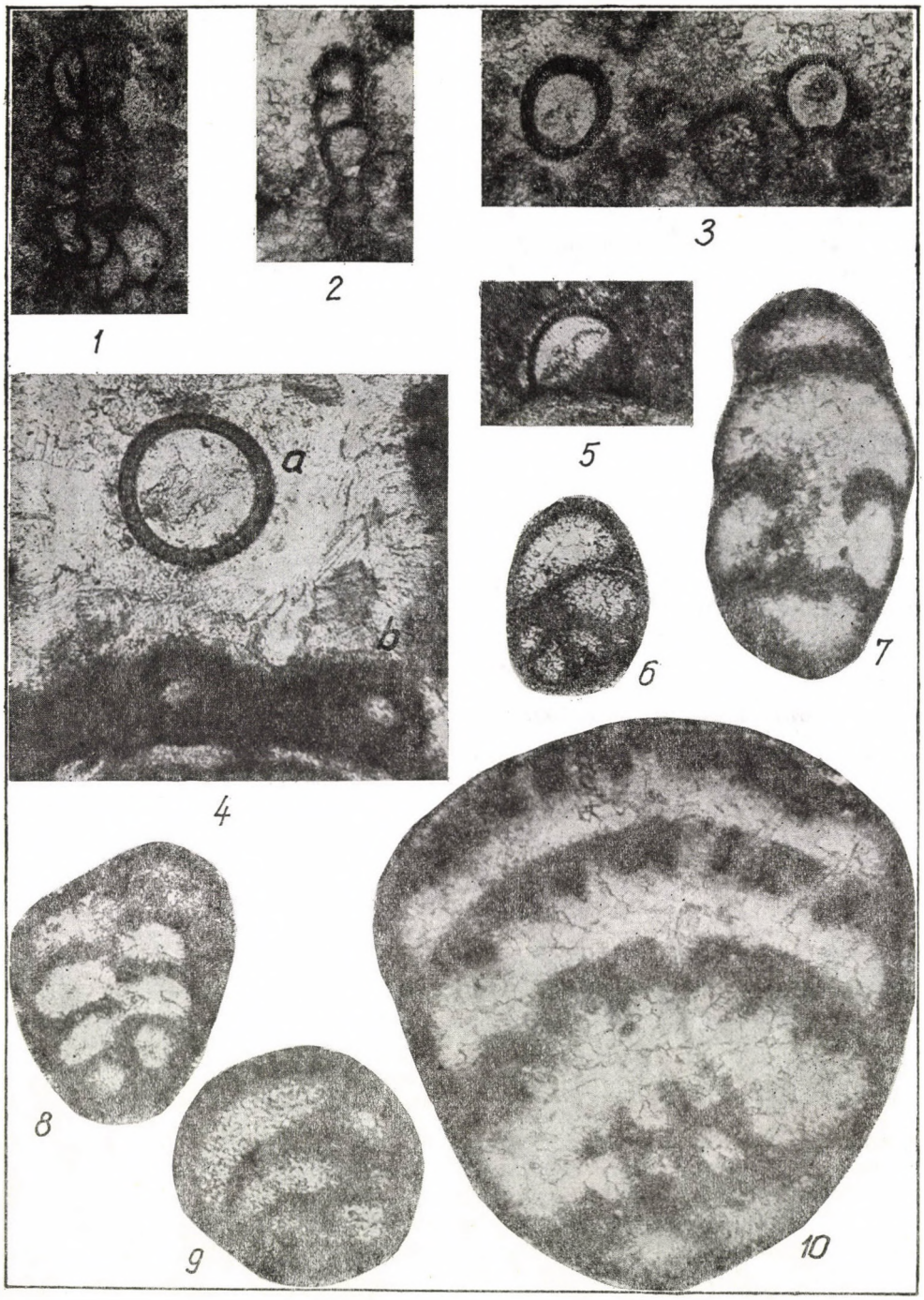

Plate I

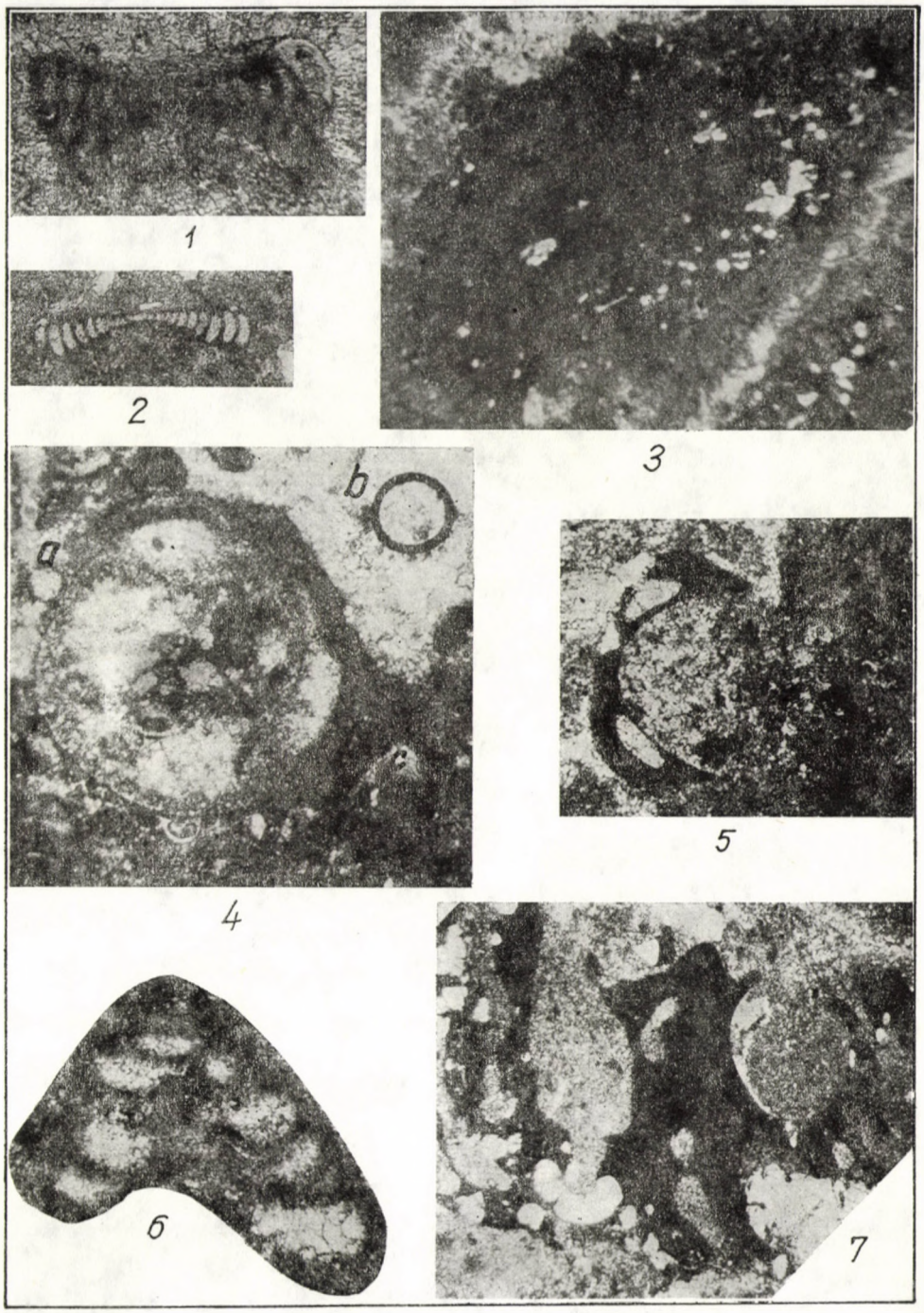

Plate II

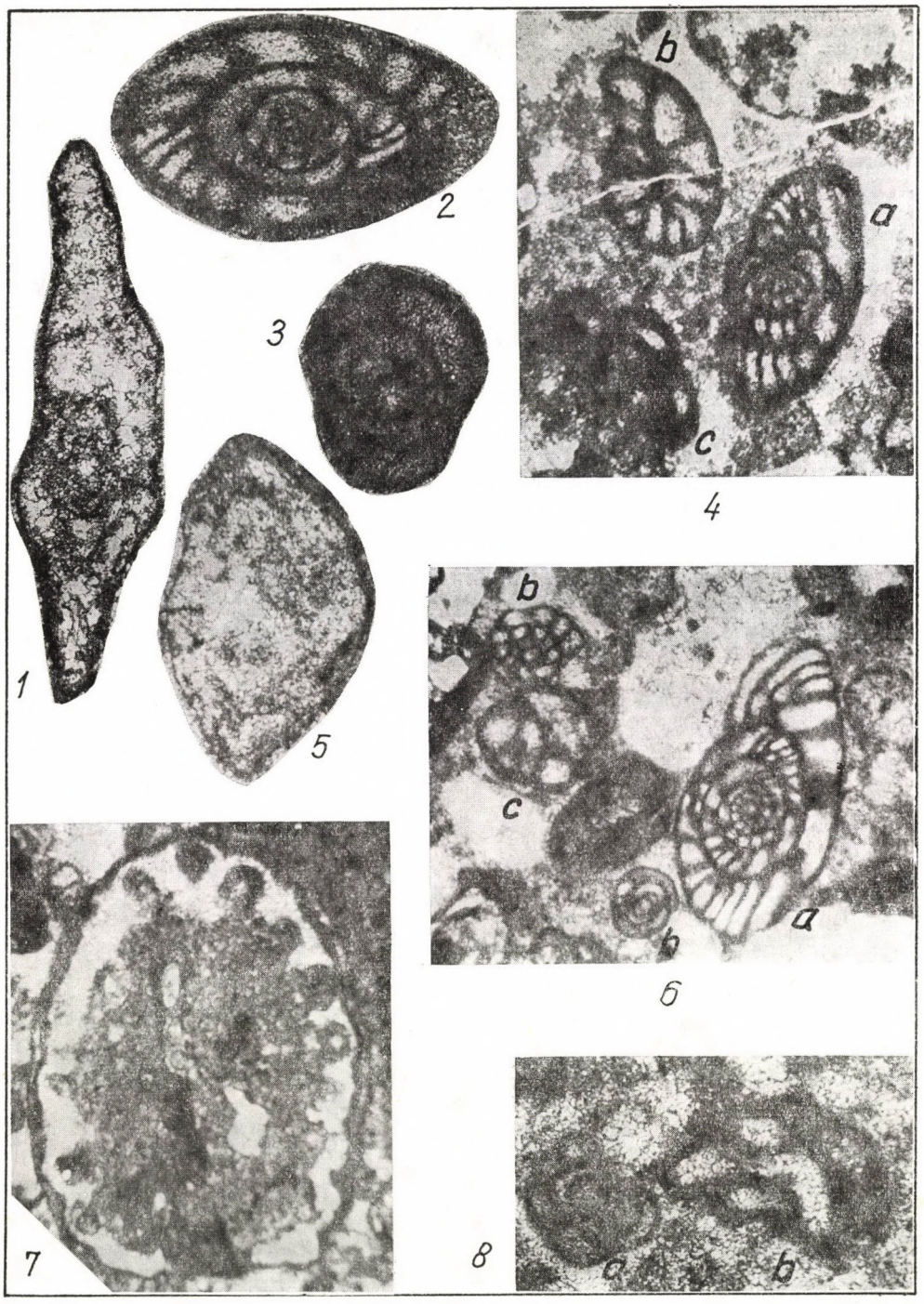

Plate III

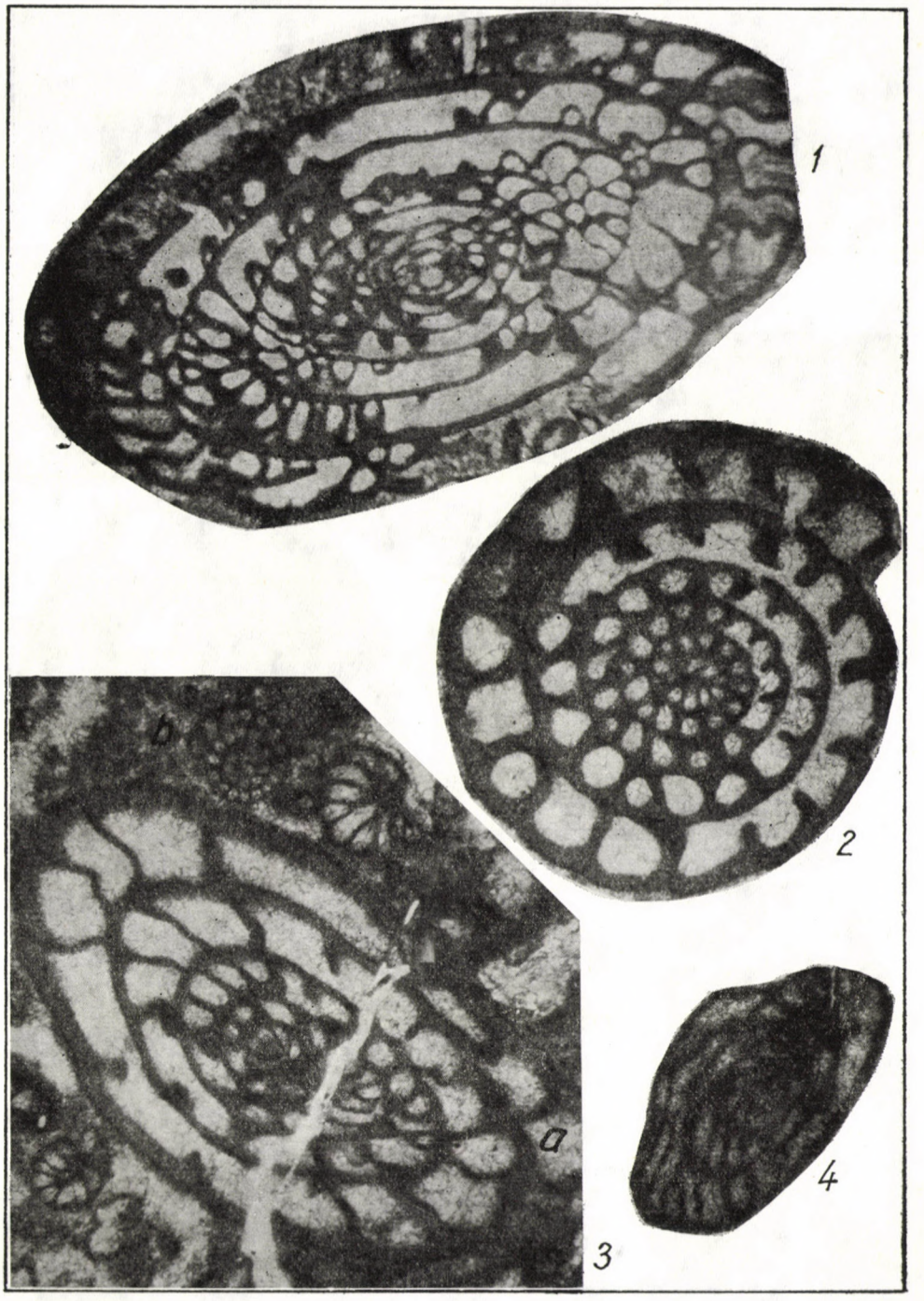

Plate IV

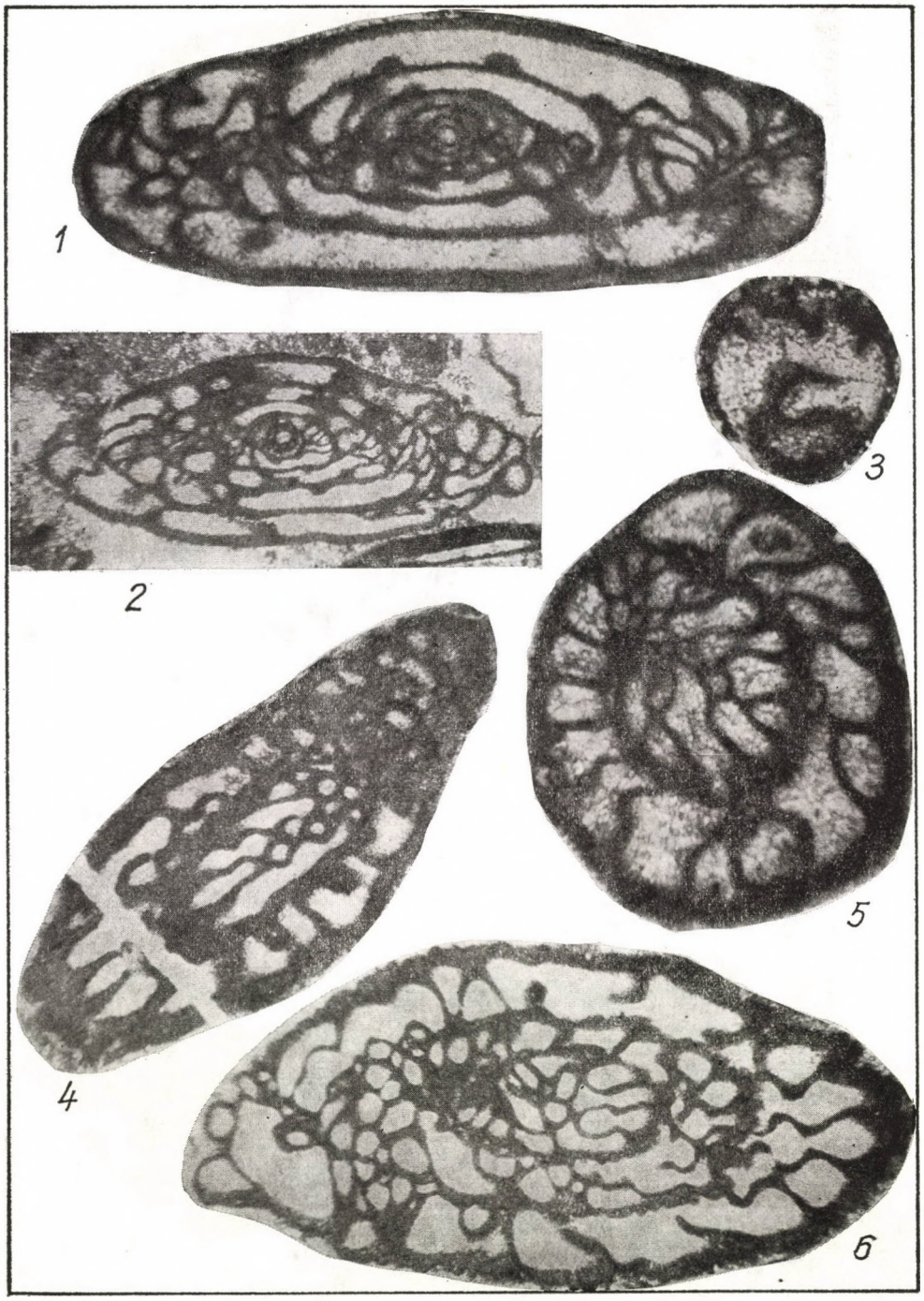

Plate V

Acta Geologica Academiae Scientiarum Hungaricae, Vol. 24 (1), pp. 129—147 (1981)

ZUR ANWENDUNG LITHO- UND BIOSTRATIGRAPHISCHER GLIEDERUNGEN IM ROTLIEGENDEN

Von

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HUNGARIAN GEOLOGICAL INSTITUTE, BUDAPEST

Die Anwendung litho- und biostratigraphischer Methoden im mitteleuropäischen Rotliegenden wird diskutiert. Für den Thüringer Wald wird die folgende lithostratigraphische Abfolge vorgeschlagen:

Abteilung	Gruppe	Formation	Formationsglied	Bank, Lage
Rotliegendes	Oberrotliegendes	Tambacher		
	Unterrotliegendes	Rotteröder		
		Oberhöfer		
		Goldlauterer		
		Manebacher	Z. B. flözführende »Zone«	Z. B. 1. Flöz
		Gehrener		

Bisher wurden einige lithostratigraphische Einheiten von manchen Autoren als biostratigraphische Einheiten angesehen. Daraus resultierten schwerwiegende Fehler bei der Korrelation des Rotliegenden mit Abfolgen in den Südalpen und den U.S.A. Biostratigraphische Datierungen lithostratigraphischer Einheiten berühren nicht die lithostratigraphisch fixierten Grenzen und verändern nicht den lithostratigraphischen Charakter dieser Einheiten in einen biostratigraphischen. Manche Autoren betrachten die Abteilung Rotliegendes als eine biostratigraphische Einheit, weil einige biostratigraphische Datierungen in dieser lithostratigraphischen Einheit durchgeführt wurden. Diese Autoren veränderten auch in einigen Fällen die Grenze zwischen dem Unter- und Oberrotliegenden wegen vager biostratigraphischer Daten (Tetrapodenfährten). Aber das Rotliegende bleibt selbstverständlich trotz irgendwelcher biostratigraphischer Datierungen eine lithostratigraphische Einheit. Die Begriffe Autunian und Saxonian sollten außerhalb Frankreichs nicht verwendet werden. Autunian und Saxonian stimmen nicht mit dem Unterperm bzw. Mittelperm überein. Ein großer Teil des Autunian gehört zum Karbon, unabhängig von den zahlreichen verschiedenen Definitionen des Autunian.

Die Gliederung und vor allem die Korrelation des kontinentalen Permokarbon (Rotliegenden) von Mitteleuropa gehört auch heute noch zu den schwierigsten Problemen der Stratigraphie des Phanerozoikum in Europa. Darüber können auch nicht scheinbar perfekte, oft mit großer Selbstsicherheit vorgetragene großräumige Korrelationen hinwegtäuschen, zumal diese z. T. mit faziell ähnlichen Sedimenten Nordamerikas vorgenommen wurden, deren biostratigraphische Datierung vielfach ebenfalls unsicher ist. Besonders die lithofaziell (einschließlich ihrer biofaziellen Ausbildung) sehr unbeständigen unterrotliegenden Sedimente mit ihrem raschen und starken horizontalen und vertikalen Fazieswechsel lassen sich selbst innerhalb eines großräumigen Ablagerungsgebietes oftmals nur schwer korrelieren. Für die bisher vorliegenden Korrelationen fehlen oftmals die faziesunabhängigen Bezugspunkte, mit denen man beweisen kann, ob eine als isochron angesehene lithostratigraphische Grenze auch wirklich isochron ist oder nur eine diachrone Faziesgrenze markiert. Als wichtige lithostratigraphische Marken werden feinklastische dunkle Schichtglieder, ausgedehntere Kohlenbildungen (z. B. in der oberen Manebacher Formation), Hornsteinlagen, Wechsel von grauen und roten Farbtönen, Konglomeratlagen, Vulkanite und Tuffhorizonte angesehen. Vor allem der Wechsel grauer und roter Farbtöne und die Konglomeratlagen sind über größere Entfernungen deutlich diachrone Marken. Die als Bezugshorizonte verwendeten Vulkanite sind z. T. intrusiv (vgl. KUNERT, 1978) und dann als lithostratigraphische Bezugshorizonte ungeeignet. Tuffhorizonte bieten in der oberen Goldlauterer und Oberhöfer Formation des Thüringer Waldes gute Markierungen für die lithostratigraphische Gliederung (vgl. ANDREAS in ANDREAS—HAUBOLD, 1975). Allerdings setzt die genaue Ansprache der im Gelände z. T. schwer auffindbaren Tuffhorizonte detaillierte petrographische Untersuchungen voraus. Die regionale Reichweite der Tuffkorrelierungen ist vergleichsweise gering, wobei in den Randgebieten der Verbreitung der betreffenden Tuffhorizonte die Unsicherheiten bei der richtigen Ansprache der Horizonte wachsen.

Die biostratigraphischen Grundlagenuntersuchungen stehen erst im Anfangsstadium. Die Mehrzahl der Fossilgruppen ist paläontologisch noch nicht oder nur unzureichend untersucht worden. Noch seltener liegen exakte biostratigraphische Aussagen vor, die nicht nur taxonomische Bearbeitungen und Revisionen, sondern auch die Kenntnis der Evolution der betreffenden Fossilgruppen voraussetzen, die eine unabdingbare, aber bisher meist nicht beachtete Grundlage der biostratigraphischen Gliederung des Rotliegenden ist. Die *Floren* fast aller wichtigen Rotliegendvorkommen Europas bedürfen der taxonomischen Revision (vgl. z. B. BARTHEL, 1976). Erst dann können für viele Arten genaue Reichweiten angegeben werden. Die biostratigraphisch wichtigen *Sporomorphen* werden künftig nicht nur die Gliederung großer Teile des Rotliegenden, sondern z. T. auch Korrelationen mit der marinen Gliede-

rung erlauben. So konnte mit Sporomorphen nachgewiesen werden, daß das Rotliegende nicht nur dem Unterperm entspricht, sondern im Oberkarbon beginnt (vgl. DOUBINGER, 1974) und bis ins Mittelperm reicht (vgl. KOZUR, 1978b). Da der mittelpermische Anteil des obersten Rotliegenden immer oberhalb der pfälzischen Bewegungen liegt, sind die bisher mit der saalischen Phase in Zusammenhang gebrachten Umstellungen der Paläogeographie in Wirklichkeit auf pfälzische Bewegungen zurückzuführen (z. B. die Bewegungen vor der Ablagerung der mittelpermischen Val Gardena-Formation). Die pfälzische Phase liegt also nicht zwischen Perm und Untertrias, sondern zwischen dem Unter- und Mittelperm. Auch die Altersstellung der Melaphyr-Gruppe der Choč-Decke (Westkarpaten) konnte durch neuere palynologische Untersuchungen geklärt werden (PLANDEROVÁ *in* VOZÁROVÁ—VOZÁR—BAJANÍK et al., 1979 und persönliche Mitteilung). Unmittelbar unterhalb der basischen Eruptiva der ersten Eruptionsphase tritt eine wohl asselische »Autun«-Assoziation (oberstes Karbon) auf. Zwischen der ersten und zweiten Eruptionsphase kommt eine eindeutig unterpermische Assoziation mit reichlich *Cordaitina* vor und in den postvulkanischen permischen Schichten oberhalb der zweiten Eruptionsphase finden sich mittel- oder oberpermische Sporomorphen. Der bisher als oberpermisch angesehene basische Vulkanismus der Choč-Decke beginnt somit nahe der Karbon/Perm-Grenze und endet innerhalb des Unterperm. Er besitzt damit das gleiche Alter wie die maximale vulkanische Aktivität in allen anderen tektonischen Einheiten der Westkarpaten sowie in den varistischen Molassebecken Mittel- und Westeuropas. Auch die Vulkanite des Mecsek-Gebirges (Gyűrűfű-Quarzporphyr-Formation) weisen nach Sporomorphen dieses Alter auf (persönliche Mitteilung von Frau BARABÁS-STUHL). Beträchtliche stratigraphische Bedeutung für die Gliederung des kontinentalen Permokarbon haben auch die *Blattodea* (Insecta) erlangt (vgl. MÜLLER, 1975, SCHNEIDER, 1978a, b). Besondere stratigraphische Bedeutung wird künftig auch den *Conchostraken* zukommen, bei denen wenig unterhalb der Karbon/Perm-Grenze in allen Rotliegendvorkommen Europas einschneidende Änderungen des Artenspektrums zu beobachten sind. Dabei wird die karbonische Assoziation mit *Pseudestheria limbata/L. lallyensis* (im oberen Teil vor allem *Pseudestheria ? autunensis*) durch eine Assoziation mit dominierend *Liostheria* n. sp. und anderen *Limnadoidea* abgelöst. Bei den *Muscheln* und *Ostracoden* (quantitativ häufigstes, aber bisher kaum beachtetes Faunenelement) als weitere biostratigraphisch bedeutsame Invertebraten des kontinentalen Permokarbon stehen taxonomische Grundlagenuntersuchungen noch aus. Mit Ausnahme der durch BOY (1972, 1976, 1978) ausgezeichnet bearbeiteten *Amphibien* gilt das auch für die Vertebraten, insbesondere für die häufigen und sicher biostratigraphisch verwertbaren *Fische*. Die *Tetrapodenfährten* erlangten u. a. durch die Arbeiten von MÜLLER (1954, 1959, 1970), SCHMIDT (1959), HAUBOLD (1971a, b, 1973, 1974) und FICHTER (1976) im Rotliegenden stratigraphische

Bedeutung. Sehr bedenklich ist aber die Verwendung hypothetischer Aussagen über die mögliche Zuordnung der Tetrapodenfährten zum osteologischen System für biostratigraphische Aussagen. HAUBOLD (1973) lehnt die Zuordnung eines Teils der Cornberger Tetrapodenfährten zu den Therapsida durch SCHMIDT (1959) mit dem Hinweis auf ein unterpermisches Alter des obersten Oberrotliegenden ab und HAUBOLD—KATZUNG (1975) schließen dann aus dem Fehlen von Therapsidenfährten im Cornberger Sandstein auf sein unterpermisches Alter und überhaupt auf ein unterpermisches Alter des obersten Rotliegenden — ein perfekter Kreisschluß. Die Datierung der Fährtenfaunen des Cornberger Sandsteins (oberstes Oberrotliegendes) durch SCHMIDT (1959) als unteres Oberperm (= Mittelperm) hat sich als richtig erwiesen (vgl. KOZUR, 1978b und in Druck a, b, c). Eine nur auf Fährtenfaunen beruhende überregionale Korrelation der Rotliegendsedimente mit Abfolgen außerhalb Mittel-Europas mußte zwangsläufig zu Fehlern führen, die überall dort klar sichtbar werden, wo die korrelierten Schichtenfolgen auch andere Fossilien enthalten (Südalpen, U.S.A.). So wurde die zuvor stets mit dem Oberrotliegenden korrelierte Val Gardena Formation der Südalpen bei HAUBOLD—KATZUNG (1975) in das »Autun« umgestuft, also mit dem Unterrotliegenden korreliert, obwohl in der mittleren Grödener Formation zahlreiche *Nautiloidea* sowie vereinzelte *Ammoniten* auftreten, die eine Einstufung in das höhere Mittelperm anzeigen. Die *Sporomorphen*-Assoziation, die aus dem gleichen stratigraphischen Bereich stammt wie die Tetrapodenfährten, enthält reichlich *Lueckisporites virkkiae* und zeigt damit an, daß sie wesentlich jünger als selbst das sehr stark nach oben ausgedehnte »Autun« im Sinne von HAUBOLD—KATZUNG (1972) ist. Die Umstufung dieser Schichten in das »Autun« ist umso unverständlicher, als die grundlegenden Sporomorphen-Untersuchungen durch KLAUS (1963) in diesem stratigraphischen Bereich eine der Grundlagenarbeiten für die Sporomorphen-Taxonomie und Stratigraphie des Oberperm (einschließlich des Zechsteins) darstellt. Während die untere Val Gardena-Formation mit dem oberen Oberrotliegenden korreliert werden kann, ist die mittlere und obere Val Gardena-Formation etwas jünger als selbst das oberste Rotliegende und dabei zumindest zum größten Teil etwas älter als der Zechstein (vgl. KOZUR, 1977, 1978b, in Druck a, b, c). Biostratigraphisch bedeutsam könnten auch noch die *Arthropodenfährten* des Rotliegenden werden. Die bisher aus verschiedenen Horizonten des Rotliegenden vorliegenden Funde (vgl. BOY, 1976, WALTER, 1978) weichen deutlich voneinander ab. Es muß geklärt werden, ob hier ökologische Faktoren eine Rolle spielen oder ob es sich um altersbedingte Unterschiede handelt. Im letzteren Falle hätten die Arthropodenfährten in Zukunft eine beträchtliche biostratigraphische Bedeutung. Viele andere hier nicht genannte Fossilgruppen, wie Charophyten, Hydromedusen und Reptilien können zwar wegen ihrer Seltenheit sicher nicht die Grundlage für biostratigraphische Gliederungen im Rotliegenden bilden, für einzelne stratigraphische

Bereiche und vor allem für die überregionale Korrelation des Rotliegenden aber durchaus bedeutsam sein bzw. werden.

Exakte biostratigraphische Datierungen der verschiedenen lithostratigraphischen Einheiten des Rotliegenden werden erst dann in großem Umfang möglich sein, wenn der *gesamte Fossilinhalt* des Rotliegenden einer grundlegenden paläontologischen Bearbeitung unterzogen wird und auf dieser Grundlage die stratigraphische (nicht die faziell bedingte) Reichweite der wichtigsten Fossilien aufgezeigt werden kann.

Auch die Auswertung paläomagnetischer Daten könnte vor allem im Oberrotliegenden und im kontinentalen Zechstein wichtige Angaben für chronostratigraphische Korrelationen liefern (vgl. DACHROTH, 1976). Zunehmende Bedeutung werden auch radiometrische Altersbestimmungen an den Vulkaniten des Rotliegenden erlangen.

Wegen des Mangels an verwertbaren biostratigraphischen Daten kann man in letzter Zeit verstärkt die Tendenz beobachten, daß biofazielle Daten der Lithostratigraphie wie biostratigraphische Daten zu Korrelationen verwendet werden. Dies führte zu Fehlkorrelationen. Besonders deutlich ist diese Tendenz in den Arbeiten von HAUBOLD und HAUBOLD—KATZUNG ausgeprägt. HAUBOLD (1977, S. 131, 132) sagte nämlich: »Die lithologischen Serien und Teilserien des Thüringer Waldes sind hinreichend gesicherte chronostratigraphische Einheiten ... biostratigraphische Abschnitte und besonders ihre Grenzen können nur in den kartierbaren chronostratigraphischen Einheiten und deren konkreten lithologischen Ausbildungen aufgezeigt werden. Die Lückenhaftigkeit und Faziesbindung der Fossilien verlangt aus praktischen Gründen eine „Vermengung" von lithostratigraphischen und biostratigraphischen Gliederungen.« — Lithostratigraphie wird hier praktisch mit Chronostratigraphie gleichgesetzt. Das ist theoretisch falsch und führt praktisches in den permokarbonischen Rotliegendablagerungen mit ihrem starken lateralen und vertikalen Fazieswechsel zu zahlreichen Fehlern. Als Beispiel seien die Datierungen der Schichtenfolge in den Lokalitäten Crock und Breitenbach (Thüringer Wald) genannt. Die flözführenden Schichten von Crock werden bei HAUBOLD (1977) aus lithostratigraphischen Erwägungen in die Goldlauterer Schichten (Formation) eingestuft. Sie sind aber nach biostratigraphischen Kriterien eindeutig ein zeitliches Äquivalent der oberen Manebacher Formation von Manebach und anderen Lokalitäten. Die Schichtenfolge von Breitenbach stufte HAUBOLD (1977) dagegen in die Manebacher Schichten (Formation) ein. Nach biostratigraphischen Kriterien sind diese Schichten aber nicht älter, sondern jünger als die flözführenden Schichten von Crock, d. h. jünger als die obere Manebacher Formation bei Manebach. Da die lithostratigraphisch korrelierten Einheiten als chronostratigraphische Einheiten angesehen werden, bedeutet dies, daß nun in der unteren Goldlauterer Formation (als »chronostratigraphische« Einheit) auch die Fossilien aus der Lokalität Crock auf-

geführt werden und in der oberen Manebacher Formation (ebenfalls als vermeintliche »chronostratigraphische« Einheit) auch die Fossilien der Lokalität Breitenbach. Dadurch entsteht nun für die obere Manebacher und die untere bis mittlere Goldlauterer Formation eine »biostratigraphische« Einheit, obwohl zwei faunistisch-floristische Einheiten enthalten sind und DOUBINGER (1956) die Grenze zwischen ihr Autunien inférieur und Autunien supérieur gerade zwischen diese beiden biostratigraphischen Einheiten gelegt hat, so daß hier sogar eine bedeutsame biostratigraphische Grenze innerhalb des Rotliegenden vorliegt. Da die lithostratigraphischen Einheiten praktisch als chronostratigraphische Einheiten angesehen werden und die biostratigraphischen Einheiten nicht nach den Änderungen des floristischen oder faunistischen Inhalts, sondern nach der Begrenzung eben dieser lithostratigraphischen Einheiten abgegrenzt werden, kann man nicht nachweisen, ob die Begrenzungen der ausgehaltenen lithostratigraphischen Einheiten im gesamten Ablagerungsgebiet zeitlich synchron oder diachron sind und Fehlkorrelationen, wie im Falle von Crock und Breitenbach können nicht erkannt werden. Wenn die Grenzen »biostratigraphischer« Einheiten mit den Grenzen derjenigen lithostratigraphischen Einheiten definiert werden, in denen sie enthalten sind, wie das in den Arbeiten von HAUBOLD und HAUBOLD—KATZUNG stets praktiziert wird und wie das bei HAUBOLD (1977) ausdrücklich gefordert wird, dann handelt es sich dabei nicht um biostratigraphische Einheiten, sondern es werden bestenfalls biofazielle Einheiten ausgeschieden und verfolgt. Das ist keine Abwertung der Methode oder der Bedeutung solcher Untersuchungen. Biofazielle Parameter sind wichtige Daten für die Charakterisierung lithostratigraphischer Einheiten. Dagegen fehlt solchen »biostratigraphischen« Einheiten, deren Unter- und Obergrenze mit den Grenzen der zuvor ausgeschiedenen oder auskartierten lithostratigraphischen Einheiten definiert werden, jede *biostratigraphische* Definition ihrer Grenzen und sie sind damit als biostratigraphische Einheiten ohne Existenzberechtigung. Sie können daher *nicht* für überregionale Korrelationen verwendet werden. ***Grundlage für die Biostratigraphie sind nicht die Fossilien an sich, sondern die Evolution der Fossilien***, ein Grundsatz, der nicht oft genug betont werden kann.

Bei der Aufeinanderfolge lithostratigraphischer Einheiten ist der zeitliche Umfang eventuell auftretender Lücken zwischen solchen Einheiten für deren lithostratigraphische Definition belanglos und er kann auch mit lithostratigraphischen Methoden nicht geklärt werden. Definiert man aber die Grenzen »biostratigraphischer« Einheiten mit den Grenzen derjenigen lithostratigraphischen Einheiten, in denen sie enthalten sind, so liegt gerade im Vorhandensein einer großen zeitlichen Lücke zwischen zwei lithostratigraphischen Einheiten eine große potentielle Fehlerquelle für die Korrelationen, die mit Hilfe solcher vermeintlichen biostratigraphischen Einheiten vorgenommen werden. In einem solchen Falle folgen nicht nur die lithostratigraphischen Ein-

heiten, sondern auch die »biostratigraphischen« Einheiten unmittelbar aufeinander, wie das auch ganz allgemein tabellarisch und in den Ausführungen zu diesen Gliederungen dargestellt wurde. Hier hilft auch nicht weiter, wenn bei klar erkennbaren Lücken (z. B. bei einer schwach diskordant auflagernden lithostratigraphischen Einheit) eine minimale Lücke zwischen zwei »biostratigraphischen« Einheiten angegeben wird, deren zeitlicher Umfang nach dem Ermessen des jeweiligen Autors angesetzt wird, da mit lithostratigraphischen Methoden nur das Vorhandensein einer Lücke, nicht aber ihre relative oder absolute Zeitdauer bestimmt werden kann. Schon allein deshalb können die lithostratigraphischen Einheiten des Thüringer Waldes keine »hinreichend gesicherten chronostratigraphischen Einheiten« sein, wie HAUBOLD (1977) annimmt. Auch hierfür ein konkretes Beispiel. Zwischen den jüngsten Schichtgliedern des Unterrotliegenden und der Eisenacher Formation des Thüringer Waldes liegt eine große zeitliche Lücke, die das obere Sakmarian, das gesamte Artinskian und das untere Leonardian (unteres Kungurian) umfaßt. In diesem langen Zeitraum haben je nach Fossilgruppe eine bis mehrere weitere wirkliche biostratigraphische Einheiten Platz und selbst in den kontinentalen Rotliegendsedimenten mit ihrem Mangel an kurzlebigen Fossilien befindet sich hier mindestens noch eine weitere biostratigraphische Einheit (*Supaia*-Flora des »Saxonien inférieur«, Tetrapodenfährten-Assoziation mit *Nanipes*, *Devipes*, *Diversipes* etc. aus der »zone de transition« und dem »Saxonien inférieur« des Lodève-Beckens). Für diese große biostratigraphische Einheit war aber zwischen den Oberhöfer Schichten s. l. (einschließlich der Rotteröder Formation) und der Eisenacher Formation kein Platz mehr und so paralellisierte HAUBOLD (1973) die Faunen und Floren der »zone de transition« und des »Saxonien inférieur« des Lodève-Beckens mit dem Bereich von der unteren Goldlauterer Formation bis zu den oberen Oberhöfer Schichten s. l., während HAUBOLD—KATZUNG (1975) unter Annahme einer zeitlich geringfügigen Lücke zwischen der Rotteröder und Tambacher Formation die »zone de transition« in die obere Goldlauterer bis untere Oberhöfer und das »Saxonien inférieur« größtenteils in die obere Oberhöfer und Rotteröder Formation und zu einem geringen Teil in die Lücke zwischen der Rotteröder und Tambacher Formation einstuften. Als Konsequenz dieser Korrelierung wurden bei HAUBOLD—KATZUNG (1975) auf Tabelle 12 die lithostratigraphischen Einheiten des Thüringer Waldes als »Zeitskala« (!) benutzt und *Nanipes* für den Bereich der oberen Goldlauterer bis zur Rotteröder Formation sowie *Devipes* und *Diversipes* für den Bereich der oberen Goldlauterer bis zur basalen Oberhöfer Formation angegeben, wo sie gar nicht vorkommen. Eine Folge davon wiederum ist die Korrelation der Hermit Shale von Arizona mit der oberen Goldlauterer und basalen Oberhöfer Formation, obwohl der Hermit Shale viel jünger ist.

Durch die falsche Anwendung biofazieller Einheiten als »biostratigra-

phische« Einheiten könnte der Eindruck entstehen, daß die Biostratigraphie ein untaugliches Mittel für die Korrelation der kontinentalen permokarbonischen Ablagerungen Mitteleuropas sei, da diese angeblich biostratigraphischen Korrelationen in den meisten Fällen fehlerhafter sind als die bisherigen Vorstellungen zur Altersstellung der betreffenden Schichten. In Wirklichkeit beruhen solche Fehler in den angeblich biostratigraphischen überregionalen Korrelationen aber gerade auf der Vernachlässigung wichtiger biostratigraphischer Grundlagen und der Verwendung lithostratigraphischer (biofazieller, irrtümlich als »biostratigraphisch« angesehener) Einheiten für überregionale Korrelationen, was schon laut Definition der Lithostratigraphie nicht möglich ist.

Um die nomenklatorische Verwirrung bei der Zuordnung stratigraphischer Einheiten im Rotliegenden abzubauen und künftig jene Fehlkorrelationen, die nur auf der Nichtbeachtung stratigraphischer Grundregeln beruhen, zu vermeiden (der gegenwärtige Kenntnisstand läßt ohnehin selbst bei genauer Beachtung der stratigraphischen Richtlinien noch viel Raum für Fehlkorrelationen), wurde von KOZUR (April 1978a, Anhang zu 1978d, in Druck a, c) eine stricte nomenklatorische Trennung von lithostratigraphischen und biostratigraphischen Einheiten durchgeführt. Für den Thüringer Wald wurde eine Einteilung in Gehrener, Manebacher, Goldlauterer, Oberhöfer, Rotteröder und Tambacher Formation sowie der Eisenacher Formation für das getrennte Sedimentationsgebiet des Eisenacher Rotliegenden vorgenommen, wobei die von BACKHAUS u. a. (1977) gegebene Empfehlung, bei den Formationsbezeichnungen an altbewährten Namen festzuhalten, berücksichtigt wurde. Als Formationsglieder (vielleicht sollte man hier die kürzere englische Bezeichnung *member* verwenden) könnten z. B. in der Manebacher Formation das Grundkonglomerat, die untere Sandstein-»Zone«, die flözführende »Zone« und die obere Sandstein-»Zone« fungieren, als Bank, Lage z. B. die einzelnen Flöze oder Zwischenmittel der Flöz member (flözführende »Zone«). Gehrener bis Rotteröder Formation gehören zur Unterrotliegend-Gruppe der Rotliegend-Abteilung. Nach der biostratigraphischen Datierung gehört z. B. die Goldlauterer Formation zum Karbon-System (Oberkarbon-Serie, Asselian-Stufe) und die Rotteröder Formation zum Perm-System (Unterperm-Serie, Sakmarian-Stufe). HAUBOLD—KATZUNG (Oktober 1978) legten ebenfalls einen Vorschlag zur Anwendung stratigraphischer Richtlinien im Rotliegenden vor. Leider sind den Autoren dabei eine ganze Reihe von Fehlern unterlaufen, die schon in ihren vorangegangen Arbeiten zu beobachten waren. Unter Berufung auf BACKHAUS u. a. (1977), schreiben HAUBOLD—KATZUNG (1978, S. 439): »Die Zone umfaßt demnach keineswegs die Lebensdauer einer Art.« Das ist in dieser absoluten Fassung falsch und wurde bei BACKHAUS u. a. (1977) auch nicht in dieser Form publiziert. Eine *range-Zone* — und die exaktesten biostratigraphischen Zonen sind range-Zonen — umfaßt sehr wohl die Lebens-

dauer einer Art, nur ist es in Ermangelung kurzlebiger, weit verbreiteter Leitformen nicht immer möglich, mit range-Zonen zu arbeiten. Sie können dann z. B. durch *concurrent range-Zonen* ersetzt werden, wo die Zone nicht durch den Lebensbereich einer Art, sondern durch den Bereich des gemeinsamen Vorkommens zweier oder mehrerer Arten definiert wird. Das Beispiel, daß HAUBOLD—KATZUNG (1978, S. 439—400) gegen die Definition einer Zone mit der Lebensdauer einer Art anführen, ist allerdings sehr leichtfertig: »Bei dieser mitunter diskutierten Praktik wäre das Rotliegende — wenig sinnvoll! — als *Callipteris conferta*-Zone zu bezeichnen.« Von einer *Callipteris conferta*-Zone, die das gesamte Rotliegende umfaßt, könnte auch bei einer *C. conferta* range-Zone niemals die Rede sein, da *C. conferta* ab der Tambacher Formation des Thüringer Waldes nicht mehr vorkommt.

Als Formationen sehen HAUBOLD—KATZUNG (1978) z. B. die unteren Manebacher, die oberen Manebacher, die unteren Oberhöfer und die oberen Oberhöfer Schichten an. Das hat u. a. den Nachteil, daß z. B. eine obere Oberhöfer Formation so zweideutig ist (oberer Teil der Oberhöfer Formation oder, wie im Sinne von HAUBOLD—KATZUNG, 1978 — eine Formation, die die oberen Oberhöfer Schichten umfaßt), daß für alle diese Formationen neue Namen geschaffen werden müßten. Diese Flut von neuen Namen würde keines der ohnehin zahlreichen Probleme der Rotliegendstratigraphie lösen, dafür aber zahlreiche neue Probleme aufwerfen; sie würde zur Instabilität der eingebürgerten lithostratigraphischen Gliederung führen. Die als Abteilung von HAUBOLD—KATZUNG (1978) vorgesehenen Begriffe Graumolasse, Buntmolasse und Rotmolasse sind mehr oder weniger genetische Bezeichnungen und daher zur Bezeichnung lithostratigraphischer Einheiten ungeeignet. Sie sind auch überflüssig, da mit dem Rotliegenden bzw. dem Unter- und Oberrotliegenden bereits übergeordnete lithostratigraphische Begriffe existieren. Diese Begriffe aber finden sich bei HAUBOLD—KATZUNG (1978) unter der Rubrik »Biostratigraphie«. Während es sich bei den lithostratigraphischen Einheiten im Sinne von HAUBOLD—KATZUNG (1978) mit Ausnahme der ungeeigneten Bezeichnungen Grau-, Bunt- und Rotmolasse um Fragen der Ranghöhe schon bestehender lithostratigraphischer Einheiten handelt, unterliefen HAUBOLD—KATZUNG (1978) bei den biostratigraphischen Einheiten grundsätzliche Fehler. Ein Subsystem Rotliegendes kann doch wohl kein ernst zu nehmender Beitrag zur geowissenschaftlichen Forschung sein, allein schon deshalb, weil dieses Subsystem des Perm das höhere Oberkarbon bis Mittelperm umfaßt und aus der Einführung *eines* Subsystems konsequenterweise die Einführung eines *zweiten* Subsystems resultiert, für das nur der Zeitraum des Oberperm s. str. (also einer Serie) und die Bezeichnung Zechstein übrig bliebe. Davon ganz abgesehen ist das Rotliegende seit seiner Originaldefinition (von VELTHEIM, 1823—1829, vgl. KOZUR, 1978d) eine lithostratigraphische Einheit. Daran ändert auch nichts die Tatsache, daß sich viele Forscher mit

der Frage der Datierung des Rotliegenden umfassen. Keine lithostratigraphische Einheit hat einen unbegrenzten chronostratigraphischen Umfang. Doch keine biostratigraphische Datierung macht aus einer lithostratigraphischen Einheit eine biostratigraphische Einheit. Als Beispiel seien die Reiflinger Kalke der alpinen Trias genannt. Sie stellen als lithostratigraphische Einheit einen bestimmten Faziestyp dar (Beckenfazies), auf den hier nicht eingegangen werden soll. Ihre biostratigraphische Datierung ist u. a. mit Hilfe von *Conodonten* bei Vorliegen entsprechender Untersuchungen bis ins Detail möglich. Aus dem Vorliegen der lithostratigraphischen Einheit Reiflinger Kalk bzw. aus dem lateralen und vertikalen Wechsel mit anderen lithostratigraphischen Einheiten, wie z. B. dem Wettersteinkalk, lassen sich wichtige Rückschlüsse auf die paläotektonische Position eines bestimmten Gebietes, Beckenkonfiguration etc. ziehen. In den verschiedenen Gebieten kann der Reiflinger Kalk je nach der paläotektonischen Position des untersuchten Gebietes durchaus verschiedene stratigraphische Reichweiten haben. Er kann nur das Unterladin umfassen, er kann aber auch bis zum Unterkarn reichen. Dagegen gibt es keine oberkarnische oder jüngere bzw. mittelanisische oder ältere Reiflinger Kalke. Der Reiflinger Kalk hat also einen viel engeren chronostratigraphischen Umfang als das Rotliegende und er kann mit einer Genauigkeit biostratigraphisch datiert werden, wie das im Rotliegenden wohl in absehbarer Zeit nicht möglich sein wird. Doch niemand wird auf die Idee kommen, die Reiflinger Kalke als biostratigraphische Einheit, etwa als Stufe zu bezeichnen, obwohl sie maximal kaum einen größeren Zeitraum repräsentieren als eine triassische Stufe. Die gleiche Problematik liegt beim Rotliegenden vor. Wenn HAUBOLD—KATZUNG (1978, S. 440) schreiben: »Bei der Benennung dieser Abteilungen ist zunächst zu berücksichtigen, daß das Rotliegende zumindest seit dem 2. Heerlener Karbonkongreß biostratigraphisch definiert ist (JONGMANS—GOTHAN, 1937), die Bezeichnungen Unterrotliegendes und Oberrotliegendes für die mittlere und obere Abteilung also nicht verwendet werden können«, dann entspricht das nicht den Tatsachen. Bei JONGMANS—GOTHAN (1937) wird in einem zusammenfassenden Diskussionsbeitrag u. a. auch das Problem der Permbasis im kontinentalen Permokarbon diskutiert. Einleitend zu dieser Problematik schreiben JONGMANS—GOTHAN (1937, S. 24, 25): »GOTHAN leitet diese Diskussion mit den folgenden Bemerkungen ein: Auch von der Grenze Stephan-Perm ist bereits beim vorigen Kongreß die Rede gewesen. Die Schwierigkeiten für die Paralellisierung des Perms im ganzen liegen zum großen Teil — auch die Begrenzung des Perms nach oben — darin begründet, daß die Charakterisierung dieser Formation ihren Ausgang von Mitteleuropa genommen hat, wo wir die fazielle Zweiteilung in einen unteren Teil, das Rotliegende, und in einen ganz anders entwickelten, oberen Teil haben, den Zechstein . . . Man kann da noch gar keine allgemeinen Regulationen geben, und jeder wird von Fall zu Fall nach bestem Willen und Gewis-

sen und unter Ausnutzung der erreichbaren Tatsachen sein Urteil fällen müssen.« Von einer biostratigraphischen Definition des Rotliegenden kann also gar keine Rede sein, nicht einmal eine exakte biostratigraphische Datierung des Rotliegenden wurde vorgenommen, die auch nichts am Charakter des Rotliegenden als lithostratigraphische Einheit geändert hätte. Es wird von einer faziellen Zweiteilung des Perms in Mitteleuropa gesprochen, das Rotliegende also ausdrücklich als *fazielle* und damit — wie schon zuvor und danach üblich — als lithostratigraphische Einheit angesehen. Jede allgemeine Festlegung zur Datierung und Korrelation des Rotliegenden und Zechstein wird entsprechend dem damaligen Kenntnisstand bewußt vermieden. Spätere Versuche einer biostratigraphischen Datierung des Rotliegenden oder von Teilen desselben ändern nichts an der Abgrenzung des Rotliegenden und schon gar nicht am Charakter als lithostratigraphische Einheit. Sonst könnte es im Phanerozoikum bald keine lithostratigraphischen Einheiten mehr geben, weil diese eines Tages u. a. mit biostratigraphischen Methoden datiert wurden oder werden, ohne daß sie deshalb zu biostratigraphischen Einheiten werden. Die konventionelle lithostratigraphische Abgrenzung des Rotliegenden bzw. des Unter- und Oberrotliegenden, die in ihrem bisherigen Umfang in den meisten Gebieten bereits seit langem auskartiert sind, wird von biostratigraphischen Datierungen nicht berührt. Daher ist z. B. jede Umstufung traditionell als Oberrotliegendes angesehener Schichten ins Unterrotliegende wegen vermeintlicher oder wirklicher Altersgleichheit mit dem Unterrotliegenden in einem anderen Ablagerungsgebiet abzulehnen, da jede Definition lithostratigraphischer Einheiten von ihrem Alter unabhängig ist. Um wieder bei dem oben erwähnten Beispiel aus der alpinen Trias zu bleiben: Es wäre widersinnig und der angestrebten geologischen Aussage völlig abträglich, wenn man einen oberladinischen Wettersteinkalk als Reiflinger Kalk bezeichnen wollte, nur weil an anderen Stellen die Reiflinger Kalke auch das Oberladin umfassen. Ebenso falsch und der geologischen Auswertung abträglich wäre es, die traditionell als Oberrotliegendes angesehene Hornburger Formation (wichtigster Teil des Saxonian in der ursprünglichen Definition des Saxonian in dessen Typusregion bei DE LAPPARENT, 1893, vgl. KOZUR, 1978d) in das Unterrotliegende (»Autun«) umzustufen, wie das HAUBOLD (1973) auf Grund von Fährtenfaunen vorschlug. Die als voroberrotliegende Phase definierte saalische Phase wurde von STILLE eindeutig an die Basis der zum Oberrotliegenden gezählten Hornburger Formation gestellt (vgl. KUNERT, 1970). Eine Umstufung der Hornburger Formation in das »Autun« (Unterrotliegende) würde die saalische Phase in ihrer Typusregion entweder in eine inner-unterrotliegende Phase umfunktionieren oder sie müßte entgegen den Vorstellungen von STILLE in ihrer Typusregion in ein höheres Niveau verlegt werden. Ganz unabhängig davon, was man von der saalischen Phase hält (vgl. KOZUR, in Druck a, c) kann man solche leichtfertigen Änderungen der Grenzen lithostratigra-

phischer Einheiten wegen biostratigraphischen Datierungen nicht vornehmen und schon gar nicht dort, wo diese lithostratigraphischen Einheiten Grundlage für die Definition einer tektonischen Phase sind. Genauso falsch wäre es zu fordern, die Oberhöfer und Rotteröder Formation ins Oberrotliegende zu stellen, nur weil altersgleiche Schichten im Saar-Nahe-Gebiet schon ins Oberrotliegende eingestuft wurden, als im Thüringer Wald noch gar keine Rotliegendgliederung vorlag. Desgleichen kann man nicht die seit vielen Jahrzehnten auskartierte und für das Saar-Nahe-Gebiet praktikable Unter- /Oberrotliegend-Grenze so verlegen, daß sie mit derjenigen des Thüringer Waldes zeitlich übereinstimmt, was ohnehin nie exakt zu realisieren wäre, da zwischen dem Unter- und Oberrotliegenden im Thüringer Wald eine große zeitliche Lücke klafft.

Als biostratigraphische Serien schlagen HAUBOLD—KATZUNG (1978) das »Autun« (Unterrotliegendes) und »Saxon« (Oberrotliegendes) vor. Unter- und Oberrotliegendes als lithostratigraphische Gruppen können, wie oben ausgeführt, niemals biostratigraphische Serien sein. Auch »Autun« und »Saxon sind als biostratigraphische Serienbezeichnungen völlig ungeeignet. Es kann und wird im Perm nur die Serien Unterperm, Mittelperm und Oberperm bzw. bei Zusammenfassung des Mittel- und Oberperm eine Zweigliederung in Unterperm und Oberperm s. l. geben. Autunian und Saxonian sind unklar definierte Einheiten, die ursprünglich als lithostratigraphische Einheiten ausgeschieden, später aber teils in einem sehr vagen biostratigraphischen Sinne, teils im lithostratigraphischen Sinne als Äquivalente des Unter- und Oberrotliegenden verwendet wurden, wobei die Grenze zwischen Autunian und Saxonian häufig an eine als »saalisch« bezeichnete Diskordanz oder deutliche Schichtlücke gelegt oder mit dem Einsetzen durchweg roter Sedimente festgelegt wurde. Wie bei KOZUR (1978d und in Druck c) ausgeführt wurde, existieren je nach Autor recht unterschiedliche Fassungen von Autunian und Saxonian, auch bei den gleichen Autor können sehr große Unterschiede in der Fassung dieser Einheiten auftreten. So liegt z. B. nach HAUBOLD—KATZUNG (1975) die Obergrenze des »Autun« in den Südalpen an der Obergrenze der Val Gardena Formation und damit oberhalb des »Saxon« von Mitteleuropa obwohl die »Autun«-Obergrenze in beiden Gebieten nach den gleichen Kriterien (Tetrapodenfährten) bestimmt wurde. Die Grenze zwischen den lithostratigraphischen Einheiten Autunien und Saxonien, die DE LAPPARENT (1893) bei seiner ursprünglichen Definition des Saxonien angab, lag zufällig in den meisten Profilen an einer biostratigraphischen Grenze. Diese Autunien/Saxonien-Grenze ist die einzige bisher vorgeschlagene Grenze zwischen diesen beiden Einheiten, die in verschiedenen Gebieten biostratigraphisch erkennbar und damit biochronologisch korrelierbar ist. Sie wurde bei ANDREAS—HAUBOLD (1975) in Unkenntnis der Arbeit von DE LAPPARENT (1893) als Grenze zwischen unteren und oberen »Autun« angesehen, also auch von diesen Autoren als biostratigraphi-

sche Grenze anerkannt. Dagegen ist die von HAUBOLD—KATZUNG (1972) festgelegte »biostratigraphische« Autun/Saxon-Grenze mit einem »Typusprofil« im Thüringer Wald keine biostratigraphisch erkennbare Grenze. Sie beruht auf dem Neueinsetzen von *Tambachichnium schmidti* und *Palmichnus tambachensis* in der Lokalität Bromacker bei Tambach. Aus der riesigen Fährtenfülle der Lokalität Bromacker liegen nur insgesamt 4 (? 3) Fährtenplatten dieser beiden Arten vor. Ohne Zweifel wären sie wohl kaum je gefunden worden, wenn die Lokalität Bromacker nicht über Jahrzehnte hinweg solche großen Mengen wohlerhaltener Fährten geliefert hätte. Alle jene häufigen Fährten aus der Tambacher Formation kommen aber schon im »Autun« im Sinne von HAUBOLD—KATZUNG vor. Beim Fehlen der überaus seltenen, möglicherweise lokalen Formen *T. schmidti* und *P. tambachensis* müßten alle gleichaltrigen Faunen aber ins »Autun« eingestuft werden. So außerordentlich seltene Arten, die nur auf eine einzige Lokalität beschränkt sind und selbst dort zu allerseltensten Zufallsfunden gehören, erfüllen natürlich in keiner Weise die Anforderungen an ein Leitfossil, denn sonst könnte man auch *Archaeopteryx lithographica* als Leitform für *Malm zeta* ansehen. Solange die von HAUBOLD—KATZUNG als Leitformen ausgewählten Arten nur aus einer Lokalität bekannt sind, können sie nicht zur Definition der »Saxon«-Basis im Sinne von HAUBOLD—KATZUNG (1972) wendet werden; es könnte sich theoretisch sogar um außerordentlich seltene endemische Formen handeln. Wie die meisten anderen Autunian/Saxonian-Abgrenzungen, hat auch jene nach HAUBOLD—KATZUNG (1972) keinerlei biostratigraphische Absicherung, nur wurde dies von den anderen Autoren im Unterschied zu HAUBOLD—KATZUNG (1972 und folgende Arbeiten) auch nicht behauptet. Selbst wenn man die einzige biostratigraphisch korrelierbare Autunian/Saxonian-Grenze, d. h. die ursprüngliche, bei DE LAPPARENT (1893) definierte Grenze, bei der Datierung von Autunian und Saxonian zugrunde legt, wären Autunian und Saxonian nicht als biostratigraphische Serienbezeichnungen geeignet, weil sie weder in dieser noch in irgendeiner anderen publizierten Fassung mit dem Unter- bzw. Mittelperm übereinstimmen. In seiner Typusregion (Autun-Becken, Frankreich) gehört das Autunian zum höheren Gzhelian und Asselian und damit ausschließlich oder teilweise (je nach der Zuordnung des Asselian zum Oberkarbon oder Unterperm) zum Oberkarbon. In der Fassung bei HAUBOLD—KATZUNG setzt es im Mittleren Gzhelian, also ebenfalls weit innerhalb des Oberkarbon, ein. Das Autunian hat lediglich in Frankreich als Lokalbezeichnung eine Berechtigung (vgl. auch BOUROZ, 1978, S. 496). Außerhalb Frankreichs sollte es ebenso wenig verwendet werden, wie die Bezeichnung Saxonian, schon allein deshalb, weil beide Begriffe so hoffnungslos »verdefiniert« sind, daß man stets Autor, Publikationsdatum und Ablagerungsgebiet angeben muß, wenn man wissen will, welche der zahlreichen unterschiedlichen Fassungen des Autunian und Saxonian gemeint ist, da der Umfang des Autunian und Saxonian je nach

Autor unterschiedlich definiert wird und auch bei den gleichen Autoren in verschiedenen Publikationen und vor allem in verschiedenen Ablagerungsgebieten unterschiedliche Fassungen des Autunian und Saxonian verwendet werden. Außerdem können diese Begriffe keinen Platz in der biostratigraphischen Hierarchie beanspruchen, da es sich weder um permische Serien noch um Stufen handelt. Wenn HAUBOLD—KATZUNG (1978) ihr »Typusprofil« der »Autun«/»Saxon«-Grenze (im Sinne von HAUBOLD—KATZUNG, 1972) im Thüringer Wald in ein Referenzprofil umfunktionieren, so muß man dazu sagen, daß es weder die Ansprüche an ein Typusprofil (die Typusregionen des Autunian und Saxonian sind bekannt und liegen in anderen Gebieten), noch an ein Referenzprofil erfüllt. Ein biostratigraphisches Referenzprofil soll die Grenzen biostratigraphischer Einheiten markieren. Das ist aber nur möglich, wenn in einem möglichst vollständigen Profil die biostratigraphischen Einheiten unmittelbar aufeinander folgen. Selbst wenn man bei kontinentalen Abfolgen hinsichtlich der Vollständigkeit der Profile oft einige Abstriche machen muß, kann man feststellen, daß im Thüringer Wald gerade im kritischen Bereich zwischen dem Unterrotliegenden (»Autun«) und Oberrotliegenden (»Saxon«) eine große Schichtlücke existiert, die mehr als eine permische Stufe umfaßt. Darüber hinaus wäre das im Sinne von HAUBOLD—KATZUNG (1972) biostratigraphisch faßbare »Saxon« nur im Steinbruch Bromacker durch saxonische Indexfossilien charakterisiert. Dort aber ist die nächst ältere biostratigraphische Einheit, die abgegrenzt werden soll, nicht aufgeschlossen. Selbst in einem kompletten Grenzprofil, das sowohl reiche Fossilfunde im obersten Unterrotliegenden als auch im tieferen Oberrotliegenden (einschließlich der von HAUBOLD—KATZUNG ausgewählten »Leitformen« für das untere Saxonian, siehe oben) enthielte und das im Thüringer Wald nirgends existiert, wäre der oberste Fossilhorizont des Unterrotliegenden noch durch eine mächtige fossilfreie Schichtenfolge des unteren Oberrotliegenden vom ersten Fossilhorizont des Oberrotliegenden getrennt (zusätzlich zu der großen Schichtlücke unterhalb der Tambacher Formation). Selbst ein solches ideales, im Thüringer Wald nicht existierendes Profil wäre als Referenzprofil auf Grund der geologischen Situation gänzlich ungeeignet. Ein aus lithostratigraphisch korrelierten Teilprofilen zusammengesetztes Gesamtprofil, das im kritischen Bereich nirgends in fossilführender Abfolge zusammenhängend aufgeschlossen ist und dessen lithostratigraphische Korrelationen z. T. noch umstritten sind, kann natürlich erst recht kein biostratigraphisches Referenzprofil sein, nicht einmal dann, wenn die Schichten fossilreich wären und die obere »biostratigraphische« Einheit (also das basale »Saxon« nach HAUBOLD—KATZUNG, 1972 mit *T. schmidti* und *P. tambachensis)* auch außerhalb des Steinbruchs Bromacker nach biostratigraphischen Kriterien zu erkennen wäre. Aus diesem Dilemma gibt es auch keinen Ausweg, indem man — wie bei HAUBOLD—KATZUNG (1978) — das »Autun« und »Saxon« als biostratigraphische Serien des Perm

deklariert und auf diese Weise das »Autun« und »Saxon« von ihren Typusregionen löst. Dies geschah wohl in Anbetracht der Tatsache, daß bei der Festlegung der neuen »Autun«/»Saxon«-Grenze durch HAUBOLD—KATZUNG (1972) der Umfang dieser Einheiten in ihren Typusregionen nicht berücksichtigt wurde, weil den Autoren nicht einmal die Arbeit von DE LAPPARENT (1893) bekannt war, wo das Saxonian aufgestellt und die Autunian/Saxonian-Grenze definiert wurde. Das geht auch aus der Umstufung der Hornburger Formation von Saxonian in das Autunian durch HAUBOLD (1973) hervor, da die Hornburger Formation den größten Teil des Saxonian in dessen Typusregion repräsentiert (vgl. KOZUR, 1978d). In diesem Sinne sind auch die folgenden Angaben bei HAUBOLD—KATZUNG (1978, S. 441) zu werten: »Als Serien sind Autun und Saxon nicht stratotypisch belastet — das gilt im übrigen auch für das Perm (!) — sondern biostratigraphisch definiert mit den bekannten Belegen in den Referenzprofilen.« Was von dem »Referenzprofil« der »Autun«/»Saxon«-Grenze zu halten ist, wurde oben dargelegt. Wo sich weitere solche Referenzprofile, die nicht eine biostratigraphische Serie, sondern deren gegenseitige Abgrenzung markieren können, liegen sollen, ist unklar. Da die »Leitformen« für das untere Saxon, *T. schmidti* und *P. tambachensis*, nur in der Lokalität Bromacker als größte Seltenheiten vorkommen, könnte auch nur dort ein Referenzprofil liegen, obwohl es selbst dort nicht existiert (siehe oben). Wie wenig die angeblich biostratigraphischen Serien »Autun« und »Saxon« mit wirklicher Biostratigraphie zu tun haben, soll an folgendem Zitat bei HAUBOLD—KATZUNG (1978, S. 441) erläutert werden: »Dieses Unterperm zerfällt zweckmäßig auch in Unter- und Mittelperm, die entsprechenden Serien sind Unteres und Oberes Rotliegendes. Sie stimmen mit den biostratigraphischen »Abteilungen« Autun und Saxon überein, definiert von HAUBOLD & KATZUNG.« Nachdem nun auch HAUBOLD—KATZUNG (1978) von ihrer bisherigen falschen Meinung abweichen, daß das Rotliegende nur bis zur Obergrenze des Unterperm reicht (die Unhaltbarkeit dieser Auffassungen wurde bei KOZUR, 1977, 1978d eindeutig nachgewiesen), behandeln sie die wirklichen biostratigraphischen Serienbegriffe Unter- und Mittelperm in einer Weise, die mit Biostratigraphie nichts zu tun hat. Daß Unter- und Oberrotliegendes eindeutige lithostratigraphische Begriffe sind und daher keine biostratigraphischen Serienbezeichnungen sein können, wurde schon ausführlich dargelegt, ebenso, daß sich dazu »Autun« und »Saxon« im Sinne von HAUBOLD—KATZUNG (1972) wie auch in jeder anderen Fassung nicht im geringsten eignen. Die Gleichsetzung dieser Begriffe mit dem Unter- und Mittelperm ist eine Leichtfertigkeit, die die Korrelation des Rotliegenden genauso belastet, wie die bisherige leichtfertige Gleichsetzung des Rotliegenden mit dem Unterperm bzw. die Gleichsetzung der Obergrenze des Unterrotliegenden mit der Obergrenze des Unterperm durch die gleichen Autoren. Erstens zerfällt, wie schon lange bekannt ist und bei KOZUR (1977) noch einmal ausdrücklich hervorgehoben wurde, das

Oberperm s. l. in Mittelperm und Oberperm s. str. und nicht das Unterperm in Unterperm s. str. und Mittelperm, wie HAUBOLD—KATZUNG angeben. Zweitens umfaßt das »Autun« im Sinne von HAUBOLD—KATZUNG (1972) in seinem Minimalumfang (Beginn mit dem Einsetzen von *Callipteris*, Obergrenze von Region zu Region verschieden, laut Definition Einsetzen von *Tambachichnium schmidti* und *Palmichnus tambachensis)* den Bereich vom Mittleren Gzhelian bis zum Artinskian (höheres Oberkarbon bis Beginn des mittleren Unterperm, vgl. KOZUR, 1977 und in Druck a, c). Die unterste permische Serie würde also, ganz gleich, ob in der endgültigen Definition der Karbon/Perm-Grenze das Asselian zum Karbon (faunistische begründet) oder zum Perm (faunistisch kaum zu begründen) gestellt wird, zu einem beträchtlichen Teil zum Karbon gehören. Drittens gehört die Tambacher Formation des Thüringer Waldes zum Unterperm, so daß die Gleichsetzung von »Saxon« im Sinne von HAUBOLD—KATZUNG (1972) und Mittelperm falsch ist. Zum (tieferen und mittleren) Mittelperm gehört nur das oberste Oberrotliegende (z. B. Cornberger Sandstein, Grenzkonglomerat, Eislebener Formation) sowie von HAUBOLD—KATZUNG (1975) zum »Autun« gezählte Schichten außerhalb Mitteleuropas (nur die Val Gardena-Formation der Südalpen, die dem gesamten Mittelperm entspricht).

Bei den biostratigraphischen »Stufen« (bei HAUBOLD—KATZUNG, 1978 wurde die biostratigraphische Stufe A_t aufgeführt) handelt es sich um eine Einteilung nach HAUBOLD (1977). Diese »Stufen« im Sinne HAUBOLDs sind aber keine biostratigraphischen, sondern lithostratigraphische Einheiten. Ihre Grenzen werden im allgemeinen mit den Grenzen lithostratigraphischer Einheiten definiert, und zwar mit den Formationsgrenzen im Sinne von HAUBOLD—KATZUNG (1978). Ihre »biostratigraphische« Charakterisierung besteht in der Aufzählung des Fossilinhalts bestimmter lithostratigraphischer Einheiten. Es handelt sich also um biofazielle Einheiten, deren Grenzen an lithostratigraphischen Grenzen liegen. Die Einführung von biostratigraphischen Stufen im Rotliegenden Mitteleuropas muß verwundern, da im Perm bereits biostratigraphische Stufen existieren (Sakmarian, Artinskian etc.), so daß es bei der Korrelation des Rotliegenden nicht darauf ankommt, weitere Stufen neu zu benennen, sondern das Rotliegende muß so datiert werden, daß es mit der bestehenden Stufengliederung korreliert werden kann. Die Stufe A_t nach HAUBOLD (1977) umfaßt laut Definition die obere Gehrener und untere Manebacher Formation, die nach KOZUR (1978c, in Druck a, c) in das höhere Gzhelian eingestuft werden können. Selbst bei einer Zuordnung des Asselian zum Perm, die der Autor nicht für richtig hält (vgl. KOZUR, 1977, 1978b, c, in Druck b, c) würde die »permische Stufe« A_t auf jeden Fall zum Oberkarbon gehören.

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MAGYAR TUDOMÁNYOS AKADÉMIA KÖNYVTÁRA

К ВОПРОСУ ПРИМЕНЕНИЯ ЛИТО- И БИОСТРАТИГРАФИЧЕСКИХ СХЕМ РАСЧЛЕНЕНИЯ КРАСНОГО ЛЕЖНЯ (РОТЛИГЕНД)

Г. КОЗУР

Резюме

Рассматривается применение литостратиграфических и биостратиграфиских методов в отложениях красного лежня (ротлигенд) на территории Средней Европы. Автор предлагает следующую последовательность красного лежня на территории Тюрингии:

надгруппа	группа	формация	подразделение	пласт
	Верхний красный лежень	Тамбах		
Красный лежень	Нижний Красный лежень	Роттероде Обергоф Гольдлаутер Манебах Герен	угленосное подр.	угольный пл. I

До сих пор некоторые литостратиграфические единицы некоторыми авторами рассматривались как биостратиграфические. В связи с этим возникли серьезные ошибки в корреляции толщ красного лежня с различными толщами в Южных Альпах и США. Биостратиграфические возрастные определения не влияют на границы, проведенные по литостратиграфическим признакам и не изменяют характер этих единиц, не превращая их в биостратиграфические. Некоторые авторы рассматривали надгруппу «красный лежень» как биостратиграфическую единицу, так как в этой литостратиграфической единице делались некоторые биостратиграфические возрастные определения. Эти авторы в некоторых случа ях также изменили литостритиграфическую границу между группами нижнего и верхнего красного лежня, исходя из совсем ненадежных и неточных биостратиграфических данных (отпечатки ног тетраподов). Однако, при всех этих биостратиграфических возрастных определениях красный лежень все же остается определенно литостратиграфической единицей. Термины отэнский и саксонский ярусы не рекомендуется применять за пределами Франции. Большая часть отэнского яруса принадлежит к каменоугольной системе, независимо от многочисленных определени отэнского яруса.

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^{57}Fe MÖSSBAUER ANALYSIS OF CHRYSOTILE ASBESTOS FROM VARIOUS MINING REGIONS

By

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^{57}Fe Mössbauer spectroscopy has been used for studying the oxidation- and coordination state of iron in chrysotile asbestos from various mining regions in Canada, Rhodesia, USSR and Yugoslavia. It has been found that both the Fe^{2+} and Fe^{3+} ions occupy only the octahedral positions in the chrysotile crystal structure and that the Fe^{2+}/Fe^{3+} ratio depends strongly on the mining region.

Mössbauer spectra have shown that the samples contain also magnetite.

Introduction

Asbestos is the general term used to describe a number of minerals which can be separated into long silky fibers. These minerals are found mainly in the serpentine (sheet silicate) and the amphibole (chain silicate) groups. The principal varieties of asbestos which have industrial usefulness are shown in Fig. 1. The chemical composition of asbestiform minerals varies somewhat according to the deposit localities. Asbestiform minerals are usually characterized by a close association with other minerals. Also, they can incorporate small concentrations of other elements in the mineral crystal lattice.

Serpentine minerals are divided into three main groups — chrysotiles, lizardites and antigorites. All have the chemical composition of $Mg_3(Si_2O_5)(OH)_4$, but varying adjustments of the misfit between the tetrahedral and octahedral sheets. Chrysotile has a tubular structure, and the tube-like fibers are the result of a curvature of the 001 crystal plane parallel to the *a* and *b* crystallographic axes. This asbestiform mineral plays an important role in industry. Minerals which are frequently associated with chrysotile asbestos are magnetite, chromite, brucite, calcite, dolomite and awaruite. Nickel and iron ions can replace magnesium ions by isomorphic substitution.

The application of Mössbauer spectroscopy to iron containing silicate minerals has been well established [1, 2, 3]. In the first place, the Mössbauer parameters, such as isomer or center shift and quadrupole splitting, can be correlated with the oxidation state and the coordination environment of iron in silicates. In some cases, it is possible to distinguish Fe^{2+} in structurally

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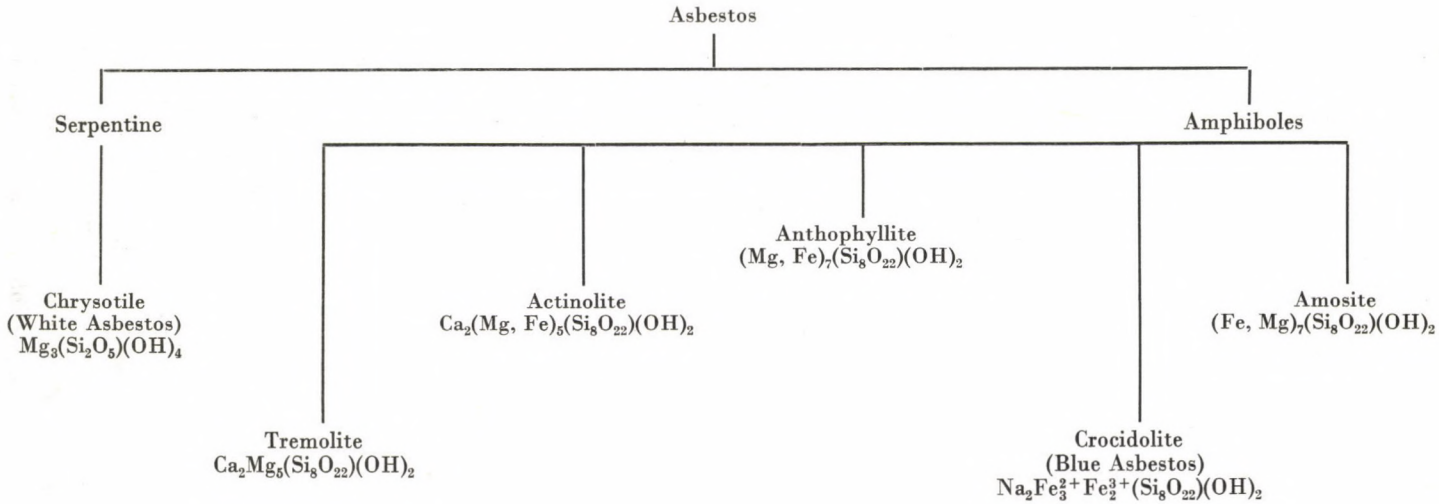

Fig. 1. Principal varieties of asbestos of industrial importance

different positions as having the same coordination environment [4]. The ratio of the intensities of the appropriate Mössbauer lines can be used to determine the Fe^{3+}/Fe^{2+} ratio. Mössbauer spectroscopy has an application in geological studies concerning the conditions of crystallization and the subsequent thermal history of natural crystals and rock-forming minerals [5, 6].

In this paper, the results of the Mössbauer analysis of chrysotile asbestos from several mining regions will be discussed.

Experimental

Specimens of chrysotile asbestos were taken from several mining localities in Canada, Rhodesia, USSR and Yugoslavia. Prior to the Mössbauer experiments, the samples were studied using X-ray diffraction, IR spectroscopy and DTA. It was established that all the chrysotile asbestos contained a small quantity of lizardite as an associated serpentine mineral. Quantitative chemical analysis was performed by standard analytical methods. Mössbauer spectra were recorded in absorption geometry at room temperature. A ^{57}Co(Pd) source was used.

The evaluation of the spectra was performed by the least square fitting computer program. The values of isomer shift (δ) are given relative to iron.

Results and Discussion

Generally, asbestiform minerals contain iron impurities. Iron can be present in the form of iron oxides or incorporated in the mineral crystal lattice. In this work, comparative ^{57}Fe Mössbauer analysis of chrysotile asbestos were performed in order to determine the following: a) the independent iron containing mineral components, and b) the oxidation states and coordination numbers of the iron ions incorporated in the chrysotile crystal structure. The chemical composition (%) of the samples is shown in Table 1. The iron contents vary from 5.21% to 8.24%, as Fe_2O_3 expressed. A characteristic ^{57}Fe Mössbauer spectrum of chrysotile asbestos is shown in Fig. 2 (sample 8A). Mössbauer parameters obtained by the computerized evaluation of the spectra are shown in Table 2.

The samples contain magnetite, since on the spectra two sextets characteristic for Fe_3O_4 appear. On the basis of the central part of the spectra, information can be obtained on iron ions incorporated into the chrysotile lattice. The low velocity range of the spectra can be resolved very well if the sample does not contain magnetite (the inner lines of magnetite disturb the evaluation of the paramagnetic components). Sample 5A (Stragari, Yugoslavia) is of this

Table 1

Chemical composition (%) of chrysotile asbestos

Sample	Moisture	Ignition loss at 1000 °C	SiO_2	Fe_2O_3	MgO	CaO	Al_2O_3	Co	Ni	K
4D-Lake 7A-Canada	0.73	14.31	39.11	5.41	39.90	0.62	1.80	0.26	0.16	0.36
BC-110-4D 8A-Canada	0.84	14.50	38.89	6.61	44.84	0.20	1.50	0.29	0.15	0.35
98 9A-Rhodesia	1.13	15.57	36.71	6.01	40.29	0.22	1.59	0.35	0.19	0.46
4450 10A-Rhodesia	0.79	16.01	36.31	8.24	39.42	1.20	—	0.30	0.22	3.14
P-3-60 11A-USSR	0.95	14.70	39.82	6.65	40.43	0.24	—	0.37	0.23	0.37
P-3-70 12A-USSR	0.83	15.53	38.32	7.14	37.96	1.33	2.06	0.31	0.22	0.36
Stragari 5A-Yugoslavia	2.14	16.33	39.24	5.74	39.23	0.29	1.47	0.39	0.15	0.57
Petrovo S. 6A-Yugoslavia	0.90	14.89	39.48	5.21	39.04	0.20	1.62	0.38	0.19	0.50

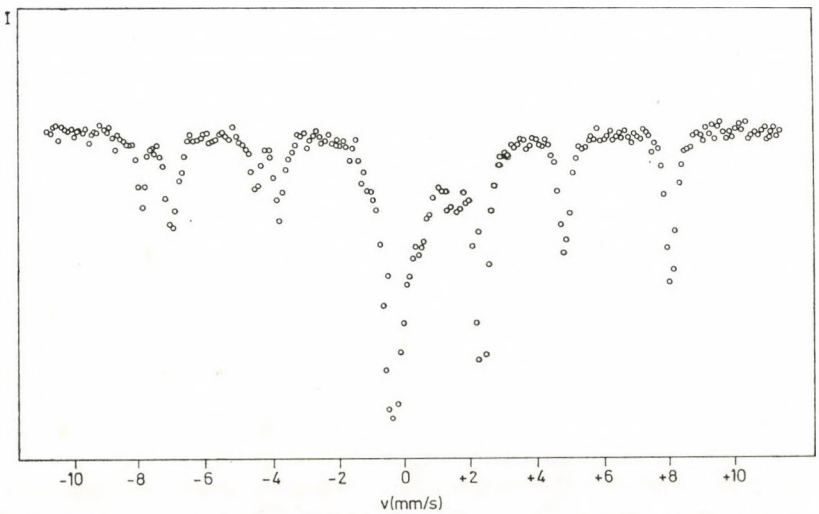

Fig. 2. Room temperature ^{57}Fe Mössbauer spectrum of chrysotile (sample 8A)

Table 2

^{57}Fe Mössbauer parameters of chrysotile asbestos at room temperature

Sample	Lines	Isomer shift δ* mm s^{-1}	Quadrupole splitting ΔE* mm s^{-1}	Line width Γ mm s^{-1}	Inner magnetic field $H^*_{5/2}$ A m^{-1} (kOe)	Area of the lines, A** %	Fe(II)/ Fe(III)
4D-Lake	M_1	0.310	0.004	0.312	3.93×10^7 (493.4)	57	
7A-Canada	M_2	0.719	0.000	0.396	3.68×10^7 (462.2)		
	Q_1	1.169	2.749	0.297	—	26	1.53
	Q_2	0.413	0.592	0.530	—	17	
BC-110-4D	M_1	0.304	0.004	0.296	3.94×10^7 (495.5)	50	
8A-Canada	M_2	0.731	0.001	0.353	3.70×10^7 (464.7)		
	Q_1	1.137	2.885	0.358	—	32	1.78
	Q_2	0.339	0.698	0.501	—	18	
98	M_1	0.286	0.009	0.457	3.92×10^7 (492.5)	42	
9A-Rhodesia	M_2	0.742	0.027	0.478	3.72×10^7 (466.9)		
	Q_1	1.184	2.696	0.409	—	14	0.32
	Q_2	0.458	0.562	0.556	—	44	
4450	M_1	0.305	0.005	0.131	3.93×10^7 (492.4)	15	
10A-Rhodesia	M_2	0.721	0.010	0.365	3.69×10^7 (464.0)		
	Q_2	0.348	0.556	0.579	—	85	0.00
P-3-60	M_1	0.349	0.004	0.289	3.94×10^7 (495.5)	36	
11A-USSR	M_2	0.727	0.009	0.456	3.70×10^7 (464.7)		
	Q_1	1.191	2.683	0.361	—	35	1.19
	Q_2	0.480	0.476	0.610	—	29	
P-3-70	M_1	0.290	0.016	0.379	3.94×10^7 (495.5)	42	
12A-USSR	M_2	0.741	0.006	0.458	3.70×10^7 (464.7)		
	Q_1	1.185	2.714	0.365	—	29	1.00
	Q_2	0.381	0.540	0.560	—	29	
Stragari	Q_1	1.165	2.709	0.310	—	72	2.57
5A-Yugoslavia	Q_2	0.330	0.494	0.580	—	28	
Petrovo S.	M_1	0.270	0.014	0.460	3.94×10^7 (495.5)	34	
6A-Yugoslavia	M_2	0.706	0.067	0.353	3.69×10^7 (464.0)		
	Q_1	1.186	2.684	0.388	—	49	2.88
	Q_2	0.450	0.560	0.591	—	17	

* The uncertainty is in $\delta = \pm0.005$ mm s^{-1}, $\Delta E = \pm0.005$ mm s^{-1} and $H_{5/2} = \pm$ $\pm6.4\times10^4$ A m^{-1} (0.8 kOe).

** A is the area of a line expressed as a percentage of the whole area calculated on the basis of spectrum at room temperature.

kind. The inner part of the spectra can be resolved for two quadrupole doublets. The doublet with the values $\delta = 1.13—1.19$ mm s^{-1} and $\Delta E = 2.6—2.9$ mm s^{-1} represents iron(II) in octahedral positions. The values of δ and ΔE for iron(II) in tetrahedral positions are less [2]. Since Fe^{2+} and Mg^{2+} ions are isomorphous, it is obvious that Mg^{2+} is substituted by Fe^{2+} in the brucite layer of the chrysotile. The second doublet with the values of $\delta = 0.33—0.48$ mm s^{-1} and $\Delta E = 0.45—0.70$ mm s^{-1} corresponds to iron(III). The line width Γ varies between 0.50 mm s^{-1} and 0.60 mm s^{-1}. This line width is rather broad and the possibility of the presence of two or more doublets (superposition) with similar parameters cannot be excluded. In clay minerals (also layer silicates), Fe^{3+}-ions in tetrahedral position have Mössbauer parameters of $\delta/\Delta E = 0.22/0.33$ [7]. The substitution of Si^{4+} in the tetrahedral position by Fe^{3+} can also be supposed in the case of chrysotile asbestos. If Mg^{2+} is substituted by Fe^{3+} ions, an excess of charge (1+) appears and the Si^{4+} should be substituted with Me^{3+}-ions (generally it can be the Fe^{3+}) owing to the compensation of the charges. In the case of the substitution of Si^{4+} by Fe^{3+}, the Mössbauer spectrum should be characterized by Fe(III) being in tetrahedral position.

MALYSHEVA et al. [8] have also supposed the Fe^{3+} to be in tetrahedral position in the course of a Mössbauer investigation of some Ural serpentines. They obtained $\delta/\Delta E = 0.31/0.31$ mm s^{-1} values. During the preparation of this publication, two papers on Mössbauer spectroscopy of serpentine minerals have appeared. BLAAUW et al. [9] published the results of the Mössbauer analysis of some Canadian chrysotiles from several mining areas. They concluded that Fe^{2+} occupies only the octahedral (Mg) position in chrysotile, whereas Fe^{3+} is distributed between the octahedral and the tetrahedral (Si) positions. For the isomer shift (relative to metallic iron) and the quadrupole splitting of Fe^{2+}, they measured $\delta/\Delta E = 1.12/2.65$ mm s^{-1}, for the Fe^{3+} the corresponding values are: $\delta/\Delta E = 0.34/0.45$ mm s^{-1} in the octahedral position and $\delta/\Delta E = 0.20/0.34$ mm s^{-1} in the tetrahedral position. ROZENSON et al. [10] have shown that in antigorites only one type of octahedral site is occupied by Fe^{2+} or Fe^{3+} ions. The Mössbauer spectra of antigorites, lizardites and chrysotiles all showed one Fe^{2+} doublet at room temperature with values of $\delta = 1.12—1.15$ mm s^{-1} and $\Delta E = 2.70—2.79$ mm s^{-1}. In the chrysotiles examined by us, it seems that Fe^{3+} ions occupy two octahedral positions. Interpretation of the Fe^{3+} lines was difficult, owing to the large linewidth of the doublets and the possible presence of Fe^{3+} with tetrahedral coordination.

