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Preface to the Hungarian–Spanish Intergovernmental S & T Cooperation Programme: Project No. 21. Paleokarst and Raw Materials

Introduction

One of the biggest challenges facing our society is the supply of drinking water. The posession of drinking water has become one of the major concerns in many parts of the world. Historically it has been the source of numerous national and international conflicts and future disputes over water supplies cannot be excluded.

Hungary and Spain have large resources of subsurface water, thanks to their geology and their favorable geographic and climatic conditions. Given the current imbalance between water resources and quality of drinking water (a direct consequence of industrial pollution and exploitation) there is a pressing need to take steps to ensure supplies of drinking water for future generations.

The surface and subsurface paleokarst systems in Hungary and Spain constitute a resource that is of considerable value to the respective populations. In Hungary about 10% of the subsurface paleokarst systems are linked to thermal water resources and approximately 30% to petroleum reserves. All bauxite, limestone and dolomite resources and a substantial part of manganese ores are also related to paleokarst systems. In Spain the Valencia Trough is a Neogene oil-producing basin located in the western Mediterranean, in which the recoverable reserves are estimated at 250 mmbbo. The reservoirs, the oil of which is sourced from Lower Miocene black shales, are made up of karstified and fractured upper Jurassic carbonate rocks. The caves have an extraordinary value in terms of minerals, fossils, flora and fauna. The surface occurrences of these paleokarst systems play a crucial role in the formation of the landscape and in the evolution of the biosphere. Needless to say, these systems are increasingly important to the environment.

Consequently, the paleokarst systems enjoy a huge natural potential. Any exploitation of this potential should be carried out taking other resources into consideration so as not to upset the natural balance. Mining, industrialisation and urbanisation have, together with pollution and changes in the climate, resulted in a considerable drop of the karst water table. Therefore, the principal aim of the present studies is to draw attention to the need for protection and to the rehabilitation of the paleokarst aquifer systems.

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Supervisors: National Committee for Technological Development, Office of International Projects (Hungary, H-1052, Budapest, Szervita tér 8), and Ministry of Foreign Affairs, General Direction of Relations for Culture and Science (Spain, 41 José Abascal, Madrid)

Duration of the project: 1992–1995

Aims and Methodology: A comparative study of the paleokarst potential in Hungary and Spain within the framework of proper national paleokarst research. Systematic exchange of results and experiences by mutual visits of the Hungarian and Spanish researchers.

Scientific Results: Joint papers published by the Hungarian Academy of Sciences in the Acta Geologica Hungarica in the form of a collection of case studies. The first three papers of the present issue were elaborated in the framework of the project. Publishing of further articles is planned in later Acta numbers.

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Dr. László Korpás & Dr. Ramon Salas

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Platform margin palaeokarst development from the Albian of Gorbea (Basque-Cantabrian region, N Iberia)

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Palaeokarst structures at different scales occur in inner and outer settings of the Albian Gorbea rimmed carbonate platform. The best palaeokarst development occurs on the platform margin environment, from where we describe three principal horizons: Axular, Aitzgorrigain and Itxingote. The first two (Axular and Aitzgorrigain) are characterised by cavernous porosity and laminated carbonate infills. Their development occurred in two respective single stages during short-term subaerial exposure. The last palaeokarst (Itxingote) is characterized by channel and vug porosity and miscellaneous infills, including quartz sandstones. This palaeokarst developed multiepisodically in a complex way throughout the middle Albian (3.5 m.a.).

During karst evolution, sedimentary wedges composed of autochthonous and allochthonous carbonates as well as siliciclastics were deposited in adjacent slope and basin environments, mainly as lowstand deposits.

The palaeokarsts originated by subaerial exposure caused by relative sea level falls. Synsedimentary tectonic activity is closely related to relative sea level changes; moreover, it influenced palaeokarst location, morphology and development.

Introduction

The identification and study of karst features and infilling sediments in modern and ancient carbonate depositional environments have been an important research subject in carbonate sedimentology during the last decades. Palaeokarst features are important because of their sedimentological interest, as indicators of sea level falls and subaerial exposure of platforms, and their economic interest, since many reservoirs and ore deposits are related to such structures. Their identification and characterization helps in determining controls on shallow marine carbonate deposition, and in modeling karst-associated diagenesis and related porosity distribution.

Many examples of palaeokarst are known from the fossil record (e.g. James and Choquette 1988; Bosak et al. 1989). Few examples of palaeokarst from the Basque-Cantabrian basin were known until recent years. These include poorly documented palaeokarsts associated with ore deposits (e.g. Vadala et al. 1981;

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García-Mondéjar and Pujalte 1983; Herrero 1989). As a result of recent studies on the Lower and Middle Albian Gorbea carbonate platform, several examples of subaerial exposure and palaeokarstic features have been identified in different settings (Gómez-Pérez et al. 1992; Gómez-Pérez 1994). Our aim here is first to establish the depositional and diagenetic evolution of these palaeokarsts and then to attempt to explain the controls on their development in relation with the evolution of the Gorbea Platform. These results can be useful in the study of economic deposits and may be applied to other palaeokarst developments of the Basque-Cantabrian Basin. They also may serve for comparison with other case histories around the world.

Geological setting

Located 30 km southeast of Bilbao in the Basque-Cantabrian basin (N Iberia – Fig. 1), the Gorbea platform (11 km wide, 550 m thick) evolved as a partially isolated bank during the Late Aptian to the Late Albian. This bank was linked to the north to the Duranguesado carbonate platform (Fernández- Mendiola 1987; García-Mondéjar 1990; Gómez-Pérez 1994), and was surrounded by shallow siliciclastic environments to the south and the southeast and by inter-reefal basin sediments to the west and northwest (Fig. 1).

The Gorbea palaeoenvironments evolved from ramp to rimmed carbonate platform from the Aptian to the Early Albian (Fig. 2). A major unconformity occurred on the platform during the Middle Albian, and during the early Late Albian a residual carbonate bank formed before shallow marine siliciclastics finally buried the underlying carbonates (Fernández-Mendiola et al. 1993; Gómez-Pérez 1994).

The Gorbea platform attained lengths ranging from 4 to 10 km in a SE–NW direction. The sedimentary facies development was markedly asymmetrical mainly due to tectonic fault-block movements. The underlying fault-block basement allowed development of restricted platform interior facies in the southeastern side of the bank. Progradation of the bank occurred from SE to NW. In the northwestern end reef development took place at the platform margin.

On proximal carbonate platform settings depositional karsts capping shallowing-upward cycles are a relatively common feature. More significant palaeokarsts developed in the early Albian rimmed platform stage at the northwestern margin of the Gorbea platform.

Platform margin facies consist of floatstones to wackestones with a highly diverse fauna including rudists, brachiopods, echinoderms, gastropods, benthic foraminifera and other skeletal fragments. Encrusting or flourishing forms of *Bacinella irregularis* algae are also very common and are attributed an important role in the creation of an elevated rim at the platform margin. Platformward restricted water circulation resulted in deposition of shallow-water/low-energy miliolid, peloidal and fine bioclastic wackestones and packstones. Basinward

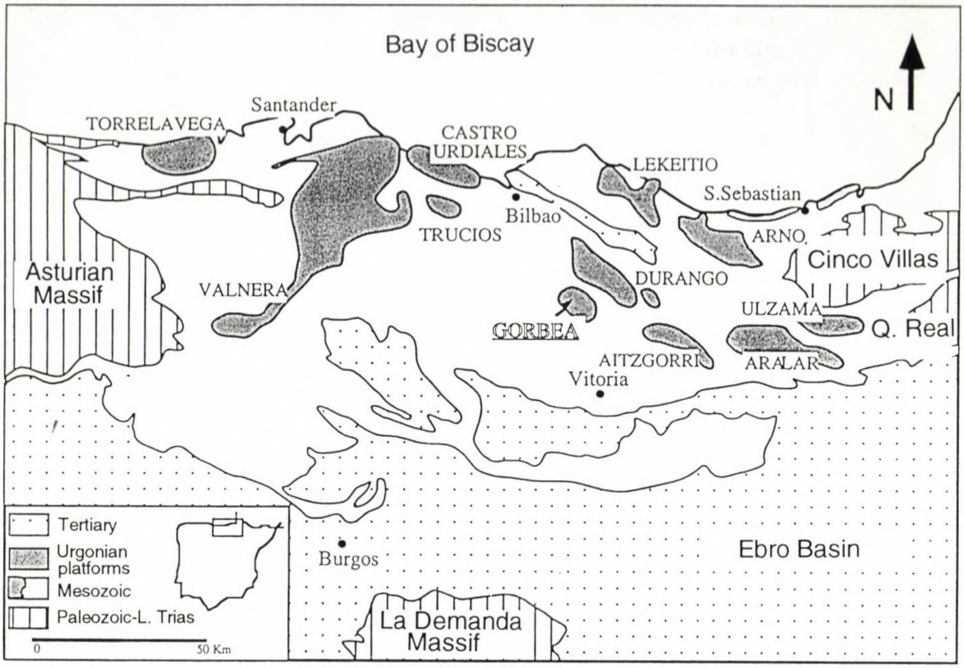


Fig. 1
Outcrop map of the Basque-Cantabrian Basin with indication of the Aptian–Albian platform carbonates and the analyzed Gorbea platform

depth increased rapidly so that platform margin facies graded to slope clinoforms made of coral boundstones and skeletal grainstones with original dips of up to 35 degrees. At the toe of the slope the clinoforms pinched out and interfingered with low energy autochthonous marls and marly limestones as well as allochthonous wedges including carbonate megabreccias and olistoliths (Fig. 3).

Palaeokarstic features at different scales occur in innermost platform and platform margin settings. The raised shallow water platform margin was exposed preferentially during relative falls of sea level and therefore major palaeokarstic features developed in this setting. We will focus on three palaeokarsts found at different stratigraphic horizons at the platform margin: the Axular, Aitzgorrigain and Itxingote palaeokarsts, respectively (Fig. 2). We will also recognize minor dissolution and microkarst features related to these and other exposure events. A fourth structure (Petrondegi) reported previously (Gómez-Pérez et al. 1992) was reinterpreted, based on its morphology, orientation and infill, as a submarine neptunian dike with likely(?) subaerial enlargement (Gómez-Pérez 1994) and is not described in this paper. The description of the characteristics and diagenetic history of each of the three major palaeokarsts follows, from older to younger.

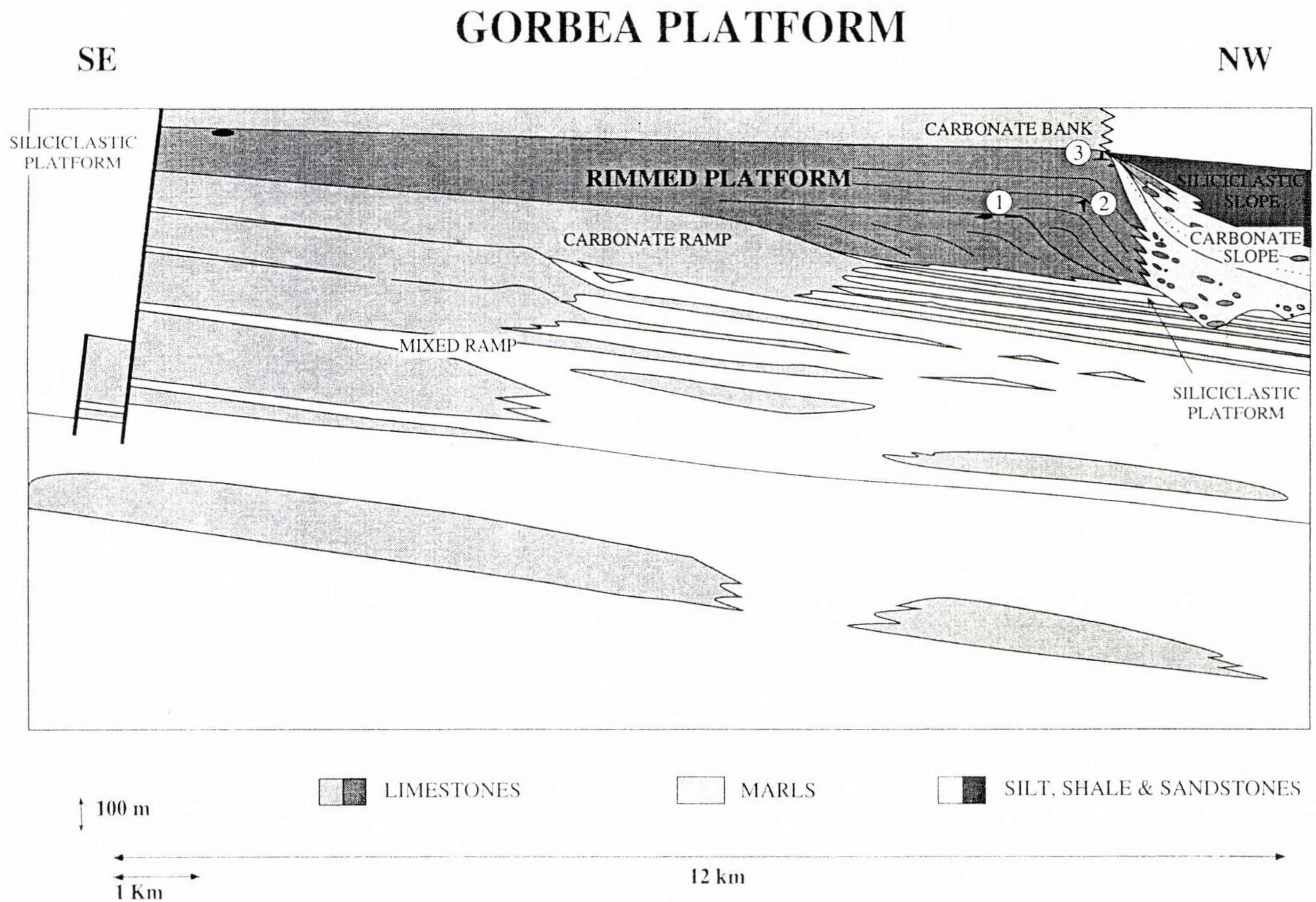


Fig. 2
Cross-section of the Gorbea carbonate platform. The rimmed platform and slope stages and the location of the studied palaeokarsts have been highlighted: 1. Axular; 2. Aitzgorrigain; and 3. Itxingote

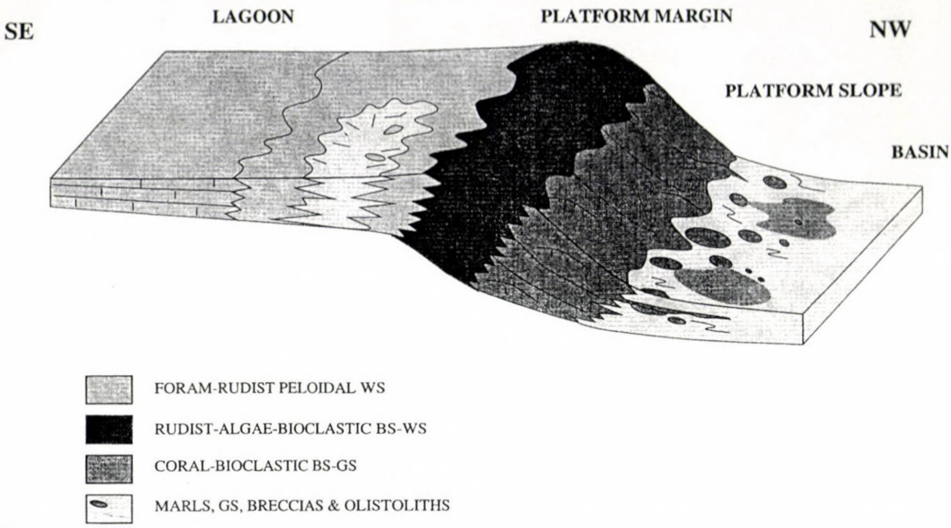


Fig. 3
Facies distribution model for the Lower Albian rimmed platform stage. Dissolution features are common in lagoon and platform margin settings. Water depth at the platform margin was few tens of meters (10–25 m) during times of platform flooding. Slope relieve about 130 meters. Grainstones in the slope and toe of the slope are winnowed carbonate detritus

Axular palaeokarst

The Axular palaeokarst occurred at the beginning of a stage of carbonate platform progradation (point 1 in Fig. 2). Prior to exposure the platform margin was not well developed and clinoforms were relatively gentle. Associated with platform exposure and palaeokarst development, erosion took place on slope settings, accentuating the depositional platform-to-basin relief. After this event a subsequent aggrading rimmed platform developed fully (Fig. 3).

The platform exposure and slope erosion with which the Axular palaeokarst is associated relate to an early Albian tectonic event documented in the area by synsedimentary faulting (Gómez-Pérez et al. 1993) and along the Basque-Cantabrian basin (e.g. Rosales et al. 1991, 1994; Aranburu et al. 1991). Block tectonics promoted the evolution from wide ramp settings to smaller carbonate platforms with tectonically-controlled steep margins that episodically supplied allochthonous carbonates to the adjacent basins.