In this work, the resolution of the central part of the spectra for three doublets Fe^{2+}_{oct}, Fe^{3+}_{oct} and Fe^{3+}_{tet} has also been attempted. However, χ^2 values were not lower in comparison with the fitting for two doublets and the Mössbauer parameters obtained for Fe^{III} in tetrahedral position were nonrealistic, if they are compared with known values in the literature for Fe^{3+}_{tet}. On the

basis of these results, it cannot be concluded with certainty that Fe^{III} substitutes for Si^{IV} in the chrysotile samples. The broadening of the iron(III) lines can be explained by the variation in the iron(III) chemical environment, and this reflects a higher defect concentration around the Fe^{III} ions.

On the basis of the line areas, distribution of iron between the Fe(II) and the Fe(III) state has been achieved. It can be shown from Table 2 that the Fe(II)/Fe(III) ratio in chrysotile asbestos depends strongly on the mining area. The Fe(II)/Fe(III) ratio in minerals reveals important information on the partial pressure of oxygen during crystallization, which is a parameter of geological significance [5]. The comparatively high iron(II)/iron(III) ratios, e.g. in the case of the Yugoslav chrysotile asbestos, refer to the formation of these minerals under high pressure. The low iron(II) content of the sample from Rhodesia (the 10A sample does not contain any Fe(III)) permits the conclusion that these minerals are formed under relatively low pressures and highly oxidative conditions.

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ОПРЕДЕЛЕНИЕ ^{47}Fe ХРИЗОТИЛОВОГО АСБЕСТА ИЗ РАЗЛИЧНЫХ ГОРНОПРОМЫЩЛЕННЫХ РАЙОНОВ АНАЛИЗОМ МЕССБАУЕРА

И. НАДЬ-ЦАКО—С. МУСИЧ—А. ВЕРТЕШ—О. ХАДЖИЯ—З. ДРАГЧЕВИЧ—ЛАХОДНИ-ШАРЦ

Резюме

Спектроскопический метод ^{57}Fe Мёссбауера применялся для изучения состояния оксиления и координационного числа железа в хризотиловом асбеста из различных районов Канады, Родезии, СССР и Югославии. При этом установили, что как ионы Fe^{2+}, так и ионы Fe^{+} занимают только лишь октаэдрические позиции в кристаллической структуре хризотила, а также, что соотношение Fe^{2+}/Fe^{3+} сильно зависит от горнопромышленного района. Спектры Мёссбауера показали, что образцы также содержат магнетит.

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MECHANISM OF DISTRIBUTION OF Fe^{2+}/Fe^{3+} IN NATURE MAGNETITE MINERAL

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The natural magnetite minerals, Fe_3O_4 of various places have been analysed and the distribution of Fe^{2+} and Fe^{3+} interpreted in terms of Crystal Field Stabilization Energy[1] (CFSE) and isomorphous replacement.[2] The CFSE factor appears to hold good.

Introduction

In magmatic segregation deposits of iron oxides or sulphides, the typical minerals are magnetite (Fe_3O_4) instead of wilstite (FeO), and Pyrrhotite ($Fe_{1-x}S$) instead of troilite (FeS). Magnetite, Fe_3O_4 is ferrous ferric mineral.

Experimental Results

For the purpose of analysis, volumetric, gravimetric and colorimetric methods have been adopted. The analytical results are presented in the Table 1.

Table 1

Cations	A	B	C	D	Average, %
Fe^{2+}	24.40	24.37	24.43	24.41	24.40
Fe^{3+}	41.12	41.09	41.15	41.14	41.12
Ti^{4+}	1.59	1.57	1.61	1.60	1.59
Mn^{2+}	0.77	0.74	0.80	0.77	0.77
V^{3+}	0.72	0.70	0.74	0.73	0.72
Cr^{3+}	0.27	0.25	0.29	0.27	0.27
Ni^{2+}	0.078	0.076	0.080	0.077	0.078

(A, B, C and D represent different samples of the same mineral.)

Discussion

Crystal field stabilization energy (CFSE) have been calculated here to characterize magnetite spinel as:

Magnetite — Fe_3O_4

Fe^{2+}	Fe^{3+}	Fe^{3+}
(Tet)	(Oct)	(Oct)

Normal spinel

Fe^{2+}—d^6 system, Fe^{2+} (d^6) tetrahedral

CFSE $= 6\,Dq$
$= 6 \times 4/9\,Dq$
$= 8/3\,Dq$
$= 2.66\,Dq$

Fe^{3+}—d^6 system, Fe^{3+} (d^5) octahedral

CFSE $= 0\,Dq$
Total CFSE $= 2.66\,Dq + 0 + 0$
$= 2.66\,Dq$

Fe^{3+}	Fe^{2+}	Fe^{3+}
(tet)	(oct)	(oct)

Inverse spinel

Fe^{3+}—d^5 system, Fe^{3+} (d^5) octahedral

CFSE $= 0\,Dq$

Fe^{2+}—d^6 system, Fe^{2+} (d^6) octahedral

CFSE $= 4\,Dq$

Fe^{3+}—d^5 system, Fe^{3+} (d^5) tetrahedral

CFSE $= 0\,Dq$
Total CFSE $= 0 + 4\,Dq + 0$
$= 4\,Dq$

Since the CFSE of Fe^{2+} is higher in case of octahedral and therefore, it prefers octahedral site and the structure of magnetite is an inverse spinel. As the CFSE for Fe^{3+} is zero and that of the Fe^{2+} in octahedral site is — 4 Dq, it is quite expected that the distribution of Fe^{2+} in the mineral will be more than Fe^{3+}. The chemical analyses tabulated in the Table 1 shows that the average percentage of Fe^{2+} is 24.4% and Fe^{3+} is 41.12%.

Thus it appears that the distribution does not obey the CFSE approach. However, if it is considered that the percentage of Fe^{3+} is due to distribution in two different sites, the atom to atom ratio $Fe^{2+}/Fe^{3+} = 24.4 : 20.5$ i.e., the percentage of Fe^{2+} is greater than the percentage of Fe^{3+}. The CFSE

for Ni (d^8) is — 12 Dq in octahedral, and — 6 Dq in tetrahedral. The size of Ni^{2+} (ionic radius 0.68 Å) is closer to Fe^{2+} (0.76 Å) and there is more CFSE for Ni^{2+} in octahedral, it substitutes for Fe^{2+} in the amount of Ni^{2+} when made equivalent, Fe^{2+} turns out to be 0.07%. Due to size closer to Fe^{2+} (0.76 Å) Mn^{2+} (0.80 Å) substitutes Fe^{2+} and the amount of Mn^{2+} when made equivalent to Fe^{2+} turns out to be 0.8. Thus the total Fe^{2+} is about (24.4 + + 0.07 + 0.8) = 25.27%. V^{3+} is a d^2-system and since it has an octahedral radius (0.61 Å) almost identical with that of Fe^{3+}, it tends to follow iron in mineral formation. The CFSE for V^{3+} is — 6 Dq in octahedral and 5.3 Dq in tetrahedral and of it substitutes Fe^{3+} in octahedral site.

Similarly Cr^{3+} (d^3) has CFSE = —12 Dq in octahedral and 5.3 Dq in tetrahedral and so substitutes Fe^{3+} in octahedral site. The Fe^{3+} equivalent of Cr^{3+} and V^{3+} turns out to be (0.45 + 0.84) = 1.29%. Thus the total Fe^{3+} turns out to be 21.8%. The substitution for Fe^{3+} (by Cr^{3+}, V^{3+}) is more than the substitution for Fe^{2+} (by Mn^{2+} and Ni^{2+}). Even if Ti^{4+} is taken to substitute Fe^{3+}, the Fe^{3+} turns out to be (21.8 + 1.9) = 23.7%; and the amount of Fe^{2+} is thus larger due to CFSE of Fe^{2+} relative to Fe^{3+}. The percentage of Ti^{4+} is larger and it indicates that Ti^{4+} may be present as ilmenite. Fe^{2+} $Ti^{4+}O_3$ where Fe^{2+} as well as Ti^{4+} have been substituted by Fe^{3+} in equal amounts. If this is taken then the atom to atom percentage of Fe^{2+} will be more in magnetite relative to Fe^{3+}. The above discussions clearly demonstrate the applicability of CFSE in nature.

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О МЕХАНИЗМЕ РАСПРЕДЕЛЕНИЯ Fe^{2+}/Fe^{3+} В ЕСТЕСТВЕННЫХ МИНЕРАЛАХ МАГНЕТИТА

Р. Н. П. ШИНА—И. РОЙ

Резюме

Анализировались естественные минералы магнетита — Fe_3O_4 — и распределение Fe^{2+} и Fe^{3+} интерпретировался как энергия стабилизации поля кристаллизации и изоморфное замещение. Коэффициент энергии стабилизации поля кристаллизации является, по-видимому, оправданным.

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DOLOMITIZATION, DE-DOLOMITIZATION, RE-CALCITIZATION IN HYDROTHERMAL CONDITIONS (50°—300° C)

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An extended literature is devoted to the metasomatic processes of calcareous rocks. Laboratory model experiments in recent decades have shed light on a number of episodes of transformation (that of a calcite lattice into a dolomite one-, $MgCO_3$-, $MnCO_3$-, $FeCO_3$—$ZnCO_3$-, $CdCO_3$-, $CoCO_3$-, etc.). Some contradictory facts, requiring further investigation, can be met with in connection with the birth of a dolomite lattice ("dolomitization"). At this moment an a priori "primary syngenetic" formation of dolomite under "normal" sedimentological conditions still remains problematic. The dolomites of "shabka" areas also represent a post-calcite epigenic formation, and not even the dolomite detectable in modern chalk deposits in the Great Hungarian Plain is contemporaneous with the calcite. Laboratory model experiments by the author have approached the question formulated in the title via studies of the "solid phase: solution system" = $CaCO_{3(s)} : MgCl_{2(aq)}$. The chemical and crystal chemical characteristics of phase changes are evaluated, of the physico-chemical parameters of the process it is primarily the influence and role of temperature that are analyzed. The model experiments were performed in a confined (closed) system — an autoclave. The chemical changes, crystal chemical and X-ray diffraction data of the resulting crystal phases are summarized in tables. The relative changes of the δ_O and δ_C isotopes of phases generated at different temperatures are worthy of attention.

The following binary systems have been examined:

a) $CaCO_3$(calcite): $MgCl_{2(aq)}$;

b) $CaMg(CO_3)_2$ — pulverulent dolomite: $MgCl_{2(aq)}$

$T° = 50, 75, 100, 125, 150, 200$
$225, 250, 300, 350$ °C.

The magnesium concentration of the solutions has been chosen so that the ratios as to the calcium in the solid phase should be: Ca : Mg = 70 : 30; 50 : 50; 30 : 70.

The formation of dolomite (protodolomite), magnesite-barringtonite ("dolomitization", "magnezitization") in the binary systems just outlined is described.

Almost the entire Ca^{2+} position in the calcite lattice may be exchanged. The phases were nonstoichiometric, for the most part deficient in anions. In the $CaCO_{3(s)}$: : $MgCl_{2(aq)}$ system the optimum temperature of dolomite genesis was 175 °C, above this figure mainly $MgCO_3$ (magnesite lattice) was formed. Most intensive exchange of Ca^{2+} positions in (pulverulent) dolomite is enhanced by *low* concentrations of Mg in solution, while Mg-rich solutions exert an opposite influence, so that a maximum of 30% of the dolomite will, at the most, be converted into magnesite. The transformation of dolomite into magnesite is the process reflecting the essence of the "dedolomitization" mechanism.

c) $MgCO_{3(s)}$-magnesite : $CaCl_{2(aq)}$;

d) $CaMg(CO_3)_2$-pulverulent dolomite : $CaCl_{2(aq)}$.

Temperature and solution concentrations were similar to the case of binary systems. The $MgCO_{3(s)} : CaCl_{2(aq)}$ system consisted of calcite, aragonite, vaterite, dolomite and triclinic barringtonite. The quantitative participation of these and the stacking of crystal phases are dependent on temperature ("recalcitization"?).

While examining the $CaMg(CO_3)_2$ system it was sought to follow with attention the "dedolomitization-recalcitization" process and its lattice structure mechanism. A restricted rate of formation of calcite, aragonite, vaterite (?) and $MgCO_3 \cdot 2\ H_2O$ and $MgCO_3$ dependent on temperature and the Ca^{2+} concentration of the solution was observed, but these mineral phases were formed by mechanisms other than the restack-

ing of the dolomite lattice (e.g. by replacement of magnesium by calcium), as they precipitated from the solution and settled on the surface of intact dolomite rhombohedra.

Restacking of dolomite under natural circumstances — at low temperature — into calcite (= "dedolomitization") can hardly take place, at higher temperatures the possibility for this is reduced even owing to crystal chemical causes.

Introduction

A large number of papers have discussed the metasomatic processes of Ca-rich rocks. Experimental works have provided reliable explanations for many different aspects of this phenomenon (i.e. the alteration of the calcite structure to $MgCO_3$, $MnCO_3$, $FeCO_3$, $ZnCO_3$, $CdCO_3$, $CoCO_3$ etc.). Several experimental — though occasionally contradictional — data are available for the formation of dolomite (dolomitization). At present there is no reasonable explanation for the primary syngenetic deposition of dolomite — in normal sedimentary environments. The problem presented in the title of this paper is now approached by the evaluation of solid-phase-solution systems, like $CaCO_{3(s)}$—$MgCl_{2(aq)}$, etc. It evaluated the chemical- and crystal-chemical aspects of the phase alterations. Amongst the physicochemical characteristics of the processes the paper is focussed mainly on the effect and role of temperature variations. The experiments have been carried out in closed system, in autoclave. The results of approximately 230 experiments, the data of chemical alterations, X-ray diffraction analyses are presented in separate plots in order to avoid the voluminous listing of all data, which otherwise might have doubled the length of the paper.

Among these data the variations in the δ_O and δ_C relative isotopic ratios of the resulted phases in different temperature levels are also presented.

The following binary systems have been investigated:

a) $CaCO_3$ (calcite) : $MgCl_{2(aq)}$

b) $CaMg(CO_3)_2$ (powdery dolomite) : $MgCl_{2(aq)}$

T = 50, 75, 100, 125, 150, 175, 200, 225, 250, 300, (350) °C.

The concentration of Mg in the solution is chosen to provide Ca/Mg = = 1 : 1,60; 70 : 30; 50 : 50; and 30 : 70 ratios when compared to the amount of Ca contained in the solid phase.

The systems above simulate the formation of dolomite (Ca-dolomite, proto-dolomite) and magnesite-barringtonite ("dolomitization", magnesitization).

Almost all of the Ca^{2+} positions in the calcite lattice might be replaced in these systems. The resulted phases were non-stoichiometric, mostly anion-depleted. The optimal temperature for the dolomite formation in the $CaCO_{3(s)}$: $MgCl_{2(aq)}$ system was 175 °C. Above this temperature mainly $MgCO_3$ (magnesite structure) has been formed. More intensive replacement of the Ca^{2+} positions of dolomite (powdery) has been achieved by dilute

Mg-containing solutions, while Mg-rich solutions adversely affected this process, since only 30 per cent of the dolomite could have been converted to magnesite.

The alteration of the dolomite structure to magnesite is in fact a typical case of "de-dolomitization", though this term as is shown later, is generally — and uncorrectly — regarded as synonymous with "re-calcitization".

c) $MgCO_{3(s)}$ (magnesite) : $CaCl_{2(aq)}$

d) $CaMg(CO_3)_2$ (powdery dolomite) : $CaCl_{2(aq)}$

In these experiments the temperature ranges were the same as were applied in *a* and *b* tests, the concentrations of the solutions were chosen between 1.0—2.5 M, with 2 : 1, 1 : 1 and 1 : 2 cation ratios.

e) "Calcareous dolomite": 2.5 M $CaCl_{2(aq)}$

The $MgCO_{3(s)}$: $CaCl_{2(aq)}$ system has been composed of calcite, aragonite, vaterite, dolomite and triclinic $MgCO_3 \cdot 2H_2O$ phases. The proportions of the different crystalline phases as well as the ordered or disordered character of their structures is largely temperature dependent ("re-calcitization"?).

The $CaMg(CO_3)_2$: $CaCl_{2(aq)}$ system was investigated in order to stimulate the "de-dolomitization" process in the way described in the literature, and to reveal its characteristics. In these experiments calcite, aragonite, vaterite (?) $MgCO_3$ and $MgCO_3 \cdot 2H_2O$ have been formed in amounts mainly influenced by the temperature and the ratio of cations in the solid-liquid system. Larger scale development of the de-dolomitization process in natural conditions is hardly probable owing to the adverse effect of increasing temperature. The processes of re-calcitization and de-dolomitization are mainly functions of the structural ordering of the initial crystalline phase, and may be also related to the facts, that the magnesite has larger structural energy and smaller unit cell parameters than the calcite, and the enthalpy changes occur at smaller temperatures than in calcite.

The crystalline phases of the investigated systems are the most frequently studied minerals of both the carbonate sediments and the rock alteration associated by the hydrothermal ore deposition ("dolomitization", "magnesitization"). There are many controversial points in the interpretation of their genesis and their formation during chemical sedimentation is one of the most critical problem of the sedimentary petrology. The main difference between the structures of dolomite-magnesite and the more frequent calcite is in the more co-valent character of the Mg–O bonds and the 30—50% ionic-covalent character of the Ca–O bonds. The possibility of the replacement of the Mg^{2+} in the dolomite by Fe^{2+}, Mn^{2+}, Pb^{2+}, Co^{2+}, Ca^{2+} and Zn^{2+} ions is very restricted (Zn-dolomite, Pb-dolomite), the replacement of Ca^{2+} by Mg^{2+}, Fe^{2+}, Mn^{2+} ions might resulted in intense changes of the unit cell volume.

D. L. Graf,—J. R. Goldschmith (1955) synthetized Mg-calcite at room temperature and found it to be unstable. At about 600 °C the calcite structure

may contain 9% $MgCO_3$ (J. R. GOLDSCHMITH, 1955), or 2% as suggested by other authors.

The solubility of Mg^{2+} in (Ca,Mg)O is larger than that of the Ca^{2+} in (Mg,Ca)O, therefore $Ca,Mg(CO_3)_4$ (huntite) is more stable than $Ca_3Mg(CO_3)_4$. The konite, $Ca_3Mg(CO_3)_4$ (RETZIUS, 1975) is in fact an intermediate phase, which occurs as a mixture of dolomite and magnesite in natural conditions, and the leesbergite $CaMg_2(CO_3)_3$ is also a solid solution, not an individual phase.

MEDLIN, W. L. (1959) described the presence of the following phases in the $MgCl_2$—$CaCl_2$+urea+CO_2 system:

120°—150 °C aragonite
120°—200 °C = 5 $MgO(CO_2)_4 \cdot 5\ H_2O$
200° = mainly dolomite
200°—250°C = calcite, dolomite, magnesite and basic Mg-carbonate

In the presence of small amount of NaCl the formation of dolomite has shifted to 210—230 °C. With 6—7% NaCl the temperature ranges of the crystalline phases have been found as follows:

150 °C = aragonite, basic Mg-carbonate
150°—210 °C = dolomite, small amount of basic Mg-carbonate
230°—350 °C = calcite, dolomite, magnesite

The total work has included about 230 experiments in systems having Ca : Mg = 1; 1.60, and 70 : 30, 50 : 50, 30 : 70 ratios, in the 25°—300°—(350°) C temperature range.

I. The Problem of "Dolomitization—Magnesitization"

I/1. $CaCO_{3(s)}$ — Calcite + $MgCl_2$ — H_2O System

(The dolomitization and magnesitization of the calcite)

I/1-a $CaCo_{3(s)} : Mg^{2+} = 1 : 1.60$

According to the crystal-chemical evaluations the introduction of Mg into the calcite structure is a function of the concentration of the solution, and showed an increase from 175°—200 °C, with a simultaneous decrease of the amount of CaO in the crystalline phase. The most relevant change in the Mg : Ca 1 : 1.60 system begins from 150 °C, when the rate of replacement of the calcium becomes almost equal with the rate of fixation of magnesium in the structure, and this process is relected by the formation of a new crystalline phase.

According to the chemical, microscopic and X-ray data, the fixation of magnesium in the calcite structure up to 100° C does not destroy the initial structure. Between 50° and 175° C Ca-dolomite, $MgCO_3 \cdot 2H_2O$ and huntite (?)

appeared in the system as intermediate phases. The destruction of the calcite structure takes place between 175° and 200° C, when $MgCO_3$ and $CaMg(CO_3)_2$ begin to develop, with associated changes in the grain-size distribution of the initial material (calcite).

Stoichiometric evaluations of the analyses show the dominance of calcium. This does not mean, however, the exclusive existence of mono-phase crystalline structures. The trend of the solid stage cation exchange can be traced as follows:

$$
\begin{array}{rcl}
25^\circ\,C = & - \quad - & = CaCO_3,\ CaMg_3(CO_3)_4\ ?, \\
50^\circ\,C = & Ca_{0.987}Mg_{0.011}CO_3 & = CaCO_3,\ CaMg_3(CO_3)_4\ ?, \\
75^\circ\,C = & Ca_{0.987}Mg_{0.021}CO_3 & = CaCO_3,\ CaMg_3(CO_3)_4\ ?, \\
100^\circ\,C = & Ca_{0.962}Mg_{0.036}CO_3 & = CaCO_3,\ CaMg_3(CO_3)_4, \\
 & & \quad CaMg(CO_3)_2,\ MgCO_3, \\
125^\circ\,C = & Ca_{0.947}Mg_{0.036}CO_3 & = CaCO_3,\ CaMg(CO_3)_2, \\
 & & \quad MgCO_3, \\
150^\circ\,C = & Ca_{0.642}Mg_{0.419}CO_3 & = CaCO_3,\ MgCO_3,\ CaMg(CO_3)_2, \\
175^\circ\,C = & Ca_{0.253}Mg_{0.885}CO_3 & = CaCO_3,\ MgCO_3,\ CaMg(CO_3)_2, \\
200^\circ\,C = & Ca_{0.137}Mg_{1.017}CO_3 & = CaCO_3,\ MgCO_3,\ CaMg(CO_3)_2, \\
250^\circ\,C = & Ca_{0.369}Mg_{0.757}CO_3 & = MgCO_3,\ CaMg(CO)_2, \\
300^\circ\,C = & Ca_{0.216}Mg_{0.947}CO_3 & = MgCO_3,\ CaMg(CO)_2.
\end{array}
$$

In the stable isotope compositions of the crystalline phases the variation of the δ_O values is apparently similar to that of the δ_C, which, on the other hand, can be related to the variations of the $(CaCO_3)$—Ca^{2+} and Ca^{2+}/Mg^{2+} compositional variations of the "mother-phase". There is no acceptable explanation for the large standard deviation of the δ_O and δ_C values in the natural dolomite and $MgCO_3$ — this can not be attributed to isotope dilution or fractionation during dissolving. In those phases developed in the 50—300 °C range the δ_O—δ_C values are decreasing with temperature. Their relationship

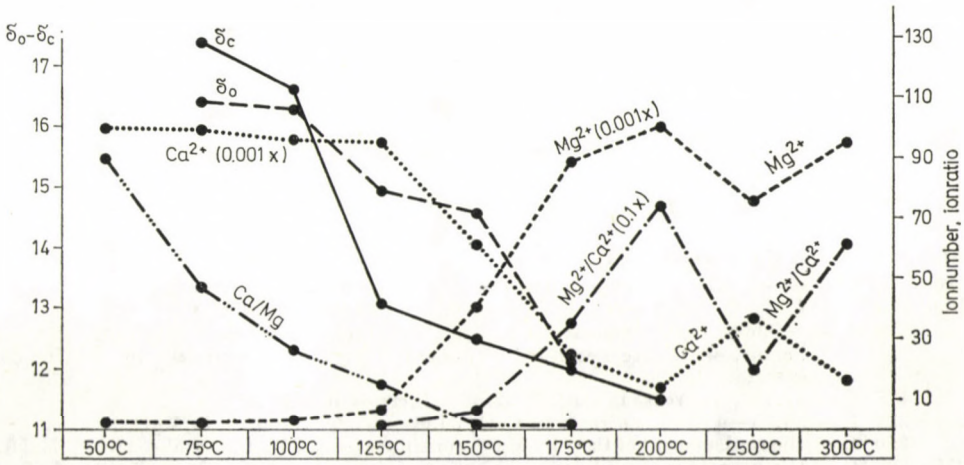


Fig. 1. The plot of δ_O and δ_C variation versus temperature

with the Ca^{2+}, Mg^{2+} and Ca/Mg values are plotted in Fig. 1. The similarity of isotopic variability during the experiments points to the necessity of the investigation of isotopic-geochemic approach, as well as further investigations.

I/1-b $CaCO_{3(s)}$—Mg^{2+} = *70 : 30*

The chemical analyses and evaluations of the resulted phases are summarized in Fig. 2 and Fig. 3.

b-1. The Chemical Analysis Data (Figs 2—3)

50—100 °C = very slight changes in CaO—MgO%.
125 °C = −1.20% CaO+0.13% MgO compared to the initial composition.
150 °C = 3.35% CaO decrease, 2.97% MgO increase,
175 °C = abrupt CaO loss, and similar increase in MgO. −35.39% CaO and +16.99% MgO compared to the initial composition,
200—225 °C = overall 47.25% CaO decrease and +23.13% MgO increase,
250 °C = −55.04% CaO, +25.16% MgO,
275 °C = −49.82% CaO, +23.41% MgO,
300 °C = theoretically to total amount of calcium has been replaced, the solid phase contains only 1.23% CaO. The crystalline phase is virtually pure $MgCO_3$, with 47.87% MgO. ($MgCO_3 \cdot 2\,H_2O$; MgO content in literature: 45.43% MgO.)

b-2. Crystalline Phases

50—150 °C = $CaCO_3$ (calcite) is dominant, $MgCO_3 \cdot 2\,H_2O$ (barringtonite) is in traces, $CaMg_3(CO_3)_4$ (huntite) is indicated up to 125 °C.
175 °C = $CaCO_3$ (calcite) and dolomite in equal amounts, $MgCO_3$ and $MgCO_3$ are accessories,
200 °C = the dolomite shows the highest degree of ordering, its amount is larger than $MgCO_3$, the $MgCO_3 \cdot 2\,H_2O$ is subordinate, calcite has not been observed.
225—250 °C = the $MgCO_3$ is predominant, its amount is 3—5% greater than that of dolomite. $MgCO_3 \cdot 2\,H_2O$ is persistent accessory, the presence of $CaMg_3(CO_3)_4$ — huntite — is questionable.
175—300 °C = mostly $MgCO_3$, with 1.23% CaO impurities, the particles are encrusted by $MgCO_3 \cdot 2\,H_2O$.

The basic Mg-carbonate ($MgCO_3 \cdot 2H_2O$) was present in each samples in small quantities, and according to the scanning-EM images, it has formed epigenetically after the completion of the experiments, thus it can not be considered as intermediate phase between the dolomite and $MgCO_3$ (magnesite).

I/1-c $CaCO_{3(s)}$—Mg^{2+} = *50 : 50*

c-1. The Chemical Analysis Data (Figs 2—3)

50— 75 °C = no significant change in the CaO—MgO percentage distribution,
100—125 °C = 1.91—2.10% decrease in CaO, +0.25—0.61 increase in MgO,
150 °C = −9.60% CaO decrease, +8.42% MgO increase, marked structural change begins, with rate becoming exponential at 175 °C, similarly to the 70 : 30 system (Fig. 3),
175 °C = 37.73% decrease in CaO, +32.31% increase in MgO,
200—225 °C = 39.35—44.50% decrease in CaO, 34.00—36.56% increase in MgO,
250 °C = the replacement of Ca^{2+} by Mg^{2+} continues (−53.34% CaO and +45.47% MgO),
275—300 °C = the process is slowed down. With increasing amount of Ca^{2+} in $MgCO_3$ (7.67—29.80%) dolomite is started to form again (300 °C) in amounts of 8—10%.

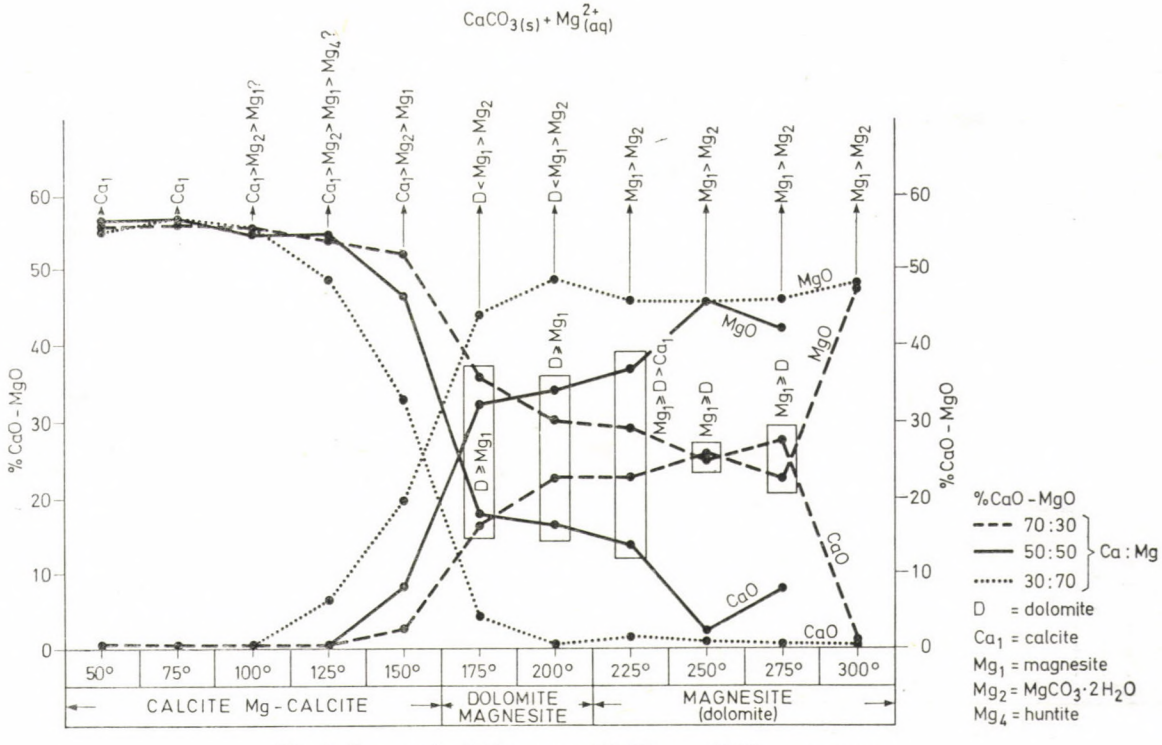


Fig. 2. Summarized diagram of $CaCO_{3(s)}-Mg^{2+}{}_{(aq)}$ systems

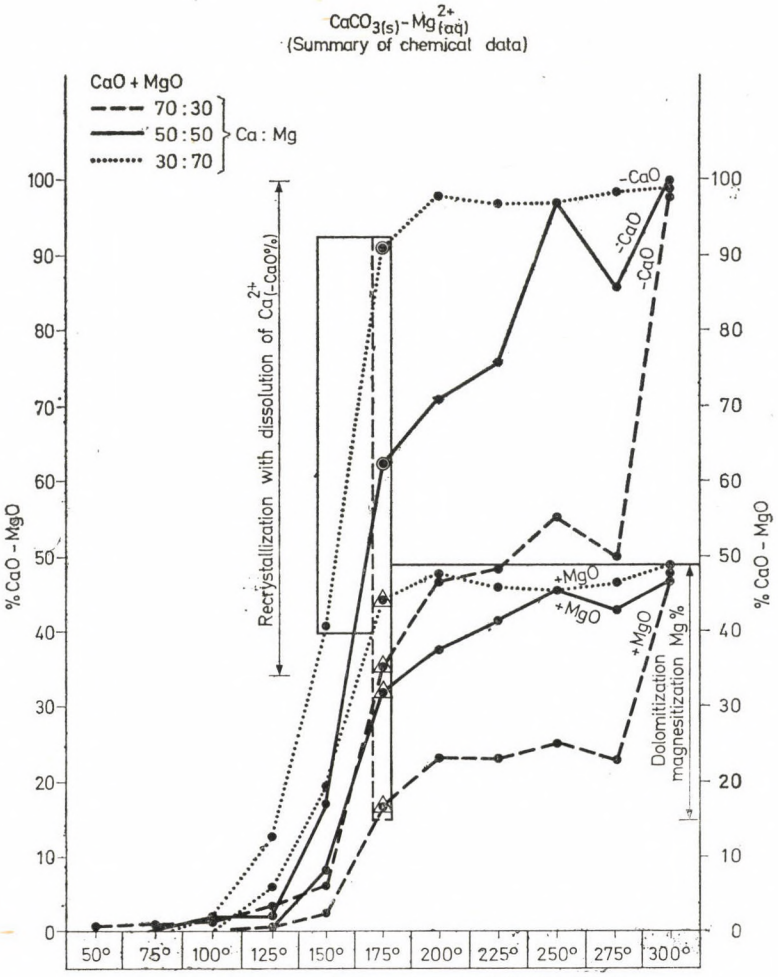


Fig. 3. Diagram of data pesented in Table 1

In the Ca : Mg = 50 : 50 system the rate of the Mg^{2+} fixation increases with temperature, similarly to that of 70 : 30 system. It is optimum values have been found at 250° C, while in the Ca : Mg = 70 : 30 it was at 300 °C. The —46.86% CaO loss and 22.48% MgO increase which have been measured at 300 °C has yet to be checked.

c-2. Crystalline Phases

50 °C = exclusively $CaCO_3$ (calcite),
75 °C = $CaCO_3$ (calcite), $MgCO_3 \cdot 2\,H_2O$ (barringtonite),
100—150 °C = calcite > $MgCO_3$ (magnesite) > $MgCO_3 \cdot 2\,H_2O$. The formation of the magnesite starts at 100° C compared to the 175 °C temperature level in the 70 : 30 system,
175 °C = $CaMg(CO_3)_2 \geqslant MgCO_3 > MgCO_3 \cdot 2\,H_2O >$ calcite.

Note: In the 70 : 30 system the dolomite was in equilibrium with calcite, while here it is equal to $MgCO_3 + CaCO_3$.

225° C = $MgCO_3$ > calcite > $MgCO_3 \cdot 2\,H_2O$. Dolomite completely disappears,
250—275° C = this temperature range has been shown to be optimal for the formation of $MgCO_3$ along with epigenic $MgCO_3 \cdot 2\,H_2O$, without any dolomite or calcite,
300° C = $MgCO_3$, $CaMg(CO_3)_2$ + $MgCO_3 \cdot 2\,H_2O$,

The new stage of formation of dolomite indicates the higher stability of its formation in the 175—300° C temperature range in closed systems.

$$I/1\text{-}d\ CaCO_{3(s)}—Mg^{2+} = 30 : 70$$

d-1. The Chemical Analysis Data (Figs 2—3)

50—100 °C = CaO loss and MgO gain appeared to be similar to the previously described systems (0.06—0.15%),
125 °C = the first stronger change: 12.70% CaO loss and +6.26% MgO increase,
150 °C = 40.80% decrease and +19.61% increase in the CaO and MgO contents, respectively,
175 °C = the composition of the crystalline phase: MgO = 44.08%, CaO = 4.87%. The formation of $MgCO_3$ approaches to optimal rate.
225—275 °C = Average values for the crystalline phase: MgO = 45.77%, CaO = 1.49%,
300 °C = $MgCO_3$ + $MgCO_3 \cdot 2\,H_2O$ + $CaMg_3(CO_3)_4$ has formed, with a composition of:

CaO = 0.71%
MgO = 48.09%
CO_2 = 50.00%

The optimum temperature for the formation has been found to be 300 °C in the most dilute system (70 : 30) of $CaCO_{3(s)}$—$MgCl_{2(aq)}$ compared to 250 °C in the 50 : 50 and 175—300 °C in the 30 : 70 systems.

d-2. Crystalline Phases

50—100 °C = $CaCO_3$ (calcite), $MgCO_3 \cdot 2\,H_2O$,
125—150 °C = $CaCO_3$ (calcite), $MgCO_3$, $MgCO_3 \cdot 2\,H_2O$,
175 °C = $MgCO_3$, dolomite, $MgCO_3 \cdot 2\,H_2O$,
200 °C = $MgCO_3$, $MgCO_3 \cdot 2\,H_2O$ + dolomite (?),
225—275 °C = $MgCO_3$ + $MgCO_3 \cdot 2\,H_2O$, calcite traces at 225 °C,
300 °C = $MgCO_3$ + ($MgCO_3 \cdot 2\,H_2O$ + Ca, $Mg(CO_3)_4$?) .

BABČAN, J. (1977) has described the lower limit of the magnesite formation between 30—50 °C, of the dolomite at <30 °C in natural conditions. In our tests with the calcite : Mg^{2+} systems we did not observe these transitions above 100 °C. The alteration of dolomite to magnesite might, however, start as early as 50 °C. The calcite—dolomite, magnesite unit cell parameters have been computed from XRD data. For XRD analysis NaCl inner standard have been used. The computed cell parameters have shown random variation in comparison with data found in the literature. We took GOLDSCHMITH, J. R. and GRAF, D. L., HEARD, H. G. data (1961) for comparison:

	a_0	c_0	c_0/a_0
calcite	4.990 Å	17.061 Å	3.419 Å
magnesite	4.633 Å	15.016 Å	3.241 Å
dolomite (defective lattice)	4.8050 Å	16.045 Å	3.3392 Å
dolomite (cation ordered)	4.8079 Å	16.010 Å	3.3299 Å

The calcite crystals of the $CaCO_{3(s)}$—$Mg^{2+}{}_{(aq)}$ = 70 : 30 system have unit cells with elongated c_0 parameters — these data are significantly larger than that of GOLDSCHMITH et al. (1961). The largest increase in the unit cell volume has been observed in the 100—125 °C temperature range (= 370.607—371.904), while marked decrease occurred in the higher temperature range.

The lattice-defective dolomite, which has been formed from 175 °C, has unit cells of elongated a_0 parameters and deformed along c_0. At increased temperatures the co-existing $MgCO_3$ has a_0 and c_0 and a_0/c_0 values indicative to deformed cells.

The calcites from the $CaCO_{3(s)}$—$Mg^{2+}{}_{(aq)}$ = 50 : 50 system have larger a_0 and c_0 unit cell parameters than those referred in the literature. The dolomite, which has been formed as the most abundant phase at 175 °C, shows only slight differences from the cation-ordered cells (Δ = —0.0009 Å), with small elongation according to the a_0 and c_0 directions. The $MgCO_3$, which formed at 100 °C, is markedly similar to literature data, the greatest deviations have been observed in the magnesites formed at 125—150 °C and 175 °C (c_0/a_0 = +0.26 Å and —0.079, —0.058 Å). With increasing temperature the c_0/a_0 parameters of the magnesite show increasing similarities with the deformed cell types.

The unit cell deformation of the calcite phases become larger in the $CaCO_{3(s)}$—Mg^{2+} = 30 : 70 system, with largest deviations in the a_0—c_0 parameters (mainly due to c_0 elongation) of the crystals formed at 150 °C.

A few d/Å data are available, a small number, which does not permit the proper evaluation of the dolomite crystals formed at 175 °C (Ca : Mg = = 30 : 70 system). The a_0 lengths of the $MgCO_3$ unit cell are shorter than of those referred in the literature, and these are mainly distorted, shortened in all types, according to their c_0/a_0 values.

The CaO losses and MgO gains in the percentage of the initial amount of $CaCO_3$ in the systems Ca : Mg = 70 : 30, 50 : 50 and 30 : 70 are summarized in Table 1. In order to make these data more relevant we plotted them in Fig. 3 along with the respective crystalline phases.

The results of experiments in the Ca : Mg = 1 : 1.6 and Ca : Mg = = 70 : 30, 50 : 50, 30 : 70 systems showed coincidence with other experimental results (MEDLIN, W. J. 1959) and provided further evidences for that hypothesis which questions the possibility of deposition of syngenetic dolomites along with limestones (at room temperatures).

At room temperature the possibility of adaptation of the Mg ions in the calcite lattice is highly restricted, therefore at depositional temperatures the Mg impurities of the limestone might exist in form of unstable — intermediary huntite, or "proto-dolomite". Consequently, at higher temperatures the limestone may alter diagenetically to dolomite. Above 100 °C, by the effect of Mg^{2+} containing brines, solutions (depending on the ionic concentration of the

Table 1

Ca^{2+}/Mg^{2+}	70:30		50:50		30:70	
C°	CaO%	MgO%	CaO%	MgO%	CaO%	MgO%
50°	−0.14	+0.34	+0.46	+0.14	−0.39	+0.14
75°	−0.80	+0.30	+0.30	+0.03	−0.07	+0.06
100°	−1.23	+0.21	−1.91	+0.025	−1.57	+0.15
125°	−3.55	+0.15	−2.10	+0.61	−12.82	+6.26
150°	−6.12	+2.97	−17.36	+8.42	−40.89	+19.61
175°	−35.49	+16.99	−67.42	+32.31	−91.32	+44.08
200°	−46.29	+23.23	−70.39	+34.00	−98.31	+48.12
225°	−48.38	+23.12	−75.66	+36.57	−96.65	+45.83
250°	−55.10	+25.16	−97.25	+45.47	−97.29	+45.24
275°	−49.90	+23.41	−86.35	+42.24	−98.09	+46.25
300°	−97.81	+47.87	(46.86)	(+22.48)	−98.73	+48.09

Mg) the limestone can alter to dolomite or magnesite by the "compaction" of the calcite lattice. Such geological and geochemical conditions might have existed in the presence of submarine exhalations of Mg^{2+}-rich brines of relatively constant temperatures.

The rhythmic alternation of dolomite and limestone beds may be attributed to the following factors:

a) the "dolomite-beds" were originally Mg-rich limestones with disordered Ca/Mg calcites ("proto-dolomites") and these rocks have subsequently recrystallized to dolomite;

b) the fine-grained and lattice defective calcite particles in the limestone beds have had favourable crystal surfaces $(10\bar{1}4)$ to adapt the magnesium, with subsequent ordering to dolomite.

Living organisms (like calcareous algae) may contain 25—29% $MgCO_3$ in form of basic Mg-carbonate (CHAVE, K. E., 1952). It is probable that the unstable varieties of this basic Mg-carbonate may alter to dolomite diagenetically or by cation exchange in lime-rich environment, like Ca-rich beds, colonies of algae, limestone and basic $MgCO_3$ layers formed from brines (like the epigenetic dolomites after calcite at Shabkha, the dolomite encrustations around sand grains in fluvial sands at the Danube–Tisza Mid-region.

I/2. Powdery Dolomite + $MgCl_2$ — H_2O System

The experiments have been carried out on powdery dolomite samples, obtained from Pilisvörösvár. The chemical analysis of this material is as follows:

SiO_2	0.18%
Al_2O_3	0.22%
Fe_2O_3	0.08%
CaO	30.16%
MgO	21.60%
Na_2O	1.55%
K_2O	0.04%
$-H_2O$	0.05%
CO_2	46.41%
	100.29%

After subtracting the SiO_2, Al_2O_3, Na_2O, K_2O and $-H_2O$ values (−2.12% total), its composition:

CaO	30.72%
MgO	22.00%
CO_2	47.28%
	100.00%

giving a composition of $Ca_{0,5477}Mg_{0,5456}C_{1,0743}O_3$, which is in fact a cation-depleted dolomite. The cation-defective (mainly depleted in Mg^{2+}) mother-phase has had "cation-disordered" dolomite lattice.

After comparison with literature data it can be suggested, that:

magnesite = a) if contains mainly Fe^{2+} Ca^{2+} Mn^{2+} impurities, then in 50% cation depleted;

b) if contains Ca^{2+} Fe^{2+} Mn^{2+}, then in 50% anion depleted.

dolomite: a) 75% of the samples show cation excess (anion loss) with traces of Fe^{2+} and Fe^{2+}; $Fe^{2+}-Mn^{2+}$ trace components;

b) 25% of the samples show anion excess, with $Fe^{2+} + Mn^{2+}$ impurities. It seems that "pure" dolomite is mainly depleted in cations.

The experiments have been carried out in the following systems:

a) $CaMg(CO_3)_{2(s)} + MgCl_{2(aq)}$; Ca : Mg = 70 : 30
b) $CaMg(CO_3)_{2(s)} + MgCl_{2(aq)}$; Ca : Mg = 50: 50
c) $CaMg(CO_3)_{2(s)} + MgCl_{2(aq)}$; Ca : Mg = 30 : 70

T = 50°, 75°, 100°, 125°, 150°, 175°, 200°, 250°, 300 °C

I/2-a. Powdery Dolomite : Mg^{2+} = 70 : 30 (Figs 4—5)

a-1. The Chemical Analysis Data

50—150 °C = deviations less than 2%, 0.79 CaO-loss and 0.70% MgO gain average values,

175 °C = optimum temperature of the dolomite formation, fundamental change in the $MgCl_2 + CaCO_3$ system, with 17.95% CaO loss (58.43% of the total), and 16.13% MgO increase (73.32% of the total),

200 °C = the cation exchange has slowed down, with 6.27% CaO loss and 6.25% MgO gain values,

225—300 °C = on the average 29.94% CaO has been exsolved (−97.46%) and 25.25% MgO has been fixated in the structure of the powdery dolomite. At 300° C almost the total CaO content has been replaced (= 99.15%) by MgO, which has increased from 22.0% to 48.61%.

a-2. Crystalline Phases

50—100 °C = dolomite is dominant, with 1—2% of $MgCO_3$ and $MgCO_3 \cdot 2\,H_2O$,
175 °C = the amount of $MgCO_3$ is approximately equal to that of the dolomite + + $MgCO_3 \cdot 2\,H_2O$. In the $CaCO_3 + MgCl_2 = 70 : 30$ system the temperature level for the formation of dolomite was the lowest among all observed,
200 °C = dominance of $CaMg(CO_3)_2$, the amount of the $MgCO_3 \cdot 2\,H_2O$ is subordinate compared to $MgCO_3$,
225—300 °C = the $MgCO_3$ is the prevailing crystalline phase, the dolomite and the $MgCO_3 \cdot$ $\cdot\, 2\,H_2O$ are only accessorial.

I/2-b. Powdery Dolomite : Mg^{2+} = 50 : 50 (Figs 4—5)

b-1. The Chemical Analysis Data

50—150 °C = 0.56% CaO decrease, 0.45% MgO increase are the mean values,
175 °C = compared to the 70 : 30 system, the amount of the cation exchange is only one-third. CaO decreased with 5.90%, MgO increased with 6.52%,
200 °C = CaO = 15.86% (51.63% decrease) and MgO 14.45% (65.68%) increase. Reaction velocity increases rapidly from 200 °C, but its values between 200° and 250 °C do not reach those of the 70 : 30 system,
225 °C = 21.09% of CaO loss and 19.22% of MgO gain has been observed in the resulted product (= 68.35% CaO decrease and 87.36% MgO increase compared to the initial composition),
250 °C = the total compositional change comprises 70.15% CaO decrease and 86.82% MgO increase compared to the initial material (powdery dolomite),
275—300 °C = the cation exchange is almost complete; 275 °C: CaO = 0.91% (= −97.04% CaO, MgO = 40.59% (+84.50%); 300 °C: CaO = 0.25% (−99.19%); MgO = = 47.66% (+116.36%).

b-2. Crystalline Phases

50—150 °C = dolomite is dominant, $MgCO_3$ and $MgCO_3 \cdot 2\,H_2O$ are accessories. Calcite has formed only at 125 °C,
175 °C = dolomite is predominant, $MgCO_3$—$MgCO_3 \cdot 2\,H_2O$ + aragonite are accessories,
200—225 °C = $MgCO_3$ + dolomite + $MgCO_3 \cdot 2\,H_2O$ (at 200 °C) + $MgCO_3 \cdot 3\,H_2O$ — nesquehonite (225 °C) + aragonite (225 °C),
250 °C = $MgCO_3$ + dolomite; the calcite, aragonite and $MgCO_3 \cdot 2\,H_2O$ are subordinate,
275 °C = $MgCO_3$ + 1—2% dolomite + $MgCO_3 \cdot 2\,H_2O$,
300 °C = 95—96% $MgCO_3$ + 4—5% $MgCO_3 \cdot 2\,H_2O$ + $MgCO_3 \cdot 3\,H_2O$ + $CaMg(CO_3)_2$ impurities.

I/2-c. Powdery Dolomite : Mg^{2+} = 30 : 70 (Figs 4—5)

The variations in CaO and MgO contents can be classified into five and four groups respectively.[1]

c-1. The Chemical Analysis Data

Average values of changes in CaO content:

50— 75 °C (1)	= the amount of CaO replaced	0.97%	(= 3.15%)
100—125 °C (2)	= the amount of CaO replaced	0.52%	(= 1.69%)
150—175 °C (3)	= the amount of CaO replaced	1.77%	(= 5.76%)
200—225 °C (4/a)	= the amount of CaO replaced	9.65%	(= 31.41%)
	and	10.75%	(= 34.99%)
250 °C (5)	= the amount of CaO replaced	6.18%	(= 20.12%)
275—300 °C (4/b)	= the amount of CaO replaced	9.92%	(= 32.27%)
	and	10.29%	(= 33.59%)

[1] Average values has been calculated from the chemical analysis data for approximately similar temperature levels and these values have been accepted as representatives for the given temperature ranges.

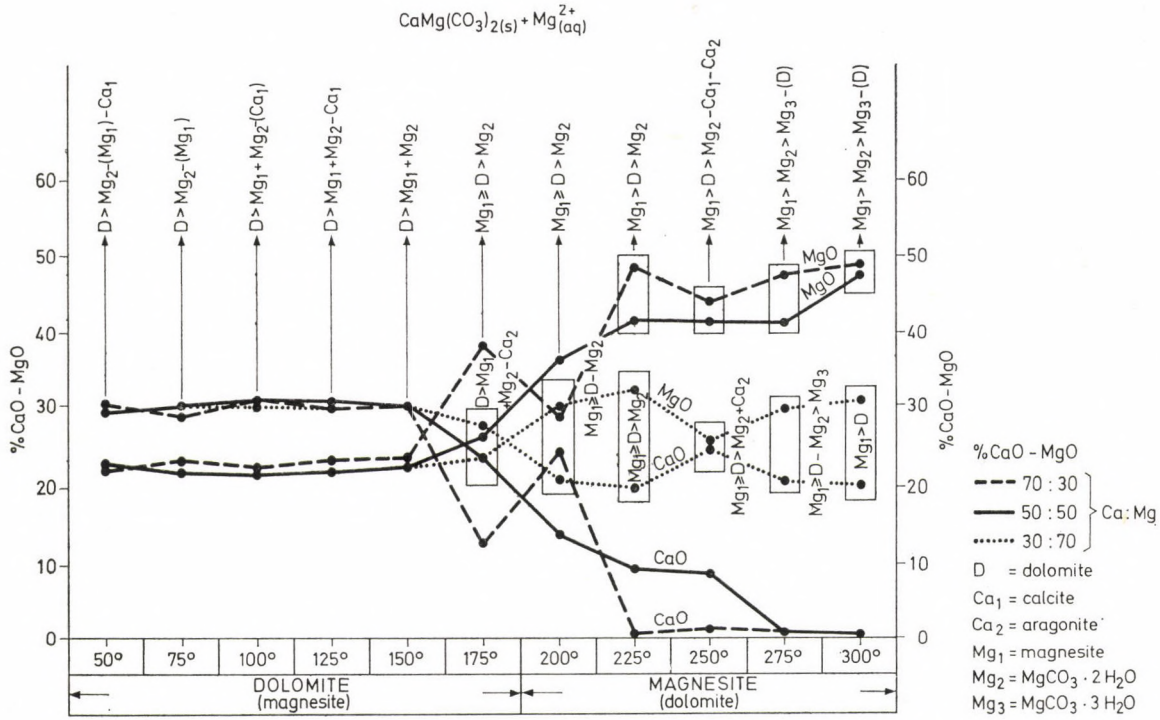


Fig. 4. The CaO—MgO distributions and crystalline phases of the dolomite $+Mg^{2+}_{(aq)}$ systems

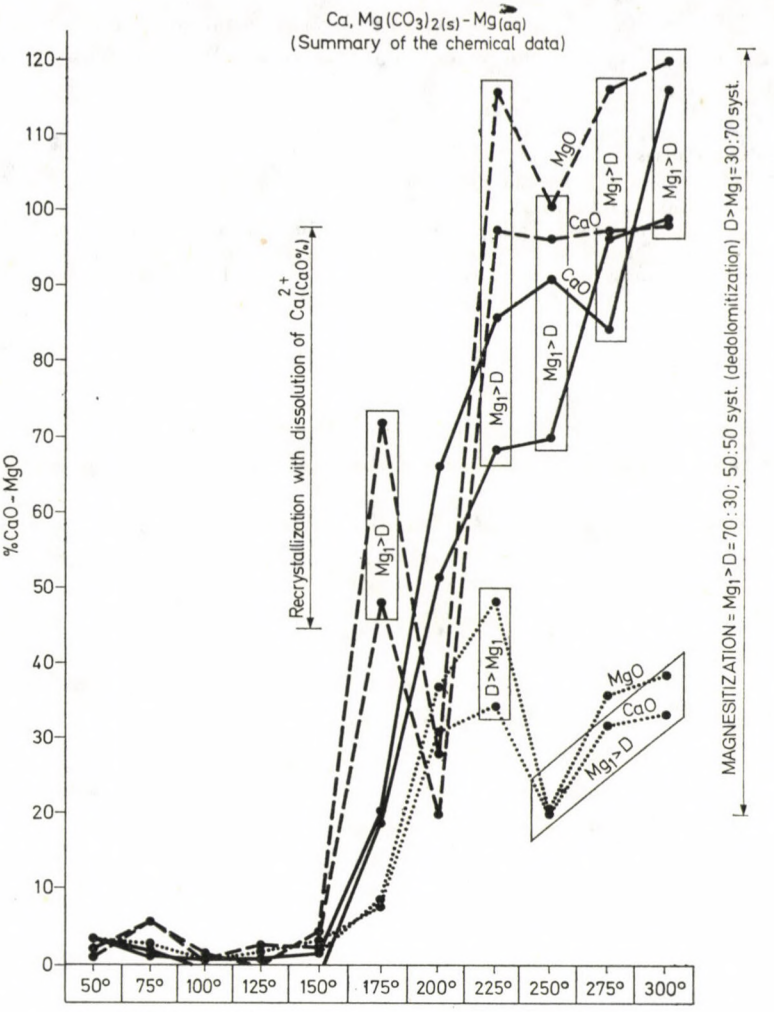


Fig. 5. Diagramatic representation of data shown in Table 2

Average values of changes in MgO contents:

50—175 °C (1)	= the amount of MgO fixated	0.74%	(+3.37%)
200 °C (2/a)	= the amount of MgO fixated	8.22%	(+37.36%)
225 °C (3)	= the amount of MgO fixated	10.94%	(+49.72%)
250 °C (4)	= the amount of MgO fixated	4.51%	(+20.50%)
275—300 °C (2/b)	= the amount of MgO fixated	8.29%	(+37.68%)

Surprisingly, the high Mg^{2+} concentration ($30 : 70 = Ca^{2+} : Mg^{2+}$) adversely affects the replacements of Ca^{2+} in the dolomite, an observation similar to that of the $50 : 50 = Ca^{2+} : Mg^{2+}$ system.

The geochemical—crystal-chemical significance of the dolomite—Mg^{2+} system is in the observation which indicated that dilute Mg^{2+} – solutions

(Mg^{2+} : Ca^{2+} = 30 : 70) are favourable for the total replacement of the Ca^{2+} ions in the dolomite, and the process can become complete from 225 °C theoretically.

c-2. Crystalline Phases

50—175 °C = dolomite is dominant, $MgCO_3$ has formed at 50—175 °C, calcite at 75—125 °C, $MgCO_3 \cdot 2\ H_2O$ at 75—175 °C,

200—250 °C = the amount of $MgCO_3$ does not reach 50% of the total dolomite. $MgCO_3 \cdot 2\ H_2O$ has formed at 200°, 225° and 250 °C, aragonite at 250 °C.

175—300 °C = dolomite predominance with associated 15—18% $MgCO_3$. $MgCO_3 \cdot 2\ H_2O$ has formed at 275—300 °C, $MgCO_3 \cdot 3\ H_2O$ at 275 °C.

The results of the experiments with the dolomite—$MgCl_2$ system have brought interesting conclusions:

a) the most diluted solutions (Ca : Mg = 70 : 30) are the most favourable for the $Ca^{2+} \rightarrow Mg^{2+}$ exchange, where at

225 °C = 0.75% CaO
250 °C = 1.25% CaO
275 °C = 0.75% CaO
300 °C = 0.36% CaO

has been found to be associated with the $MgCO_3$ (the CaO impurity of the natural $MgCO_3$ normally falls between 0.33—1.66%).

b) In the Ca : Mg = 50 : 50 system (powdery dolomite) the optimal cation exchange reaction rate has fallen in the range of 275—300 °C, with $MgCO_3$ 0.91% and CaO 0.25% participation.

c) The replacement of Ca^{2+} is hindered by high Mg^{2+} concentrations in the solution, and only one-third of the Ca^{2+} content of the powdery dolomite can be replaced. Dolomite remains the dominant phase, while $MgCO_3$, $MgCO_3 \cdot 2H_2O$, $MgCO_3 \cdot 3H_2O$, calcite and aragonite are subordinate.

I/3. The Genetic Interpretation of the Calcite — Mg^{2+} and the Dolomite — Mg^{2+} Solution Systems

a) Mg^{2+} rich hydrothermal solution promotes the alteration of limestone to magnesite.

b) The Ca^{2+} exchange of dolomite (powdery dolomite) may be taken place by extremely dilute Mg^{2+} containing hydrothermal solutions. The source of Mg^{2+} can be the release of Mg^{2+} during the serpentinization of basic, ultrabasic rocks, or the migration of hypersaline brines, Mg^{2+} rich evaporate-formed solutions the deeper zones.

c) The "de-dolomitization", as a process of cation exchange in a solid phase : solution system can be encountered only with respect to the dolo-

mite : Mg^{2+} system. Reversed process can not be undergone, or its possibility is very small (dolomitization of magnesite — discussed later). The process is similarly irreversible as the well-known calcite : Mg^{2+} systems.

Table 2

Ca^{2+}/Mg^{2+}	70 : 30		50 : 50		30 : 70	
T C_s	CaO%	MgO%	CaO%	MgO%	CaO%	MgO%
50°	−2.08	+1.95	−3.55	+3.55	−3.15	+2.77
75°	−5.63	+5.68	−2.05	+1.64	−3.13	+2.86
100°	−0.20	+1.14	−0.94	+1.14	−1.05	+0.36
125°	−2.64	−0.55	−0.55	+1.14	−2.34	+2.50
150°	−2.24	+4.50	−1.73	+2.73	−2.54	+3.41
175°	−58.43	+73.32	−19.21	+20.55	−8.98	+8.32
200°	−20.41	+28.41	−51.62	+66.14	−31.41	+37.36
225°	−97.56	+120.82	−68.65	+87.36	−34.99	+49.73
250°	−95.93	+100.73	−70.45	+91.36	−20.12	+20.50
275°	−97.56	+116.50	−97.04	+84.50	−32.29	+36.32
300°	−98.83	+121.05	−99.19	+116.63	−33.59	+39.00

In Table 2 we summarized to CaO losses and MgO increases in the crystalline phase formed in the dolomite + Mg^{2+} systems, and compared to the CaO and MgO values of the powdery dolomite (30.72 and 22.00% respectively). The variations of data, along with the indication of crystalline phases produced, are plotted in Fig. 5.

II. "De-dolomitization", "Re-calcitization"
II/1. $MgCO_3$ (Magnesite) + $CaCl_2$ — H_2O System

(the re-calcitization of the magnesite)

We used white crystalline magnesite from St. Kathrein (Bruck an der Mur — Austria) in order to ascertain the proper identification of the chemical changes in the models; chemical and X-ray analyses have been completed on the standard samples (analysed by Mrs. Nenes L. and Guzy K.):

SiO_2 = 0.81%
Al_2O_3 = 0.42%
Fe_2O_3 = 0.99%
MgO = 47.02%
CaO = 0.64%
Na_2O = 0.11%
K_2O = 0.04%
CO_2 = 49.96%
H_2O = 0.04%
Total = 100.03%

The theoretical composition of the magnesite after subtracting the SiO_2, Al_2O_3, Fe_2O_3, Na_2O, K_2O contents (2.41%):

$$\begin{aligned} MgO &= 48.17\% \\ CaO &= 0.65\% \\ CO_2 &= 51.18\% \\ \hline \text{Total} &= 100.00\% \end{aligned}$$

This gives a stoichiometric composition of $Mg_{1,195}Ca_{0,011}C_{1,163}O_3$, indicating a cation-depleted structure.

The experiments have been carried out by 1.0 and 2.5 M $CaCl_2$ solutions, in which $Mg^{2+} : Ca^{2+} = 2 : 1$; $1 : 1$ and $1 : 2$ cation ratios have been chosen.

II/1-a. $MgCO_{3(s)} : Ca^{2+} = 2 : 1$ (1.0 M solution)

a-1. The Chemical Analysis Data (Fig. 6)

With respect to CaO the new crystalline phases can be grouped to distinct compositional ranges:

1. 50—100 °C = 0.86—3.21% CaO
2. 125—175 °C = 4.47—4.60% CaO
3. 200—225 °C = 5.39—5.38% CaO
4. 250—300 °C = 7.54—7.73% CaO

Comparing the average values to the magnesite standard:

1. 50—100 °C = + 2.31% CaO
2. 125—175 °C = + 4.02% CaO
3. 200—225 °C = + 4.74% CaO
4. 250—300 °C = + 7.00% CaO

increase has been recorded. These values can rather be attributed to different $CaCO_3$ varieties overgrown on magnesite crystals than Ca fixation in the positions of the magnesite lattice.

By comparing the MgO content to $CaCO_3$ by non-equipartial averages, a negative correlation has been observed:

1. 50—100 °C = 46.45% = 1.72% MgO deficiency
2. 125—175 °C = 43.68% = 4.49% MgO deficiency
3. 200—225 °C = 41.15% = 7.02% MgO deficiency
4. 250—300 °C = 38.69% = 9.48% MgO deficiency

The analyses indicate that more Mg^{2+} is exsolved from the magnesite than the amount of Ca in independent phases (Ca_1, Ca_2, Ca_3, D) produced from the "mother-solution".

The CO_2 content of the crystalline phases moderately decreases with the increase of temperature (anion-deficiency). A sharp decrease was observed at 200 °C (= 4.42% CO_2), with contemporaneous increase of the H_2O and Cl content. This is partly due to the formation of basic $MgCO_3 \cdot 2H_2O$. The Cl is probably behaves as substituting anion, since despite of its good solubility, washing treatments of several weeks duration could not remove it from the system.

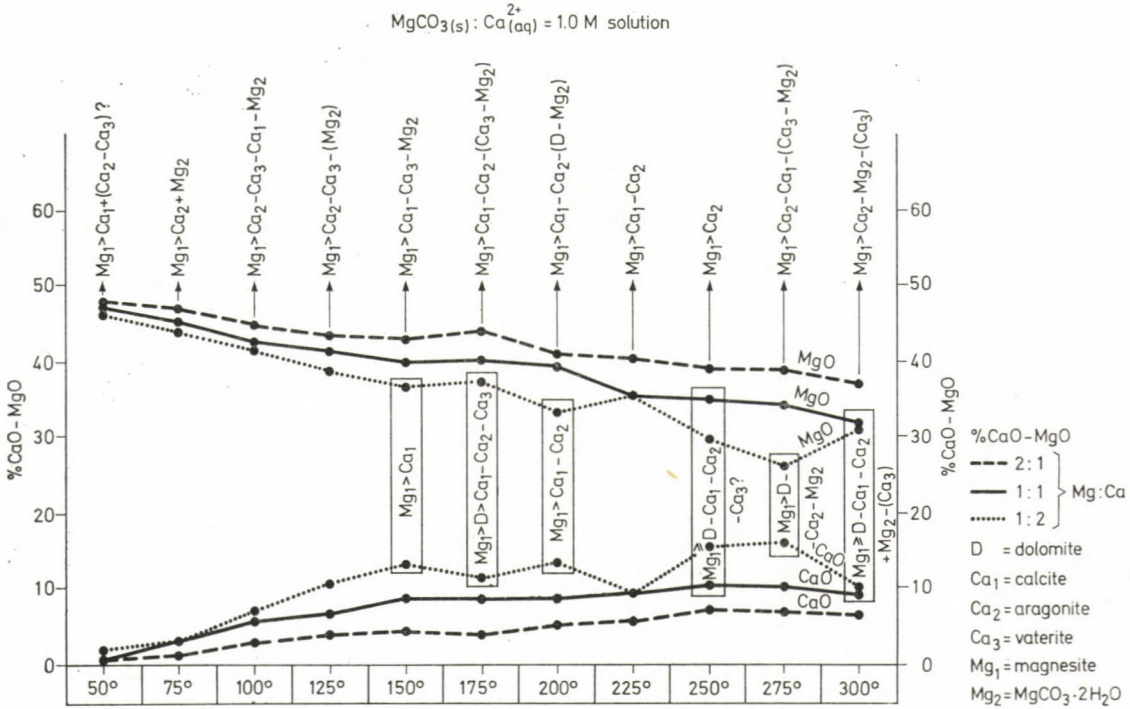


Fig. 6. **CaO—MgO distributions in the crystalline phases (magnesite $+Ca^{2+}$ = 1.0 M solution, Mg : Ca = 2 : 1, 1 : 1, 1 : 2)**

II/1-b. $MgCO_{3(s)} : Ca^{2+} = 1 : 1$
(1.0 M solution)

b-1. The Chemical Analysis Data (Fig. 6)

If we consider the cation ratio of the solid phase and the solution, in theory the total amount of magnesium can be replaced by calcium. Compared to the 2 : 1 system significant changes have been observed in the chemical compositions of the crystalline phases:

a) First, in the 50—75 °C interval 0.89 and 3.38% CaO surplus have been measured at 50 °C and 75 °C respectively. In the 100—300 °C range the amount of CaO increases with one per cent for every 25 °C temperature rise.

b) At 50 °C the loss of MgO is larger than in the Mg : Ca = 2 : 1 system. From 75 °C upwards the rate of decrease is larger than 1%. In the 150—175 °C and 225—250 °C intervals the decrease of MgO content has shown similar amounts, while between 275° and 300 °C the MgO loss was again over 1%.

The average compositions showed the following deviations from the original magnesite standard:

CaO:
1. 50—100 °C = + 2.71% CaO excess
2. 125—175 °C = + 7.38% CaO excess
3. 200 °C = + 7.96% CaO excess
4. 225 °C = + 9.79% CaO excess
5. 250—300 °C = + 9.04% CaO excess

MgO:
1. 50—100 °C = − 3.09% MgO deficiency
2. 125—175 °C = − 7.45% MgO deficiency
3. 200 °C = − 9.35% MgO deficiency
4. 225 °C = −12.72% MgO deficiency
5. 275—300 °C = −14.67% MgO deficiency

The H_2O contents of the crystalline phases above 175 °C have increased over 1% due to the formation of basic Mg-carbonate. However, a simultaneous loss of CO_2 have been observed, but it is not comparable with the increasing of Cl.

II/1-e. $MgCO_{3(s)} : Ca^{2+} = 1 : 2$
(1.0 M solution)

c-1. The Chemical Analysis Data (Fig. 6)

With the further increase of cation ratios in both the solid phase and the solution the variations of the chemical compositions have shown larger differentiations:

CaO:
1. 50—100 °C = + 3.49% excess
2. 125—175 °C = +11.25% excess
3. 200 °C = +12.98% excess
4. 225 °C = + 8.64% excess
5. 250—300 °C = +13.47% excess

MgO:
1. 50—100 °C = − 3.95% deficiency
2. 125—175 °C = −10.54% deficiency
3. 200 °C = −14.51% deficiency
4. 225 °C = −12.56% deficiency
5. 250—300 °C = −18.22% deficiency

The distribution of H_2O:

50— 75 °C = 1.23—0.91%
100—175 °C = 0.80—0.52%
200—275 °C = over 2%

These values indicate the increasing rate of formation of basic Mg-carbonate. The distribution of CO_2 shows small variations only up to 175 °C, but its concentration drops to less than 39—40% to 200 °C. The chloride content of the crystalline phase — with the exception of the product at 50 °C — is in the order of 0.1%, increasing to 0.4% in the higher temperature intervals.

By evaluating the analyses of the three systems with different cation ratios (2 : 1, 1 : 1, 1 : 2 = Mg : Ca) in a given temperature level, the following relationships can be seen:

1. The MgO content of the crystalline phases at the same temperature level is related to the initial cation ratios. The MgO decrease is three-fold in the Mg : Ca = 1 : 2 system than in the 2 : 1 system. In other words, the larger is the Ca^{2+} concentration in the solution, the higher is the MgO loss.

2. The total CaO content of the crystalline phases is roughly proportional with the Mg : Ca ratios in the solid-phase : solution system. The larger is the initial Ca^{2+} concentration in the solutions the higher is the CaO content in the crystalline phases. If the concentration of the Ca^{2+} ions in the solution is increased, so that Ca^{2+} does not enter into the structure, the removal of Mg^{2+} is also intensified without further epigenetic crystal formation.

The CaO—MgO distributions and trends in the crystalline phases of the Mg : Ca = 2 : 1, 1 : 1, 1 : 2 systems (1 M solutions) are shown in Fig. 6.

II/1-a. $MgCO_{3(s)}$: Ca^{2+} = 2 : 1 Crystalline Phases
(1.0 M solution)

50 °C = $Mg_1 \gg Ca_1 > (Ca_2-Ca_3)$?;
75 °C = $Mg_1 \gg Mg_2-Ca_2$;
100 °C = $Mg_1 \gg Ca_2-Ca_3-Mg_2$;
125 °C = $Mg_1 \gg Ca_2-Ca_3-Mg_2$;
150 °C = $Mg_1 \gg Ca_1-Ca_2-Mg_2$;
175 °C = $Mg_1 \gg Ca_1-Ca_2$;
200 °C = $Mg_1 \gg D + Ca_1-Ca_2$;
225 °C = $Mg_1 \gg Ca_1-Ca_2$;
250 °C = $Mg_1 \gg Ca_2$;
275 °C = $Mg_1 > Ca_1-Ca_2-(Ca_3-Mg_2)$;
300 °C = $Mg_1 \gg Ca_2-Mg_2-(Ca_3)$?.

$D = CaMg(CO_3)_2$; $Mg_1 = MgCO_3$; Ca_1 = calcite(trigonal); Ca_2 = argonite (rhombic); Ca_3 = vaterite (hexagonal), $Mg_2 = MgCO_3.2H_2O$(triclinic).

The results show that $MgCO_3$ is the most stable dominant phase through the whole temperature range (50—300 °C). Dolomite has formed only at 200 °C, but its amount is less than expected from the $MgCO_3$ content of the initial phase. Some Mg^{2+} positions of the magnesium-carbonate crystals are

replaced by Ca^{2+}. The exsolved Mg^{2+} crystallizes in form of separate epigenic phase along with Ca^{2+}. The Ca_2—aragonite structure has formed at every temperature level but 150 °C, the vaterite between 100—150 °C, the Ca_1—calcite at 50°, 150°, 175°, 200°, 225°, 275 °C temperatures. Basic Mg-carbonate ($MgCO_3 \cdot 2H_2O$ = barringtonite) has formed at 100—150 °C, 300 °C. Its presence is uncertain at 275 °C, and it was not found in the 50—75 °C and 175—250 °C temperature intervals. The accessorial phases in association with the $MgCO_3$ "mother-phase" are crystallized products of the ions of the solution, epigenetic precipitations on the initial phase, and did not take part in the s.str. "metasomatic" process. This question is discussed in details in the geological interpretation.

II/1-b. $MgCO_{3(s)} : Ca^{2+} = 1 : 1$ Crystalline Phases
(1.0 M solution)

50 °C = $Mg_1 \gg Ca_2$;
75 °C = $Mg_1 \gg Mg_2 - Ca_2$;
100 °C = $Mg_1 \gg Ca_2 - Ca_1 - (Ca_3)$?;
125 °C = $Mg_1 \gg Ca_2 - (Mg_2)$?;
150 °C = $Mg_1 \gg Ca_1 - Ca_2 - Ca_3$;
175 °C = $Mg_1 \gg Ca_1 - Ca_2 - (Ca_3 - Mg_2)$?;
200 °C = $Mg_1 \gg Ca_2 - Mg_2$;
225 °C = $Mg_1 \gg Ca_1 - Ca_2$;
250 °C = $D \geqslant Mg_1 - Ca_1 - (Ca_2)$?;
275 °C = $Mg_1 > Ca_2 - Mg_2 - D$;
300 °C = $Mg_1 \gg D - Mg_2 - (Ca_3)$?;

$Mg_1 = MgCO_3$; D = dolomite; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterite.

The higher Ca^{2+} concentration does not promote the structural ordering of the $MgCO_3$, since $MgCO_3$ continues to be the dominant phase. It is clear from the analysis data, that gradual increase of CaO content in the crystalline phases is shown, with maximum values in the Mg : Ca = 1 : 2 system. The Ca_2—aragonite — has formed at every temperature level, the Ca_1–calcite structure has been present at 150°, 175°, 225°, 250 °C temperatures. The hexagonal vaterite was observed at 150 °C (Ca_3), its presence was uncertain at 100°, 175° and 300 °C. Dolomite has crystallized at 250° and 275 °C, but its amount was subordinate to magnesite.

II/1-c. $MgCO_{3(s)} : Ca^{2+} = 1 : 2$ Crystalline Phases
(1.0 M solution)

50 °C = $Mg_1 \gg Ca_1 - Ca_2 - (Ca_3)$?;
75 °C = $Mg_1 \gg Ca_2$;
100 °C = $Mg_1 \gg Ca_1 - Ca_2 - Ca_3$;
125 °C = $Mg_1 \gg Ca_2 - Ca_3 - Mg_2$;
150 °C = $Mg_1 \gg Ca_1$;
175 °C = $Mg_1 \gg D - Ca_1 - Ca_2 - Ca_3$;

$$200\ °C = Mg_1 \gg Ca_1 - Ca_2;$$
$$225\ °C = Mg_1 \gg Ca_1 - Ca_2;$$
$$250\ °C = D \gg Mg_1 - (Mg_2 - Ca_1 - Ca_3);$$
$$275\ °C = Mg_1 > Mg_2 - D;$$
$$300\ °C = D \gg Mg_1 - Ca_1 - Ca_2.$$

$Mg_1 = MgCO_3$; D = dolomite; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterite.

The XRD data indicate that at certain temperatures the higher Ca^{2+} concentrations are favourable for the development of dolomite structure (175°, 250°, 275°, 300 °C). In cases the amount of dolomite exceeds that of the "mother-phase" magnesite (250°, 300 °C). This is not considered as a result of structural re-arrangement of the magnesite lattice, rather the crystallization of dissolved ions in form of new epigenetic phases, aggregates, crystal nuclei, separated from the $MgCO_3$.

The formation of magnesite from powdery dolomite, however, has been promoted by dilute Mg^{2+} solutions, and the process has been completed without the dissolution of the initial phase.

The microphotographs of the crystalline precipitations are shown in Plate VI. Photos: 23—24, Plate VII. Photos: 26—28 and Plate VIII. Photos: 29—32.

Upon these results it is suggested that the cations, which have been present, or subsequently entered into the solution, are crystallized in separate crystal phases, in cases overgrowing the rhombohedral planes of magnesite crystals. Crystals with euhedral habits, pseudo-hexagonal prisms, laths are frequently seen (e.g. aragonite). In the majority of cases the absorbed, dissolved magnesite rhombohedra are observed along with xenomorfic encrustations of $CaCO_3$ modifications and/or basic Mg-carbonates.

At higher temperatures (200—300 °C) small lamellae have formed on the surfaces of the magnesite crystals (Plate VI. Photos: 23—24), which are in fact aggregates of precipitated dolomite and basic Mg-carbonate. In these aggregates the remnants of the original Mg-carbonate rhombohedra can still be recognized (Plate VII. Photo: 25).

II/1-d. $MgCO_{3(s)} : Ca^{2+} = 2 : 1$
(2.5 M solution)

d-1. The Chemical Analysis Data (Fig. 7)

Compared to the 1.0 M $CaCl_2$ system, the crystalline phases show a more balanced compositional series:

CaO: 50—125 °C = +3.23 excess
150—175 °C = +4.08% excess
200—225 °C = +5.25% excess
250—300 °C = +6.49% excess

MgO: 50—125 °C = −1.77% deficiency
150—175 °C = −4.04% deficiency
200—225 °C = −6.21% deficiency
250—300 °C = −8.25% deficiency

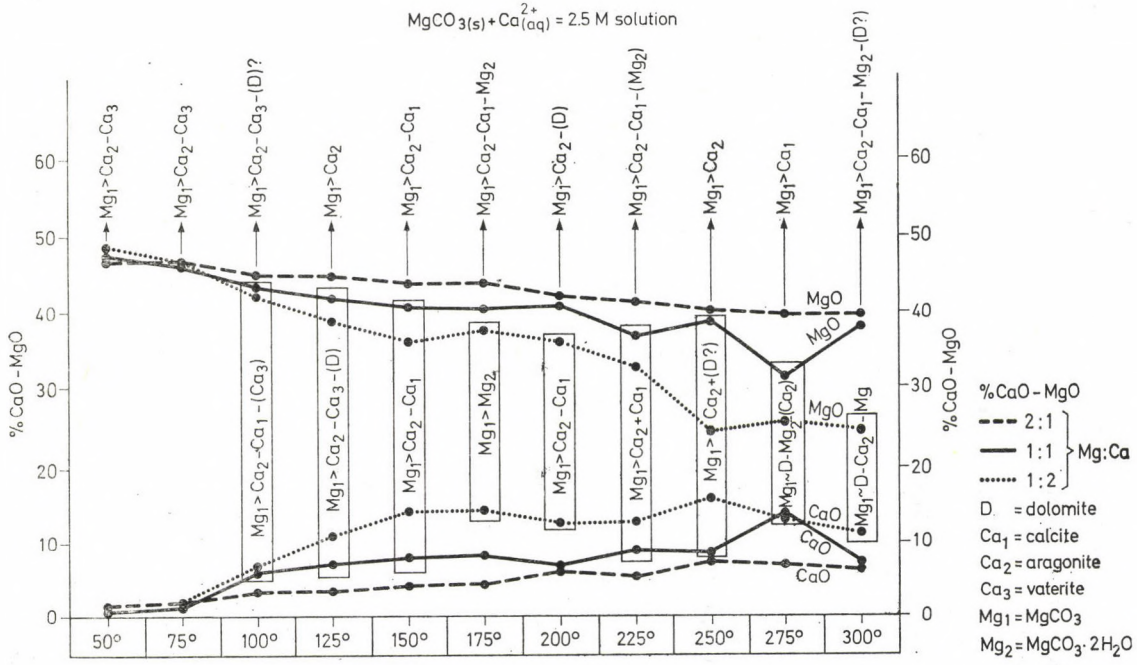


Fig. 7. The variations of the CaO—MgO values in the $MgCO_{3(s)} + Ca^{2+}{}_{(aq)} = 2:1, 1:1, 1:2 = Mg:Ca$ systems (2.5 M solutions)

The CaO versus MgO comparison indicates that in the 50—175 °C interval the amount of removed (dissolved) magnesium is equal to the amount of the precipitated calcium. Above 200 °C the MgO loss is about 1—2% larger than the CaO excess.

II/1-e. $MgCO_{3(s)} : Ca^{2+} = 1 : 1$
(2.5 M solution)

e-1. The Chemical Analysis Data (Fig. 7)

New temperature intervals are required for the subdivision of the chemical data:

CaO: 50— 75 °C = + 0.99% excess
100—200 °C = + 6.72% excess
225—250 °C = + 8.43% excess
275—300 °C = +10.15% excess

MgO: 50— 75 °C = — 1.10% deficiency
100—200 °C = — 6.48% deficiency
200—250 °C = —10.39% deficiency
275 °C = —16.65% deficiency
300 °C = —10.09% deficiency

The increasing average loss of magnesium is coupled by the increase of the amount of calcium in the new crystal phases. In the 50—200 °C interval the increase of CaO is proportional with the amount of dissolved magnesium, but above 225 °C 2.36—3.27% more magnesium is dissolved than the amount of calcium crystallized in solid phase.

II/1-f. $MgCO_{3(s)} : Ca^{2+} = 1 : 2$
(2.5 M solution)

f-1. The Chemical Analysis Data (Fig. 7)

As has been observed in the more dilute systems, the larger were the cation ratios in the solid phase ($MgCO_3$) versus solution, the more proportional the trends of chemical compositions of the precipitated crystal phases were:

CaO: 50— 75 °C = + 1.32% excess
100 °C = + 6·58% excess
125—225 °C = +12.72% excesses
250 °C = +15.56% excesses
275—300 °C = +12.08% excess

MgO: 50—75 °C = — 0.76% deficiency
100 °C = — 5.98% deficiency
125—225 °C = —11.96% deficiency
250 °C = —18.34% deficiency
275—300 °C = —17.76% deficiency.

The higher rate of dissolution of magnesium is indicated by the higher average values.

The data indicate that in the 50—75 °C interval the CaO—MgO distribution of the crystal phases in solutions with low Ca^{2+} concentrations is two-three times lower than the CaO content of these products, which have been formed in Ca : Mg = 2 : 1 solution systems.

The variation of the CaO—MgO values are shown in Fig. 7.

II/1-d. $MgCO_{3(s)}$: Ca^{2+} = 2 : 1 Crystalline Phases
(2.5 M solution)

50 °C = $Mg_1 \gg Ca_2$—Ca_3;
75 °C = $Mg_1 \gg Ca_2$;
100 °C = $Mg_1 \gg Ca_2$—Ca_3—(D?);
125 °C = $Mg_1 > Ca_2$;
150 °C = $Mg_1 > Ca_2$—Ca_3;
175 °C = $Mg_1 > Ca_2$—Ca_1;
200 °C = $Mg_1 > Ca_2$;
225 °C = $Mg_1 \gg Ca_2$—Ca_1—(Mg_2?);
250 °C = $Mg_1 > Ca_2$;
275 °C = $Mg_1 \gg Ca_2$;
300 °C = $Mg_1 \gg$ D(?)—(Ca_2—Mg_2)?

$Mg_1 = MgCO_3$; D = dolomite; Ca_2 = aragonite; Ca_3 = vaterite, Ca_1 = calcite; $Mg_2 = MgCO_3 \cdot 2\,H_2O$.

The distributions and compositions of the crystalline phases undoubtedly show the restricted replaceability of the Mg^{2+} positions by Ca^{2+}. In some systems (e.g. 1 : 1 and 1 : 2 = Mg : Ca) the removal of Mg^{2+} (0.66 Å = = AHRENS) from the lattice of the "mother-phase" is increasing with temperature. These positions can be filled by larger Ca^{2+} ions (0.99 Å = AHRENS) only in those cases where vacancies and defects adjacent to the Mg^{2+} positions provide increased space in the lattice structure. The possibility of the formation of dolomite as intermediate phase after magnesite is highly restricted. The different $CaCO_3$ phases are not metasomatic products but epigenetic precipitations from the solution.

In the Mg : Ca = 2 : 1 system rhombohedral Ca-carbonate (calcite) has formed only at 175° and 225 °C temperatures, while the orthorhombic variety (aragonite) was much more dominant. The hexagonal $CaCO_3$ (vaterite) was identified at 50°, 100°, and 150 °C. The uncertainty of the existence of $MgCO_3 \cdot 2H_2O$ was surprising, a significant amount of the dissolved magnesium has formed chloride-complexes, and remained in dissolved form.

II/1-e. $MgCO_{3\,(s)}$: Ca^{2+} = 1 : 1 Crystalline Phases
(2.5 M solution)

50 °C = $Mg_1 \gg Ca_2$—Ca_3;
75 °C = $Mg_1 \gg Ca_2$—Ca_3;
100 °C = $Mg_1 \gg Ca_2$—(Ca_3)?;
125 °C = $Mg_1 \gg Ca_2$—Mg_2—D;
150 °C = $Mg_1 > Ca_2$;

$$175\ °C = Mg_1 \gg Ca_1-(Ca_2-Mg_2)?;$$
$$200\ °C = Mg_1 > Ca_2-D;$$
$$225\ °C = Mg_1 > Ca_2;$$
$$250\ °C = Mg_1 \gg Ca_2(D)?;$$
$$275\ °C = D \approx Mg_1-Ca_2?;$$
$$300\ °C = Mg_1 \gg (Ca_1-Ca_2).$$

$Mg_1 = MgCO_3$; D = dolomite; Ca_1 = calcite, Ca_2 = aragonite, Ca_3 = vaterite.

The products of this system are characterized by the aragonite-$CaCO_3$, and hexagonal vaterite-$CaCO_3$. Rhombohedral $CaCO_3$ calcite-structure has formed at 175° and 300 °C. The d/Å lines of dolomite at 125 °C and 250 °C are similar to that of proto-dolomites. Its amount at 275 °C was equal to that of magnesite. Small amounts of basic Mg-carbonate were observed at 125 °C and 175 °C (4—5 per cent).

II/1-f. $MgCO_{3(s)} : Ca^{2+} = 1 : 2$ Crystalline Phases (2.5 M solution)

There were only slight differences from the above discussed systems:

$$50\ °C = Mg_1 \gg Ca_3-Ca_2;$$
$$75\ °C = Mg_1 \gg Ca_2-Ca_3;$$
$$100\ °C = Mg_1 \gg Ca_2-Ca_1;$$
$$125\ °C = Mg_1 \gg Ca_2-D-Ca_3;$$
$$150\ °C = Mg_1 > Ca_2-Ca_1;$$
$$175\ °C = Mg_1 > Mg_2;$$
$$200\ °C = Mg_1 > Ca_1-Ca_2;$$
$$225\ °C = Mg_1 \gg Ca_1-Ca_2;$$
$$250\ °C = Mg_1 \gg Ca_2;$$
$$275\ °C = Mg_1 > Mg_2(D?);$$
$$300\ °C = Mg_1 \approx D-(Ca_2-Mg_2).$$

$Mg_1 = MgCO_3$; D = dolomite; $Mg_2 = MgCO_3 \cdot 2\ H_2O$; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterite.

At 300 °C the amount of dolomite and magnesite was similar. At 175° and 250 °C only two phases have been recorded in the system, the magnesite was associated by basic Mg-carbonate and aragonite respectively. Aragonite has appeared to be the significant $CaCO_3$ variety, while calcite and vaterite was formed only at certain temperature levels.

The microphotographs of the resulted crystal phases are shown in Plates VI—VIII.

In the $CaCO_{3(s)} + Mg^{2+}{}_{(aq)}$ system the gradual replacement of calcite rhombohedra by $MgCO_3$ has been observed, due to structural positioning of Mg^{2+}.

Since the migration of Ca^{2+} into the solution was faster than the fixation of Mg^{2+}, voids had formed in the calcite rhombohedra and, according to the decrescency theory, these voids have been filled in with $MgCO_3$ rhombohedra without the contemporaneous dissolution of the "mother-phase"

($CaCO_3$). (Photos: 13, 15.) In transitional case $MgCO_3$ overgrowth has framed the calcite rhombohedron. This phenomenon has not been observed during the microscopic investigations of these latter systems.

In the 2.5 M $CaCl_2$ solution systems similar phenomena to those of the more dilute (1.0 M) solution systems were recognised. The differences were mainly due to the structural ordering and grain size distribution of the associated Ca_1, Ca_2, Ca_3 and Mg_2 phases.

The most common features are the differently shaped Ca_2-Ca_3 crystalline particles overgrown on the rhombohedral planes of the $MgCO_3$ (Photos 26, 27, 28). The aragonite (Ca_2) is frequently developed in fibrous, pinnate, lath-like habit (Photo 29).

At higher temperatures (i.e. at 250 °C) the re-crystallization of $MgCO_3$ was also seen in a few cases, with fine grained aragonite overgrowth (Photos 31, 32).

The results of the experiments with the $MgCO_{3(s)}$—$CaCl_{2(aq)}$ systems can be summarized as follows:

a) The metasomatic alteration to calcite (re-calcitization) is highly improbable, for two reasons: 1. the Mg—O bonds in the $MgCO_3$ are mainly covalent, the Ca—O 1/3, 1/2 bonds are of weaker ionic nature. 2. Because of the larger ionic radii, structural fixation of Ca^{2+} would only be possible at larger temperatures by dilatation of the lattice structure. This is of very small magnitude in the temperature range of the present experiments, it does not reach the size of the Ca^{2+} radius. With increasing temperature the Mg-ions can be increasingly removed from the rhombohedral planes, but only from the restrictedly dislocated planes (in solutions of constant pH values). These Mg ions remain in the solution in form of chloride complexes. A smaller fraction of the magnesium is epigenetically fixated in form of unstable $MgCO_3 \cdot 2H_2O$.

In lithological sense the $MgCO_3$ might dissolved in the above conditions though its alteration to dolomite — and subsequently to calcite — is in general confined to closed geological systems of higher temperatures.

II/2. Powdery Dolomite — $CaCl_2$ — H_2O System
("de-dolomitization, re-calcitization")

II/2-a. Powdery Dolomite : Ca^{2+} = 2 : 1
(1.0 M solution)

a-1. The Chemical Analysis Data (Fig. 8)

The same powdery dolomite from the vicinity of Budapest has been used for standard as for the dolomite + $MgCl_{2(aq)}$ system. The chemical and XRD analyses data are summarised in Fig. 8.