Palaeokarst Morphology

The Axular palaeokarst consists of several meter-scale caves and vugs, very likely originally connected to each other. The palaeokarst crops out mainly in a limited area of 25x4m. Scattered features occur, however, in a wider area around the main outcrop. Karstic development is preferentially horizontal and the maximum measured depth for the main cavity is around 6 m, but outcrop is limited by modern erosion. Cavity morphologies are irregular and wall surfaces are mainly smooth but also irregular in places (Fig. 4a). Below the subaerial exposure surface pervasive dissolution and recrystallization of limestone fabrics are common.

Based on its position and morphology this palaeokarst is classified as a platform margin phreatic cave. These caves are horizontally developed, and presumably formed in a very active zone where two primary areas of dissolution converged: the top of the phreatic lens and the mixed marine-fresh water lens (i. e. Mylroie and Carew 1990; Raeisi and Mylroie 1995). Analog modern platform margin caves are described in the Bahamas (e.g. Mylroie 1988; Vogel et al. 1990).

Palaeokarst infill

Cavity infill consists of a fining-upward sequence. It contains bedded rudstones toward the base and laminated mudstones and grainstones toward the top (Fig. 4a). Scarce host rock micritic clasts are embedded in the calcarenite fill, and postdate spar cements on the cavity walls. These cements expand to centimeter-scale grikes connected to the main cave. The infill beds often onlap the cavity walls.

Laminated infill is interpreted as a cave sediment formed in a marine phreatic subaquatic environment. Low and high-energy conditions alternated during cavity fill. High-energy conditions prevailed in some of the conduits during the initial periods of cave infilling. During these high-energy phases coarse material composed of bioclastic and spar grains would have filtered down from the sediment surface, and then been redistributed and deposited in the caves. During calm periods carbonate mud settled or even formed in situ induced by cyanobacterial activity. Analog laminated infills have been described in a Paleozoic palaeokarst of the Alps (Schönlaub et al. 1991) and Hungary (Juhász et al. 1995; Korpás et al. 1996), and interpreted as phreatic in origin.

Diagenetic evolution

The evolution of this palaeokarst occurred as a single episode. This implies a relative sea-level drop during which the platform was exposed and dissolution occurred, followed by a sea-level rise during which the cavity was infilled. Spar cement precipitation probably occurred in a phreatic environment following dissolution in vadose to phreatic mixed-water conditions. As sea level rose the cavity was flooded by marine waters. In this setting laminated sediment deposition took place in a semirestricted quiet environment agitated only during storms. Coarser sediments toward the base of cavities were probably related to initial marine transgression, with energy decreasing as the sea flooded the entire cave system. Late secondary diagenesis caused local stiliolitization and recrystallization of the cave-infill contact.

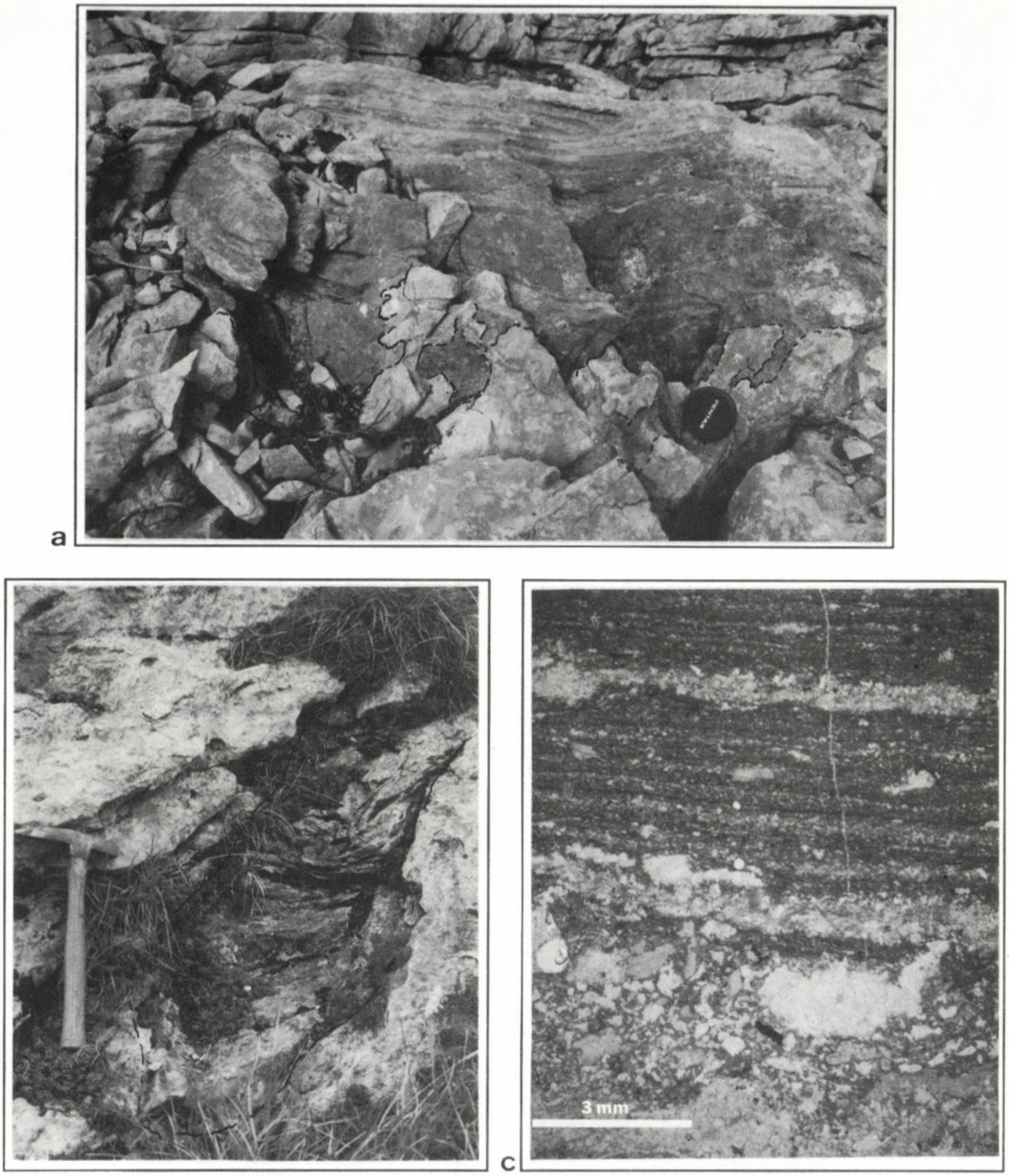


Fig. 4
Field photographs and microfacies of the Axular and Aitztorrigain palaeokarsts. a) Axular palaeokarst. Irregular contact of host rock and karst infilling. Note the transition from coarse rudstone at the base to laminated mudstone-grainstone upwards. b) Aitzgorrigain Palaeokarst. Microphotograph of grainstone-mudstone laminated facies. c) Aitzgorrigain Palaeokarst. Secondary cavity with "cul de sac" laminated infill

Aitzgorrigain palaeokarst

Following the unconformity with palaeokarst at Axular, an aggrading rimmed platform developed. Basement faults controlled the position of this platform margin, governed depth and subsidence and prevented progradation of the shallow water carbonates. The platform grew upward with vertical stacking of successive platform margins. The platform aggradation alternated with periods of platform exposure, with erosion on the slope and redeposition of allochthonous carbonates at the toe of the slope. Events of subaerial exposure are documented by karstic features at different scales. The most important of these features is the Aitzgorrigain palaeokarst (Fig. 2) firstly reported and briefly described in a previous study (Gómez-Pérez et al. 1992).

Palaeokarst morphology

The Aitzgorrigain palaeokarst consists of a major cavity 10–30 m wide and 40 m deep, with preferential vertical development (Fig. 5). Secondary meter-scale cavities are related to the main cave. Cave morphology is irregular but the walls are smooth. The subaerial exposure horizon is a flat surface with small-scale minor dissolution features.

In spite of its position at the platform margin the morphology of this cave differs from horizontal platform margin caves such as the Axular palaeokarst. Its anomalous morphology for a platform margin cave is interpreted as the result of joint control, with dissolution following a vertical pattern influenced by a larger-scale relative sea-level fall than in Axular. It has been classified as a "localized palaeokarst" or palaeokarst with a localized drainage system (term from Bonte 1963).