Three temperature intervals have been used for the evaluation of chemical variations:

CaO: 1. 50—125 °C = +1.17% excess
2. 150—275 °C = +0.21% excess
3. 300 °C = —1.36% deficiency

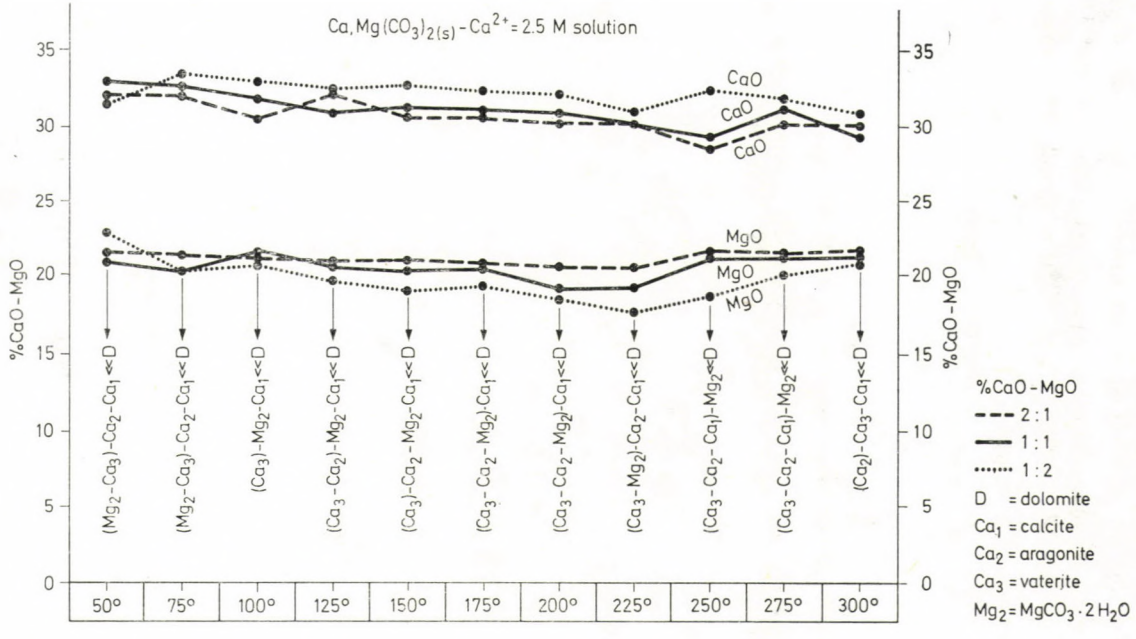


Fig. 8. The CaO—MgO variation in the dolomite + Ca^{2+} (1.0 M) systems (Mg : Ca = 2 : 1, 1 : 1, 1 : 2)

MgO: 1. 50—125 °C = —0.68% deficiency
2. 150—275 °C = —0.96% deficiency
3. 300 °C = —2.70% deficiency

The CaO content of the dolomite structure varies only slightly above 150 °C. The removal of magnesium is somewhat larger: under 1% between 50° and 125 °C, around 1% in the 150—275 °C range, 2.70% maximum value at 300 °C.

II/2-b. Powdery Dolomite : Ca^{2+} = 1 : 1
(1.0 M solution)

b-1. The Chemical Analysis Data (Fig. 8)

There is no significant change in the CaO distribution pattern in the 50—300 °C range. The MgO contents can be grouped into five temperature intervals:

CaO: 1. 50—275 °C = +1.62% excess (= —1.50% MgO)
2. 300 °C = —0.68% deficiency

MgO: 1. 50— 75 °C = —0.71% deficiency
2. 100—175 °C = —1.25% deficiency
3. 200—225 °C = —2.55% deficiency
4. 250—275 °C = —1.49% deficiency
5. 300 °C = —3.69% deficiency

According to these results the magnesium shows greater solubility than calcium, and dissolves from the mother-phase to a larger extent in this system (200—225 °C = —2.55%, 300 °C = —3.69% MgO).

II/2-c. Powdery Dolomite : Ca^{2+} = 1 : 2
(1.0 M solution)

c-1. The Chemical Analysis Data (Fig. 8)

Five and six temperature intervals can be chosen for the representation of the CaO and MgO values, respectively:

CaO: 1. 50—100 °C = +1.96% excess
2. 125—150 °C = +3.23% excess
3. 175 °C = +0.56% excess
4. 200—275 °C = +2.11% excess
5. 300 °C = —2.32% deficiency

MgO: 1. 50— 75 °C = —1.15% deficiency
2. 100—150 °C = —2.32% deficiency
3. 175 °C = —0.73% deficiency
4. 200—225 °C = —3.22% deficiency
5. 250—275 °C = —2.69% deficiency
6. 300 °C = —4.32% deficiency

At larger Ca^{2+} concentrations (i.e. Mg : Ca = 1 : 2) the crystallized phases of the systems are richer in CaO (with the exception of 300 °C), the MgO deficiencies are also higher.

It is well seen in Fig. 8 that at different cation-ratios and temperature intervals the CaO—MgO distributions of the crystallized phases roughly

follow the cation ratio of the solid phase (powdery dolomite) and the solutions; if the solution is richer in Ca^{2+}, then the resulted crystalline phases are also richer in calcium, and contain less MgO.

II/2-a. Powdery Dolomite : Ca^{2+} = 2 : 1 Crystalline Phases
(1.0 M solution)

$$
\begin{aligned}
50\ ^\circ C &= D \gg Ca_1-Ca_2-(Mg_2)?;\\
75\ ^\circ C &= D \gg Mg_2-(Ca_2-Ca_3);\\
100\ ^\circ C &= D \gg Ca_1-Ca_2-Ca_3-Mg_2;\\
125\ ^\circ C &= D > Ca_1-Ca_2;\\
150\ ^\circ C &= D > Ca_1-Ca_2-Mg_2;\\
175\ ^\circ C &= D \gg Ca_1-Mg_2-(Ca_2-Ca_3)?;\\
200\ ^\circ C &= D \gg Ca_1-(Ca_2-Ca_3-Mg_2);\\
225\ ^\circ C &= D \gg Ca_1-(Ca_2-Ca_3-Mg_2);\\
250\ ^\circ C &= D \gg Mg_2-(Ca_1);\\
275\ ^\circ C &= D \gg Mg_2-Ca_2;\\
300\ ^\circ C &= D \gg (Mg_2-Ca_2)-(Mg_1-Ca_1)?;
\end{aligned}
$$

D = dolomite; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterite; $Mg_2 = MgCO_3 \cdot 2\ H_2O$; $Mg_1 = MgCO_3$.

The chemical and XRD data indicate that in the applied experimental conditions the high Ca concentration in the solution only slightly affects the re-arrangement of the dolomite structure. There is only restricted cation removal (mostly magnesium) from the surface of the stable dolomite lattice, and this is not followed by any significant fixation of Ca^{2+} from the solution. It is probable that the dolomite may alter to calcareous dolomite ("proto-dolomite") which surrounds the crystals of the original phase (powdery dolomite). The Ca^{2+} and Mg^{2+} ions are fixated from the solution as epigenetic phases. The "de-dolomitization" and "re-calcitization" in this case means that the surface of dolomite rhombohedra are encrusted by calcite, aragonite, vaterite in form of thin films or small aggregated overgrowths.

The experiments have not supported the theory of solid phase metasomatic process by cation exchange. Unlike the $MgCO_3 + CaCl_2 - H_2O$ systems, the calcite is more dominant $CaCO_3$ variety here than the aragonite. Calcite — $CaCO_3$ was observed up to 250 °C, the presence of vaterite (hexagonal $CaCO_3$) was indicated only by 2—3 d/Å lines.

II/2-b. Powdery Dolomite : Ca^{2+} = 1 : 1 Crystal Phases
(1.0 M solution)

$$
\begin{aligned}
50\ ^\circ C &= D \gg Ca_1-Mg_2;\\
75\ ^\circ C &= D \gg Ca_1-Ca_2-Mg_2-(Ca_3?);\\
100\ ^\circ C &= D \gg Ca_1-Ca_2;\\
125\ ^\circ C &= D \gg Mg_2-Ca_1-Ca_2-(Ca_3?)\\
150\ ^\circ C &= D \gg Mg_2-Ca_1-Ca_2;\\
175\ ^\circ C &= D \gg Ca_1 + ?;\\
200\ ^\circ C &= D \gg Ca_1-(Ca_2-Mg_2)?;
\end{aligned}
$$

$$225\ °C = D \gg Ca_1-(Ca_3-Mg_2)?;$$
$$250\ °C = D \gg Mg_2-(Ca_2)?;$$
$$275\ °C = D \gg Mg_2-Ca_2-(Ca_3)?;$$
$$300\ °C = D \gg Mg_2-Ca_2-(Ca_3?);$$

D = dolomite; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterite; $Mg_2 = MgCO_3 \cdot 2\ H_2O$.

In comparison with the previous model the main difference is in the proportions of the epigenetic accessorial components. The importance of aragonite and vaterite decreases. In the 50—225 °C range the calcite is the more significant $CaCO_3$ variety, while aragonite becomes dominant from 250 °C upwards.

II/2-c. Powdery Dolomite : Ca^{2+} = 1 : 2 Crystalline Phases
(1.0 M solution)

$$50\ °C = D \gg Ca_1-Ca_2-Mg_2;$$
$$75\ °C = D \gg Ca_1-Mg_2-(Ca_3?);$$
$$100\ °C = D \gg Ca_1-Ca_2-Mg_2-(Ca_3?);$$
$$125\ °C = D \gg Ca_1-(Ca_2);$$
$$150\ °C = D \gg Mg_2-(Ca_1-Ca_2)?$$
$$175\ °C = D > Ca_1-Mg_2-(Ca_2-Ca_3)?;$$
$$200\ °C = D \gg Ca_1-Ca_2-(Ca_3?);$$
$$225\ °C = D > Ca_1-(Ca_2-Mg_2)?;$$
$$250\ °C = D \gg Mg_2-(Ca_2-Ca_3)?;$$
$$275\ °C = D \gg Mg_2-(Ca_1-Ca_3?);$$
$$300\ °C = D \gg (Ca_1-Ca_2).$$

D = dolomite; C_1 = calcite; C_2 = aragonite; Ca_3 = vaterite; $Mg_2 = MgCO_3 \cdot 2\ H_2O$.

These data, together with the chemical compositions, indicate that in given conditions the solubility of the powdery dolomite is very low. Complete structural alteration by metasomatic cation exchange does not take place. The presence of $CaCO_3$ varieties in natural dolomite horizons, layers (like Anisian or Norian dolomite sequences) can be attributed to two factors:

1. dehydration of $Ca(HCO_3)_2$ by ascendent-descendent solutions;
2. the effect of the Ca^{2+} ions, which have originated by the partial dissolution of dolomite in closed geological systems — or in open systems comprising relatively large CO_2 partial pressures. The excess calcium is mainly of external origin and does not come from the mother-phase (dolomite).

II/2-d. Powdery Dolomite : Ca^{2+} = 2 : 1
(2.5 M solution)

d-1. The Chemical Analysis Data (Fig. 9)

The average values can be grouped into six and four intervals with respect to the CaO and MgO values respectively:

CaO: 1. 50— 75 °C = +1.34% excess
2. 100 °C = +0.02% excess
3. 125 °C = +1.41% excess
4. 150—225 °C = −0.31% deficiency
5. 250 °C = −2.43% deficiency
6. 275—300 °C = −0.55% deficiency

MgO: 1. 50—100 °C = −0.64% deficiency
2. 125—225 °C = −1.19% deficiency
3. 250—275 °C = −0.58% deficiency
4. 300 °C = −0.26% deficiency

The CaO content of the powdery dolomite at 100 °C, and 150—225 °C shows only small variations, the major part of the Ca^{2+}—Mg^{2+}, which have mobilized from the dolomite, remains in the solution.

II/2-e. Powdery Dolomite : Ca^{2+} = 1 : 1
(2.5 M solution)

e-1. The Chemical Analysis Data (Fig. 9)

CaO: 1. 50—100 °C = +1.84% excess
2. 125—225 °C = +0.13% excess
3. 250 °C = −1.35% deficiency
4. 275 °C = +0.62% excess
5. 300 °C = −1.32% deficiency

MgO: 1. 50— 75 °C = −1.35% deficiency
2. 100 °C = −0.64% deficiency
3. 125—175 °C = −1.46% deficiency
4. 200—225 °C = −2.91% deficiency
5. 250—300 °C = −1.73% deficiency

A CaO anomaly is visible by the comparison of chemical compositions of the "powdery dolomite": Ca^{2+} = 2 : 1 and 1 : 1 systems. The 150—300 °C temperature interval in the Mg^{2+} : Ca^{2+} = 2 : 1 system is characterized by 0.31—2.43% CaO deficiency in the precipitated crystal phases despite of the presence of excess Ca^{2+} in the solution. The MgO loss, however, is somewhat smaller. Calcium deficiency has occurred at 250 °C and 300 °C in the Mg^{2+} : Ca^{2+} = 1 : 1 system (with the increasing Ca concentration in the solution). The CaO increase in the other temperature intervals (except for 50—100 °C and 225 °C) was lower than the CaO values of the "powdery dolomite": Ca^{2+} = 1 : 2 system, where a continuous calcium increase has been observed.

II/2-f. Powdery Dolomite : Ca^{2+} = 1 : 2
(2.5 M solution)

f-1. The Chemical Analysis Data (Fig. 9)

CaO: 1. 50 °C = +1.06% excess
2. 75—100 °C = +2.55% excess
3. 125—200 °C = +1.63% excess

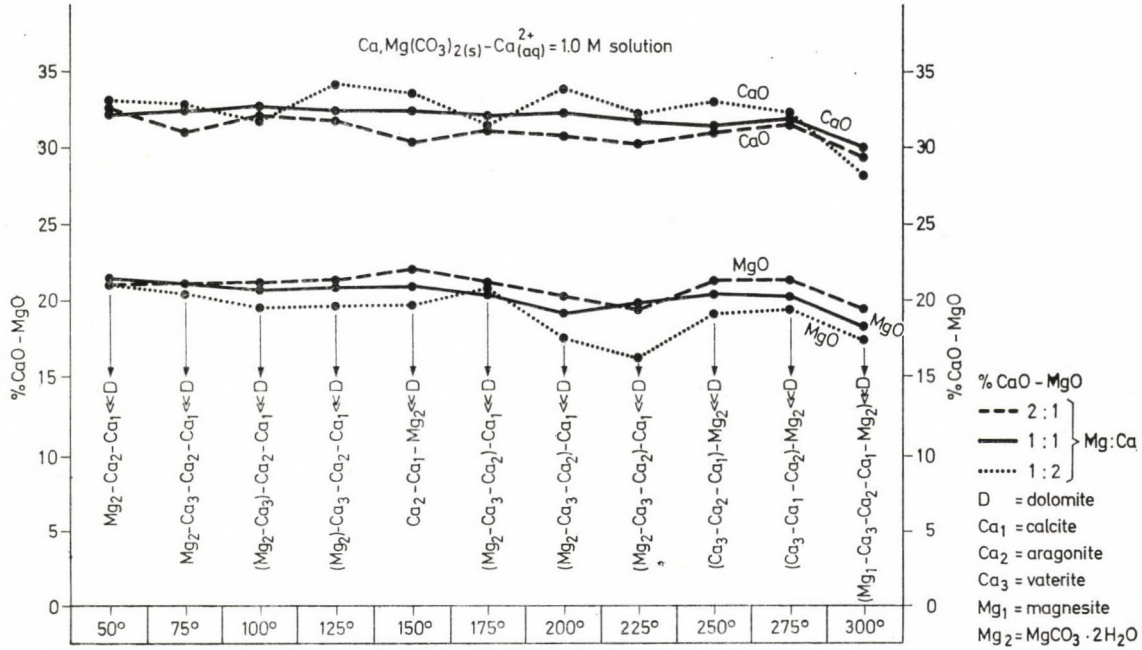


Fig. 9. Dolomite + Ca^{2+} = 2.5 M solution. The variation of CaO—MgO contents in the Ca : Mg = 2 : 1, 1 : 1, and 1 : 2 systems

4. 225 °C = +0.13% excess
5. 250 °C = +2.05% excess
6. 275—300 °C = +0.52% excess

MgO: 1. 50—100 °C = —1.24% deficiency
2. 125—200 °C = —1.94% deficiency
3. 225 °C = —4.40% deficiency
4. 250 °C = —3.35% deficiency
5. 275—300 °C = —1.65% deficiency

The significant differences between the two components CaO and MgO were observable from 125 °C, the maximum MgO deficiency was measured at 225 °C and 250 °C.

The variations of the MgO—CaO contents are plotted in Fig. 9.

II/2-d. Powdery Dolomite : Ca^{2+} = 2 : 1 Crystalline Phases
(2.5 M solution)

50 °C = D ≫ $Ca_1-Ca_2-(Ca_3-Mg_3)$?;
75 °C = D ≫ $Mg_2-Ca_1-Ca_2-(Ca_3)$?;
100 °C = D ≫ $Ca_1-Mg_2-(Ca_3)$;
125 °C = D ≫ $Ca_1-Mg_2-(Ca_2-Ca_3)$?;
150 °C = D ≫ $Ca_1-Ca_2-Mg_2$;
175 °C = D ≫ $Ca_1-Mg_2-(Ca_2-Ca_3)$?;
200 °C = D ≫ $(Ca_1-Ca_3-Mg_2)$?;
225 °C = D ≫ $Ca_2-Ca_1-(Ca_3-Mg_2)$?;
250 °C = D ≫ Mg_2;
275 °C = D ≫ $Mg_2-(Ca_1-Ca_2-Mg_1)$;
300 °C = D ≫ $Ca_1-Ca_3-(Ca_2)$?.

D = dolomite; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterite; $Mg_2 = MgCO_3 \cdot 2\,H_2O$; $Mg_1 = MgCO_3$.

Typical to the system is the presence of calcite at every temperature level (except for 250 °C). The presence of aragonite and vaterite is in most cases uncertain (50°, 100°, 125°, 175°, 225°, 300 °C). Basic Mg-carbonate has been observed at 100°, 125°, 175° and 250°, with only weak intensities at 275 °C. At this latter temperature $MgCO_3$ was also present.

II/2-e. Powdery Dolomite : Ca^{2+} = 1 : 1 Crystal Phases
(2.5 M solution)

50 °C = D ≫ $Ca_1-Ca_2-(Mg_2)$?;
75 °C = D ≫ $Ca_1-Ca_2-(Ca_3)$?;
100 °C = D ≫ $Ca_1-Mg_2-(Ca_3)$?;
125 °C = D ≫ $Ca_1-Ca_2-Mg_2$;
150 °C = D ≫ $Ca_1-Mg_2-(Ca_2-Ca_3)$?;
175 °C = D ≫ $Ca_2-Ca_1-(Ca_3)$?;
200 °C = D ≫ $Ca_1-Ca_2-(Ca_3-Mg_2)$?;
225 °C = D ≫ $Ca_2-Ca_1-(Ca_3)$?;
250 °C = D ≫ $Mg_2-(Ca_3)$?;
275 °C = D ≫ Mg_2;
300 °C = D ≫ $Ca_2-Ca_3-(Ca_1)$?.

D= dolomite; Ca_1 = calcite; Ca_2= aragonite; Ca_3 = vaterite; $Mg_2 = MgCO_3 \cdot 2\,H_2O$.

Calcite (Ca_1) has developed at every temperature level up to 225 °C, but there was no independent $CaCO_3$ phase at 250—275 °C (except for weak traces of vaterite). At 300 °C aragonite and vaterite has formed, with uncertain traces of calcite. Basic Mg-carbonate has crystallized at 250° and 275 °C in significant quantities. It was subordinate, however, in the other temperature intervals.

II/2-f. Powdery Dolomite : Ca^{2+} = 1 : 2 Crystal Phases
(2.5 M solution)

50 °C = D $\gg$ $Ca_1-Ca_2-(Ca_3-Mg_2)$?;
75 °C = D $\gg$ $Ca_1-Ca_2-(Ca_3-Mg_2)$;
100 °C = D $\gg$ Ca_1-Ca_3;
125 °C = D $\gg$ $Ca_1-(Ca_2-Mg_2)$?;
150 °C = D $\gg$ $Ca_1-Mg_2-(Ca_2-Ca_3)$?;
175 °C = D $\gg$ $Ca_1-(Ca_2-Ca_3)$?;
200 °C = D $\gg$ $Ca_1-(Ca_3-Mg_2)$?;
225 °C = D $\gg$ $Ca_1-Ca_2-(Ca_3-Mg_2)$?;
250 °C = D $\gg$ $Mg_2-(Ca_2-Ca_3)$?;
275 °C = D $\gg$ $Mg_2-(Ca_1-Ca_2)$;
300 °C = D $\gg$ $Ca_2-Ca_3-(Ca_1)$?.

The distribution of the crystalline phases is essentially similar to the previous model (Mg : Ca = 1 : 1). The calcite is the dominant accessorial component, the amount of Mg_2 was significant at 250—275 °C.

The chemical, X-ray and microscopic analysis of the powdery dolomite + + $CaCl_2$ + H_2O system has not supported the "de-dolomitization—re-calcitization" theory. The results suggest, that in natural conditions with parameters similar to that of the model experiments this process does not take place.

According to the microscopic analysis, there was no significant Ca^{2+} fixation subsequent to the slight dissolution and the resorption of the dolomite. It is more probable that the "normal" dolomite alters to so-called "proto-dolomite" (= calcareous dolomite), by the dissolution of that amount of Mg^{2+} which exceeds the amount of Ca^{2+}. This proto-dolomite precipitates on the surface edges of the cation-ordered dolomite rhombohedra (R) in form of thin films. The major part of Ca^{2+} is fixated in the epigenetic carbonates (Ca_1—Ca_2). Among these the Ca_2 has had well-developed crystals (Photos: 26, 29).

II/3. "Calcareous Dolomite" + $CaCl_2$ — H_2O System

"Calcareous dolomite" (Pilisvörösvár) has been used as starting material. Its chemical composition (analysed by Mrs. V. Tolnay):

SiO_2	0.19%
Al_2O_3	0.13%
Fe_2O_3	0.04%
FeO	0.06%
CaO	32.03%
MgO	20.90%
K_2O	0.04%
Na_2O	0.21%
H_2O	1.42%
CO_2	45.51%
	100.55%

SiO_2	0.19%
Al_2O_3	0.13%
Fe_2O_3	0.04%
K_2O	0.04%
Na_2O	0.21%
H_2O	1.42%
	2.03%

100.55 − 2.03 = 98.52%

After subtraction:

CaO	32.53%	= 0.580 (atomic quotient)
MgO	21.21%	= 0.526 (atomic quotient)
FeO	0.06%	= 0.001 (atomic quotient)
CO_2	46.19%	=
	99.99%	

Its stoichiometric composition: $Ca_{0.580}Mg_{0.526}Fe_{0.001}CO_3$.

The chemical data indicate Mg deficiency in the calcareous dolomite. The experiments have been carried out only with the dolomite : 3 ml 2.5 M $CaCl_{2(aq)}$ system, which is approximately equal to Ca : Mg = 2 : 1.

II/3-a. The Evaluation of Chemical Analyses of the "Calcareous Dolomite" + Ca^{2+} system (Fig. 10)

CaO:	50 °C	= 33.23%	= +0.70% excess
	75 °C	= 32.25%	= −0.28% deficiency
	100 °C	= 33.86%	= +1.33% excess
	125 °C	= 33.67%	= +1.14% excess
	150 °C	= 33.80%	= +1.27% excess
	175 °C	= 33.98%	= +1.45% excess
	200 °C	= 33.52%	= −0.99% deficiency
	225 °C	= 33.19%	= +0.66% excess
	250 °C	= 34.09%	= +1.56% excess
	275 °C	= 26.09%	= −6.44% deficiency
	300 °C	= 30.04%	= −2.49% deficiency

There was no significant change at 100—175 °C interval, varying CaO values were observed at 200—225 °C. Marked CaO deficiency was measured at 275 °C and 300 °C.

MgO:	50 °C	= 20.30%	= −0.91% deficiency
	75 °C	= 21.15%	= −0.06% deficiency
	100 °C	= 20.15%	= −1.06% deficiency
	125 °C	= 19.99%	= −1.22% deficiency
	150 °C	= 19.52%	= −1.69% deficiency
	175 °C	= 19.68%	= −1.53% deficiency
	200 °C	= 19.39%	= −1.82% deficiency
	225 °C	= 19.42%	= −1.79% deficiency
	250 °C	= 18.75%	= −2.46% deficiency
	275 °C	= 25.21%	= +4.00% excess (?)
	300 °C	= 18.31%	= −2.90% deficiency

The CaO—MgO distributions are plotted in Fig. 10.

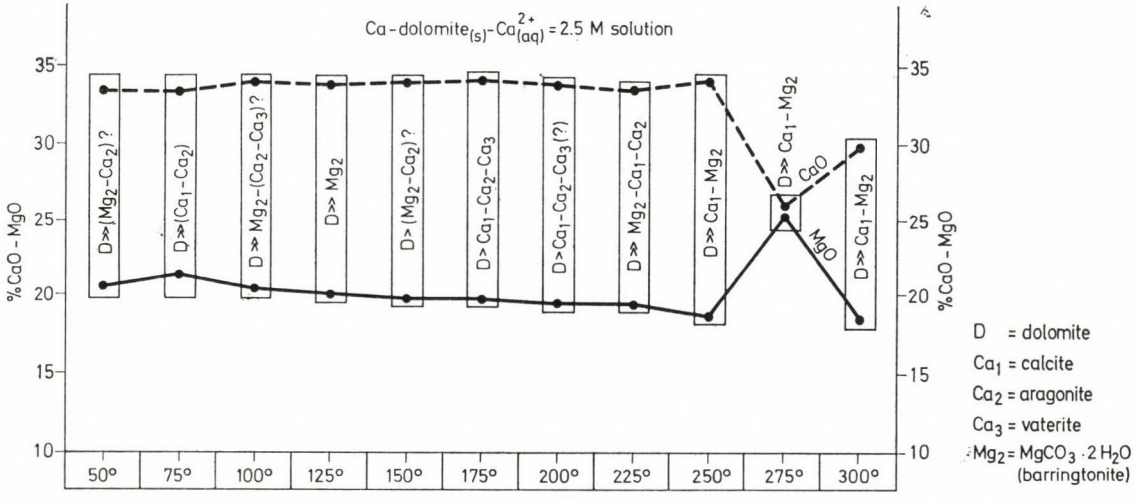


Fig. 10. "Ca-dolomite" + $Ca^{2+}_{(aq)}$ (2.5 M solution) system, CaO—MgO distributions and crystal phases

The chemical data have proven, that even a Ca^{2+} rich and Mg^{2+} depleted dolomite does not tend to "de-dolomitise" or "re-calcitise" by fixation of Ca^{2+}. Significant, large alterations have undergone between 275—300 °C, but even in this case the calcareous dolomite ("proto-dolomite") approaches the "normal dolomite" composition by Ca^{2+} loss. The 2.46—2.90% Mg^{2+} loss at 250° and 300 °C remained in dissolved form, except for a fraction which has fixated as $MgCO_3 \cdot 2H_2O$. The +4.00% MgO at 275°C should be considered cautiously since there were only uncertain traces of $MgCO_3 \cdot 2H_2O$ among the crystalline phases.

II/3-b. "Calcareous Dolomite" : Ca^{2+} System Crystal Phases

$$
\begin{aligned}
50\ ^\circ C &= D \gg Mg_2-(Ca_2)?;\\
75\ ^\circ C &= D \gg (Ca_1-Ca_2);\\
100\ ^\circ C &= D \gg Mg_2-(Ca_2-Ca_3)?;\\
125\ ^\circ C &= D > Mg_2;\\
150\ ^\circ C &= D > Mg_2-Ca_2)?;\\
175\ ^\circ C &= D > Ca_1-Ca_3-Ca_2;\\
200\ ^\circ C &= D > Ca_1-Mg_2-Ca_3)?;\\
225\ ^\circ C &= D \gg Mg_2-Ca_1-Ca_2;\\
250\ ^\circ C &= D \gg Ca_1-Mg_2;\\
275\ ^\circ C &= D \gg Ca_1-Mg_2;\\
300\ ^\circ C &= D \gg Ca_1-Mg_2.
\end{aligned}
$$

D = dolomite; Ca_1 = calcite; Ca_2 = aragonite; Ca_3 = vaterites; Mg_2 = $MgCO_3 \cdot 2H_2O$.

The calcareous dolomite — similarly to the powdery dolomite — does not alter to calcite by metasomatic processes at larger temperatures, though it is possible that the Ca^{2+} which has been mobilized by partial or complete dissolving is fixated in situ in form of carbonated. These epigenetic phases are mainly composed from calcite, the presence of aragonite and vaterite is either uncertain or subordinate. Similar role has been played by the $MgCO_3 \cdot 2H_2O$. Based on these facts "re-calcitized" dolomite does exist, but it is not a product of cation exchange metasomatism, rather asynchronous precipitation from a multi-phase system.

III. Morphogenetic Evaluation of the Crystalline Phases

Due to the extremely small size of the particles (10—30 microns) optical microscopy could only rarely used. SEM methods proved to be suitable for the observation of morphological and structural features, providing sufficiently high magnification. A typical case of such morphological features is shown on 1—2 photos, about the $CaCO_{3(s)}$—Fe^{2+} system. A similar morphogenetic mechanism is shown on the photos of the $CaCO_{3(s)}$—$Mg^{2+}{}_{(aq)}$ system — 300 °C (15,000× magnification).

The cation exchange of the solid phase—solution system (metasomatism) is a multiple process. In theory the total amount of the R^{2+} ions of the solid phase can be replaced without the dissolution of the "mother-structure":

$$Ca^{2+} \rightarrow Fe^{2+}$$
$$Ca^{2+} \rightarrow Mn^{2+}$$
$$Ca^{2+} \rightarrow Mg^{2+}$$

In cases the possibility of such an exchange is restricted owing to the differences in ionic radii, bonding forces and stability of the mother structure. For example, the Ca^{2+} ions of the calcite can wholly be replaced by Mg^{2+}, Fe^{2+} or Mn^{2+} ($Ca^{2+} = 0.99$ Å, $Mg^{2+} = 0.66$ Å or $Ca^{2+} = 0.99$ Å, $Fe^{2+} = 0.74$ Å), but in reversed direction the cation exchange is restricted:

$$MgCO_{3(s)} \rightarrow Ca^{2+}$$
$$CaMg(CO_3)_2 \rightarrow Ca^{2+}{}_{(aq)} \text{ (in place of } Mg^{2+})$$

In certain bands (rhombohedral planes) the intensity of the cation exchange is larger. At the same time other planes of the same belt may remain intact. The former ones are in fact dislocated, electro-statically unbalanced lattice-planes, which are arranged along the rhombohedral planes of the mother-phase. In these cases the resorption and cation exchange phenomena are concentrated to these bands, or, in case of incoherent lattice defects, to random points on the crystal plane (rhombohedron). This is resulted in higher rate of replacement and migration of that cation, which is more suitable for replacement. Due to the slower rate of fixation of the replacing cation, voids are starting to be formed in the crystal (rhombohedron). The skeleton crystal (the initial calcite mother phase) is surrounded by the rhombohedral frame of the new phase ($FeCO_3$) of a few microns thickness. The voids are then filled stepwise by the rhombohedra of the new phase ($MgCO_3$, $FeCO_3$) in analogous manner to that of the decrescency theory.

On the first picture a void-rich rhombohedron skeleton of $FeCO_3$ is shown (19,500×), which has been formed in closed system in three years from $CaCO_{3(s)}$—$Fe^{2+}{}_{(aq)} = 1 : 2$ system at room temperature. In the second picture an almost completely infilled $FeCO_3$ rhombohedron is seen, formed from the same system after longer treatment (5 years — closed system). On its surface α—FeOOH+vaterite encrustation has formed.

The picture 3 shows $CaCO_3$ (calcite) rhombohedra, which formed from $CaCO_{3(s)}+Mg^{2+}{}_{(aq)} = 70 : 30$ system at 50 °C. The crystallites on the surface of the rhombohedral planes are of triclinic "barringtonite" ($MgCO_3 \cdot 2H_2O$).

At the same temperature (50 °C), with increased Mg^{2+} content of the solution ($CaCO_{3(s)}$—$Mg^{2+}{}_{(aq)} = 50 : 50$) the formation of $MgCO_3 \cdot 2H_2O$ is also accelerated. Resorbed, dislocated patches fields are shown on the surface of a calcite crystal (Photo: 4. 9000×).

MAGYAR TUDOMÁNYOS AKADÉMIA KÖNYVTÁRA

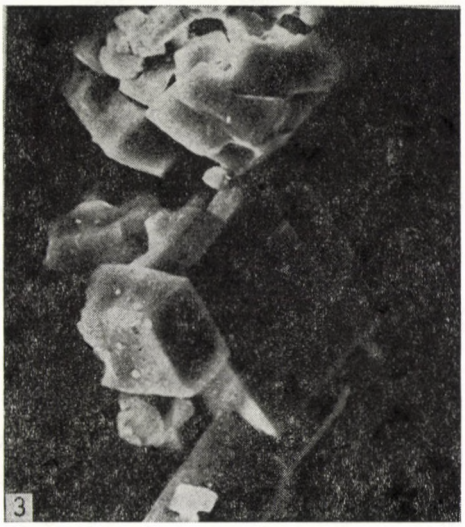


Plate I

1. $CaCO_{3(s)}-Fe^{2+}{}_{(aq)} = 1 : 2$; room temperature
   $FeCO_3$ — 19,500× — 3 years
2. $CaCO_{3(s)}-Fe^{2+}{}_{(aq)} = 1 : 2$; room temperature
   $FeCO_3$ — 15,000× — 5 years
3. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 70 : 30$; 50 °C
   Calcite rhombohedra; 6000×
4. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 50 : 50$; 50 °C
   Calcite + $MgCO_3 \cdot 2\,H_2O$; 9000×

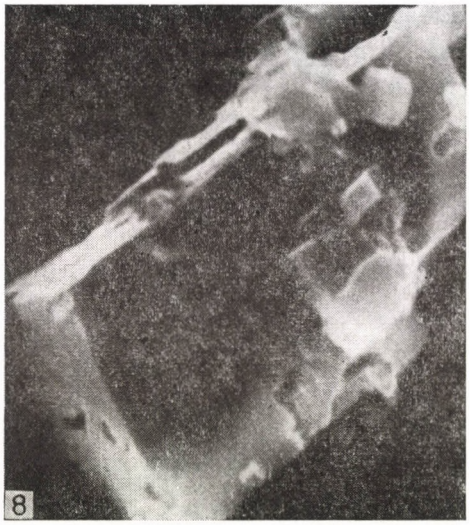
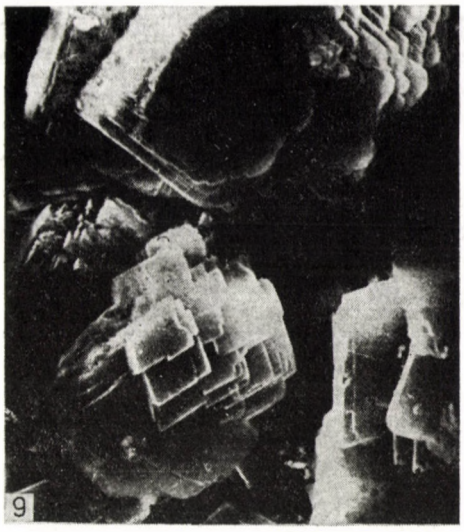
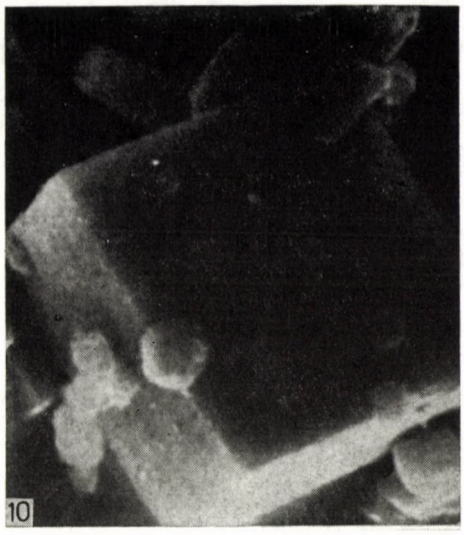


Plate II

5. $CaCO_{3(s)}-Mg^{2+}_{(aq)} = 70 : 30$; 125 °C
   Vuggy calcite rhombohedron, gradually infilling with $MgCO_3$; 3000 ×
6. $CaCO_{3(s)}-Mg^{2+}_{(aq)} = 30 : 70$; 150 °C
   $MgCO_3-CaMg(CO_3)_2$ rhombohedra, 7100 ×
7. $CaCO_{3(s)}-Mg^{2+}_{(aq)} = 70 : 30$; 175 °C
   Skeleton crystals, vuggy rhombohedra of dolomite, $MgCO_3 \cdot 2\ H_2O$ lamellae; 9000 ×
8. $CaCO_{3(ss)}-Mg^{2+}_{(aq)} = 70 : 30$; 200 °C
   Dolomite rhombohedra, 27,000 ×





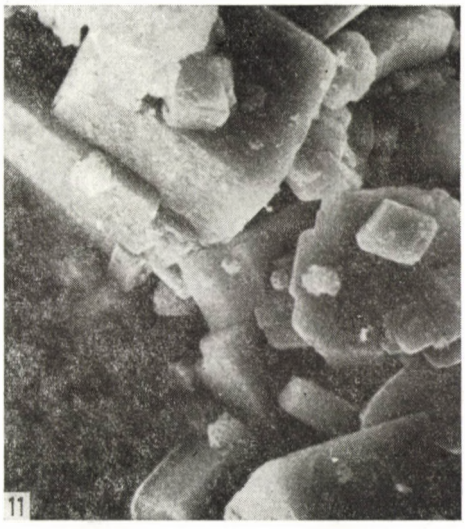


Plate III

9. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 30 : 70$; 200 °C
Crystal aggregate composed of $MgCO_3$ rhombohedra of different sizes; 7200 ×
10. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 30 : 70$; 200 °C
$MgCO_3$ rhombohedron, with $MgCO_3 \cdot 2\,H_2O$ crystals on its surface; 8600 ×
11. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 30 : 70$; 225 °C
$MgCO_3$ rhombohedra; 8200 ×
12. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 50 : 40$; 275 °C
Vuggy $MgCO_3$ rhombohedra, with traces of resorption; 8400 ×

A larger unfilled void is seen in the calcite rhombohedron of the $CaCO_{3(s)}$-$Mg^{2+}{}_{(aq)}$ system, treated at 125 °C. The definite bands and lamellae on the lower edge of the crystal are composed of $MgCO_3$ (Photo: 5; 3000×).

The smaller calcite crystals ($\mu\emptyset$) which have formed at 150 °C from $Ca^{2+} : Mg^{2+} = 30 : 70$ system, are altered to $CaMg(CO_3)_2$ and $MgCO_3$. The larger rhombohedra (40—50 microns) contains voids, and encrusted partially by a thin film of $MgCO_3$—$CaMg(CO_3)_2$ (Photo: 6; 7100×).

A dolomite (= "proto-dolomite") crystal, which formed at 175 °C from $Ca^{2+} : Mg = 70 : 30$ system, is seen in Photo: 7. The crystals, with poorly developed contours are of epigenetic $MgCO_3 \cdot 2H_2O$ (barringtonite) though $MgCO_3$ is also present.

$CaMg(CO_3)_2$ — dolomite — rhombohedra have formed with dislocated, voidy bands in the $CaCO_{3(s)}$—$Mg^{2+}{}_{(aq)} = 70 : 30$ system at 200 °C (Photo: 8; 27,000×).

The increasing Mg^{2+} concentration and temperature of the solution promotes the formation of $MgCO_3$, and hinders that of the $CaMg(CO_3)_2 \cdot$ $\cdot MgCO_3$ with 0.95% CaO impurity has formed in the $CaCO_{3(s)}$—$Mg^{2+} =$ $= 30 : 70$ system at 200 °C (Photo: 9; 7200×).

An $MgCO_3$ rhombohedral twin intergrowth with epigenetic $MgCO_3 \cdot 2H_2O$ encrustation is seen in Photo: 10; 8600×.

$MgCO_3$ rhombohedron with 1.88% CaO impurity has formed in the $CaCO_{3(s)}$—$Mg^{2+}{}_{(aq)} = 30 : 70$ system, with $MgCO_3 \cdot 2H_2O$ crystals on its surface (Photo: 11; 8200×).

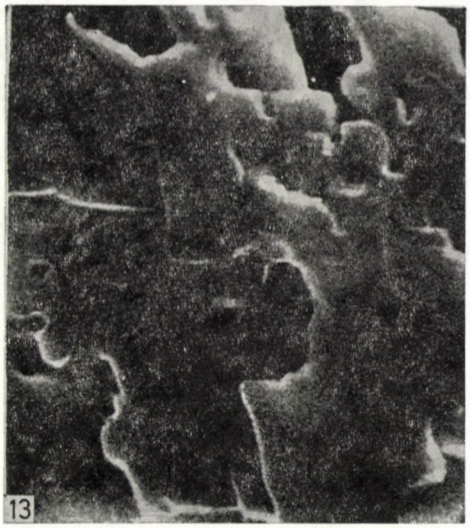
Vuggy rhombohedral skeleton crystals of $MgCO_3$ have been formed in the $Ca^{2+} : Mg^{2+} = 50 : 50$ system at 275 °C (Photo: 12).

A void in a rhombohedron is partially filled by $MgCO_3$; $Ca^{2+} : Mg^{2+} =$ $= 50 : 50$ system, 275 °C (Photo: 13; 19,000×). Euhedral, well-preserved $MgCO_3$ hombohedra of the same system are seen in Photo 14).

A rhombohedral shaped void, consisting of $MgCO_3$ skeletal rhombohedra and exhibiting small scale resorption is seen in Photo: 15; 15,000×).

The rhombohedral dislocation planes of the dolomite became vuggy and lamellar due to partial dissolving (powdery dolomite + $Mg^{2+}{}_{(aq)} = 70 : 30$ system, 50 °C). $MgCO_3 \cdot 2H_2O$ aggregates are found on the surface of the lamellae. At increased temperatures the $(h0\bar{h}1)$ planes are loosening, the dolomite becomes more and more lamellar by partial dissolving of the dislocated lattice planes. The formation of $MgCO_3 \cdot 2H_2O + MgCO_3$ is also increasing, though these are still epigenetic precipitations, independent from the dolomite (Photo: 16; 7200×).

Only slight changes are seen in temperature and Mg^{2+} ion concentration of the solution is increased. Dolomite + $Mg^{2+} = 50 : 50$ system, 100 °C. (Photo: 17; 7200×) (See also Photos: 16 and 18.)



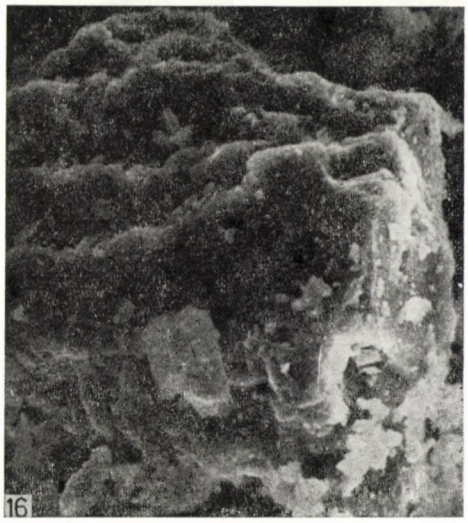
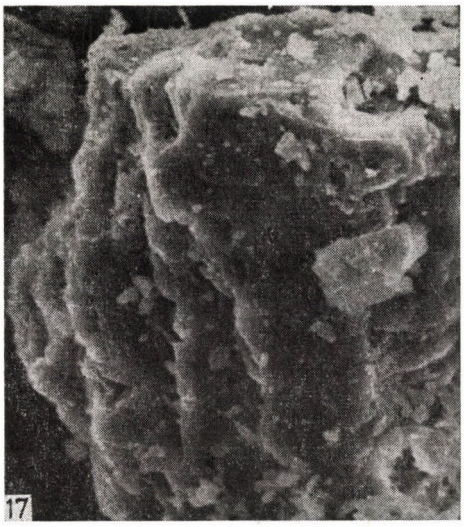
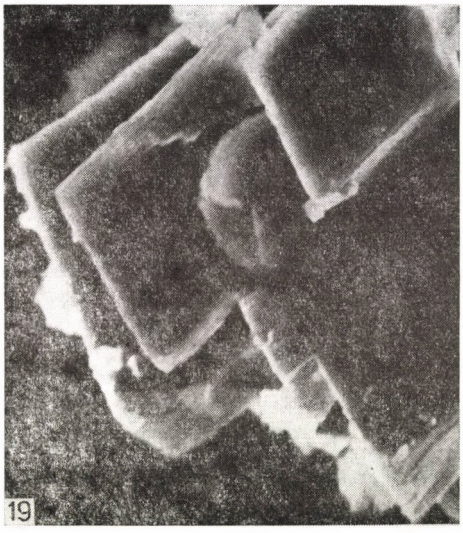


Plate IV

13. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 50 : 50$; 275 °C
Vuggy $MgCO_3$ rhombohedron with step-like infilling; 19,000×
14. $CaCO_{3(s)}-Mg^{2+}{}_{(aq)} = 50 : 50$; 275 °C
Intact $MgCO_3$ rhombohedron; 7500×
15. $CaCO_{(s)}-Mg^{2+}{}_{(aq)} = 30 : 70$; 275 °C
A void in a $MgCO_3$ rhombohedron, with well visible gradual infilling. Resorption marks are seen on the $MgCO_3$ polyhedra; 15,000×
16. Powdery dolomite $-Mg^{2+}{}_{(aq)} = 70 : 30$; 50 °C
The alteration starts on the dislocated planes of the dolomite rhombohedron, the crystal becomes lamellar; 7200×





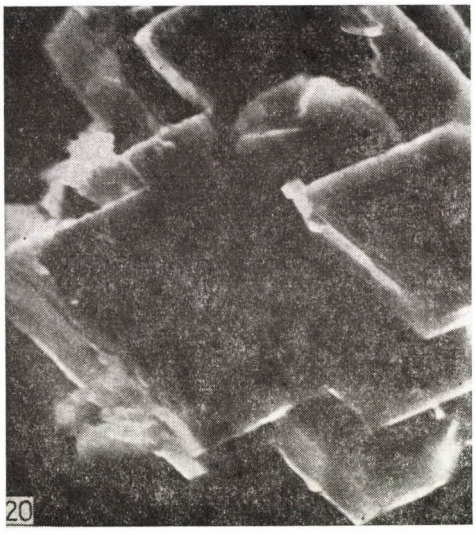
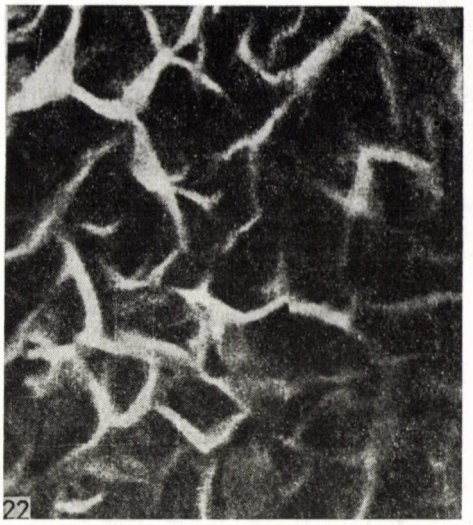
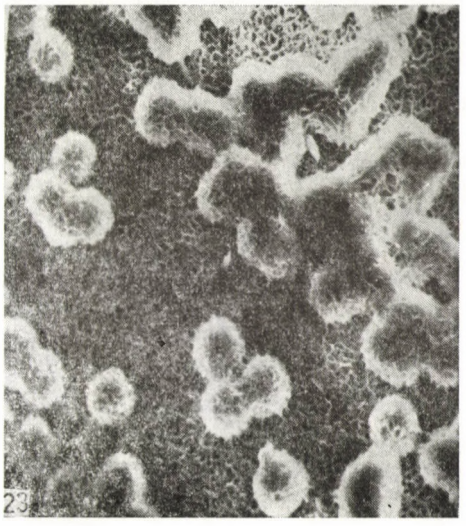


Plate V

17. Powdery dolomite $-Mg^{2+}_{(aq)} = 50 : 50$; 100 °C
Similar to picture 16; 7200 ×
18. Powdery dolomite $-Mg^{2+}_{(aq)} = 50 : 50$; 125 °C
Similar to pictures 16—17; 7200 ×
19. Powdery dolomite $-Mg^{2+}_{(aq)} = 30 : 70$; 175 °C
Dolomite + $MgCO_3$ rhombohedra; 9800 ×
20. Powdery dolomite $-Mg^{2+}_{(aq)} = 30 : 70$; 175 °C
Dolomite + $MgCO_3$ rhombohedra; 16,100 ×





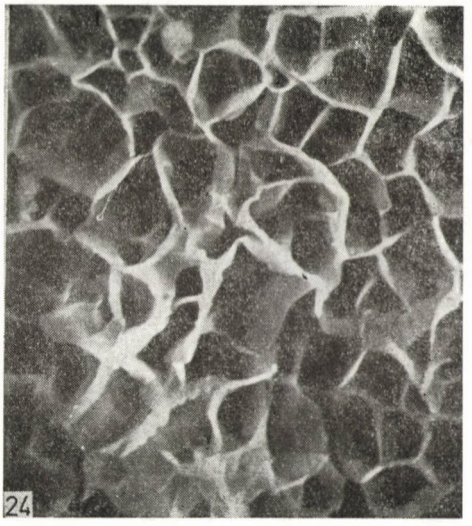
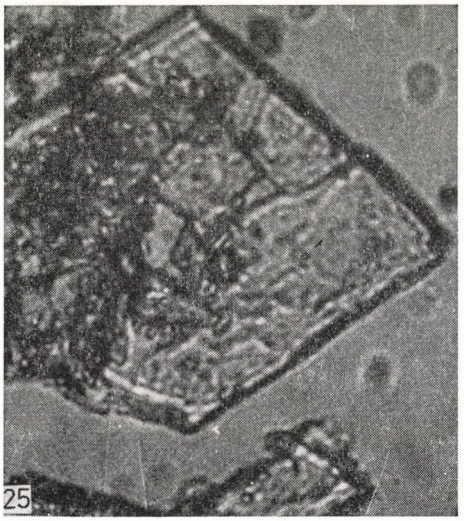
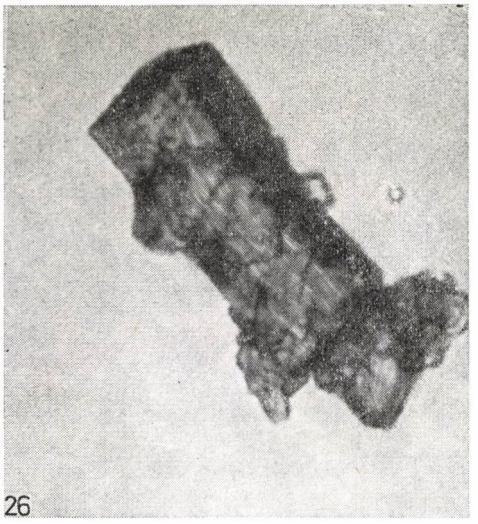
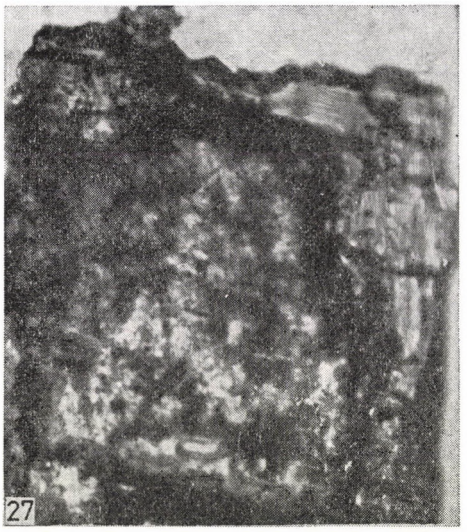


Plate VI

21. Powdery dolomite $-Mg^{2+}_{(aq)} = 70 : 30$; 225 °C
MgCO$_3$ rhombohedra, with resorbed surfaces; 9000 ×
22. Powdery dolomite $-Mg^{2+}_{(aq)} = 70 : 30$; 225 °C
Cellular crystallites of $MgCO_3 \cdot 2\,H_2O$ on the surface of a MgCO$_3$ rhombohedron; 30,000 ×
23. $MgCO_{3(s)} - Ca^{2+}_{(aq)} = 1 : 2$; 250 °C
Calcite-aragonite aggregates on the surface of magnesite (1.0 M Ca^{2+} solution); 4800 ×
24. $MgCO_{3(s)} - Ca_{(ay)} = 1 : 2$; 250 °C
Cellular lamellae of $MgCO_3 \cdot 2\,H_2O$. The system contains calcite and aragonite too; 30,000 ×







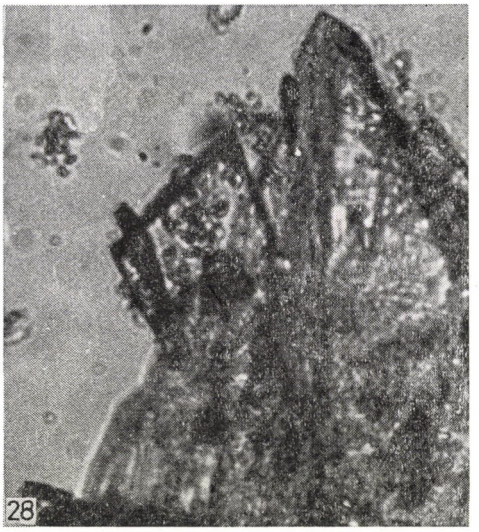


Plate VII

25. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 2 : 1$; 50 ° C
(2.5 M solution)
Aragonite and vaterite aggregates on the surface of a magnesite cleavage rhombohedra 150×

26. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 1 : 2$; 50 °C
(2.5 M solution)
Aragonite; 573×

27. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 1 : 2$; 125 °C
Epigenetic dolomite perigrowth on the edge of a magnesite rhombohedron (2.5M solution) 157×

28. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 2 : 1$; 175 °C
Epigenetic overgrowth on the surface of magnesite with calcite and aragonite precipitations. (2.5 M solution). 157×

The temperature level of the dolomite stability is 175 °C. The crystals of the powdery dolomite are rearranging to well-formed $MgCO_3$ + dolomite rhombohedra by replacement of Ca^{2+} by Mg^{2+}. Dolomite + $Mg^{2+}{}_{(aq)}$ = = 30 : 70 system, 175 °C (Photo: 19; 9800×). The same rhombohedra are seen with larger magnification in Photo 20; 16,100×.

$MgCO_3$ crystals with 0.75% CaO content have formed from the dolomite + $Mg^{2+}{}_{(aq)}$ = 70 : 30 system at 225 °C, with marks of resorption on their surface (Photo: 21; 9000×). Epigenetic $MgCO_3 \cdot 2H_2O$ crystallites are seen on the surface of the previously mentioned rhombohedra (Photo: 22; 30,000×).

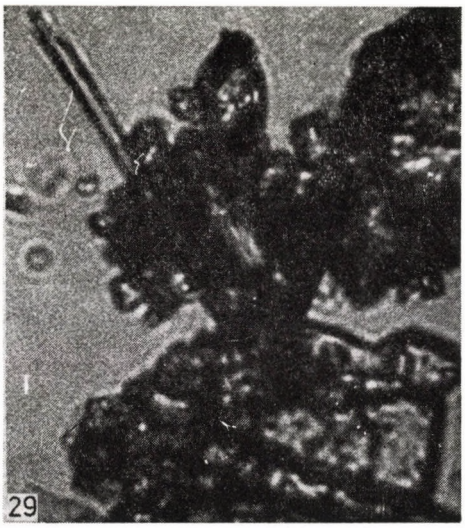
The "re-calcitization" experimental systems consisted of independent separate phases. Mostly dissolution was observed, recrystallization by cation exchange has not been found. Calcite aggregates and aragonite lamellae, which have been formed on the surface of a $MgCO_3$ (magnesite) rhombohedra ($MgCO_{3(s)}$ + $Ca^{2+}{}_{(aq)}$ = 1 : 2) are seen on Photo: 23; 4800×. Material from $MgCO_{3(s)}$ + $Ca^{2+}{}_{(aq)}$ = 1 : 2 system at 250 °C is seen in the next picture. The thin cellular forms are composed of $MgCO_3 \cdot 2H_2O$ + calcite + aragonite (Photo: 24; 30,000×).

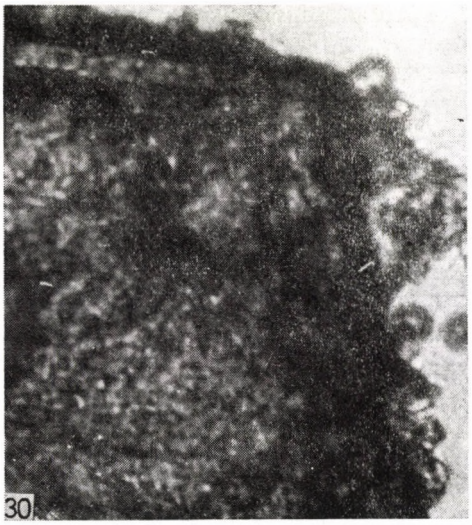
For the investigation of morphological variations in the "de-dolomitization–re-calcitization" systems normal optical microscopy has been used. Though smaller magnifications were obtained, the morphogenetic features were still well seen.

Intact magnesite cleavage rhombohedron from $MgCO_{3(s)}$ — $Ca^{2+}{}_{(aq)}$ = = 2 : 1 system at 50 °C. The aggregates on its surface are composed of a mixture of aragonite and vaterite (Photo: 25; 150×). At the same temperature, in the Mg^{2+} : Ca^{2+} = 1 : 2 system $CaCO_3$ (aragonite) crystals were formed, completely isolated from the magnesite "mother-phase" which is seemed to be intact. The $[CO_3]^{2-}$ which has been mobilized from the magnesite planes, promoted the formation of aragonite. Since the treated mother phase has contained 0.80% Cl after several washings, it is possible, that chlorine can be fixated in the lattice, though this hypothesis had been questioned by a few authors. The aragonite is seen in Photo: 26; (573×).

A magnesite rhombohedron ($MgCO_{3(s)}$ + $Ca^{2+}{}_{(aq)}$ = 1 : 2, 125 °C) is seen in the next picture, with clean dolomite overgrowth of somewhat higher birefringence (proto-dolomite) (Photo: 27; 157×).

Solid phase rearrangement similar to that of between Mg^{2+} solutions and calcite, dolomite structures, was not observed even at larger temperatures. Possibly the size if Ca^{2+} ions are too large to replace the partially mobilized Mg^{2+}. It is rather crystallizes as separate epigenetic phase, independent from the mother-phase. Magnesite rhombohedron with perimorphous overgrowth, separate epigenetic crystalline phases (aragonite + calcite) from the $MgCO_{3(s)}$+ + $Ca^{2+}{}_{(aq)}$ = 2 : 1 system at 175 °C is seen on Photo 28. At the same temperature the Mg : Ca = 1 : 1 system shows similar features, the CaO content





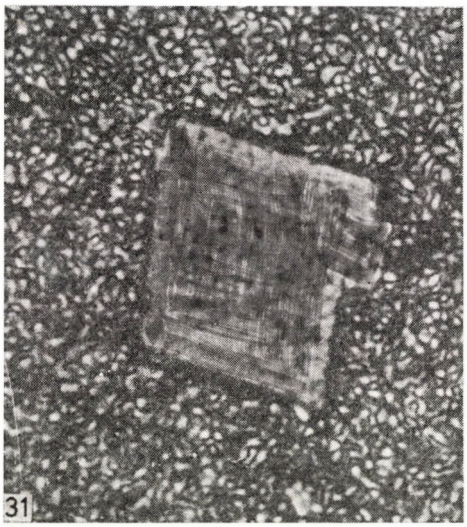


Plate VIII

29. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 1:1$; 175 °C
Fibrous aragonite crystals on the surface of a magnesite crystal (2.5 M solution). 246×

30. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 1:1$; 200 °C
Dolomite perigrowth on a magnesite rhombohedron, with overgrown aragonite (2.5 M solution); 157×

31. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 1:1$; 275 °C
Epigenetic crystal aggregate of $MgCO_3$ + dolomite + aragonite, with magnesite (?) — dolomite (?) rhombohedron inclusion (2.5 M solution); 275×

32. $MgCO_{3(s)}-Ca^{2+}{}_{(aq)} = 2:1$; 300 *s*C
Epigenetic $MgCO_3$ rhombohedron on the magnesite "motherphase". 250× (2.5 M solution)

of the crystalline phase is proportional to the Mg/Ca ratio of the total system. Though +4.73% CaO excess was measured in the 2 : 1 system, it has been almost doubled in the 1 : 1 system (+8.29% CaO), which has been crystallized in epigenetic calcite and aragonite. Aragonite and calcite aggregates are seen on the surface of a magnesite rhombohedron (Photo: 29; 246×).

A similar pattern is seen in the next photo, where a magnesite crystal is surrounded by dolomite overgrowth, with aragonite aggregates on its surface. $MgCO_{3(s)}$ — $Ca^{2+}{}_{(aq)}$ = 1 : 1; 200 °C. (Photo: 30; 157×)

A mosaic-like aragonite + $MgCO_3$ + dolomite aggregate, with a MgCO — — (?) — $CaMg(CO_3)_2$ — (?) crystal is seen in Photo 32. (Mg : Ca = 1 : 1; 275 °C.) It was initially dissolved, and crystallized after cooling (275×).

Clean epigenetic $MgCO_3$ rhombohedron was formed from the initial magnesite phase at 300 °C. A part of the Ca^{2+} content of the solution was stabilized in form of calcite + aragonite structures, the latter are not visible in the photo (Photo: 32).

In the powdery dolomite + $Ca^{2+}{}_{(aq)}$ system a similar phenomenon to that of $MgCO_{3(s)}$ + $Ca^{2+}{}_{(aq)}$ system was observed. The Mg^{2+} positions can be restrictedly removed from the lattice, mainly controlled by the dislocation fields.

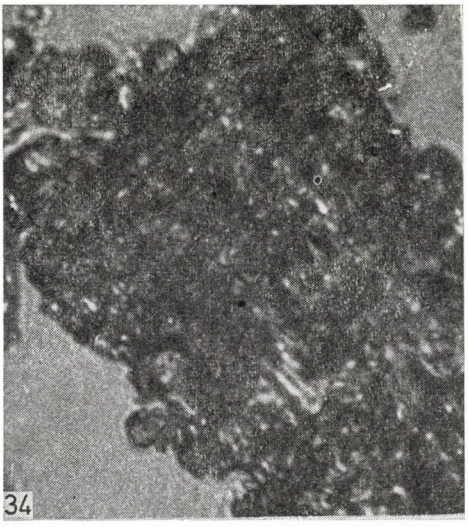
Powdery dolomite — $Ca^{2+}{}_{(aq)}$ = 1 : 2 system, 150 °C: fresh, unaltered dolomite rhombohedra along with epigenetic calcite, $MgCO_3 \cdot 2H_2O$, aragonite-vaterite (?) aggregates (Photo: 33; 275×).

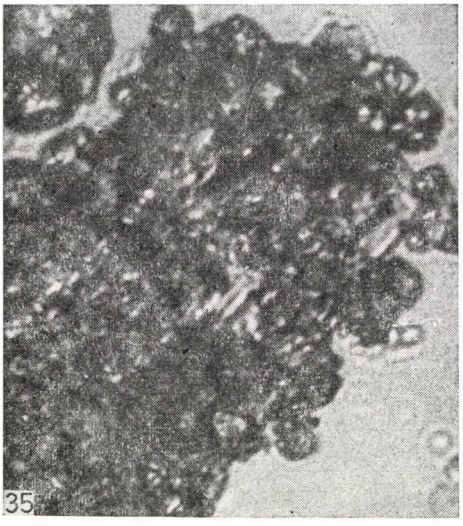
Well visible epigenetic overgrowth of calcite, aragonite, vaterite (?) is seen on the surface of dolomite rhombohedra. The presence of $MgCO_3 \cdot 2H_2O$ was proven by XRD analysis. Powdery dolomite — $Ca^{2+}{}_{(aq)}$ = 1 : 1 system, (Photos: 34, 35). (The first photo illustrates the material treated at 175 °C the second one at 200 °C.)

The last photo provides further evidence against the solid phase recrystallization of magnesite-dolomite by cation exchange. As far as Ca^{2+} replacement is concerned, "de-dolomitization" or "re-calcitization" does not take place. The dolomite "mother-phase" is sharply separated from the calcite-aragonite-vaterite overgrowth on the photo (Photo: 36).

The geological and mineralogical interpretation of the results are summarized as follows:

1. $CaCO_3$—calcite (limestone)—Mg^{2+} fixation = dolomite
   = dolomitization
2. $CaCO_3$—calcite (limestone)—Mg^{2+} fixation = magnesite
   = magnesitization
3. $CaMg(CO_3)_2$—dolomite—Mg^{2+} fixation = magnesite
   = magnesitization
   = de-dolomitization





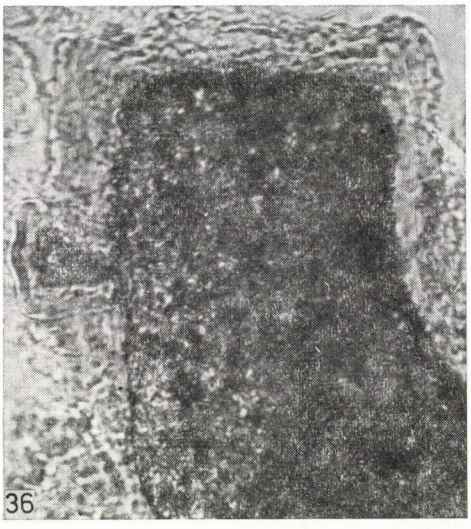


Plate IX

33. Powdery dolomite $-Ca^{2+}_{(aq)} = 1 : 2$; 150 °C
Dolomite rhombohedra with epigenetic $MgCO_3 \cdot 2\ H_2O$ — calcite (— aragonite—vaterite) precipitations; (2.5 M solution); 275 ×

34. Powdery dolomite $+Ca^{2+}_{(aq)} = 1 : 1$; 175 °C
Calcite—aragonite—vaterite (?) aggregates on dolomite (2.5 M solution); 275 ×

35. Powdery dolomite $+Ca^{2+}_{(aq)} = 1 : 1$; 200 °C
Similar to picture 34. (2.5 M solution); 275 ×

36. Powdery dolomite $+Ca^{2+}_{(aq)} = 1 : 1$; 300 °C
Epigenetic frame of calcite — vaterite — aragonite surrounding a dolomite rhombohedron (2.5 M solution); 275 ×

4. $CaMg(CO_3)_2$—dolomite—Ca^{2+} fixation	highly restricted due to lattice structure = no de-dolomitization or recalcitization
5. $MgCO_3$—magnesite—Ca^{2+} fixation	highly restricted due to lattice structure = no "de-magnesitization" or "re-calcitization"
6. $MgCO_3$—magnesite—Ca^{2+} fixation	highly restricted due to lattice structure = no "de-magnesitization" or" re-calcitization

The de-dolomitization term can be applied only on those processes, which can be described by the dolomite : Mg^{2+} system, in which the dolomite may alter to magnesite owing to the effect of Mg-rich solutions (de-dolomitization: stratiform magnesite deposits). This process, however, is irreversible in theory. The Ca fixation of the dolomite is of very small scale, the dolomitization is practically an irreversible process [i.e. $CaMg(CO_3)_2 \rightarrow CaCO_3$], the dolomite can not re-calcitize, de-dolomitize upon the effects of Ca-rich solutions. Therefore, the "de-dolomitization" term is valid only for "dolomite-Mg^{2+}" systems, and can not be used in "dolomite-Ca solution" systems. The calcite components in a dolomite sequence are epigenetic precipitations from solutions, and not crystallized products of a replacement process in the dolomite, in which the Mg would replace the Ca.

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О ПРОЦЕССАХ ДОЛОМИТИЗАЦИИ, «ДЕДОЛОМИТИЗАЦИИ», «ПЕРЕКАЛЬЦИТИЗАЦИИ», ПРОИСХОДЯЩИХ В ГИДРОТЕРМАЛЬНЫХ УСЛОВИЯХ

Я. Кишш

Резюме

Метасоматическим процессам известковых пород посвящена обширная литература. Эксперименты по моделям, проведенные в лабораторных условиях в течение последних десятилетий, проливали свет на ряд эпизодов и моментов преобразования (кальцитовой решетки в доломитовую, в решетки и т. д.). При этом можно встретиться с некоторыми противоречивыми и требующими проведения дальнейших исследований, фактическими данными, касающимися образования доломитовой решетки («дедоломитизация»). В настоящее время а-приори пе «первичное сингенетическое» образование доломитов в «нормальных» условиях осадконакопления все еще является нерешенным вопросом. Доломиты районов типа «шабка» также являются эпигенными образованиями по калциту, а доломиты, обнаруживаемые в современных известковистых илах Большой Венгерской низменности, также не одновозрастны с кальцитом.

Лабораторные эксперименты по моделям попытались подойти к решению проблемы, сформулированной в заглавии, путем изучения «системы твердая фаза: раствор = $CaCO_{3(s)}$: $MgCl_{2(aq)}$. Оцениваются химические и кристаллохимические характеристики фазовых преобразований, причем из физико-химических параметров процесса анализируются прежде всего влияние и роль температуры. Эксперименты были проведены в закрытой системе — автоклаве. Химические изменения и кристаллохимические, а также рентгеновско-диффрактометрические данные образовавшихся кристаллических фаз обобщены в виде таблиц. Относительные изменения изотопов δ_O и δ_C в фазах, полученных в условиях реаличных температур, заслуживают особого внимания.

Изучены были следующие бинарные системы:

а) $CaCO_3$ (кальцит): $MgCl_{2(aq)}$

б) $CaMg(CO_3)_2$ (пылистый доломит): $MgCl_{2(aq)}$

$T° = 50, 75, 100, 125, 150, 200, 225, 250, 300, 350$ °C.

Концентрация магния в растворе была подобрана, так, чтобы соотношения с кальцием твердой фазы были бы следующие: Ca : Mg = 70 : 30; 50 : 50; 30 : 70.

Описывается образование доломита (протодоломита) и магнезит-баррингтонита («доломитизация», «магнезитизация») в условиях вышеизложенной бинарной системы.

Позиции Ca^{2+} кальцитовой решетки могут быть полностью замещены. Фазы были нестоихиометрическими, большей частью лишенными аниона или страдающими дефицитом его. В системе $CaCO_{3(s)}$: $MgCl_{2(aq)}$ оптимальная температура образования доломита составляла 175°C, выше этой температуры большей частью образовалось $MgCO_3$ (магнезитовая решетка). Замещению позиций Ca^{2+} (пылистого) доломита способствуют сильно разбавленные растворы концентрата Mg, богатые магнием растворы вызывали противоположный эффект, причем по крайней мере 30% от общего количества доломита превращается в магнезит. В сущности механизм «дедоломитизации» выражается преобразованием доломита в магнезит.

в) $MgCO_{3(s)}$ — магнезит: $CaCl_{2(aq)}$;

г) $CaMg(CO_3)_2$ — пылистый доломит: $CaCl_{2(aq)}$.

Температура и концентрации растворов были аналогичны бинарным системам *а* и *б*.

Система $MgCO_{3(s)}$: $CaCl_{2(aq)}$ состояла из фаз кальцита, арагонита, ватерита, доломита и триклинального баррингтонита. Количественное присутствие этих фаз и упорядоченность кристаллических фаз обусловливаются температурой («перекальцитизация»?).

Изучение системы $CaMg(CO_3)_2$: $CaCl_{2(aq)}$ потребовало следить за процессом «дедоломитизация-перекальцитизация», точнее за механизмом перестройки кристаллических решеток. Нами наблюдалось образование кальцита; арагонита, ватерита (?), $MgCO_3 \cdot 2H_2O$ и $MgCO_3$ в количестве ограниченном в зависимости от температуры и концентрации Ca^{2+} в растворе, но эти минеральные фазы формировались не в результате перестройки доломитовой решетки (например путем замещения магния кальцием), а выделялись из раствора и осаждались на поверхности ненарушенных ромбоэдров доломита.

В природных условиях — при пониженной температуре — перестройка доломита в кальцит («дедоломитизация») вряд ли может произойти в условиях повышенной температуры возможность этого ограничена даже по кристаллохимическим причинам.

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SYNCHISITE-(Nd) FROM GREBNIK BAUXITE DEPOSIT (YUGOSLAVIA)

By

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Electron probe analysis of a bauxite sample with the highest content of the rare earth elements from Grebnik bauxite deposit revealed the presence of *synchisite-(Nd)*, a mineral of the bastnaesite group. The first analysis of this mineral is presented and the genesis discussed.

Introduction

The geochemical study of the Mediterranean karstic bauxites revealed the concentration of many trace elements (Ni, Co, Cu, Mn, Mo, Y, La—Lu, Pg, Zn, Hg) at the bottom of the deposits and in the contact with the foot wall limestone. The enrichment "per descensum" of these elements is considered as contemporaneous with removal of silica, i.e. with the bauxitization process in a karstic environment (MAKSIMOVIĆ, 1976).

The concentration of the rare earth elements (REE) was observed in Grebnik bauxite deposit, Yugoslavia. In this deposit neodimium was concentrated the most in bauxite sample G.12 just above the foot wall limestone (MAKSIMOVIĆ and ROALDSET, 1976). The study by electron probe analysis of the G.12 bauxite sample is presented in this paper.

Analytical method

The study of minerals of the REE in bauxite samples was carried out by means of the JEOL JXA-5 electron probe at an accelerating voltage of 25 kV. Y, La, Pr, and some lanthanide elements having even atomic numbers (Ce, Nd, Sm, Gd, Dy, Er, Yb) were determined. Generally the Lα lines were selected for measurements, except for Pr, for which Lβ was used. In the case of Gd (Lα) an intensity correction method was applied, because of the disturbing position of La (Lγ 2) and Ce (Lγ 1) lines. The pure metal REE standards were used, supplied by Pierce Inorganics, Rotterdam. The ZAF correction was done according to the correction program given by DUNCOMB and JONES (1969).

Results and discussion

The major constituents of the Grebnik deposit are diaspore and hematite. The section through the deposit and the concentrations of the lanthanides along the vertical profile were given by MAKSIMOVIĆ and ROALDSET (1976). The highest content of total lanthanides (Σ Ln) was found in the lowermost part of the deposit, in sample G.12 with Σ Ln = 4262 ppm. Neodimium, with 1107 ppm, was the most concentrated REE in this sample. It indicated the presence of a Nd-rich mineral of the REE.

In order to determine the Nd mineral, all samples of bauxites collected along the vertical profile were studied by electron probe. However, only in

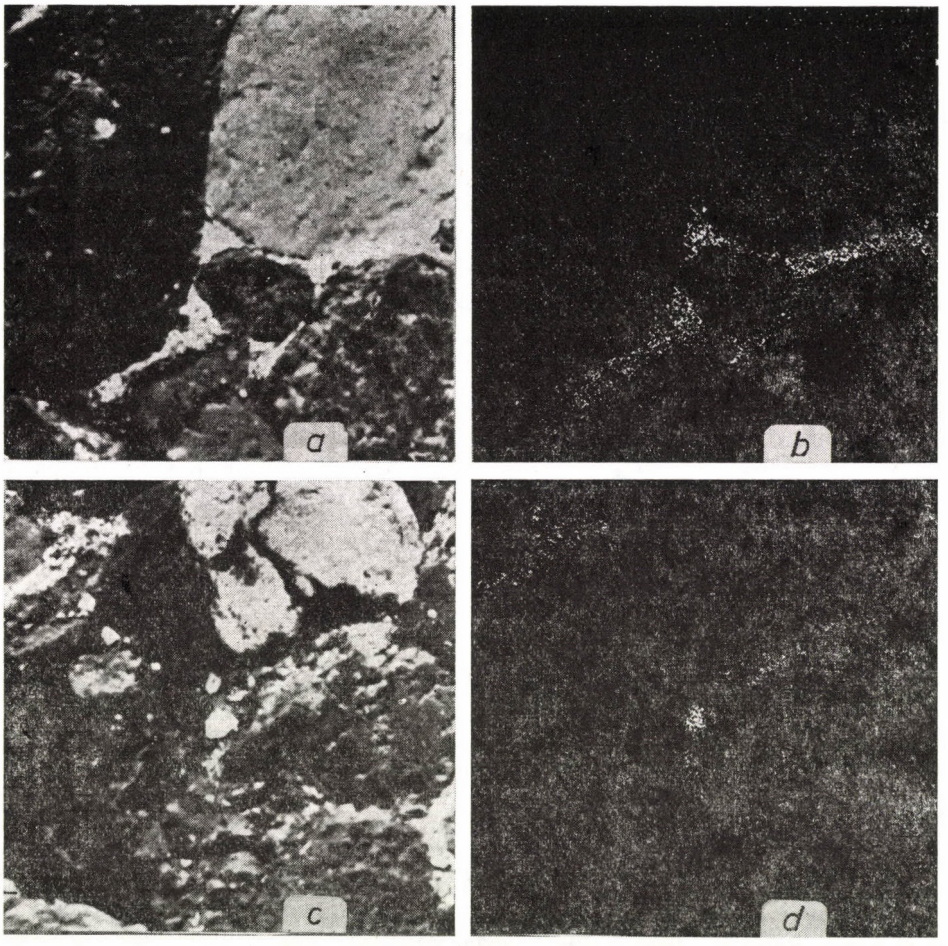

Fig. 1. Synchisite-(Nd) in diasporic bauxite from Grebnik, sample G.12. Electron pictures: a) and c) 6 mm = 33 μm; b) X-ray picture of picture a): Nd_{La}; d): X-ray picture of picture c): Nd_{La}

sample G.12, above the foot wall limestone, a Nd-rich mineral of the bastnaesite group was discovered using a combination of back-scattered electron images and searching by the characteristic wave length of the lanthanides, particularly of neodimium (MAKSIMOVIĆ and PANTÓ, 1978).

Nd-rich bastnaesite generally occurs between oolites as secondary space fillings and micropore fillings in the cement (Figs 1 and 2). Patches of this mineral are often composed of very fine micropore fillings, less than 1 μm in diameter, showing no sharp boundaries.

The electron probe analysis of the Nd-rich mineral was made on three homogeneous space fillings of the diameter of about 30×30 μm. Two of these three grains are shown in Figs 2 and 3. The average analysis of the three measurements is given in Table 1.

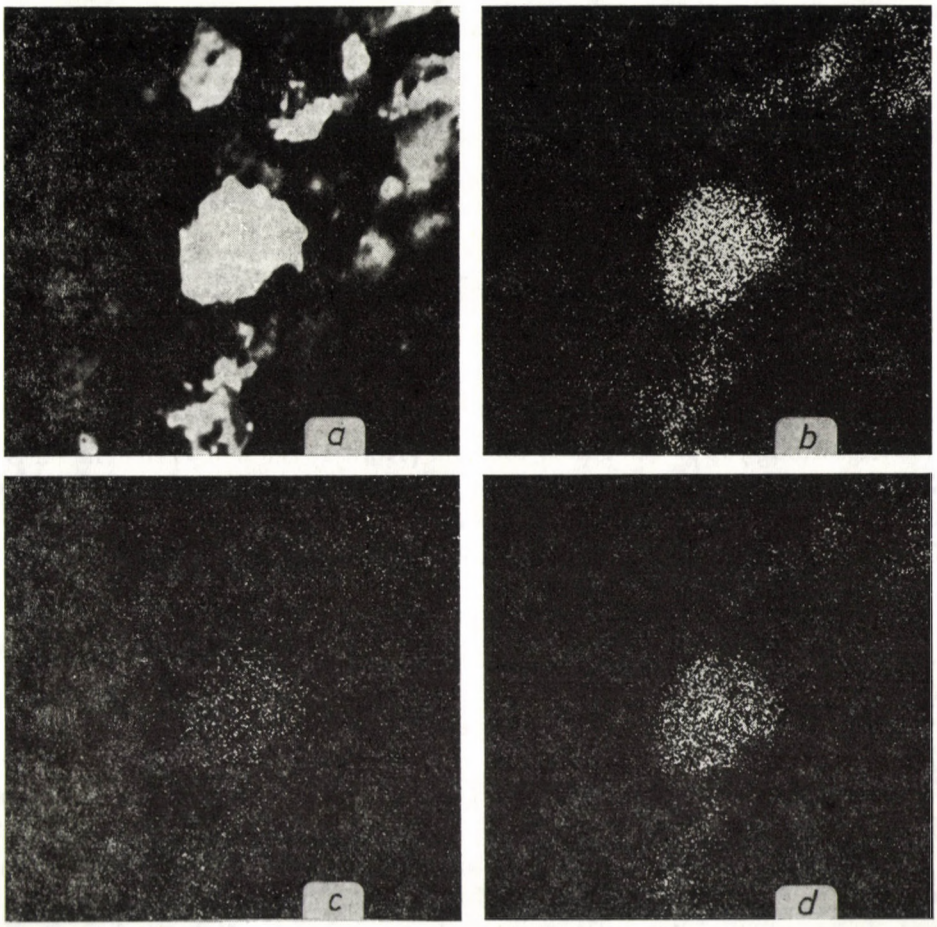

Fig. 2. Synchisite-(Nd) in sample G.12. Electron picture: a) 6 mm = 8.3 μm; X-ray picture of picture a): b) $Nd_{L\alpha}$; c) $Ce_{L\alpha}$; d) $La_{L\alpha}$

Table 1

Analysis of synchisite-(Nd) by electron microprobe (%)		Mol. ratios	Number of ions on the basis of 6 (O)
La_2O_3	15.5	0.0476	0.29
Ce_2O_3	2.4	0.0073	0.04
Pr_2O_3	4.4	0.0133	0.09
Nd_2O_3	19.0	0.0565	0.36
Sm_2O_3	3.6	0.0103	0.06
Gd_2O_3	2.5	0.0069	0.04
Dy_2O_3	1.4	0.0037	0.02
Y_2O_3	3.8	0.0168	0.10
CaO	18.2	0.3245	1.05
F	6.1	0.3210	1.04
Total	76.9		
Less O for F	−2.5		
	74.4		
CO_2*	25.6	0.5818	1.90
	100.0		

* CO_2 was added to 100.0% of the total
Calculated formula:
$Ca_{1\cdot05}(Nd_{\cdot36}La_{\cdot29}Y_{\cdot10}Pr_{\cdot09}Sm_{\cdot06}Ce_{\cdot04}Gd_{\cdot04}Dy_{\cdot02})_{1\cdot00}(CO_3)_{1\cdot90}F_{1\cdot04}$

According to the calculated formula (Table 1) the analyzed mineral is a Nd-rich member of the bastnaesite group, which corresponds to *synchisite-(Nd)*. The calculated formula is close to the ideal formula of synchisite, Ca(Ce, La) $(CO_3)_2F$, but with neodimium as the most abundant REE.

The X-ray diffraction study of the G.12 bauxite sample was not successful in detecting the reflections of the synchisite-(Nd) even after using various separation techniques.

Synchisite-(Nd) was first discovered by VORMA et al. (1966), in Pyörönmaa pegmatite in SW Finland by X-ray diffraction and by qualitative tests. The analysis presented in Table 1 is the first analysis of this mineral. The Y-rich member, synchisite-(Y) also was described in the literature (FLEISCHER, 1975, 1978).

The recent geological and mineralogical study of the bauxite-bearing area in the Grebnik Mountain revealed the presence of relics of the Cretaceous weathering crust on ultramafic rocks, and on the rocks of diabase-chert formation, built up mostly of shale. The weathered shale and diabase were the sources

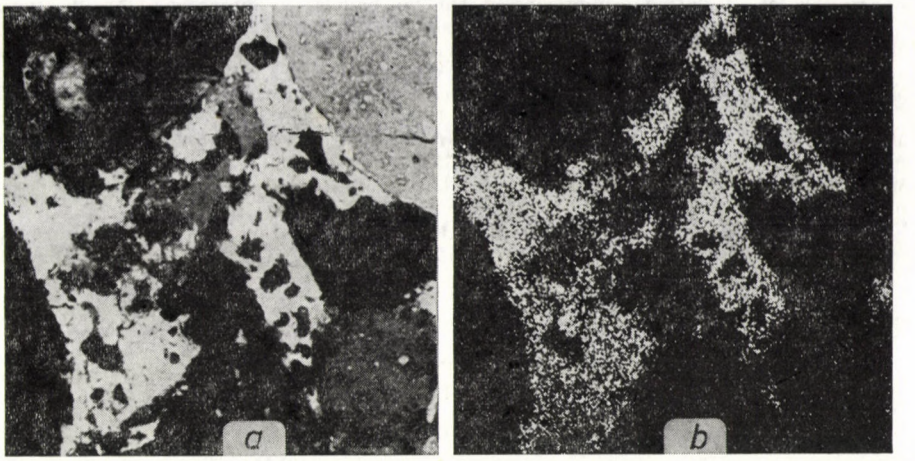

Fig. 3. A large grain of synchisite-(Nd) in sample G.12. Electron picture: a) 6 mm = 33 μm; X-ray picture of picture a) : b) $Nd_{L\alpha}$

of alumina, yttrium and lanthanides (MAKSIMOVIĆ, 1980). In the red, kaolinic clay from the weathering crust on shale, the total lanthanides in the form of adsorbed ions amounts 45.7% of the total. It indicates that yttrium and lanthanides were transported from the weathered shale and diabase into adjacent karstic depression partly as adsorbed ions on clay particles. During the bauxitization process they were readily removed per descensum and concentrated above the foot wall limestone, which acted as an effective geochemical barrier for descending solutions.

Conclusion

The first data on the chemical composition of synchisite-(Nd) were obtained by electron probe analysis. The calculated formula is close to the ideal synchisite, Ca (Ce, La) $(CO_3)_2F$, but with neodimium as predominant rare earth element.

Synchisite-(Nd) occurs in the lowermost part of the Grebnik karstic bauxite deposit. It has been formed during the bauxitization of the red, kaolinic clays, which were washed down into a karstic depression from the Cretaceous weathering crust on shale and diabase.

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(Nd)-СИНХИЗИТ ИЗ БОКСИТОВОГО МЕСТОРОЖДЕНИЯ ГРЕБНИК (ЮГОСЛАВИЯ)

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Резюме

Исследование электронным микрозондом образца боксита наи-вытшего содержания РЭЭ из месторождения Гребник выявило присуствие в нем (Nd)-синхизита, минерала из группы бастнезита. Приводится первый анализ этого минерала и обсуждается его происхождение.

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FORMATION OF BAUXITE FROM KHONDALITE

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The authors report the alteration of sillimanite to bauxite during normal coursy of weathering. Gibbsite is the product of alteration. Replacement of sillimanite b.e gibbsite, or a mixture of goethite and gibbsite or rarely by goethite is also reported

Bauxite deposits are found to occur in workable quantities in Anantagiri Hill Region, 80 km away from Visakhapatnam Port (17°42′30″ N, 83°17′40″ E) situated on the East Coast of India. The host rocks for these bauxite occurrences are high grade metamorphic rocks, the garnet-sillimanite gneisses otherwise called as Khondalites. During the course of systematic investigations on the genesis and geochemistry of these bauxite deposits, a number of microsections of these rocks showing various stages of bauxitization have been examined. The samples for these investigations were collected from different places and from different depths. In the fresh unaltered host rocks, feldspar is the dominant mineral followed by sillimanite which is the only member of the aluminous silicate Polymorphic group present. The individual crystals of sillimanite are prismatic, elongated and sometimes needle-shaped, with a cleavage parallel to (010). These are usually found in association with garnets and sometimes as inclusions in them. Fibrolitic sillimanite with rhombic shape is found as scattered concentrations. The crystals have cracks perpendicular to cleavage directions. Most of the grains are optically positive; but, however, optically negative crystals are not uncommon. Extinction is usually straight but sometimes inclined. It has an optical axial angle around 27° with values of α and γ 1.66 and 1.68, respectively.

Alteration

The mineral, sillimanite shows initial stages of alteration after major portions of feldspars and garnet are subjected to alteration. In rare cases, its alteration is almost simultaneous with that of garnet, the smaller grainsize of the mineral in this case perhaps facilitating this earlier alteration. The process of alteration proceeds along periphery as well as along cracks (Fig. 1),

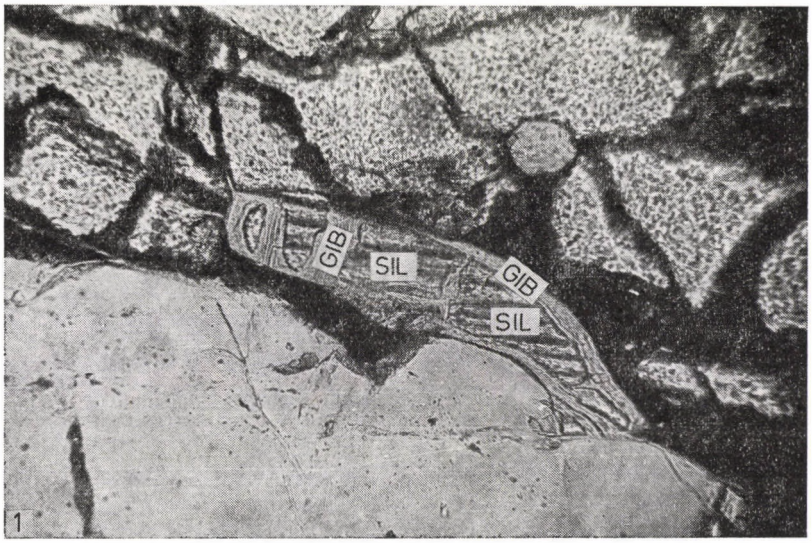

Fig. 1. Sillimanite (SIL) altering to gibbsite (GIB) along cracks and periphery. Plane light, ×40

colourless and sometimes ferruginous solutions bringing about the alteration. The cleavage directions have not been preferred by the solutions bringing about alteration. Complete alteration of sillimanite is a rare feature.

Product of Alteration

The alteration product derived from sillimanite is one or more of the following: a) some isotropic material; b) fine-grained gibbsite; c) mixture of goethite and gibbsite; d) clay mineral. The isotropic material can be submicroscopic-aluminium hydroxide or halloysite. The X-ray diffraction study of these samples has not proved the presence of halloysite. Moreover, a gradual conversion of this isotropic material to fine-grained gibbsite has been noticed (Fig. 2). Therefore, the isotropic material is believed to be submicroscopic aluminium hydroxide.

The fine-grained gibbsite derived from sillimanite shows wavy extinction. In advanced stages of alteration, small relicts of irregular grains of sillimanite are found embedded in fine-grained gibbsite. These relicts are usually in optical continuity (Fig. 3). In some cases, the matrix of these relict grains is goethite which is a result of alteration by ferruginous solutions.

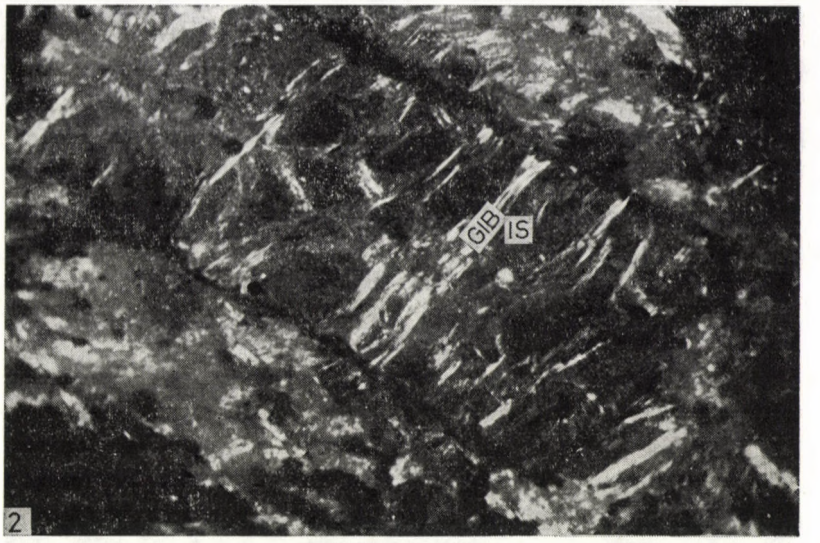

Fig. 2. Transformation of Isotropic material (IS) to gibbsite (GIB) (Sillimanite alteration). Crossed Nicols, ×10

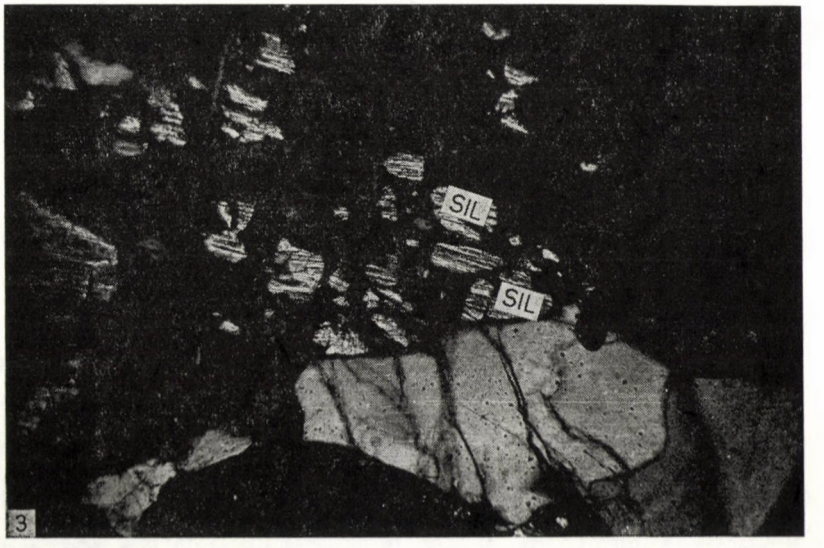

Fig. 3. Detached pieces of sillimanite (SIL) with optical continuity. Crossed Nicols, ×40

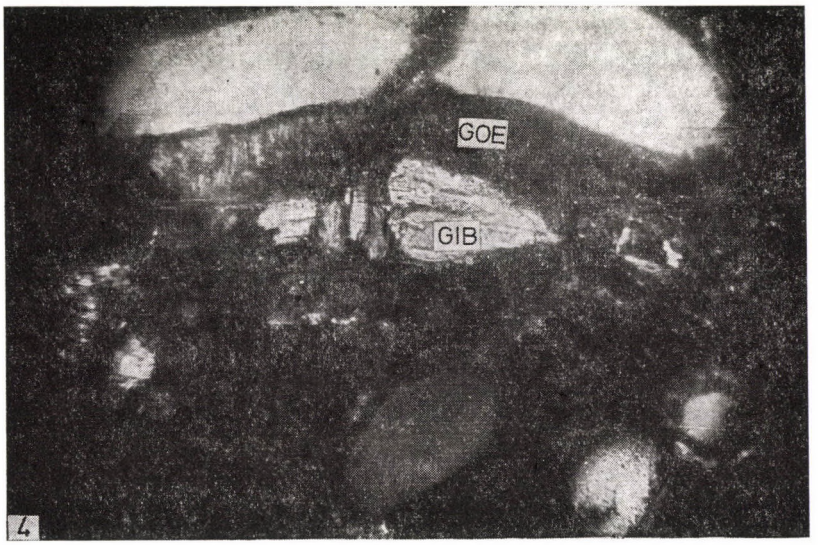

Fig. 4. Pseudomorph of gibbsite (GIB) and goethite (GOE) after sillimanite. Crossed Nicols, × 50

Replacement of Sillimanite

During the advanced stage of bauxitization when solution activity dominates over decomposition and disintegration, relict sillimanite as well as the products of sillimanite alteration are subjected to replacement by ferruginous as well as by aluminous solutions, with the result that assemblages of gibbsite and goethite are pseudomorphous after sillimanite with the general shape of the grain well-preserved (Fig. 4). During this stage, replacement of relict sillimanite is more prominent than alteration of individual sillimanite. During such replacement, in some cases, the sillimanite relicts have suffered some movement from their original positions as shown by their haphazard distribution in gibbsite and goethite matrix.

Discussion

The mineral, sillimanite is usually highly resistant to weathering. However, in the present area, it is found to be altering and in a few rocks examined even completely [1, 2]. The alteration is believed to have been brought about by a process of hydrolysis, and it may be represented by the following equation:

$$2\ Al_2SiO_5 + 10\ H_2O \rightarrow 4\ Al(OH)_3 + 2\ H_4SiO_4$$

The reacting solutions in the form of circulating waters passing through the cracks of sillimanite have brought about this alteration (Fig. 5). Some-

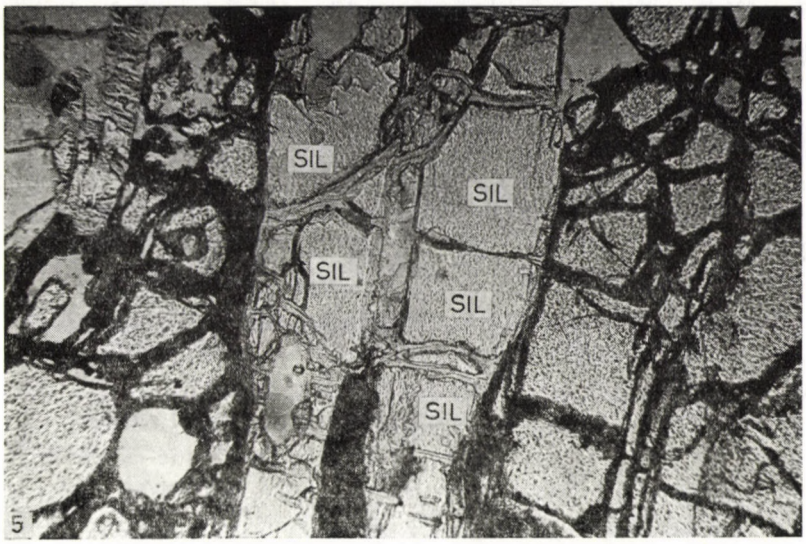

Fig. 5. Alteration of sillimanite along cracks. Plane light, ×10

times, the initial product of alteration is submicroscopic aluminium hydroxide which shows definite indications of transition to fine-grained crystalline gibbsite. There are instances where the submicroscopic aluminium hydroxide is absent and the mineral sillimanite seems to have directly given rise to crystalline gibbsite during alteration. It is possible that in such cases, the intermediary aluminium hydroxide was completely converted to crystalline gibbsite. In this connection, the observation made by FIELDS and SWINDALE (1954) that all the primary minerals must pass through the amorphous stage in a transition to secondary crystalline material is worthy to note.

The reacting solutions break up the Si—O tetrahedra of sillimanite structure and leach away the silica that is dissolved in the reacting solutions thereby leaving aluminium hydroxide concentrate in the residue as a colloidal precipitate which gradually is converted to a crystalline state. Under poor drainage conditions, the released silicic acid may react with aluminium hydroxide to form kaolinite, whereas good drainage conditions facilitate the formation of gibbsite, by removing silicic acid rapidly from the site. If alteration has been brought about by ferruginous solutions, secondary gibbsite and secondary goethite are formed as layers around a sillimanite core. The well-preserved shapes of altered sillimanite grains indicate that during alteration, the shapes of the material grains were not obliterated. Alteration has proceeded along the cracks which are the stress planes but not along the more compact cleavage planes.

In addition to the normal process of bauxitization by hydrolysis, sillimanite is found to have undergone replacement also mostly by gibbsite or by a mixture of goethite and gibbsite and rarely by goethite. Such type of replacement is noticed only in the advanced stages of weathering. Absence of this type of replacement in the initial stages is believed to be due to the fact that at this stage, the waters are charged more with reacting constituents than with replacing constituents.

Both the altered portion and the unaltered core of the sillimanite grain are found to have undergone replacement, the former to a better degree in which replacement is nothing but replacement of earlier formed gibbsite or goethite or both. A mere exchange of material takes place under quiet conditions as a result of which well-crystallized gibbsite is found in the place of microcrystalline gibbsite or goethite formed from sillimanite during alteration. In the case of unaltered core, the aluminium hydroxide-rich solutions, on coming in contact with the unaltered mineral surface disengage the cation Al^{3+} from its structure, which goes into solution immediately. Silica must have been released evidently not in the form of Si—O tetrahedra but as silicic acid, as otherwise the Si—O tetrahedra would have been available for the formation of hydrated silicates. Simultaneous with the release of cation, Al^{3+} and leaching of silicic acid, aluminium hydroxide already present in the solution might have been emplaced in the form of well-developed crystals of gibbsite. The solutions are still rich in aluminium hydroxide, since fresh cation of Al^{3+} released from sillimanite structure has been added to the solutions.

Acknowledgements

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ОБРАЗОВАНИЕ БОКСИТОВ ИЗ ХОНДОЛИТА

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Резюме

Рассматривается преобразование хондолита (силлиманита) в бокситы в процессе нормального выветривания. Гиббсит является продуктом превращения. Кроме того, авторы обращают внимание на замещение силлиманита гиббситом или смесью гётита и гиббсита или же, реже, гётитом.

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ACCUMULATION OF BAUXITIC MATERIALS IN GUANTÁNAMO PROVINCE, CUBA

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The principal geomorphologic types of the accumulation zones of more or less bauxitized sediments in the Sierra Guaso and in the Sierra de Caujeri are: karstic depressions, tectonically controlled depressions, and mesetas. The chemical and mineralogical composition of 31 samples is presented in tables and figures. The Sierra Guaso area is characterized by the predominance of gibbsite as alumina bearing mineral accompanied by boehmite, while the Sierra de Caujeri by the exclusive enrichment of boehmite. A secondary enrichment of iron in the Sierra Guaso samples has been established. The possible source rocks (green schists, volcanites etc.) and the theorically possible phases of the complex geochemical process of allitization and bauxitization are discussed and compared with the make-up of actual lateritic crusts of weathering. A more detailed exploration of the two areas by means of a regular network shallow core drilling is proposed.

1. Antecedents

Earlier search for bauxite in Cuba resulted in fairly poor estimates as to the possibilities (BRAMLETTE, 1943; MAT'L, 1967), with the single exception of the San Francisco area in the West of the country (KUDRIAVTSEV, V.—MESSINA, V., 1969). In the early seventies, however, during the complex investigation of the weathering crusts in the former province of Oriente, a lot of data emerged pointing to the possibility of bauxite formation and even to the presence of bauxitic materials (BUGUELSKIY, Y. Y., 1972; FINKO, V. I.—PETROV, V. P.—COUTIN, D. P., 1973; BUGUELSKIY, Y. Y.—FORMELL, F., 1973 and 1974; FINKO, V. I.—KORIN, I. Z.—COUTIN, D. P., 1974; GUERASIMOV, E.—GURIEV, G.—KOSTAROV, G.—SAUNDERS, E., 1974; FINKO, V. I.—LITZAREV, M. A.—SAMOTOIN, N. D.—COUTIN, D. P., 1974). From 1973 on, BUGUELSKIY, Y. Y. and FORMELL, F. distinguished two types of weathering crust, and, accordingly, two possible genetic types of bauxite: "in situ" bauxite in primary lateritic crust of weathering, and redeposited bauxite in depressions of limestones.

The Cuban-Hungarian Geological Mapping Party of the Institute of Geology and Paleontology of the Cuban Academy of Sciences, surveying the former province of Oriente at a scale of 1 : 250,000, dedicated particular attention to the sampling in areas of possible occurrence of the afore-mentioned two types.

In areas, where the crust of weathering developed on mafic rocks (gabbroids) and schists, the alteration is of kaolinitic composition, with only extremely local enrichment of gibbsite. The alteration of rhyolite tuffs also has produced kaolinitic materials.

During the mapping operations, quite a number of karstic depressions filled up with allitic clays were discovered in the Guantánamo region, in the mountains called Sierra del Guaso y Sierra de Caujerí.

The contours were checked on aerial photographs, and sampling was done manually from the surface, in some cases by means of hand-drills down to one metre. The results were described in an unpublished report (NAGY, COUTIN, BREZSNYÁNSZKY, GYARMATI and RADÓCZ, 1975) and presented to the VIIIth Geological Meeting of the Caribbean in Curazao (by FORMELL, GYARMATI and JAKUS, 1977). The results have been incorporated and complemented by the laboratory investigation of some new samples by two Hungarian bauxite experts (E. DUDICH, 1976 and E. SZABÓ, 1977 in their unpublished reports).

2. Geological Setting

2.1. The principal geomorphological units of the Guantanámo region are the following.

1. The Bayate Basin situated between the 2. southern slope of the Sierra de Cristal and of the 3. Sierra Guaso; 4. the SW part of the Toa Range, 5. the Caujerí Valley, 6. the Sierra Caujerí, 7. the Sierra Maquey, 8. the Guantánamo Basin, and in the central part, 9. the Sierra de Canasta.

The Bayate Basin is a genuine sedimentary basin, while the Guantánamo Basin and the Caujerí Valley are of erosional origin.

The Sierra de Canasta is of 440 m and the Sierra Maquey of 320 m maximum elevation a.s.l., while the height range of the Sierra Guaso and Sierra Caujerí is between 500 and 750 m. No wonder, that the latter two are characterized by depressions filled up with allitic sediments (Fig. 1).

Most of these depressions are of karstic origin; some of them (e.g. Naranjal and Limonar) are rather soft undulations of a gently karstified limestone meseta, and one is known to be a tectonically controlled depression (Table 1).

2.2. *Geological formations*

1. *Serpentinitized ultramafic rocks* are tectonically superimposed (as overthrusts and nappes) to the volcano-sedimentary sequences of the Cretaceous (Aptian-Turonian) and to the metamorphic mass of the region. Mostly peridotites and harzburgites, they are widespread N and NW of the Sierra Guaso, and a restricted outcrop is known (Convento Hill) in the Caujerí Valley. Geo-

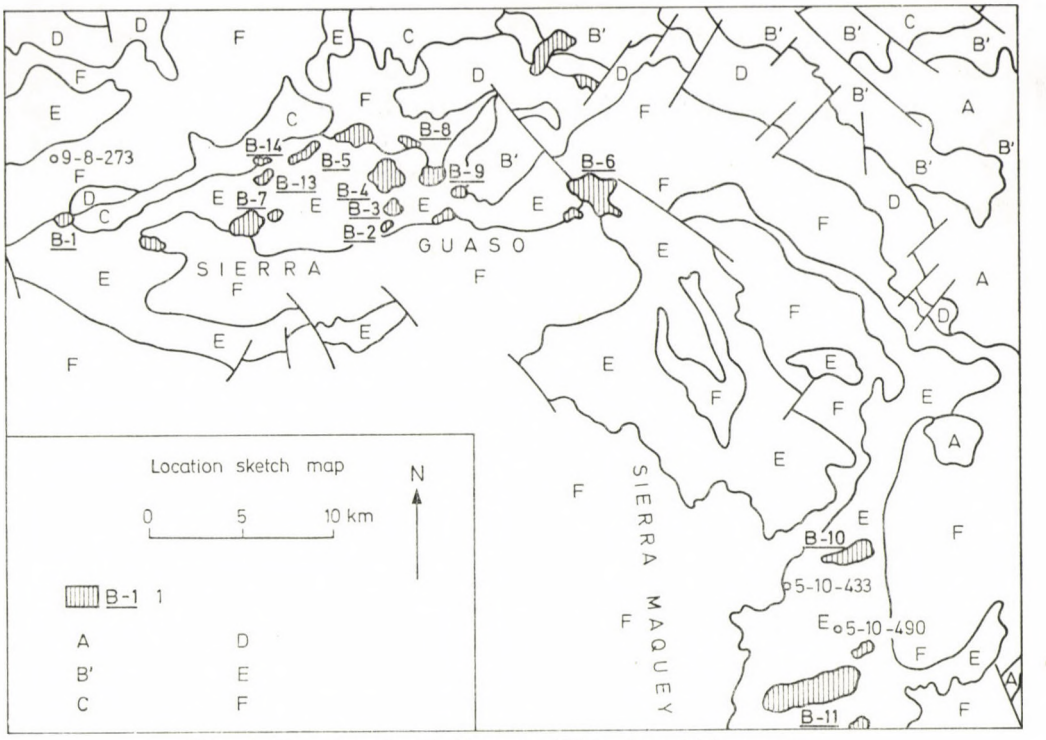

Fig. 1. Location sketch Map. 1. Allitic (bauxitic) outcrops.
A = metamorphic rocks; B' = ultramafic rocks; C = Cretaceous volcanites; D = Paleogene volcanites; E = limestones (Pg_2^2, N_1^1); F = marls, sandstones, conglomerates (Pg_2^3, N_2)

Table 1
Occurrences of allitic sediments

No. in Fig. 1	Denomination	Area, km²	Type
B-1	Concepción del Cuzco	0.6	karstic depression
B-1	Dos Mujeres I	0.2	karstic depression
B-3	Dos Mujeres II	0.5	karstic depression
B-4	Alto del Mango	1.0	karstic depression
B-5	Guayabal del Peru	1.0	karstic depression
B-6	La Prenda	2.5	karstic depression
B-7	Monte Alto	1.2	karstic depression
B-8	Manguito	0.3	uncertain
B-9	Hondones	0.8	karstic depression
B-10	Naranjal	1.2	meseta
B-11	La Mesa	4.0	tectonically controlled depression
B-13	Potosi	0.2	uncertain
B-14	Santo Domingo	0.1	uncertain
9-8-273	Limonar	0.1	meseta

physical data suggest the presence of ultramafic bodies in the basement of the Guantánamo Basin, too.

2. *The metamorphic massif* comprises the Eastern border of the Caujerí Valley, one sector of the Toa Range, and the western end of the Sierra del Purial. It consists of green schist type rocks which are products of a regional metamorphism; mostly metavolcanites of originally mafic and intermediary composition. The deep basement is supposed to be made up by amphibolites and other rocks of more advanced grade of metamorphism. The metamorphic formations are overlain (if covered) by basal conglomerates of various sedimentary formations of Paleogene age.

3. *The volcano-sedimentary sequence* of Cretaceous age is made up by tephroid rocks (tufites, subordinate tuffs, agglomerates and lavas) of intermediary to mafic composition. In some limestone interbeddings Aptian-Turonian microfossils have been found. The sequence is interpreted as the product of an arc-type, prolongated submarine volcanism. The underlying rocks are unknown. Overlying, one encounters terrigenous sediments of Maestrichtian age and in a few cases various Paleogene rocks. Outcrops of this sequence are situated at the borders of the Bayate Basin and in the Toa Range.

4. *The terrigenous sequence of Maestrichtian age* consists of Picota-type conglomerates and Micara-type sandstones, both being of molassoid facies. They are overlain by the basal conglomerates of the Paleogene volcano-sedimentary sequence or by younger sediments. Outcrops have been mapped at the base of the southern slope of the Sierra de Cristal, and N and E of the Bayate Basin.

5. *The Paleogene volcano-sedimentary sequence* starts with a calcareous basal conglomerate (Gran Tierra type) and consists mainly of tephroid tuffs and tufites, with intercalated marls and conglomerates. The volcanogenic rocks are of acidic and intermediary composition, with common alteration phenomena (zeolite and bentonite formation). The partly overlying formations are Middle Eocene limestones or younger sediments. Outcrops are known at the NW border of the Bayate Basin (Achotal), on the N slope of the Sierra Guaso (La Tagua) and in a narrow strip in the SW part of the Toa Range (Bernardo).

6. *The Middle Eocene limestone formations* are subdivided into a well-stratified limestone type and into a compact, occasionally thick-bedded limestone series (Charco Redondo type), which is of more transgressive character. They are covered by Upper Eocene or more recent sediments. Smaller outcrops are situated around the Bayate Basin and larger ones in the central and W parts of the Sierra Guaso. Moreover, the Sierra de Canasta is made up almost entirely by these formations.

7. *Basal conglomerate, type San Ignacio* represents the basal member of the Upper Eocene series in the E slope of the Caujerí Valley. It passes vertically to the sandstones and siltstones of the San Luis formation.

8. *The terrigenous sequence of the Upper Eocene* (the San Luis formation) consists of alternating sandstones and siltstones with lenticular intercalations of conglomerate. This molasse-type formation covers large areas in the Guantánamo Basin and in the Caujerí Valley. Isolated outcrops occur at the NW margin of the Bayate Basin, in the Bernardo and Palenquito areas, and in the SW part of the Toa Range. If covered, it is overlain by the basal conglomerate of the Oligo-miocene siltstone series or by younger sediments.

9. *The siltstone sequence of the Oligo-miocene* represents a sedimentary cycle starting with the basal conglomerate (Cilindro type), continues with badly stratified siltstones (Maquey type) and ends with thickly stratified, porcelain-like limestones (Yateras type). The age is probably Upper Oligocene—Burdigalian. It is covered unconformably by Pliocene and Quaternary conglomerates, and breccias (Jamaica type and Centeno type).

2.3. *The bauxitic accumulations*

The depressions filled up with allitic sediments occur on the top of the Sierras Guaso and Caujerí. The karstified bedrocks belong either to the Charco Redondo formation or to the Yateras type limestone.

Possible source rocks of the primary weathering crust are most likely the items 1, 3, 4, 5 and 9a.

B-1 Concepción del Cuzo: Bedrock: Charco Redondo limestone. Borders: W and SE Charco Redondo limestone; NE: the Cretaceous volcano-sedimentary rocks; N: white sandy marls of the San Luis formation.

B-2 Dos Mujeres I. Bedrock and borders: Charco Redondo limestone.

B-3 Dos Mujeres II. Bedrock and borders: Charco Redondo limestone.

B-4 Alto del Mango. Bedrock: in the W: Charco Redondo limestone; in the E: Yateras type limestone. The same is valid for the borders.

B-5 Guayabal del Peru, La Tagua. Bedrocks and country rocks are Yateras type limestone in the S, the Maquey type siltstone series in the N and E and the Paleogene volcano-sedimentary sequence in the W.

B-6 La Prenda, La Clarita. Bedrock: Yateras type limestone. Borders: the same, with the exception of the SW part, where Maquey type siltstones are outcropping.

B-7 Monte Alto. Bedrock: Charco Redondo limestone in the SE and Yateras type limestone in the NW. Borders: the same limestones.

B-8 Manguito. Bedrock and border: Maquey type siltstone sequence.

B-9 Hondones. Bedrock: Yateras type limestone. Borders: the same, except for the N part, where appear siltstones of Maquey type.

Table 2

Results of the chemical and X-ray diffraction analysis of 31 allitic samples from Guantánamo province

No. %	Al_2O_3	SiO_2	Fe_2O_3	TiO_2	L.o.I.	Total	Module	Boehmite		Illite		Goethite		Quartz	Sampling No.
								Gibbsite	Kaolinite	Chlorite		Hematite		Calcite	
Sierra Caujeri (7 samples)															
1	42.6	18.5	21.9	2.6	13.4	99.3	2.32		47	16	9	28	tr.		5-10-437a 4.5 cm
3	43.2	17.1	22.0	2.5	12.3	97.1	2.53		37	5	10	5	21 12	9	5-10-433
4	42.9	18.2	22.4	2.7	11.2	97.4	2.36		34		16	10	— 39 —	tr.	5-10-437a
5	30.0	33.5	16.1	1.9	12.1	93.6	0.90		15	18	25		35	7	5-10-490
6	30.4	34.5	18.0	2.4	12.1	97.4	0.88			43			— 57 —		5-10-491-1
7	24.1	26.6	13.9	1.7	18.8	85.1	0.91			24		tr.	32		40	5-10-491-2
8	31.2	35.4	17.8	2.3	11.3	98.0	0.88			50	tr.		50			5-10-491-3
Sierra Guaso (22 samples)															
15	37.4	18.2	21.7	2.4	18.6	98.3	2.05	42	5	22			31 tr.			B-4-1
18	26.6	25.3	29.6	2.2	14.3	98.0	1.05			21			73	6		B-6-1
19	37.5	18.9	19.5	1.8	20.1	97.8	1.98	++		+	+	tr.	++	+		B-6-2
20	35.7	22.5	21.0	2.4	17.2	98.8	1.59	17	9	16	19	tr.	39			B-4-2
21	37.6	15.1	25.4	2.5	18.0	98.6	2.49	27	18	16	tr.		37	5		B-4-3
23	28.9	24.7	25.9	3.6	15.2	98.3	1.17	20			14	16	25 11	13		B-5-2
29	28.6	30.8	23.7	2.4	14.3	99.8	0.92			45	21		34			B-9-1
30	26.1	29.8	29.5	1.1	12.9	99.4	0.88	8		21	21		30 11	8		B-9-2
31	25.0	25.7	34.8	1.9	12.9	100.3	0.97			21			66 14			B-9-3
33	29.6	22.5	27.6	2.0	16.9	98.9	1.32	13		22			65			B-6-3
34	20.3	20.3	42.8	1.3	12.3	97.0	1.00				tr.		61	33		B-6-4
35	26.5	26.4	30.6	1.8	14.4	99.7	1.00			39			61			B-6-5
36	26.2	20.3	34.6	2.6	15.6	99.3	1.29	tr.		31	19		50			B-6-6
37	26.8	18.7	35.9	2.1	16.5	100.0	1.43	9		26	7		56			B-6-7
41	32.4	32.3	16.5	2.3	15.4	98.9	1.00	13		17	13		50	7		B-1-4
46	28.4	26.9	26.9	2.6	13.1	97.9	1.06	15	tr.	23			62	tr.		B-3-2
49	25.2	27.5	32.8	1.2	12.1	98.9	0.92						— 33 —	2		SE-12
50	25.5	28.9	31.6	1.2	11.6	98.8	0.88			not	analyzed					SE-13
51	25.3	21.2	34.2	1.4	13.8	95.9	1.19			not	analyzed					EDC-40
464	34.7	16.8	27.7	1.2	18.3	98.7	2.05			not	analyzed					
463	28.4	27.9	25.7	0.9	15.8	98.7	0.98			not	analyzed					
462	24.9	27.8	28.3	0.8	14.1	95.9	0.90			not	analyzed					
Baracoa area (2 samples)															
47	41.4	16.4	22.7	2.4	16.8	99.7	2.52			not	analyzed					3-6-343×B
48	25.4	26.3	31.7	1.2	15.9	100.5	0.96			not	analyzed					6-6-82×B 2 m

B-10 Naranjal. Bedrock and borders: Yateras type limestone.

B-11 La Mesa: Yateras type limestone only.

B-13 Potosi. Bedrock: Yateras type limestone. Borders: the same, with the exception of a restricted sector in the SE, where the Charco Redondo limestone appears on the surface.

B-14 Santo Domingo. Bedrock: Yateras type limestone. Borders: chiefly the same; in the NW — the Cretaceous volcano-sedimentary sequence.

9-8-273 Limonar. Bedrock: calcareous siltstones of Maquey type. Here the allites are covered by a thin layer of soil.

Several other small depressions have been mapped (and identified on aerial photographs) in the Sierra Guaso and in the Sierra Caujerí, of which no samples and laboratory analyses are available at present. All these are represented on Fig. 1 (without numeration) and would require further field work.

3. Chemical and Mineralogical Composition

A total of 70 samples have been investigated in the laboratories of the Hungarian Geological Institute in Budapest. (Traditional wet chemical analysis, DTA-DTG-TG analysis using a MOM type derivatograph, and X-ray diffractometry done by a Phillips diffractometer.)

In Table 2 the results concerning altogether 31 samples are presented, which have turned out to be of allitic nature (i.e. their module or $Al_2O_3 : SiO_2$ ratio is higher than 0.87). The data of 2 samples taken by E. Szabó (analysed in the Central Laboratory of the Hungarian Bauxite Prospecting Company at Balatonalmádi), 1 sample taken by E. Dudich (analysed in Santiago de Cuba) and 3 samples taken by Y. Y. Buguelskiy and F. Formell, are also included.

Some typical samples are briefly described in Table 3.

It is obvious from the data, that there is a remarkable difference as to their mineralogical composition between the samples taken in the Sierra Caujerí and those collected in the Sierra Guaso. However, one should keep in mind that the allitic sediments of the Sierra Caujerí are represented by a few samples only.

Anyway, the allites of the Sierra Caujerí are characterized by relatively high boehmite contents, and by the absence of gibbsite. As relates to the clay minerals, the amount of illite in some cases surpasses that of kaolinite and they are accompanied by chlorites. Along with goethite, hematite is also an important mineral of iron.

In the allites of the Sierra Guaso, gibbsite is the main aluminium bearing mineral, the boehmite is subordinate. In the two other groups, kaolinite and goethite, respectively, are predominating.

Table 3

Brief macroscopic description of some allitic samples

Sierra Caujerí area	
N° (2) 5-10-433	Ochreous, friable material of red colour, cementing limestone pebbles
(1 and 4), 5-10-437	*(B-10)* Compact ochereous material, of vivid red colour, cementing angular limestone fragments
(5) 5-10-490-1	Argillaceous "terra rossa", of yellowish pink colour, containing scarce quartz grains
6) 5-10-491-1	*(B-11)* "Terra rossa" of argillaceous character, plastic, of pink colour
(7) 5-10-491-2	*(B-11)* Compact "terra rossa", relatively hard, of yellowish-pink colour
(8) 5-10-491-3	*(B-11)* Very argillaceous "terra rossa", plastic, of vivid red colour
Sierra Guaso area	
(41) *(B-1-4)*	Argillaceous substance, with some quartz, of vivid red colour
(46) *(B-3-2)*	Argillaceous substance of red and yellowish red colour, with dark brown pisolites
(20) *(B-4-2)* and (21) *(B-4-3)*	Slightly plastic argillaceous substance of red and yellowish red colour, with reddish brown pisolites
(23) *(B-5-2)*	Ochreous-terrous, friable substance of red colour with violet patches, with no pisolites, but with a few small quartz grains
(18) *(B-6-1)* and (19) *(B-6-2)*	Ochreous, friable substance of reddish yellow colour
(33) *(B-6-3)*	Slightly argillaceous substance with many reddish brown pisolites
(34) *(B-6-4)*	Compact, red, very ferruginous material of conglomeratic texture and with quartz pebbles
(35) *(B-6-5)*, (36) *(B-6-6)*, (37) *(B-6-7)*	Ochreous-argillaceous substances, friable, with numerous pisolites of reddish brown colour
(29) *(B-9-1)*	Argillaceous, slightly plastic substance of yellowish red colour, with some pisolites
(30) *(B-9-2)*	Argillaceous, somewhat terrous substance of yellowish red colour, with rare pisolites
(31) *(B-9-3)*	Variegated argillaceous material (red-yellow, type "tiger"), with ferrugineous concretions
9-8-273	Argillaceous substance of yellowish red colour, with a few pisolites
Baracoa region (East)	
(47) 3-6-343×B	Slightly argillaceous substance of red colour, without pisolites
48) 6-6-82×B	Argillaceous substance of dark red colour, with brown ferruginous patches.

For both regions, the binary diagrams of the 5 main constituents (Al_2O_3, SiO_2, Fe_2O_3, TiO_2, loss on ignition) have been plotted (Figs 2 and 3).

In both areas have been found some samples to have an $Al_2O_3 : SiO_2$ ratio (= the silica module) higher than 2.0 and Al_2O_3 content of about 40 weight %, which according the Soviet norms could be labelled as "bauxite". However, having adopted the industrial lower limit value of 2.6 (in vigour now in Hungary) we call them only allites.

The titanium module (the $Al_2O_3 : TiO_2$ ratio) seems to be much more constant in the case of the Sierra Caujerí (between 14 and 17) and much more

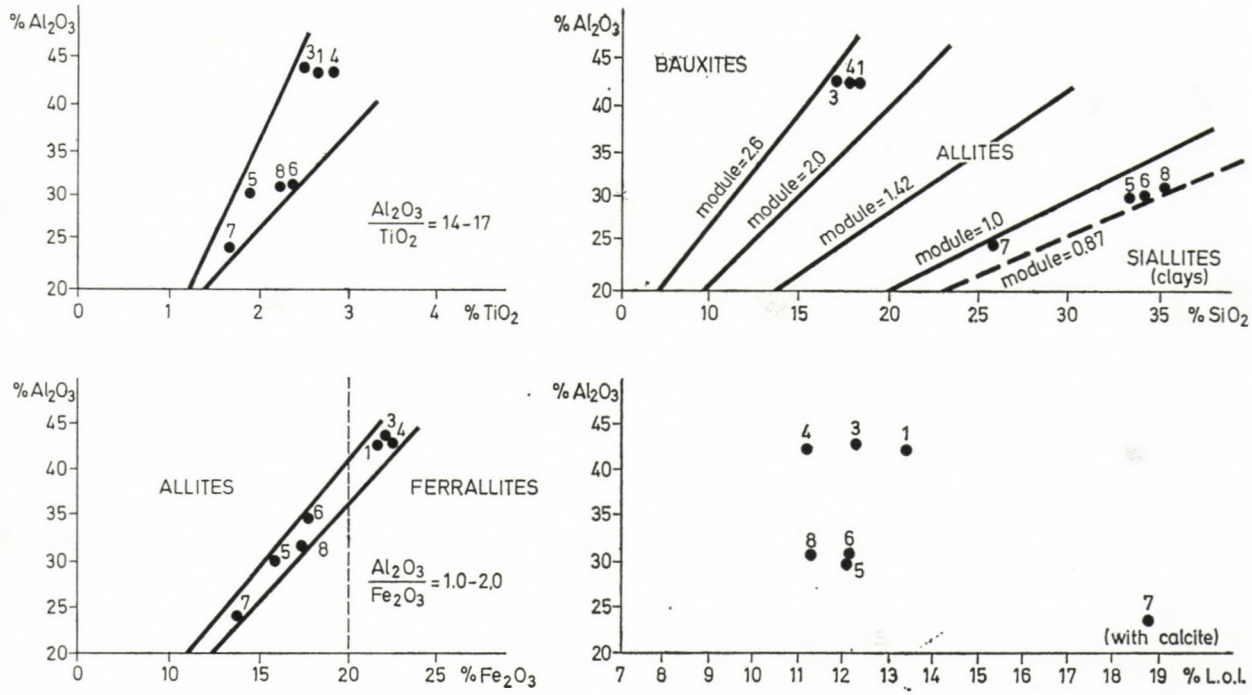

Fig. 2. **Binary diagrams of 7 allite samples (Sierra Caujerí)**

Table 4

Chemical composition of some possible

No.	Locality Type of rock w%	Al_2O_3	SiO_2	Fe_2O_3	FeO	TiO_2	MnO	MgO
(42)	La Tinta green schist	15.1	51.9	2.6	5.1	0.6	0.2	7.4
(43)	Cajobabo Carbonatized metavulcanite	11.3	36.8	3.2	3.7	0.7	0.3	4.6
(52)	Ses de Baracoa schist	18.0	49.6	2.8	6.0	1.3	0.2	6.9
(55)	Between La Tinta and Asunción amphibolite	16.1	43.3	3.0	5.0	0.1	0.2	15.5
(41)	Guira de Jauco amphibolite	15.5	52.0	2.6	4.5	1.6	0.2	6.4
(50)	Vertientes serpentinite	8.5	45.7	0.6	3.8	0.2	0.2	25.8
(5)	N of San Luis dioritic quartz porphyr	15.2	54.0	5.2	5.5	0.8	0.1	3.7
(46)	Toa Tange andesite	13.2	50.7	2.6	8.0	1.7	0.2	7.8
(49)	Vertientes altered dacite tuff	13.2	61.7	6.6	0.9	0.9	0.2	3.4
(47)	Toa Range weathered rhyolite tuff	12.3	64.1	2.8	0.5	0.2	0.05	2.7
(760)	N-Oriente gabbro	16.7	47.9	6.6	2.7	0.3	0.2	11.4
3689	Olivine gabbro	18.1	45.9	7.0	3.0	0.3	0.2	12.2
6699	Gabbro-diabase	15.9	49.3	10.9	1.9	0.1	0.1	7.5
551	Serpentinitized harzburgite	0.7	36.8	7.7	1.4	tr	0.3	40.3
5902	Serpentinitized pyroxenite	0.7	40.6	8.6	2.3	tr	0.3	41.4
756	Serpentinitized lherzolite	0.7	38.3	8.6	2.9	tr	0.2	41.4
	Cayoguán, Moa Chromite	31.0	0.2	2.6	10.4	0.06	0.03	17.4

variable in the Sierra Guaso (in the range from 7.7 to 20). This heterogeneity may be due either to differences in the source rocks, or due to a more advanced geochemical differentiation, or eventually to both.

The Sierra Guaso samples are much more ferruginous. Indeed, the majority of the samples taken in this area are ferrallites (with iron oxide contents higher than 20%). The Al_2O_3 and Fe_2O_3 contents are opposed to each other in variation (their ratio varying from 0.5 to 2.1). This is most likely due to a secondary accumulation (enrichment) of iron. On the contrary, in the Sierra Caujerí a marked positive correlation (Al_2O_3 : Fe_2O_3 1.8—2.0) could be established.

The loss on ignition values in function of the alumina content are of irregular pattern for the Sierra Caujerí. This fact can be easily interpreted in terms of a mixed mineralogical composition of the allites as mentioned above. On the contrary, in the Sierra Guaso the loss of ignition displays a rather strong positive correlation with the alumina content, in accordance with the gibbsitic-goethitic (more hydrated) composition of the samples.

source rocks in the former Province Oriente

CaO	Na_2O	K_2O	$-H_2O$	$+H_2O$	CO_2	P_2O_5	Cr_2O_3	$\frac{Al_2O_3}{SiO_2}$	$\frac{Al_2O_3}{Fe_2O_3 FeO}$	$\frac{Al_2O_3}{TiO_2}$
9.3	2.5	0.4	0.4	4.9	0.04	0.05		0.29	2.0	25.2
21.0	1.5	0.7	0.04	3.3	13.4	0.06		0.30	1.6	16.1
5.2	3.0	1.9	0.05	4.9	0.03	0.2		0.36	2.0	13.8
12.8	1.9	0.2	0.07	1.5	Ø	0.02		0.37	2.0	161.0
11.7	3.8	0.2	0.14	1.5	0.18	0.09		0.30	2.2	9.7
8.4	0.3	0.1	0.15	6.3	0.03	0.01		0.18	1.9	42.5
6.9	2.8	1.0	2.6	2.0	0.07	0.16		0.28	1.4	19
8.4	4.1	0.2	0.1	3.2	0.07	0.1		0.26	1.2	7.8
3.6	2.2	1.3	0.5	4.7	0.03	0.16		0.21	1.8	14.7
1.7	0.8	1.3	6.5	7.6	tr	0.02		0.19	3.7	61.5
12.8	1.8	0.2	0.4			0.03		0.34	1.8	55.7
14.1	1.6	0.2	0.8			tr		0.39	1.8	60.3
10.6	2.8	0.3	1.7			tr		0.32	1.2	159.0
0.2	0.4	Ø	13.8			tr		0.02	0.08	
0.6	0.9	Ø	51.1			tr		0.02	0.06	
0.4	0.6	Ø	9.0			tr		0.02	0.06	
0.02	Ø	Ø	0.10		Ø	0.06		155.0	2.4	

In Fig. 4 the samples of both areas are plotted in a triangular diagram. They are concentrated in the field of allites and ferrallites. It is obvious that there are two "branches" of geochemical differentiation: the relative enrichment of alumina (bauxitization) and the secondary accumulation of iron (ferrallitization), the latter being more accentuated in the Sierra Guaso.

4. Discussion of the Genesis

Let us discuss briefly the possible source rocks for the two areas separately.

Sierra Caujerí. The bedrock is the Yateras type limestone of Burdigalian (Early Miocene) age. Possible source rocks are the green schists and, subordinately, the serpentinitized ultramafic rocks. At present, these are separated from the karsted meseta by an erosional valley.

Sierra Guaso. The bedrocks are the Charco Redondo limestone (of Middle Eocene age) and, subordinately, the limestones of the Yateras and Maquey

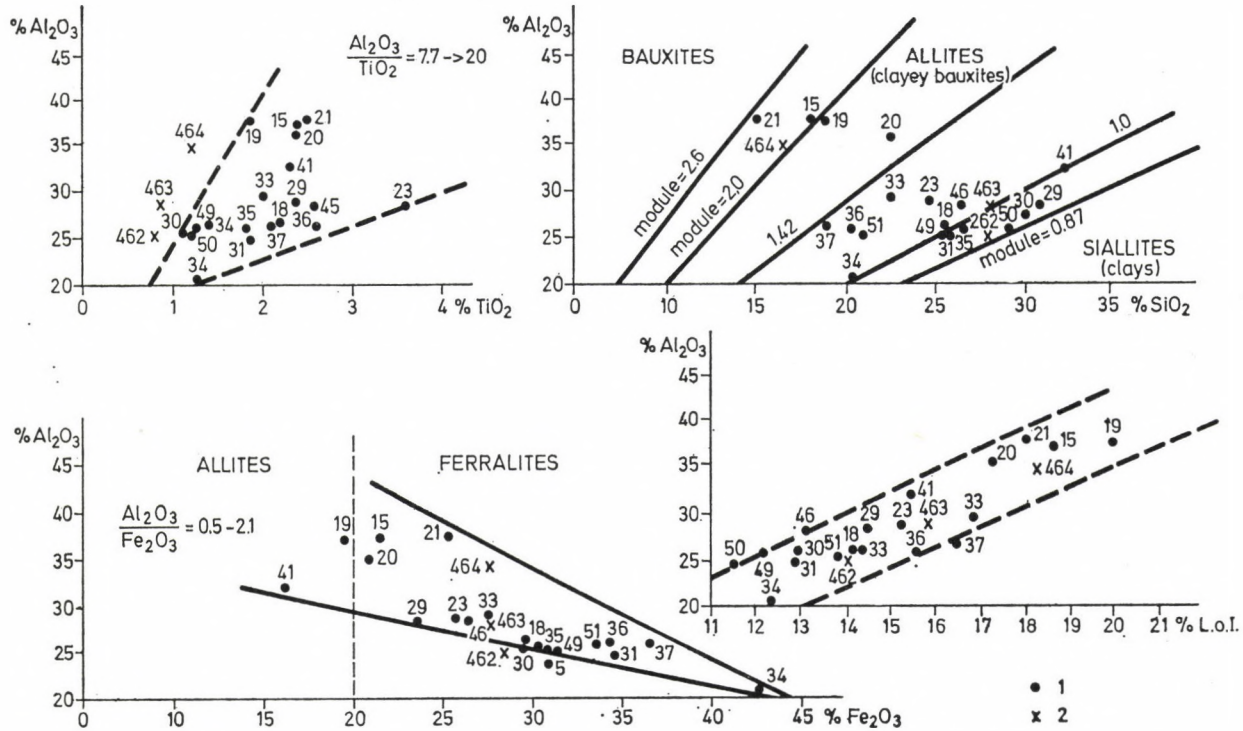

Fig. 3. Binary diagrams of 22 allite samples (Sierra Guaso). 1. NAGY et al. 1975—76; 2. BUGUELSKIY—FORMELL, 1974

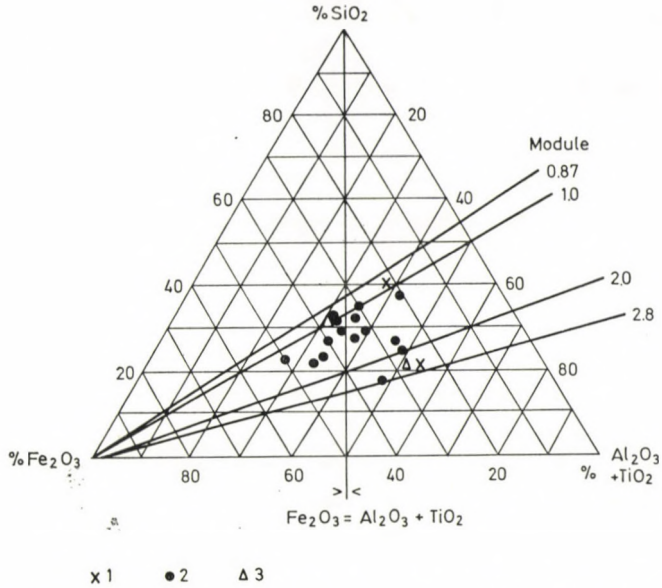

Fig. 4. Triangle diagram of 25 allite samples from the karst of Eastern Cuba. 1. Sierra Caujerí (2 dots, 3+4 samples); 2. Sierra Guaso (15 dots, 16 samples); 3. Baracoa area (2 samples)

types (of Miocene age). The possible source rocks are the various volcanic and volcano-sedimentary sequences of the Cretaceous and the Paleogene, the ultramafic rocks, and, moreover, the minomict conglomerates (consisting of mafic and ultramafic pebbles) of the Maestrichtian and the Oligo-Miocene.

Table 4 shows the chemical composition of some important igneous and metamorphized rocks of the former province of Oriente.

The geochemical process of weathering, allitization and bauxitization is complicated enough in itself. In our case it is even more sophisticated due to the intervention of transportation and deposition on one hand and to the continuation of allitization and bauxitization in the depressions of the karsted limestone depressions, on the other.

Assuming, as a working hypothesis, that the $Al_2O_3 : Fe_2O_3$ and the $Al_2O_3 : TiO_2$ rations remain essentially constant during the process, the most probable source rocks would be those the respective ratios of which are similar to those of the allitic sediments (see the two right columns of Table 4).

The contribution of the "insoluble residue" of the bedrock limestones can not be discerned, because the same igneous and metamorphic rocks had furnished the terrigenous material into the carbonatic basins in which they have been deposited.

A possible "model" of geochemical evolution has been calculated for the case of the green schists.

The chemical decomposition of the primary silicate minerals in the course of weathering under a humid, tropical climate is characterized by hydratation on one hand and by the (practically complete) leaching of the ions Mg^{++}, Ca^{++}, Na^{+}, K^{+} and by a partial removal of silica on the other (Table 5).

Similar calculations could be done for other rocks of Table 4 as well, but all this would be nothing more than theorical speculation.

For this reason, we preferred to present in Table 6 some data on three laterite profiles of Cuba (only one of them, that of La Farola in the Sierra Purial, refers to the region of study).

It can be seen that in the complete profile of Madruga the development of a latosol type soil produces the parallel enrichment of both Al_2O_3 and TiO_2 (4 to 5 times) and the complete leaching of magnesium, with the silica content remaining constant.

In the two other cases, which represent the formation of ferruginous laterites, Al_2O_3 and TiO_2 are less increased (2 to 4 times only), while the iron content is augmented 7 to 8 times.

For the formation of high-grade bauxite ores intermediary processes between these two extremes would be needed, and preferably other source rocks than serpentinites or other ultramafic rocks which are already originally rich in iron and other heavy metals.

These prerequisites, and also the other indispensable geomorphologic and climatical conditions did exist during the Neogene and Quaternary (and exist even now) in the Guantánamo region. (See the idealized sketch originally constructed by E. Szabó for the Hungarian bauxites, Fig. 5.)

The analogy with the important bauxite deposits of Jamaica also support the assumption that in the karst of Guantánamo province may exist industrial accumulations of bauxitic materials. Recalling the same analogy (Fig. 6), the highest-grade ore is situated usually at the centre of the depression filling (not at the surface).

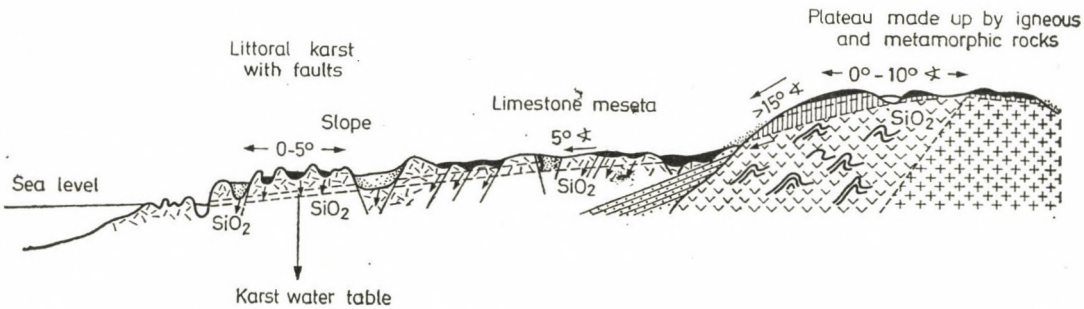

Fig. 5. **Leaching of silica as a function of topography (a theoretical sketch, after E. Szabó)**

Table 5

A theoretical example of possible transformation of green schist into allite and bauxite

w%	Al_2O_3	SiO_2	Fe_2O_3	TiO_2	L.o.I.	Al_2O_3:SiO_2 ratio	Process
Green schist	15.1	51.9	7.7	0.6	5.3	0.3	
1. phase: oxidative leaching and hydration	22.5	38.7	11.4	0.9	11.6	0.6	weathering
2. phase: oxidative leaching and hydration	30.7	26.4	15.6	1.2	16.0	1.2	allitization (formation of gibbsite
3. phase: oxidative leaching and hydration	34.2	14.7	17.4	1.3	26.7	2.3	accompanied by boehmite)
4. phase: dehydration	40.7	17.5	20.7	1.6	19.8	2.3	
5. phase leaching and dehydration:	46.1	9.9	23.4	1.8	18.1	4.7	bauxitization
Enrichment factor:	3.1	0.2	3.0	3.0	3.4		

Explication: In each of the first three successive phases 50 weight percents of SiO_2, MgO, CaO, Na_2O, K_2O are leached and the relative water content is increased by 50%. In phase No. 4 geothite and gibbsite are dehydrated by 40% and the remnants of MgO, CaO and Na_2O are removed. (See the data of Samples No. 15 and 21). In phase No. 5 50 relative % of silica is leached and the water content is reduced.

Table 6

Chemical and emission spectrographic data on three laterite profiles (by E. DUDICH, 1977)

	%									ppm						
	Al_2O_3	SiO_2	Fe_2O_3	FeO	TiO_2	MnO	MgO	CaO	Cr_2O_3	Ni	Co	V	Ga	Zn	Cu	Mo
Madruga, soil profile																
Zone A	10.1	41.6	34.7		0.20	0.4	0.7	0.3								
Zone B	9.3	37.8	41.7		0.10	0.3	0.5	0.4								
Zone C	2.6	53.2	13.0		0.06	0.2	22.4	0.2								
Serpentinite	2.2	39.4	7.4		0.05	0.1	34.9	0.2								
La Mulata, profile. Ferruginous laterite	6.4	6.2	65.6	0.5	0.10	0.7	2.0		2.1	600	400	250	16	600	160	10
Weathered serpentinite	3.8	38.7	12.8	0.9	0.02	0.3	24.6		0.6	600	60	25	40	160	60	10
Fresh serpentinite (Cajalbana type)	1.5	39.6	8.0	1.1	0.07	0.2	32.9		0.4	400	60	25	25	160	100	10
La Farola, profile. Ferruginous laterite	7.2	3.8	70.3	0.5	0.20	0.7	0.1		1.7	1000	600	400	40	1000	250	16
Serpentinite	3.1	38.1	9.7	0.3	0.10	0.2	1.7		0.2	1000	160	25	40	160	60	10

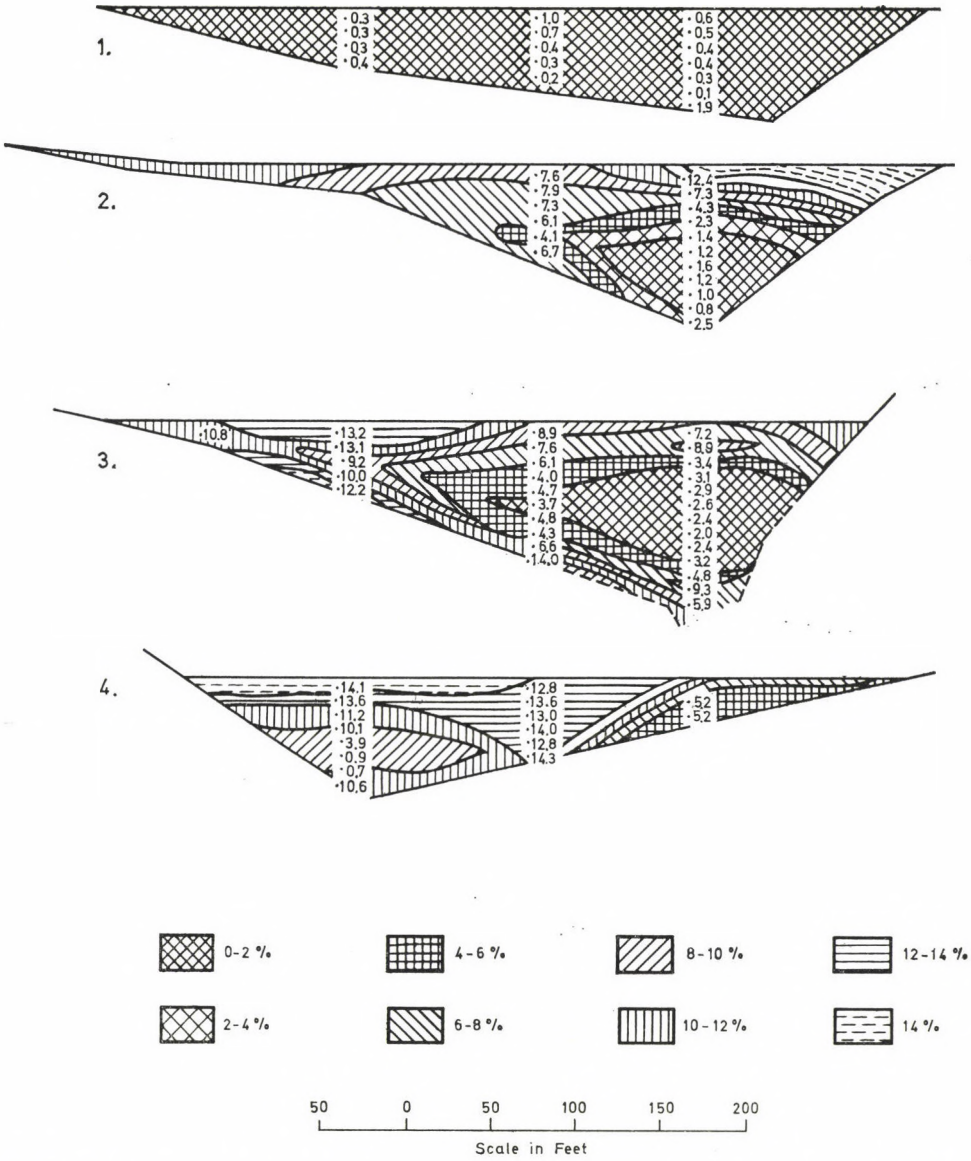

Fig. 6. Silica content isolines in four bauxite bodies of Jamaica (after U. A. ZANS, 1953). 1. Low silica (high-grade) bauxite body at Manchester (at about 2000 feet a.s.l.); 2. high-silica (low-grade) bauxite bodies at Santa Caterina (at about 450 feet a.s.l.)

5. Conclusions and Recommendations

The allitic sediments of the Sierra Caujerí seem to represent a more mature product (more dehydrated and more homogeneous) while the allites of the Sierra Guaso belong to a more immature and more heterogeneous stade, characterized also by a secondary iron enrichment.

A task of the future is to study the allites in question under microscope (in thin sections and the fractionated minerals).

A systematic sampling, extended towards the depth, may modify the ideas exposed in the present paper. The authors hope that the results will verify the presence of industrially important high-grade bauxite deposits. They recommend, once more, to carry out a regular exploration by shallow core drilling (down to 30—40 m). There is no other alternative of clearing up the perspectives of the region.

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НАКОПЛЕНИЯ БОКСИТОВЫХ ОТЛОЖЕНИЙ В ПРОВИНЦИИ ГВАНТАНАМО, КУБА

Д. П. КУТИН, Е. ДУДИЧ, Е. НАДЬ

Резюме

Главные геоморфологические типы зон накопления более-менее бокситовых отложений в областях Сьерра Гвасо и Сьерра Каухери — карстовые впадины, тектонические впадины и мезеты. По данным геологической съёмки произведённой Кубинско-венгерской съёмочной бригады КАН, аллитовые материалы залегают над известняками свита Чарко Редондо (средне-эоценного возраста), и, в меньшей мере, над известняками типа Ятерас (ранне-миоценового возраста).

Приводятся лабораторные данные о химическом и минералогическом составе 31 аллитового образца. Образцы взяты в Сьерра Каухери характеризуются чисто бемитовым составом, а в образцах из Сьерра Гвасо гиббсит преобладает над тоже присутствующим бемитом. Отмечаются различия также в распределении глинистых минералов. Вторичное обогащение железом обнаружено в образцах Сьерра Гвасо.

Проводятся химические анализы возможных материнских пород (например зелёных сланцев). Излагается модель комплексного геохимического процесса аллитизации и бокситизации и сравнивается химическими данными трёх разрезов современной латеритной коры выветривания.

Рекомендуется систематическая разведка обоих областей неглубоким керновым машинным бурением.

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REGIONAL EFFECTS ON THE DEVELOPMENT OF THEORIES ON BAUXITE GENESIS*

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The formation of bauxite deposits within the lateritic weathering crust of alumosilicate rocks is now rather well understood, though there are still some controversial details, of course. The origin of bauxite bodies situated on the more or less karsted surface of limestones and dolomites is still a matter of arduous debates.

The main theories are reviewed according to the countries in which they were formulated (see Tables 1—9). Some theories are of only historical interest, but others still have their adherents even at present. All theories but three were developed for the bauxites of France, the Dinarides and the USSR. This fact can be explained by a combination of natural (geological) and social (economic) factors interplaying in these countries. The reputation of some outstanding scientists played an important role in the particular choice of theory adopted in each country concerned. For instance, under the influence of A. DE LAPPARENT, the "terra rossa" theory was at the premium for more than a quarter of a century from the 1930 on. On the Balkan Peninsula the same theory was adopted in the 1960's upon the influence of TUĆAN, KIŠPATIĆ and DE WEISSE. In the Soviet Union A. D. ARKHANGELSKY's derivation of bauxites from chemical precipitation in aqueous solutions was the "official" theory for more than 20 years. In Hungary, E. VADÁSZ advocated the theory suggesting the bauxites to have originated from redeposition of lateritized products of weathering. His point of view dominated from 1946 to 1955, and, partly, to as late as 1966.

Since 1960 complex theories have appeared more and more frequently. An idea suggesting the necessity to develop by strictly scientific methods special genetic schemes for each particular bauxitiferous area has been brought up and has found an ever increasing number of adherents arguing that no universal theory of global validity can be recommended.

Hungarian scientist GY. BÁRDOSSY succeded in pointing out the general regularities of the major prerequisites for bauxite genesis relying on the global tectonic model. Naturally, details and peculiarities are accounted for by regional geological features.

Introduction

Hardly any other kind of ore has ever given rise to such a great variety of genetic speculations as bauxite. In fact, practically all possible geological factors have been supposed to be responsible — at least to a certain degree — for the origin of some or other bauxite deposits: hydrothermal activity, decomposition of tuffs and/or pumice, chemical alteration of practically all types of rock from basalt and granite to limestone and dolomite; aeolian and fluviatile transportation and (re)deposition, colloidal phenomena, chemical precipitation in freshwater to marine environments, biological (bacterial, vegetal and animal) activity, and even dynamometamorphism (Tables 1—9).

* (Presented at the VIII. Symposium of INHIGEO, Münster, 12th September, 1978)

Table 1

Regional and temporal distribution of theories

FRANCE		
1821	BERTHIER	aquatic transportation and redeposition of lateritic materials
1869	DAUBRET	hydrothermal origin of bauxite
1871	COQUAND	
1879	MAC GEE	laterisation due to organic acids
1881	DIEULAFAIT	aeolian transportation of lateritic products
1892	PASSARGE	universal theory of laterization
1904	DOLLFUSS	the "terra rossa" theory: the bauxites with carbonatic bedrock derive from the insoluble residue of the carbonate rocks themselves "in situ" bauxitization of marls (Ariège)
1963	COMBES	

From 1930 on, thanks to the exceptional authority of J. de Lapparent, the terra rossa theory reigned in France for a quarter of a century.

YUGOSLAVIA (Dinaric Mountains)		
1875	NEUMAYR	bauxite is the product of weathering of cosmic dust deposited on the karst-surface
1912	TUČAN and KIŚPATIĆ	detailed elaboration and divulgation of the terra rossa theory
1927	HARRASOWITZ	bauxite is a "tectonite" produced by dynamometamorphism
1931	KORMOS	the Hungarian scientist assumes, in a letter addressed to Fox, the tufogenic origin of the Istrian bauxites

To about 1960, the terra rossa theory monopolized the Balkan Peninsula.

INDIA		
1881	MALLET	the laterites are genuine sediments due to chemical precipitation
URSS		
1933	ARKHANGUELSKY A. D.	the generalized theory of chemical precipitation ("official theory" till 1950)
1935—37	ARKHANGUELSKY N. I.	secondary laterization of the materials transported onto the karst surface
1960—64	ZELENOV	modernization of the hydrothermal theory
Islets of the Pacific Ocean		
1971	TERCINIER	pumice floating on the sea gives birth to the bauxite on the coral islands.

Table 2

Application of the theories in Hungary

1821	BERTHIER	resedimentation of laterite	1946	VADÁSZ
1869—71	DAUBRET and COQUAND	hydrothermal theory	1905	SZÁDECZKY, GY. (in Transylvania)
			1948	PÁVAI VAJNA (in Transdanubia)
1881	DIEULAFAIT	aeolian transportation	1923	GYÖRGY and TELEGDI RÓTH
1904	DOLLFUSS	terra rossa theory	1933	FÖLDVÁRI and others
1931	KORMOS	weathering of tuffs	1952	GEDEON
1935	ARKHANGUELSKY N. I.	secondary laterization on the karst	1955	BARNABÁS

Table 3

Principal theories on the laterization of non-carbonate source rock

Year	Author	Region	Emphasized factor
1879	MAC GEE	France	organic acids
1889	WALTHER		nitric acid
1895	PASSARGE		universal theory
1897	LIEBRICH	Germany	
1903	DU BOIS	Surinam	sulphuric acid
1903	HOLLAND		bacterial activity
1906	MAC LAREN		capillarity
1912	COLE	Ireland	
1919	STOPNEVITCH	USSR	sulphuric acid produced by the oxidation of pyrite
1922	FERSMAN	USSR	
1927	ANSHELES	USSR	
1933	PUSTOVALOV	USSR	
1927	THIEL		bacterial activity
1937	ARKHANGUELSKY A. D.	USSR	bacterial activity
1937	YANSHIN	USSR (Siberia)	
1942	VINOGRADOV, BOJCHENKO		bacterial activity
1949—60	GORETSKY	USSR (Tula, Riasan, Perm)	bacterial activity
1955	SZÁDECZKY-KARDOSS	Hungary	oxidation (increase in the co-ordination number of Al)
1964	THEOBALD	USA (Colorado)	
1966—70	PEDRO and others		experimental studies in soil science

SZÁDECZKY-KARDOSS (Hungary) emphasized the role of oxidation in the bauxitization process in 1955. Combined theories emerge: ERHART 1962, VALETON 1965, NICOLAS 1970 in France, BUSHINSKY from 1955 in the USSR, BALKAY 1966 and BÁRDOSSY in Hungary etc.

The biggest bauxite deposits of the world are known to occur in Africa, Australia, India and Jamaica. Nevertheless, with only a few exceptions, all the theories on the origin of bauxite have been developed in three definite regions of the world: in France, in the Dinaric Mountains and in the Ural region of the USSR. Of the exceptions, two concerning the Indian laterites and the subrecent bauxites occurring on some coral islands of the Pacific, are due to Frenchmen; two principally important theories (on the geochemical nature of bauxitization and on the role played by plate tectonics) have been conceived by Hungarian scientists.

Under the general label "regional influences" (in the broad sense of the term) one can distinguish economico-social, geological and personal factors.

Table 4

Karst bauxites. Theories of redeposition of more or less laterized products of chemical weathering

Year	Author	Region	Subtype
1821	BERTHIER	France (Alpilles)	
1881	DIEULAFAIT	France	aeolian transportation
1913	LACROIX	France (Midi)	
1923	GYÖRGY	Hongrie (Halimba)	aeolian transportation
1923—27	TELEGDI-ROTH	Hongrie	aeolian transportation
1935—37	ARKHANGUELSKY I. N.	USSR	secondary laterization on the karsted surface
1937	MALIAVKIN	USSR (Tikhvin)	colloidal transportation
1940	VISHNIAKOV	USSR (Tikhvin)	transportation of detrital materials
1942	KOROTOV, STOLIAROVA	USSR (Ural)	limnic deposition
1946—57	VADÁSZ	Hungary	fluvial transportation
1948	BOLDIZSÁR	Hungary	colloidal deposition in lakes and lagoons
1951	GLADKOVSKY, SHAROVA	USSR (Asian part)	detrital transportation
1954	ZANS	Jamaica	reworking of laterized andesite tuffs
1955—66	BARNABÁS	Hungary	secondary laterization on the karst
1955—71	BUSHINSKY	USSR (Ural, Kazakhstan, Siberia)	
1955	ARONIS	Greece	
1957	BÁRDOSSY	Hungary	fluvio-colloidal transportation
1959	ROCH	France (Provence)	aeolian transportation
1962	ERHART	France	bio- and rhexistasy
1963	COMBES	France (Ariège)	"in situ" laterization of marls
1964	PAPASTAMATIOU	Greece	
1965	VALETON	France	
1966	BALKAY	Hungary	materials of different origin
1967	BÁRDOSSY, MACK	Greece	
1970	NICOLAS	France (Provence)	materials of different origin
1971	BÁRDOSSY	Hungary	materials of different origin
1974	MAKSIMOVIĆ	Dinaric Mountains	
1975	SZABÓ E.	Hungary	fluvial reworking of plateau-type laterite

1. Economico-social Factors

No wonder that the first and most widely accepted (however, strictly opposite to each other) theories were conceived and perfectioned in France, which is the "mother country" of bauxite, a country with early developed industrial civilization and ancient traditions in geological sciences.

The bauxite bearing region of the Dinaric Mountains on the Balkan Peninsula entered the scene during the last decades of the Austro-Hungarian Monarchy, another state with old traditions in the fields of geology and mining.

The scientists of the USSR presented their new ideas somewhat later, during the great boom of industrialization of the country, which involved also the systematic exploration of numerous bauxite deposits.

Table 5

Theories of hydrothermal and volcano-sedimentary origin

Year	Author	Region
1869	DAUBRÉE	France
1871	COQUAND	France (Les Baux)
1894	GOODCHILD	Brasil (non-carbonate bedrock)
1895	HAYES	USA (Alabama, Georgia, Arkansas) (non-carbonate bedrock)
1896	CROSS	USA (Colorado), non-carbonate bedrock
1896	COLA	Ireland (non-carbonate bedrock)
1897	LIEBRICH	Germany (Vogelsberg (non-carbonate bedrock)
1905, 1908	SZÁDECZKY GY.	Transylvania (Bihar Mts)
1908	LACHMANN	Transylvania (Bihar Mts)
1930	KORMOS	Istria (combined theory)
1935	LEINTZ	Sumatra
1937	ARKHANGUELSKY A. D.	USSR
1941	LOTTI	Istria
1947	PEIVE	USSR (Northern Ural)
1948	PÁVAI-VAJNA	Hungary
1960—64	ZELENOV	Kurili Islands
1968	YASHIN	USSR
1969	PAPIU, MÎNZATU	Transylvania (Pădurea Craiului)
1969	DZOTSENIDZE	USSR (Georgia), Hungary
1970—72	GLADKOVSKY	USSR (Ural)

Table 6

Theories of chemical sedimentation

Year	Author	Region
18?1	MALLET	India, Ireland (non-carbonate bedrock)
1906	WETHERELL	India (Mysore) (non-carbonate bedrock)
1917	BURTON	India, (Seoni) (non-carbonate bedrock)
1933—37	ARKHANGUELSKY A. D.	USSR (generalized theory, for all types of bedrock)
1968	KRIVTSOV	USSR (experimental base)

2. Geological Factors (the "Regional Factor" in the Narrow Sense of the Term)

In France, the knowledge of (mostly overseas) laterite profiles, the absence of laterized igneous rocks in the vicinity of large karst bauxite areas known in the South of the country (including the name-providing locality Les Baux itself), and the age-old tradition concerning the hydrothermal origin of most ore deposits resulted in the development and repeated (often vehement) confrontation of contradictory theories.

In the Dinaric Mountains, the impressive masses of Mesozoic Limestones and the (at least in those times) complete lack of paleolateritic crusts of weathering compelled the scientists to look for other explanations than straightforward laterization and redeposition.

Table 7

"Terra rossa" theory (insoluble residue of carbonate rocks)

Year	Author	Region
1904	DOLLFUSS	France
1912	TUĆAN	Dinaric Mountains
1912	KIŚPATIĆ	Dinaric Mountains
1913	PAULS	Transylvania (Bihor Mts)
1921—25	KERNER	Dinaric Mountains (Dalmatia)
1930	DE LAPPARENT	France (Midi)
1933	FÖLDVÁRI and others	Hungary
1946	DE WEISSE	the circummediterranean region
1958	BUTTERLIN	the Caribbean region
1963	HOSE	the Caribbean region
1964—72	MARIĆ	Dinaric Mountains
1967	KOMLÓSSY	Hungary
1967	SINCLAIR	Jamaica
1971	VENDEL and others	Hungary

Table 8

Tufogenic theories (decomposition of tuffs and/or tuffites)

Year	Author	Region
1931	KORMOS	Istria
1940	TELEKI, G.	Dinaric Mountains (combined theory)
1947	GOLDICH, BERGQUIST	Haiti and Dominica
1952	GEDEON	Hungary
1962	WATERMAN	Jamaica
1966	KNIAZEVA	USSR
1967	KALIGIN	USSR (generalized theory)
1969	D'ARGENIO	Italy
1972	GLADKOVSKY	USSR

Table 9

Special theories

Year	Author	Region and/or theory
1875	NEUMAYR	Dinaric Mountains (cosmic dust)
1927	HARRASOWITZ	Dinaric Mountains (dynamometamorphism)
1949	BERG	USSR (theory of plant activity)
1971	TERCINIER	Pacific Islands (pumice and scoria floating on the sea)

In the USSR, particularly in the Ural Mountains region, Paleozoic bauxite deposits grade laterally into marine sedimentary rocks of mostly chemical origin. Moreover, areas have been discovered where karst type bauxite deposits are in direct (tectonic) contact with igneous rocks topped with ancient weathering crusts of lateritic type.

These may have been the main reasons of the regionally controlled diversity of the *birth* of theories on the origin and formation of bauxite deposits.

Let us review briefly what was the situation in Hungary. As a matter of fact, the geological setting of the Hungarian bauxite deposits is rather similar to that of Southern France. Most of the important deposits were explored after World War II with the scientific and technical assistance of Soviet experts. The geographic vicinity of the bauxite bearing areas of Yugoslavia favoured the adoption of genetic explanations developed there, in spite of some evident differences such as the different number of bauxitic horizons in the geologic column (three in Hungary in contrast to eight or ten in Yugoslavia) and the (as a rule) much wider sedimentation gaps in Hungary.

In consequence, all principal theories have found supporters in Hungary. The only exception was the concept of chemical precipitation, which was too obviously in contradiction to the available geological evidence. (Table 2)

In the time span ranging from the late thirties to about 1955 (which might be called the second stage of development of bauxite genetic theories) regionally, or even locally founded theories were applied to bauxite deposits of different types. Three of them — the theory of lateritic derivation, the "terra rossa" theory and that of chemical precipitation — made a run for becoming a universal interpretation valid for practically all karst bauxite deposits of the world.

At this point the personal factor enters the scene.

3. Personal Factors

It was nothing else than the personal authority of certain great scientists that brought about the predominance of some or other of these theories in the individual countries.

This was the case with J. De Lapparent in France, F. Tućan, F. Kišpatić and the Swiss geologist J.-G. de Weisse in Yugoslavia, A. D. Arkhanguelsky in the USSR and E. Vadász in Hungary.

In this way, at the same time (i.e. in the early fifties) the "terra rossa" theory prevailed in France and Yugoslavia, the theory of chemical precipitation in the Soviet Union and the derivation of karst bauxites from laterites in Hungary.*

4. The Stage of Comparative Analysis and Synthesis

From the late fifties onwards there is a marked trend to abandon the attitude of an overall and hasty generalization of any genetic theory. "Com-

* In 1955, an essentially new approach was made by E. Szádeczky-Kardoss, who pointed out the strongly oxidative nature of the geochemical process of bauxitization.

plex" or "combined" theories, more or less successful "genetic compromises" succeed each other in being forwarded. Nowadays, it is generally admitted that each particular bauxite deposit requires its own genetic explanation based on a multilateral investigation of the peculiar geological features of the given area.

Mutual visits of bauxite experts, comparative laboratory work, and — last but not least — the meetings of the ICSOBA (International Commission on the Study of Bauxite, Alumina and Aluminium), 1963 Zagreb, Yugoslavia, 1969 Budapest, Hungary, 1971 Leoben, Austria, 1973 Nice, France, 1977 Kingston, Jamaica, 1978 Athens, Greece, — contributed substantially to the international (and interdisciplinary) exchange of experience.

On this basis, Gy. Bárdossy, Hungary, succeeded in developing a global explanation for the temporal and spatial distribution of the bauxite deposits of the world, relying upon the concept of plate tectonics [1].

Even this idea, however, concerns only the most fundamental prerequisites of bauxite genesis. It admits that in various regions different geological factors can play the decisive role in the intricate process of bauxite formation and deposition.

Thus, seemingly contradictory theories may be in vigour for the interpretation of the origin of bauxite deposits in geologically different regions.

In the personal opinion of the present writer, this analytic-synthetic approach (the third stage of the development of the theories on bauxite genesis) — is not only justified, but it is also a due tribute to the creative mindes of those numerous scientists, who did their best to solve the puzzle of bauxite genesis. Moreover, it represents an essentially new stage as to the right appreciation of the regional influences, now almost exclusively of geological nature, avoiding both the exaggeration and the underestimation of their significance.

Conclusion

The problem of bauxite genesis is by far not only a question of purely theoretical speculation. On the contrary, it has a very practical aspect, providing a solid scientific basis for the prognostics (forecasting), prospecting and exploration of commercially important bauxite deposits.

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РЕГИОНАЛЬНЫЕ ВЛИЯНИЯ НА РАЗВИТИЕ ТЕОРИЙ О ГЕНЕЗИСЕ БОКСИТОВ

Э. ДУДИЧ

Резюме

Образование бокситовых залежей встречающихся в профилях латеритной коры выветривания алюмосиликатных горных пород достаточно хорошо изъяснено, хотя споримые детали конечно есть. Однако о происхождении бокситовых тел, залегающих на более-менее закаростованной поверхности известняков и доломитов, ведутся острые споры.

Даётся хронологический обзор главных теорий по странам их возникновения. (См. таблицы 1—9.) Некоторые теории имеют только исторический интерес, а другие имеют и в настоящее время своих сторонников. За тремя исключениями, все были разработаны по поводу бокситов Франции, Дынарид и СССР. Этот факт обьясняется содействием природных (геологических) и общественных (экономических) факторов в этих трёх регионах. Авторитет некоторых крупных учёных играл важную роль в том, что та или иная теория стала почти общепринятой в отдельных странах. Например, во Франции с 1930 года, из-за влияния А. де-Лаппарана теория «терра росса» царствовала более четверти века. На Балканском полуострове из-за авторитета Тучана, Кишпатича и Дэ Вейсе та же самая теория считать общепринятой до 60-ых годов. В СССР, теория А. Д. Архангельского о происхождении бокситов путём химического выпадения из водных растворов свыше 20 лет была «официальной» теорией бокситообразования. Во Венгрии, Проф. Э. Вадас был сторонником теории переотложения латеритзованных продуктов выветривания. Его мнение господствовало с 1946 по 1955, и отчасти даже по 1966 год.

С 1960 года все чаще появляются комплексные теории. Возникает и крепнет концепция, по которой приходится разра атывать, строго научными методами, генетические схемы для отдельных бокситоносных областей, и нельзя предлагать общую, как будто универсальную теорию.

Венгерскому учёному Дь. Бардошши удалось установить общие закономерности важнейших условий и предпосылок бокситообразования при помощи глобальной тектоники. Подробности и особенности объясняются региональными чертами геологического развития.

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CONTRIBUTION TO THE STUDY OF FOSSIL CORAL FACIES DEVELOPMENT IN EASTERN CUBA

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Coral communities of Eastern Cuba are found spread in the superimposed basins related to the three main structural complexes of the region showing rather an areal zonation: the younger formations are placed near the actual shore line while the older are located in the neighbourhood of ancient anticlinoria.

Coral species from lagoonal facies abound in the whole area, especially in the Oligomiocene sediments of the Guantánamo-Basin.

The oldest assemblages, correlated to the Antiguan and Isthmian faunas are present in the Oligomiocene sediments of the Guantánamo Basin, and to a lesser extent in those of the Bahia de Nipe-Guacanayabo Basin. This fauna displays an Indopacific character and grew primarily on a gravel bottom in a rather low-energy environment. Warmer temperatures seem to have favoured the coral development not as true reefs but as isolated rather packed biostromes.

An abrupt change took place with the beginning of the Middle Miocene probably by the existence of an interoceanic connection and the subsequent change in the littoral thermal regime. Optimal conditions reappeared during the Upper Pliocene.

The Pleistocene coral areal zonation resembles to some extent to the pattern described by DE BUISSNJÉ for the Netherlands Antilles but is less well outlined.

Fifty percent of the fossil species listed occur in the younger reefs and lagoons. The important climatic changes that have taken place on higher latitudes during Pleistocene seem to have not been reflected in the coral growth of the study area.

Introduction

During the last years of the past century, A. AGASSIZ and others visited the island of Cuba and made some geological observations of its littoral belt, namely of its eastern portion.

A. AGASSIZ devoted special attention to the coralline formations found in the socle of the elevated terraces.

Later, T. W. VAUGHAN studied the fossil coral faunas of the island, in particular those of Eastern Cuba, drawing some conclusions on their development and relationship with the geological history of the Caribbean-Gulf of Mexico area.

These investigations remained in fact interrupted up to the early seventies when H. H. KÜHLMANN studied the recent coral fauna of our country pointing out some observations on Pleistocene coral development in connection with glaciations.

During years 1966 and 1969, G. L. FRANCO performed some geological work on the Upper Tertiary and Quaternary deposits of Eastern Cuba collecting a rich fossil fauna in which figured coral specimens, although he has not then made a thorough study of these corals.

In connection with the geological surveying the ancient Oriente province in 1972 made by the Institute of Geology and Paleontology of the Cuban Academy of Sciences, particular attention was devoted to the investigations with fossil corals, specially those of Guantanamo basin.

The authors, who belonged to the Cuban-Hungarian surveying group, present here the preliminary results of their observations in Eastern Cuba.

Distribution

The fossil species were collected in the sediments of the superimposed basins fringing the three main structural units of Eastern Cuba, namely:

1. The Auras anticlinorium.
2. The Sierra de Nipe-Cristal-Baracoa zone forming a structural complex consisting of Cretaceous and Early Paleogene rocks. It is possible that the Sierra del Purial metamorphites included in it could be older.
3. The Cayman zone containing the Sierra Maestra anticlinorium.

The depositional areas are: the Guantanamo basin (with the secondary basins of Bayate and Esperanza), the Bahia de Nipe-Guacanayabo basin (with an approximate alignment NE-SW, dividing the eastern region in two inequal parts, the oriental part bigger being, the small Santiago basin (located between the Sierra Maestra anticlinorium and the Bartlett Trough), and finally the northern and southern littoral transgressive fringes.

Younger coralline formations corresponding to reefal and lagoonal assemblages appear in both transgressive fringes. Accordingly, they are located geographically farther from the anticlinoria axes than the older formations arranged in this manner according to certain areal zonation. Their fauna includes 26 species of living corals either very well preserved in the Upper Pleistocene Jaimanitas Formation and in some Holocene beach deposits or more or less recrystallized in older sediments (Upper Pliocene Maya and Guardarraya formations). Fifty per cent of the enlisted corals occur in these younger formations.

The Upper Neogene La Cruz Formation of the Santiago basin contains 15 species which from 10 are living. It is surpassed only by Maquey Formation (with 38 species) and Jaimanitas Formation (with 26 species).

Historical Evolution

Coral development in Eastern Cuba exhibits two well definite acmes: one less evoluted during the Aquitanian s.l. (*i.e.* Oligomiocene) and a second from the Upper Pliocene to Recent.

During Paleocene and Eocene times the coral development has not left remarkable clues in this part of our territory and probably in the whole insular area. In the Caribbean-Antillean region coralline sediments appears restricted to a few places: Mexico, British Honduras, Jamaica and the Gulf Coast (FROST, 1972; VAUGHAN, 1919).

In the Middle Eocene faunal assemblages prevail, according to Frost, Tethian genera.

During the Aquitanian s.l. (*i.e.* the interval Upper Oligocene—Lower Miocene, according to us and others) the corals reach in the Guantanamo basin and the middle and northern portion of the Bahia de Nipe-Guacanayabo basin (Clindro, Maquey and Yateras formations for the first; Camazan Formation for the latter) a stronger development, with a relatively great number of species and wide spatial distribution without forming true reefs but rather isolated biostromes more or less closely arranged. These differ from those observed by VAUGHAN (1919) in other places of the Caribbean-Gulf of Mexico area.

VAUGHAN mentioned, moreover, the possibility of an interoceanic connection during the Oligocene based upon the continuity of both the Culebra and Emperador formations in the isthmian zone from the Atlantic coast up to the Pacific slope and also upon the faunal similarity of the Pacific and the Atlantic coast of Panama. This connection has been supported too by A. TORRE (verbal communication) but he locates it in the Middle Miocene. In this case the entrance of a cold current could have modified the thermal regime then existing in the Caribbean-Gulf area, inhibiting coral development.

FROST explanation (1972) considers the close of this connection at the end of the Oligocene, not allowing the entrance of the East Pacific Equatorial Countercurrent.

It is interesting that the influence of Tethian genera fails in the Aquitanian s.l.: then appear many genera nowadays restricted to the Indopacific and the Mediterranean. This fact can be established almost uniformly in the whole Caribbean-Gulf area and especially in the coral fauna of Antigua Formation, in which hermatypic forms prevail.

During the Middle-Upper Miocene an abrupt faunal change occurs but scarce Indopacific and Mediterranean faunal elements persist for some time.

In spite of the relative abundance of hermatypic corals, no true reefal development can be observed in the area of study. Some living coral species appear (*Montastrea annularis*, *Acropora cervicornis*, *Porites porites*, *Solena-*

Table 1

Age	Formations / Species	*Acropora panamensis*	*A. sabudensis*	*Agathiphyllia anguillensis*	*A. antiguensis* (1)	*A. splendens* (1)	*A. tenuis* (1)	*Antiguastrea cellulosa*	*A. cellulosa silecensis*	*Astrocoenia decaturensis*	*A. guantanamensis*	*A. meinsori*	*Diploastrea crassolamellata*	*Diploria antiguensis*	*Goniopora decaturensis*	*Manicina willoubiensis*	*Montastrea costata*	*M. imperatori*	*Pironastrea antiguensis*	*Pocillopora arnoldi*	*P. guantanamensis*	*Porites panamensis*
Q_4	Recent																					
Q_3	Jaimanitas																					
N_2-Q_1	Maya, Guardarraya																					
N_2	Baracoa																					
N_1^3	Jucaro																					
N_1^2a-?N_1^3	La Cruz																			...		
N_1^2a-N_2^3	Manzanillo																					
N_1^2a	Cabo Cruz																					
N_1^2	Jagüeyes																					
N_1^1	Camazán		+					+	+		+		+				+		+			+
P_{g3}-N_1^1	Bitiri (?)		+								+											
P_{g3}-N_1^1	Yateras, Maquey	+	+	○	×	+	○	×		○	○	○	○	+	○	+	+	×	○		×	
P_{g3}-N_1^1	Cilindro	+	+					+			+		+							+		
		I. OLIGOCENE-LOWER MIOCENE																				

[1] = *Cyathomorpha in* VAUGHAN (1919).

strea bournoni, etc.). The number of species is considerably minor than in Aquitanian. In general the coral development diminishes assume a restricted and irregular character.

From the Upper Pliocene onwards the reefal facies begins to appear in the Cuban seas, reaching probably its major development from the Upper Pleistocene up to date.

The neotectonic activity in the Caribbean-Gulf area, which left remarkable clues in the Cuban territory (*e.g.* the terraces *en échelon* of Maisi and Cape Cruz), appears to have controlled the erosional rhythm, and therefore the amount of detritic materials transported to the surrounding seas, affecting rather periodically the ecological *optimum* for coral growth.

	Siderastrea conferta	*Stylophora canalis*	*Trochoseris meinseri*	*Dichocoenia stokesi*	*D. tuberosa*	*Diploria labyrinthiformis*	*Goniopora jacobiana*	*Manicina areolata*	*Montrastrea annularis*	*M. cavernosa*	*M. limbata*	*Pocillopora baracoaensis*	*Porites asteroides*	*P. baracoaensis*	*P. furcata*	*P. porites*	*Siderastrea siderea*	*Solenastrea bournoni*	*S. hyades*	*Stephanocoenia michelini*	*Stylophora affinis*	*S. granulata*	*Thysanus excentricus var.*	*Acropora cervicornis*	*A. palmata*	*Agaricia agaricites*	*Cladocora arbuscula*	*Diploria clivosa*	*D. strigosa*	*Favia fragum*	*Meandrina meandrites*	*Mycetophyllia lamarckiana*	*Siderastrea radians*
				○		○		○	○	○			○		○		○	○	○	○				○	○	○	○	○	○	○	○	○
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II. MIDDLE-UPPER MIOCENE AND PLIOCENE | III. QUATERNARY

Symbology

+ collected by the group
○ referred to in literature
× collected by the group and referred to in literature

Ecological Considerations

Coral assemblages found from the Aquitanian s.l. to Miocene are chiefly represented by species corresponding to lagoonal facies.

On the thick layers of coarse detritus corresponding to the beginning of molassic deposition in the Guantanamo basin (Cilindrol Formation) can be observed an early coral development often observed as incrusting forms *(e.g. Astrocoenia guantanamensis)* or nodular forms *(Stylophora canalis)*.

Major development is observed in the basal part of Maquey Formation (Vega Grande Member) gradually decreasing upwards with the beginning of Middle Miocene deposition.

These assemblages are ecologically characterized by their malacological fauna and by the abundance of large specimens of *Lepidocyclina* (*L. gigas* group), referable to the *Corallinacea-Lepidocyclina* community of BÁLDI (1973). The genera included in this assemblage are widespread in their majority not only in the study area but in the Caribbean-Gulf area within the same stratigraphic range.

The relatively low energy of the lagoonal environment hindered probably the continued suspension of finer detritals, important constituent of the Guantanamo basin sediments allowing enough the aeration, food circulation and light penetration needed for survival of coral.

In the Bahia de Nipe-Guacanayabo basin, coral development was not so important as in the Guantanamo basin. Nevertheless it exhibited relative abundance of fossil species in some places: *e.g.* the sediments Gamazán Formation, corresponding in a greater extent to the Aquitanian *s.l.* (Only two of the enlisted species existing in Gamazán Formation have not been found in the Guantanamo basin deposits.) It looks reasonable to suppose some peculiar microclimatic conditions in particular the thermal regime of the sea bottom to explain this relative inhibition.