Palaeokarst infill

Infill of the Aitzgorrigain palaeokarst consists largely of laminated sediments very much like those from the Axular palaeokarst (Fig. 4). Along with these, barren mudstones without any biologic or sedimentary structures also occur. Cavity rim spar cements postdated by the infilling sediments are rare and thin.

Laminated facies consist of mudstone layers with interbedded grainstones. Grainstone layers consist of rudist and echinoderm skeletal and spar debris and are millimeter to centimeter-scale, graded and often with erosive bases (Fig. 4c). Laminated facies are often convolute or vertical, and their relations to karst walls are both onlapping or following their morphology. Less common structures are centimeter scale convolute lamination, bioturbation, microslumps and internal unconformities. Rare dolomitic mudstone clasts collapsed from the walls, indicate early dolomitization of the host limestone.

Deposition of laminated facies occurred in alternating low and high energy conditions. Periods of low energy resulted in carbonate mud deposition. These finely laminated or homogeneous muds suggest a very quiet subaquatic

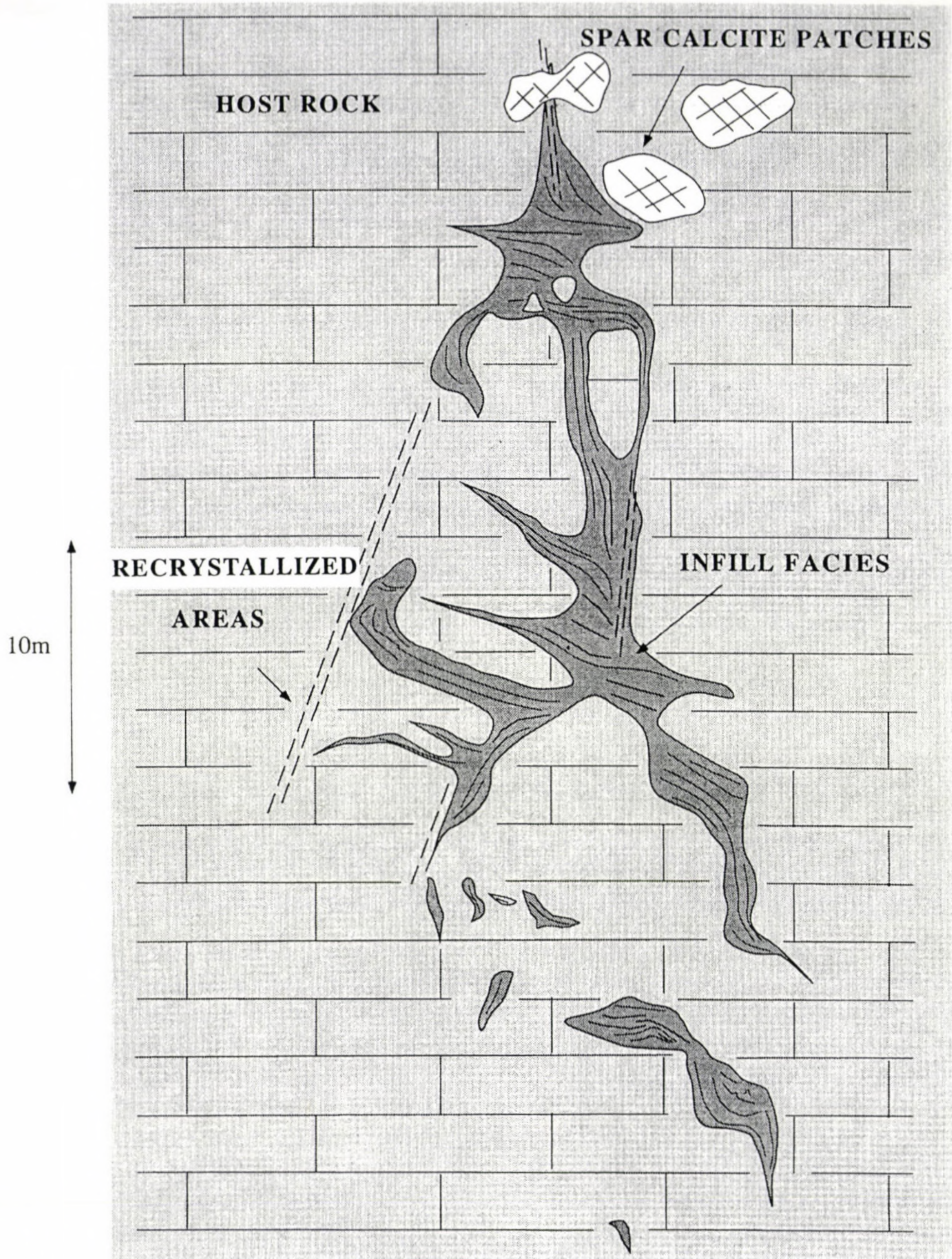


Fig. 5
Sketch of the Aitzgorrigain palaeokarst. The host limestone consists of rudist-skeletal wackestone to floatstone. Laminated infill either follows or onlaps cavity walls

restricted environment. During periods of high energy debris of skeletal grains infiltered in the cavity and were reworked and redeposited as graded beds, possibly during storms. Some of the spar fragments in the grainstone layers are interpreted as wall cements reworked and redeposited inside the cavity.

Modern analogs of storm influenced calcarenite infilling dissolutional cavities have been documented from the Bahamas Blue Holes (Smart et al. 1988; Bourrouilh-Le Fan 1997). Antigravitational cavity fill laminae suggest organic influence in deposition, probably by algae or cyanobacteria. A modern example of laminated micrite crusts formed in marine cavities influenced by biological activity of algae or bacteria and independent from the photic zone is described by Brachert and Dullo (1991). Laminated sediments similar to these are also known in modern algal mats (Park 1976).

Preinfill spar cements are interpreted as phreatic, probably meteoric. Coarse spar cements are widespread towards the top of the cavity, and very likely late in origin. They are interpreted as deposited in late diagenetic stages, infilling remaining porosity relict after burial of the karst. These cements are also found in fractures subparallel to the walls of the cavity, suggesting late reactivation of joints that controlled palaeokarst location and development.

Diagenetic evolution

Evolution of this palaeokarst occurred, as for the Axular palaeokarst, in a single stage of dissolution during subaerial exposure, followed by cavity infill as the sea reinundated the platform. Smoothness of the cavity walls would be related to erosion during rising sea level. Infilling occurred in a subaequous phreatic environment, probably during transgression following the sea level fall, and in restricted cavities with discrete connections to the sedimentary surface. Complete occlusion of porosity occurred during burial diagenesis.

Itxingote palaeokarst

The Itxingote palaeokarst is located on top of the rimmed carbonate platform system and associated to a major unconformity (Gómez-Pérez 1994). It developed during the Middle Albian and implies a sedimentary hiatus of 3.5 m.a. On top of the lower Albian rimmed platform system lie unconformably the Upper Albian carbonates of a residual bank system (Fig. 2). Recording the time involved in the unconformity, a wedge of deeper water slope and basinal siliciclastic and carbonate sediments infilled the adjacent basin (Gómez-Pérez 1994, Gómez-Pérez et al. 1994). In fact, at least three unconformities represented by megabreccias and sandstone turbidites are included in this wedge. Exposure and evolution of the platform at this time was related to activity of synsedimentary faults, which resulted in the break up of the platform in at least two uplifted blocks separated by a graben where shallow marine siliciclastic sediments accumulated (Fig. 6).

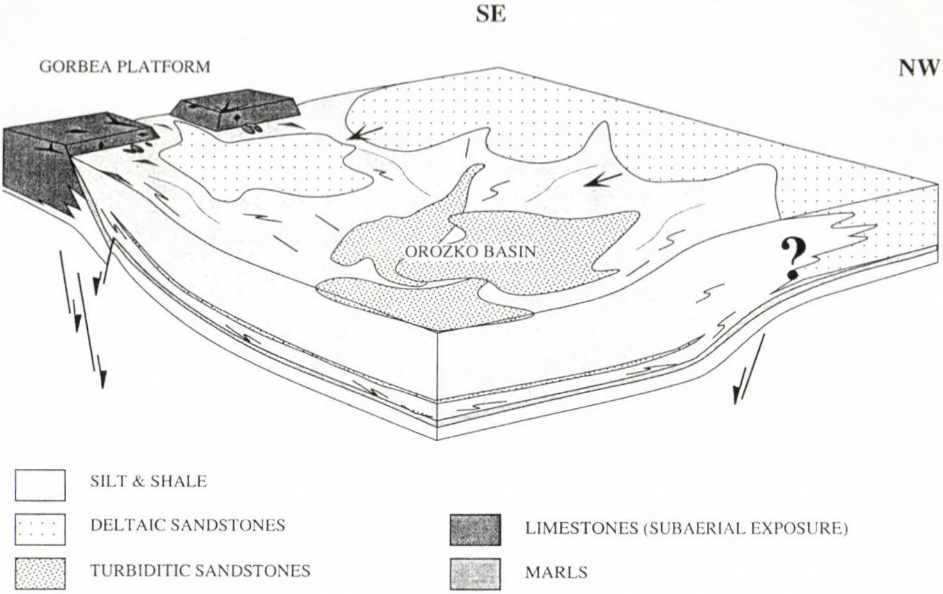


Fig. 6
Three-dimensional model of the Gorbea platform during development of the Itxingote palaeokarst. Block tectonics resulted in subaerial exposure and karsting of the tilted blocks

Palaeokarst morphology

Major dissolution features of the Itxingote palaeokarst deepen 42 m in platform margin facies. However the exposure surface has been removed by modern erosion, therefore total thickness of the interval with palaeokarstic features is unknown. Main dissolution features include channels and vugs. In the uppermost 13 m of the palaeokarst, channels have a prevailing vertical dimension. They resulted from fracture enlargement by dissolution and are centimeter to decimeter wide. They are perpendicular to the platform margin to trend and follow regional tectonic directions, paralleling faults in the vicinity. Vugs found associated to channels or disperse in the rock are irregular in shape and are also centimeter to decimeter in scale. At 13 m below the exposure surface horizontal channels dominate and a pervasively dissolved 50 cm thick horizon with very irregular dissolutional morphology occurs. Below this horizon dissolution vugs are small and disperse.

Meteoric dissolution occurred at several different stages. These are difficult to recognize at outcrop scale and are more evident at the microscopic scale, although they tend to be superimposed. Superimposing dissolution features have also been described at a bigger spatial and temporal scale by Goggin (1995).

Palaeokarst infill

Infilling of the Itxingote palaeokarst is complex. Miscellaneous products occluding porosity include sediments and cements accumulated at different stages and in different diagenetic environments. These palaeokarst fills were analyzed petrographically and under cathodoluminescence (Fig. 7).

Four principal events of dissolution-infilling have been inferred after initial deposition and micritization of grains in the marine environment. The first event of dissolution is followed by several phases of rim cementation and geopetal mud sedimentation. The initial fill consists of localized peloidal microbreccias and more abundant laminated peloidal mudstones with ostracod shells and rare bioclasts. These are followed by non-luminescent rim cement, mud and crystal silt at the bottom of cavities, and final geopetal mud (Fig. 8a). At the top of microdissolution cavities and in minor vugs where geopetal sediments did not accumulate, the infilling sequence consists of one or several generations of rim cements, showing banded bright and non luminescent zones.

Peloidal sediments, as lime mud, are interpreted as deposited in a restricted environment in submarine caves. Some examples of peloidal and internal micrite formed in modern and ancient submarine caves have been described elsewhere (i.e. Reid 1987; Chafetz 1986). Rim cements are also interpreted as marine phreatic. However, the presence of both vadose crystal silt (term from Dunham 1969) and non to bright luminescent cements suggests changes in the pore waters, indicating meteoric diagenesis in the fresh water phreatic environment (Fig. 8b).

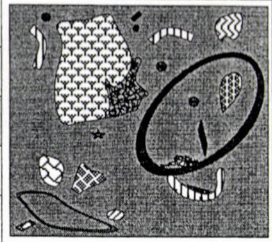
After a second stage of dissolution, deposition of geopetal dolomite occurred in voids and vugs. The dolomite formed in meteoric phreatic to mixed water. Sandstone sedimentation began during the latest stages of dolomite formation, and continued simultaneously to the precipitation of younger clear blocky spar in fractures and vugs. Clear spar cement is non to dull luminescent and precipitated in a marine phreatic diagenetic environment (Fig. 9a).

A third stage of dissolution occurred locally superimposed to older dissolutional features, and was followed by development of very dark dull luminescent cave speleothems. These were fractured immediatly afterwards, then partially dissolved, and finally cemented by rim to blocky spar cements in a phreatic (fresh water?) environment (Fig. 9b). A second event of geopetal dolomite sedimentation formed very likely in the mixed water zone, and it was followed by precipitation of coarse blocky spar that infilled almost all remaining porosity in a marine phreatic to burial environment.

Diagenetic evolution

As it has been described, almost 20 stages of porosity evolution have been recognized associated to this palaeokarst, including 4 stages of dissolution-fracturing (Fig. 7). The recorded event succession shows the complex evolution

	STAGE / PROCESSES	PRODUCTS	SEDIMENTARY / DIAGENETIC ENVIRONMENT
1	**DEPOSITION**	Rudis-coral-bioclastic Wackestone	Marine subtidal Moderate energy shallow water carbonate platform margin
1a	**Micritization/ Perforation**	Micritic coatings, pelletization, borings	Marine phreatic, stagnant water
2	**DISSOLUTION I**	Irregular, affecting bioclasts and lime mud: vugs	Mixed or fresh water phreatic
3	**Internal sedimentation**	Lithoclastic-peloidal microbreccia	Marine phreatic, very shallow, restricted
3a	**Internal sedimentation**	Peloidal laminated lime mud, bacterial?	Marine phreatic, very shallow, restricted
4	**Cementation**	Bladed turbid rim spar, NL	Fresh water phreatic to mixed water
5	**Internal sedimentation**	Geopetal lime mud / Geopetal Crystal silt, BL	Fresh water vadose?
6	**Cementation**	Bladed turbid spar. Rim. NL	Fresh water phreatic?
7	**Internal sedimentation**	Geopetal lime mud	Marine phreatic
8	**Cementation**	Turbid bladed rim	Marine phreatic
4-8	**Cementation**	Turbid bladed spar. Banded NL-BL (1-4)	Marine phreatic to mixed and fresh water?
9	**DISSOLUTION II**	Irregular, no selective; vugs, pipes, grikes	Fresh water vadose
10	**Cementation**	Geopetal dolomite & Dolomitic silt	Mixed zone?
11	**Sedimentation**	Quartz sandstones	Fresh water? to marine phreatic
12	**Cementation**	Equant turbid spar. banded BNL to dark DL	Fresh water? to marine phreatic
13	**DISSOLUTION III**	Dissolution vugs & grikes	Fresh water vadose
14	**Cementation**	Speleothems: Brown radial & concentric calcite. Orange to dark DL	Fresh water vadose
15	**Fracturing/ reworking**	Collapse & fragmentation of cements & geopetal deposition	Shallow burial
16	**Cementation**	Turbid bladed spar. NL.	Phreatic, fresh water?
17	**Cementation**	Geopetal dolomite/ dolomite cement	Mixed zone?
18	**Cementation**	Turbid to equant spar. NL to banded DBL-DL	Burial
19	**Fracturing**	Angular fractures	Burial
20	**Cementation**	Bl Dolomite cement NL Clear spar	Burial
21	**Compaction**	Stylolites	Deep burial



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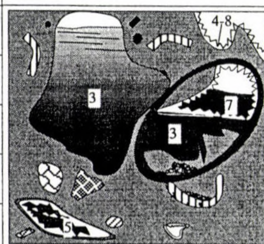

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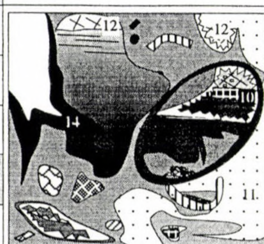

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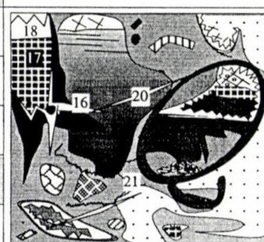

15-21

Fig. 7
Porosity evolution of the Itxingote palaeokarst. The recorded events and products and their characteristics are shown to the left. Porosity creating events are highlighted in gray. The right column shows synthetic sketches of the karst evolution