Upper Pliocene coral growth marks the true reefal development in the area with the predominance of typical forms of a high energy environment like *Acropora palmata* found sometimes in growing position in the barrier reef zone. Although coral growth is also remarkable in the lagoon with forms like *Montastrea annularis*.

Hard percussion-sounding whitish reefal limestone belonging to this interval appear today forming the socle of the elevated terraces of the southern coast of the ancient Oriente Province. They are included in the Maya Formation.

The Upper Pleistocene coral assemblages present sometimes a not always well outlined areal zonation that keep resemblance to the "biozones" (subfacies) described by DE BUISONJÉ (1974) in the Netherlands Antilles, namely: barrier-reef zone (*Acropora palmata* zone) representing a turbulent high energy environment; the *Montastrea annularis* zone representing the lagoonal conditions and finally the *Siderastrea* zona corresponding to the inner border of the littoral ensemble.

The abrupt climatic changes that have taken places in the higher latitudes during the Pleistocene appears to have been not reflected, at least remarkably, in the coral growth of the study area although the existing gap between the Maya and Jaimanitas formations could offer us the key of an important inhibition in coral growth during the Early and Middle Pleistocene.

Conclusions

1. The coral development in the Caribbean-Gulf area exhibits two acmes: one during the Aquitanian s.l. and another from the Upper Pliocene up to the present, probably due to changes in the thermal regime of the seabottom.

2. True reefal growth can be observed only from the Upper Pliocene upwards.

3. Younger and older coralline formations are distributed rather zonally: younger coral fringes are farther from ancient structures than the older formations.

4. In the Oligomiocene Guantanamo-Basin sediments the lagoonal corals show a strong influence. They form no reefs but isolated more or less closely arranged biostromes.

5. There are no clear clues in the study area reflecting the influence of the abrupt climatic changes during the Pleistocene.

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К ВОПРОСУ ИССЛЕДОВАНИЙ РАЗВИТИЯ ИСКОПАЕМЫХ КОРАЛЛОВ В ВОСТОЧНОЙ ЧАСТИ КУБЫ

Г. Л. ФРАНКО—Э. НАДЬ—ДЬ. РАДОЦ

Резюме

Сообщества кораллов Восточной Кубы распространены в наложенных бассейнах, связанных с тремя главными структурными комплексами рассматриваемой области. Они обнаруживают скорее;площадную зональность: более молодые образования расположены вблизи современной береговой линии, в то время как более древние развиты около древних антиклинориев.

На территории рассматриваемого района изобилуют виды кораллов, приуроченные к лагунным фациям, это особенно справедливо для олиго-миоценовых отложений Гвантанамского бассейна.

Самые древные комплексы, сопоставимые с фаунами Антигвии и Истмии, встречаются в олиго-миоценовых отложениях Гвантанамского бассейна, в меньшей мере — в отложениях такого же возраста, развитых на территории Бахияде—Нипе—Гваканаябо. Эта фауна имеет индо-пацифический характер и она проживала в преимущественно галечниковых донных условиях в водной среде пониженного энерегетического режма.

Более высокие температуры благоприятовали, пов-видимому, развитию кораллов не в виде подлинных рифов, а скорее в форме изолированных, но более-менее густых биостром.

Резкое изменение произошло с началом среднего миоцена, что было обусловлено, по-видимому, существованием межокеанического сообщения и последующим изменением термального режима прибрежной зоны. Оптимальные условия восстановились в позднеплиоценовое время.

Площадная зональность коралловых сообществ плейстоцена напоминает в некоторой степени условия, описанные Де-Байсонье для Нидерляндских Антилл, но она менее четко выражена.

Пятьдесят процентов перечисленных ископаемых видов кораллов встречается в более молодых рифах и лагунах.

Значительные климатические изменения, имевшие место на более высоких щиротах в плейстоценовое время, по-видимому, не сказывались на произрастании кораллов в пределах рассматриваемого района.

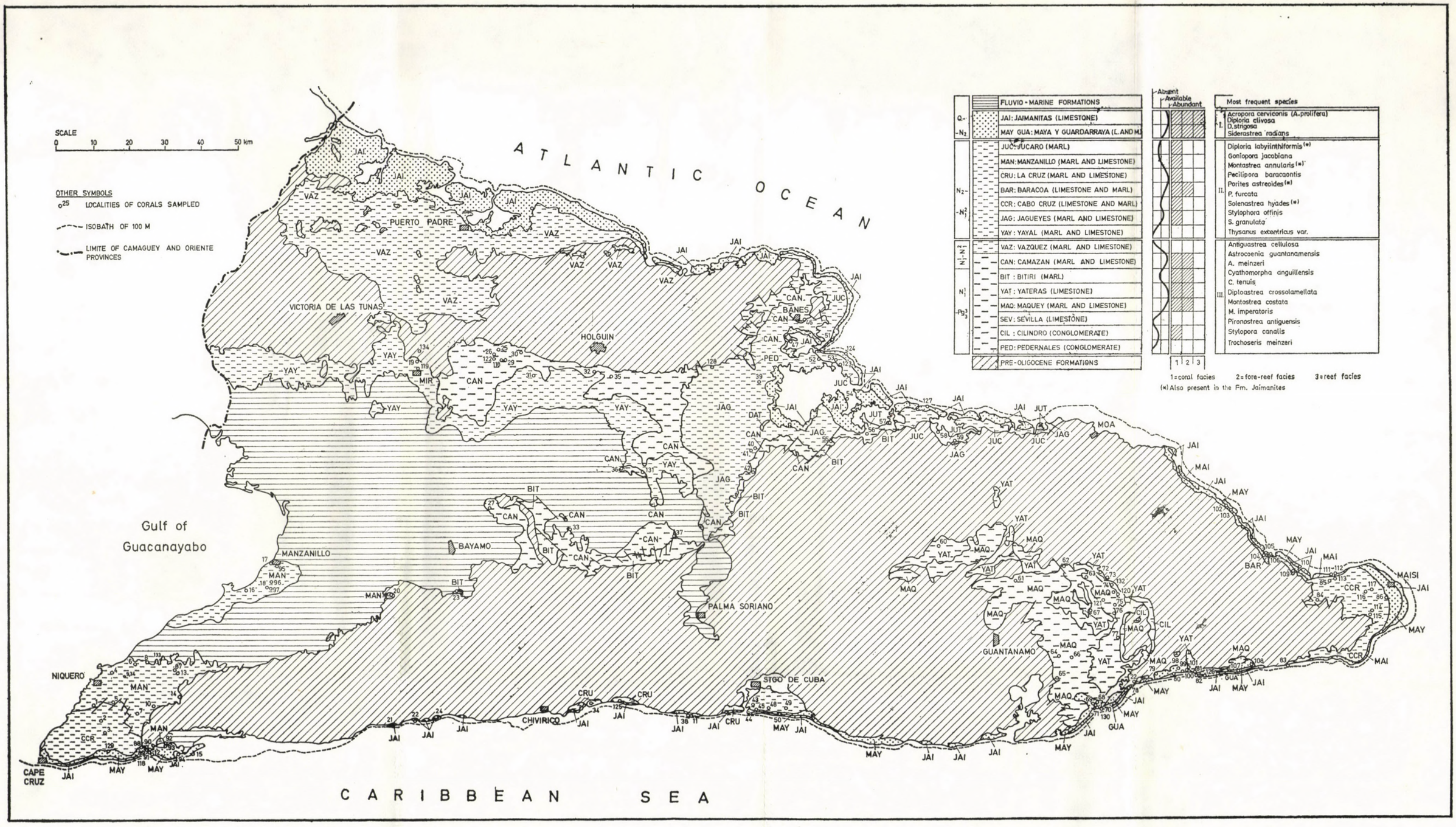

Fig. 1. Distribution of the coralline formations in Eastern Cuba. Compiled by: G. L. Franco—E. Nagy and Gy. Radócz. Relying on results of the Cuban-Hungarian team, Geological-Paleontological Inst. of the Academy of Sciences, Cuba

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TRANSITIONAL VERY LOW- AND LOW-GRADE REGIONAL METAMORPHISM OF THE PALEOZOIC FORMATIONS, UPPONY MOUNTAINS, NE-HUNGARY: MINERAL ASSEMBLAGES, ILLITE-CRYSTALLINITY, $-b_0$ AND VITRINITE REFLECTANCE DATA

By

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The Middle Devonian—Middle Carboniferous formations of the Uppony Mountains lying at the northern border of the Igal-Bükk Alpine mobile belt form a monoclinal folded structure. The sedimentation is characterized by the gradual decrease of the proportion of the calcareous sediments to their total disappearence as well as the parallel increase of that of the detrital and chemical siliceous sediments. The volcanic activity established in several stratigraphic horizons that had produced mostly basaltic lavas and pyroclastics can be explained by the deep-faults connected with the subsidence of the sedimentary basin.

On the basis of the mineral assemblages, petrotextural characteristics, illite-crystallinity, vitrinite reflectance data, chloritoid formation, glauconite-celadonite/stilpnomelane isograde and anthracite-graphite "co-existence" the regional metamorphism of the Paleozoic of the Uppony Mountains occured at a temperature of about 300 °C. This corresponds to the boundary between the anchi- and epi-zones. The b_0 lattice parameter distribution of sericites ($\bar{x} = 8.996$ Å, $n = 34$, $s = 0.010$) indicates low-pressure recrystallization, i.e. of about 2.5 kilobar. There is no significant difference between the metamorphic grades of the Middle Devonian and Middle Carboniferous formations. This, as well as the continuous sedimentation conditions, further the lack of angular unconformity rule out the possibility to assume effects of the Hercynian (Bretonian or Sudetan) tectonometamorphic phases. On the basis of the metamorphic grade of other Paleozoic and Mesozoic formations of the area the Alpine (Cretaceous) age of the regional dynamothermal metamorphism can be assumed.

Introduction

The Paleozoic formations of the Szendrő and Uppony Mountains lying in the northeastern part of Hungary are superficial parts of the anchi-epimetamorphic belt extending from the Graz Basin in the Eastern Alps through the Little Plain, Balaton Highland up to the Spissko-Gemerské Rudohorie (Western Carpathians) (Szádeczky-Kardoss et al., 1969, 1976; Jantsky, 1976; Szepesházy, 1973). The difference in the tectonic behaviour of the parts of the belt is indicated by the difference in grades of transformation of the Mesozoic formations within the Alpine cycle (Árkai and Viczián, 1975). In the southern part of the Paleozoic anchi-epimetamorphic region, the Alpine

mobile belt lying in NE-SW direction (the "Igal-Bükk eugeosyncline", WEIN, 1969) and bordered by the so-called Darnó-line shows the most complete sequence in the Bükk Mountains with the remnants of "oceanic basement-like ophiolitic" magmatites (SZÁDECZKY-KARDOSS, 1978; ONOUHA, 1978). Southeast of the Darno Tectonic Zone, the northwestern border of the Alpine mobile belt is constituted by the Devonian (Silurian?) and Middle Devonian-

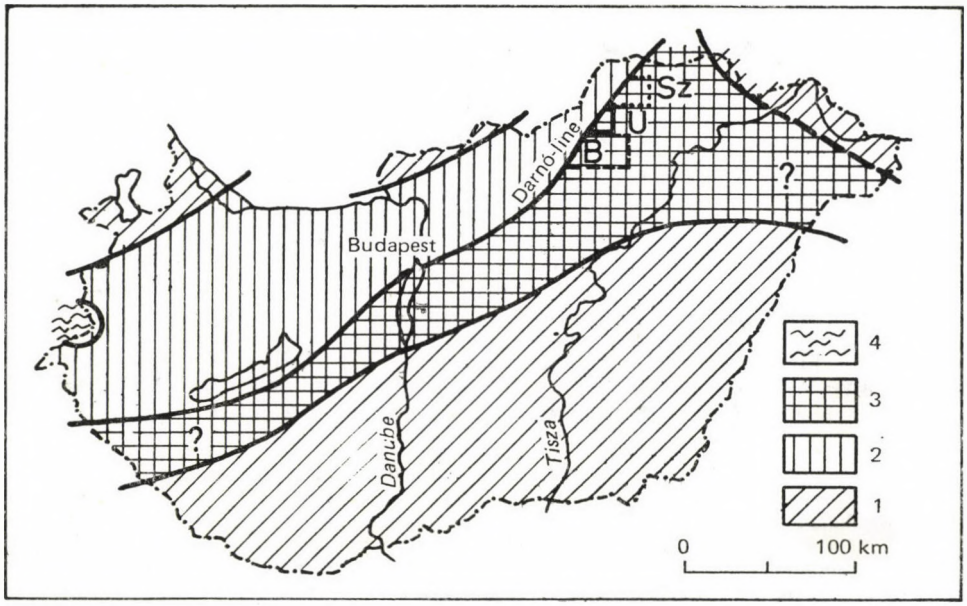

Fig. 1. Geological situation of the Uppony, Szendrő and Bükk Mountains. 1. Medium- and low-grade polymetamorphic (Early Baikalian—Hercynian—Alpine) crystalline basement; 2. very low- and low-grade, generally Hercynian metamorphic Paleozoic basement, with non-metamorphic Mesozoic; 3. very-low- and low-grade metamorphic Paleozoic, with very low-grade metamorphic Mesozoic (Alpine mobile belt); 4. Penninicum of the Kőszeg Mountains; SZ = Szendrő Mountains. U = Uppony Mountains; B = Bükk Mountains

Middle Carboniferous formations of the Szendrő resp. Uppony Mountains (Fig. 1). Based on the petrological features of these metamorphic formations the changes of intensity of the different tectono-metamorphic cycles can be determined in relation to the tectonic evolution. This is aimed by the investigation of the stratotypes of the Szendrő, Uppony and Bükk Mountains under the guidance of FÜLÖP (1976) and sponsored by the Central Geological Office of Hungary.

The aim of this paper is to characterize the Paleozoic of the Uppony Mountains from the metamorphic petrological point of view and to compare it with the data on the Szendrő Mountains published earlier (ÁRKAI, 1977).

Geological Setting

The geological sketch map (Fig. 2) was prepared after BALOGH (1964), demonstrating also the lithostratigraphic units distinguished by FÜLÖP (1976).

The Paleozoic of the Uppony Mountains forms a monoclinal folded structure of southeastern dip. Its northwestern and southeastern borders are tectonic. In the NW, in the Darnó Tectonic Zone the Paleozoic sequence was overthrusted to Triassic and Oligocene-Miocene formations. In the SE it contacts the overlying Triassic rocks also along overthrust planes (PANTÓ, 1954; BALOGH, 1964).

The Paleozoic sequence of Uppony is divided into three formations. The oldest one is the 400 m thick light-grey compact crystalline limestone

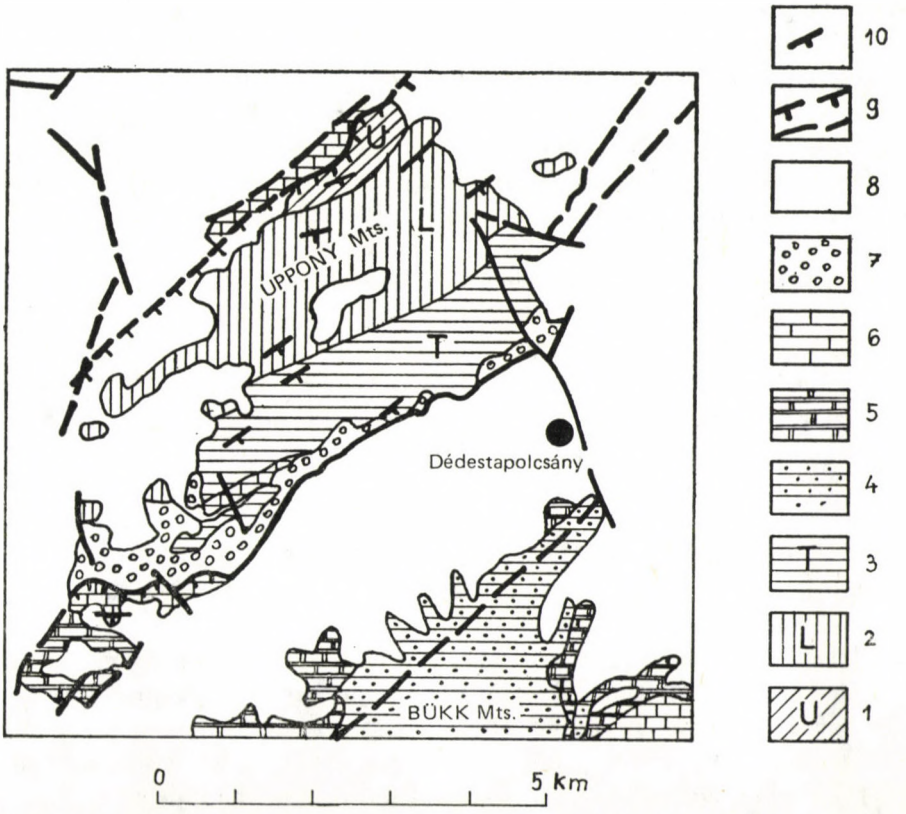

Fig. 2. Geological sketch of the Uppony Mountains (after BALOGH, 1964), simplified. 1. Uppony Limestone Formation; 2. Lázbérc Formation; 3. Tapolcsány Formation; 4. Middle and Upper Carboniferous slate, sandstone with limestone lenses; 5. Permian variegated slate and sandstone (Lower Permian) and bituminous limestone (Upper Permian); 6. Triassic: Anisian dolomite (metasomatic iron ore), limestone; Ladinian limestone, diabase and tuff, dolomite; 7. Cretaceous: Senonian limestone conglomerate of Gosau type; 8. Tertiary sedimentary rocks; 9. overthrust, fault; 10. strike and dip

sequence of the Uppony Limestone Formation (I). This is overlain by the Lázbérc (II) and Tapolcsány (III) Formations of about 2000 m total thickness. The Lázbérc Formation consists of the alternation of thin-banked lamellar limestone, lime-schist and slate with subordinate intercalation of meta-sandstone and meta-tuffitic limestone. The Tapolcsány Formation is characterized by the repeated alternation of slate, meta-sandstone, meta-chert as well as of metabasalt and -tuff (BALOGH and KÖRÖSSY, 1974). Along the Pre-Senonian overthrusting zones effects of Fe-Mg-carbonatic metasomatism can be observed (PANTÓ, 1954). In the slate and meta-chert strata of the Tapolcsány Formation traces of sulfidic ore (pyrite-chalcopyrite) are common, in the weathering zone with oxi-hydroxide, sulfate-phosphate mineralisations (ELSOLTZ and SELMECZI, 1977).

The age of the Paleozoic sequence poor in fauna has been uncertain. SCHRÉTER (1954) assigned the Formation I. to the Devonian, the Formations II. and III. to the Lower and Upper Carboniferous resp. Assuming a single sedimentary cycle PANTÓ (1954) and BALOGH (1964) dated the whole sequence as Lower Carboniferous. On the basis of uncertain Hystrichosphaerides and Graptolites remnants the Paleozoic sequence of Uppony was assigned to the Ordovician and Silurian by NAGY and SZEPESHÁZY (1971) and later by BALOGH and KÖRÖSSY (1974) and correlated with the two lower formations of the Szendrő Mountains though the correlation of the two mountains was ruled out by JÁMBOR (1961) relying upon petrological evidence.

According to the latest Conodonta investigations Formation I. is of Middle Devonian, Formation II. of Upper Devonian—Lower Carboniferous, Formation III. of Middle Carboniferous age (KOZUR and MOCK, 1977). Accordingly, this sequence would be a stratigraphic link indicating continuous sedimentation between the (Silurian?)-Devonian limy-detrital sequence of the Szendrő Mountains and the Paleozoic-Mesozoic sequence of the Bükk Mountains starting with the Middle Carboniferous clastic strata. The lower part of the Uppony Limestone Formation and of the Lázbérc Formation can be correlated in age with the Szendrőlád Limestone and Borsod Limestone Formations of the Szendrő Mountains.

The Paleozoic sequence of Uppony is hardly known from the metamorphic-petrological point of view. Only generalities resp. qualitative references concern its epimetamorphic character: "epizonal dynamometamorphism" (NOSKE—FAZEKAS, 1973); "low-grade metamorphism" (KOZUR and MOCK, 1977). Its metamorphic degree is weaker than that of the Paleozoic of Szendrő but stronger than that of the Paleozoic of the Bükk Mountains (JÁMBOR, 1961; BALOGH, 1964).

The Triassic volcanogenic-sedimentary sequence of the Bükk Mountains is characterized by Alpine regional dynamometamorphism of pumpellyte-prehnite-quartz facies (ÁRKAI, 1973). The (Silurian?)-Devonian sequence of

the Szendrő Mountains was affected by low-medium-pressure low temperature greenschist facies metamorphism of uncertain age (Hercynian?-Alpine?). The metamorphic grade of the upper horizons corresponds to the boundary between the anchi- and epizones (ÁRKAI, 1977). On the basis of the color alteration index of conodonts (CAI) the metamorphic grades of the Paleozoic of Uppony and of the Triassic formations of the Bükk Mountains is the same. Consequently, the age of metamorphism of the Triassic formations of the Bükk Mountains is Alpine (KOZUR and MOCK, 1977) and the Hercynian orogeny had no or only rather slight effect in this belt.

The method based on the qualitative observation of the colour and recrystallization of conodonts is unsuitable to determine the metamorphic grade just due to other factors (e.g. organic matter content, etc.). Thus, for determining an eventual Hercynian metamorphic effect, the metamorphic-petrological investigation of the Middle Devonian—Middle Carboniferous sequence of Uppony is of decisive significance.

Sampling, Analytical Methods

The strato-types of the Paleozoic of Uppony, the boundaries of the formations resp. the profile discovering these (road-cut of the Lázbérc storage lake), the core samples of four boreholes (Dédestapolcsány-5, 8, 9 and Uppony-10) of about 1500 m length, altogether more than 300 samples were investigated. To collect the surficial samples ELSOLTZ, L. and PÉRÓ, Cs. assisted. The relatively continuous sampling rendered possible the statistical data processing.

The X-ray diffractometric records were made from whole rock and from carbonate-free powder samples (after treating the samples with 3% cold hydrochloric acid) by means of a Philips Micro Müller 111 equipment.

(Recording conditions of the desoriented and oriented samples: CuK_α radiation, acceleration voltage: 42 kV, amperage: 32 mA, proportional counter, graphite monochromator, divergency and detector slit: 1°, goniometer speed: 2°/min resp. 0.5°/min; paper speed: 1600 mm/h, time constant: 2.)

The computation of the semi-quantitative mineral composition was carried out by the direct method of NÁRAY-SZABÓ and PÉTER (1967) using also the data of BÁRDOSSY (1966, 1970) and of RISCHÁK and VICZIÁN (1972). Results were controlled by the solution residue data, occasionally by DTG-analyses and the mineralogical recalculation of the chemical analyses.

The structural order characteristic of the illite-muscovite series with respect to the metamorphic grade was measured by the half-width of the (002) base reflexion after KÜBLER (KÜBLER index) in mm, resp. by the relative half-width in order to eliminate the disturbing effects of methodic conditions (WEBER, 1972):

$$Hb_{rel} = \frac{Hb_{illite(002)}, \text{mm}}{Hb_{quartz(100)}, \text{mm}} \times 100$$

The boundaries between diagenetic zones and anchi-zone resp. between the anchi- and epizones were determined from the data of samples of known metamorphic grade (ÁRKAI, 1977).

To measure the illite crystallinity, sample preparation was standardized: coarse grinding of 5 minute, solution by diluted hydrochloric acid in case of carbonate samples, fine-

grinding producing grain size ranging from 30 to 60 microns. The standard deviations depending on sample preparation and instrumental errors are in case of $n = 10$ sample (measurement), and at 2°/min goniometer speed: KÜBLER index: $s = 0.05-0.18$; Hb_{rel}: $s = 4-10$. Indices were measured for whole rock samples since the average grain size of the investigated samples is >10 μm, and the decrease of grain size produces half-width increase below 2 μm only. In the course of statistical evaluation of the indices obtained in this way the effect of detrital micas can be eliminated.

The determination of the $d_{(060)}$ lattice parameter of sericites was carried out in desoriented powder samples using the quartz content of the samples as an internal standard.

The electron microprobe tests were made by NAGY, G. with a instrument of JEOL JXA-5 type.

The random vitrinite reflectance was measured by means of the reflectance standard of the glass prism series produced by the Bituminous Coal Research Inc, in the light of 548 nm wavelength.

Lithology

On the basis of the average mineral composition of the main rock types of the Uppony Paleozoic formations a continuous and unidirectional change in the sedimentation conditions can be determined (Table I, Fig. 3).

The *Uppony Limestone Formation* is characterized by reductive calcareous sediment of predominantly chemical, subordinately biogenic origin. The supply of detrital (clayey) matter proved to be very low (less than 2%).The detrital matter is characterized by the predominance of quartz and phyllosilicates and by the small quantity of feldspars. On the basis of this, as well as of the low chlorite/sericite ratio this formation is rather similar to the Borsod Limestone of the Szendrő Mountains.

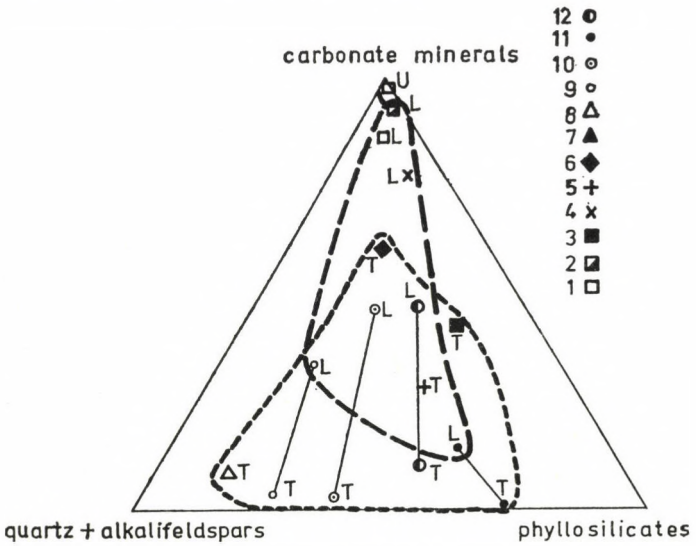

Fig. 3. Average mineralogical composition of the main rock types of the Uppony Paleozoic formations. 1. limestone; 2. metasomatic dolomitized limestone; 3. clayey limestone; 4. metatuffitic limestone; 5. metabasalt; 6. carbonate schist; 7. sideritic meta-chert; 8. meta-chert; 9. metasandstone; 10. slate (clayey, silty); 11. slate; 12. banded slate (clayey, silty). U = Uppony Limestone Formation; L = Lázbérc Formation; T = Tapolcsány Formation

The *Lázbérc Formation* characterized by the variety of sedimentation conditions developed from the previous sequence by continuous transition. It is constituted by chemically precipitated calcareous, resp. detrital (pelitic-psammitic) sediments and their mixtures of varying ratio as well as by strata consisting of the rhythmic alternation of them.

In the lower part of the Lázbérc Formation metatuffitic limestone (previously "tuff-stripped limestone" "limestone with diabase tuff") intercalations are found. In this series the lime-material of sedimentary origin predominates. The dispersed (probably halmyrolitically argillified volcanic fine-detrital matter accumulated syngenetically in the calcareous mud, forming thin strips. The metatuffitic strips do not contain primary igneous minerals or relict structures relating to volcanic rocks. The volcanic origin can be probabilized only on the basis of the high chlorite and sericite content of the strips as well as of the oxidative character (hematite) which sharply differs from the reductive character of the formation in general. (The lack of feldspars and the small amount of quartz suggest basic volcanic origin.) The metatuffitic limestone is very similar to the cipolino-like crystalline limestone of the Szendrő Mountains which is probably also of metatuffitic origin.

There is a continuous sedimentation transition from the Lázbérc Formation to the overlying Tapolcsány Formation. In the Tapolcsány Formation the lime content of the pelitic sediments is gradually decreasing upwards and the siliceous and fine-detrital sediments containing bacterial pyrite, siderite and organic matter and deposited in strongly reductive environment, become predominant. These are overlain by coarser-detrital (sandy) sediments. Pelitic, aleuritic, siliceous, subordinately sandy sediments containing syngenetic basaltic (andesitic) volcanic intercalation constitute the terminal member of the formation. Evidence for the syngenetic volcanic activity is as follows: the lack of contact effects, and the joint occurrence of lava and lapilli-tuff-like volcanites in eight horizons, in a thickness of 5—8 m on the average and 15 m as a maximal value.

The gradual decrease of the carbonate content (Fig. 3) as well as the gradual predominance of the fine-detrital and siliceous sediments indicate deeper, pelagic conditions. Deep faults related to the subsidence might initiate the submarine volcanism of predominantly basaltic composition.

Regional Metamorphism

Texture and Structure

Due to the metamorphic recrystallization the difference between the detrital constituents (predominantly quartz, subordinately feldspars, muscovite and muscovite-chlorite aggregates formed after biotite (Plate I/2)) and

the fine-grained clayey, carbonatic cement was gradually eliminated except the coarser-detrital rocks.

In the meta-sandstones poor in carbonates "spiny like textures" (radially arranged quartz-sericite fibres) can be observed around the quartz grains (Plate I/1). This indicates anchi-zonal metamorphism which can be paralleled with the pumpellyite-prehnite-quartz facies (KOSSOVSKAYA and SHUTOV, 1963; FREY, 1970).

The groundmass of the rocks of aleuritic origin recrystallized to a quartzite-like texture containing oriented sericite scales.

The crenulation cleavage is commonly characteristic of the slates (Plate I/3). The recrystallization related to the progressing burial generated sericite orientation parallel to the sedimentary stratification (I. clevage plane). In the subsequent dynamometamorphic phase micro-folds and from their evolution shearing cleavage planes were generated (II. cleavage direction). The two recrystallization phases do not indicate necessarely two metamorphic phases of different age (since no paragenetic differences exist). It seems to be more probable that this phenomenon reflects different evolution phases of one tectonometamorphic cycle.

The stylolitic structure is characteristic of the limestones. Grain size varies in strips, the direction of elongation of the recrystallized carbonate grains is parallel with the II. cleavage plane. This can be particularly well observed in the strips of the metatuffitic limestone containing sericite, chlorite, quartz and hematite (Plate I/4). The grain size of limestones depends on the quantity of the detrital and organic matter as well as on their dispersity, thus it cannot be applied even to the qualitative characterization of the metamorphic grade.

Mineral Assemblages

The Uppony Paleozoic metamorphites of sedimentary origin do not contain critical mineral parageneses suitable to the more exact determination of the physical conditions of recrystallization. The metamorphic mineral parageneses of the main rock types are as follows:

crystalline limestone: quartz, sericite (±chlorite, pyrite, albite, potash feldspar, calcite, dolomite-ankerite, siderite);

slate (clayey-silty): quartz, chlorite, sericite, albite, potash feldspar (± calcite, dolomite-ankerite, siderite, pyrite, hematite);

metasandstone: quartz, albite, sericite, chlorite (±calcite, potash feldspar, dolomite-ankerite, siderite, pyrite, hematite, chloritoid).

Of course of minerals listed above quartz and part of feldspar are of detrital origin: depending on the grain size, these are not or only slightly recrystallized. The major part of the dolomite-ankerite and siderite was

Table 1

Average mineral composition and sericite parameters of the main rock types of the paleozoic formations of Uppony Mountains

	Semiquantitative Mineral Composition (Weight percent)																				Sericite parameters											
																					Crystallinity indices								b_0 (Å)		d_{002} (Å)	
																					Kübler (mm)				Weber							
																					OR.		DESOR		OR.		DESOR					
		Quartz	Albite-oligoclase	Potash feldspar	Sericite	Glauconite-celadonite	Stilpnomelane	Chlorit	Chloritoid	Kaolinite	Montmorillonite	Clay mineral of mixed layer structure	Calcite	Dolomite-ankerite	Siderite	Pyrite	Hematite	Goethite	Jarosite	Rutile	n	$\bar{x}$	n	$\bar{x}$	n	$\bar{x}$	n	$\bar{x}$	n	$\bar{x}$	n	$\bar{x}$
III. Tapolcsány Formation																																
metabasalt	18	8	18	3	5	7	4	25	—	ny	—	—	10	18	ny	1	1	—	—	ny	7	2.96	8	3.06	5	111	8	108	7	9.026	8	9.972
slate	20	18	7	2	30	—	—	40	ny	—	—	ny	1	ny	ny	2	ny	ny	—	ny	6	3.18	19	3.63	7	119	17	128	17	8.993	18	9.975
banded slate (clayey, silty)	10	32	15	3	20	—	—	24	—	—	—	—	ny	3	2	1	—	—	—	—	8	3.19	10	3.30	9	119	10	116	9	9.001	9	9.988
metasandstone	17	53	16	2	14	—	—	11	—	ny	—	ny	1	3	ny	ny	—	ny	—	ny	6	3.01	15	3.43	6	113	15	121	14	8.997	15	9.956
carbonate slate	4	19	—	1	14	—	—	5	—	—	—	—	—	13	37	11	—	—	—	—	2	2.80	3	3.19	2	105	3	112	3	9.045	3	10.013
sideritic meta-chert	7	36	—	—	16	—	—	15	—	—	—	—	—	—	27	6	—	—	—	—	4	2.95	6	3.43	4	110	6	121	1	9.018	7	9.981
meta-chert	22	69	1	1	10	—	—	3	—	ny	—	—	2	4	2	8	—	—	ny	—	8	3.14	16	3.46	8	118	16	122	16	9.015	17	10.005
II. Lázbérc Formation																																
slate	9	20	8	1	27	—	—	29	—	—	—	ny	14	ny	—	1	—	—	—	—	3	2.84	6	3.29	3	106	7	116	6	8.999	7	9.963
banded slate (clayey, silty)	2	11	12	1	12	—	—	17	—	—	—	—	47	—	—	—	—	—	—	—	1	3.08	1	3.66	1	115	1	129	1	8.994	1	9.964
metasandstone	5	36	12	ny	11	—	—	7	—	ny	—	—	34	—	—	ny	ny	ny	—	—	3	2.65	3	3.31	3	99	1	118	3	9.006	4	9.959
metasomatic-dolomitic limestone	9	3	ny	ny	5	—	—	—	—	—	—	—	29	62	—	—	—	1	—	—	1	2.81	3	3.13	3	105	3	110	3	9.014	3	9.984
meta-tuffitic limestone	6	7	—	ny	9	—	—	7	—	—	—	—	74	—	—	—	3	—	—	—	3	2.30	8	2.99	3	86	8	105	4	9.017	8	9.968
limestone, lime-schist	41	5	2	ny	4	—	—	2	—	ny	—	ny	87	—	—	ny	ny	—	—	—	18	2.79	33	3.23	19	105	33	114	32	9.010	34	9.968
I. Uppony Limestone Formation																																
crystalline limestone	14	1	—	—	1	—	—	—	—	—	—	—	98	—	—	ny	—	—	—	—	10	2.97	13	3.42	10	111	13	120	12	9.010	13	9.970

OR = oriented } sample
DESOR = desoriented

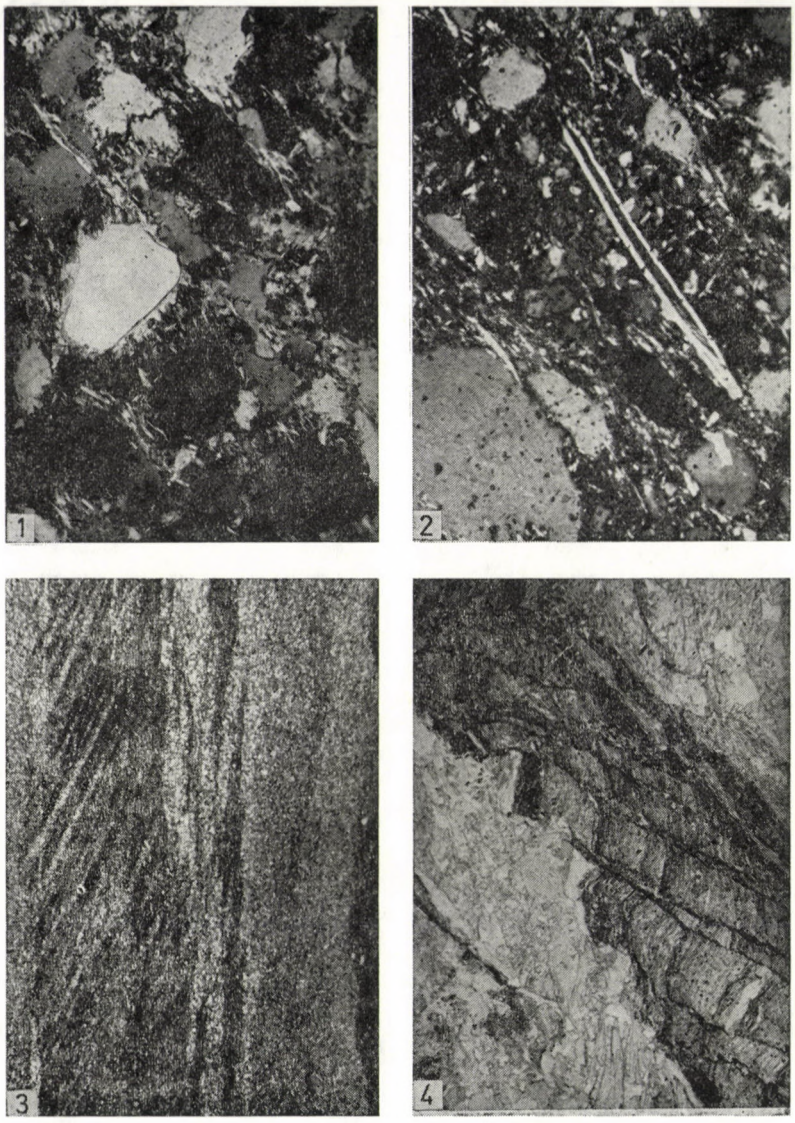

Plate I

1. "Spiny like texture" (radially arranged quartz-sericite fibres) around detrital quartz grains in metasandstone. U-II/I sample, + Nicols M = 30×
2. Muscovite-chlorite aggregate formed after detrital biotite in metasandstone. Dt-8. 256.7 m, + Nicols, M = 115×
3. Crenulation cleavage in slate. — Dt-8. 38.0 m. 1 Nicol, M = 30×
4. Folded metatuffitic strip with crenulation cleavage in limestone. U-380. 1 Nicol, M = 30×

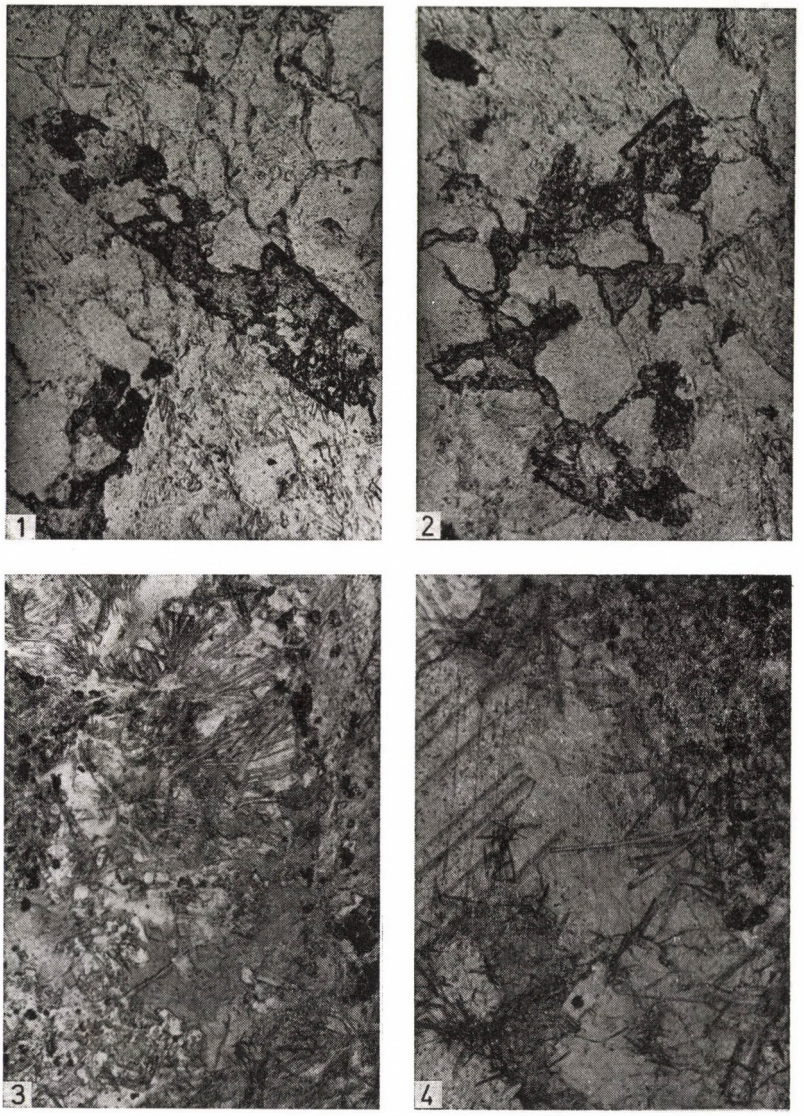

Plate II

1. **Chloritoid** grains filling the intergranular space between detrital quartz grains in **metasandstone. U-II/I.** 1 Nicol, M = 115 ×
2. **See U-II/1;** 1 Nicol, M = 115 ×
3. **Components of** metabasalt-tuff groundmass: stilpnomelane, glauconite-celadonite, **calcite, titanite. Dt-8.** 188.0 m, 1 Nicol, M = 115 ×
4. **Amygdaloidal** metabasalt with calcite, glauconite-celadonite and stilpnomelane **filling. Dt-8.** 188.8 m, + Nicols, M = 30 ×

formed by subsequent metasomatic effect. The effect of Superficial weathering is reflected by the kaolinite, the mixed layer-structure clay mineral of 8 and 12 Å reflection, by the goethite and the rarely occurring jarosite and gibbsite.

From the metamorphic-petrological point of view the chloritoid-bearing metasandstone (Tapolcsány Formation, Rágyincs Valley) is of special interest. Its mineral composition is: 80% quartz, 8% sericite, 5% chloritoid, 2—2% potash feldspar and chlorite, 1—1% albite, kaolinite and clay mineral of mixed-layer structure. The detrital origin of chloritoid was assumed by NOSKE-FAZEKAS (1973) though its grain size is much greater than that of the detrital heavy minerals (zircon, rutile, etc.) of the rock.

According to the microscopic and microprobe observations (Plates II/1—2; IV/1—3) chloritoid is a mineral produced by anchi-metamorphic recrystallization, filling the space between the quartz grains in the intergrowing "spiny-like" sandstone structure characteristic of the anchi-zone. In addition to the xenomorphic chloritoid grains subordinately idiomorphic, sometimes polysynthetically twinned grains can also be observed.

The chemical composition of the chloritoid grains is homogeneous. No difference has been found between the core parts of the grains and the marginal parts believed previously to have been replaced by quartz:

U-II/I sample	Centre	Margin
	of the chloritoid grain	
SiO_2	24.3	25.2
Al_2O_3	39.8	40.0
FeO*	22.2	22.6
MgO	2.0	2.0
CaO	0.1	0.1
Total:	88.5	90.1 w%

* Total Fe as FeO; microprobe analysis by NAGY, G.

The projection points of the chemical composition of the chloritoid-bearing metasandstone given in Table II fall within resp. in the close neighbourhood of the fields found to be characteristic of chloritoids by HOSCHEK (1967).

In the phyllites of the Aar Massif FREY and WIELAND (1975) determined the reaction: pyrophyllite + chlorite → chloritoid + quartz + H_2O. This isoreaction grade is assigned to the epi-zone. According to MIYASHIRO (1973) other possibilities for chloritoid formation are as follows:

kaolinite + chlorite → chloritoid + quartz + H_2O;
hematite + chlorite → magnetite + chloritoid + quartz + H_2O;
paragonite + chlorite + quartz → chloritoid + albite + H_2O.

Table 2

Chemical composition of some characteristic rock types of the paleozoic formations of the Uppony Mountains

	1.	2.	3.	4.	5.	6.
SiO_2	56.89	31.04	39.94	72.85	61.53	88.36
TiO_2	0.96	2.40	3.68	0.56	0.56	0.24
Al_2O_3	19.26	11.83	12.72	11.90	13.33	6.11
Fe_2O_3	1.65	1.41	1.94	0.51	—	0.03
FeO	6.76	8.16	9.10	2.46	6.85	0.89
MgO	1.00	7.98	9.48	1.00	0.50	0.50
CaO	2.09	13.54	9.37	2.09	4.18	0.70
MnO	tr.	0.11	tr.	tr.	0.07	tr.
Na_2O	1.08	0.62	2.26	1.86	0.24	0.50
K_2O	3.35	1.80	2.65	2.40	2.50	0.50
$-H_2O$	—	—	0.13	—	0.33	0.06
$+H_2O$	4.17	2.58	4.14	1.71	4.25	0.98
CO_2	1.04	17.26	3.93	1.88	4.32	0.24
P_2O_5	0.12	0.41	0.16	0.07	0.34	0.05
Σ	98.37	99.14	99.50	99.29	39.00	99.16 weight per cent

1. Banded slate (clayey, silty), Dt-8. 81.2 m; 2. metabasalt, Dt-8. 156.6 m; 3. metabasalt tuff, Dt-8. 188.0 m; 4. metasandstone, Dt-8. 296.0 m; 5. meta-chert (clayey), Dt-8. 493.7 m; 6. chloritoidic metasandstone, U-II/I, Rágyincs Valley.

Though in the medium- and high-pressure metamorphites the chloritoid is more common (SEIDEL et al., 1975), chloritoid is known to develop also under low-pressure conditions. No data are available on the lower temperature stability limit of the chloritoid. It occurs predominantly in formations of green-schist facies but may also occur in the (very low-grade zone) anchi-zone (KÜBLER, 1975; WINKLER, 1976).

No observation data are available on the formation reaction of the chloritoid found in the metasandstone of the Uppony Mountains. On the basis of the illite-crystallinity, vitrinite reflectance and textural features of the neighbouring formations, the formation of chloritoid can most likely be ranged in the high-temperature part of the anchi-zone, near the boundary between the anchi- and epi-zones.

The mineral assemblage of the metatuffitic limestone of the Lázbérc Formation consist of calcite, sericite, chlorite, quartz, hematite ($\pm$ potash feldspar).

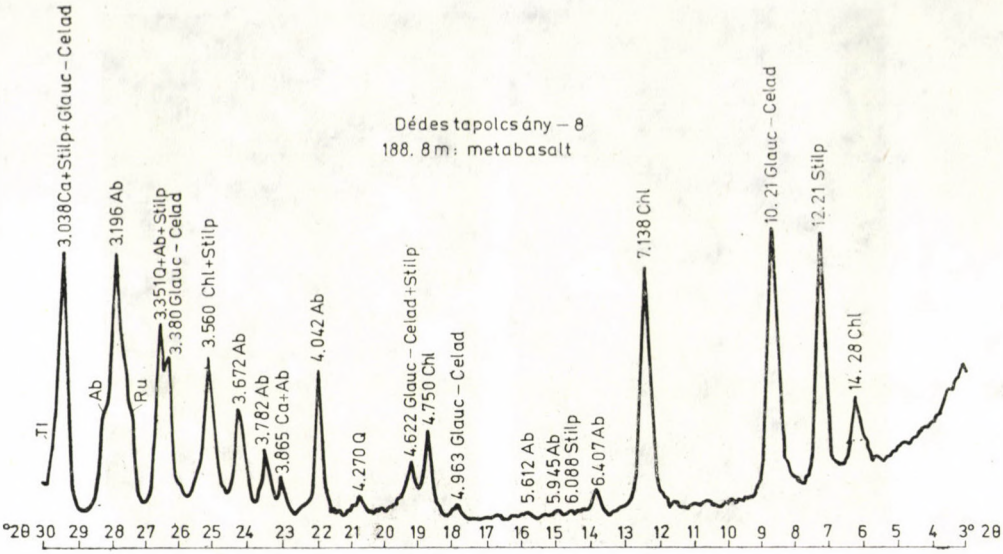

Fig. 4. **Diffractogram of metabasalt containing glauconite-celadonite and stilpnomelane. Ab = albite; Ca = calcite; Chl = chlorite; Glauc-celad = glauconite-celadonite; Q = quartz; Stilp = stilpnomelane**

In the meta-basalt and tuff rocks of the Tapolcsány Formation three mineral assemblages could be distinguished:

1. dolomite-ankerite, chlorite, albite, quartz, sericite, potash-feldspar (± calcite, siderite, pyrite, hematite, rutile, stilpnomelane);

2. glauconite-celadonite, stilpnomelane, chlorite, albite, calcite (± potash-feldspar, ilmenite);

3. chlorite, glauconite-celadonite, calcite, albite, dolomite-ankerite, potash feldspar.

The 2nd and 3rd mineral assemblages might develop from a more basic initial material than the first one.

Since in the fine-detrital volcanic rocks metamorphic schistosity can be observed and the illite crystallinity of the metavolcanites is close to that of the associated rocks, the regional metamorphic origin of the stilpnomelane is highly probable. Glauconite-celadonite was formed presumably during the low temperature transformation of volcanites, i.e. before the metamorphism. The fine-grained glauconite-celadonite forming rarely scaly heaps and the fibrous stilpnomelane developed from it occur in the groundmass of the lava rocks, replacing the phenocrysts, in the lapilli and cement of volcanic tuffs and it is found most frequently in the calcitic-chloritic hollows resp. alveoles (Plates II/3—4, III/1 and IV/4—5). A characteristic diffractogram of the metabasalt is shown in Fig. 4. The chemical composition of the glauconite-celadonite and stilpnomelane determined by means of microprobe Table 3.

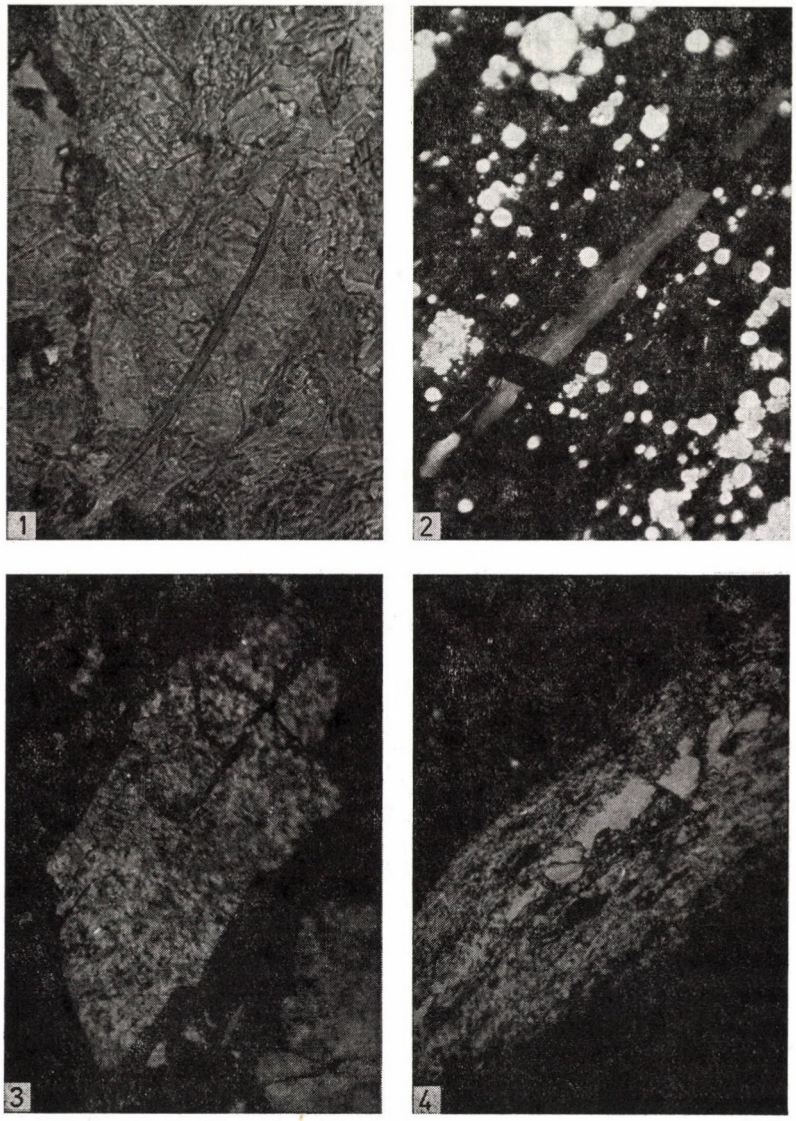

Plate III

1. Metabasalt tuff, groundmass: stilpnomelane, glauconite-celadonite and chlorite. Dt-8. 188.0 m, 1 Nicol, M = 450×
2. Fractured anthracite strip in bacterio-pyritic environment. Meta-chert. Dt-8. 480.0 m, without Nicols, M = 540×
3. Metabituminite with mosaic-reflexion. Clayey meta-chert, Dt-8. 396.5 m. Without Nicols, M = 290×
4. "Coexisting" anthracite (bigger fractured grains) and graphite in clayey meta-chert. Dt-8. 396.5 m. without Nicols, M = 800×

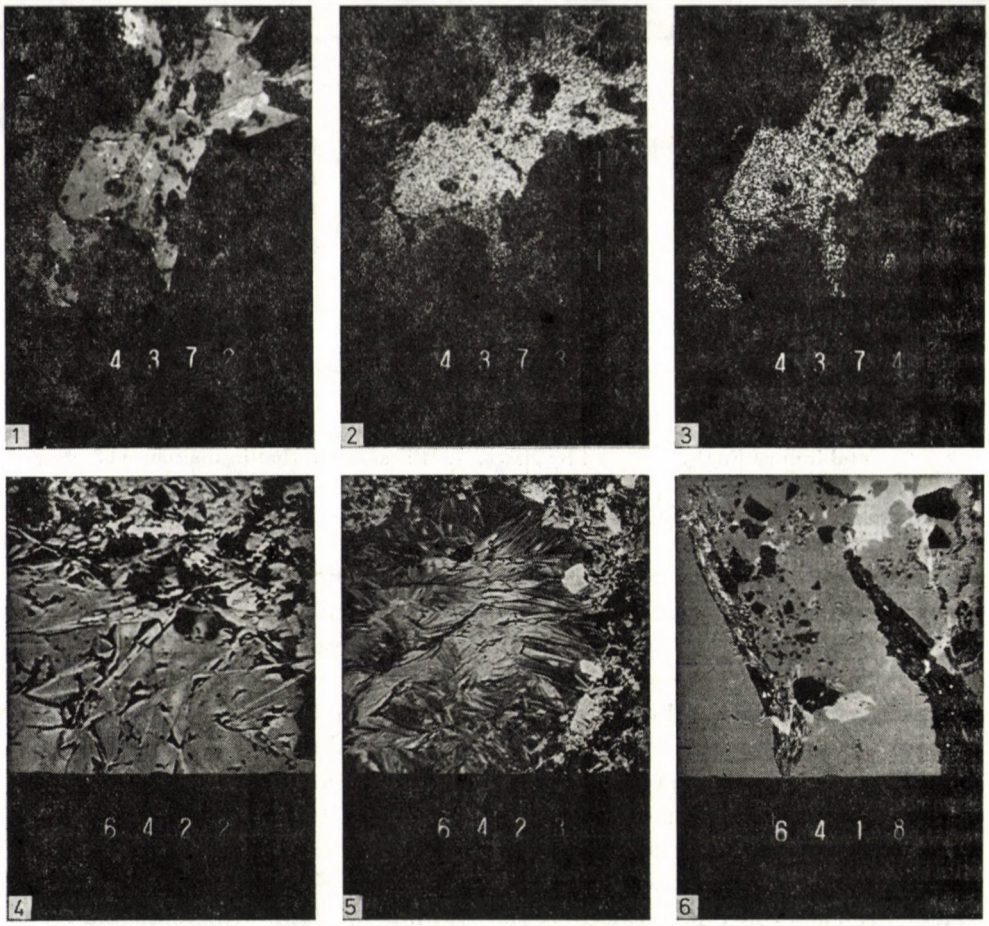

Plate IV

1. Composition image of chloritoid grain. U-II/I. M = 160×
2. $Al_{k\alpha}$ picture of IV/1
3. $Fe_{k\alpha}$ picture of IV/1
4. Glauconite filling in metabasalt tuff (Dt-8. 188.0 m). Composition. M = 160×
5. Elongated crystals of stilpnomelane in the filling of metabasalt tuff (Dt-8. 188.0 m). Composition, M = 320×
6. Coexisting anthracite and graphite in clayey meta-chert (Dt-8. 396.5 m), Composition M = 300×

The optical features (weak, yellowish-green, greenish-brown pleochroism) and the X-ray diffractometric parameters of the glauconite-celadonite are similar rather to the celadonite resp. glauconite mineral, its chemical composition is of transitional character. The determination of stilpnomelane is unambiguous on the basis of the strong reflexion at 12 Å occurring in the diffractograms.

Table 3

	Glauconite-celadonite (Picture IV/4)	Stilpnomelane (Picture IV/5)
SiO_2	56.0	45.2
Al_2O_3	5.7	10.5
Fe_2O_3**	23.1	17.7
MgO	10.2	15.1
MnO	0.4	—
CaO	0.7	0.1
K_2O	2.1	8.5
Total	98.2	97.1 w%

** Total iron as Fe_2O_3. Analyzed by NAGY, G. It is assumed that in the course of analysis a part of crystalline water was lost.

The formation of stilpnomelane is possible in the most different pressure ranges though it is most common in the high pressure low temperature metabasites, iron-rich sandstones and metapelites (MIYASHIRO, 1973). It is stable both in the pumpellyite-prehnite-quartz and in the greenschist facies (WINKLER, 1976). In the iron-oolitic glauconite-bearing metapelites of the Helveticum of the Glarus Alps the stilpnomelane was formed under anchi-zone conditions according to the reaction: glauconite + chlorite + quartz → stilpnomelane + potash feldspar + H_2O + O_2. The estimated minimum parameters of the isoreaction grade are: 1.6 kilobar and 260 °C (FREY et al., 1973; FREY, TEICHMÜLLER and MULLIS, 1978).

In the metabasalt of Uppony the stilpnomelane developed from the glauconite-celadonite mineral. It is remarkable that the stilpnomelane being of smaller potash content than the glauconite-celadonite accompanies the disappearance of potash feldspar and Ca-Mg-Fe-carbonate and the appearance of calcite. Thus the participation of these minerals in the reactions can also be assumed. The data available so far are insufficient to determine exactly the reaction, nevertheless the isograde determined by the common occurrence of glauconite-celadonite and stilpnomelane indicate recrystallization conditions characteristic of the middle-upper part of the anchi-zone corresponding to pumpellyite-prehnite-quartz facies.

Illite Crystallinity

The Kübler-indices of the sericites of the Uppony Paleozoic rocks are of unimodal distribution (Fig. 5a-b). This would indicate regional metamorphism corresponding to the boundary of the anchi- and epi-zones. Their maxima fall to the epizone, near the anchi-epi-boundary. The same conclusion can

be drawn from the distribution of the relative half-widths (Fig. 5c-d). The illite crystallinity indices of the Paleozoic formations of the Szendrő Mountains suggest epi-zonal (greenschist) metamorphism. The metamorphic grade of the Paleozoic of Uppony as a whole, is lower than that of the whole of the Szendrő Paleozoic.

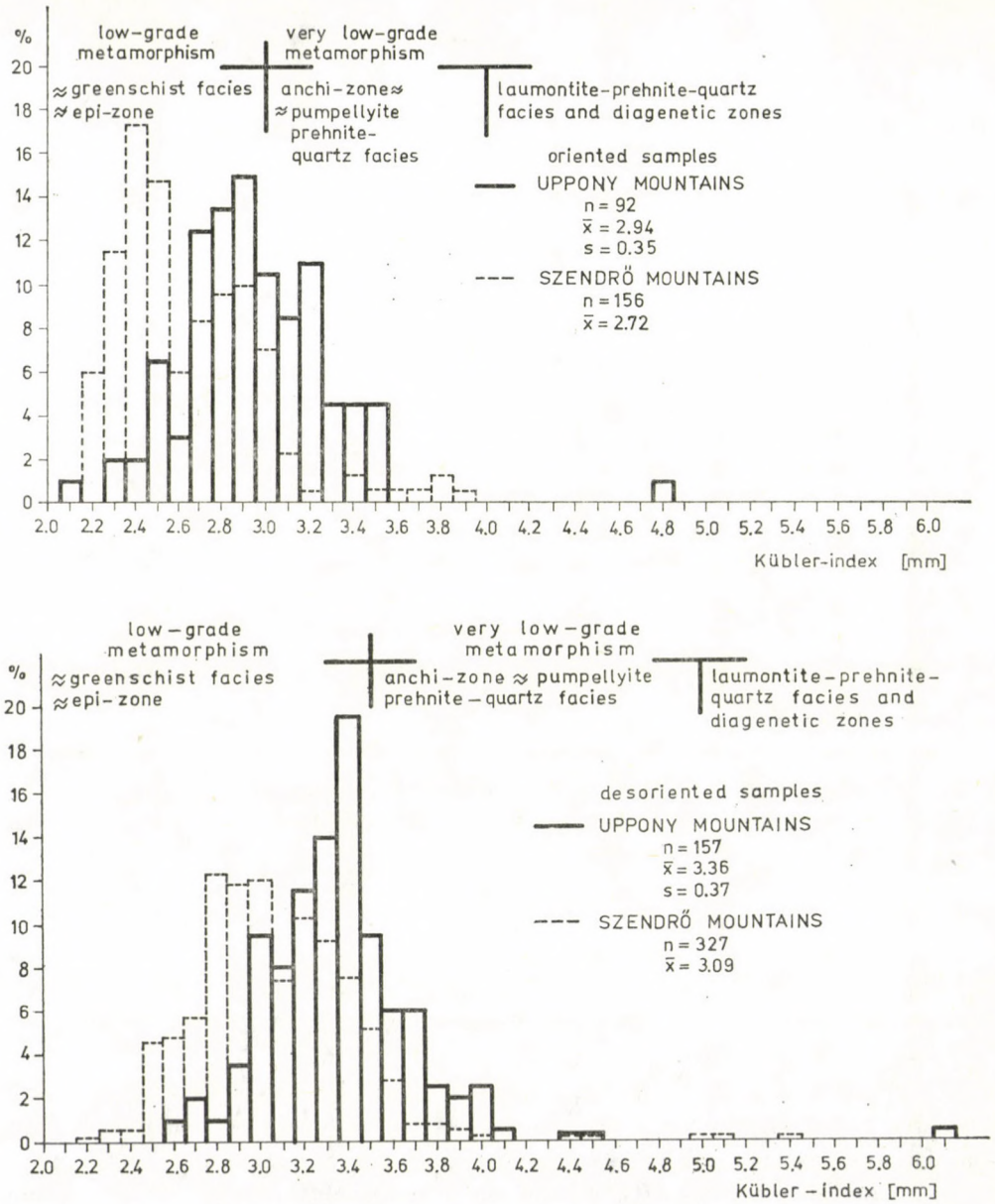

Fig. 5/a—b. Frequency distributions of illite crystallinity indices

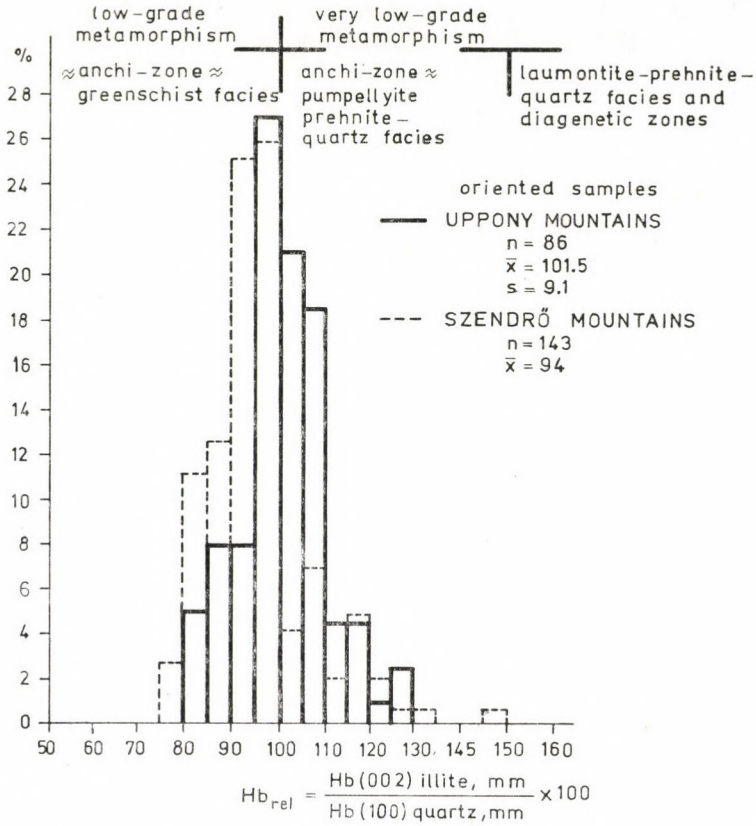

Fig. 5/c. Frequency distributions of illite crystallinity indices. WEBER — indices

No considerable differences have been found in the average values of crystallinity indices of the various Paleozoic formations of Uppony. The indices of the Lázbérc Formation revealing somewhat stronger transformation can be explained by lithological reasons.

In the Lázbérc Formation the illite crystallinity decreases in the following succession: metatuffitic limestone—metasandstone—limestone—slate. This tendency can be explained partly by the effect of the originally different degradation states of the 10 Å phyllosilicates in the different lithofacies which can be recognized already in spite of recrystallization, partly by the different porosity and permeability of the individual rock types.

Similar reasons may be responsible for the decrease of illite crystallinity in the Tapolcsány Formation, the succession of decrease being carbonate schist—metabasalt and tuff—sandstone—meta-chert—slate.

The illite crystallinity of the slate and metasandstone types of the Tapolcsány Formation which represent a higher stratigraphic horizon is some-

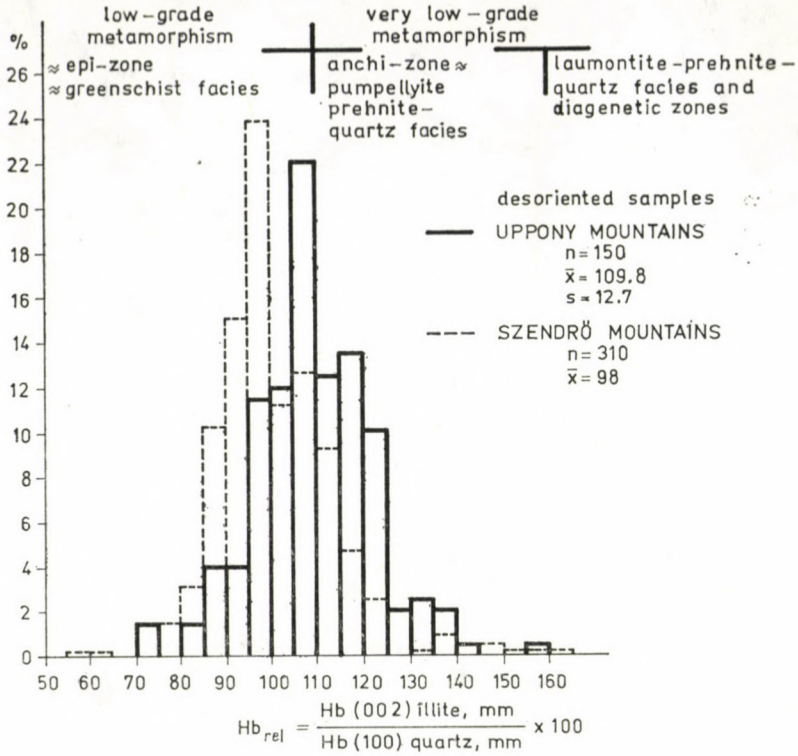

Fig. 5/d.

what lower than the average values of the corresponding rock types of the underlying Lázbérc Formation. It should be emphasized that the illite crystallinity of metabasalt and tuff intercalations is essentially the same or somewhat better than that of the associated sedimentary rocks. This is an evidence to the same regional metamorphic transformation of both the sedimentary rocks and volcanites.

b_0 *and* d_{002} *Parameters of the Sericites*

The b_0 frequency distribution of the Uppony Paleozoic is unimodal (Fig. 6). The average value is somewhat lower than that of the Szendrő Paleozoic.

Also the frequency distribution of the I_{004}/I_{002} intensity ratios (Fig. 7) is similar to the distribution of the samples of Szendrő. This suggests a unimodal and according to the ESQUEVIN (1969) diagram predominantly muscovitic, subordinately phengitic composition.

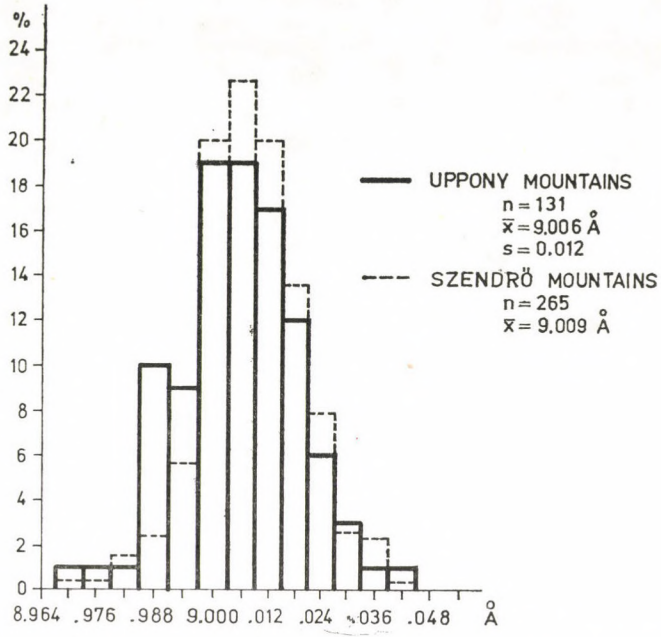

Fig. 6. Frequency distribution of the b_0 lattice parameter of sericites

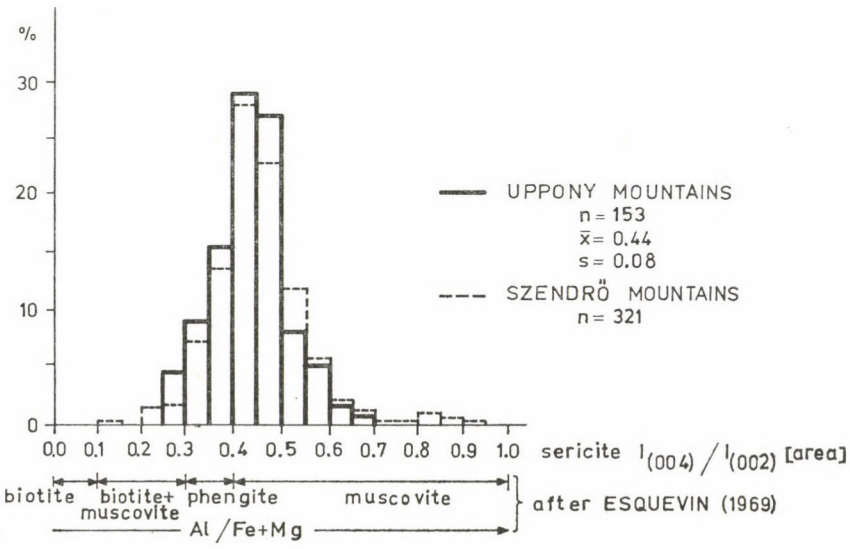

Fig. 7. Frequency distribution and composition of the I_{004}/I_{002} intensity ratios of sericites

The average b_0 values of the formations are dependent on the lithological composition and do not characterize the pressure conditions of the metamorphism (Fig. 8).

While no systematic difference is found in the d_{002} values of the different rock types, the b_0 average values of sericites of the calcareous rocks of sedimentary origin are considerably higher than those of the low carbonate rocks of detrital origin. (The measurements were carried out on carbonate-free residues of the samples, thus the disturbing effect of carbonate reflexions could be avoided. The same relationship, which is opposite to the statements of GEYSSANT et al. (1973) was established also in the Szendrő Paleozoic (ÁRKAI, 1977). This relationship may be in connection with higher Mg/Al ratio of limestones as compared to the detrital rocks.

The higher b_0 value of the metatuffitic limestones and metabasalts may also be due to petrochemical reasons. While in this case first of all the Mg-Al replacement should be taken into account, the extremely high b_0 values of meta-cherts and carbonate schists can be explained by the high Fe/Al ratio, thus by the considerable Fe-Al substitution in the sericites.

To determine the pressure conditions of the regional metamorphism, the slates of low carbonate (less than 20%) content, of quartz, sericite, chlorite,

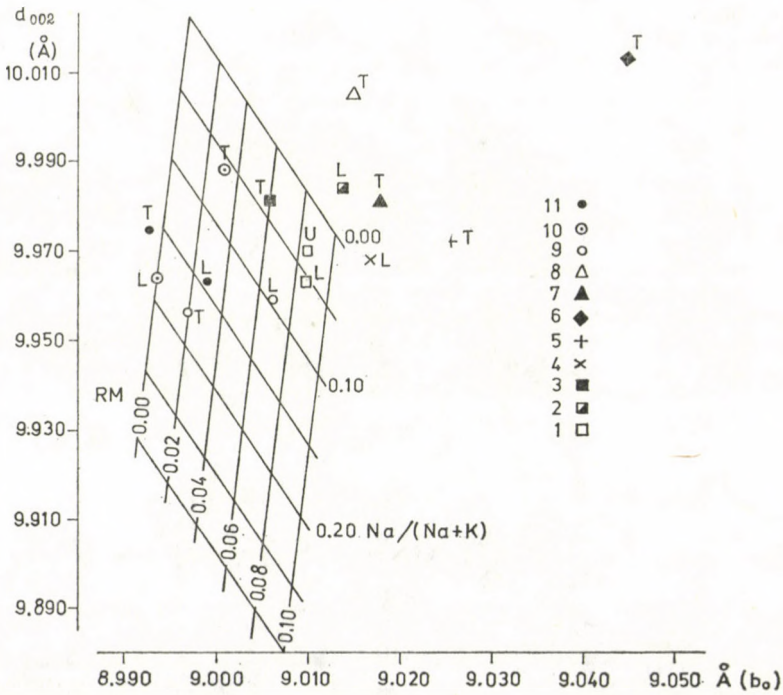

Fig. 8. b_0 and d_{002} values of sericites of the main rock types. The Na/Na + K and RM values (RM = $^1/_2$ Fe_2O_3 + FeO + MgO) after GUIDOTTI and SASSI (1976). Legend see on Fig. 3

± albite, potash feldspar, pyrite mineral assemblage was used. Since the method had been elaborated to low temperature greenschist facies metapelites, and the metamorphism of the Uppony Paleozoic reached only the boundary of the anchi- and epi-zones, the pressure estimation on the basis of the b_0 values may be only informative.

The average b_0 value of the chosen slate samples of Uppony ($\bar{x} = 8.9965$ Å, $n = 34$, $s_x = 0.010$, $s_{\bar{x}} = 0.0018$) indicates low pressure metamorphism. The cumulative distribution curve of the b_0-values indicates lower pressure than suggested by the distribution of the Szendrő metapelites (Fig. 9), and reveals a transitional position between the low pressure Bosost and the Central East Dalradian type indicating the low-medium pressure boundary.

Out of the b_0 pressure data of the metamorphites of the Alps and Carpathians, the metamorphites of the Uppony Mountains are similar to the phyllites of the Eastern Carpathians and Eastern Alps of Hercynian age but they indicate a slightly higher pressure (Fig. 10).

In the Alpine and Carpathian region the low pressure metamorphism is characteristic of the Hercynian tectonometamorphic cycle. On the contrary, the low pressure does not indicate necessarily Hercynian age since the pressure

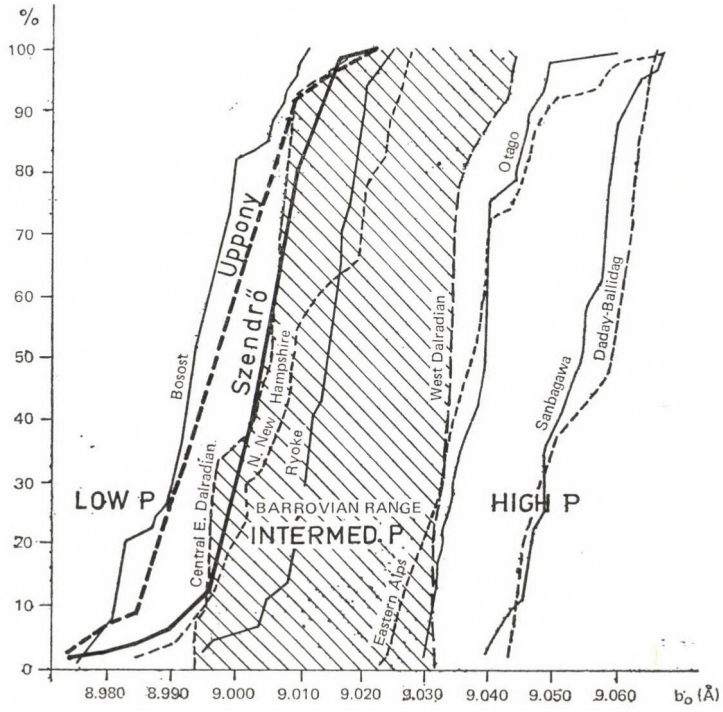

Fig. 9. Cumulative frequency curves of b_0 parameters of muscovites from low-temperature greenschist facies metapelites of different pressure range (after SASSI and SCOLARI, 1974)

conditions of metamorphism of the Hercynian showed also considerable fluctuations (KRAUTNER, SASSI et al., 1976) on the one hand, and the Alpine tectonometamorphic cycle consists of different recrystallization phases of different pressure ranges depending on the tectonic state of evolution, on the other (NIGGLI et al., 1978).

Vitrinite Reflectance, "Co-existing" Anthracite and Graphite

The clayey and silty slate and meta-chert types of the Lázbérc and Tapolcsány Formations contain great amounts of fine-dispersed coalified plant remnants. The organic matter as well as the spherical bacteriopyrite of stripped distribution parallel with the stratification, further the less frequent chalcopyrite are indicative of reductive environment of sedimentation (Plate III/1).

Three types of the coalified plant remnants can be distinguished: anthracite, metabituminite with mosaic-reflectance and graphite.

Anthracite is usually of finely disperse (less than 1 μm on the average), less frequently it forms greater grains fractured by deformation effects (Plate III/2).

Metabituminite which is a transformation product of the mobilized but not migrated bituminous substance of the rock occurs rarely but always in the cavities of carbonate crystals. In polished sections it appears as an aggregate of grains of strong bireflexion and less than 1 μm producing thus the characteristic mosaic-reflectance (Plate III/3). Its reflectance surpasses that of the syngenetic vitrinites in the anthracite state.

In the samples of the Tapolcsány Formation the following average random reflectance values were determined (due to the small number of measurement, i.e. 10—60, averages can be considered to be of informative nature only):

Sample	Rock	$\bar{R}$%	n	s
Dt-8. 373.0 m	meta-chert,anthracite	5.24	14	0.35
Dt-8. 373.0 m	meta-chert, metabituminite	5.64	22	1.00
Dt-8. 396.5 m	meta-chert, anthracite	4.43	30	0.40
Dt-8. 396.5 m	meta-chert, metabituminite	6.02	56	0.52
Dt-8. 466.6 m	slate, anthracite	6.29	14	0.68
Dt-8. 480.0 m	slate, anthracite	5.62	12	0.43

The average value of the homogeneous anthracite grains is $\bar{R}_0 = 5.17\%$. The illite crystallinite indices of the same samples indicate the boundary of the anchi- and epi-zones.

According to different authors the boundaries of the diagenetic zones/anchi-zone and anchi-zone/epi-zone) can be characterized by the following vitrinite reflectance data resp. volatile contents:

Author	Diagenetic zones/anchi-zone		Boundary of anchi-epi-zones		
	R%	VM%	R%	VM%	
FREY and NIGGLI, 1971	2.2*	10.0	3.7*	4.0	(R_0)
WEBER, 1972; WOLF, 1972	4.0	4.0*	6.0	2.5*	(R_{max})
KISCH, 1974	2.5*	8.0	6.0*	2.5	(R_{max})
KÜBLER, 1975	2.6—2.8	7.0*	4.0—4.2	3.5*	(R_0)
FREY, TEICHMÜLLER and MULLIS, 1978	2.4—3.1	3.5—5.5*	4.7—5.6	2.8—2.0*	(R_{mean})

* Estimated values after the KÖTTER diagram.

The average value ($\bar{R}_0 = 5.17\%$) of the measurements concerning the Uppony samples falls to the interval of the anchi-epi-boundary of FREY, TEICHMÜLLER and MULLIS supported by mineral-paragenetic resp. illite crys-

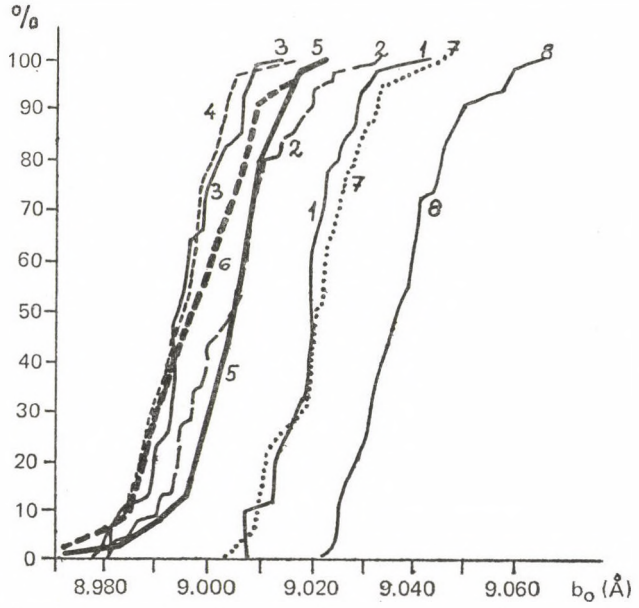

Fig. 10. Cumulative frequency distributions of b_0 parameters of muscovite in the metapelites of the Alps and the Carpathians (SASSI and SCOLARI, 1974; KRAUTNER, SASSI et al., 1975, 1976) as well as in the low-grade Paleozoic metamorphic pelites of Northeastern Hungary. 1. Dalslandian (Early Baikalian) Rebra-Barnar series of the Eastern Carpathians; 2. Assyntic (Late Baikalian) Tölgyes Series of the Eastern Carpathians; 3. Hercynian Repedea, Tibau and Rusaia Series of the Eastern Carpathians; 4. Hercynian phyllites of the Eastern Alps; 5.phyllites of the Szendrő Mountains; 6. slates of the Uppony Mountains; 7. Hercynian phyllites, Poiana Rusca Massif, Southern Carpathians; 8. Alpine phyllites, Eastern Alps

tallinity data. Consequently, though there are considerable differences in the literature data, on the basis of vitrinite reflectance data the metamorphism corresponding to the anchi-epi-zone boundary can be established. This agrees fairly well with the results obtained by the illite crystallinity indices.

In the parts of shearing boundaries between rock masses of different mechanical thus of different dynamic behaviour (limestone and slate) as well as in micro-dimensions: along the sheared cleavage planes the major part of anthracite was transformed into folded strips of graphite (average grain width: < 1 μm, max.: 40 μm). At the same time, between the shearing strips the anthracite has been preserved in the stress-free parts (Plates III/4 and IV/6).

The "co-existence" of anthracite and graphite has been unknown so far in Hungarian rock samples. A similar phenomenon was described by DIESSEL and OFFLER (1975) from the metamorphic series of Mt. Lofty (Southern Australia) where this "co-existence" suggests conditions corresponding to the chlorite zone of the greenschist facies.

According to our observations the appearance of graphite is caused first of all by the local increase of shearing-stress; consequently, the "co-existence" may occur already at the boundary of the anchi- and epi-zones.

On the basis of the X-ray diffractogram of the organic matter concentrate decarbonated and desilicified by HCl and HF the coal matter is in the graphite-d_2 (medium disordered graphite) state according to the graphite structural order-scale elaborated by LANDIS, 1971 (Fig. 11). This fact also proves that in addition to the microscopically well crystallized graphite the quantity of disordered anthracite is also remarkable.

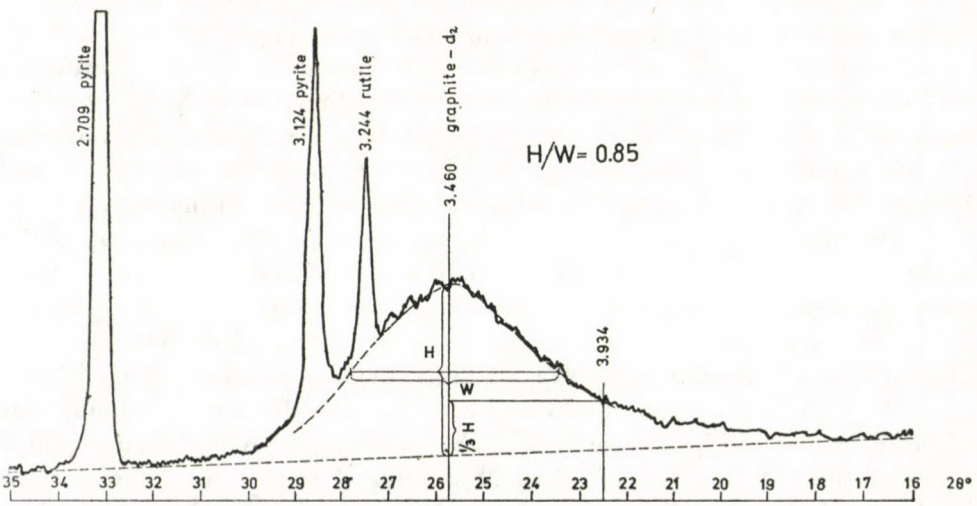

Fig. 11. Diffractogram of the dispersed organic matter of a meta-chert sample, obtained by HCl + HF treatment (Dédestapolcsány-8. bore 370.0 m). Characterization of the (002) reflexion of graphite according to LANDIS (1971)

Conclusions

In the Uppony Mountains of monoclinal folded structure the proportion of calcareous sediments gradually decreases while the quantity of fine-detrital siliceous sediments increases since the Middle Devonian up to the Middle Carboniferous. The equilibrated predominantly carbonatic sedimentation of the Middle Devonian Uppony Limestone Formation is replaced in the Upper Devonian—Lower Carboniferous Lázbérc Formation by the frequent alternation of calcareous-clayey-sandy sediments of different ratio. The increase in mobility of the geosyncline is indicated by the basic (?) tuffitic intercalations. In the Middle Carboniferous Tapolcsány Formation the total absence of calcareous sediments and the predominance of the detrital (mostly clayey subordinately silty, sandy) and siliceous sediments suggest the deepening of the sedimentary trough. The deep-faults related presumably to these movements generated the submarine volcanic activity manifested in several horizons and represented by predominantly basaltic lava and tuff.

Between the Middle Devonian and Middle Carboniferous formations continuous sedimentation can be assumed. No angular unconformity can be established between the individual formations, resp. between their parts.

On the basis of the illite crystallinity indices, vitrinite reflectance data, rock-textural features, mineral assemblages, presence of chloritoid, of the glauconite-celadonite—stilpnomelane isograde and of the anthracite-graphite "co-existence" the regional metamorphism of the Paleozoic formations of the Uppony Mountains proved to have occured at a temperature corresponding to the boundary of the anchi- and epi-zones (the boundary of the very low- and low-grade metamorphic grade; about 300 °C according to WINKLER, 1976).

The slight differences between the illite crystallinity values of different rock types indicate that recrystallization did not eliminate the primary differences of the illites of different degradation grade of the individual lithofacies.

No significant difference in the metamorphic grade has been found between the Middle Devonian and Middle Carboniferous formations.

This fact, as well as the continuous sedimentation, the deepening of the sedimentary trough and the lack of angular unconformity exclude the possibility to assign the regional dynamothermal metamorphism of the sequence to the Bretonian phase (Devonian-Carboniferous boundary) or to the Sudetan phase (Visean-Namurian boundary) of the Hercynian orogenic cycle.

Only slight differences could be stated between the metamorphic grades of the Paleozoic of Uppony and the anchi-zonal (pumpellyite-prehnite-quartz facies) Triassic formations of Bükk Mountains. For this reason, and since in the Bükk Mountains a sedimentary sequence is known characterized by continuous sedimentation from the Middle Carboniferous up to the Upper Triassic, it seems to be improbable that the metamorphism of the Uppony

Paleozoic would have been connected to some phase of the Hercynian cycle younger than Sudetan. It is more likely that the Igal-Bükk mobile belt including also the Paleozoic of the Szendrő, Uppony and Bükk Mountains, endured an Alpine dynamothermal metamorphic effect syngenetic with the folding related to the Cretaceous collision plate movements.

Because of the lack of erosional or angular unconformities the Hercynian recrystallization, if there had been any, might have been only of burial character and regarding its intensity, it might not exceed the metamorphic grade of the Alpine phase.

The low-medium transitional pressure range (2.5—3 kilobar) and the low pressure range (about 2.5 kilobar) established on the basis of b_0 lattice parameter distribution of sericites in the Szendrő resp. Uppony Mountains do not contradict to this fact since the pressure conditions of metamorphism depend on the tectonic régime and not on the geological age.

The low pressure metamorphism indicates that the part of the crust now situated southeast of the Darnó tectonic line did not take part in the Alpine subduction processes. The anchi-zonal metamorphism of the Triassic of the Bükk Mountains, the lack of the Jurassic or Cretaceous formations (resp. their continental character), or when their erosion is assumed, the small thickness of the Jurassic—Lower Cretaceous sediments require to take into consideration also nappe-forming processes contributing to the pressure and temperature conditions of the metamorphism.