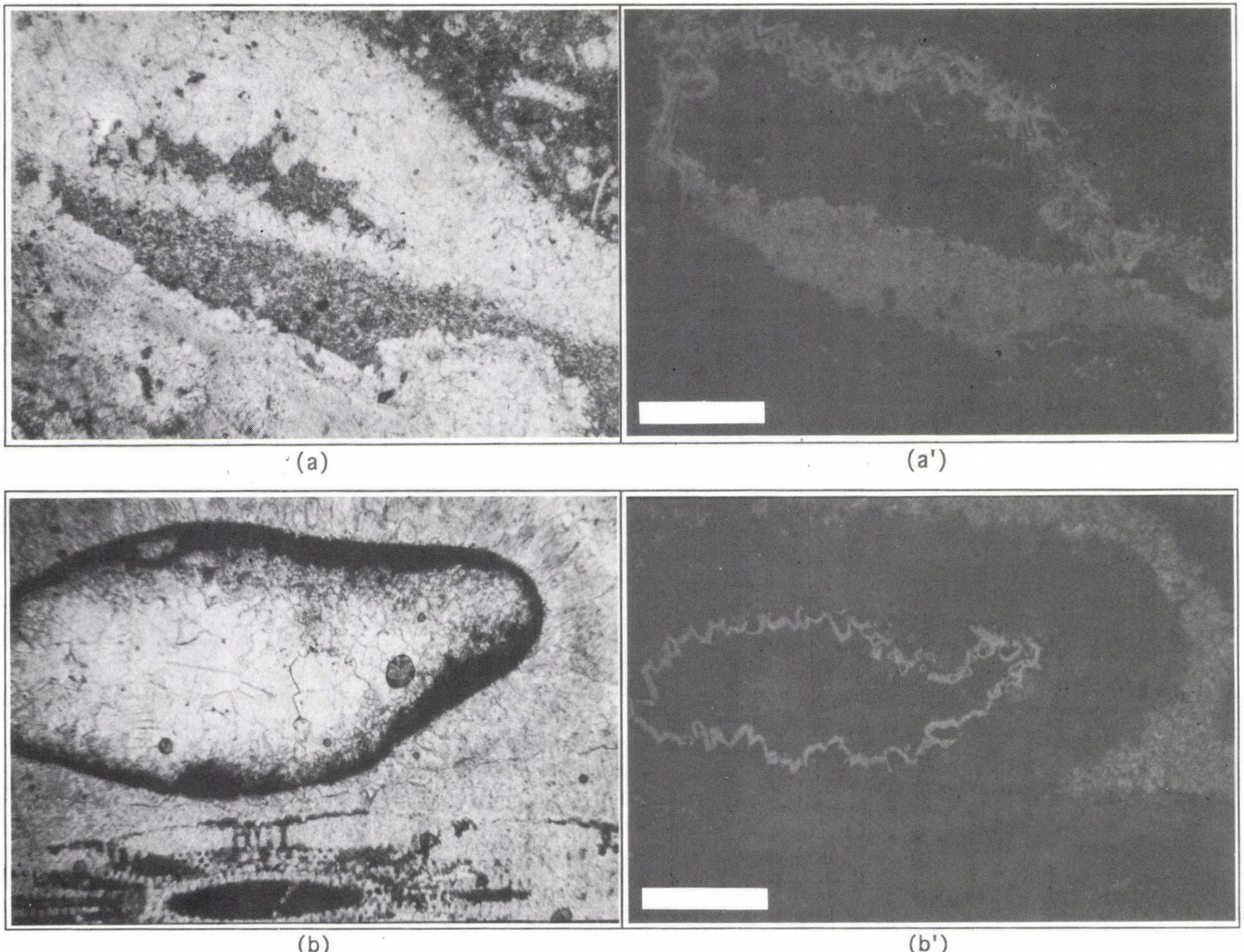

Fig. 8
Transmitted light and cathodoluminescence couples for some of the Itxingote palaeokarst cements and internal sediments. a–a′: Bladed rim cements (NL and banded BL–NL) interrupted by crystal silt (BL) and lime mud deposition at the bottom of a microcavity. After dissolution stage I. b–b′: Dissolved bioclast infilled by blocky luminescent marine cements (after dissolution I) and non luminescent to bright luminescent spar cements (after dissolution II). (NL – non-luminescent, BL – bright luminescent, DL – dull luminescent)

of the Itxingote palaeokarst, with multiple processes taking place over a long period of episodic karst dissolution and infilling.

The more remarkable dissolution features formed in the second dissolution event and are infilled by sandstones. Vertical dissolution features are interpreted as formed in the vadose environment, and horizontal features are attributed to the top of the meteoric phreatic water lens. Minor dissolution occurred below this horizon, probably in the fresh water phreatic environment. Because of its platform margin position, the top of the fresh water lens and the mixed water lens would have coalesced, and changes in diagenetic environment from marine phreatic to vadose fresh water would have been rapid. Change downdip from fresh water to marine phreatic was probably restricted to a narrow vertical zone. The last stage of dissolution-infill of the palaeokarst is areally restricted and major features have been recognized only in one of the tectonically uplifted blocks.

Controls on palaeokarst development

Relative sea level changes

Platform margin exposure occurred during relative sea level falls. In these periods carbonate production was interrupted, and platform areas were exposed to subaerial dissolution and diagenesis. Lowering of wave base produced increasing energy in the slope, and instability resulted in upper slope erosion of semiconsolidated sediment and downslope resedimentation on the toe of the slope and basin margin settings.

Estimated sea level falls related to the described palaeokarsts, considering an initial depth of about 10 meters for platform margin facies, are a minimum of 16 meters for the Axular palaeokarst, about 50 meters for the Aitzgorrigain palaeokarst, and a minimum of 23 meters for the Itxingote palaeokarst (limited outcrops for the first and the last palaeokarsts do not allow a better evaluation).

Relative sea level changes in the Gorbea area are, in many cases, demonstrably driven by active tectonics, clear in the case of the Axular and Itxingote karst-related unconformities, and very likely in the Aitzgorrigain and other unconformities (Gómez-Pérez 1994).

Climate

A humid climate in the Basque-Cantabrian basin during the Early Albian has been reported in previous works (García-Mondéjar 1979; Pascal 1984; Yusta 1993). This humid and warm climate was conditioned by the intertropical position, around 25–30 degrees north, of the area in the Early Albian (i.e. Lloyd 1982).

The presence of palaeokarst and their nature and characteristics support a humid climate. Percolation of dissolving fluids that results in karst-related

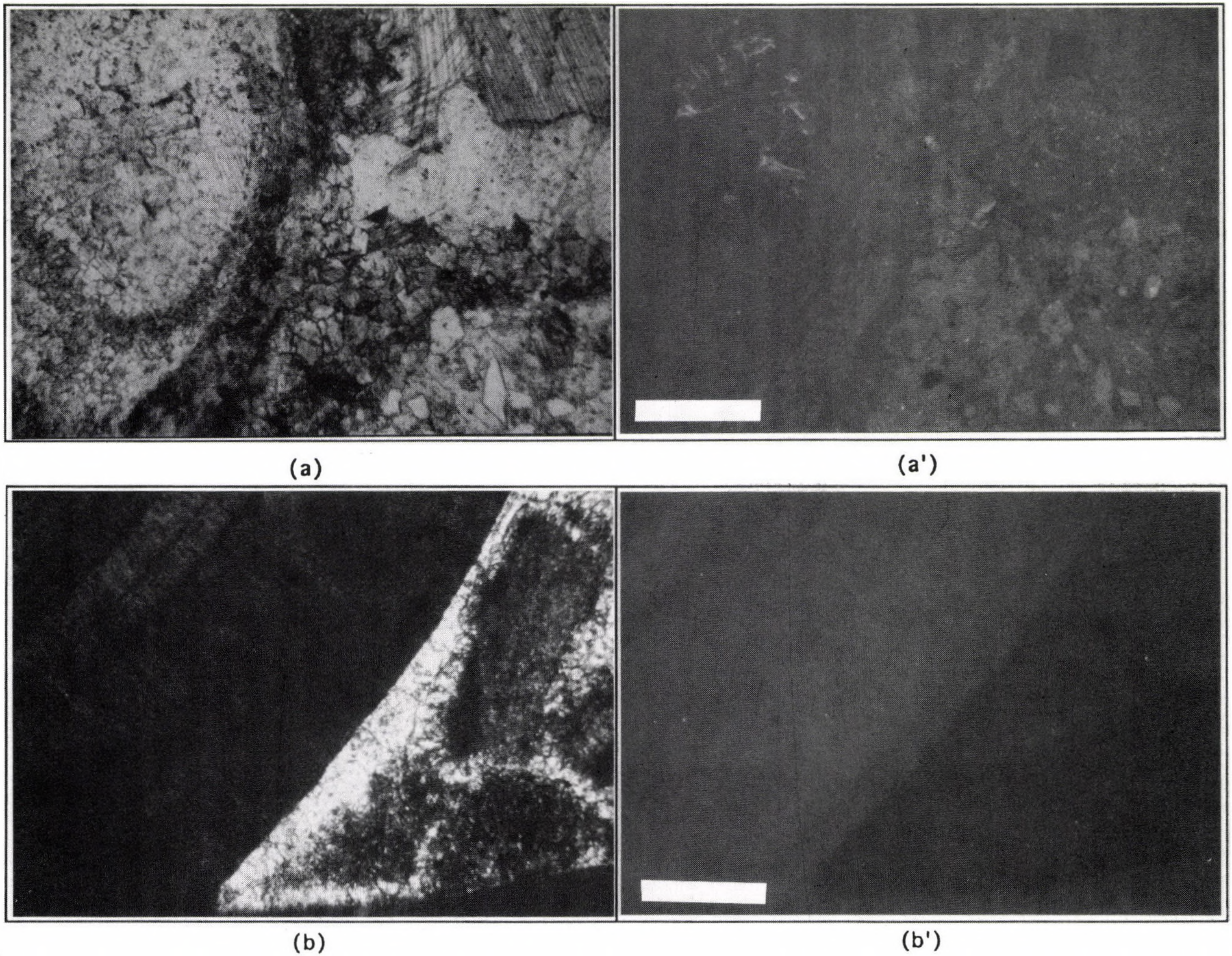

Fig. 9
Transmitted light and cathodoluminescence couples for some of the Itxingote palaeokarst cements and internal sediments. a–a′: Dolomite (reddish luminescence) and blocky spar with sandstone grains infilling porosity after dissolution stage II. b–b′: Speleothems (dark DL) and blocky rim (NL) cementing reworked clasts, after dissolution stage III. (NL – non-luminescent, BL – bright luminescent, DL – dull luminescent)

features, as well as precipitation of cements, are characteristic indicators of high pluviometry regimes.

Other indirect indicators of humid conditions are the absence of calcretes, evaporites or other signals of arid climate, and the presence of large amounts of siliciclastic deposits carried into the adjacent basin and deposited as prodeltaic and deltaic wedges.

Finally, the association of bauxitic deposits (Fernández-Mendiola 1987) to the Itxingote paleokarst unconformity in the inner platform (Gómez-Pérez 1994) further supports this interpretation.

Changes in nature of the cavity fills are related to sediment availability. During the Axular and Aitzgorrigain karst formation siliciclastic sediments were not available, as terrigenous input into the basin was absent. During the Itxingote karst development siliciclastic input from land areas located to the SSE was very significant. Increase in terrigenous input during the Middle Albian is related to tectonic reactivation in the source area in the entire Basque- Cantabrian Basin. This reactivation is recorded as uplift, breakup and exposure of carbonate platforms. However, a local change in climate in the Middle Albian implying increase of rainfall and more humid conditions with respect to the Early Albian is not discarded.

Duration of exposure

An important control in palaeokarst development is the extent of time of subaerial exposure during which dissolution occurred. In the case of the studied palaeokarsts different magnitudes are envisaged. For the Axular and Aitzgorrigain palaeokarsts a biostratigraphically resolvable time gap is not detected, and subaerial exposure in each case is supposed to have lasted a few thousand years. Development of karst in such cases occurred in a single stage during a relatively short period of time, so that dissolution concentrated at one site and pervasively acted upon it.

For the Itxingote palaeokarst a gap of about 3.5 million years associated to the unconformity has been biostratigraphically determined. Subaerial exposure during that period was episodic and alternated with phases of platform submergence and no sedimentation, during which siliciclastic sediments bypassed the platform and accumulated in the adjacent basin and in karstic cavities. This longer-lasting and discontinuous exposure resulted in multiple amalgamated and widely distributed dissolutional features. Superimposition of these features are attributed to old conduits constituting pathways for new generations of dissolving fluids.

Synsedimentary tectonics

Synsedimentary tectonics, as previously remarked, is considered one of the main factors controlling the karst development in the Gorbea area. This activity is related to extensional movements originated by the opening of the Bay of Biscay and the North Atlantic during the Mesozoic.

The Axular and Itxingote palaeokarsts are related to tectonically driven unconformities (Gómez-Pérez 1994). This is also probably true for the Aitzgorrigain palaeokarst (Gómez-Pérez 1994).

Karst morphology is closely associated to tectonic structures in the case of Itxingote, where major dissolution features resulted mainly from enlargement of tectonically originated joints. Karstification occurred on the margins of uplifted blocks and was related to synsedimentary faults (Fig. 9). Various blocks display different histories of diagenetic development.

In the case of the Aitzgorrigain palaeokarst, its morphology and location were very likely also related to tectonic fractures and joints developed at the platform margin. Enlargement of joints and fractures is considered the cause of the predominant vertical dimension of the karst and its localized occurrence.

Conclusions

Three palaeokarst horizons are described from an Albian rimmed carbonate platform system (Gorbea platform). These include two early Albian localized palaeokarsts (Axular and Aitzgorrigain) with laminated carbonate infill, and one middle Albian more extensive and mature palaeokarst with sandstone infill (Itxingote). The three of them occur at the platform margin setting.

Localized palaeokarsts occurred during short periods of exposure (probably at the scale of thousand of years) and low tectonic activity. The Axular palaeokarst is a typical platform margin, preferentially horizontally developed palaeokarst. The Aitzgorrigain palaeokarst is predominantly vertical and was probably controlled by dissolution along joints. More widespread karstification related to the Itxingote palaeokarst occurred mainly by enlargement of fractures during a high active tectonic stage and episodic exposure for a long period of time (3.5 m.a.).

The nature and morphology of the palaeokarsts are related to dissolution patterns driven largely by tectonic factors. The different kinds of infills relate upon sediment availability. Controls in palaeokarst development, finally, are relative sea level changes, local tectonic factors, climate and duration of exposure. Relative sea level changes are closely associated to tectonic activity.

Acknowledgements

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The karstification potential of the aquifers in the Val d'Aran (Catalonia)

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The present paper is set forth as a reflection on the concept of karstification potential and the different factors allowing its definition.

The discussion is based on the information acquired from in the major aquifers in the limestones of the Val d'Aran, as far as discharge, water temperature and variables of the CO_2–H_2O-carbonate system are concerned. These experimental data correspond to discrete measurements (fortnightly and monthly) carried out in five springs (corresponding to four karstic systems) between 1988 and 1991.

All hydrodynamic, hydrogeothermal and hydrogeochemical features reveal important differences in the degree of karstification of the aquifers drained by each one of the springs. Thus, the Joèu, Lastoar and Tèrme-Pila systems show a meaningful degree of karstification, while the Aigüèira system is essentially fissural, without deep karstification (weak karstification potential). These differences are reflected in Mangin's hydrodynamic classification and in the observed variability of the temperature and mineralization data (frequency distribution). All the studied waters have very low mineralization and CO_2, a condition which is closely related to the fact that the corresponding basins are in alpine and sub-alpine climatic domains with their slopes covered by snow during a great part of the year, with limited development of well-configured soils and, as a result, little CO_2 available for the water. Conversely, the various aquifers differ in their main type of recharge. The only system with essentially autochthonous recharge, of the diffuse type (unary system) is that of Aigüèira. All the others have a clear allochthonous component (impermeable outcrops drained by the aquifer) conferring on them a marked binary character which leads to the conclusion that the localized infiltration through swallow holes has a clearly relevant role. In this sense the most extreme case is the Joèu in which 85% of the surface of its basin is impermeable.

Hence it is shown that the limited karstification of the Aigüèira aquifer must be related to a clear predominance of slow infiltration, while the high potential observed in the Joéu must be the result of the important role played by the rapid infiltration through swallow holes.

Therefore, it clearly appears that at least in the mountain aquifers of the Val d'Aran the role of CO_2 (bio-climatic factor) in the karstification potential is very relative and that the hydrodynamic factor (water potential) is one of paramount importance.

Key words: degree of karstification, karstification potential, experimental karst, CO_2, hydrodynamics, autochthonous karst, allochthonous karst

Experimental karsts and evaluation of the degree of karstification

Background

Intensive research in karsts, experimental systems or basins and also in the so-called experimental perimeters has been an adequate framework for hydrological studies of aquifers in carbonate rocks (Shuster and White 1971;

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Drake and Harmon 1973; Mangin 1975; Atkinson 1977; Bakalowicz 1979; Botton 1984; Gunn 1986; Bonacci 1987). These studies have been especially relevant in France but also in other countries such as Switzerland, the United Kingdom, Italy, the former Yugoslavia, Canada, the U.S.A. and New Zealand, amongst others.

In what could be called the Anglo-Saxon school or contribution, some works of great interest stand out because they are the first attempts to systemize the functioning and the structure of carbonate aquifers. Those of greatest interest are Shuster and White (1971) and Atkinson (1977). These authors define a wide spectrum of carbonate aquifers between two extreme types: those of diffuse flow and those of conduit flow. Later works study these concepts in greater depth (Ewers 1982; Quinlan and Ewers 1985; Hobbs and Smart 1986). More recently the monographs of Dreybrodt (1988), White (1988), Ford and Williams (1989) and White and White (1989) stand out.