On the basis of the petrological data of the Uppony Paleozoic it is probable that the Alpine metamorphic effect established earlier for the Triassic formations of the Bükk Mountains (Árkai, 1973) caused also the recrystallization of the Paleozoic basement of the Igal-Bükk mobile belt.

In spite of the conclusions drawn from the conodont investigations (Kozur and Mock, 1977) there is a difference between the metamorphic grades of the Paleozoic and Triassic formations. In the stratigraphic profile the metamorphic grade decreases from the Devonian to the Triassic from the low-medium pressure low temperature greenschist facies (low-grade metamorphism) to the low pressure anchi-zone (pumpellyite-prehnite-quartz facies, very-low-grade metamorphism). This difference, however, can be sufficiently explained by the depth differences of burial related to the stratigraphic position and in accordance with the statements of Kozur and Mock (1977) it does not support the assumption of a Hercynian dynamothermal metamorphism.

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МИНЕРАЛЬНЫЕ АССОЦИАЦИИ, ДАННЫЕ СТЕПЕНИ КРИСТАЛЛИЗАЦИИ ИЛЛИТА, $-b_0$ И СТЕПЕНИ УГЛЕФИКАЦИИ ПАЛЕОЗОЙСКИХ ПОРОД ГОР УППОНЬ (СВ ВЕНГРИЯ), ПОДВЕРГШИХСЯ ПЕРЕХОДНОМУ МЕТАМОРФИЗМУ НИЗКОЙ И ВЕСЬМА НИЗКОЙ СТЕПЕНИ

П. АРКАИ—З. А. ХОРВАТ—М. ТОТ

Резюме

Среднедевонские-среднекаменноугольные образования гор Уппонь, расположенных на северной границе Игало—Бюкской подвижной зоны, образуют моноклинальную складчатую структуру. Условия осадконакопления характеризуются постепенным уменьшением доли известковистых осадков, и в конечном счете их полным отсутствием, а также параллельным увеличением обломочных и химогенных кремистых осадков. Вулканическая деятельность, выявленная в нескольких стратиграфических горизонтах и создавшая в основном базальтовые лавы и пирокласти́ты, была обусловлена, по-видимому, глубинными разломами, приуроченными к опусканию осадочного бассейна.

Судя по минеральным ассоциациям, петроструктурным особенностям пород, степени кристаллизации иллита, метаморфозу глауконито/селадонито-стильпномеланового изограда, а также образованию хлоритоидов и сосуществованию антрацитов и графитов, региональный метаморфоз палеозойских отложений гор Уппонь произошел при температуре около 300°C, что соответствовало гравнице между анхи- и эпизональным метаморфизмом. Распределение параметра b_0 кристаллической решетки серицитов указывает на перекристаллизацию пород в условиях пониженного давления ($\bar{x} = 8.996$Å, $n = 34$, $s = 0{,}010$), то есть на около 2,5 килобар. Нет какой либо значительной разницы по степени метаморфизма между среднедевонскими и среднекаменноугольными отложениями. В связи с этим, а также ввиду беспрерывности осадконакопления и углового несогласия между ними исключена вероятность предположения герцинской, бретонской или судетской тектоно-метаморфических фаз. На основании степени метаморфизма других палеозойских и мезозойских пород рассматриваемой зоны можно предположить альпийский (меловой) возраст регионального динамометаморфизма.

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Rb–Sr ISOTOPIC STUDIES ON GRANODIORITIC ROCKS FROM THE MECSEK MOUNTAINS, HUNGARY

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New isotopic age determinations carried out with the rubidium-strontium method on granodioritic basement rocks from the Mecsek Mountains in Southeastern Transdanubia give further support to assumptions on a polyphase development of the Mecsek crystalline. As shown by initial Sr isotopic ratios, the protolith of the granodioritic assemblage of polymetamorphic-anatectic origin must have been strongly basic in its composition. Granitization processes commenced about 430 million years ago, thus the (sedimentary) protolith assemblage subjected to granitization must have been at least Lower Silurian in its age. The scatter of individual model ages obtained on total rock samples reflects the polymetamorphic-polytectonized character of the basement, and suggests that following a primary granitization process secondary events might have led to the total or partial recrystallization of the rocks in question. An event at about 335 million years (Early Carboniferous) seems to be of special importance, and is characterized by potash metasomatism having affected more or less the entire basement mass. The interpretation of this event as the onset of regional emergence is supported by the fact that biotite ages indicate a lowering of ambient temperature below the blocking temperature of the biotite Rb–Sr isotopic system.

The tectonic development of the basement crystalline as reflected in the isotopic age data closed at about 270 ± 20 million years with processes causing retrograde changes in the crystalline as a whole, but leading to the development of dyke rocks in the tectonically active parts of the basement mass.

Introduction

Age relationships within the crystalline basement of the Mecsek Mountains has been a matter of debate in Hungarian geology for a long time, with several questions still unsolved. Contradictions and inconsistencies among different opinions regarding age relationships within the lower structural stages of the area are especially significant in relation to the evolution in time of the processes leading to the anatectic development of the crystalline core of the mountain with a regional extension well exceeding the limits of surface occurrences of granitic-granodioritic rocks. The origin of debates traces back partly to different views about the way of formation of the granitic rocks in question. Similar weight should be given, however, to the fact that information obtained by different authors in different parts of the mountain, as well as the use of different sample material threw light on different aspects of the processes having influenced the rock assemblages of the crystalline area. Thus, each aspect gained different weight in the synthetizing work of individual authors.

In recent years the exploratory, research and mapping work carried out in the Mecsek area extended our knowledge upon the crystalline basement by a considerable amount. More accurate information than before is now available on the areal extent of the granitized region, as well as on its tectonic relationships both within the region and on a broader regional scale. In connection with the age relationships, however, no significant step forwards has been made. On the contrary, with accumulating information the number of open questions even increased. For example, because of the lack of adequate stratigraphic control, the age problem arises as a crucial question when possible genetic connections with metamorphic rocks of the surrounding areas are examined. Increased knowledge on the structural build-up of the basement, the discernment of its tectonically dissected character with units featuring even some lithological dissimilarities raised, for example the basic question; whether a regional heterogeneity in primary (protolith) age and/or granitization age could be assumed or not. Previous isotopic dating (OVCHINNIKOV et al., 1951; KOVÁCH—BALOGH—SÁMSONI, 1968) could not answer all these questions, which fact served as an impetus to try to collect additional age evidence by Rb-Sr dating based on a more extended sample material.

Views on Age Relationships in the Mecsek Crystalline

Although, since the discovery of SZÁDECZKY-KARDOSS (1959) — that the formation of the Mecsek granites should be explained by migmatization connected to regional metamorphism — there is an essentially unanimous agreement among different authors about the general character of the granitization processes in this region, views about the temporal development of the granitization process, as well as with respect to the age and character of the protolith assemblage are widely different. Evidently, these two questions are interrelated; the geological age of the protolith assemblage serves as a lower limit for the possible time of granitization. Similarly, assumptions (or data) on the granitization process itself should be interpreted as an upper limit for the possible age of the precursor sediments.

Previous assumptions (VADÁSZ, 1960) based still on an assumed analogy with the Velence Mountain granites regarded the Mecsek crystalline as a Hercynian intrusion. Later — already on the basis of the migmatitic derivation recognized by SZÁDECZKY-KARDOSS (1959) — several authors tried to compile a self-consistent picture on the igneous-metamorphic-tectonic development of the region. CSALAGOVITS (1964) regarded the Mecsek granitoides as products of two subsequent orogenic cycles. According to him, in the first period of granitization attached to the Caledonian orogeny synorogenic granodiorites were formed ("anatexis I"). During the Early Carboniferous a second

phase characterized by potassium metasomatism and by the intrusion of microgranitic and aplitic dykes closed the process of granitization ("anatexis II").

Similarly, GHANEM and RAVASZ-BARANYAI (1967) regard the crystalline basement rocks of the Mecsek Mountains as products of several superimposed processes distinctly separated in time. According to these authors, the marginal volcanic-sedimentary association of an assumed Precambrian geosyncline has undergone progressive metamorphism culminating in local migmatization still in the Precambrian. During the Hercynian orogeny, granitic melts ascending in the fissures of this (still basic) assemblage led to the development of the porphyroblastic character of the granodiorites, contemporaneously with the intrusion of aplitic dykes. They attribute the cataclastic fabric of the granodiorites to Hercynian movements. This evolutionary scheme thus regards the granitic-granodioritic rocks of the Mecsek Mountains as a polymetamorphic-polytectonic assemblage.

Based on the investigation of feldspars, BUDA (1969) classified the granodiorites of the region into two groups. Granitoid rocks of anatectic origin might have been derived from geosyncline sediments of Precambrian (?) age through amphibolite-grade regional metamorphism in the main phase of the Hercynian orogeny. Subsequently, in the late orogenic phase through potassium metasomatism the microcline-rich granodiorites, as well as aplite and microgranite dykes of late-kinematic character have developed.

The comprehensive work of JANTSKY (1974) based on detailed mapping and interpretation work took a standpoint quite different in assessing the time of granitization in the Mecsek Mountains. Classifying the basement formations into three tectonic stages, he regarded the granitoid rocks as of ultrametamorphic origin, forming a continuous genetic sequence with migmatitic rocks in the border zone of the central granodioritic mass. The ultrametamorphic assemblage including the granodioritic core with its aplitic dykes is treated as Lower to Middle Proterozoic in age, metamorphosed and even diaphtorized during the Riphean orogeny. According to his views, following the Riphean orogeny no further progessive metamorphism manifesting itself in newly formed mineral associations has ever occurred in the area. The effects of the Hercynian orogeny are regarded as insignificant ones, and out of the igneous rocks of the area only the serpentinites of the Görcsöny Ridge, subordinate aplitic dykes bordering the Carboniferous sedimentary areas and products of the (subsequent) Permian quartz porphyry volcanism are attributed to it.

SZEDERKÉNYI (1974a, b) divided the basement formations of the Mecsek Mountains and its surrounding areas on lithological and tectonic grounds. According to his summarizing conclusions, the composite basement consists of genetically and lithologically defined units delimited by Early Paleozoic structural lines. The oldest basement formation is represented by the meta-

morphites of the Görcsöny Ridge, the progressive series of which leads into the granitic mass of Mórágy with gradually increasing signs of anatectic granitization. The pre-metamorphic age of the metamorphic sequence and hence that of the protoliths of the granitized region might have been Precambrian or Lower Palaeozoic. He took no definite standpoint with respect to the age of granitization. In the Paleo-Hercynian tectonic substage characterized by rigid block tectonics regional emergence has occurred, accompanied by subordinate magmatic activity (development of microgranitic dykes and aplites) only. The igneous history of the basement is closed by the (Permian) quartz porphyry volcanism attributed to the Neo-Hercynian tectonic substage.

Recent findings (SZEDERKÉNYI, 1977) support previous assumptions that the Lower Paleozoic (?) phyllites of the Ófalu area should be regarded as the protolith of the granodioritic crystalline. This assumption is strongly supported by the discovery of crystalline limestones of the Ófalu sequence in the form of inclusions within the granitized rocks nearby (SZEDERKÉNYI, personal comm., 1978). If supported by further evidence, this finding would strongly speak against an assumed Precambrian age of granitization.

Isotopic age determinations carried out so far on the granitic rocks and their mineral constituents did not allow to draw unanimous conclusions with respect to the polyphase history of igneous development of the region. OVCHINNIKOV et al. (1961) obtained K-Ar ages between 335 and 260 million years on total rocks and biotites. KOVÁCH, BALOGH and SÁMSONI (1968) report a Rb-Sr isochron age of 284 $\pm$ 10 m. y. for biotites, and conclude that the last magmatic-tectonic event (except the Permian volcanism) has occurred in the Late Carboniferous. According to subsequent re-evaluation, however, the samples taken at that time as biotites contained also chloritic micas in considerable amount developed from biotite in a retrograde metamorphic process. Thus, the age attributed to them most probably refers to the last tectonic event accompanied by low-grade metamorphism in the given area. Determinations on total rock samples raised the possibility of assuming a previous event at about 440 m. y., as well. Some results seemed to support the conception of Precambrian descent, but the strongly altered character of these samples with exceptionally high model ages makes probable that the Precambrian ages obtained for them are accidental ones only, due to the redistribution of Rb and/or strontium.

In the light of the previous discussion it seemed highly desirable to carry out further measurements even on the reconnaissance level on carefully selected total rocks and collect additional isotopic data pertinent to the estimation of the time of granitization, which seems to be the crucial question in deciphering the metamorphic-tectonic development of the Mecsek crystalline. In the course of the present investigations we aimed at the determination of possible age limits for the regional metamorphism leading to granitization;

to establish whether there were any discernible differences in age between the individual phases of igneous activity in the Mecsek area, as well as between the individual units of the basement separated by distinct tectonic zones.

Experimental Methods

In determining the basic experimental data used in model age calculations (concentrations of Rb and Sr, isotopic ratios of strontium) mass spectrometric procedures were applied. Samples of 0.2 g weight were dissolved in platinum crucibles in the mixture of HF and $HClO_4$ on a hot plate. Following dissolution the dried samples were dissolved in a small amount of 3 n HCl. The solid residue observed in some cases consisted of zircon and some ore minerals (i.e. rubidium- and strontium-free phases) only, which were filtered off and were discarded.

Concentrations of Sr and Rb have been determined by stable isotope dilution (see e.g. SMALES and WAGER, 1960) in separate runs by adding highly enriched ^{87}Sr and ^{87}Rb spikes to the samples prior to dissolution. Pure strontium samples for mass spectrometry were prepared on Dowex 50 W x 12, 200—400 mesh cation exchange columns. Sr samples obtained in this way were converted into a nitrate form and loaded onto the side filaments of a three-band ion source with a rhenium central filament. All mass spectrometric measurements were carried out with an MI-1309 type mass spectrometer. Discrimination effects in the Sr isotopic ratio runs were corrected through normalizing to the $^{86}Sr/^{88}Sr = 0.1194$ reference ratio. In the course of the present series of measurements, errors in the $^{87}Sr/^{86}Sr$ ratios were in the order of 0.3%, the concentration values are correct within about 2%. The errors assigned to the individual model ages were calculated on the basis of the RMS errors of the underlying experimental data. In calculating the model ages the radioactive decay constant of ^{87}Rb was taken as $\lambda = 1.39 \cdot 10^{-11}\ yr^{-1}$.

Results and Discussion

The experimental isotopic data obtained by the methods sketched above are summarized in Table I together with the calculated model ages based on an initial $^{87}Sr/^{86}Sr$ ratio of 0.7055 (as for the justification of this value see later). Bearing in mind the reconnaissance character of the present work, the detailed description of the samples has been omitted, giving sampling localities only. No detailed petrographic investigation has been carried out on most of the samples all the more because we did not aim at discerning or disproving possible age relations between individual samples in details, but concentrated our efforts to obtain information on processes having affected the region in question as a whole.

Table 1

Compilation of analytical data

Serial No.	Locality or drilling	Rb μg/g	Sr μg/g	^{87}Rb μg/g	^{86}Sr μg/g	$^{87}Rb/^{86}Sr$ atomic ratio	$^{87}Sr/^{86}Sr$ atomic ratio	T m.y.
	A. Total rock samples							
1.	Erdősmecske, quarry ("inclusion")	266	299	74.0	28.9	2.57	0.7175±0.0012	336±52
2.	Erdősmecske, quarry ("inclusion")	316	302	87.9	29.2	3.02	0.7179±0.0012	295±46
3.	Erdősmecske, quarry ("inclusion")	279	267	77.7	25.8	3.02	0.7185±0.0012	309±45
4.	Erdősmecske, quarry	226	419	62.9	40.5	1.56	0.7159±0.0026	479±117
5.	Erdősmecske, quarry	199	407	55.4	39.4	1.41	0.7131±0.0029	388±164
6.	Erdősmecske, quarry	340	381	94.6	36.8	2.58	0.7175±0.0028	334±88
7.	Erdősmecske, quarry	230	310	64.0	30.0	2.15	0.7163±0.0012	361±42
8.	Mórágy, municipal quarry	138	198	38.4	19.1	2.01	0.7157±0.0038	365±153
9.	Mórágy, municipal quarry	164	201	45.6	19.4	2.36	0.7160±0.0013	320±59
10.	Alsónána-1; 235.0 m	238	388	66.2	37.5	1.77	0.7137±0.0010	333±32
11.	Mágocs-1; 369.5 m	286	210	79.6	20.3	3.92	0.7116±0.0015	295±42
12.	Újmecsekalja 1; 114.0 m	225	618	62.6	59.8	1.06	0.7100±0.0023	306±157
13.	Szalatnak-3; 235.0 m	226	555	62.9	53.7	1.17	0.7101±0.0017	319±135
14.	Szalatnak-3; 563.0 m	157	438	43.7	42.4	1.03	0.7108±0.0026	377±210
15.	Nyugatszenterzsébet-1; 27.0 m	309	220	86.0	21.3	4.05	0.7278±0.0021	396±60
16.	Nyugatszenterzsébet-1; 81.6 m	273	222	76.0	21.5	3.55	0.7248±0.0030	391±70
17.	Gyűrűfű-9800; 298.0 m	232	464	64.6	44.8	1.44	0.7114±0.0015	295±73
18.	Korpád-9006; 682.0 m	168	147	46.8	14.2	3.30	0.7183±0.0010	279±27
19.	Korpád-9015; 520.0 m	283	148	78.8	14.3	5.54	0.7282±0.0012	295±14
	B. Results obtained biotites							
20.	Erdősmecske, quarry (4)	725	32.8	202	3.176	63.73	1.0156±0.0043	352±25
21.	Erdősmecske, quarry (4)	733	14.0	204	1.356	150.84	1.4701±0.0085	365±41
22.	Erdősmecske, quarry (6)	596	10.4	166	1.010	164.69	1.4496±0.0086	325±38
23.	Szalatnak-3; 563. m (14)	388	51.1	108	4.900	22.10	0.7752±0.0009	227±59
	C. Total rock samples — dyke rocks							
24.	Erdősmecske, quarry (felsite)	469	10.4	131.5	1.010	139.60	1.3424±0.0013	354±25
25.	Erdősmecske, quarry (aplite)	212	124	59.0	12.0	4.94	0.7273±0.0015	317±31
26.	Erdősmecske, quarry (aplite)	244	200	67.9	19.3	3.52	0.7206±0.0029	309±60
27.	Nyugatszenterzsébet-2; 35.4 m (aplite)	271	88.8	75.4	8.05	8.61	0.7364±0.0013	252±25
28.	Helesfa-2; 275.0 m (microgranite)	316	177	87.9	17.1	5.16	0.7299±0.0022	340±34
29.	Pellérd-N-2; 14.4 m (microgranite)	434	13.4	120.8	1.25	93.64	1.0838±0.0063	291±44
30.	Pellérd-N-4; 18.0 m (microgranite)	416	17.3	115.8	1.67	69.49	0.9655±0.0024	269±40

The first part of Table I (Nos 1—19) shows results obtained on total rock samples ranging in composition from granodiorite to syenite porphyry. Samples listed under No. 20—23 are biotite samples separated from total rocks, the number of which is given in parantheses and refers to the serial number of the corresponding total rock sample. The third part of Table I (Nos 24—30) lists measurements carried out on dyke rocks (microgranites and aplitic dykes).

In order to facilitate an easy comparison of isotope systematics in dyke rocks and their host granites the results are plotted in Figs 2 and 1, respectively, in a Nicolaysen diagram (NICOLAYSEN, 1961). It is conspicuous at the first sight that in the case of the Mecsek rocks the isochron concept is not applicable to either assemblage on a regional scale. It is remarkable enough, however, that the points representing the measured isotopic ratios occupy a wedge-shaped field in the $^{87}Sr/^{86}Sr$—$^{87}Rb/^{86}Sr$ diagram.

It is quite reasonable to assume that the protolith assemblage of the granitized region could not have had large regional inhomogeneities with respect to the isotopic composition of strontium, due to its basic character. Thus, the distribution as seen in Fig. 1 should be attributed to secondary effects having destroyed any isochrone arrangement developed in the course of a primary granitization-ultrametamorphic process, i.e. which acted as

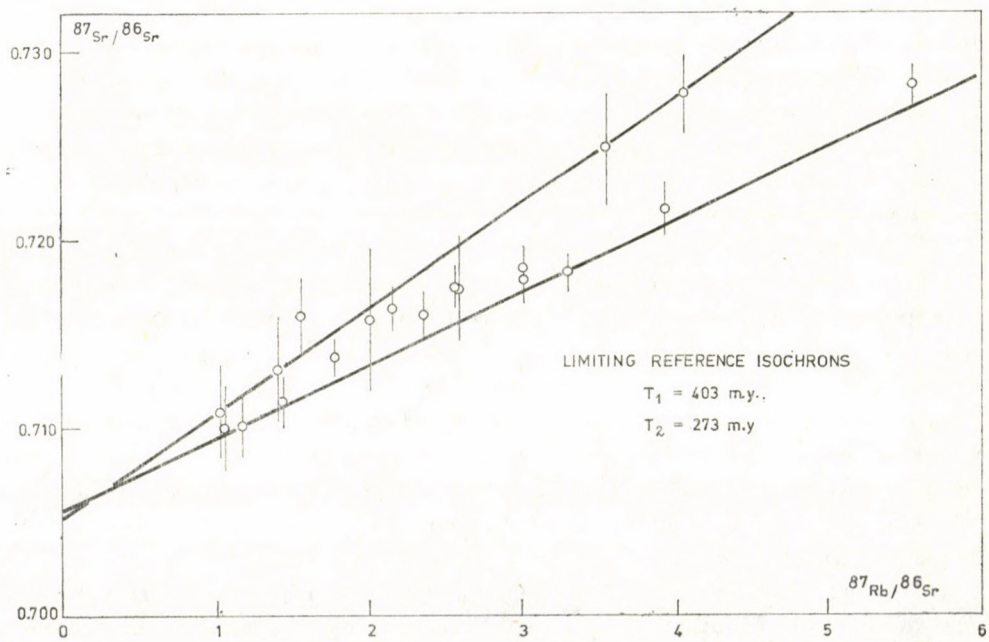

Fig. 1. Plot of rubidium and strontium isotopic data for the Mecsek Mountains granodiorites (dyke rocks excluded, data obtained on total rock samples)

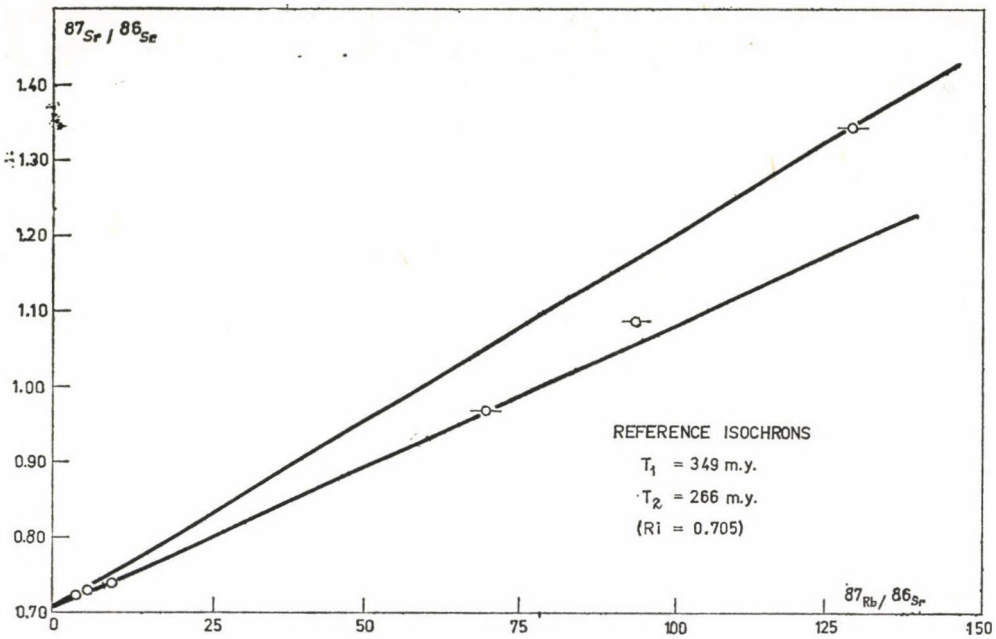

Fig. 2. Rb—Sr isochron plot of dyke rocks from the Mecsek Mountains (aplites and microgranites, data obtained on total rock samples)

processes breaking the closed system character of the granitized assemblage and enabling the migration of rubidium and/or strontium. The measured distribution of sample points in the diagram of Fig. 1 however, can be delimited by two straight lines, which if regarded as reference isochrons might supply upper and lower limits in time for these secondary processes which led to the (partial) redistribution of Rb and Sr.

A similar distribution has been observed by HURLEY and FAIRBAIRN (1970) in orogenic rocks of the Northern Appalachian province. These authors concluded that this distribution is the result of several distinct, superimposed intrusive or metasomatic events, time limits for which should be given by ages deduced from the limiting isochrons. The strictly wedge-like distribution assures at the same time that initial $^{87}Sr/^{86}Sr$ ratios attributable to the individual samples (or to isochrons hidden in the general distribution and representative for any subsystem within the orogene) are quite similar, so the individual model ages based on this assumed common initial ratio are directly comparable with each other.

In the present case the two limiting reference isochrons of Fig. 1 yield ages of 403 and 273 million years, respectively, which according to the previous way of interpretation supply age limits for the orogenic development of the region, at least as far as it is manifested in the Rb-Sr isotopic system of the

granitized mass. (The concept of "orogenic development" is taken here, of course, as a series of events causing and reflected in the migration of Rb and/or Sr.)

The coincidence of the younger age limit deduced in this way with the age value given by KOVÁCH et al. (1968) and interpreted now as the age of the last (retrograde) metamorphic event is quite conspicuous, and supports the previous conclusion that following regional emergence during the Late Carboniferous the basement mass responded to all subsequent stresses by fracture tectonics only.

The limiting reference isochrons supply initial $^{87}Sr/^{86}Sr$ ratios close to each other, thus in calculating the individual model ages as listed in Table I the average of these two values (0.7055) has been used as reference. The actual value of the common initial ratio deduced in this way is in compliance with the assumption that the granitoid rocks of the mountain were developed in the course of the ultrametamorphic granitization of a basic volcanic-sedimentary geosyncline assemblage.

The interpretation of the second time limit of about 400 million years deduced from the slope of the upper reference isochron is by far not so simple and unambiguous. From the general characteristics of the process of granitization follows that eventual local inhomogeneities present in the protolith might have been "smoothed out" by migration phenomena exceeding in their scale the sample dimensions, thus the age supplied by the upper limiting isochron should not be attributed to the protolith, but rather to the granitization process itself. If the upper part of the wedge-shaped distribution comprised samples whose model ages would have been increased by chance throughout the subsequent events, instead of a limiting isochron a "diffuse" upper limit would be expected. Thus, we attribute real significance to the age deduced from the upper limiting reference isochron, and ascribe it to the declining period of a "primary" granitization process, i.e. to that time when following thermal activization the geochemical systems represented by our samples became for the first time closed systems with respect to the migration of Rb and Sr. According to this interpretation, the granodioritic suite of the Mecsek Mountains started to develop not later than 400 million years ago. It follows, of course, that the protolith assemblage of the migmatitic complex must have been not younger than Silurian, allowing ample time for subsidence.

It should be mentioned here, that similar distributions of isotopic ratios are known for regions which have undergone multiple igneous, metamorphic and tectonic reactivation following a primary granitization process. We refer here, for example, to results of age determinations carried out on granitic-granodioritic rocks of the "Eastern basement" of New England in Massachusetts, U.S.A. (KOVÁCH, HURLEY and FAIRBAIRN, 1977) where the reality of age deduced from the upper limiting reference isochron of a similar wedge-

shaped distribution has been confirmed by a real isochron age obtained on samples from a single locality.

The actual time of primary granitization might be somewhat higher than the previously determined age. Another approach would be to compare results obtained on samples from a single locality. In this respect, samples from the drilling Nyugatszenterzsébet-1 seemed to be the most promising, being characteristic of granites least affected by secondary events. The individual model ages based on an initial Sr isotopic ratio of 0.7055 are 391 and 396 million years for the two samples respectively. Assuming a co-genetic origin (to which the petrographic similarity of the samples and their spatial proximity gives, a sound basis) the common isochron age of the two samples turns out to be 432 $\pm$ 30 million years with an initial ratio of 0.7035. This initial ratio already corresponds to actual values found in tholeiitic basalts of the oceanic realm, supporting again previous assumptions on the geosyncline origin of the protolith suite. At the same time, bearing in mind the actual Rb/Sr ratios of the samples, this low value rules out the possibility that the observed common initial ratio is the result of a rejuvenation event, because in such a case necessarily higher $^{87}Sr/^{86}Sr$ initial ratios would have been developed. Thus, we conclude that primary granitization in the Nyugatszenterzsébet area might have commenced not sooner than about 430 million years ago, i.e. not sooner than during the Early Silurian. Accordingly, the primary age of the protolith assemblage might be Lower Silurian at least. Allowing ample time for subsidence to a level corresponding to the thermodynamic conditions of the amphibolite facies necessary for granitization, a somewhat older age (Cambrian-Ordovician, or even Precambrian?) is highly probable.

The next discernible event following the process of primary granitization is reflected by the isotopic ages measured on biotites from granodioritic-granitic rocks of the Erdősmecske quarry, with an average of 347 $\pm$ 24 million years. The reality of this age is supported by K/Ar data obtained on the same samples. The average of the K/Ar ages is 334 $\pm$ 11 million years (BALOGH, K. and ÁRVA-SOÓS, E., personal communication). These concordant data demonstrate unambiguously that following this datum no regional metamorphism has occurred in the Mecsek area and that the ambient temperature has not exceeded about 300 °C on a regional scale.

The significance of the event around 335 m.y. gains further support through the average Rb-Sr model age of 333 $\pm$ 37 million years obtained on total rocks from the Erdősmecske area. Members of the "older" generation of dyke rocks supplied similar ages: for a felsitic sample from Erdősmecske 354 $\pm$ 25 m.y., for the Helesfa microgranite 340 $\pm$ 34 m.y. ages have been obtained. Ages of aplitic rocks from the Erdősmecske quarry also exceed 300 m.y. All these lead to the obvious conclusion that at about 335 m.y. (i.e. during the Early Carboniferous) an important event characterized by

tectonic activization, increase of temperature on a regional scale, as well as by the intrusion of dyke rocks has affected the granodioritic basement of the Mecsek Mountains. This event caused the "rejuvenation" of the granodioritic rocks developed during the first stage of granitization, producing the "mixed" model ages characteristic for the Mecsek area. Since this event resulted in the pronounced migration of Rb and/or strontium as well, it is highly probable that the development of porphyroblastic microcline typical for the rocks supplying transitional ("mixed") model ages is to be attributed to potash metasomatism accompanying this event of activization.

It looks somewhat curious that in spite of the concordant Rb/Sr and K/Ar ages clustering around 335 m.y. for biotites, several total rock samples yielded model ages well below this datum down to 270 million years. Similarly, in the group of dyke rocks a "young" generation also occurs. The lowest ages correspond to the age determined by Kovách et al. (1968) on chloritic biotites, and coincide with the lower limit of K/Ar ages determined on total rock samples (Ovchinnikov et al., 1961).

A suitable key for the interpretation of these data is given by the fact, that these "young" total rock ages have been obtained on samples which either originate from the tectonized border zone of the basement crystalline with its highly fractured and/or mylonitized rocks (Újmecsekalja-1, Mágocs-1) or represent dyke rocks intruding the mylonite zone. Moreover, some of the samples stem from the tectonically dissected block of the Western Mecsek affected also by extensive volcanic activity during the Early Permian (drillings No. 9800, 9006, resp. 9015). Thus, recurrent tectonic activity during the Sudetian-Asturian orogenic phase in the Late Carboniferous leading to regional emergence and dissection into separate blocks of the Mecsek crystalline should be considered as the cause of both the extensive retrograde metamorphism observed on a regional scale and the local rejuvenation/partial recrystallization of some rocks in the tectonically most active zones, as well as of the intrusion of some dyke rocks under tensional stress conditions.

Without going into details of large-scale correlation we should like only to mention here that the age values attributed to the main periods of activization in the Mecsek crystalline as deduced in the previous discussion correspond well to isotopic ages obtained on crystalline rocks of the Northern and Eastern Carpathians (cf. Bojko, 1975).

Conclusion and Summary

Confronting information deduced from isotopic ages as interpreted in the present paper with other evidences on the development in time of the granitoid rocks of the Mecsek region, the following evolutionary scheme might

be constructed for the main events shaping the crystalline basement of the Mecsek mountains.

Granitic-granodioritic rocks of the Mecsek crystalline have been developed through the polyphase granitization of geosyncline rocks probably of Lower Paleozoic age. The evolution of the granitization process as reflected by the isotopic age relationships started in the Silurian (430—400 m.y., Taconian orogeny?) with several periods of reactivation during the Paleozoic era. During the Early Carboniferous (335 m.y.) the granodioritic basement has undergone extensive changes in the course of progressive processes characterized by migration phenomena in a volatile-rich environment (metasomatism), by the recrystallization of mineral phases incl. biotite, and by the development of dyke rocks during intensive tectonic activization attributed to the Bretonian orogenic phase. Following regional emergence, during the Late Carboniferous in the tectonically most affected regions of the crystalline mass already dissected into separate blocks some granitoide rocks have been locally recrystallized/rejuvenated in the course of a renewed tectonic activization accompanied by subordinate dyke intrusions as well. This reactivation, however, caused only retrograde mineral changes in the crystalline mass as a whole (270—285 m.y.).

This evolutionary scheme based on isotopic age data and geological evidence shows thus the gradual consolidation of the crystalline basement of the Mecsek Mountains throughout the entire Paleozoic, reaching final consolidation with the ceasing of acid volcanism of final character in the Lower Permian.

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ИССЛЕДОВАНИЯ ГРАНОДИОРИТОВ ИЗ ГОР МЕЧЕК (ВЕНГРИЯ) ИЗОТОПНЫМ МЕТОДОМ Rb—Sr

Э. ШВИНГОР—А. КОВАЧ

Резюме

Новые определения возраста гранодиоритовых пород фундамента в горах Мечек на юго-востоке Задунайского края, проведенные методом рубидия-стронция, подтвердили предположение о многофазовом развитии мечекских кристаллических пород. Как показывают исходные изотопные соотношения стронция, протолит в гранодиоритовой ассоциации минералов полиметаморфического-анатектического происхождения был, по-видимому, сильно основным по составу. Процессы гранитизации начались примерно 430 млн лет назад. Таким образом, сообщество (осадочных) протолитов, подверженных гранитизации должны иметь минимум нижнесилурийский возраст. Рассеивание отдельных модельных возрастов, полученных для образцов горной породы в целом, отражает полиметаморфический-политектонический характер фундамента и указывает на возможность такого варианта, что после перивчной гранитизации вторичные события могли привести к полной или частичной перекристаллизации рассматриваемых пород. Событие, имевшее место примерно 335 млн лет назад (ранний карбон), имело по всей вероятности особое значение и характеризуется калиевым метасоматозом, которому более или менее подвергалась вся масса фундамента. Интерпретацию этого события как наступления регионального поднятия поддерживает факт, что полученные для биотита возрастные данные указывают на понижение окружающей температуры ниже «блокирующей» температуры изотопной системы Rb—Sr биотита.

Тектоническое развитие кристаллического фундамента по изотопным возрастным данным завершилось примерно 270 ± 20 млн лет назад процессами, вызвавшими ретроградные изменения в кристаллическом фундаменте в целом, но вместе с тем приведшими к образованию даек в тектонически активных частях массы фундамента.

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GENESIS OF THE HUNGARIAN GRANITOID ROCKS

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The granitoid rocks of Hungary classified into two main groups. The closed association with regional metamorphic rocks and lower temperature mineral assemblage of granitoids indicate an origin by anatexis and K-metasomatism in South Hungary. The granitoids of Balaton—Velence fault zone are characterised by higher temperature mineral assemblage with contact-metamorph aureole indicate magmatic origin.

The granitoids have been classified into two main groups according to mineralogical and petrological characters, tectonic occurrences and age of origin.

I. Synkinematic anatexite and latekinematic K-metasomatic granitoids (kata-mesozone).

II. Postkinematic quartz-diorites and granites (epi-zone).

I. *The Synkinematic Anatexite and Latekinematic K-metasomatic Granites*

These rocks are situated in South Hungary (Fig. 1). Large part of this granitic and metamorphic belt, except of the Mecsek Mountains, is covered by thick sediments. The granitoids can be divided into three subgroups according to the mineralogical and chemical characters.

a) Basites occur as enclaves in the synkinematic granodiorites, in and around Mecsek Mountains, Danube—Tisza Interfluve and Békés basin. They are rich in biotite, amphibole and plagioclase (An_{35}). Microcline is very rare and occurs as a late crystallized mineral. Q, Or, Ab norms are between 55—70% (Table 1). The rocks are rich in Mg and Ca (Fig. 2).

b) Synkinematic porphyroblastic granodiorite occurs in and around Mecsek Mountains, Danube—Tisza Interfluve and Békés basin. The anorthite content of plagioclase is high in the granodiorite of Mecsek Mts. The relatively disordered structure and predominant complex-twins of plagioclases indicate a higher temperature of formation. The anorthite content decreases from Mecsek Mts. towards the area surrounding of Mecsek Mountains and Danube—Tisza Interfluve. The decrease of anorthite content is followed by a higher degree

Table 1

South Hungarian granitoids and associated rocks

Basic-Enclaves or Resistites
(Q + Ab + Or = 55—70%)

Occurrence	Normative ratios (CIPW)						Cation ratios (NOCKOLDS)									Ab/An	No. Sampl. + total
	Q	Ab	Or	An	Ab	Or	Total Fe	Total Alk.	Mg	Total Fe	Mg	Ca	K	Na	Ca		
Mecsek Mountains	18.7	34.4	46.9	17.3	35.1	47.6	23.2	49.4	27.4	33.5	39.1	27.4	39.9	32.7	27.4	2.7	10
Surroundings of Mecsek Mts.	18.0	39.2	42.8	11.9	44.1	44.0	19.9	55.3	24.8	28.7	34.9	36.4	32.5	36.4	31.1	5.5	5
Danube—Tisza Interfluve	38.1	30.4	31.5	17.8	40.4	41.8	20.0	59.3	20.7	34.9	36.0	29.1	38.5	39.5	22.0	2.3	1
Békés basin	33.5	40.5	26.0	20.3	47.8	31.9	23.8	58.5	17.7	39.2	28.6	32.2	29.1	45.8	25.1	2.9	8
Average	27.0	36.2	36.8	16.8	41.9	41.3	21.7	55.6	22.7	34.1	34.7	31.2	35.0	38.6	26.4	3.3	24
Synkinematic anatexite (Granodiorites) (Q + Ab + Or = 70—80%)																	
Mecsek Mountains	34.6	31.7	33.7	13.4	41.4	45.2	18.8	61.9	19.3	36.3	37.5	26.2	41.9	40.7	17.4	3.6	7
Surroundings of Mecsek Mts.	19.3	42.0	38.7	10.2	46.1	43.7	16.8	66.9	16.3	36.7	35.5	27.8	38.5	45.7	15.8	4.5	16
Pécs-7	32.5	39.4	28.1	10.2	53.0	36.8	19.8	60.9	19.3	39.2	38.4	22.4	33.9	50.9	15.2	8.2	4
Danube—Tisza Interfluve	46.2	25.8	28.0	6.4	45.0	48.6	27.6	63.3	9.1	58.7	19.0	22.3	43.4	42.4	14.2	8.2	6
Békés basin	38.1	39.2	22.7	9.9	56.4	33.7	17.7	69.4	12.9	41.7	30.5	27.8	30.8	54.6	14.6	9.5	11
Average	34.1	35.7	30.2	10.0	48.4	41.6	20.1	64.5	15.4	42.5	32.2	25.3	37.7	46.9	15.4	6.8	44

Latekinematic K-metasomatic granites
(Q + Ab + Or = 80—100%)

Mecsek Mountains	37.4	28.7	33.9	7.3	42.5	50.2	9.0	86.8	4.2	42.2	22.1	35.7	48.3	43.3	8.4	8.8	8
Surroundings of Mecsek Mts.	29.1	34.2	36.7	2.4	47.4	50.2	8.9	83.4	7.7	41.9	31.1	27.0	43.4	49.5	7.1	21.4	10
Pécs-7	35.3	31.2	33.5	6.0	45.6	48.4	9.6	80.3	10.1	33.8	35.9	30.3	45.3	45.1	9.6	8.2	2
Danube—Tisza Interfluve	38.7	30.5	30.8	5.4	46.3	48.3	8.9	87.6	3.5	47.2	18.1	34.7	45.8	46.5	7.7	16.7	10
Békés basin	36.5	32.9	30.6	7.3	48.0	44.7	11.7	82.0	6.3	41.0	23.0	36.0	41.8	47.6	10.6	7.2	7
Average	35.4	31.5	33.1	5.7	45.9	48.4	9.6	84.0	6.4	41.2	26.0	32.8	44.9	46.4	8.7	12.4	37

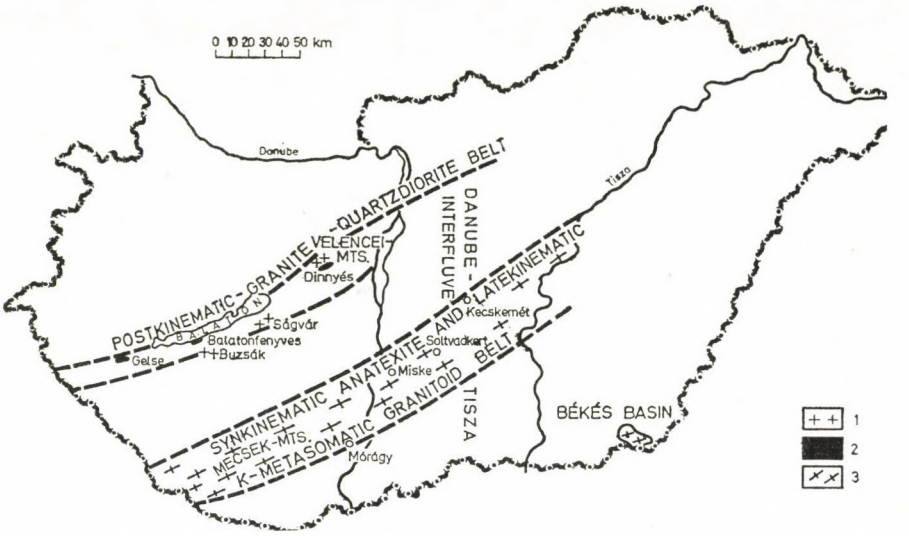

Fig. 1. Map showing the granitoid occurrences of Hungary. 1. Granite and granodiorite; 2. quartzdiorite; 3. anatexitic granodiorite and granite

of ordering and decreasing amount of complex-twins indicating a lower temperature of formation.

Porphyroblastic potassium feldspar is common constituent with more or less ordered structure (maximum or nearly maximum triclinicity). They are rich in inclusions of biotite and relict plagioclase and sometimes show cross-hatched twins. The crystallization temperature of microcline and plagioclase was about 550—590 °C (at 5 kb) counted according to WHITNEY and STORMER's method (1977). The K-feldspars occur in the groundmass too. Their triclinicities differ from the porphyroblasts. In Mecsek Mts. the triclinicity is medium in groundmass and maximum in porphyroblast. In Danube—Tisza interfluve microcline has higher degree of triclinicity in groundmass and porphyroblast has the same degree of ordering than in Mecsek Mts. Around Mecsek Mountains the double triclinicities are rare (average $\Delta = 0.65$). Perthite of K-feldspar is coarser in granodiorite of Mecsek Mountains than in the Danube—Tisza Interfluve.

Sum of Q, Or, Ab norms is between 70—80% (Table 1). The rocks are characterized by high alkali content (Fig. 2).

c) Latekinematic granites occur as dykes and small batholiths in the synkinematic granodiorite, they are pinkish, fine-grained, equigranular and rarely contain biotite. Perthitic microcline of high triclinicity ($\Delta = 0.90$) is common. The plagioclase has an ordered structure and its higher Ab content indicates a low temperature of formation (480 °C). The cross-hatched microcline pseudomorphs after plagioclase suggests K-metasomatic origin.

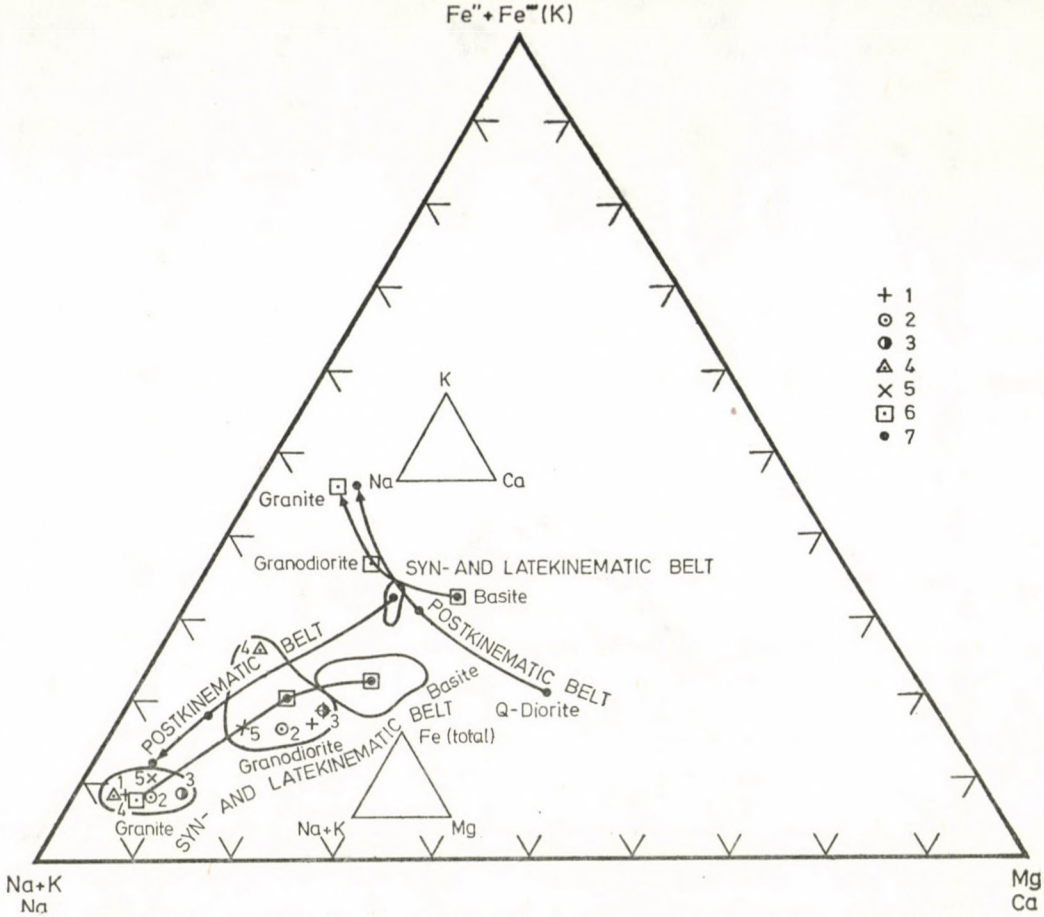

Fig. 2. Plots of K—Na—Ca and $(Fe^{2+} + Fe^{3+})$—(Na + K)—Mg. 1. Mecsek Mountains 2. surroundings of Mecsek Mts.; 3. Pécs-7 (borehole); 4. Danube—Tisza Interfluve; 5. Békés basin; 6. average (South Hungary); 7. average (Balaton—Velence fault-zone)

Sum of Q, Or, Ab norms is between 80—100% (Table 1). The rocks are rich in alkalies due to the alkali metasomatism (Fig. 2).

The synkinematic granodiorites of Mecsek Mountains are rich in *An* norms (Table 1), whereas granodiorites of Danube—Tisza Interfluve and Békés basin (Table 1) are rich in Q norms indicating that the former crystallized under higher pressure then latter.

The higher *An* content of the latekinematic granites of Mecsek Mts. and high Q content of granites of Danube—Tisza Interfluve indicate a relative higher pressure and temperature conditions in the Mecsek Mts. compare with the Danube—Tisza Interfluve (Table 1).

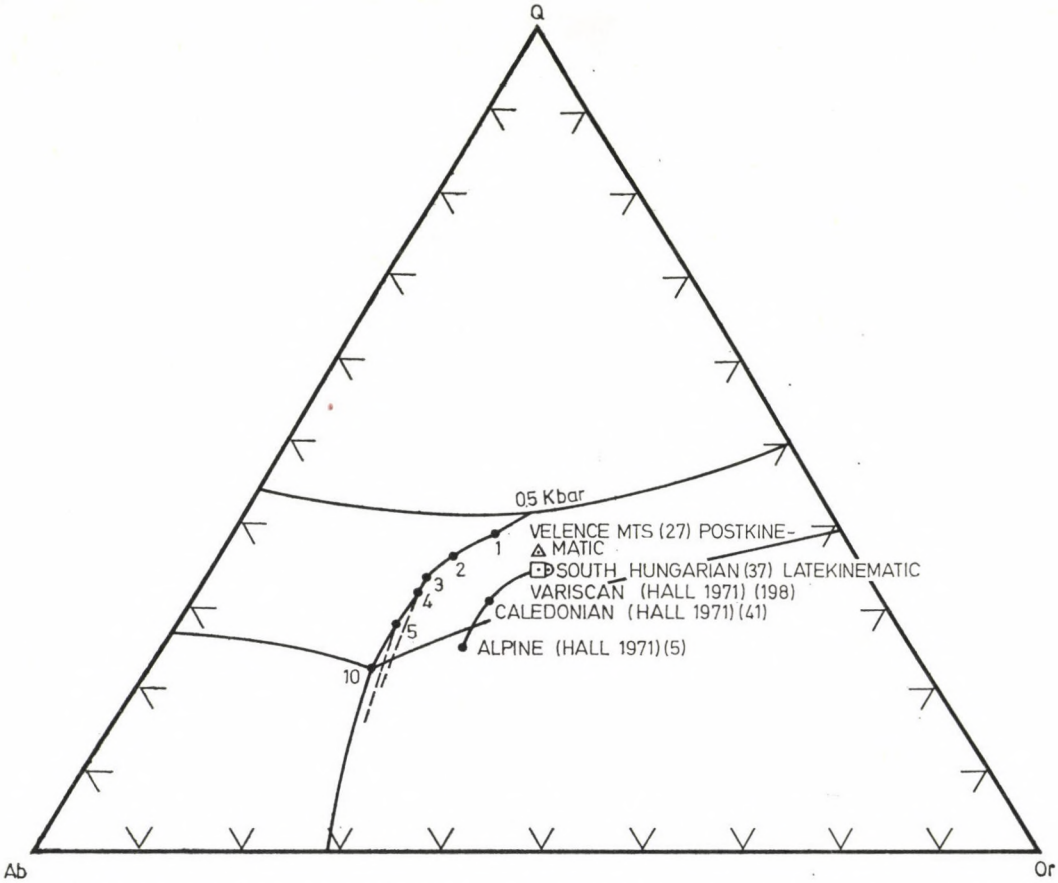

Fig. 3. **Average composition of granites of different ages in the system Q—Or—Ab—H_2O. Points indicate isobaric minima at water pressure between 0.5 and 10 Kbar. (after: TUTTLE and BOWEN, 1958 and LUTH et al., 1964). The figures in brackets indicate the number of samples**

The average Q, Ab, Or norms of the granites are strikingly similar to the West European Variscan granites (HALL, 1971, 1972, Fig. 3), thereby suggesting similar age of origin. This is also supported by the isotopic age determination (KOVÁCS etc., 1968).

Petrogenesis

An eugeosyncline had been filled up with sediments and basic volcanites which were regionally metamorphosed. The deepest part of the geosyncline subjected to anatexis along the belt of Mórágy—Miske—Soltvadkert—Kecske-

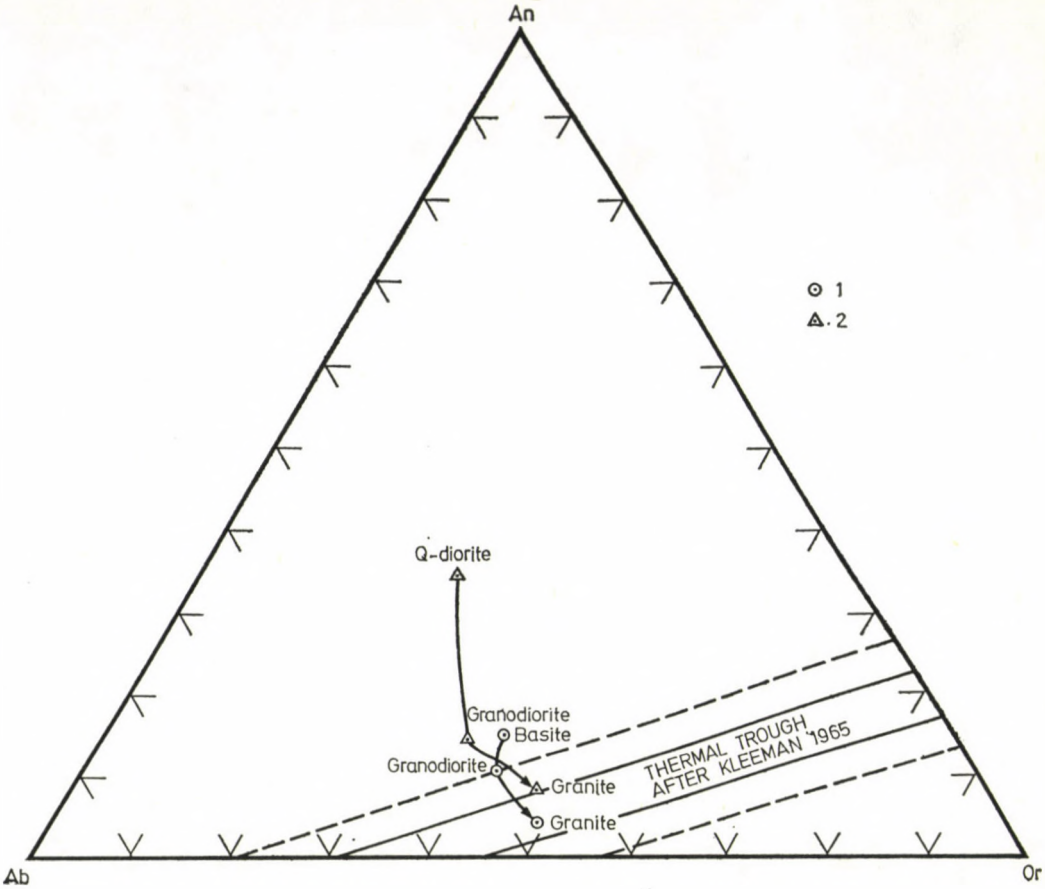

Fig. 4. **An—Ab—Or proportions of Hungarian granitoid rocks. 1. Average composition of South Hungarian granitoids; 2. average composition of granitoids occurring along the Velence—Balaton fault zone**

mét. The temperature and pressure conditions prevailed during the anatexite formation correspond with the amphibolite facies. This synkinematic anatexis took place in a closed system with the presence of volatiles, at elevated temperature and pressure caused by the subsidence of sediments.

Further subsidence of the rigid synkinematic rocks (anatexite granites) caused K mobilization in the groundmass which migrated to the newly formed fissures. The intensity of mobilization was greatest in the deepest parts of the geosyncline. This K-migration transformed the granite into granodiorite in the Mecsek Mts. The transformation of granite into granodiorite in the Danube—Tisza interfluve was incomplete due to lower intensity of K-migration.

Table 2

Postkinematic intrusives of Balaton—Velence fault zone

Q-diorites

Occurrence	Normative ratios (CIPW)						Cation ratios (NOCKOLDS)									Ab/An	No. Sampl. + total
	Q	Ab	Or	An	Ab	Or	Total Fe	Total Alk.	Mg	Total Fe	Mg	Ca	K	Na	Ca		
Boreholes (1. 2.)*	25.6	45.5	28.9	35.6	39.4	25.0	29.1	50.5	20.4	34.1	23.8	42.1	21.9	36.5	41.6	1.1	4
Velence Mountains	26.8	42.7	30.5	16.8	47.8	35.4	34.2	46.4	19.4	51.4	28.0	20.6	31.6	45.1	23.3	3.7	2
Granodiorites																	
Buzsák (boreholes)	35.6	36.3	28.1	15.3	47.8	36.9	18.2	62.2	19.6	35.2	37.9	26.9	34.5	47.2	18.3	3.2	2
Velence Mountains	37.4	32.6	30.0	11.6	42.5	45.9	18.5	73.2	8.3	48.2	23.3	28.5	44.1	43.1	12.8	3.2	6
Granites																	
Velence Mountains	37.2	31.3	31.5	8.7	45.2	46.1	12.4	82.2	5.4	49.0	18.9	32.1	44.4	46.3	9.3	7.4	27

* 1. Dinnyés; 2. Gelse

The mobilization of K, Na, Si etc. produced the late-kinematic K-metasomatic microgranites and aplites which filled up the fissures of the granodioritic body.

II. *Postkinematic Quartz-diorites and Granites*

Two igneous rock types occur along the Balaton—Velence fault-zone.

a) Quartz-diorite (tonalite) has been reached in the boreholes at Dinnyés, Gelse and also found as enclaves in the granite of the Velence Mts. (Fig. 1). Anorthite rich plagioclase (An_{50}) is the predominant constituent with zoned texture. Microperthitic orthoclase is rare. Quartz sometimes shows wavy extinction. Biotite is mostly chloritized. Hornblende is rather common and zircon, titanite, apatite and rutile occur as accessory minerals.

Chemical differences have been observed between the quartz-diorite from boreholes (Dinnyés, Gelse) and the enclaves of the Velence Mts. The enclaves are rich in Or and poor in An norms indicating K-metasomatism caused by the huge mass of K-rich granitoid pluton (Table 2).

b) Granite-granodiorite intrusions have been studied in Velence Mts., Ságvár and Buzsák areas (Fig. 1). The average anorthite content of plagioclases are similar (An_{30}). Complex-twins are prevailing and Al/Si ordering is medium degree.

The K-feldspar has disordered structure. The temperature of formation was about 630—650 °C counted according to I. C. STROMER's (1975) method.

The average chemical composition of these rocks are granitic, but minor deviation has been observed. The granodiorite occurs in the borehole of Buzsák and in the margin of the granitic pluton of Velence Mts. (Table 2).

Granites of the Velence Mts. are rich in Q and poor in An norms (Table 2) indicating lower pressure conditions of formation compared with the South Hungarian latekinematic granitoids (Fig. 3 and Fig. 4). Their age, determined by isotopic (Sr/Rb) method, is Variscan (310—330 million years). They are chemically very similar to the West-European Variscan granites (Fig. 3).

There are no direct evidences to establish the relationship existing between quartz-diorites and granites, however, enclaves in the Velence granites suggest that the earlier intrusion was quartz-diorite and it was followed by granodiorite-granite.

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К ГЕНЕЗИСУ ВЕНГЕРСКИХ ГРАНИТОИДОВ

ДЬ. БУДА

Резюме

Гранитоиды Венгрии относят к двум главным группам. Приуроченность гранитоидов к регионально метаморфизованным породам и их низкотемпературный минеральный парагенез на территории Южной Венгрии свидетельствуют об образовании в результате анатектических и калиевометасоматических процессов. Гранитоиды зоны разломов Балатон—Веленце характеризуются минеральным парагенезом повышенной температуры образования с контактово-метаморфическим ореолом, что свидетельствует о магматическом происхождении пород.

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THE GEOLOGY AND CHEMISTRY OF THE GRANITE SOUTH WADI SIWIQAT EL SODA, EASTERN DESERT, EGYPT

By

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The present work deals with the geology and chemistry of the granite mass south Wadi Siwiqat El Soda of the Eastern Desert of Egypt. Chemical study of the granite includes complete chemical analysis of four samples beside the quantitative determination of some trace elements like Rb, Sr, Y and Zr. The petrochemical characteristics of the granite are worked out.

The most important result of this work is a change in the identity of the investigated granite on geological maps. Thus instead of being regarded as old grey granite and granodiorite, this rock proved to be a member of the younger low-Ca pink granites. Evidences from the distribution of major elements, petrochemical characteristics and trace elements are discussed.

Introduction

The granite mass under study is located between latitudes 25°08′ and 25°10′ N and longitudes 33°40′ and 33°42′ E. It is found to the south of Wadi Siwiqat El Soda which is a branch of Wadi El Miyah lying in the Eastern Desert of Egypt (Fig. 1). This mass occupies an area about 56 km^2 and it is bordered from the north by slates and serpentine rocks while the western side of the mass is in contact with Nubian sandstone. To the south and east of the mass the area is covered by sandy plains.

The granite body south Wadi Siwiqat El Soda is subjected here to both geological and chemical study. The chemical work includes complete chemical analysis of four samples covering the main types of this granite body, besides certain trace elements (Rb, Sr, Y and Zr) are quantitatively analysed as well. This study helped to identify the character of the granite.

Sometimes Egyptian granites suffer from a serious problem of misidentification and many misidentified bodies are indicated on the geological maps. The present studied granite is an example of those misidentified bodies. Such bodies should be subjected from time to time to geological, petrographical and chemical studies in order to elucidate their exact nature.

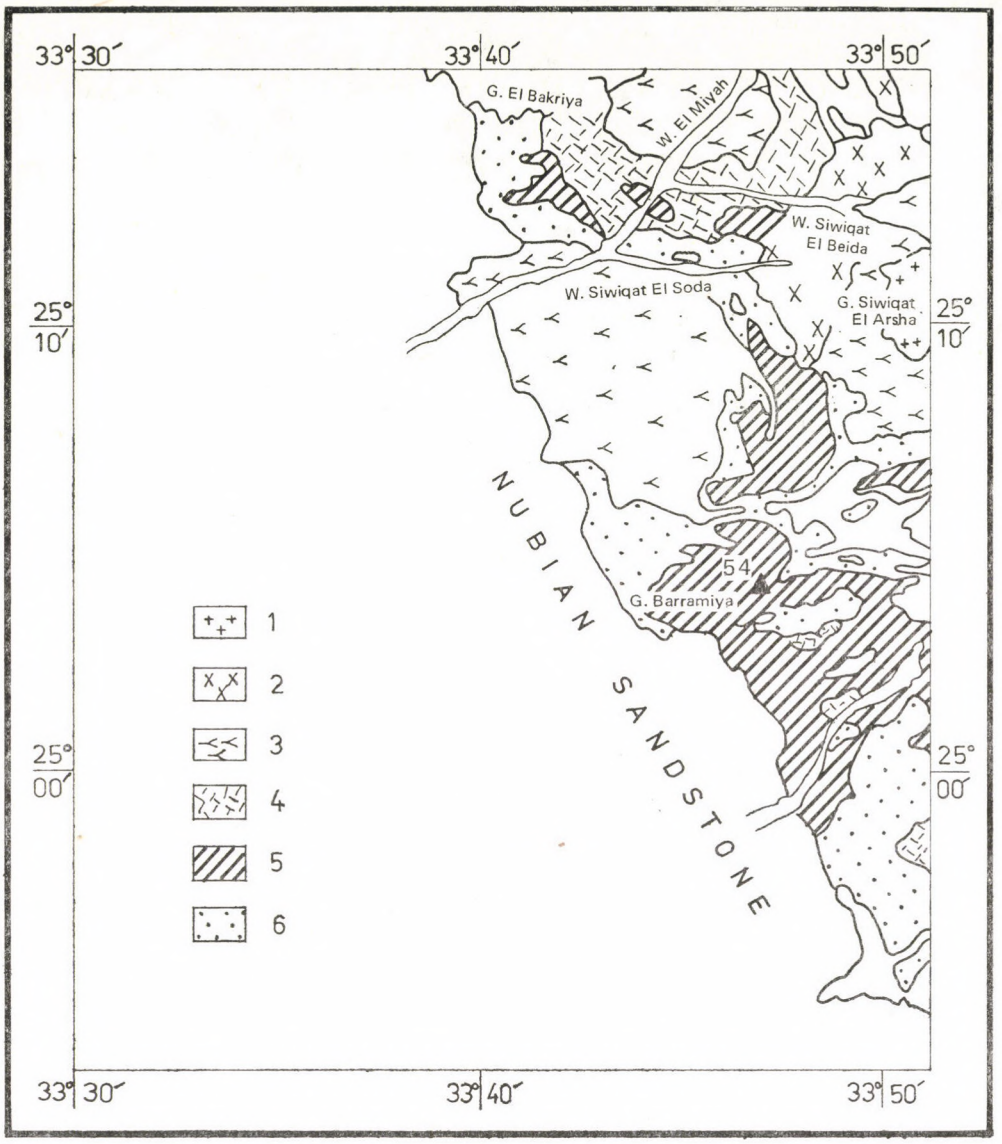

Fig. 1. Geological map of the granite South Wadi Siwiqat El Soda. 1. Muscovite granite; 2. pink granite; 3. grey granite and granodiorite; 4. epidiorites and diorites; 5. serpentines and related rocks; 6. schist-mudstone-greywacke series

Geology of the Granite

The northern part of the granite mass is found as isolated outcrops while the southern part is somewhat more massive and it is found as two nearly circular outcrops in contact with each other and occupying an area about 15 km^2

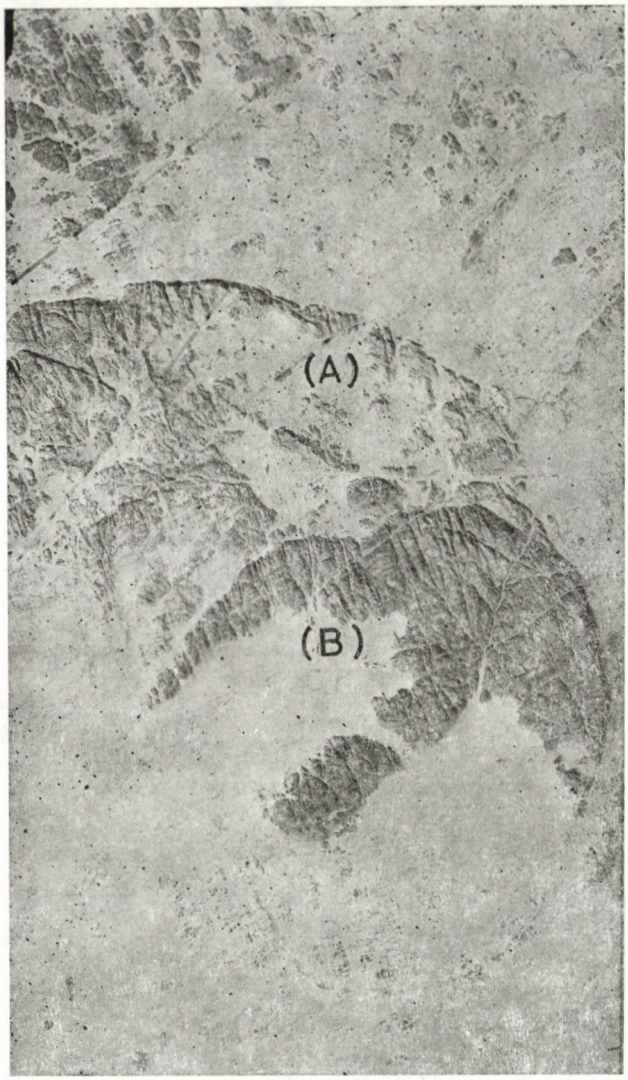

Fig. 2. The granite mass South Wadi Siwiqat El Soda

(outcrops A + B, Fig. 2). The two southern outcrops are found swamped in a sandy plain which surrounds them from all sides. Each outcrop has a more or less circular outline but the northern outcrop occupies a larger area. The hand specimens representing the two outcrops show marked difference. The hand specimen collected from the northern outcrop has light grey color while that representing the southern outcrop has reddish pink color.

The photogeologic study of the mass shows that the northern part appears on aerial photograph as dark grey isolated outcrops with low relief and fine

grained surface texture. The two outcrops to the south of the mass show marked differences between each other. These differences can be described in terms of surface texture and topography, drainage and photolineaments.

The two southern outcrops ($A + B$) have fine grained surface texture but the surface of the northern outcrop (A) has finer texture as compared with that of the southern outcrop (B). The topography of the southern outcrop (B) is higher than that of the northern outcrop (A).

The drainage in the two outcrops is mainly of the dendritic type but the density of drainage varies greatly in each outcrop. Measurement of the density of drainage per unit area which is 1 km^2 in each outcrop shows that the northern outcrop (A) has lower density amounting to 12 drainage lines per km^2 while the southern outcrop (B) has its density 26 drainage lines per km^2 (Fig. 3).

Photolineaments in the granitic mass represent mainly joints and probable faults. The main trends of these lineaments are N 15° W, N 25° W and N 65° W. In addition to these trends there are three less prominent trends which are N 55° E, N 55° W and N 65° E. These lineaments are represented mainly in the two outcrops ($A + B$) to the south of the granitic mass (Fig. 4). Most of these lineaments are found in outcrop (B) and constitute 62% of all lineaments present in the whole mass ($A + B$). The main trends in outcrop (A) are N 65° W, N 15° W and N 65° E, constituting 67.5% of all lineaments present in this outcrop. On the other hand, lineaments in outcrop (B) trend in the directions N 25° W, N 55° E and N 15° W. Probable faults are few in number but most of them are found in outcrop (A) with the trends N 40° E—N 60° E and N 60° W— N 70° W.

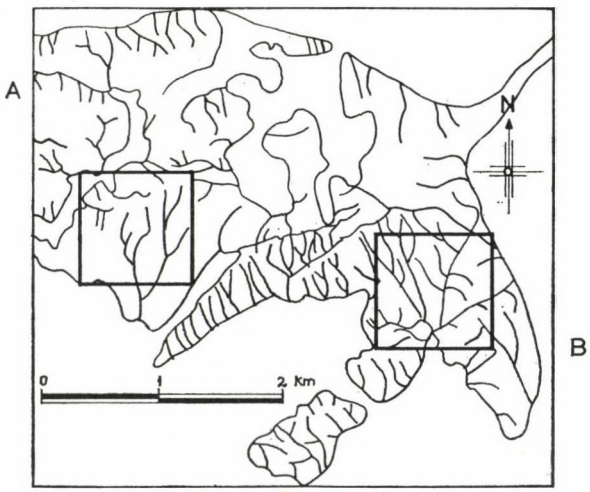

Fig. 3. **Drainage in granite South Wadi Siwiqat El Soda. Density of drainage is 12 drainage lines per 1 km^2 in mass (A) and 26 drainage lines per 1 km^2 in mass (B)**

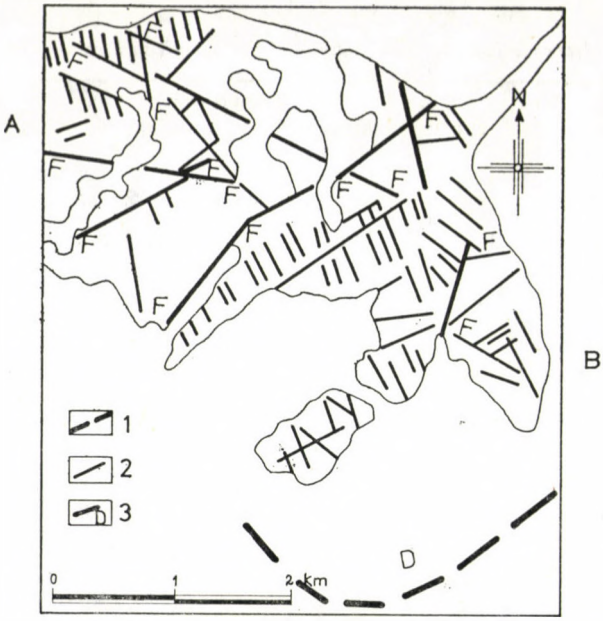

Fig. 4. Photo lineaments in granite South Wadi Siwiqat El Soda. 1. Inferred fault; 2. fracture line; 3. dykes

Four granitic samples are collected from the two southern outcrops ($A + B$). Two samples with greyish color are collected from outcrop (A), one with pink color from outcrop (B) and one sample only with pinkish color from the contact between the two outcrops. These samples covered the main varieties of granitic rocks present in the investigated area.

The two samples *A1* and *A2* look greyish in color, somewhat friable, with varying biotite content but this mineral is more enriched in sample *A2*. Sample *B* looks pink in color with the presence of normal amount of mica. Sample *C* which is taken from the contact zone between the two granite masses *A* and *B* looks pinkish in color, has a low mafic mineral content and is finer in grain size than the pink variety.

Chemistry of the Granite

Major Elements

The four samples collected from the granite south Wadi Siwiqat El Soda are subjected here to complete chemical analysis for major and minor elements. Analytical techniques are described elsewhere (El Sokkary, 1970). Results of these analyses expressed in weight per cent of the oxides are shown in Table 1

Table 1

Complete chemical analyses (wt.%) of the investigated granitic rocks

Oxide	Weight per cent						
	A1	*A2*	*B*	*C*	Low-Ca	Typical pink	Typical grey
SiO_2	73.32	73.53	74.02	73.12	74.29	74.08	67.84
Al_2O_3	13.01	13.26	13.01	14.03	13.61	13.14	14.34
Fe_2O_3	0.39	0.44	0.21	0.22	2.03*	0.30	—
FeO	0.63	1.02	0.39	0.31	—	1.13	3.53
MnO	0.01	0.04	0.04	0.04	0.05	0.02	0.05
MgO	0.41	0.30	0.41	0.36	0.27	0.29	2.32
CaO	1.82	0.95	1.63	1.63	0.71	0.82	1.95
Na_2O	4.30	3.85	4.03	4.00	3.48	4.24	3.65
K_2O	3.92	4.52	4.04	4.58	5.06	4.49	3.02
TiO_2	0.02	0.02	0.01	0.01	0.20	0.25	0.76
P_2O_5	0.04	0.01	0.01	0.01	0.14	0.05	0.18
H_2O^+	1.06	1.09	1.06	1.05	—	0.62	1.33
H_2O^-	1.02	0.68	0.82	0.50	—	0.10	0.06
Total	99.95	99.71	99.68	99.86	99.84	99.53	99.03

* Total iron is calculated as Fe_2O_3

Table 2

CIPW norms of the investigated granite rocks

Normative mineral	*A1*	*A2*	*B*	*C*
Or	23.75	27.45	24.50	27.60
Ab	39.50	35.45	37.10	36.55
An	4.68	4.75	5.58	6.88
C	—	0.37	—	—
Wo	1.72	—	1.04	0.50
Il	0.02	0.02	0.02	0.02
Mt	0.42	0.47	0.23	0.24
Hm	—	—	—	—
En	1.16	0.86	1.16	1.02
Fs	0.72	1.34	0.50	0.36
Ap	0.08	0.03	0.03	0.03
Q	27.95	29.25	29.83	26.82

which compares them with corresponding values of the low-Ca granite of TUREKIAN and WEDEPOHL (1961), typical pink granite and typical grey granite from Egypt (EL SOKKARY, 1970). Table 2 gives the CIPW norms of the investigated granitic rocks, while Table 3 shows their NIGGLI values.

Table 3

NIGGLI *values of the investigated granitic rocks*

NIGGLI value	*A1*	*A2*	*B*	*C*
al	43.22	45.63	44.99	46.39
fm	8.14	9.73	6.65	5.61
c	11.01	5.95	10.27	9.82
alk	37.63	38.68	38.09	38.16
mg	0.43	0.27	0.54	0.54
k	0.38	0.44	0.40	0.43
si	414.07	430.18	435.15	410.99
p	0.09	0.03	0.04	0.03
ti	0.07	0.08	0.05	0.04

Comparison of the analyses of the four investigated rocks with those of the reference low-Ca granite reveals the presence of good coincidence between the two sets of data particularly evident from the contents of: SiO_2, Al_2O_3, total Fe_2O_3, MnO, MgO, CaO, Na_2O and K_2O. However, total iron seems to show somewhat lesser values while CaO tends to develop somewhat higher values in the analysed rocks. Nevertheless the four analysed specimens from Siwiqat El Soda develop the typical character of low-Ca granites. Thus the granites from Siwiqat El Soda are low-Ca granites.

It should be noted that analyses of the four granite specimens are very close to each other which means that all of them represent one granite type, namely low-Ca granite. Color variation of the specimens from grey to pink shades is not significant and the granite mass, at least its southern parts, looks homogeneous.

On comparing the analyses of the four analysed granites from Siwiqat El Soda with those of typical pink granite from Egypt as given by EL SOKKARY (1970), there appears close similarity between the two sets of data. This is clear from the contents of: SiO_2, Al_2O_3, Fe_2O_3, FeO, MnO, MgO, CaO, Na_2O, K_2O and H_2O. The alkali elements Na_2O and K_2O have strikingly close values in the two sets of data. The ferrous content (FeO) tends to be lower while the CaO content tends to be higher in the analysed rocks. Despite of these minor variations, the present rocks represent members of typical pink granites occurring in various places in the Eastern Desert of Egypt.

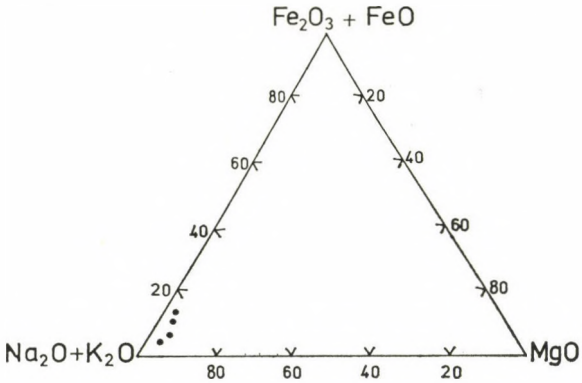

Fig. 5. The ternary relation ($Na_2O + K_2O$)—MgO—(Fe_2O_3 + FeO) in the studied granites

Figure 5 assures that the studied granites are impoverished in the ferromagnesian elements (Fe_2O_3, FeO, MgO) and are enriched in the alkali elements ($Na_2O + K_2O$), moreover they occupy the same position previously assigned for Egyptian pink granites (EL SOKKARY, 1970).

Fig. 6 shows the alkaline nature of the studied granites as points representing them lie almost mid-way between Na_2O and K_2O ends, moreover these granites tend to be relatively impoverished in lime (CaO).

How close are the analysed granites from Siwiqat El Soda to typical grey or high-Ca granites? As a matter of fact the analysed rocks can not represent typical Egyptian grey granites because their analyses are quite different. The analysed granites show decisively higher amounts of SiO_2 and K_2O with concomitant lowering trends in FeO, MgO and CaO with respect to typical grey granite. The reference high-Ca granite of TUREKIAN and WEDEPOHL

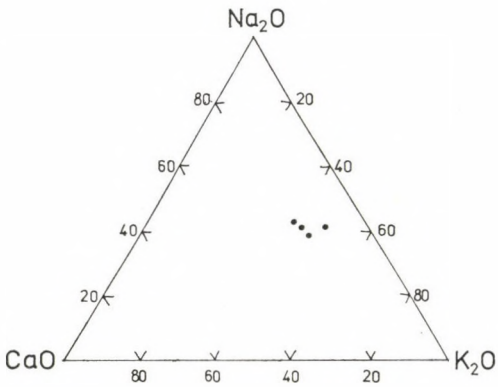

Fig. 6. The ternary relation CaO—K_2O—Na_2O in the studied granites

(1961) which is essentially a granodiorite has CaO content that equals 3.54% which is quite higher than the corresponding values of the analysed rocks.

In conclusion, the four specimens from Siwiqat El Soda are low-Ca granites being members of typical pink granite, they can not represent typical grey granites. With their analyses close to each other, they represent members of a more or less homogeneous granite mass. Referring to the geological map of Fig. 1, it is seen that the granite mass lying to the south of Wadi Siwiqat El Soda is marked as grey granite and granodiorite–a matter which is obviously unfounded and being against the results of the present investigation. The character of this granite mass should be corrected on the geological maps (e.g. EL RAMLY, 1972) as pink granite and not grey granite and granodiorite.

Petrochemical Characters

Table 2 shows that normative plagioclase somewhat exceeds normative orthoclase in the four analysed samples. On the other hand, the average An content of these normative plagioclases is found to be 13% An i.e. the plagioclase is either an oligoclase or may be an albite as the anorthite content is very close to the upper limit of that of albite (10% An). This plagioclase excess may be due to the perthitic nature of the present potash feldpars or may be due to the presence of much alkali feldspars i.e. both K and Na bearing feldspars. However, the granite of Siwiqat El Soda might be subjected to a process of albitization. Perthites of the vein type are reported in the pink granite mass of Wadi Siwiqat El soda (SABET and ZAATOUT, 1955). These might be formed through a process of albitization.

Normative quartz ranges between 26.82—29.83% in the four analysed samples with an average value 28.46%. The distribution of normative mafic minerals follows closely the content of biotite in the four samples. Thus sample *A2* with abundant biotite has 2.20% mafic minerals, sample *A1* with lesser biotite has 1.88% of these minerals, while the pink granite sample *B* has its mafic minerals 1.66%, sample *C* with scarce biotite has 1.38% normative mafic minerals.

The ternary relation between the normative feldspars Or—Ab—Anis superimposed on HIETANEN chemically classified diagram (1963), Fig. 7, from which it is concluded that the investigated granites from Siwiqat El Soda are classified as normal granites.

With respect to NIGGLI values, the four rocks from Siwiqat El Soda show similar values of si, k and alk. The fm and mg values of these rocks show certain fluctuations: the fm value ranges between 5.61—9.73 while the mg value ranges between 0.27—0.54 reflecting fluctuations in the content of the ferromagnesian minerals.

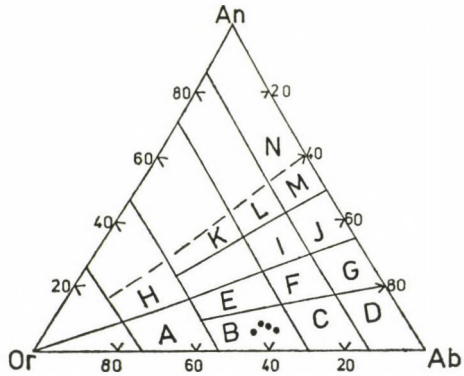

Fig. 7. The ternary relation between the normative minerals Or—Ab—An in the studied granites. A = kali granite; B = granite; C = granite-trondhjemite; D = trondhjemite; E = = quartz monzonite; F = monzonite; G = tonalite; H = calcic granite; I = granodiorite; J = quartz diorite; K = calcic monzonite; L = grano gabbro; M = gabbro; N = mafic gabbro

Trace Elements

Table 4 gives the contents in *ppm* of four trace elements, namely Rb, Sr, Y and Zr in the analysed four samples. These are compared with corresponding values of: (1) The reference low-Ca granite of TUREKIAN and WEDEPOHL (1961), and (2) Typical pink and grey granites as given by EL SOKKARY (1970).

Table 4

Trace elements (ppm) of the investigated granitic rocks

Element	*A1*	*A2*	*B*	*C*	Low-Ca	Typical pink	Typical grey
Rb	165	134	195	221	170	66	76
Sr	145	183	132	51	100	56	578
Y	14	19	17	21	40	22	6
Zr	113	113	97	69	175	137	234

The assigned trace elements in the four samples are comparable with those of the reference low-Ca granite especially Rb and Sr, being comparable as well with those of the typical pink granite particularly Y and Zr. These comparisons reflect that the granite from Siwiqat El Soda is low-Ca pink granite. With respect to typical grey granite from Egypt, the investigated four granites show higher Rb and Y with a simultaneous drop in Sr and Zr. This assures that the granite from Siwiqat El Soda cannot be typical grey granite.

Figure 8 shows the relation between Sr and Rb from which it is seen that some granites e. g. sample *C* (with scarce mafic minerals) are enriched in Rb and depleted in Sr which might suggest the presence of a weak differentiation trend within these granites.

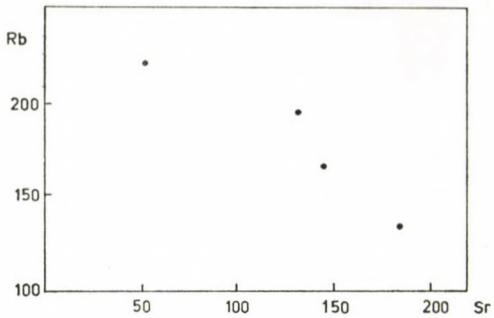

Fig. 8. Sr versus Rb in the studied granites

Discussion

The present study showed that the granites south Wadi Siwiqat El Soda, Eastern Desert, are found as two nearly circular outcrops in contact with each other. The northern outcrop has light grey color while the southern outcrop has reddish pink color. The topography of the southern outcrop is higher than that of the northern outcrop.

The distribution of major elements, the petrochemical characteristics and the trace element pattern of the investigated granites all point towards the fact that these rocks are low-Ca granites being members of typical pink granite, they cannot represent typical grey granite. The character of this granite mass should be corrected on the geological maps as being younger pink granite and not the old grey granite and granodiorite.

Since analyses of samples representing the two circular granite masses are close to each other and they essentially indicate one granite type, therefore differences alone in photogeological characters between these two granite masses are not conclusive in identifying each. The granite with grey color from the northern mass has relatively high percentage of biotite and its feldspars are lighter in color, a matter which may give differences in photographic tone between this granite and the pinkish southern variety with lower biotite content and with feldspars colored reddish-pink.

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ГЕОЛОГИЧЕСКОЕ СТРОЕНИЕ И ХИМИЧЕСКИЕ УСЛОВИЯ ГРАНИТОВ НА ЮГЕ ВАДИ-СИВИКАТ-ЕЛЬ-СОДА В ВОСТОЧНОЙ ПУСТЫНЕ ЭГИПТА

Е. М. ЭЛ-ШАЗЛИ—А. А. ЭЛ-СОККАРИ—М. М. АЛЫ—А. Ф. КАМЕЛ

Резюме

В настоящей работе рассматриваются геологическое строение и химические условия гранитного массива в южной части Вади-Сивикат-Ель-Сода в Восточной пустыне Эгипта. Химическое изучение включает в себя полный серийный анализ четырех образцов, помимо количественного определения некоторых рассеянных элементов, а именно — Rb, Sr, Y и Zr. Определены петрохимические характеристики гранитов.

Важнейшим результатом данной работы является изменение идентичности исследованных гранитов на геологических картах. Таким образом, в отличие от их отнесения раньше к древним серым гранитоидам и гранодиоритам, эти породы теперь оказались членом ряда более молодых, розовых гранитов с пониженным содержанием кальция. Приводится фактический материал относительно распространения главных элементов, петрографических характеристик и рассеянных элементов.

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ON THE OCCURRENCE OF SPHERULITIC STRUCTURE IN BASIC DYKE FROM DONI AREA, INDIA

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A basic dyke containing unoriented and unusually large sized xenoliths, traverses coarse grained amphibolites of Doni area. The study reveals that i) quartz dominates amongst the xenoliths, ii) felspar xenoliths alone have developed spherulitic structure, iii) the basic dyke is a basalt and iv) it acquired the xenoliths in it by forcibly intruding into a pre-existing vein of quartz.

Field Characters

The xenolith bearing basic dyke is situated at about 2 kms. south of Doni village (lat: 15°17′ N and long: 75°43′30″ E.) of Mundargi taluka, Dharwad district, Karnataka state. It is contained in survey of India Toposheet No. 48M/11. The dyke trends N5OE—S5OW, while the coarse grained amphibolitic country rocks strike N35E—S35W with a dip of 20° due S55E. The dyke is 150 mts. in length, 6—7 meters wide at its northeastern end, while it pinches to a width of half a meter or so at its southwestern end: it is thus wedge shaped in outline. The xenoliths are mainly of quartz and these vary in size from a few millimeters to as large as 30 centimeters in length; and from a few millimeters to 6 centimeters in width. Xenoliths are so numerous and of diverse shapes and sizes that at first sight the rock resembles a breccia (Plate 1, Figs 1 and 2). Xenoliths are present throughout the dyke and are unoriented (Plate 1, Figs 1 and 2).

Petrography

The dyke rock is dark gray in colour and megascopically xenoliths alone are identifiable. In thin sections the rock shows a fine grained texture (Plate 2, Fig. 2) and is seen to be made up dominantly of plagioclases and clino pyroxenes. Magnetite is present in small amount. Subhedral laths of plagioclases display typical multiple twinning. Central portion of the laths show alteration. The laths are unoriented. Clino pyroxenes are euhedral to subhedral in shape and are bigger in size than the laths of plagioclases. Clino pyroxenes show simple twinning and a pleochroism in shades of green and pale green colours. Magnetite grains are unhedral and are opaque.

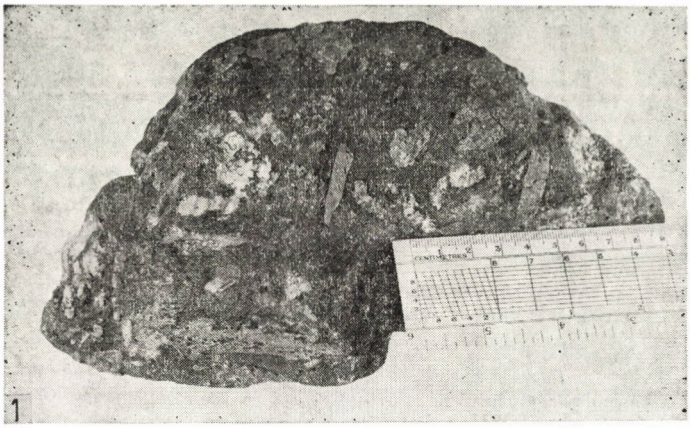
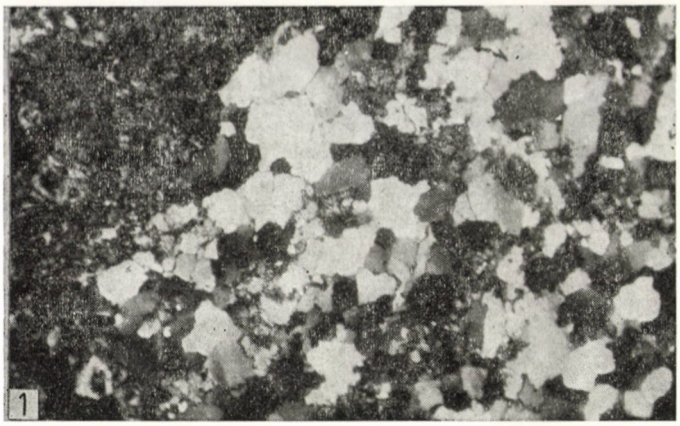
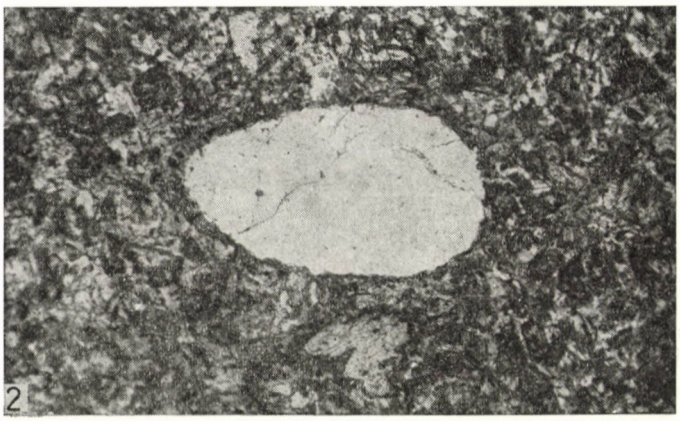


Plate I

Figs 1, 2. Hand specimens showing "breccia-like" distribution of xenoliths. Note unoriented nature of xenoliths ($^1/_3$ natural size)

Fig. 3. Microphotograph of xenoliths. Big white minerals at the base are feldspar, which show diffused contact with the matrix. Note granular, small pieces of quartz on the lower side of central, lenticular quartz xenolith. Note sharp contact of quartz with the matrix. (Crossed Nicols 3.5 ×)





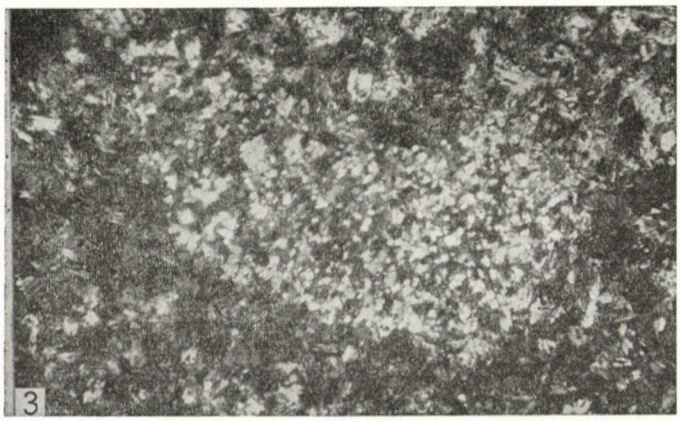
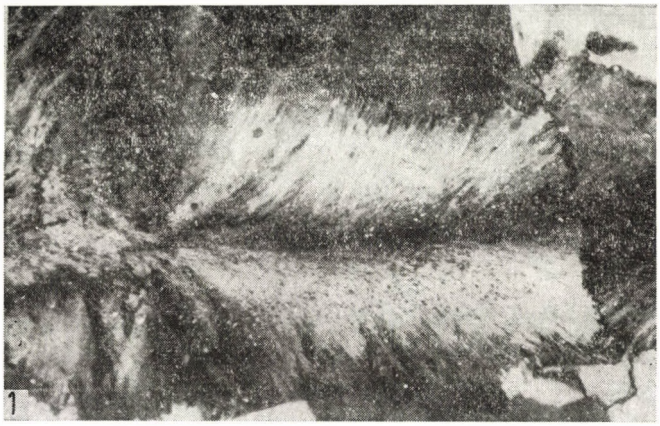


Plate II

Fig. 1. Microphotograph of a very big quartz xenolith. Quartz grains show mosaic texture and granulation. (Crossed Nicols 3.5×)

Fig. 2. Microphotograph of an oval shaped quartz xenolith. Note reaction rim around it. The matrix displays basaltic texture. (Crossed Nicols 10×)

Fig. 3. Microphotograph of a quartz xenolith. Note complete granulation of quartz. (Crossed Nicols 10×)



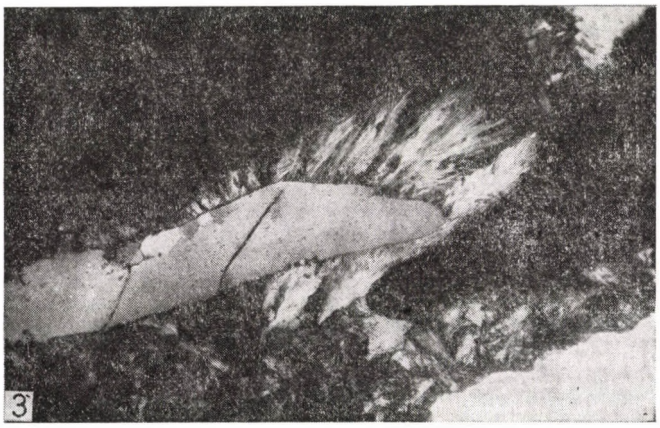


Plate III

Fig. 1. Microphotograph of spherulitic structure. Note singular black strip (extinction) within the mineral. (Crossed Nicols 45 ×)

Fig. 2. Microphotograph of spherulites. In some complete "cross" is seen, in others, it is half developed. Note restriction of spherulitic structure to felspar xenolith only. (Crossed Nicols 10 ×)

Fig. 3. Microphotograph showing laths of felspar radiating out from a stem of quartz. In the lower right hand side, spherulitic structure can be observed. (Crossed Nicols 10 ×)

Typical basaltic texture is not readily discernible. Laths of plagioclases stop against the plates of pyroxenes which indicates that the pyroxenes have crystallised earlier to plagioclases.

Xenolithic structure is noticed even in thin sections (Plate 1, Fig. 3 and Plate 2, Fig. 2). Xenoliths are numerous, generally lenticular in shape and occur as aggregates as well as individual grains. Quartz dominates amongst the xenoliths, those of felspars (Plate 1, Fig. 3) and sometimes of calcite are also seen. Quartz shows mosaic texture. The contact between the quartz xenoliths and the surrounding rock alone is sharp (Plate 1, Fig. 3). Tiny granules of pyroxenes are seen to be developed around xenoliths of quartz (Plate 2, Fig. 2). In some xenoliths, quartz is granular and is very fine grained (Plate 2, Fig. 3), in some others granulation is noticed only one side (Plate 1, Fig. 3). At places the quartz xenolith appears to be recrystallized (Plate 2, Fig. 1). Undulose extinction is displayed by quartz.

The most important feature is the development of spherulitic structure. This is restricted to the xenoliths felspar. In some cases, complete circle of spherulite is developed, in others a part of the circle is produced (Plate 3, Figs. 1 and 2). At many places laths of felspar are seen to radiate out from a grain of quartz (Plate 3, Fig. 3), and such a growth is seen to result is spherulitic structure.

Discussion

Mineralogical composition and textural features described above indicate that the basic dyke is a basalt. Xenoliths encountered in the dyke are of different dimensions, unoriented, and are not noticed outside the dyke. Majority of the xenoliths are of quartz and some are of felspars. These are therefore acidic in composition. These xenoliths are either "cognate" or are acquired by the basaltic magma from an extraneous source. It is quite clear that these xenoliths could not belong to the "cognate type", because these are acidic in composition, and such minerals are not produced as the first product of crystallization of the basaltic magma". Alternative source for these xenoliths is their derivation from an extraneous source. It is significant to note that veins of quartz are very frequently traversing rocks adjacent to the dyke. All the field and petrographical characters of the basaltic dyke suggest that it intruded into a pre-existing vein of quartz. Granular quartz, granulation observed on the periphery of quartz xenoliths, undulose extinction of quartz, diverse sizes and shapes of xenoliths, non oriented nature of the xenoliths, all these features together clearly point to the forceful injection of the dyke into the pre-existing vein of quartz. As a result of this, the quartz vein was shattered to small and big pieces which are now found as xenoliths within the basic dyke. This also satisfactorily explains the restriction of the xenoliths to the dyke rock alone.

The basaltic magma probably had no time or was not hot enough or both, to assimilate the xenoliths of quartz and of felspars to any extent. It, however, brought about recrystallisation of quartz and development of reaction rim around xenoliths of quartz. It also produced mosaic texture in the quartz xenoliths and spherulitic structure in those of felspars. GOKHALE and KOPPAD (1973) have described the occurrence of a similar dyke which contains xenoliths of quartz and felspars. They have pointed out that the spherulitic structure is produced only in the felspar xenoliths found in that dyke. The dyke under study also could be held responsible for the development of spherulitic structure in it. However, the present dyke differs from the one described by GOKHALE and KOPPAD (op. cit.) in containing considerably large sized xenoliths and these xenoliths being found throughout the dyke.

In conclusion it may be said that i) the dyke is unique in having rather large sized xenoliths of quartz, ii) the xenoliths being acquired by forcibly intruding into a pre-existing vein of quartz, iii) the crystallisation of the basaltic magma has produced the spherulitic structure in the felspar xenoliths alone.

Acknowledgements

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О ПРОЯВЛЕНИЯХ СФЕРОЛИТОВОЙ СТРУКТУРЫ В ДАЙКЕ ОСНОВ НОГО СОСТАВА В РАЙОНЕ ДОНИ В ИНДИИ

Н. В. ГОКХАЛЕ и С. К. ПУРАНИК

Резюме

Грубозернистые амфиболиты района Дони в Индии пересекаются дайкой основного состава, включающей в себя неориентированные ксенолиты необычайно больших размеров. В результате проведенных работы было установлено, что I) среди ксенолитов преобладает кварц; II) сферолитовая структура развита только лишь в полевошпатовых ксенолитах; III) дайка основного состава представлена базальтами и IV) она приобрела ксенолиты в ней путем форсированного внедрения в кварцевую жилу, которая там существовала раньше.

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A HYPOTHESIS ABOUT THE TERTIARY VOLCANIC ACTIVITIES OF THE MÁTRA MOUNTAINS, NE HUNGARY

By

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The authors have attempted to summarise the Tertiary evolution of the Mátra Mountains by means of re-evaluation of available field data adding their recent results of geological mapping, drilling and geophysics.

New data are given for the pre-Tertiary basement of the Mátra, followed by a short outline of the Upper Eocene volcanic evolution.

Neogene formations, especially the volcanics, are interpreted by their structural and stratigraphical setting. A reconstruction is given for the Western Mátra by adapting the previously published single volcano and caldera model of Szádeczky-Kardoss (1959). New aspects for further explorations are provided by the connection of hydrothermal activities and acid volcanism with a geophysically indicated acid intrusion. Similar, but more tentative reconstruction is provided for the Eastern Mátra.

1. Introduction

The Mátra Mountains is one of the largest unit of the Inner Carpathian volcanic arc as well as the largest Tertiary volcanic range of Hungary. Among many others, Mauritz (1909), Noszky (1927), Szádeczky-Kardoss (1959), Vidacs (1964), Kubovics (1970) were the most significant contributors to our geological knowledge about this area. In the last two decades detailed geological and geophysical surveys have been carried out, providing a great number of new data for both the surface and subsurface geology. Varga and others (1975) have given a comprehensive summary of these informations in their Mátra monography. More recently the Recsk ore explorations supplied numerous new drill-hole data for the NE-region of the Mátra.

The present study is attempting to reconstruct the volcanic evolutions in the different phases with the aid of re-evaluation and new arrangement of the existing data as well as the incorporation of more recent results.

2. The Basement

Up to the latest years only xenoliths have given scarce and ambiguous informations of the prevolcanic basement for the most part of the Mátra. These xenoliths indicated metamorphic basement in the Western Mátra, gran-

itoids in the Central parts. Mesozoic limestones — shales, cherts and spilites have been known in boreholes in the Eastern Mátra (VARGA, 1965).

Recently direct evidences were provided by several bore holes for the characteristics of the basement at Nagybátony (Nb-324), and in the Kékes—Disznókő area (R-92, R-103, R-109), showing Central-Mountains type Mesozoic sediments, limestones, shales, dolomites and siltstones. Granitoid rocks have also been found, but of much younger age, as later will be discussed.

These data indicate that in the known part of the Mátra two essentially different basement structures are in contact with each other: the folded Mesozoic of the Eastern Mátra and the faulted Mesozoic structural belt of Central-Mountain type in the Central- and Western Mátra.

The two basements are separated from each other by a regional-scale deformation-zone — known as Darnó-zone — and its continuation beneath the Tertiary of the Mátra, as revealed by geophysics (SZALAY, 1975).

This structural feature have been proved to be a controlling factor in each of the later events in the evolution history (ZELENKA, 1973).

3. Paleogene Volcanism and Sedimentation

The post-Triassic uplifting of the Mesozoic formations has been followed by a long period, during which the whole Mátra area was a part of the mainland, subjected to intense erosion and weathering, with no trace of marine sedimentation or volcanic processes in the Jurassic and Cretaceous.

The Laramian—Pyrenean phase has resulted in intense structural deformations, represented by strong rifting within the Darnó zone, uplifting and underthrusting along the basement boundary, with simultaneous subsidence and marine ingression, shoreline type sedimentation of Upper Eocene age west of the Darnó line. These movements have also initiated an intermediate volcanic activity in submarine environment, related closely to the Darnó zone, in the Recsk region.

The volcanic cycle itself comprise one semi-continuous eruptive period with four sub-stages. Its age is undoubtedly determined by the fossil-rich underlying, intermingling and overlying sediments, as in the Nummulites fabianii horizon of the Upper Eocene.

The first, initial substage is represented by entirely subaqueuos volcanics, like lava flows, lava agglomerates, tuffs, mixed volcanogenic breccias, peperites, witnessing frequent alternations of phreatic explosions and forceful effusions. The rocks are typically biotite-hornblende andesites, exhibiting strong influence of the carbonate rich basement in form of assimilation and contamination. The simultaneous structural process is a marked subsidence of both flanks of a central uplifted zone in which the centre of these eruptions could probably be

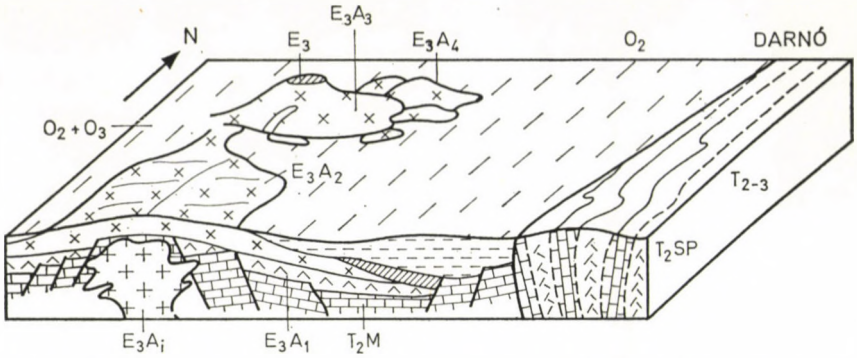

Fig. 1. **Block diagram of the Upper Eocene volcanic district of Recsk. T_{2-3}: Triassic folded limestone-shale series (Bükk-type). T_2SP: Spilites, diabases. T_2M: Triassic limestone, quartzite, shale. E_3: Upper Eocene limestones, marls, clays. E_3A_1: Upper Eocene lower biotite-hornblende-andesite series (1. substage). E_3A_2: quartz-biotite-hornblende andesites, dacites (2. substage). E_3A_3: biotite-hornblende andesites, (3. substage). E_3A_4: biotite-pyroxene-hornblende andesite dykes, extrusions, laccoliths (4. substage). O_2— O_3: Middle and Upper Oligocene siltstones, clays**

located. The peripherial character of these subsidences is evidenced by the laterally increasing amount of primary and reworked pyroclastic material.

The second substage has gradually developed by the step-by-step decrease of assimilation and contamination as well as the build up of a relatively typical stratovolcanic character. The course of structural evolution has not been changed though by the end of this substage the rate of accumulation exceeded over the magnitude of subsidence. This has resulted in a gradual shift from submarine to subaerial environment. Final eruptions in this period gave rise to the formation of ignimbrites, typical subaerial depositions.

The chemical character of these volcanics were shifted towards the more acidic range, giving place to dacites, andesito-dacites instead of the original andesites (Földessy, 1975). The eruptive centre of this phase falls north from that of the initial stage along the axis of the central horst. Our earlier studies have pointed to the genetic relationship between this volcanic substage and the emplacement of the shallow intrusion in Mesozoic country rocks in the depth (Baksa et al., 1977).

By further northward shifting of the eruptive centre, the third substage have also produced a stratovolcanic sequence of biotite — hornblende andesites and pyroclastics of much smaller regional extension than the previous substages, partly overlapping the earlier volcanic sequences. This substage has related to the most intense hydrothermal alterations in the whole series.

The fourth, final stage was in relation with the development of a central explosive caldera in the third stage strato-volcano, and resulted in the formation of radial and irregular dyke-pattern and small laccoliths within and around the caldera. Reef limestones and other lacustrine sediments mark the begin-

ning of a relatively quick subsidence of the volcanic area. This subsidence have reached its maximum by the middle Oligocene, when the largest part of the Eocene volcanics have been covered by marine sediments and only the central horst has escaped transgression.

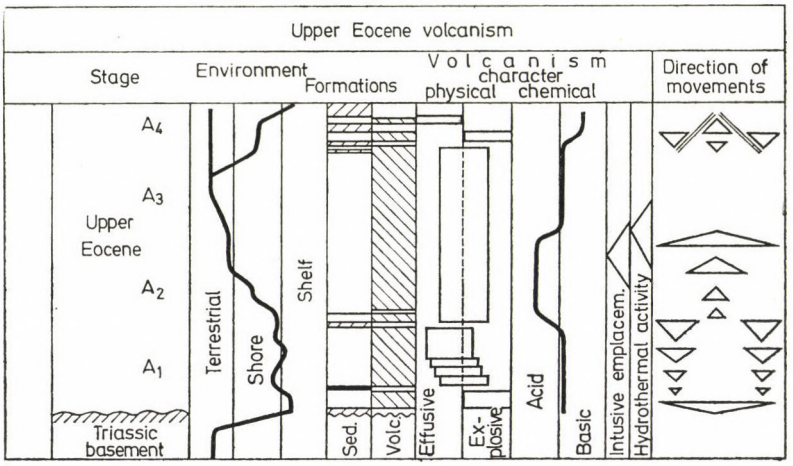

Fig. 2. **Relationships of the Upper Eocene geological events. At the direction of movement arrows represent uplifting and subsidence; double lines indicate development of fault systems**

The Upper Eocene volcanic activity has associated with very significant mineralizations in connection with the shallow intrusive porphyric body and its skarn environment, giving rise to the localisation of porphyric copper ores, skarnous copper ores, and metasomatic lead-zinc replacement ores in the intrusives and altered country rocks. The third substage of volcanism has produced intense hydrothermal alterations as well as formation of stockwork copper ores, while exhalative-sedimentary copper mineralization has developed in the caldera area.

4. Neogene Volcanism and Sedimentation

The Neogene volcanic area is isolated areally from the Paleogene with only slight overlap at the Disznókő—Kékes area. The pre-Neogene evolution of this area was ended by the gradual uplifting of the whole area after the shallow-water sedimentation in the Middle and Upper Oligocene, with clays, marls, sandstones and conglomerates. This sedimentation has been restricted to the Western and Central Mátra, while the eastern parts remained part of the mainland until the earliest Miocene.

In this recent model we separate the Western and Eastern Mátra by the southward continuation of the Darnó zone. This separation is supported by both structural differences and present morphology and drainage.

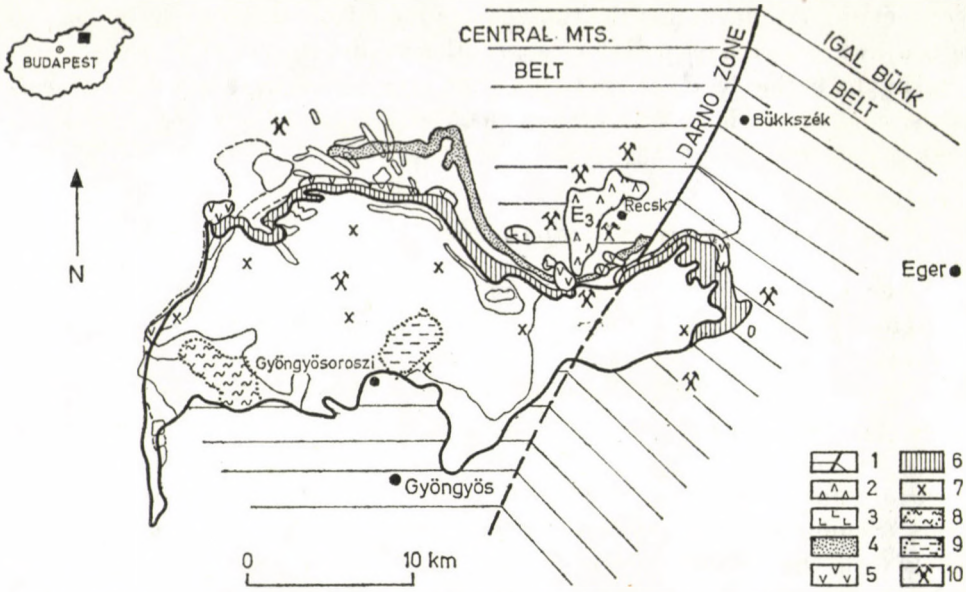

Fig. 3. Tertiary volcanics of the Mátra Mts. 1. Basement formations. 2. Upper Eocene biotite-hornblende andesites. 3. Eggenburgian andesites. 4. Lower Rhyolite Tuffs (Eggenburgian—Ottnangian). 5. Carpathian pyroxene andesites. 6. Middle Rhyolite Tuffs (Carpathian—Badenian). 7. Badenian pyroxene andesites rhyolites. 8. Diatomites (Caldera stage sediments). 9. Hydroquartzites (Caldera stage)

As a whole, the Western Mátra has preserved its original volcanic structure, so well established conclusions can be drawn from morphology, areal pattern of different volcanic formations, and the alteration pattern. The Eastern Mátra, however, has suffered stronger tectonic deformations in the post-volcanic times, and its primary morphological elements have been largely deteriorated. But, as a result of overall southward tilting, this area provides a great number of informations for the Early Neogene evolution in the northern parts, which otherwise could not be seen in the Western Mátra (Fig. 3).

4.1. *Eggenburgian—Ottnangian Stage*

The beginning of the Neogene is marked by the development of a relatively small scale volcanism. There are clear indications of isolated pyroxene andesite flows and related pyroclastics in the Eastern Mátra (Pálbikk, Rákhalom, etc.) which were overlain by the pumiceous rhyolite — dacite tuffs belonging into the Lower-Rhyolite-Tuff horizon (Földessy, 1973). Most authors agree in a trough-like explosive centre in the NE-foreground of the upper Neogene volcanics. Their opinion is based mostly on the size distribution of pumice fragments. Oil wildcat drillings in the SE-foreland of the Mátra

(Verpelét — 1) indicate that the thickness of these tuffs is significantly increasing southwards, perhaps pointing to a southern volcanic centre.

In the higher levels of the Ottnangian stage terrestrial and fluvial sediments, coal seams have deposited in the Western Mátra, while sedimentation gap has been recorded in the Eastern Mátra.

4.1. *Carpathian Volcanism and Sedimentation*

The Carpathian stage has brought an overall subsidence and marine transgression in the whole Mátra area, with the formation of siltstones, clays, sandstones of the Slir formation. In this marine environment extensive pyroxene andesite volcanism has taken place and produced a thick strato volcanic pile in the western parts, relatively thin lava flows, contemporaneous pyroclastics in the east. Its submarine character is evidenced by several features, like frequent glauconite in the groundmass (Óvár), connection with marine sediments (Somhegy), widespread argillitization (Rózsaszállás). The surface data, mainly from the western flank of the Western Mátra, and scarce drill hole data (Gyöngyösoroszi — 2) indicate, that the Carpathian andesites probably formed a central-type single large stratovolcano, with a centre near to Világoshegy, which has finished its activity by an explosive caldera stage, which have removed the most part of the volcanic superstructure, and deposited this material as fine grained andesite tuffs, intermingling with the Carpathian sediments. Fissure-type monogenic small volcanoes and laccoliths mark this period of activity in the Eastern Mátra (Somhegy, Csákánykő, Györke).

The andesites have been followed by a different type of volcanism, giving rise to the formation of rhyolite, rhyodacite tuffs, known as the Middle Rhyolite Tuff Horizon. Very little is known about the formation of these tuffs, which extend well over the area of the Mátra Mountains, to the northwestern side and southern foreland of the Bükk Mountains. Their uniform thickness and grain-size distribution throughout the Mátra indicates an external source. However, several rhyolite-dacite extrusions have been described in previous studies (Mezősi, 1950; Varga, 1975). Some of these have been reinvestigated, and found that instead of extrusive features, they exhibit typical ignimbrite character. With this in mind, we can conclude that this volcanic phase has taken place in subaerial conditions and its centre should not have fallen too far from the Eastern Mátra.

4.2. *Badenian Stage*

The present morphology and geology of the Mátra is largely dominated by the Badenian volcanic forms and structures. Three main phases of volcanism has been known in the Badenian stage, two andesitic and one rhyolitic ones.

Each of these incorporates several lava and pyroclastite varieties as well as compositional and textural types, reflecting different steps in the course of the volcanic evolution.

Basically two main groups of theories have been developed concerning the Badenian volcanism of the Mátra; SZÁDECZKY-KARDOSS and his colleagues have worked out a single stratovolcano model with a collapsed central caldera (SZÁDECZKY-KARDOSS, 1959). This model has been later modified by KUBOVICS (1971), with the assumption of a compaction caldera development. In the contrary, VARGA and others (1975) have described a volcanic range model, multiple eruptive centres along fissure systems. We, in our recent hypothesis, separate the eastern and western parts of the Mátra, mainly because of their structural differences, and attempt to give a morphological as well as structural approach to their genetic interpretation.

4.21. *Western Mátra*

In accordance with the earlier assumptions of SZÁDECZKY-KARDOSS (1959), the Western Mátra exhibits striking features of a large stratovolcano, with well developed caldera structure in its central parts. Several field evidences support this theory, which otherwise could not be fit into a volcanic range model.

The first of these is provided by the uniformity of the topographic position, thicknesses and petrology of the immediate basement for the Badenian volcanics, i.e. the uniform character and topography of the Middle Rhyolite Tuff Horizon. This should be hardly possible in the case of structurally controlled multiple-centre volcanism, where considerably larger horizontal and vertical displacements should have been expected.

The Badenian volcanic series is subdivided into a lower pyroxene andesite series (Ma_2) and an upper more basic pyroxene andesite series (Ma_3), which have been separated by the rhyolitic volcanism. The second evidence for the single stratovolcano model is given by the thickness distribution of the Ma_2 volcanics. This pattern strongly resembles to the double Somma-ring as described by RITTMANN (1962). In the outer ring, i.e. in the assumed basement of the stratovolcano, 200—400 m thicknesses have been recorded. In the inner ring drill hole intersections and seismic data revealed 900—1300 thicknesses. This arrangement is concentric around the central caldera (Fig. 4). Finally, within the caldera the andesites exhibit approximately 400 m average thickness, reflecting the uplifted nature of this central zone.

If the areal distribution of scoriaceous pyroclastics is considered, a further evidence can be obtained for the single volcano model. These pyroclastics are limited to the outer ring area — i.e. the outer flanks of the volcanic cone. The proportion of the pyroclastics compared to the lava flows is ranging

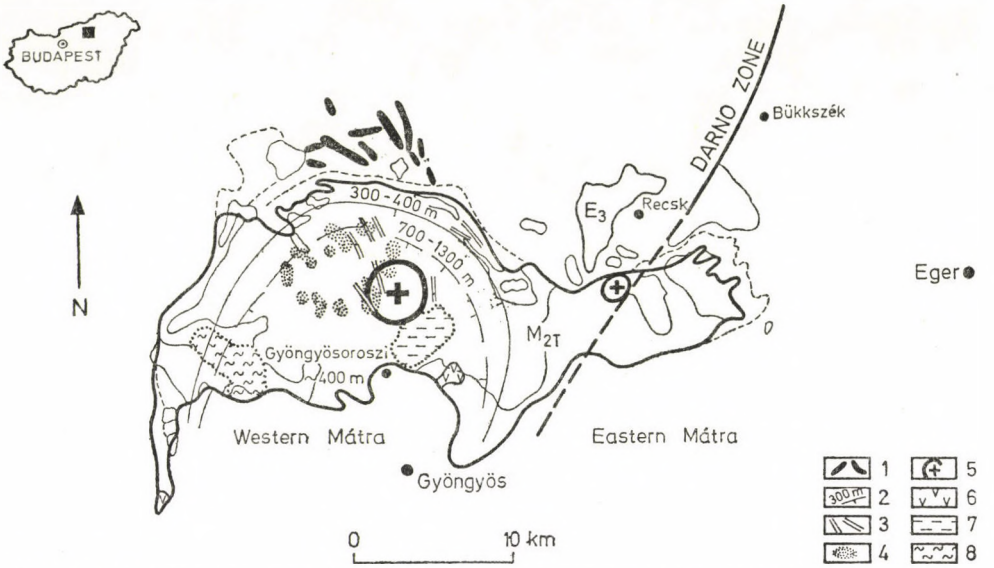

Fig. 4. Badenian volcanism of Mátra. 1. Radial and concentric dyke systems. 2. Approximate isopach controls of volcanics. 3. Hydrothermal vein systems. 4. Areas of intense potash-metasomatism. 5. Surface projection of geophysically indicated intrusives. 6. Rhyolitic extrusives; 7. Area of intense solfatara activity. 8. Diatomites

between 60—80 per cent in the outer ring, and remains below 40 per cent in the caldera area.

The recognized dyke systems can also be fit into this pattern, with remnants of recognizable concentric, radial and conic arrangements, in the northwestern side of the area.

The great variability of petrological and textural features apparently contradict to this uniformity of volcanic structure. KUBOVICS (1971) have encountered 14 different pyroxene andesite varieties, but has not offered reliable genetic explanation for their relationship. These varieties include mainly textural, grainsize differences, and different rates of oxidation, though their basic petrological similarities, like mineralogical and chemical compositions, by no means exist. This problem needs further detailed investigations, in order to fit the petrological pattern into the structural model. However, we prefer regional scale relationships to relatively minor and local differences, and consider this petrological variability of secondary importance compared to their structural connections.

The Lower Badenian volcanism has been followed by a series of structural and volcanic processes which finally have led to the development of the central caldera. The existence of this caldera is primarily evidenced by its excellently preserved morphology, but also by the above listed features, outlining a Somma-type structure.

The age of the caldera is reliably determined by the spacely connected hydrothermal activity, which has already not affected the younger andesite series (Ma_3). This places the development of the caldera stage to the interval between the formation of the two andesite series.

For the formation of this caldera explosion or erosional processes can be suggested. Collapsion or compaction is unprobable since the pre-volcanic basement is considerably uplifted within the caldera, and this is coupled with significant loss of material from this area. Extensive K anomalies in the southern foreland of the Western Mátra have probably associated with the erosion and transport of the hydrothermally altered volcanics from the caldera area to the peripheries.

In the caldera stage several processes have taken place which permit the time-and-space reconstruction of its development.

Rhyolite domes, extrusions and associated pyroclastics have long been recognised in the caldera area, but these were considered as the youngest of the volcanic phases. In Fig. 4 is well seen that their distribution is limited within the caldera contours. On the other hand, these rhyolites, unlike the final andesites, have been regionally affected by hydrothermal alterations, consequently these are probably older than the fresh, unaltered andesites.

The regional hydrothermal activity, another characteristic feature of the caldera area, has resulted in intense rock alterations as well as several small Pb—Zn vein-type mineralizations. The most significant of these alterations, in close association with the ore-mineralization, is the potassium metasomatism. Airborne radiometric surveys have revealed several K-enriched areas, showing the regional extent of this alteration process (GÉRESI and WÉBER, 1970). The previous studies of these K-rich varieties pretended either orthomagmatic source (SZÉKY-FUX and HERMANN, 1951) or hydrothermal leaching (KUBOVICS, 1971). The 10—13 per cent K content, however, can not be attributed solely to porphyric feldspars or relative enrichment from a standard andesite composition, but to a very intense metasomatism of both felsic and mafic components, producing 60—80 per cent K feldspar in the whole rock mass.

The recent seismic measurements might provide a reliable key for the genetic explanation of these coincidences. In the northern side of the inner zone of the caldera seismic refraction profiles have indicated an intrusive body surrounded by Neogene sedimentary country rocks at 1000—1500 m depth, with average velocities of 5500 m/sec, characteristic to acid plutonic rocks. If we assume the existence of such an acid intrusive body, which has been emplaced simultaneously with the caldera development, all of the recognised surfacial phenomena can be linked together as remote effects of this emplacement.

The final stage of the caldera development is marked by extensive solfatara activity and contemporaneous marine ingression into the caldera area,

during which lacustrine and fresh-water diatomites have deposited at Gyöngyöspata and Szurdokpüspöki, geysirites at several places, like Asztagkő or Gyöngyössolymos. The interbedded rhyolite tuffs in these rocks also support the contemporaneity of caldera formation and the rhyolitic volcanism.

The final andesite volcanism produced more basic pyroxene-olivine andesites. The volcanism is confined to the rims of the caldera structure, and its mode was probably strongly effusive, resulting in the formation of relatively uniform sheets of basaltic andesite lavas with subordinate pyroclastics. These sheets have later separated into parts by erosional effects.

By the end of the (Tortonian) Badenian the present volcanic structure and morphology has been almost completely developed. Post-volcanic movements have strongly affected the southern forelands, where significant subsidences have taken place, indicated by the 600—700 m throwdown of the Badenian andesites, now overlain by Pannonian sediments.

4.2.2. *Eastern Mátra*

This area exhibit more complex features, consequently more tentative approach to its genesis is only possible. Field data clearly outline its integrity as one single volcanic structure, but from this point two interpretations are equally possible:

(1) Present surface data delineate an elongated, WSW-ENE trending fissure controlled volcanism.

(2) Drill holes in the southern, south-western periphery recorded significant Badenian andesite thicknesses under younger Neogene sedimentary cover. This fact may lead to the conclusion, that the present surfacial extension of the Eastern Mátra Badenian volcanics represent only the northern, northwestern part of a considerably larger, probably central type stratovolcano.

Like the western volcanic structure, this one has also been affected by late Neogene faulting, which has also shaded its original superstructure.

The lower Badenian pyroxene andesites are much alike as those in the Western Mátra. The pyroclastic layers can well be traced thoughout the southern surface outcrops, thus further supporting the single eruptive centre model. The mud-flows and lahars near Markaz give the western slopes of this volcanic structure.

While the existence of a caldera structure has not been clearly established, drilling evidences are available for the existence and approximate position of a Neogene intrusive facies. This has been provided by R92, an exploratory borehole which had been set to a gravity and geomagnetic anomaly, which coincided with a seismic uplift. The hole has intersected the two Badenian andesite series and found a contact zone of andalusite-diopside hornfelses from Neogene and Oligocene sediments. These have been intruded by dykes of

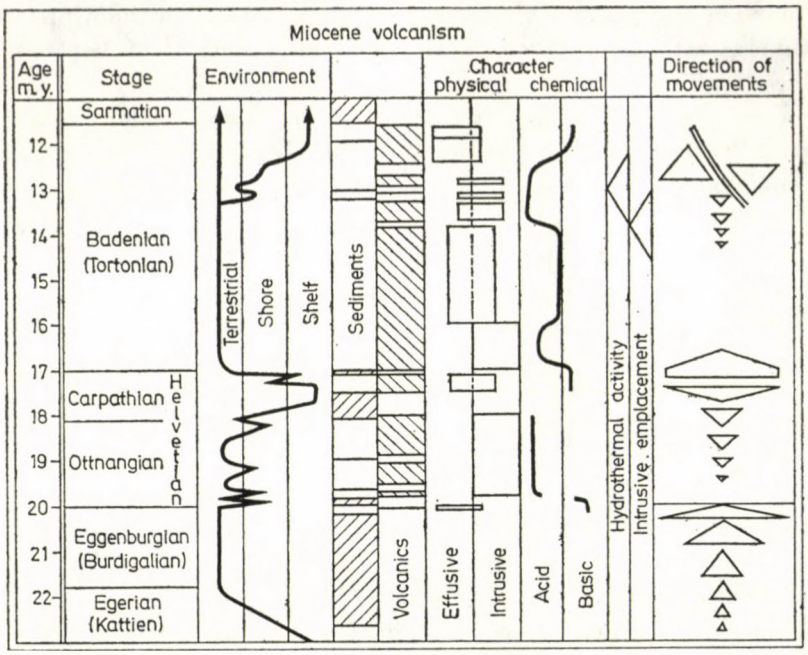

Fig. 5. **Relationships of the Miocene geological events**

andesite and holocrystalline quartz diorites. Seismic profiles have shown a potential intrusive emplacement west from this bore-hole. The only K-anomaly in Eocene volcanics (at Ilona-völgy) is related to a small dyke of granitoid origin — a possible remote effect of this intrusive mass.

Absolute age data are available for the intrusive rocks, giving an age of 13.8 ± 2.0 m.y. which is younger than the Lower Badenian volcanics (15.7 ± 2.7 m.y.) but older than the upper basaltic andesites (11.8 ± 0.8 m.y.) (BALOGH, 1975).

The final basic andesites show similar characteristics to those in the Western Mátra, with relatively thin lava flows and pyroclastics (Fig. 5).

5. Conclusions

The pre-Tertiary basement of the Mátra indicates the existence of a major deformational zone across the Mátra, in the continuation of the Darnó zone. The structural differences between the two units separated by this zone have been maintained throughout the Tertiary period.

The first Tertiary volcanic activity is clearly connected with the movements along the Darnó zone, and resulted in a four stage volcanism, strating

in submarine, and vanishing in terrestrial conditions in the Upper Eocene. The volcanism has coupled with the emplacement of andesite–diorite intrusives in Triassic sedimentary rocks. The significant mineralization at Recsk were connected to the localisation of the intrusive mass and the subsequent alteration processes. On the surface these processes have been associated by the formation of an explosive caldera, and volcanogenic-copper-pyrite mineralization.

The Neogene volcanism is not controlled by the Darnó zone itself, though the influence of this structural zone is relevant in the Neogene volcanic stages too. Three main periods of volcanism has been recognised in the Neogene, in the Eggenburgian–Ottnangian, in the Carpathian and in the Badenian stages. Each of these includes an andesitic and a rhyolitic phase. Through these three phases, the initial rhyolite predominance has been changed towards the andesite side. The first two stages reflect alternating marine–terrestrial environments, while the third one has taken place in terrestrial environment.

Generally speaking, the Badenian structure of the Mátra Mountains includes two volcanic structures, the Western Mátra and the Eastern Mátra, each of them exhibiting a marked stratovolcanic character.

The most significant events in the Western Mátra have been associated by the development of a central caldera subsequent to the lower Badenian andesite volcanism. This caldera stage is connected with acid volcanism, intense hydrothermal alterations, K-metasomatism and potentially with the emplacement of a shallow intrusive.

The scarcity of data does not allow similar reconstruction for the Eastern Mátra. Drill hole data indicate, that much of its Badenian volcanic structure is now covered by younger sediments, and its volcanic character can only be interpreted together with the covered subsided parts. On its northern side drill-hole evidences became available concerning the existence of intrusive formations, which related to the Lower Badenian volcanics.

Based on the above mentioned hypothesis for the caldera development, with the connection of the vein type Pb—Zn mineralization and K-metasomatism to intrusive activity, these considerations provide new dimensions for the exploration potential of the Mátra Mountains.

Acknowledgements

The authors wish to express their thanks to Szalay István (ELGI) for his valuable discussions on geophysical results, and Fábián Béla (OKGT) for providing detailed informations on their drill-hole logs in the studied area.

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ПОПЫТКА РЕКОНСТРУКЦИИ ТРЕТИЧНОГО ВУЛКАНИЗМА ГОР МАТРА (СВ ВЕНГРИЯ)

Ч. БАКША—Й. ЧИЛЛАГ—Й. ФЕЛДЕШШИ и Т. ЗЕЛЕНКА

Резюме

Авторы настоящей работы попытались восстановить ход третичного геоисторического развития гор Матра на основании переоценки имеющихся полевых наблюдений, геологосъемочных материалов, буровых данных и геофизических материалов.

Они сообщают новые данные о древнем фундаменте гор Матра. Затем они дают краткое изложение вулканической деятельности, имевшей место в позднеэоценовое время.

Оценка неогеновых образований — преимущественно вулканитов — была проведена на основании учета структурных и стратиграфических особенностей района. Модифицировав модели, установленной Э. Садецки-Кардошшем (1959), в основе которой легла концепция вулкана и кальдеры с единственным центром извержения, авторы статьи дают новую реконструкцию истории Западной Матры.

Предполагая существование связи между гидротермальной деятельностью и вулканизмом кислого состава, с одной стороны, и образованием кальдеры и присутствием геофизически выявленного кислого интрузивного тела, с другой, авторы открывают новые перспективы для дальнейших исследований.

Аналогичные реконструкции даются и для Восточной Матры, но они основываются на ряде гипотез.

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STATISTICAL ANALYSIS OF MAJOR ELEMENT PATTERNS IN BASALTIC ROCKS OF HUNGARY

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Discriminant diagrams (PEARCE, 1976) have been employed to establish the most likely tectonic setting of four different basaltic rock-associations in Hungary. The young Upper Pliocene-Pleistocene basalts (Balatonfelvidék, Cserhát) have marked within plate basalt affinities the eruptions of which are connected with recent mantle-diapirism and continental rifting. The Lower Cretaceous basic rocks (Mecsek Mts.) plot also as within-plate basalts and therefore they are interpreted as witnesses of a Lower Cretaceous rifting process. The Mesozoic diabases of the Bükk Mts. show ocean-floor basalt affinities and it has been argued that they were emplaced tectonically along the most important tectonic line of the Pannonian Basin. Finally, the basaltic andesites of the Carpathians show shoshonitic affinities and they could have been generated by a late Oligocene early Miocene subduction process occurred in the Carpathian region.

Introduction

Ever since the plate tectonic model of tectogenesis has gained a general acceptation among earth scientists, and our understanding as to the magma genesis has become as deep as never before, petrologists have been seeking for an eventual correlation between the element patterns of lavas and the tectonic environment in which they were erupted. In this respect several successful methods, especially in the case of basic volcanic rocks have been proposed, e.g. that of PEARCE—CANN (1973) which uses minor element patterns, as well as PEARCE's (1976) method which on the contrary makes use of major element patterns to establish the most likely tectonic settings of basic volcanic rocks. A similar approach to the second one has been proposed by NISBET—PEARCE (1977), the greatest difference being that they used major element patterns of clinopyroxene instead of whole rocks.

In the present paper I employed Pearce's method, that is I calculated discriminant functions on the basis of major element data to represent visually the separations of four well-defined basaltic rock associations of Hungary. Details and theoretical background of the calculations with a complete list of references can be found in Pearce's work. In the following, geology and petrology of the rock associations will be shortly discussed.

Geology and Petrology of the Basic Rock Associations

A. *Young Alkali Basalts*

The first basaltic formation to be considered is that of the young basalts erupted partly north of the Lake Balaton (Balatonfelvidék) and partly in the Cserhát Hills (Fig. 1). However, basaltic eruptions of the same volcanic activity can be identified well beyond the frontiers of the country, so the Cserhát basalt area is continuing to the North in Slovakia and there is another basalt area near Graz in Austria and several others in Transylvania. The volcanic areas are mainly composed of lava flows, scoriacious lavas and tuffs. At several localities they contain nodules of peridotite, megacrysts, and xenoliths of cumulate origin. Lavas show typical alkaline affinities, they are fresh, although at some localities zeolitization is widespread. Differentiation processes could not have operated to any large extent, since the chemistry of individual lava flows is rather uniform (MAURITZ et al., 1948). The age of the eruptions is Upper Pliocene-Pleistocene, and it seems that the Cserhát basalts are younger (~2 m.y.) than those of the Balaton area (~3—5 m.y., BALOGH KADOSA's personal communication). Seeing that since Upper Pliocene time only a few million years have passed we can expect that tectonic conditions at that time could not have been very different from that prevailing at present. The essential

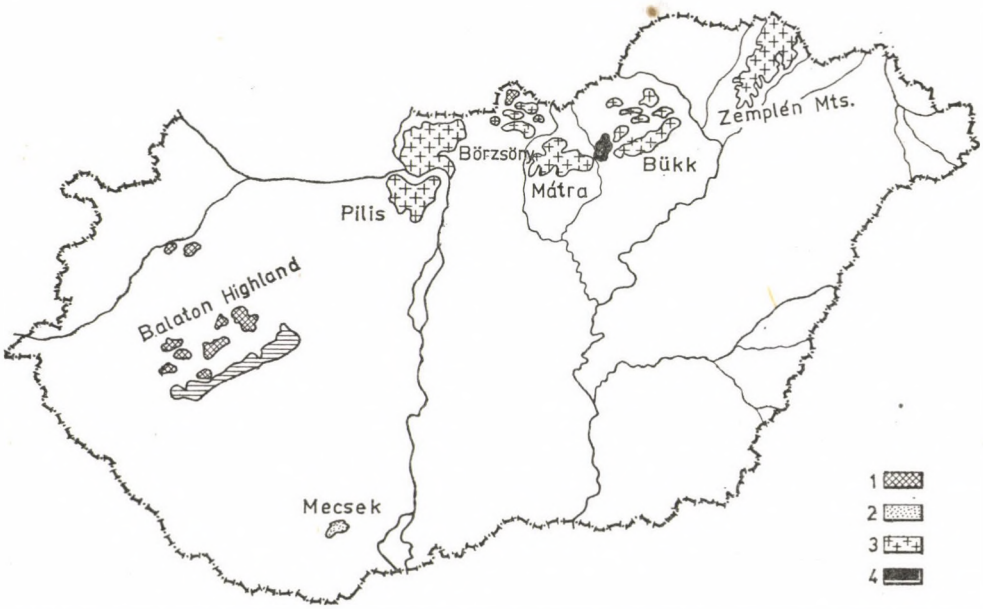

Fig. 1. Volcanic regions of Hungary. 1. Young basalts; 2. Lower Cretaceous basalts; 3. Miocene andesites; 4. Mesozoic diabases

characteristics in the Pannonian Basin are as follows: the presence of a continental-type crust, though thinner than in a normal case, block tectonism that results in a highly variable degree of sinking in different parts of the basin.

B. *Alkali Basalt-trachybasalt-trachyandesite-phonolite Series*

Members of this rock series were described in details by MAURITZ (1913) as a "trachydolerite-phonolite" association. They were erupted in the Mecsek Mountains (Fig. 1) where they form numerous dykes, stocks, sills and probably laccoliths. The age of the beginning of eruptions can be defined unambiguously, as the igneous bodies cut the Upper Jurassic sediments and synchronous volcanic activity can be deciphered in the Berriasian ($\sim$ 136 m.y.) at the base of the Cretaceous (PANTÓ, 1961).

Apparently differentiation processes must have been operating on a large scale, since limburgite-like cumulative rocks and complementary trachybasalts, trachyandesites and phonolites are well represented in this area. It is obvious that the parent magma must have had a marked alkalic affinity to be able to yield such highly undersaturated differentiated products. Rocks of this igneous series are by no means so fresh than those of the preceding section: zeolitization is very intense, especially in the case of the more acid members, other alteration products such as calcite, sericite and chlorite are not rare.

C. *Diabase-gabbro-oreperidotite Association*

This rock association is situated in the western part of the Bükk Mountains (Fig. 1) and it is composed mainly of a NE-SW striking diabase body about 10 km in length and 2—4 km in width. Smaller bodies of gabbro and peridotite with an unusually high content of magnetite and ilmenite were formed at lower levels, very probably due to gravitative differentiation. Very subordinate acid differentiation products, nests of gabbro pegmatite, nests and veins of albite granite occur in the gabbro itself. The whole series was intruded in a slightly metamorphosed shale of Triassic age. The age of the intrusion itself could not be determined exactly, it may be Triassic or Cretaceous as well.

The petrography of the rock association was described at a considerable length by SZENTPÉTERY (1953). There are some localities where the rocks are quite fresh, the mineralogy and chemistry of which enables us to conclude that the parent magma must have been tholeiitic in character.

D. *Basaltic Andesites*

As it is indicated by the name, these rocks are by no means true basalts though they approach the basaltic composition. They represent the most basic members of the andesite-dacite-rhyolite association of the Intracarpathian Volcanic Chain. In Hungary the Pilis, Börzsöny, Cserhát, Mátra and the Zemplén Mountains (Fig. 1) are built of this rock association. At the western part of the chain rocks are about 15 m.y. old, to the East, in the Zemplén Mts. they are only 12 m.y. old and in Transylvania they are even younger. There are many records of basalts accompanying andesites in the case of island arc-type volcanism, and this kind of basalt can be distinguished chemically from basalts of other tectonic settings. In the following, using the discriminant diagrams of Pearce, I intend to test whether or not the basaltic andesites in question fall in one of the fields of volcanic arc basalts.

Choice of Analyses

PEARCE (1976) has shown that a small number of analyses may be sufficient if we need to classify lavas of unknown origin by the aid of his discriminant function diagrams. Where it was possible I strictly respected his recommendations: analyses of samples described as altered or with $FeO/Fe_2O_3 < 0.5$ were rejected, the total of all major element oxides including H_2O was between 99 and 101 per cent everywhere, and the range of $CaO + MgO$ was between 12 and 20 per cent. From the young alkali basalts described under section A, as well as from the diabases (Section C) a higher number of analyses was published from which 30 and 22 have been chosen. Analyses of the young alkali basalts were chosen so as to represent all the basalt eruptions of the country. Though it is obvious that the discriminant diagrams can be used exclusively for lava rocks, I plotted there discriminant functions of some gabbros associated with diabases (Section C) too. Since there are transitional rocks between diabases and gabbros of the Bükk Mountains, the eventual compositional shifting caused by increasing gabbroic character may thus be deciphered.

Unfortunately, there is only a smaller number of analyses published from the Lower Cretaceous alkali basalts of the Mecsek Mountains. Thus, I had to plot rocks with greater and lesser amount of $MgO + CaO$ than the normal basaltic range, too. In this way, compositional shifting caused by the differentiation could be evaluated. As these rocks are partly zeolitized and chloritized, oxidation rate and H_2O is higher than in fresh basalts.

Finally the basaltic andesites are much lower in MgO than are average basalts and the sum of $MgO + CaO$ is around the lower limit (12 per cent), in some cases only approaching this value.

The list of analyses and the sources of data are summarized in Table 1.

Table 1

Sources of major element analyses used in the discriminant diagrams

Location	No. of Analyses	References
Mecsek Mts.	8	MAURITZ (1913)
	2	TAKÁCS (1933)
	2	SZÉKY-FUX (1952)
Balatonfelvidék and Cserhát Mts.	2	POJJÁK (1944)
	4	REICHERT (1925)
	2	EMBEY-ISZTIN (1977)
	2	VITÁLIS (1911)
	12	MAURITZ—HARWOOD (1936, 1937, 1938)
	8	MAURITZ et al. (1948)
Bükk Mts.	25	SZENTPÉTERY (1953)
	7	EMBEY-ISZTIN (unpublished)
Valley of Bódva	4	PANTÓ—FÖLDVÁRI-VOGL (1950)
Mátra Mts.	4	KUBOVICS (1970)
Zemplén Mts.	2	SZÉKY-FUX (1970)
Börzsöny Mts.	4	PANTÓ (1970)
Pilis Mts.	2	KORPÁS et al. (1967)

Basalts of Different Tectonic Settings

In PEARCE's diagrams there are six magma types that can be defined on the basis of modern plate tectonic processes according to PEARCE—CANN (1973):

1. Ocean-floor basalts (OFB) are erupted at diverging plate margins, either within large oceans (mid-oceanic ridges) or within the small ocean basins behind island arcs.

2. Island arc tholeiites (LKT) are erupted on oceanic crust at converging plate margins. They are typically erupted close to the deep-ocean trenches.

3. Calk-alkali basalts (CAB) are erupted on continental crust at converging plate margins, and on oceanic crust well behind the deep-ocean trenches.

4. Shoshonites (SHO) are erupted at converging plate margins, either a long way behind the deep-ocean trenches in mature arcs, or in post-orogenic situations after subduction has ceased.

5. Ocean-island basalts (OIB) are erupted as ocean islands within ocean basins, largely within plates but occasionally as islands on ridge crests.

6. Continental basalts (CON) are erupted through continental crust, largely at continental rifts in a within-plate setting.

In the discriminant diagrams ocean-island and continental basalts form only one field (WPB) that is the field of within-plate basalts, since they cannot be separated on the basis of their major element chemistry. This fact may prove that the nature of the crust, oceanic or continental does not modify these basaltic rocks by chemical interaction of the magma and wall-rocks.

Discussion of the Results

The calculated discriminant functions (F_1, F_2, F_3) have been plotted one against the other in Figs 2, 3 and 7.

Young Alkali Basalts

Figure 2 testifies that projection points of the young alkali basalts fall almost invariably in the WPB (within-plate basalt) field, owing to the low SiO_2 and high TiO_2 concentrations of these rocks, so typical of WPB-basalts (Table 2). In respect of major element content, the young alkali basalts are

Table 2

Means and standard deviations of major element oxides

	No. = 30 young alk. basalts	No. = 12 Cretaceous alk. basalts	No. = 22 diabases	No. = 10 gabbros	No. = 12 basaltic andesites
SiO_2	47.19	44.83	48.64	44.76	53.92
TiO_2	2.04	2.91	1.86	3.73	1.02
Al_2O_3	15.69	15.11	15.47	15.81	18.03
FeO	9.60	10.91	11.62	14.53	7.18
MgO	7.45	6.36	5.96	6.06	3.31
CaO	9.24	8.38	9.11	9.46	8.47
Na_2O	3.76	4.21	3.74	2.52	2.73
K_2O	1.92	1.16	0.21	0.16	2.07
		Standard Deviations			
SiO_2	1.71	2.84	1.94	1.88	1.98
TiO_2	0.47	1.15	0.55	1.10	0.22
Al_2O_3	1.40	2.47	1.85	1.95	1.66
FeO	1.08	2.13	2.65	2.15	0.92
MgO	1.11	3.97	0.79	1.17	1.09
CaO	0.75	2.68	1.22	1.52	0.92
Na_2O	0.69	3.21	0.70	0.69	0.45
K_2O	0.53	0.50	0.16	0.15	0.67

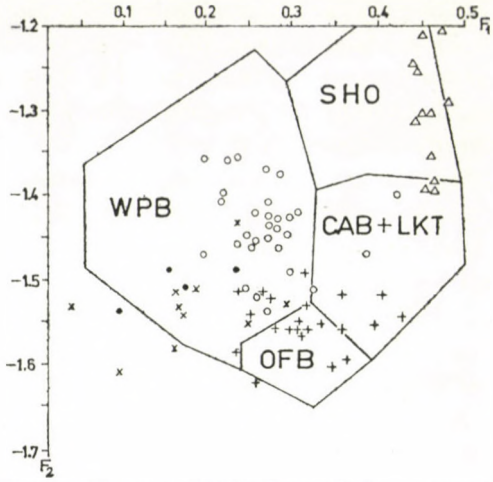

Fig. 2. Discriminant diagram (PEARCE, 1976). Open circles: young alkali basalts, triangles: basaltic andesites; crosses: diabases; x: gabbros, dots: soda gabbros from the Valley of the River Bódva

Table 3

CIPW norms of the magma-types calculated on the basis of average concentration values given in Table 2

		Young basalts	Cretaceous basalts	Diabases	Basaltic andesites
Q		—	—	—	6.26
or		10.04	6.23	1.03	12.76
ab		14.65	19.58	29.03	24.05
an		18.31	17.66	23.1	29.28
ne		7.68	7.42	—	—
di	Fe	3.33	3.57	3.65	3.16
	Mg	5.79	4.87	3.89	2.93
	Ca	9.63	8.78	7.72	6.17
hy	Fe	—	—	2.31	5.91
	Mg			2.41	5.65
ol	Fe	9.92	11.03	10.9	—
	Mg	15.35	13.79	10.39	—
mt		1.88	1.94	2.35	1.48
il		3.42	5.1	3.12	2.37

$Fe^{3+}/Fe^{2+} = 0.15$ theoretical value was used in every calculation.

rather homogenous judging from the small values of standard deviations. The CIPW norms (Table 3) calculated from the means of major element data (Fe^{3+}/Fe^{2+} was arbitrary taken to be equal to 0.15) show a typical ol-ne composition and since ne >5 per cent, the composition may be regarded as basanitic according to RINGWOOD's nomenclature.

Alkali basalts of continents and oceanic islands could not be distinguished from each other by statistical procedures (PEARCE, 1976). The chemical features of the alkalic basaltic liquids are largely dependent of two factors: the physicochemical conditions of the partial melting (P, T, P_{H_2O}, P_{CO_2}) and the subsequent fractionation of the crystallizing magma. Indeed, lavas of the young alkaline basaltic series in Hungary seem to represent either primary melts, that is mantle-derived magmas that have been unmodified by crystal fractionation or by other processes, or in a lesser amount, liquids that have been modified by olivine and perhaps pyroxene fractionation. The identification of basaltic liquids as primary or derivated may be carried out by using three criteria (e.g. FREY et al., 1978). First, the occurrence of spinel lherzolite nodules in a basalt proves that the magma has ascended from depth > 30 km without crystal fractionation or chemical interaction with wall-rock environments. In Hungary, most of the peridote nodules come from basaltic tuffs, however, lavas of Bondoróhegy, Badacsony (Balatonfelvidék), as well as Eresztvény and Magyarbánya (Cserhát) do contain such nodules (EMBEY-ISZTIN, 1976 A; 1976 B; 1977; 1978) so they can be regarded as primary by the first criterion. Secondly, the evidence from mantle-derived rocks (lherzolite xenoliths from basalts and kimberlites, diapiric intrusives of high P and high T peridotites, peridotite bodies exposed in modern mid-oceanic ridge and old ophiolite complexes) indicates that the least refractory upper mantle compositions have 100 $Mg/Mg + Fe^{2+}$ (Mg-values) around 88—89. If such a source material becomes melted (up to about 30 per cent) the basaltic melt must have Mg-values $\sim$68—75 using $K_{D(ol/liq)}^{(Fe/Mg)} = 0.3$ published by ROEDER—EMSLIE (1970). Considering this criterion in the case of the young alkali basalts, we can say that the basalt of Badacsony with a Mg-value of 67—70, and that of Sághegy (68—71), Nagyláz (70), Sarvaly (68), Pécskő (71) can be identified as primary partial melts. The basalt of Pálhegy (Pauliberg) with a very high concentration of MgO ($\sim$12 per cent) and NiO ($\sim$300 ppm, MAURITZ et al. 1948, SCHARBERT's personal commun.) must be regarded as a very primitive partial melt. On the contrary, basalts of Tátika (Mg = 51—67), Haláp (64—65) and Kabhegy (63—68) as well as some Cserhát basalts are to be considered as fractionated liquids. The depth of the fractionation cannot be defined with certainty, it may have occurred either in the mantel itself or at lower crustal levels.

The third criterion which is likely useful in distinguishing between primary melts and derivative magmas involves compatible trace elements such

as Sc, Cr, Co, and Ni. Since these elements have solid/liquid partition coefficients greater than unity for at least one of the major mantle phases, their concentrations in a magma are very sensitive to fractional separation of mafic minerals. Consequently, rocks having the highest concentrations from these elements have likely experienced very little fractionation or not at all. Especially Ni abundances have been used to identify magmas with very low fractionation rate of olivine (GAST, 1968; KAY et al., 1970; ALLÈGRE et al., 1977; FREY et al., 1978). Evidence from NiO concentrations published by MAURITZ et al. (1948) is not always in accordance with the one we can obtain from Mg-values. While the Pálhegy basalt has both a high Mg-value and NiO abundance (300 ppm) and the Tátika basalt with a low Mg-value only 0—100 ppm NiO, the Haláp basalt though lower in Mg-value contains also 300 ppm NiO. The basalt of Kabhegy has an intermediate amount of NiO (200 ppm).

Summing up, the result of the discriminant analysis has shown that the young alkali basalts of Hungary are typical WPB basalts that were erupted through a continental-type crust not too different from the present one, judging from geological and geophysical evidences. The basaltic magma has not been modified by chemical interactions with wall-rocks to any sensible extent. Among the lavas some represent primary melts, while others, liquids modified by olivine and perhaps pyroxene fractionation.

Lower Cretaceous Alkali Basalts

Only four of these rocks have MgO + CaO between 12 and 20, therefore analyses with more and less MgO + CaO have also been plotted (by different symbols) to evaluate the effect of differentiation (Fig. 3). It is remarkable that all the four "true" basalts fall in the WPB-field and so do some of the trachyandesitic differentiates though a part of them lies in the shoshonitic (SHO) field of the diagram. On the contrary the cumulative-type rocks (with more MgO + CaO than 20 per cent) have very low values of F_2 with respect to their F_1-values, consequently they do not fall in any of the fields, however, their vicinity to the WPB-field is significative. The analyses roughly follow two lines, one with increasing F_1 and F_2 and another one with slightly increasing F_1 and decreasing F_2. The first line represent the trachyandesitic trend of differentiation which leads to the highly undersaturated phonolites, the second one may be a complementary differentiation process essentially due to olivine and clinopyroxene accumulation. Considering all these facts, we can quite safely say that the parental magma of the Lower Cretaceous volcanism must be defined as a WPB-basalt. Since PEARCE gives a set of scaled eigenvectors too, we are able to evaluate the contribution of a given major element to the total discriminating power expressed by F_1, F_2, and F_3. For example, the largest scores on the second function (F_2) are for MgO (—0.74) and K_2O (+0.70).

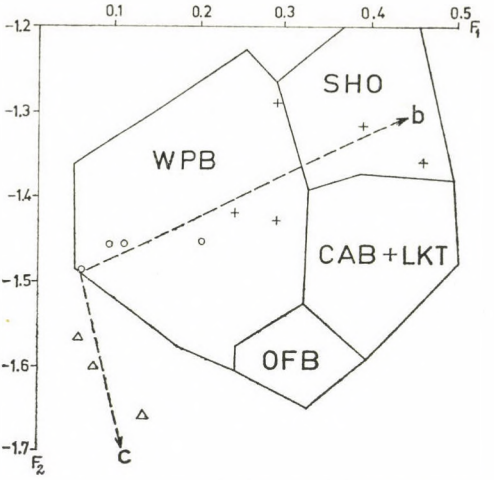

Fig. 3. **Discriminant diagram (PEARCE, 1976) for the Mecsek volcanic rocks. Open circlest "true" basalts; triangles: cumulative rocks; crosses: trachyandesites; lines b and c represen: two complementary differentiation trends**

Therefore the high magnesium and low potash cumulative rocks have typically low F_2-values and the high potash and low magnesium trachyandesitic rocks high F_2-values.

The calculated mean of the analyses (Table 2) can provide only a rough estimate as to the composition of the parental magma. The standard deviations are very high since the compositional range of the rock analyses is also large. The "mean" composition is that of an alkali basalt very low in SiO_2 and high in TiO_2. The calculated norms (Table 3) represent a typical ol-ne characterized basanite. The alkaline nature of the Mecsek basalts and trachyandesites is also obvious if we plot them in the diagram of MACDONALD and KATSURA (1964). It is remarkable that all the analyses fall to the left-hand side of the Hawaii alkalic-tholeiitic division line (Fig. 4), and none of them fall in the encircled area (Field of analyses from mid-Atlantic Ridge at 45° (AUMENTO, 1968). In the triangle Alk-Fe-Mg (Fig. 5) the differentiation trend of the Mecsek alkali basalt-trachyandesite-phonolite series is very similar to the Hawaii alkalic trend of differentiation, however, in the triangle Na—K—Ca (Fig. 6) differentiation trend seems to be disturbed. Especially intermediate and highly differentiated members were modified by some process of soda-enrichment. This process was probably a zeolitization, since according to MAURITZ (1913) the phonolites and intermediate rocks contain a great amount of natrolite and analcite.

100 Mg/Mg + Fe^{2+} (Mg-values are) generally too low except for the cumulative rocks, consequently even the most "primitive" basalts have been modified by olivine fractionation. No xenoliths of peridotite was described

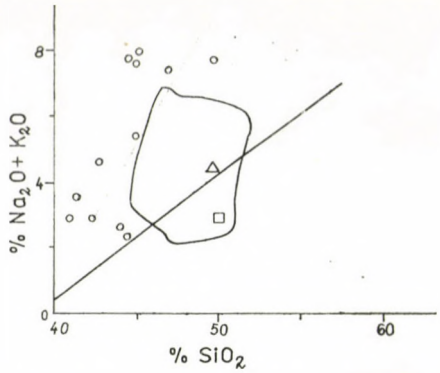

Fig. 4. Alkali-silica diagram for the Mecsek basaltic and trachyandesitic rocks (open circles). Alkali-tholeiitic division line after MACDONALD—KATSURA (1964), encercled area is the field of analyses from mid-Atlantic Ridge at 45° N after AUMENTO—LONCAREVIĆ (1969), AUMENTO (1968). Triangle is average spilite from POLDERVAART (1955). Square: average oceanic tholeiite after ENGEL et al. (1965)

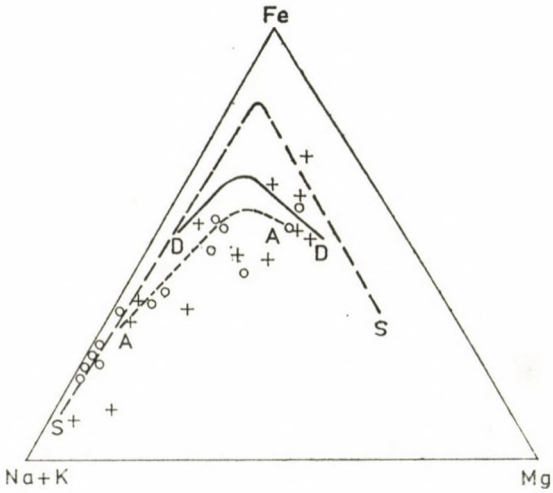

Fig. 5. Alk—Fe—Mg diagram. Open circles: alkaline basaltic → trachyandesitic → phonolitic rock series of the Mecsek Mts.; crosses: diabases and acid differentiates of the Bükk Mts.; S-S: differentiation trend in the Skaergaard intrusion after WAGER—MITCHELL (1951); D-D: diabase → granophyre series, Dillsburg, Pennsylvania after HOTZ (1953); A-A: Hawaiian alkaline basalt → trachyte series after NOCKOLDS—ALLEN (1954)

from these rocks. P_{O_2} of the crystallizing magma was probably high, leading to an early separation of iron ore, which is thought to be the first phase to separate on textural grounds (see: Mauritz, 1913).

The high differentiation rate of this rock series is in sharp contrast with the non-differentiated young alkali basalts discussed in the preceding section.

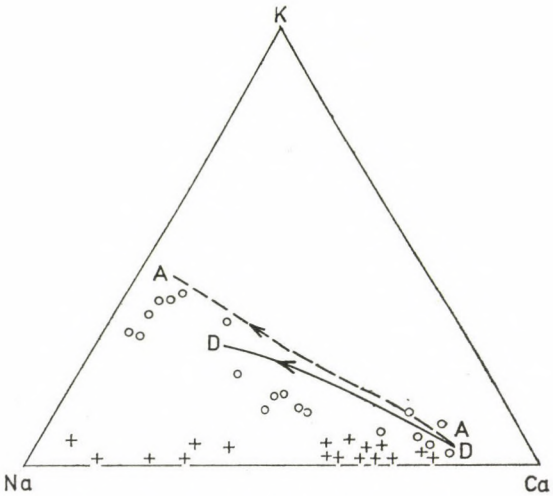

Fig. 6. Na—K—Ca diagram. Symbols are the same as in Fig. 5

Diabase-gabbro-oreperidotite series of the Bükk Mts.

Analyses of diabases have been plotted on the discriminant function diagrams in Figs 1 and 7. Both diagrams reveal that the analyses have clear OFB (ocean-floor basalt) affinities. The gabbros plot as within-plate basalts in Fig. 1 and island arc tholeiites (LKT) in Fig 7, and this inconsistency proves that the original constraints must be respected and the use of the discriminant diagrams should be restricted to lava rocks. The means and standard deviations

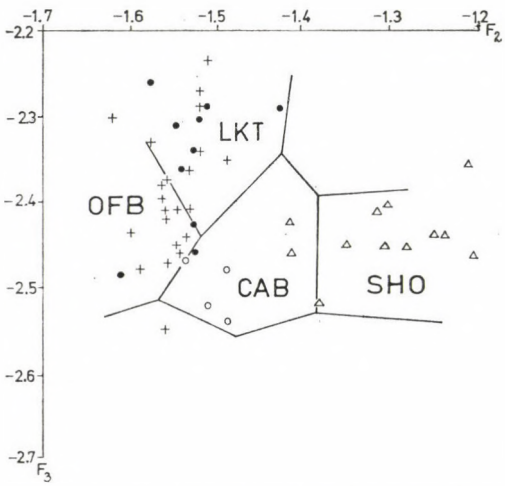

Fig. 7. Discriminant diagram (PEARCE, 1976). Crosses: diabases; dots: gabbros; open circles: soda gabbros; triangles: basaltic andesites

of the eight oxides are given in Table 2. The low potassium content is remarkable and this is one of the characteristics of ocean-floor basalts. Since potassium has the highest score (+0.70) on the discriminant function F_2, ocean-floor basalts plot on the lowest part of the diagram. Other features, however, show certain differences between the Bükk diabases and average ocean-floor basalts. Notably the high MgO and CaO concentrations of ocean-floor basalts are contrasted with relatively low concentrations in the diabases. On the other hand, the diabases are higher in total Fe than are ocean-floor basalts. We may think that the Bükk diabases were modified by olivine and clinopyroxene fractionation. Mg-values are indeed rather small (52—61). The separated olivine and clinopyroxene crystals accumulated at lower levels of the igneous mass forming small peridotite bodies. Evidence for the presence of cumulative texture in the largest peridotite body was given by NOSKE-FAZEKAS—EMBEY-ISZTIN (1978).

The CIPW-norms (Table 3) show that we are dealing with a hy-ol rock-composition that may be designated by the name of olivine tholeiite according to Ringwood's nomenclature. In the diagram of MACDONALD and KATSURA (Fig. 8) diabases plot at either side of the Hawaii alkalic-tholeiitic division line and in the field of mid-Atlantic Ridge basalts. The differentiation trend of the Bükk rock association is a peculiar one. Since the concentrations of potassium are very low, the differentiation proceeds along the join Na—Ca of the triangle Na—K—Ca (Fig. 6). A smooth descent line cannot be ascertained in Fig. 5, though it shows some tholeiitic affinity.

Considering the OFB-affinity of the Bükk diabases it seems appropriate to examine whether spilitization could have played a role in the chemistry of these rocks. Low potash and high soda concentrations are characteristic of spilites, however, there is only a slight deficiency in Ca and none in Al, though the latter is considered to be very mobile during spilitization (CANN, 1969; MELSON VAN ANDEL, 1966), finally there is no excess Si. Consequently diabases of the Bükk Mts. could not have been modified by spilitization to any greater extent.

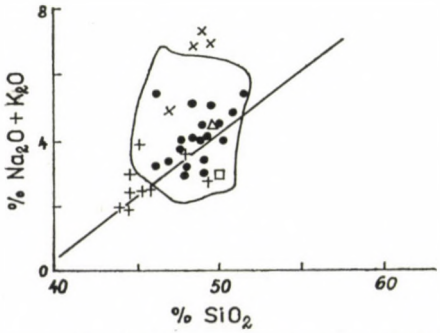

Fig. 8. Alkali-silica diagram after MACDONALD—KATSURA (1964). Dots: diabases; crosses: gabbros; x: soda gabbros. For other symbols and explications see: Fig. 4

Summing up, we can say that the mesozoic diabases of the Bükk Mts. are olivine tholeiitic in composition and they show OFB-affinities. The abyssal tholeiitic nature of these rocks was earlier emphasized by ONUOHA (1977) and NOSKE-FAZEKAS — EMBEY-ISZTIN (1978). The primary magma must have been modified by olivine and clinopyroxene fractionation and modifications due to spilitization could not be ascertained.

Basaltic Andesites

The sum of MgO + CaO in the analyses plotted in Figs 2 and 7 is only 11—13 per cent, so they are close to the lower limit of basaltic composition. Both in Figs 2 and 7 they largely plot as shoshonites. An interesting feature of these rocks is that while their F_1 and F_3-values remain essentially constant, F_2 is changing considerably. Alteration of the rocks could produce this effect (PEARCE, 1976) but only analyses of fresh rocks were used. On the other hand, we can explain the effect from the antipathetic variation of K_2O and MgO, oxides that have the largest scores on the second discriminant function (+0.70 and —0.74, respectively).

The 100 MgO/Mg + Fe^{2+} (Mg-values) of the basaltic pyroxene andesites are low, because of the low concentrations in MgO. In this respect they are very similar to the amphibole and biotite andesites and even to the dacites which they are intimately associated with. However, a primary and probably an undifferentiated nature cannot be categorically excluded for these magmas since experimental work by NICHOLLS (1974), KUSHIRO (1974) and others indicate that hydrous peridotite can yield andesitic melts at water pressures of 10—15 kbar. At the same time andesites and dacites can be considered as partial melts derived from eclogite or amphibolite (subducted oceanic crust; GREEN—RINGWOOD, 1968; T. GREEN, 1972).

The chemical composition of the basaltic andesites from different volcanic mountains are rather homogeneous and CIPW-norms calculated on the basis of mean values reveal a hy-qu characterized quartz tholeiitic composition (Table 3).

Tectonic Implications

It has been suggested, that present-day geophysical peculiarities (high heatflow-values, Moho discontinuity and LVZ in elevated position, gravitational anomalies etc.) of the Pannonian Basin may be explained if we suppose the presence of a thermal mantle diapir below this territory. According to SZÁDECZKY-KARDOSS (1971) and STEGENA et al. (1975) and GÉCZY et al. (1975) the mantle diapir is a consequence of the late Oligocene—early Miocene subduction of the oceanic basement of the Carpathian flish trough. In the late

Miocene and especially during Pliocene times the Pannonian Basin was formed by subsidence. At the same time the earlier compressional tectonism gave way to an essentially extensional tectonic regime. Thus, the youngest tectonic evolutionary stage of the Pannonian Basin is similar to that of continental rift zones, such as Auvergne and the Rhein Graben in Europe or the Baikal district in Siberia. The presence of primitive peridotite nodule-bearing alkali basalts with marked WPB-affinities are ubiquitous in the rift zones the generation of which must be connected with the mantle diapirism. Some features of the lherzolite xenoliths and especially an amphibolite/lherzolite composite nodule seem to support this suggestion (EMBEY-ISZTIN, 1976/a ; 1977; 1978).

The basaltic (pyroxene) andesites along with other andesites, dacites and rhyolites could have been formed during the Miocene due to the subduction mentioned above. The SHO-affinities of basaltic andesites may be in accordance with this suggestion. Since the oceanic or suboceanic area subducted could not have been too large in the case of the Carpathian arc, judging from geological conditions, magmatic evolution may have reached a mature stage characterized by SHO-affinity very soon or else magmas of earlier stages with LKT and CAB-affinities were completely lacking.

The discriminant analysis and features of the differentiation trend strongly suggest a within-plate origin for the Mecsek volcanic suite. This suggestion is supported by the geological and sedimentological considerations proposed by VADÁSZ (1935) who claimed that the Mecsek area was a sinking continental platform at the time of the volcanic eruptions. Thus the existence of a rift-zone in late Jurassic and early Cretaceous times seems to be established. Whether or not this rifting passed into an oceanic stage of evolution remains a question to be solved.

It must be mentioned, however, that the present interpretation is at variance with the ideas proposed by SZEPESHÁZY (1977) who thinks that the Mecsek volcanic alkali basalt association is a member of an ophiolitic suite continuing to NE below the Great Hungarian Plain. The lack of mantle-derived ultrabasic rocks, pillow lavas, cherts and basalts with OFB-affinities in this zone does not seem, however, to strengthen his views.

Finally, the results of the present paper suggest an ocean-floor origin for the diabase-gabbro complex of the Bükk Mountains. I should like, however, to emphasize that this suggestion does not necessary imply that the Bükk volcanic complex should be a "typical" ophiolite zone (e.g. ONUOHA, 1977) and even less the existence of another typical ophiolitic zone buried in greater depth between Zagreb and the Bükk Mountains (e. g. SZEPESHÁZY, 1977). It must be emphasized that mantle-derived ultrabasic rocks are completely lacking in the Bükk Mts., and though at some (very restricted) places pillow-like structures can be seen, the exact nature of this structure has never been investigated. Cherts on the top of diabases are also lacking and the relation of

the igneous and sedimentary rocks is a typical intrusion with a marked contact metamorph aureole in the sedimentary member. The last feature seems to be lacking in the case of really typical ophiolite suites such as the Troodos Massif in Cyprus. These difficulties and available evidence on the nature of the igneous activities detected in the basement of the Pannonian Basin does not favour the hypothesis of two subparallel ophiolite zone (supposed to be in more or less original situation!) running through Hungary. On the other hand, it has been recently proposed that the Carpatho—Pannonian region was formed as a result of a collision of two microplates of European and African origin respectively (GÉCZY, 1973; CHANNEL and HORVÁTH, 1976). Considering this hypothesis we can say that the Mecsek rift-volcanism is situated just at the border of the European microplate, thus it may be a witness of the beginning of rifting, that is, the individualization of this small portion of lithosphere as a microplate. The Bükk diabase-gabbro complex along with other basic (in Slovakia also ultrabasic) occurrences are situated on the African microplate along a major tectonic line, the Balaton — Darnó line which is separating the Austroalpine unit from the Southern Alpine—Dinaride unit. It seems thus, that tectonic emplacement of basaltic rocks with a possible ocean-floor origin or ultrabasic rocks probably of the same origin (both of them may perhaps represent desintegrated parts of an earlier ophiolite complex) took place along this tectonic line, at least in the northern part of the Pannonian region. It can not be categorically excluded that this emplacement had some analogous features with the mechanism of true ophiolite forming processes, but available evidence strongly suggest, however, that it could not have certainly been a "typical" one.

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СТАТИСТИЧЕСКОЕ ИССЛЕДОВАНИЕ ГЛАВНЫХ ЭЛЕМЕНТОВ ВЕНГЕРСКИХ БАЗАЛЬТОВ

А. ЭМБЕИ-ИСТИН

Резюме

На основании статистического изучения различных по возрасту типов магм основного состава из Венгрии автор статьи с помощью диаграмм Пирса (Pearce) определил наиболее вероятное тектоническое положение этих типов магм. Плио-плейстоценовые базальты Балатонского нагорья и Черхата обнаруживают сродство с магмами внутри континентальных плит, и таким же образом следует оценить и нижнемеловой вулканизм гор Мечек. В противоположность этому, толща диабазов и габбро гор Бюкк обнаруживает ярко выраженное сродство с магмами, приуроченными ко дну океанов. Базальтовые андезиты в свою очередь сродны шошонитам.

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EL-MORGAN OIL FIELD CRUDE OIL AND CYCLES OF OIL GENERATION, MIGRATION AND ACCUMULATION IN THE GULF OF SUEZ PETROLEUM PROVINCE, A. R., EGYPT

By

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Crude oils producible from the Gulf of Suez petroleum province, oil fields have variable specific gravities in different lithostratigraphic horizons from the upper Paleozoic upward to the Miocene. Some oil fields show vertical gradation from heavy oil to relatively lighter one upwards; while others show the reverse. El-Morgan oil field as the other off-shore oil fields differ from the on-shore fields in the same province as regarding their oil gravities attitudes. The study and interpretation of the geochemical characteristics and the variability of the crude oils gravity would result in two significant and distinct categories of crude oils as well as two cycles of oil generation, migration and accumulation in the Gulf of Suez petroleum province.

Introduction

This paper deals with some oil-geological investigations of El-Morgan off-shore oil field, Gulf of Suez petroleum province, A. R., Egypt. The field lies in the Gulf water between longitudes 33° 17′ and 33° 30′ E, and latitudes 28° 05′ and 28° 33′ N (Fig. 1), covering an area of about 46 km^2.

El-Morgan oil field lies some 220 km south of Suez town, about 40 km southeast of Ras-Gharib and 20 km from Ras-Shukeir where the land installations and marine terminal are set up. The field is some 13 km to the west of El-Tor village where, in the off-shore area, a rich development of coral reefs known as El-Tor Banks are present, this gave rise to the Arabic name "El-Morgan" to the field. El-Morgan oil field lies within an area that comprises a group of oil fields in the Gulf of Suez petroleum province, e.g. Belayim off-shore and on-shore (known as Belayim marine and land or Belayim west and east), Feiran and Abu-Rudeis oil fields to the north; and Amer, Bakr, Ras-Gharib, Kareem, Um El-Yusr, Kheir, Shukeir, July and Ramadan oil fields to the northwest and west (Fig. 1).

The Gulf of Suez, the Gulf of Aqaba and the Red Sea proper represent three different structural elements that are genetically related to one of the principal structural features of the African continent, the great rift valley system of east Africa (SAID, 1969). The Gulf of Suez which is situated within the stable shelf of Egypt, constitutes a zone of subsidence throughout its geological history and has been the site of an immense accumulation of sediments. The cumulative thickness of the sedimentary section is 10 000 m. and the Gulf

is essentially a taphrogeosyncline. The region possesses a unique sedimentary history when compared with the bordering regions or with Red Sea proper of which it is today a part. The Gulf of Suez basin started at least as early as the Upper Paleozoic in the form of a graben that probably encompassed present-day southern Sinai. The basin developed as a result of a master fault pattern along the oversteepened sides of two interbasin arches that are assumed to have penetrated into the major sedimentary basin of the unstable shelf of Egypt and the Levant. This situation continued to the end of Lower Eocene time. During this interval the intermittent activation of the blocks that constitutes the sides kept rejuvenating the graben.

The birth of the Gulf of Suez graben in its present form (SAID, op. cit.) dates back to the Oligocene when the Gulf represented an uplift region on top-of which the modern graben was formed in Late Oligocene time. By contrast, the Red Sea proper formed part of the stable shelf throughout its history and is a feature which developed in Lower Miocene time.

SAID (1964) stated that the movements that affected the Gulf were tensional. The Gulf is bordered by and is made up of large number of blocks that were continuously rising and sinking with varying magnitudes and intensities. Some of these blocks are so small and old at one locality (e.g. 82 km along the Suez—Ras Gharib road) two adjacent blocks, have different stratigraphic successions.

ABDEL-GAWAD (1969) stated that the Gulf of Suez area has generally been regarded as a graben structure controlled by NW—SE normal faults. There is considerable evidence, however, that it is profoundly influenced by intersecting systems of faults trending E—W, NE—SW, NW—SE and WNW—ESE. The N—S and WNW—ESE faults are particularly responsible for the Zigzag shape of the coast lines, the outline of the Pre-Cambrian shield, and major drainage lines west of Suez Gulf. The influence of these faults on the shield rocks of Sinai is equally evident. Along many of these faults lateral displacements are observed, a notable example is the left — lateral off-set of gebel Araba block in western Sinai along N—S faults are inferred across Esh El-Mellaha range west of the Gulf of Suez.

Thus this paper is intended for a regional oil geological investigation on El-Morgan field in relation to the other surrounding fields in the Gulf of Suez Petroleum Province.

METWALLI et al. (1978) investigated and elaborated the stratigraphic and tectonic setting of El-Morgan field, with special reference to the structures produced by the salt movements in the area, which affected to some extent the hydrocarbon accumulation and the source—reservoir relations and stand as a puzzling problem to the exploration activities in the Gulf of Suez region till the present. In addition, the geochemical characteristics of the crude oil producible from the two pay-zones in El-Morgan field within the Miocene

successions named: Belayim and Kareem Formations (pay-zones) are investigated in a trial to throw more light on the problem of oil migration and accumulation in the field. The geochemical characteristics of the oil produced from same nearby oil fields are also discussed in an attempt to clarify the source-reservoir relations in the Gulf of Suez petroleum province.

The Gulf of Suez area was extensively explored by surface geological and geophysical methods; exploratory drilling has been active in the province since 1838. Most of the reserves of Egypt in the Gulf of Suez area were thought to be found in the Miocene rocks, e.g. sands in El-Morgan and Belayim fields, also other relatively minor accumulations in Miocene sands and sandstones in Sudr, Um El-Yusr, Shukeir, Kareem, Feiran and Rudeis-Sidri oil fields. However, the Miocene sediments of reefal facies are oil-bearing in Ras-Gharib, Bakr, Kareem, and Gemsa oil fields.

Pre-Miocene reservoir rocks ranging in age from Eocene to Paleozoic are also oil-bearing in Hurgada, Ras-Gharib, Bakr, Amer, Kareem, Sudr, Asl, Ras-Matarma and Belayim off-shore oil fields. Accordingly, Miocene sandstones and reefal sediments together with Pre-Miocene sands and limestones are the main oil-bearing rocks. The Miocene shales, marls and evaporites form the sealing rocks for the majority of oil traps in the Gulf of Suez petroleum province.

A considerable part of the off-shore areas in the Gulf of Suez region witnessed and conquered the former seismic techniques for mapping below the Miocene salt section represented by the South-Gharib Formation. However, hopeful and expected prospects (METWALLI et al., op. cit.) are thought to be trapped in the Pre-Miocene faulted-blocks, which are proved to be oil-bearing horizons in many localities in the Gulf as a result of the exploration activities by GPC and other oil companies. Accordingly, the production of oil and gas comes from different lithostratigraphic horizons of Miocene, Eocene and/or Mesozoic or older rocks.

The General Geochemical Setting of El-Morgan Crude Oil in Relation to other Oil Fields in the Gulf of Suez Petroleum Province

Crude oils in the Gulf of Suez petroleum province oil fields have variable specific gravities in the different lithostratigraphic horizons, these may also vary within the same horizon (Tables 1—6). The crude oil gravity is the sum of the chemical changes accompanying its maturation and modification during genesis in the source rock, and/or the result of the geochemical changes resulting from migration.

SILVERMAN (1965), on the basis of the sum of the chemical changes accompanying petroleum migration, suggested that the migration of a whole

Table 1

Geochemical characteristics of crude oil from El-Morgan oil field

Well No.	1	1	1	1	2	5	6	7	7	8	8
Interval in ft.	5043 5090	5870 5920	6045 6109	6247 6322	6270 6325	6369 6416	6000 6030	5175 5226	5965 5999	5142 5160	5680 5710
Stratigraphic Horizon Hammam Faraun	Kareem							H. Faraun	Kareem	H. Faraun	Kareem
Age	Middle Miocene	Lower Miocene						M. M.	L. M.	M. M.	L. M.
Sp. gr. at 60 °F	0.911	0.885	0.878	0.875	0.873	0.894	0.891	0.878	0.871	0.869	0.882
API Gravity	23.70	28.40	29.50	30.20	30.50	26.65	27.50	29.70	30.80	26.30	28.8
Viscosity Red. I at 37.8 °C	183	71	69	61	61	117	83	68	55	118	51
Asphaltenes Wt%	6.20	1.50	1.22	1.21	1.1	5.8	1.68	1.55	0.47	3.23	1.45
Paraffin Wt%	4.51	4.48	4.5	6.73	5.0	3.2	5.2	5.88	9.92	7.5	6.7
Sulfur Wt%	2.40	1.60	1.60	1.60	1.55	1.90	1.75	1.90	1.60	2.20	1.55

Table 2

Geochemical characteristics of crude oil from Belayim on-shore oil field

Well. No.	112–21	112–23	112–31	113–10	113–1	113–2	112–3	112–1	112–4
Depth in ft.	8050	8230	7860	7380	8000	8000	7580	7110	4940
Stratigraphic Horizon	Sidri							Feiran Zeit	
Age	Middle Miocene								
Sp. gr. at 60 °F	0.95	0.927	0.939	0.910	0.918	0.919	0.924	0.92	0.913
API Gravity	17.44	21.4	18.98	24.05	22.63	22.48	20.55	22.3	23.48
Viscosity at 100 °C in E.	31.7	11.4	19.4	3.65	10.3	10.8	11.34	8.59	8.25
Asphaltenes Wt%	13.0	9.1	10.6	10.3	12.2	12.4	12.54	9.42	12.0
Paraffin Wax Wt%	2.9	2.8	3.9	3.6	3.85	2.83	4.5	2.4	3.9
Carbon residue Wt%	12.9	9.8	10.9	6.6	9.05	9.5	9.6	8.4	8.6
Sulfur content Wt%	3.59	3.30	3.58	2.60	2.81	2.85	2.89	2.20	2.12

N.B. The crude oil in Belayim on-shore (So-called Belayim land) oil field is proposed in this work to be of Miocene Source rock; and the successive attitude of the gravities of oil within depth intervals shows clearly the migration differentiation mechanism i.e. lighter oil overlies vertically relatively heavier one.

petroleum results in an increase in the contents of paraffinic hydrocarbons and a decrease in the content of asphaltenes, resins, porphyrins, non-hydrocarbons, and heavy metals (nickel, vanadium and copper), as well as a slight decrease in C^{13}/C^{12} ratio. If some or all of these changes occur, the migrated petroleum should undergo an increase in API gravity (decrease in density). He concluded that the postulated changes resulting from the secondary migration are similar, if not identical to the chemical changes occurring during maturation of petroleum. SILVERMAN (op. cit.) gave the north-south structure of the Quirequire oil field, Venezuela, which provides an example of chemical differences of crude oils, probably induced by migration. The productive zone lies in the non-marine Quirequire Formation of Pleistocene-Pliocene age, which unconformably overlies truncated, folded and faulted marine strata of Cretaceous, Eocene, Oligocene and Miocene ages.

MATHEWS et al. (1970) gave the two Moonie crude oils (Surat basin, Queensland, Australia) as an example where the Moonie lighter oil of the upper zone could have been derived from Jurassic shale, by separation migration along fractures from the lower reservoir. They reviewed that the production in the Moonie oil field is from two zones in the fresh water Precipice sandstone of early Jurassic age, where a few wells produce from the upper unit.

METWALLI (1974), cited the two Um El-Yusr crude oils (Gulf of Suez region, longitude 33° 30′ W and latitudes 38°—28° 3′ N), as an example where the upper crude in Ayun Member (Miocene) of 29.9° API has been derived from a lower one in Yusr Member (Miocene) of 21° API by separation migration along faults and unconformity surfaces up dip. The two oil zones are separated by about 300 ft of shales and marls where a few wells produce oil from the upper zone.

BARTON (1934), PRATT(1934), HLAUSCHEK (1936), BROOKS (1948 and 1949), HAEBERLE (1951), ZOBELL (1952), HUNT (1953), HENGSTEBECK (1959), ERDMANN (1961, 1964), SMITH (1963), KARTESV (1964), LOUIS (1964), SILVERMAN (1964), LONG et al. (1964), KERTAI (1962 and 1967), HOLMQUEST (1966), FLOURIA (1967), LANDES (1967), MATHEWS et al. (1970), METWALLI (1970), METWALLI et al. (1972), METWALLI (1974), and METWALLI and ABDEL-HADY (1975) and others, have all contributed on the origin of petroleum hydrocarbons and the causes of the variability of their composition.

In El-Morgan oil field two main productive zones are recorded within the Miocene Formations by GUPCO. The upper zone is in the Belayim Formation (Hammam Faraun Member) and produces oil of 23° API (Morgan well no. 1, M-1), 26° API (M-8), and 29.7° API (M-7). The lower zone is in the Kareem Formation and produces oil of 30.8° API (M-1); 28.8° API (M-8) and 30.8° API (M-7). GUPCO production section subdivided the Kareem Formation (pay-zone) in El-Morgan oil field into two territorial general portions; a northern portion and a southern one depending on their following observations:

1. There is about 400 psi pressure difference between the two portions, in spite of the fact that the original reservoir pressure is the same.

Table 3

Geochemical characteristics of crude

					Bakr
Oil Field	Bakr South				
Age	Middle Miocene	Lower Miocene	Eocene	Turonian	Turonian
Sp. gr. at 15 °C					
60 °F	0.9176	0.9190	0.9351	0.9265	0.9314
API Gravity	22.7	22.45	19.8	21.25	20.4
Sulfur content Wt%	4.11	4.00	4.75	4.03	3.33
Carbon residue Wt%	9.70	8.90	9.90	9.80	10.80
Ash content Wt%	0.03	0.03	0.02	0.03	0.13
Asphaltenes Wt%	8.55	8.10	9.70	9.40	9.10
Viscosity at :37 °C., Sec.	590	625	1520	957	1260
60 °C., Sec.	201	209	480	305	372
93 °C., Sec.	62	68	140	97	105

N.B. From Table 3, it can be seen that the stratigraphic succession of oil gravities could imply two cycles of oil migration and accumulation in Bakr oil field. An upper cycle for the Tertiary rocks and a lower cycle for Mesozoic and older oil-bearing rocks. While the miscellaneous oil gravity is that of the Devonian (Sp. gr. 1.0126; 8.2° API) which reflects a very heavy

Table 4

Some geochemical characteristics of crude oil from Ras-Gharib oil field

Age	Miocene	U. Cretaceous	L. Cretaceous	Carboniferous
Sp. gr. at 60 °F	0.895	0.895	0.920	0.903
API gravity	26.1	26.1	22.3	23.0
Viscosity Red. I at 100 °F	46.5	39.0	52.5	68.0
Paraffin wax Wt%	6.3	6.1	3.6	6.6
Sulfur content Wt%	3.1	3.1	4.2	3.7

N.B. The crude oils from Ras-Gharib oil field are of interest where an upper Cretaceous limestone is truncated and overlapped by organic fossiliferous marl and shale. This upper Cretaceous limestone trapped oil of 26.1° API, that migrated laterally into sandstone of Nukhul-Rudeis Formation (Miocene).

The source rock is proposed to be of pre-Miocene; Cretaceous age and the successive attitude of oil gravity within depth or stratigraphic intervals shows clearly the migration differentiation mechanism i.e. lighter oil overlies laterally and vertically relatively heavier one.

oils from Bakr and Amer oil fields

Oil Field							
					Bakr North		Amer
Ceno-manian	Lower Creta.	Upper Carbon.	Devonian	Lower Miocene	Eocene	Turonian	Turonian
0.9365	0.9405	0.9651	1.0126	0.9304	0.9309	0.9276	0.9197
19.6	18.95	15.15	8.2	20.6	20.5	21.05	22.3
3.41	3.70	4.60	3.87	5.13	5.15	3.81	3.50
10.60	10.63	10.60	12.60	10.00	9.70	9.50	10.00
0.28	0.28	0.11	0.07	0.02	0.01	0.03	0.03
10.28	10.35	11.50	12.31	10.15	10.05	8.20	7.00
4950	3700	17590	30.88	7.95	860	725	338
1010	860	2690	4840	2.62	290	225	—
240	150	400	1300	90	95	82	—

oil in comparison with the other oils produced from the Gulf of Suez oil fields. But the Carboniferous? oil seems to be derived by migration from the underlying one. While the Turonian oil in Amer oil field seems to bear nearly the same geochemical characteristics as the Turonian oil in Bakr oil field.

Table 5

Geochemical characteristics of crude oil from Kareem oil field

Stratigraphic Horizon	Rudeis	Rudeis — Nukhul	
Age	L. Miocene	L. Miocene	Turonian
Sp. gr. at 60 °	0.9504	0.9485	0.9733
API Gravity	17.3	17.7	13.9
Sulfur content Wt%	3.88	3.93	3.39
Viscosity Red I at 100 °F	7480	4530	22300 ?
Asphaltenes Wt%	10.00	9.29	11.90
Carbon residue Wt%	12.4	12.0	12.8

N.B. The crude oils in Kareem oil field are producible from two pay-zones, the upper one from the Lower Miocene (oil of 17.3° API) and the lower one is from Middle Cretaceous (Turonian) oil of 13.9° API i.e., the Miocene oil is of considerably varying geochemical characteristics than the Cretaceous one. The source rock can not be located in one age simply.

Table 6

Some geochemical characteristics of crude oil from Hurgada oil field

Well No.	131	136	144	144	74
Depth in ft.	1780	1500	2125	2706	1700
Age	Middle Miocene	Lower Miocene	Carboniferous		Devonian
Sp. gr. at 60 °F	0.873	0.8735	0.865	0.880	0.896
API Gravity	30.55	30.5	32.0	29.3	26.4
Viscosity Red at 50 °C in E	1.8	1.75	1.58	1.57	1.95
Ash content Wt%	0.4	0.4	0.02	0.92	0.76
Sulfur content Wt%	1.4	1.4	1.4	1.3	1.5

N.B. The Lower and Middle Miocene pay-zones bear oil of the same chemical characteristics which is relatively lighter than the underlying oil in the Carboniferous and may be older ? pay-zones.

2. The existence of a gas-cap in the northern portion only penetrated by the discovery well no. 1 (M-1).

3. The measured reservoir and production data show that oil is of higher API quality in the northern portion.

The subdivision of the field into northern and southern portions might be a reflection of the tectonic movements that affected the study area and modelled the main reservoir traps into a group of fault blocks, recognized and traced on isopachous and structural contour maps and cross sections by METWALLI et al. (1978). These fault-blocks which made up the oil reservoir and traps in El-Morgan oil field are dragged along their boundaries (METWALLI et al., op. cit.), so that they are erroneously interpreted as folds by HASSAN (1974, 1975) and KHALED (1974).

The gas cap recorded in the northern portion of the field is at a lower structural attitude, on the other hand, the wells that penetrated the Kareem Formation in the southern portion are at higher structural attitudes and the reservoir traps are without gas cap. Also, the gas-oil contact in the northern portion of the field is at —5800 ft which is in a lower attitude than that in the southern portion.

The lateral variability of crude oil gravity within the same formation in El-Morgan oil field could be simply related to the effect of lateral migration up dip from one fault block to another. The question that arises here is: Are the two El-Morgan crudes of the same source or of different sources? If these two crude oils are of the same source, we might find that the upper crude of the Belayim Formation relatively lighter than that of the Kareem pay-zone, as it is used by the well accepted rules of petroleum migration (SILVERMAN, 1965; METWALLI, 1974), but this is not the case. In El-Morgan oil field

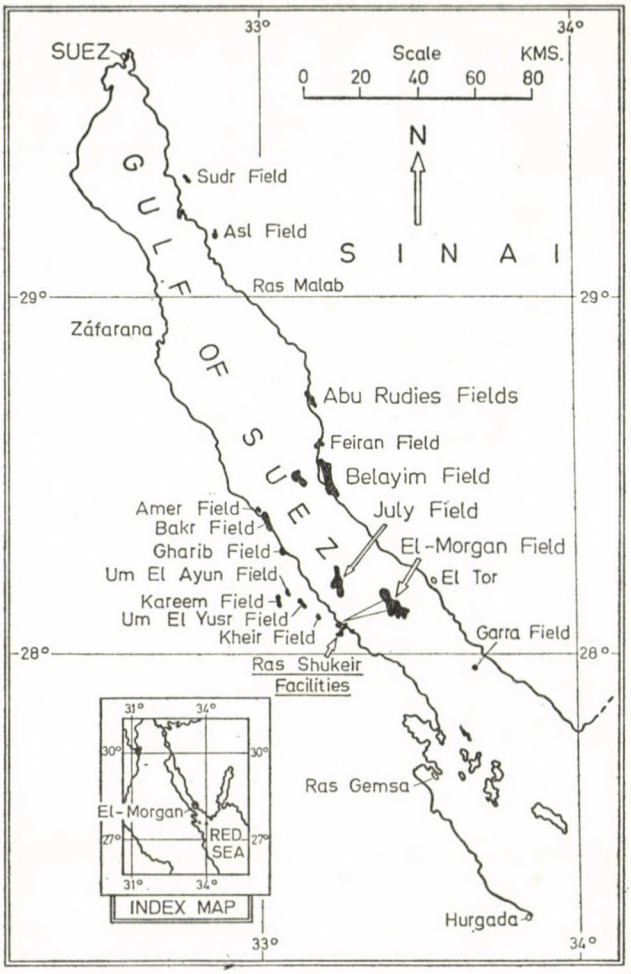

Fig. 1. Location map

a relatively heavy oil overlies a lighter one, i.e. "the deeper the oil, the lighter the oil". This observation leads to another interesting question: Are the two crude oils in El-Morgan oil field (Belayim and Kareem pay-zones) of different sources? To answer this interesting question the following must be discussed: Figs 2 and 4, which give the relation between the specific gravities against depth; and the total S wt% against depth, respectively for El-Morgan crudes, show more than one line of relation. One line is clear for the oil in the Belayim pay-zone and two intersecting lines of relation for the oil in the Kareem pay-zone. The dislocation of these lines of relation might be fairly related to different sources for Belayim and Kareem crude oils in El-Morgan oil field. This

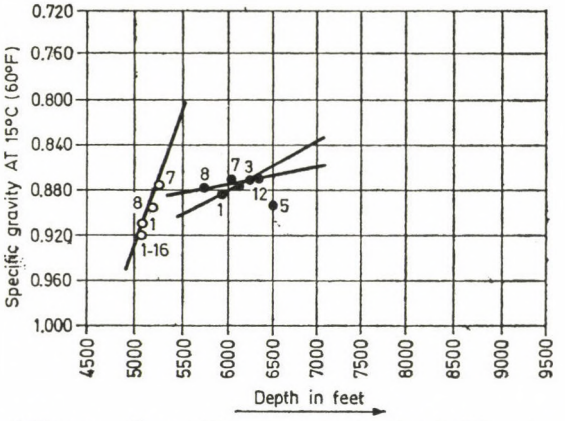

Fig. 2. **The relation of Miocene oil specific gravity to depth of oil-bearing rock, El-Morgan oil field. ○ Belayim pay-zone; ● Kareem pay-zone; ○⁷ well number**

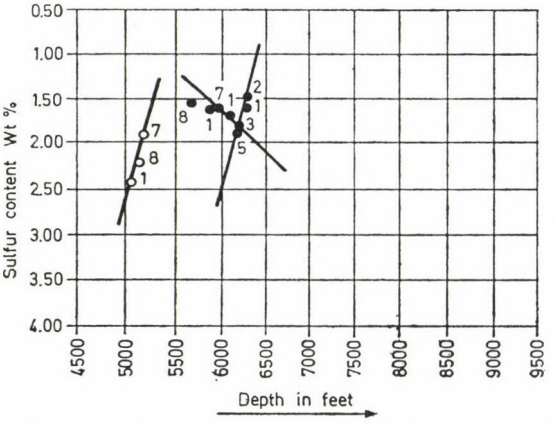

Fig. 3. **The relation of Miocene oil sulphur wt% to depth of oil-bearing rock, El-Morgan oil field**

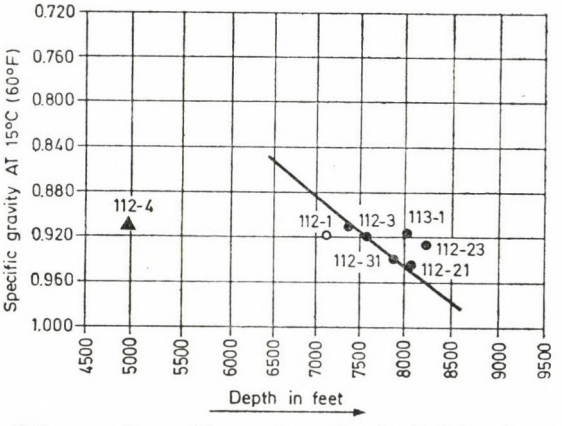

Fig. 4. **The relation of Miocene oil specific gravity to depth of oil-bearing rock, Belayim on-shore oil field. ▲ Zeit pay-zone; ○ Feiran pay-zone; ● Sidri pay-zone**

view is in harmony with HOLMQUEST (1966). The intersection of the two lines of relation of the Kareem crude oil can be considered as a probable indication of mixing of the Kareem crude oil as a result of the active faulting movements that modified El-Morgan fault-block traps.

Accordingly, the El-Morgan two main crude oils in the Belayim and Kareem Formations can be considered as derived from two probable separate sources. The two pay-zones are separated by the anhydrite of the Belayim Formation (that underlies the Belayim pay-zone) which seals the lower pay-zone, while the upper pay-zone is roofed and sealed by the thick salt of the south Gharib Formation.

The crude oil gravity, being the sum of the geochemical characteristics of petroleum during its life history, is a sensitive and important criterion. The present authors used the specific gravity and the total sulphur wt% among the general geochemical characteristics given in Tables 1—6 for comparison and interpretation of the studied oils, since these are the two criteria usually determined for most of the crudes in the Gulf of Suez petroleum province. However, they reflect other geochemical characteristics or they are related to them.

Based upon the geochemical data cited in Tables 1—6, interesting questions concerning the cause of the vertical variability of specific gravities of the crude oils in time and space, and its controlling factor or factors arise. To promote the solution of this interesting problem Figs 2, 4, 5 which show the relation between sp.gr. and depth or time of the oil-bearing rocks; and Figs 3 and 6, which show the variation of the total sulphur content with time are constructed.

The analysis and interpretation of the above tables and figures resulted in two significant and distinct categories of crude oils in the Gulf of Suez petroleum province:

(A): In the off-shore oil fields (Gulf of Suez), the general rule "the deeper the oil; the lighter the oil" is clear. In El-Morgan well no. 1 (M-1) in its two successive oil horizons within the depth interval 5000—6406 ft; four distinct specific gravities are recorded for the crudes from top downwards. These are 0.9110, 0.8848, 0.8784 and 0.8745. These four values of specific gravities in M-1 do not conform with values from other wells in the same field, e.g. M-5, M-7 and M-8. In Figure 2 three lines of relation between sp. gr. and depth can be seen for the Belayim and Kareem pay-zones (crude oils). In Fig. 5, which represents the relation between specific gravity and depth for the crude oils in El-Morgan, July and Belayim off-shore oil fields, four lines of relation of specific gravities against depth can be seen.

METWALLI et al. (1978) showed that the wells M-1 and M-8 in El-Morgan field are located in horst blocks; M-5 and M-6 are located in step fault-blocks, while M-7 is located in a graben fault-block.

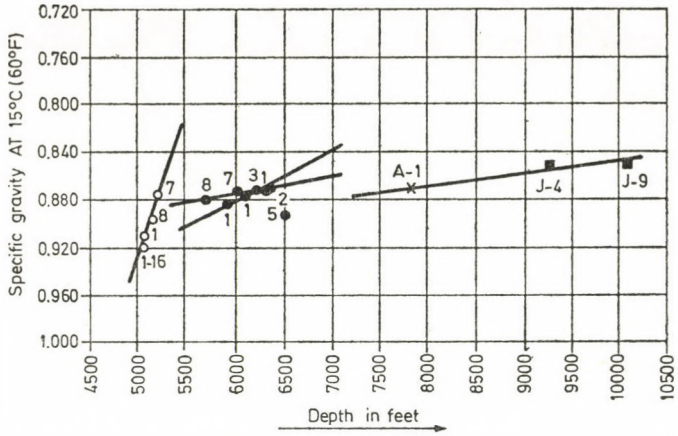

Fig. 5. The relation of Miocene and Upper Paleozoic oil specific gravity to depth of oil-bearing rock in El-Morgan, July, and Belayim off-shore oil fields, Gulf of Suez region. ○ Belayim (Hammam Faraun) pay-zone; ● Kareem pay-zone; × Rudeis pay-zone; = Nubia pay-zone; ○-1, ■ J-4, × A-1: well number

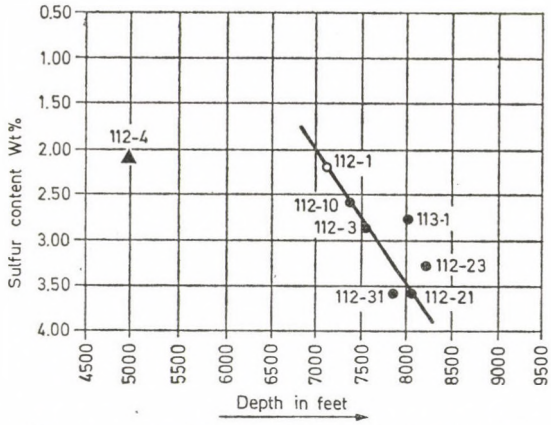

Fig. 6. The relation of Miocene oil sulphur wt% to depth of oil-bearing rock, Belayim on-shore oil field

Also July wells No. 4 and No. 9 (J-4 and J-9) in July oil field are located in one horst block. Belayim well no. 1 (A-1) in Belayim oil field is located also in a fault-block. Accordingly, from the above cited observations, one could state that the vertical variability of specific gravities of the crude oils in the Gulf of Suez petroleum province is controlled by their structural attitude, i.e. the type of fault traps either horst, graben or step faulted blocks. The distinct displacement of lines of relations (Figs 2 and 5) indicates different fault-blocks. However, in Fig. 2, the displacement of the lines of relations does not only show distinct fault-blocks, but also distinguishes the oil-bearing rocks in the Belayim pay-zone (Miocene) from that of the underlying pay-zone.

(B): In the on-shore oil fields, Gulf of Suez petroleum province, e.g. Belayim on-shore (so-called Belayim land), Um El-Yusr, Bakr, Gharib and Hurgada oil fields, the separation migration mechanism of SILVERMAN (1965), or the migration differentiation mechanism of METWALLI (1974) is clearly represented in the Gulf of Suez off-shore oil fields, i.e. a lighter crude oil overlies a relatively heavier crude oil in the reservoir traps in the same and in different ages.

The specific gravity of Belayim on-shore oils and total sulphur wt% in relation to depth are plotted in Figs 4 and 6. Also the API gravities of an example of these crudes are plotted diagramatically against time in Fig. 7, for the southern part of Bakr oil field.

It is of interest to note that El-Morgan, July and Belayim off-shore oil fields which fulfil the rule "the deeper the oil, the lighter the oil" are located in the off-shore concession, i.e. in the major graben that formed the Gulf of Suez. Oil fields of the second category, e.g. Um El-Yusr, Belayim on-shore, Kareem, Bakr and Hurgada oil fields which show the chemical differences induced by migration are located in the on-shore concessions, i.e. at the foreland sides of the Gulf of Suez major graben.

The first category is the traditionally accepted rule in petroleum geology, however, the second exceptional rule is a direct concept of the changes accompanying the phases of the secondary migration of petroleum.

METWALLI (1974) visualized the Miocene subsurface section in Um El-Yusr oil field area as a huge geochromatograph (acting as chromatographic column) for the differentiation of an upper lighter crude (29.9° API) derived from a lower relatively heavier one (21° API) across faults and unconformity surfaces. In addition, METWALLI and BASHAT (1974) stated that, the presence of numerous hydrocarbon accumulations of minor dimensions in Um El-Yusr

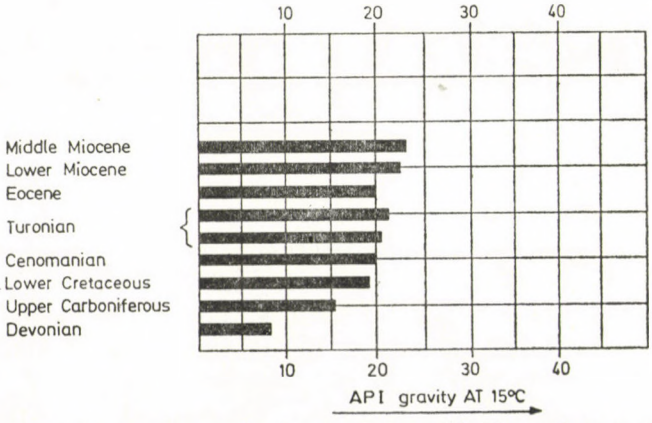

Fig. 7. Crude oil gravities and geologic time in Bakr oil field, south part; the Gulf of Suez region A. R. Egypt

oil field, which is of the typical lithologic type, and the direction of migration of oil are positive criteria for considering the Gharandal Group (Lower Miocene shales and marls) as the source rocks.

The same case of Um El-Yusr oil field can be considered for the oil-bearing Miocene rocks in Belayim on-shore and Bakr oil fields.

The oil-bearing Mesozoic rocks in Bakr, Hurgada, and Ras-Gharib oil fields, where the crude oils participated in the migration differentiation mechanism, also show the same phenomenon. However, one should be conservative in visualizing the source rocks for the pre-Miocene oil in these fields and similar concessions.

The Problem of the Source—Reservoir Relations

The production of oil and gas from the Gulf of Suez petroleum province comes as mentioned before from different iithostratigraphic horizons. Most of the production comes from the Miocene successions mainly the sandstones e.g. El-Morgan, July, Belayim, Um El-Yusr and other fields. However, other oil fields produce from reefal and crystalline limestones, e.g. Bakr and Ras-Gharib oil fields. The Eocene and Upper Cretaceous limestones and the Nubia Sandstones are oil-bearing and productive in some oil fields.

WEEKS (1952) considered that the source of the oil accumulations in the Gulf of Suez coastal region is within the Miocene sediments which were accumulated in a basin with favourable conditions for oil occurrences.

GHORAB (1960) stated that there have been two contradicting schools regarding the source of oil in the gulf of Suez region. One claims the Eocene—Upper Cretaceous as the source rocks; the other favouring the Miocene as the source rocks, in what is called the Globigerina marls of Early and Middle Miocene age.

HOLMQUEST (1966) stated that the source beds of Ras-Gharib oil field can be related readily to overlying beds, where an upper Cretaceous Limestone is truncated and overlapped by organic marl and shale.

METWALLI and BASHAT (1974) located the source rocks for the Miocene oils in Um El-Yusr oil field, Gulf of Suez region in the Gharandal Group (Lower Miocene shales and marls).

However, the oil migration and accumulation in the Gulf of Suez region seem to be controlled in most parts of the region by the evaporites (mainly salt). Oil occurs in the post South-Gharib Formation, i.e. in the Zeit Formation, in the on-shore fields only, e.g. Belayim on-shore oil field. In the off-shore fields the oil pay-zones and occurrences are restricted to pre South-Gharib Formation, i.e. pre-salt deposition and flowage, e.g. El-Morgan, July and similar structures. The South-Gharib Formation (salt and other evapo-

rites) is thicker in the major graben of the Gulf, i.e. in the off-shore area than on the fore-land sides of the Gulf. Also the lithology of the Zeit Formation changes to become sandy at the fore-land sides of the Gulf. Accordingly, the thinning of the salt of the South-Gharib Formation and the increase of the sandy facies promote and cause the accumulation of oil in the Zeit Formation but in a restricted dimension and minority of production. It is also of interest to mention that the Belayim evaporites in El-Morgan oil field become intercalated with sandstones and clastic sediments landwards, e.g. Belayim on-shore oil field. In this respect it might be cited that the change in facies of deposition might have promoted the migration and accumulation of oil from Belayim Formation to Zeit Formation.

METWALLI and ABD EL-HADY (1975) used the variability and successive characteristics of crude oil gravities as well as the successive gas reservoirs in the northern part of the Western Desert of Egypt as criteria and they classified the times of oil generation and accumulation in that part of Egypt into three main cycles: A lower cycle "I" for the Jurassic and older rocks; an intermediate cycle "II" for the Cretaceous oils; and an upper cycle "III" for the Tertiary oils. METWALLI and ABD EL-HADY (op. cit.) zones of oil generation and accumulation are supported by EL-SHAZLY et al. (1975), partly by using C^{13}/C^{12} for the distinction and limitation of the lower cycle "I" of Jurassic from the intermediate cycle "II" of Cretaceous age. However, EL-SHAZLY et al. (op. cit.) analysed the carbon isotopic data (C^{13}/C^{12}) from reservoir rocks (sandstones and limestones) ranging from Miocene to Cretaceous or older age and found that all possess the same isotopic values. Accordingly, EL-SHAZLY et al., are in favour of a single source for oil in the Gulf of Suez petroleum province mainly the Miocene rocks. However, they neglected the change in properties of oil that are expected following perculation through adsorptive media, specially the ratio of the stable carbon isotope (C^{12} and C^{13}).

SILVERMAN and EPSTEIN (1958) showed that aromatics have slightly higher C^{13}/C^{12} ratios than do the asphaltic hydrocarbons (paraffins and naphthenes) from equivalent boiling ranges. Thus, from the standpoints of selective adsorption on mineral grains the C^{13}/C^{12} ratio of petroleum should decrease slightly with increasing distance of migration (SILVERMAN, 1965).

Figure 8, shows that the crude oils produced from the different lithostratigraphic horizons in the Gulf of Suez oil fields have variable oil gravities both vertically and laterally. The vertical variations occur within the different successive pay-zones of the same or varying reservoir characteristics. The crude oils in the on-shore oil fields at the eastern and western coasts in the Gulf have their distinct oil gravities as compared with the off-shore oil fields, i.e. the general rule "the deeper the oil, the lighter the oil" is clear and represented by the off-shore oil fields. However, the separation migration differentiation of SILVERMAN (1965), or the migration differentiation mechanism

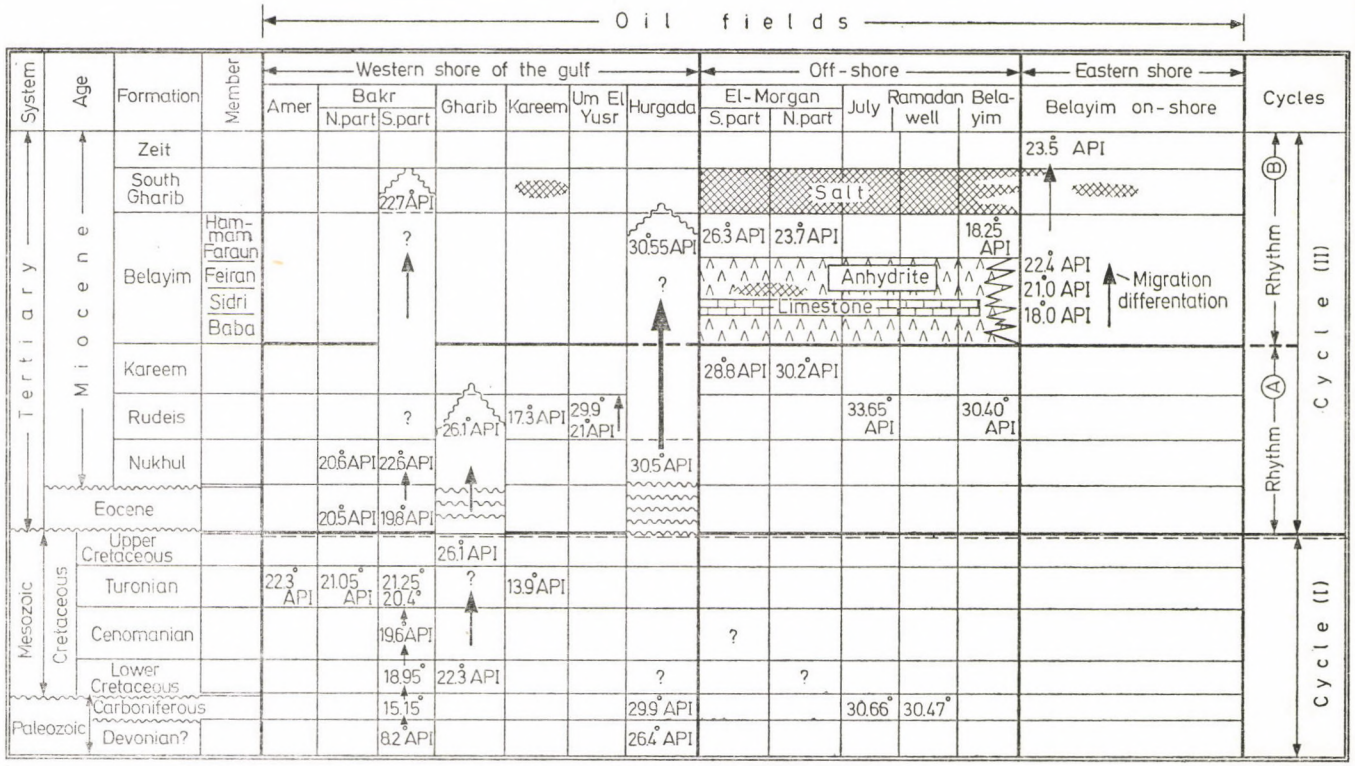

Fig. 8. **Proposed cycles of oil generation, migration and accumulation of some oil fields in the Gulf of Suez petroleum province, A. R. Egypt. (November, 1976), (By: M. Hamed Metwalli, G. Philip and El-Sayed A. A. Youssif)**

of METWALLI (1974) applies to the in the on-shore oil fields in the Gulf of Suez petroleum province.

The cited oil gravities (Fig. 8) can be seen and observed in a vertical successive geologic attitude with certain distinct deviations represented by arrows in the figure. These arrows might represent the direction of oil migration (phases of migration) within the geologic section in the Gulf of Suez petroleum province. Accordingly, the relationship between the different types of oil must be considered. The successive characteristics of oil gravities within the same and in different ages, would reflect more than one cycle of oil generation as well as migration and accumulation.

Accordingly, the following cycles are proposed in the Gulf of Suez petroleum province: A lower cycle "I" for the Mesozoic and older rocks; and an upper cycle "II" for the Tertiary crude oils. However, as can be seen from figure 8, cycle "II" might be a collective cycle or could represent more than one cycle or rhythms of the same cycle. This interesting problem created in the present work would be solved or at least clarified if additive drilling and geochemical data become available in future.

The problem of the definition and recognition of the source rocks of hydrocarbons in the Gulf of Suez petroleum province is at present the subject matter of current research by Metwalli and his co-workers. Eventually future deeper drilling in the Gulf of Suez and adjoining regions will clarify many of the concepts at present controversial. The present writers believe and are optimistic that large amounts of oil are present be struck in the thick sedimentary sequence underlying the Miocene rocks in the Gulf of Suez region based on their proposed separate cycle of oil generation, migration and accumulation for the Mesozoic and underlying rocks "Cycle I, Fig. 8".

Acknowledgements

The authors are grateful to M. T. SHAWKY Chairman and Managing Director of GUPCO who provided the oil samples of El-Morgan oil field and for A. A. HASSAN, former chief geologist of GUPCO for their indispensable help during the progress of this work. A special gratitude is to The Egyptian General Petroleum Corporation (EGPC) and for The Egyptian Central Agency for Public Mobilization and Statistics, Cairo, A. R., Egypt for permission of doing this study.

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О НЕФТЕНОСНОСТИ ПРОМЫСЛА ЭЛЬ-МОРГАН И ЦИКЛИЧНОСТИ ОБРАЗОВАНИЯ, МИГРАЦИИ И НАКОПЛЕНИИ НЕФТИ В НЕФТЕГАЗОНОСНОЙ ПРОВИНЦИИ СУЭЦКОГО ЗАЛИВА (А. Р. ЕГИПЕТ)

М. ГАМЕД МЕТВАЛЛИ—Г. ФИЛИП—ЭЛЬ-САЙЕС А. А. ЮССИФ

Резюме

Нефть, добываемая в пределах нефтегазоносной области Суэцкого залива, показывает изменение удельного веса по различным литостратиграфическим горизонтам от верхнего палеозоя вверх по стратиграфической колонке вплоть до миоцена. Некоторые нефтепромыслы обнаруживают постепенное изменение по вертикали от тяжелой нефти в низах разреза до относительно легкой в верхах, в то время как на других промыслах наблюдается как раз обратная тенденция уменьшения удельного веса вниз по разрезу. Подобно другим морским нефтепромыслам, промысел Эль-Морган отличается от наземных промыслов внутри той же самой нефтегазоносной области в отношении поведения удельного веса нефти. В результате исследований и интерпретации геохимических характеристик и изменчивости нефти по удельному весу можно выделить две значительные и выраженные категории нефти, а также два цикла формирования, миграции и аккумуляции нефти на терри- и акватории нефтегазоносной области Суэцкого залива.

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PALAEOLITUONELLA MAJZONI NOV. GEN. NOV. SP. (FORAMINIFERA) FROM A WETTERSTEIN REEF LIMESTONE IN NE HUNGARY*

By

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HUNGARIAN HYDROCARBON INSTITUTE, BUDAPEST

The *Palaeolituonella majzoni* nov. gen. nov. sp. is a new Foraminifera species which shows a frequent occurrence in the Upper Ladinian—Carnian, Norian reef limestone facies of the Alsóhegy Mount (NE Hungary).

The agglutinant Foraminifera species encountered frequently in the Upper Ladinian-Carnian, Norian Wetterstein reef limestone of Alsóhegy Mount (Northeast Hungary) is *Palaeolituonella majzoni* nov. gen. nov. sp. and belongs to the Family Ataxophragmiidae SCHWAGER.

The assemblage of microfossils accompanying it is characterized by a massive appearance of the Involutinidae and by a frequent occurrence of species belonging to the Family Miliolidae (Ophthalmidium, Galeanella) and Varistomatidae (Duostomina, Variostoma). This faunal assemblage is characteristic of the reef platform. In this zone the big, thick walled Foraminiferae are most abundant. The dominating species of this facies are *Aulotortus sinuosus* WEYNSCHENK, *Galeanella panticae* ZANINETTI et BRÖNNIMANN, *Palaeolituonella majzoni* BÉRCZI-MAKK, nov. gen. nov. sp.

The holotype of the Foraminifera *Palaeolituonella majzoni* nov. gen. nov. sp. was found in the same sample from which the new Sphinctozoa species *Verticillites triassicus* had been described by KOVÁCS, S. (1978).

The characteristic Foraminifera species of the Upper Ladinian-Carnian Norian Wetterstein reef facies in the Alsóhegy Mount (Fig. 1) are as follow:

Tolypammina gregaria WENDT
Gaudryina sp.
Palaeolituonella majzoni nov. gen. nov. sp.
Endothyra nov. sp.
Endothyranella cf. *robusta* SALAJ
Endothyranella cf. *pentacamerata* SALAJ
Agathammina sp.
Agathamminoides spiroloculiformis ORAVECZ—SCHEFFER
Ophthalmidium plectospirus ORAVECZ—SCHEFFER
Ophthalmidium exiguum KOEHN—ZANINETTI
Ophthalmidium sp.
Galeanella panticae ZANINETTI et BRÖNNIMANN
Frondicularia woodwardi HOWCHIN

* Manuscript received: 1979. 10. 19.

Fig. 1. Position of the Alsóhegy Mount in NE-Hungary, along the Hungarian—Czechoslovakian border

Mesodiscus eomesozoicus (OBERHAUSER)
Aulotortus sinuosus WEYNSCHENK
Aulotortus friedli (KRISTAN—TOLLMANN)
Duostomina alta KRISTAN—TOLLMANN
Duostomina biconvexa KRISTAN—TOLLMANN
Diplotremina sp.
Variostoma acutoangulata KRISTAN—TOLLMANN
Variostoma cf. *pralongense* KRISTAN—TOLLMANN
Variostoma sp.

Paleontological description

Phylum: **RHIZOPODA**
Classis: *Foraminifera*

In defining the taxonomic position the system of LOEBLICH, R. A. et TAPPAN, H. (1964) was used.

Family: **ATAXOPHRAGMIIDAE**
Subfamily: *Ataxophragmiinae* SCHWAGER, 1877

Genus: *Palaeolituonella* BÉRCZI-MAKK, nov. gen.

Derivatio nominis: the ancient form of the *Pseudolituonella* MARIE, 1954

Genotype: *Palaeolituonella majzoni* nov. sp.

Description: The shell is elongated, coniform and in its initial form it consists of 4 to 5 coiled chambers. Later on it turns to a linear form consisting of wide and low chambers with faint septum-rudimenta

hardly to observe even in thin sections. The one-layer wall is thick, agglutinated. The apertura is not known.

Remarks: It differs from the Pseudolituonella by the fact that the inner septa has hardly been developed, more exactly they are missing in almost every case. On the basis of the missing or underdeveloped inner-septa it can be considered as the ancient form of the Pseudolituonella.

Palaeolituonella majzoni BÉRCZI-MAKK, nov. sp.
Plate: I, Figs 1—8

Derivatio nominis: after the late dr. L. MAJZON, Hungarian micropaleontologist.

Locus typicus: Sample No. T-421(1975) from Alsóhegy Mount (Silica Nappe, NE Hungary).

Stratum typicum: Wetterstein reef limestone, Upper Ladinian-Carnian, Norian.

Holotype: In the micropaleontological collection of the Hungarian State Geological Institute.

Material: Several individuals in the following samples: 223/50, 4(1972)B, 7(1972)A, 8(1972)12, 9(1972), 9(1972)A, 11(1972)E, 19(1972)1, 35(1973)F, fig. 6. G-9(1974)A, G-23(1974), T-66(1975)A, T-66(1975)F, T-188(1975)2, T-211(1975), T-421(1975) = Holotype, T-477(1975), T-484(1975)A.

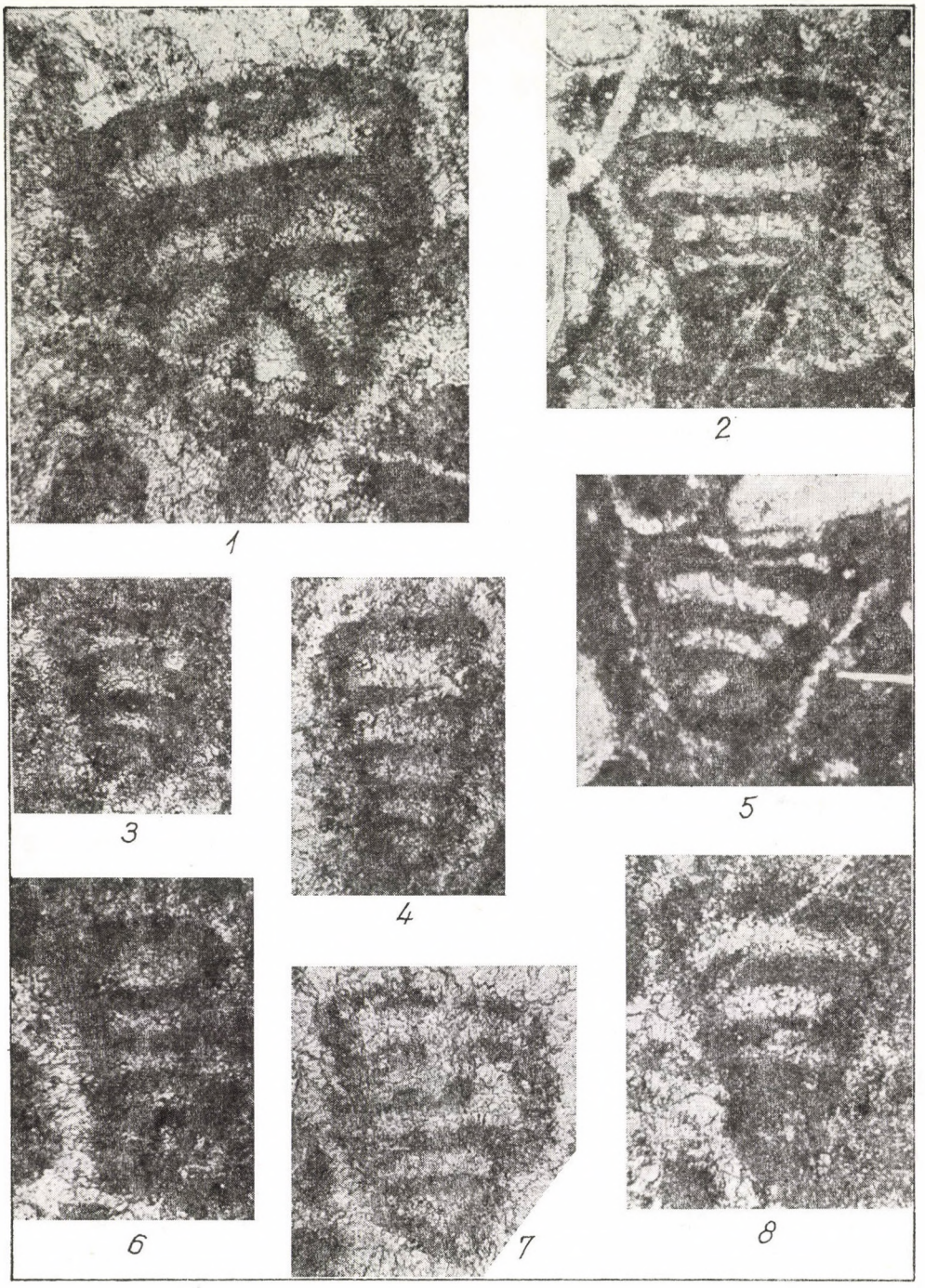
Synonyma:

1966/67. *Lituolida* — PANTIĆ, S. pl. 3, fig. 7.
1971. *Ammobaculites* sp. — HOHENEGGER, J. et H. LOBITZER, pl. 2, fig. 6.
1971/72. *Lituolidae* — PANTIĆ, S. pl. 12, fig. 4.
1978. "*Lituosepta*" sp. — SCHÄFER, P. et SENOWBARI-DARYAN, B. pl. 2, fig. 7.
1978. *Duotaxis* sp. — TRIFONOVA, EK. pl. 3, fig. 3.

Description: The shell is free, elongated forming a slightly flattened cone. In its initial stage it consists of 4—5 chambers coiled up in a trochospiral way. During its grow, the shell suddenly changes into a linear form. The chambers spread and form a low reversed truncated cone, which, however, grows during its development. Inside the chambers hardly perceptible rudiments of septa can be observed. The one layer wall is thick, agglutinated. The apertura is not known.

Dimensions of the test: The length of the shell 0.5—0.8 mm. The diameter of the coiled up initial section 0.26—0.33 mm. The thickness of the septa 0.04—0.08 mm. The thickness of the wall 0.04—0.06 mm.

Occurrence: It is very abundant in the Upper Ladinian-Carnian, Norian Wetterstein reef limestone of the Alsóhegy Mount (Silica Nappe, NE Hungary).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PALAEOLITUONELLA MAJZONI NOV. GEN. NOV. SP. — НОВАЯ НАХОДКА ФОРАМИНИФЕР ИЗ ВЕТТЕРШТЕЙНСКИХ РИФОГЕННЫХ ИЗВЕСТНЯКОВ СЕВЕРО-ВОСТОЧНОЙ ВЕНГРИИ

А. БЕРЦИ-МАКК

Резюме

Из верхнеладинских-карнийских рифогенных известняков на горе Альшохедь в северо-восточной части Венгрии описан новый вид фораминифер, часто встречающийся рассматриваемых отложениях.

Палеонтологическое описание

Phylum: **RHIZOPODA**

Classis: *Foraminifera*

При систематическом отнесении рассматриваемой формы была применена система Р. А. Лёблиха и Г. Таппана (1964).

Семейство: **ATAXOPHRAGMIIDAE**

Подсемейство: *Ataxophragmiinae* SCHWAGER, 1877.

Род: *Palaeolituonella* — *Palaeolituonella* BÉRCZI—MAKK, nov. gen.

Derivatio nominis: Древняя форма

Genotypus: *Palaeolituonella majzoni* nov. sp.

Plate I

Palaeolituonella majzoni nov. gen. nov. sp.

1. T-421 (1975) No. sample. M: 100×. Holotype
2. T-421 (1975) No. sample. M: 60×
3. 22 (1972) No. sample. M: 80×
4. T-66 (1975)F No. sample. M: 90×
5. 8 (1972) 12 No. sample. M: 70×
6. 9 (1972) No. sample. M: 90×
7. T-66 (1975)A No. sample. M: 90×
8. G-23 (1974) No. sample. M: 90×

Описание: Раковина вытянутой, конической формы, в ранней стадии завитая, состоящая из 4—5 камер. Впоследствии она становится однорядной с широкими и низкими камерами. Внутри камери имеются слабые зачатки столбиков, которые не всегда видны в шлифах. Стенка толстая, агглютинированная, однослойная. Устье неизвестно.

Примечание: От Pseudolituonella рассматриваемый род отличается тем, что внутренние столбики еще едва развиты, они почти полностью отсутствуют. Из-за недоразвитых или еще неразвитых внутренних столбиков его можно принимать за древнейшую форму ***Palaeolituonella*** **Bérczi—Makk, nov. gen.**

Фототаблица 1, фиг. 1—8

Derivatio nominis: по покойному микропалеонтологу д-ру Ласло Майзону

Locus typicus: Альшохедь (Силицкий покров, СВ Венгрия), образец Т-421 (1975)

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Голотип: в микропалеонтологической коллекции Венгерского Геологического института.

Материал: Ряд экземпляров в образцах 223/50, 4/1972/B, 7/1972/A, 8/1972/12, 9/1972, 9/1972/A, 11/1972/E, 19/1972/1, 35/1973/F, 8—9/1974/A, G-23/1974, T-66/1975/A, T-66/1975/F, T-188/1975/2, T-211/1975, T-211/1975, T-421/1975 = голотип, T-477(1975), T-484(1975).

Синонимика: 1966/67 *Lituolida* = Pantić, S. pl. 3, fig. 7
1971. *Ammobaculites* sp. = Hohenegger, J. et Lobitzer, H. Pl. 2, Fig. 6.

1971/72. *Lituolidae* — Pantić, S. pl. 2, fig. 4.
1978. "*Lituosepta*" sp. — Schäfer, P. et Senowbari-Daryan, B. pl. 2, fig. 7.
1978. *Duotaxis* sp. — Trifonova, Ek. pl. 3, fig. 3.

Описание: Раковина свободная, вытянутой, слегка уплощенной конической формы, В ранней стадии она состоит из 4—5 камер, завитых трохоспирально. В процессе роста раковина становится однорядной совме внезапно. Камеры растилаются. образуя перевернутый усеченный конус с высотой, увеличивающейся в процессе роста раковины. Внутри камер имеются зачатки столбиков, которые не всегда видимы. (Фототаблица I, фиг. 1). Стенка толстая, агглютинированная, однослойная. Устье неизвестно.

Размеры раковины: Длина раковины — 0,5—0,8 мм. Диаметр завитой ранней части — 0,26—0,33 мм. Толщина перегородок камер — 0,040—0,080 мм, толщина стенки раковины — 0,040—0,060 мм.

Распространение: Очень часто встречается в верхнеладинско-карнийских рифогенных веттерштейнских известняках горы Альшохедь (Силицкий покров, СВ Венгрия)

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THE ESTABLISHMENT OF OPTIMAL TIME SCALES AND THEIR USE

By

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In geology events may be exactly defined by the changes in the attributes of the rocks along the space. Geologists concludes from the existence of an order in space between certain geological bodies the existence of a relation of order in time between certain events. The correctness of the conclusion depends on the measure of correspondence between the applied type of spatial order and on the invariancy of the spatial order towards common geological transformations.

Several spatial and temporal relations between formalized biostratigraphic and chronostratigraphic units and their invariancy have been investigated in the general situation and when rock bodies are normal deposited.

Then optimal time scales and methods to construct those have been defined. Conditions which admit to solve problems are discussed. A method to estimate the relative age of a sample has been shown.

Methodology has been applied succesfully to Eocene sediments of Dorog Basin, Transdanubia, Hungary. Different fossil groups as nannofossils, Mollusca, Foraminifera and lithological data were together used to obtain a complex scale and diagnostic rules for sample age determination.

Introduction

In the past decade a remarkable worldwide proliferation of geological, paleontological, magnetical and geophysical data for geological events could be observed. The handling of data by computers promises the best way to synthetize elementary observations. Formal manipulations provide the possibility to set up time scales which are optimal in a certain sense. If once an exact time scale has been constructed, the position of new samples on the scale (determination of the age) can be estimated by algorithmic approach. Formalized stratigraphy introduced by the present author (DIENES I., 1974, 1977a, 1977b) is suitable to formulate and solve the above problems. The algorithmic and formal way is the single one, which may provide the correctness and verificability of the results.

The following discussion assumes that a suitable *physical time scale* (atomic, astronomic etc.) has been defined earlier and it shows backwards in the time. Any set of events ordered so, as their ages on the physical scale is called

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a time scale. We denote the *area of investigation* by T, the *points* of T by t, the *coordinates of the point* t in a normally orientated system by x, y and z. We denote the *age of origin of a point* t by $\tau(t)$. This is understood as a positive number with physical units, as year etc. The function $f_j(t)$ is defined so that its value is equal to 1 at least one specimen of the taxon j is present if the ε environment of the point t, and zero else. The value of the function $f_j(t(\tau))$ is equalling 1 if at least one living specimen was present in the moment τ and in the ε environment of the point $t(\tau)$. The function $t(\tau)$ describes the orbit of the point with present position $t(o)$.

In our studies *formal taxa* have been used. A specimen belongs to the formal taxon named by $\{\{A\}, \{B\}, \{C\}\}$, if the specimen has been identified in the document C by the author B as the taxon A. The taxon code contains information also for the conservation state and the precision of the identification.

Uncertainties in the definition of the taxa and in the recognition of fossil specimens are not discussed. In connection with our topics it seems to be necessary to assume only the possibility to define classes of fossil objects using restrictions for their attributes, and it is not denied. Erratic diagnoses only result in incorrect determination of the extent of stratigraphic units, and this effect easily can be studied by formal methods.

We define the moments:

$$\tilde{\tau}'_j = \inf_{\vartheta} \{\inf_{t \in T} \{\tau(t(\vartheta)) \mid \tilde{f}_j(t(\vartheta), = 1\}\} \tag{1}$$

$$\tilde{\tau}''_j = \sup_{\vartheta} \{ \sup_{t \in T} \{ \tau(t(\vartheta)) \mid \tilde{f}_j(t(\vartheta) = 1\}\} \tag{2}$$

the first and last and appearance of the taxon j and

$$\tau'_j = \inf_t \{\tau(t) \mid f_j(t) = 1\} , \tag{3}$$

$$\tau''_j = \sup_t \{\tau(t) \mid f_j(t) = 1\} \tag{4}$$

the age of the oldest and the youngest points, such, that contain in their ε environment at least one specimen of the taxon j.

1. *Optimal Biostratigraphic Time Scales*

In the following part the most important types of the biostratigraphic and chronostratigraphic units which have been used in the rigorous formulation of the problems are defined.

1.1. *Definition of special stratigraphic units*

$$T_{\tau}^{A\tau} = \{t \mid t \in T;\ \tau_j' \leq \tau(t) < \tau_j''\} \tag{5}$$

is the A type chronozone of the taxon j,

$$T_j^{B\tau} = \{t \mid t \in T;\ \tau(t) \in \Theta;\ \Theta = \{\tau(t) \mid t \in T_j^B\}\} \tag{6}$$

is the B type chronozone of the taxon j,

$$\overline{T}_j^{A\tau} = \{t \mid t \in T;\ t \notin T^{A\tau}\} \tag{7}$$

is the A type absence chronozone of the taxon j,

$$T_j^B = \{t \mid t_j \in T;\ f(t) = 1\} \tag{8}$$

is the B type taxon range biozone of the taxon j,

$$\begin{aligned} T_j^A = \{t \mid t \in T;\ & t_j', t_j'' \in T; f_j(t') = f_j(t'') = 1; \\ & x_j' = x_j'' = x;\ y_j' = y = y_j'';\ z_j' \leq z < z_j''\} \end{aligned} \tag{9}$$

is the A type taxon range biozone of the taxon j,

$$T_I^{BC} = \{t \mid t \in T;\ f_j(t) = 1;\ \textit{for any } j \in I\} \tag{10}$$

is the B type concurrent range biozone of the taxa j; j ∈ I,

$$\begin{aligned} T_I^{AC} = \{t \mid & z_j' \leq \tau < z_j'';\ f_j(t) = f_j''(t) = 1; \\ & t_j', t_j'', t \in T;\ x_j' = x = x_j'';\ y_j' = y = y_j';\ \textit{for any } j \in I\} \end{aligned} \tag{11}$$

is the A type concurrent range biozone of the taxa j ∈ I,

$$T_I^{AC\tau} = \{t \mid \tau_j' \leq \tau(t_j) < \tau_j'';\ t \in T;\ \text{for any } j \in I\} \tag{12}$$

is the A type concurrent range chronozone of the taxa j ∈ I,

$$T_I^{BA} = \{t \mid t \in T;\ f_j(t) = 1;\ \textit{for at least one } j \in I\} \tag{13}$$

is the B type assembly biozone of the taxa j ∈ I,

$$\begin{aligned} T_I^{AA} = \{t \mid & t \in T; t_j', t_j'' \in T;\ z_j' \leq z < z_j''; \\ & \textit{for at least one } j \in I;\ x_j' = x = x_j'';\ y_j' = y = y_j''\} \end{aligned} \tag{14}$$

is the A type assembly biozone of the taxa $j \in I$,

$$T_I^{AA} = \{t \mid t \in T;\ \tau_j' = \tau(t) < \tau_j''; \textit{ for at least one } j \in I\} \tag{15}$$

is the A type assembly chronozone of the taxa I, J, K,

$$T_{I,JK}^{O\tau,} = \{t \mid t \in T,\ \underbrace{\tau_i' \leq \tau(t) < \tau_i''; i \in I;}_{\text{positive part}}\ \underbrace{\tau(t) < \tau_j'; j \in J; \tau(t) \geq \tau_k''; k \in K}_{\text{negative part}}\} \tag{16}$$

of the definition

C type Oppel chronozone of the triplet of taxa I, J, K, ,*

* As it is usual, the relation "contains" is denoted by $\subset$, "common part" by $\cap$, "union" by $\cup$ and "element" by $\in$.

$$T_{I,J,K}^{C} = \underbrace{(\bigcup_{i \in I} T_1^A)}_{\text{positive part}} \cap \underbrace{(\bigcap_{j \in J} \overline{T}_j^A) \cap (\bigcap_{k \in K} \overline{T}_k^A)}_{\text{negative part}} \tag{17}$$

of the definition

is the C type Oppel biozone of the triplet of taxa I, J, K. Chronozones above are *apparent*. Chronozones defined by $\tilde{\tau}_j'$ or $\tilde{\tau}_j''$ are *intrinsic*. A summary of the definition of different stratigraphic units can be found in DIENES I. 1977c.

1.2. *The Relation "Containing" between Geological Bodies*

The *taxon j* is named *partially redeposited*, if $T_j^B \not\subset \tilde{T}_j^B$. *Taxon j* is named *partially redeposited into another age*, if $T^{B\tau}_j \not\subset \tilde{T}_j^{B\tau}$. Table 1 shows relation of "containing" being true between different stratigraphic units defined by the same taxon j.

It is evident, that

$$T_{I,J,K}^{O\tau} = (\bigcap_{i \in I} T_i^{A\tau}) \cap \overline{(CCT^{A\tau})} \cap \overline{(CCT^{A\tau})} \tag{18}$$

The geological body $\overline{T}$ is named *upper absence zone* of the geological body T',

$$\overline{T'} = \{t \mid t \notin T';\ t \in T;\ z \in pr_{12}\, T';\ z > \sup\{\hat{z} \mid \hat{x} = x;\ y = \hat{y};\ \hat{t} \in T'\}\} \cup$$
$$\cup \{t \mid t \notin T';\ t \in T;\ z \notin pr_{12} T'\} \tag{19}$$

The operation "projection onto the $x - y$ plane" is denoted by pr_{12}. The geological body T is named the *lower absence zone of the geological body T*.

$$\overline{T'} = \{t \mid t \in T',\ t \in T,\ z \in pr_{12} T', z < \inf\}\ \hat{z} \mid \hat{x} = x;\ \hat{y} = y;\ \hat{t} \in T'\}\} \cup$$
$$\cup \{t \mid t \notin T';\ t \in T;\ z \notin pr_{12}\, T'\} \tag{20}$$

Table 1

Truth of the relation "containing" between different geological bodies defined by the same taxon. If the geological body in the heading of the column contains the geological body in the heading of the row, then the respective item is one and zero else. Columns on the left for the general case, columns on the right if the geological bodies are not partially redeposited. Normal deposition is assumed. T is of normal deposition, when if $z > z''$, then $\tau(x, y, z) < \tau(x, y, z')$ for each $(x, y, z), (x, y, z') \in T$

This geol. body contains this body	T^A_j	$T^{A\tau}_j$	T^B_j	$T^{B\tau}_j$	$\tilde{T}^A_j$	$\tilde{T}^{A\tau}_j$	$\tilde{T}^B_j$	$\tilde{T}^{B\tau}_j$
T^A_j	1 | 1	1 | 1	0 | 0	0 | 0	0 | 1	0 | 1	0 | 0	0 | 0
$T^{A\tau}_j$	0 | 0	1 | 1	0 | 0	0 | 0	0 | 0	0 | 1	0 | 0	0 | 0
T^B_j	1 | 1	0 | 1	1 | 1	1 | 1	0 | 1	0 | 1	0 | 1	0 | 1
$T^{B\tau}_j$	0 | 0	1 | 1	0 | 0	1 | 1	0 | 0	0 | 1	0 | 0	0 | 1
$\tilde{T}^A_j$	0 | 0	0 | 0	0 | 0	0 | 0	1 | 1	1 | 1	0 | 0	0 | 0
$\tilde{T}^{A\tau}_j$	0 | 0	0 | 0	0 | 0	0 | 0	0 | 0	1 | 1	0 | 0	0 | 0
$\tilde{T}^B_j$	0 | 0	0 | 0	0 | 0	0 | 0	1 | 1	0 | 1	1 | 1	1 | 1
$\tilde{T}^{B\tau}_j$	0 | 0	0 | 0	0 | 0	0 | 0	0 | 0	1 | 1	0 | 0	1 | 1

The geological body CCT_j is named the *vertical convex cover* of the geological bodies T_j, $j \in J$, if

$$CCT_j = \{t \mid \exists j_1, j_2, t_{j_1}, t_{j_2};\ j_1, j_2 \in J;\ t, t_{j_1}, t_{j_2} \in T; f_{j_1}(t_{j_1}) = f_{j_2}(t_{j_2}) = 1;$$
$$x_{j_1} = x_{j_2} = x;\ y_{j_1} = y_{j_2} = y;\ z_{j_1} \leq z < z_{j_2}\} \tag{21}$$

The following relations are of essential importance in the use of formalized stratigraphic units and are true, if partial redeposition is excluded:

$$T^{O\tau}_{I,J,K} \subset \bigcap_X \left(\left(\bigcap_{j \in J'} CCT^B_j\right) \cap \left(\bigcap_{k \in K'} CCT^A_k\right)\right) \tag{22}$$

$$T^{O\tau}_{I,J,K} \supset \bigcup_{X'} \left(\bigcap_{i \in I'} T^A_i\right) \tag{23}$$

X contains arbitrary combinations $\{J', K'\}$ with p_j and p_k indices; $p_j = 1, \ldots, m_j$, $p_k = 1, \ldots, m_k$. X' contains arbitrary combinations I' with p_i indices, $p_i = 1, \ldots, m_i$ if $I' \cap I_s = \emptyset$ for the positive part I_s of any other investigated chronostratigraphic unit $T^{O\tau}_s$, $J' \subset J$, $K' \subset K$, $I' \subset I$.

Table 2

Code numbers of respective definitions of time order can be obtained adding 5 to each number in the Table. Summary of different precedence relations between two geological bodies T_l and T_i. T_j is above T_l in sense of the item in r-th column and s-th row, iff. quantity in the head of the column r is greater, than the quantity in the head of the rows. The number in each item denotes the code number of the definition in the text. (After I. DIENES, *1977)*

T_j is above	$t \in K \cap T, T^k$ common part of T and a z direction lines z_k														
				T^k arbitrary						if T^k has common part with T_l and T_i					
T_i in sense:	$K = T$			$K = \forall T^k$			$K = \exists T^k$			$K = \forall T^k$			$K = \exists T^k$		
if $z_j > z_i$	sup $\{z\}$	inf $\{z\}$	mean $\{z\}$	sup $\{z\}$	inf $\{z\}$	mean $\{z\}$	sup $\{z\}$	inf $\{z\}$	mean $\{z\}$	sup $\{z\}$	inf $\{z\}$	mean $\{z\}$	sup $\{z\}$	inf $\{z\}$	mean $\{z\}$
sup $\{z'\}$	3	2	—	13	12	—	23	22	—	33	32	—	43	42	—
inf $\{z'\}$	1	4	—	11	14	—	21	24	—	31	34	—	41	44	—
mean $\{z'\}$	—	—	5	—	—	15	—	—	25	—	—	35	—	—	45

1.3. *Relations of Spatial Order between Geological Bodies*

The Table 3 shows the interrelations between different relations of spatial order between geological bodies.

Table 3

How follows the spatial orderedness of two geological bodies in sense r from that in sense p. Left column in general case, right column for a vertical line form T. If T_j „p" T_i, then T_i „r" T_j

r \ p	1	2	3	12	32	
1	1	1	1	1	0	1
2	0	1	0	1	0	1
3	0	1	1	1	0	1
12	0	1	0	1	1	1
32	0	1	0	1	1	1

The Table 2 contains the short form of the definitions of the relations of spatial order denoted by "1"—"45".

1.5. *Invariancy of the Spatial Order towards Common Geological Transformations*

Let T, $\{T_i\}$, $T_i \subset T$, T', $\{i\}$, $T'_i \subset T'$ be arbitrary, fixed. Assume that $T' \subset T$, $T'_i \subset T_i$ and if $t \notin T'_i$ and $t \in T_i$ then for any $t_x \in T_i$, $z^x > z$ $y^z = y$, $x^z = x$, should be true $t^x \notin T'_i$. Then we call T'_i, $\{T''_i\}$ *to be deducible from* T, $\{T_i\}$ *by an ideal denudation.*

Let T, $\{T_i\}$, T', T^x be fixed such that $T_i \subset T$, $T^x \subset T$, T'"32"T^x, $T|T^x$"32"T^x. Let $T'_i = T' \cup (T_i \cap T^x)$ if $T_i \cap (T|T^x) \neq \emptyset$ and $T'_i = T_i$ else. Then we name T', $\{T'_i\}$ *to be deducible from* T, $\{T_i\}$ *by an ideal redeposition.*

Let T, $\{T_i\}$, T', $\{T'_i\}$, $T_i \subset T$, $T'_i \subset T'$ be arbitrary, fixed. We assume, that T has been dissected by the plane S into two disjoint, not empty parts T^x and $T|T^x$, and T' may be generated from T by moving of T^x, T^x"32"$T|T^x$ along the plane S in direction ν. If If ν_3 is negative, then we call T', $\{T'_i\}$ *to be deducible from* T, $\{T_i\}$ *by an ideal normal faulting.*

The definitions above are of formal and static character. They define relations between sets of real number triplets.

We name the *relation "p" invariant towards ideal denudation, redeposition, normal faulting*, if for any arbitrary T, $\{T_i\}$, T_i "p" T_j follows T'_i "p" T'_j if

Table 4

Attributes of different relations of precedence between geological bodies. For relations 6 — 10,

Attribute \ Relation of order		1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Antisymmetricity		0	1	1	1	1	0	1	1	1	1	0	1	1	1	1	0
Transitivity		0	1	1	1	1	0	1	1	1	1	0	1	1	1	1	0
Irreflexivity		0	1	1	1	1	0	1	1	1	1	0	1	1	1	1	0
Ascertainability		0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Precludibility		0	1	0	0	0	0	1	0	0	0	1	1	1	1	1	1
Invariancy	towards denudation	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	towards redeposition	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	towards normal faulting	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	towards reconnaissance	0	1	0	0	0	0	1	0	0	0	0	0	0	0	0	0

T', $\{T_i'\}$ is deducible from T, $\{T_i\}$ by an ideal denudation, redeposition, normal faulting. We name the *relation "p" ascertainable with knowledge of* $T^1 \subset T$, if there exists at least one r, $r \in [1, 45]$ so, that T_i "p" T_j follows from $T^1 \cap \cap T_i$ "r" $T^1 \cap T_j$ for any T_i, $T_j \subset T$.

We name the *relation "p" precludible with knowledge of* $T^1 \subset T$, if there exists at least one r, $r \in [1, 45]$ so, that $(T_i''\, p''\, T_j)$ follows from $T^1 \cap T_i$ "r" $T^1 \cap T_j$ for each T_i and T_j.

Viewing to the geological praxis, the relations, which are precludible or ascertainable by boreholes or superficial outcrops bear the greatest importance.

Geological cognition is getting downwards in the space with time. In a geological column, if any, only the top appearance of a feature may be recorded certainly, i.e. the upper boundary of the geological bodies. Therefore only the relations defined by the upper boundaries may be stated certainly, if auxiliary information is not available.

Let T, $\{T_i\}$, T', $\{T_i'\}$ be fixed, such, that $T' \subset T$, $T_i \subset T$, $T_i' := T_i \cap T'$. The *relation* "p" is named *invariant towards any reconnaissance* which is *getting on downwards* if T_i' "p" T_j' follows from T_i "p" T_j, when T' "32" $T|T'$.

Table 4 shows how invariant the most important relations of order are towards the transformations defined here.

18 — 20, 26 — 30, 38 — 40, assuming normal deposition before transformation

17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40	41	42	43	44
1	1	1	1	0	0	0	0	0	0	0	0	0	0	0	1	1	1	1	0	1	1	1	1	0	0	0	0
1	1	1	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
1	1	1	1	0	1	1	1	1	0	1	1	1	1	0	1	1	1	1	0	1	1	1	1	0	1	1	1
0	0	0	0	1	1	1	1	1	1	1	1	1	1	0	0	0	0	0	0	0	0	0	0	1	1	1	1
1	1	1	1	0	0	0	0	0	0	0	0	0	0	1	1	1	1	1	1	1	1	1	1	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0	1	0	0	1	0	1	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	0	0	0	1	1	0	0	0	0	0	0

The mentioned transformations also preserve the normal deposition.

Assume that $\widehat{T}$, $\{\widehat{T}_i^A\}$ have been formed from T, $\{T_i^A\}$ by an ideal denudation, redeposition or normal faulting. Then — following from the Table 4 — if $\widehat{T}_i^A$ "32" $\widehat{T}_j^A$, then also T_i^A "32" T_j^A comes true.

1.5. *Problem 1 of the Relative Biostratigraphic Geochronology*

Using definitions in DIENES I., 1977a and in the previous text the most essential problem of the relative geochronology (problem 1) may be formulated as follows.

Problem 1. There exists a set of outcrops and boreholes $\{T^{(x_i, y_i)}\}_{i \in I}$. Those are sampled in the points $\{t_{ik}\}_{i \in I, k \in K_i}$, and the presence or absence of the taxa $j = 1, \ldots, n$ have been recorded, i.e. the value of the functions f_j has been determined. Then we should like to find a time scale, which consists of moments of the type: τ_j', τ_j'', $\tilde{\tau}_j'$, $\tilde{\tau}_j''$. The time scale should be optimal, it should contain a maximal number of events.

Problem 1 is indefinite, and further conditions are necessary to obtain a solution. Conditions and the solution are in the following theorem.

Theorem 1. Let $T, f_k, \tilde{f}_k, \tau(t)$, $k = 1, \ldots, n$ be fixed.

Let T be a *geological body of normal deposition* that is let the age τ be a monotonous decreasing function of vertical coordinate z. Assume that:

$$\sup_t \{\tau(t) \mid (x, y) \in pr_{12} T_i^A \cap pr_{12} T_j^A\} \geq \sup_{t_j} \{\tau(t) \mid (x, y) \notin pr_{12} T_i^A \cap pr_{12} T_j^A\} \quad (24)$$

$$\inf_t \{\tau(t) \mid (x, y) \in pr_{12} T_i^A \cap pr_{12} T_i^A\} \leq \inf_t \{\tau(t) \mid (x, y) \notin pr_{12} T_j^A \cap pr_{12} T_i^A\} \quad (25)$$

Then for any (i, j), if T_j^A "32" T_i^A, then $\tau_j'' \leq \tau_i''$ and $\tau_j' \leq \tau_i'$.

Proof:

It is evident that for any (x, y), such, that $(x, y) \in pr_{12} T_i^A \cap pr_{12} T_j^A$ the relation $z > z'$ is true for any $t \in T^{(x_p, y_p)} \cap T_i^A$, $t' \in T^{(x_p, y_p)}, \cap T_j^A$.

We fix such a pair (t, t'). Then $\tau(t) < \tau(t')$ is true, since T is of normal deposition. So it follows:

$$\inf \{\inf \{\tau(x, y, z) \mid z \in pr_3 T^{(y,x)} \cap T_j^A\} \mid (x, y) \in pr_{12} T_j^A\} < \tau(t') \quad (26)$$

and so:

$$\inf \{\inf \{\tau(x, y, z) \mid z \in pr_3 T^{(x,y)} \cap T_j^A\} \mid (x, y) \in pr_{12} T_j^A\} < \inf \{\tau(t') \mid T_i^A \ni t'\} \quad (27)$$

Similarly it is true

$$\sup \{\tau(t)\} < \sup \{\tau(t')\} \quad (28)$$

Conditions (24)—(25) provide that relations (27) and (28) are true, when $x, y) \notin pr_{12} T_i^A \cap pr_{12} T_j^A$, too.

q.e.d.

Corollary 1. Let us denote

$$T_{jL}^A := \{(x, y, z) \mid (x, y, z) \in T_j^A,\ z \leq z' \text{ for any } (x, y, z') \in T_j^A \cap T^{(x,y)}\} \quad (29)$$

$$T_{jU}^A := \{(x, y, z) \mid (x, y, z) \in T_j^A,\ z \geq z' \text{ for any } (x, y, z') \in T_j^A \cap T^{0x,y)}\} \quad (30)$$

the lower and upper boundaries of the geological body T_j^A. Assume the conditions of the Theorem 1.

Then it is evident that:

$$\sup \{\tau(t) \mid t \in T_{kU}^A\} = \sup \{\tau(t) \mid t \in T_k^A\} \quad (31)$$

$$\inf \{\tau(t) \mid t \in T_{kL}^A\} = \inf \{\tau(t) \mid t \in T_k^A\} \quad (32)$$

Therefore it is true that

$$\tau'_j < \tau'_i \qquad \text{follows from } T^A_{jL} \text{ "32" } T^A_{iL} \tag{33}$$

$$\tau'_j < \tau'_i \text{ and } \tau''_j < \tau''_i \text{ follow from } T^A_{jL} \text{ "32" } T^A_{iU} \tag{34}$$

$$\tau''_j < \tau''_i \qquad \text{follows from } T^A_{jU} \text{ "32" } T^A_{iU} \tag{35}$$

$$\tau''_j < \tau'_i \qquad \text{follows from } T^A_{iU} \text{ "32" } T^A_{jL} \tag{36}$$

This theorem shows the type of spatial order that may be applied, when the time order of the respective events is to be studied. One can also easily show that

$$\tau''_j > \tau'_i \text{ follows from } T_{jM} \text{ "32" } T_{iN},\ M, N = \{L, U\} \tag{37}$$

1.6. *Solution of Problem 1*

A precedence matrix $\{\{a_{ij}\}\}$ which has been made applying relations (33), (34), (35), (36), (37) defines a network. $a_{ij} = 1$ if event j precedes event i.

Then by an algorithm for finding a maximum transitive path, a maximum ordered set of events, i.e. an approximate time scale may be selected by the algorithms of Kovács L. B., Dienes I., 1977. Each event of the scale is known explicitely. The scoring time scale is optimal in sense of the maximality of the number of events.

Another time scale may be defined applying the same set of events and defining the following relation of order between them: An event is called first which is not followed by other events and its serial number is minimal. An event is called second, if: either not followed by other events and its serial number is minimal excepting the first event, or if such does not exist, an event which is followed by one event and its serial number is minimal. Analogously may be defined event m, $2 < m \leq N$. This time scale contains N events, although the events of the scale are not known explicitly. Only an upper and lower estimation may be made for the position $k(i)$ of each event i on this scale by summing:

$$\sum_{j=1}^{N} a_{ij} \leq k(i) \leq N - \sum_{i=1}^{N} a_{ji} \tag{38}$$

However, the precedence matrix of the geological bodies T^A_{Pi}, $P \in \{U, L\}$ is not known, only its estimation by relations of type $T_{jP} \cap (\cup T^{x,y})$ "32" $T_{iQ} \cap \cap (\cup T^{x,y})$. Therefore only relation (37) may be used without limitations when temporal order is to be determined, and relations (33)—(36) only limited. The theorem guarantees the asymptotic undistortedness of the matrix made by application of the relations of type (33)—(37).

Problem also may be reformulated and then a whole family of problems of type "problem 1" may be viewed instead. We fix a set of areas $\{T_\alpha\}$, $\alpha \in \alpha$, the boreholes and outcrops of the areas $\{\{T^{(x,y)}_\alpha\}\}$, the geological bodies to be

investigated on the areas $\{\{T_{i\alpha}\}\}i \in I_\alpha \cup J_\alpha \cup K_\alpha$. Let us denote probability distributions of type $\lambda((x, y) \mid T_\alpha^{(x,y)} \cap T_{i\alpha}$ "32" $T_\alpha^{(x,y)} \cap T_{j\alpha y})$ by $\{\{p_{ij\alpha}\}\}, i, j = 1, \ldots, N$. Let P_α be a probability measure on the space α and N_α be a random variable on the space (α, P_α). Then we are interested in the search of a sequence of the indices $k_{1\alpha}, \ldots, k_{N_\alpha \alpha}$ such, that either

a) $P_\alpha(T_{k_j \alpha}$ "32" $T_{k_n \alpha}$ for at least one $j > n \mid$

$$\mid \bigcup_{(x,y)} (T_\alpha^{(x,y)} \cap T_{k_j \alpha})\text{ "32" } \bigcup_{(x,y)} (T_\alpha^{(x,y)} \cap T_{k_n \alpha})\ 1 \geq n > j \geq N_\alpha) = \max, \quad (39)$$

$$MN_\alpha > \text{const} \quad (40)$$

or

b) $P_\alpha > \text{const}, \qquad MN_\alpha = \max \quad (41)$

M is the expected-value-operator. Then this problem may be solved by different methods.

Part II

2. *Application of Optimal Time Scales*

If once an optimal time scale has been defined, it can be applied in the solution of a number of related problems. Two of them will be discussed below. DIENES I., 1978 discusses a third problem, that of automatic plotting of temporal range charts.

2.1. *The smallest biostratigraphic unit which contains and the greatest biostratigraphic unit which is contained in a chronostratigraphic unit, defined by the moments of the time scale*

The smallest chronostratigraphic units which can be defined by the moments of a time scale are those which may be defined by two neighbouring moments. As far as the chronostratigraphic units defined by a time scale can not be contoured explicitly, geologists are interested in the construction of contourable biostratigraphic units, such, which are indicated in the title of the present paragraph.

These geological bodies may be viewed as the best lower and upper estimations of the chronostratigraphic units of the discussed type.

If a constructed time scale is factually a scale, than one can find an *equivalent C type Oppel chronozone* for each chronostratigraphic unit definable by the moments of the scale. Two definitions are called equivalent if they both define the same geological body. Assume, that *maximum definitions* are

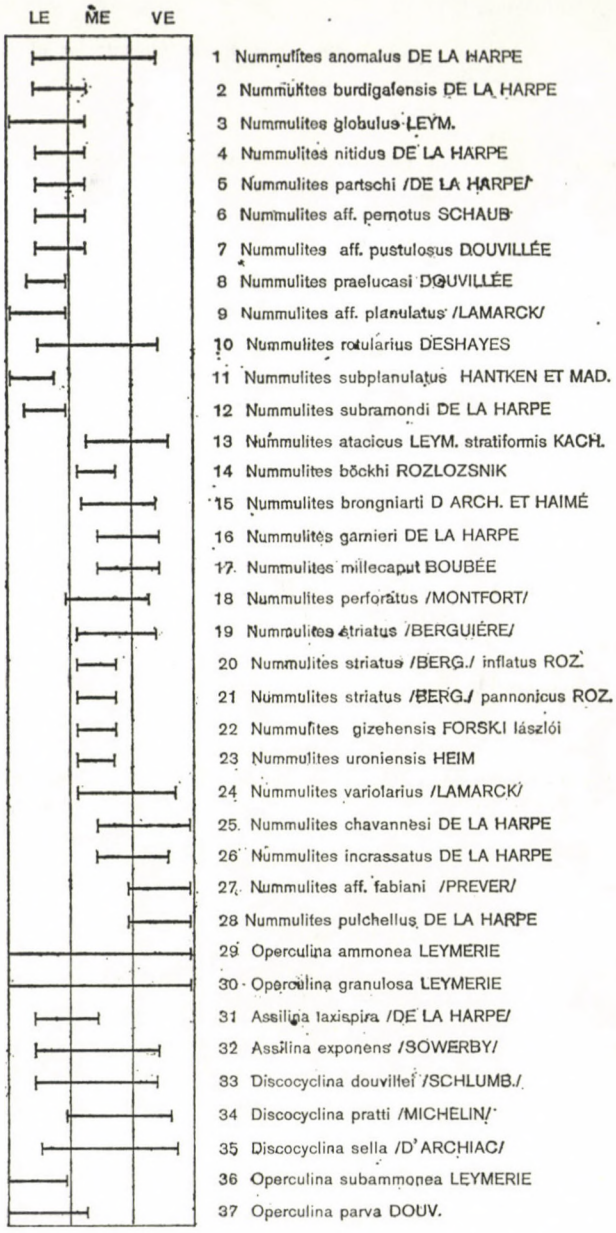

Fig. 1. **Haemera table of some important Eocene Foraminifera. M. JÁMBOR-KNESS 1973 Dorog Basin. Transdanubia, Hungary**

Table 5

Definition of a maximum number of chronostratigraphic

m	j_p	k_p	l_p
1	3, 9, 11, 29, 30, 36, 37	1, 2, 4, 5, 6, 7, 8, 10, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 25, 26, 27, 28, 31, 32, 33, 34, 35	
2	3, 8, 9, 11, 12, 29, 30, 36, 37	1, 2, 4, 5, 6, 7, 10, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 25, 26, 27, 28, 31, 32, 33, 34, 35	
3	1, 2, 4, 5, 6, 7, 3, 8, 9, 10, 12, 29, 30, 31, 32, 33, 35, 36, 37, 11	13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 25, 26, 27, 28, 34	
4	1, 2, 4, 5, 6, 7, 3, 8, 9, 10, 12, 29, 30, 31, 32, 33, 35, 36, 37	13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 25, 26, 27, 28, 34	11
5	1, 2, 3, 4, 5, 6, 7, 10, 18, 28, 29, 30, 31, 32, 33, 34, 35, 37	13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 25, 26, 27	8, 9, 11, 12
6	1, 2, 3, 4, 5, 6, 7, 10, 14, 18, 19, 20, 21, 22, 23, 29, 30, 31, 32, 33, 34, 35, 37	13, 15, 16, 17, 25, 26, 27, 28	8, 9, 11, 12, 36
7	1, 13, 14, 15, 18, 19, 20, 21, 22, 23, 24, 29, 30, 31, 32, 33, 34, 35, 10	16, 17, 25, 26, 27, 28	2, 3, 4, 5, 6, 7, 8, 9, 12, 36, 11

applied, that is each taxon is mentioned in each definition, and in the most powerful possible restrictions. In respect to the Dorog Basin, Transdanubia, Hungary M. JÁMBOR-KNESS, 1973 published the Fig. 1 where the haemera (temporal range) of the most important Nummulites have been plotted. Based on Fig. 1, 15 events and 14 different chrons and respective chronostratigraphic units can be defined. The equivalent maximum C type Oppel chronozones are the following: Fig. 1. Table 5.

Definitions above for each m are correct, independently from the correctness of the raw range chart, applied as a basis at its setting up. However, if the haemera chart were not correct, than the defined chronostratigraphic units may not fill in T, or the different chronostratigraphic units may have common points.

Two C type Oppel zones are named discernible, if deleting common taxa from the positive part of their definition, both positive-member-rests are not empty. A *C type Oppel zone T_1 is not discernible from the C type Oppel zone T_2,*

units $T^{0\tau}$ based on the haemera table by M. JÁMBOR-KNESS, *1973*

m	j_p	k_p	I_p
8	1, 10, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 25, 26, 29, 30, 32, 33, 34, 35	27, 28	2, 3, 4, 5, 6, 7, 8, 9, 11, 12
9	1, 13, 10, 16, 17, 18, 19, 24, 25, 26, 29, 30, 32, 33, 34, 35, 15	27, 28	2, 3, 4, 5, 6, 7, 8, 9, 11, 12, 14, 20, 21, 22, 23, 31, 36, 37
10	1, 10, 13, 15, 16, 17, 18, 19, 24, 25, 26, 27, 28, 29, 30, 32, 33, 34, 35		2, 3, 4, 5, 6, 7, 8, 9, 11, 12, 14, 20, 21, 22, 23, 31, 36, 37
11	1, 10, 13, 15, 16, 17, 19, 24, 25, 26, 27, 28, 29, 30, 32, 33, 34, 35		2, 3, 4, 5, 6, 7, 8, 9, 11, 12, 14, 18, 20, 21, 22, 23, 31, 36, 37
12	13, 24, 25, 26, 27, 28, 29, 30, 34		1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11, 12, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 31, 32, 33
13	25, 27, 28, 29, 30, 34		1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 26, 31, 32, 33, 34, 35, 36, 37
14	25, 27, 28, 29, 30		1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24, 26, 31, 32, 33, 34, 35, 36, 37

if each taxon in the positive member of the definition of T_1 is present also in that of T_2. *A stratigraphic unit* consisting of C type Oppel zones is named *discernible*, if all zero order component units are pair-by-pair discernible.

Chronostratigraphic units T_i^- defined by a time scale form a chronostratigraphic unit of the first order, B. If it is not discernible, then there occur chronostratigraphic units in B, that can not contain explicitly contourable biostratigraphic units, $T_i \in B$ such, which are certainly contained in T_i.

In regard of the Table 1 the chronostratigraphic units T_1, T_2 (despite of the possible errors in the range chart) are not discernible, T_1 and T_7 are discernible and T_1 is not discernible from T_2, but T_2 is discernible from T_1. The equations (22) and (23) define the largest biostratigraphic unit, which is certainly a part of a chronostratigraphic unit Tss, and the smallest biostratigraphic unit which certainly contains Tss. The same equations hold for $\tilde{T}^{0\tau}$ when taxa are not partially redeposited and T is of normal deposition.

2.2. *Estimation of the Relative Age of a Sample*

We assume that a precedence matrix is available, and a time scale has been defined. The taxa $i \in I'$ have been found in a sample around the point $t \in T$, and the moments of the scale τ_i, τ_{i+m} are searched, which satisfy $\tau_i \leq \tau(t) < \tau_{i+m}$ but $\forall\, \tau_k < \tau_i$, $\tau(t) \ngeq \tau_k$ and $\forall\, \tau_{i+m}$, $\tau(t) \nless \tau_k$. This is the dual problem of the problem which has been discussed in the paragraph 2.1. The smallest chronostratigraphic unit $\widehat{T}$ which contains a biostratigraphic unit T is to be found. This problem will be called Problem 2 of the relative biostratigraphic chronostratigraphy.

If no further observation is available about the presence of fossils in other samples of the same vertical, then the smallest zero order biostratigraphic unit $\hat{\hat{T}}$ which contains a sample, where the taxa $i \in I$ have been identified, and the lack of the taxa $j \in J$ and $k \in K$ has been established is the following:

$$\hat{\hat{T}} = (\bigcap_{i \in I'} T_i^B) \cap \overline{(\bigcup_{j \in J'} T_j^B)} \cap \overline{(\bigcup_{k \in K'} T_k^B)} \tag{42}$$

If the presence of the taxa $k \in K'$ under the sample and the presence of the taxa $j \in J'$ above the sample and the presence of the taxa $l \in L'$ under and above the sample have been observed, then also the smaller geological body

$$\hat{\hat{T}} = (\bigcap_{i \in I_1'} T_i^B) \cap (\bigcap_{l \in I_2'} T_l^A) \cap (\bigcap_{j \in K'} \overline{\overline{T_k^A}}) \cap (\bigcap_{j \in J'} \overline{\overline{T_j^A}}) \cap (\bigcap_{p \in P'} \overline{T_p^B}) \tag{43}$$

contains that sample. The taxa $p \in P'$ have not been identified at all. Furthermore we define C type Oppel chronozones, $\hat{\hat{T}}$ which certainly contain the biostratigraphic units $\hat{\hat{T}}$ defined above. So we obtain chronostratigraphic units, which certainly contain that sample. If no further observation is available about the presence of fossil taxa along the same vertical, then the following relation may be used for the age estimation:

$$\hat{\hat{T}} \subset \bigcap_{s} T_s^{O\tau}, \; T_s^{O\tau} = (\bigcap_{i \in I_s} T_i^{A\tau}) \cap \overline{(\bigcup_{i \in J_s} CCT_i^{A\tau})} \cap \overline{(\bigcup_{i \in K_s} CCT_i^{A\tau})} \tag{44), (45}$$

$$I' \cap (K_s \cup J_s) = \emptyset\,, \; K_s \subset K', \; J_s \subset J', \tag{46), (47), (48}$$

If further observations are available, then the following relations may be useful:

$$\hat{\hat{T}} \subset \bigcap_{s} T_s^{O\tau}, \text{ where over} \tag{49}$$

$$(I_1' \cup I_2') \cap (K_s \cup J_s) = \emptyset\,, \tag{50}$$

$$K' \cap J = \emptyset\,, \tag{51}$$

$$K_s \cap J' = \emptyset\,, \tag{52}$$

3. *Complex Scales*

The truth of the previous statements is not restricted to biostratigraphic units. If the analogous formal conditions are fulfilled, then any other (lithostratigraphic or mixed) unit is appropriate for setting up a complex time scale, and also the rules for the estimation of the relative age on that scale are the same.

Conclusions

The main problems of the relative geochronology, namely the establishment of optimal time scales using elementary observations and the estimation of the relative age of a sample in regard of the scale may be formulated as formal problems, and may be solved by logically correct steps accepting rational geologic suppositions formulated in a formal way.

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СОЗДАНИЕ ОПТИМАЛЬНЫХ ХРОНОЛОГИЧЕСКИХ ШКАЛ И ИХ ПРИМЕНЕНИЕ

И. ДИЕНЕШ

Резюме

События в геологоческой истории могут быть определены изменениями характеристик город в пространстве. Из существования порядка в пространстве между определенными геологическими телами можно сделать вывод о соотношении между определенными событиями в отношении порядка их времени. Правильность и временна вывода зависит от степени соответствия между примененными типами пространственного порядка и от инвариянции пространственного порядка по отношению к общим геологическим преобразованиям.

Исследовались некоторые пространственно-временные связи между формализованными био- и хроностратиграфическими единицами и их инвариянцией применительно к всеобщей ситуации и к случаю нормально отложившихся тел горных пород.

Определялись оптимальные хронологические шкалы и методы их построения. Рассматриваются условия, позволяющие решение проблем. Показан метод подсчета относительного возраста образца.

Рассматриваемая методика была с успехом применена для эоценовых отложений Дорогского бассейна (Задунайский край, ВНР). Для создания комплексной шкалы и правил диагностики для определения возраста образцов совместно применялись различные группы ископаемых организмов, например, нанноформ, моллюсков, фораминифер и литологические данные.

MAGYAR TUDOMÁNYOS AKADÉMIA KÖNYVTÁRA

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