The French contribution, from the Moulis Subterranean Laboratory, deserves to be mentioned separately. The work carried out has been developed since the 60s until the present-day, being characterized from its very beginning by a theoretical or epistemologic outlook, based on the explicit use of a systemic approach and in an experimental and interdisciplinary orientation. In this sense it can be stated that the research which has been developed has had, in the systemic paradigm, a solid and consistent methodology. This theoretical position has led to outstanding results in the entire research carried out. Their contribution can be considered as a conceptual breakthrough in karst knowledge and a remarkable advance in the theoretical karstology (Mangin 1975; Eberentz 1975; Andrieux 1978; Rouch 1978; Bakalowicz /1979/). Freixes (1993) made a critical synthesis of these studies.

The contributions correspond to different disciplines: hydrodynamics, hydrogeothermics, hydrogeochemistry and hydrobiology. The works on hydrodynamics (Mangin 1975) deal specifically with the various methodologies to define the unit, the functioning and the structure of the system. The most significant results to be mentioned are a quantitative classification based on hydrodynamic parameters (i and k parameters – Bakalowicz and Mangin 1980), a definition of the karst as a system introducing a fundamental feature in the drainage structure concept, as well as an interesting outline or conceptual model of the system or karstic aquifer. Andrieux (1972, 1976, 1978) uses temperature as a hydrogeological marker, characterizing the subterranean circulation and the heterogeneity of the saturated and unsaturated areas of the karst. Bakalowicz applies the concept of the regionalized variable to mineralization (or conductivity) with the aim of characterizing the karstification degree and of classifying the carbonate aquifers. From the starting point of hydrogeochemical research he introduces details to Mangin's hydrodynamic scheme, especially in the definition of the infiltration type (Bakalowicz 1977, 1979, 1981). These authors state that the predominance in a system of the superficial or subterranean morphology can be interpreted from the hydrodynamic and hydrogeochemical data. In this way the predominance of the surface karst is related to the slow infiltration one (and the epikarst) and the endokarst is related to the rapid infiltration (and also with infiltration through swallow holes). Generally there is a balance between both morphologies. In the overall organization of the aquifer the superficial and the deep morphologies are also closely related.

The karstification potential concept

From Mangin's contribution (1975, 1978, 1982a, 1994) an important collection of ideas and methodological and theoretical concepts can be highlighted. They present a great contribution to the explanation of karstic phenomena as a whole. This author proposes an analysis from three different points of view, which look at: 1) the morphology, or karstic landforms as a result of karstification, 2) the hydrology, starting from the water circulation as the fundamental process, and 3) the thermodynamics, as a basis for the energy which determines and controls the processes responsible for the karstification (Thermodynamics of Irreversible Processes (TIP); Prigogine 1968). The karstification potential concept has its niche in this third approach (Freixes 1995).

The aforementioned author defines karstification potential as "the set of conditions constituted by the external morphology and/or the geologic structure and to a certain degree of water and CO_2 potential". In later papers, Mangin (1994) explicitly introduces the concept of "high hydraulic gradient" as an important characteristic of the karstification potential.

From an energy point of view, karstification potential corresponds to the energy implied or introduced by hydric circulation, i.e. by karstification (karstification processes – Mangin 1982a). From this energy perspective the geologic, geomorphologic and climatic conditions leading to karst formation can again be found. Consequently the knowledge and definition of the potential driving karstification in a specific system must go through a thorough analysis, in a wide sense, of the geodynamic factors of its context. The importance of the endogenous geodynamic factors and the present tectonic activity can be determinant to the erosive potential and, as a result, to the karstification potential (Freixes 1995). This has been demonstrated by studies carried out in southwestern Asia and the Caribbean: the karstification potential depends much more on present tectonic activity than on the climatic factor (plate tectonics role in the regional context – Geze and Mangin 1980; Mangin and Bakalowicz 1990; Freixes 1993).

Entropy and organized structure in the karstic aquifer

The karstification process begins from the moment that the geodynamic conditions favour an important karstification potential. In this way the aquifer, initially of a fissural hydrogeological nature, evolves toward a heterogeneous and organized structure of voids peculiar to karstic aquifers. This process of an increased order (organization) is translated into a decrease in the entropy of the system.

These characteristics are peculiar to the open systems which evolve far from equilibrium, i.e. systems which interchange matter and energy with the environment (Prigogine 1968; Nicolis and Prigogine 1992). In the karst and karstic aquifers investigation these concepts have been introduced in several interesting Mangin works (1978, 1982a, 1994; see Freixes 1993). This author has also pointed out that the geometry of the karstic aquifers is of a fractal nature (Mangin 1986).

However, the organization of the structure of the karstic aquifers can be very diverse, depending upon the factors determining the karstification (Mangin 1978, 1982a, 1982b). In this sense the character of the systems (according to Mangin 1975, 1978 and 1982a and Walliser 1977), either binary (allochthonous recharge coming from impermeable materials) or unary (autochthonous recharge coming from the same permeable materials constituting the aquifer), is one of the aspects which can influence in a more significant way the structural geometry of the system (and of the aquifer) and also the degree of karstification development.

In the unary systems (in which the concepts of system and aquifer are coincident) the structure will differ depending upon the area considered: unsaturated or saturated (Mangin 1975; Freixes 1993). In the unsaturated zone the structure is heterogeneous with vertical drains and lateral domains of fissural structure; the saturated area is characterized by a more marked heterogeneity, with drains and annex systems. Both types of structures correspond to intense karstification but present a different hydrodynamic behaviour (Mangin 1975, 1974; Freixes 1993). In the unary examples, some characteristics of the basin may have a certain influence causing intense anisotropies in the endokarst structure, as happens with the presence of poljes corresponding to the existence of a concentrated surface runoff which, finally, recharges the aquifer through swallow holes. In the binary systems (those without equivalence between the aquifer and the system) the draining structure is of a filamentous type, from upstream to downstream, being characterized by a geometric structure which can be compared to the saturated as well as to the unsaturated zone (Mangin 1978, 1982a). Many examples can be found in the Pyrenees range. The known large endokarstic networks correspond mainly to binary systems. However, these considerations should be contemplated within the complexity of the karstification potential which is different in each example. Thus, the type of the resulting karstic structure will depend upon the relation between the different factors implied.

The karsts of the Val d'Aran

Introduction

The aim of this paper is to evaluate the degree of karstification based on the experimental data from the various aquifers in limestones of the Val d'Aran and to analyze and discuss the karstification potential of each case. The experimental data or markers used in the evaluation of the degree of karstification are flow rate, temperature and mineralization measured at the natural discharge point of the systems. The discussion will be essentially centred on two systems: the Joèu system and the Aigüèira system. Each has quite a different hydrogeologic nature, i.e. their functioning and structure are completely different. The results found in the Lastoar and Tèrme-Pila systems will also be commented on. In all cases the experimental data on temperature and mineralization correspond to discrete measurements carried out during a four-year period between 1988 and 1991.

The Val d'Aran: physical characteristics

The Val d'Aran is located in the central Pyrenees and is part of the high atlantic basin of the Garona River (Fig. 1). In the Aran domain the river has a run of 45 km and a basin of around 590 km^2 with a very remarkable altitude gradient: the maximum and minimum heights of the basin are the summit of Mt. Aneto (3404 m) and the river path in Pont de Rei (600 m) respectively.

At low and medium heights, the climate is of the atlantic montane type with relatively abundant rain (928.4 mm rainfall in Arties at a height of 1138 m) evenly distributed throughout the year. The temperatures are cold in winter (1.9 °C average) and mild in summer (16.2 °C). At higher altitudes, the climate is of the alpine and sub-alpine type: at Bonaigua station (2072 m) the yearly average temperature is 2.2 °C, with a rainfall module of 1250 mm, more than half of it being snow. During the winter months, and above 1700 m, the rainfall is almost exclusively snow (Rijckborst 1967). By spring, coinciding with the start of rising temperature, most of the snow melts causing a massive (generalized) infiltration of water into the aquifers leading to seasonal flooding. In the Joèu system, the rainfall in the upper part of the basin (Maladetes massif – at an altitude of 3000 m) can reach around 2500 mm or more (Arenillas et al. 1993).

The Aran's aquifers: physical characteristics

The Joèu, Lastoar, Tèrme-Pila and Aigüèira hydrogeologic systems constitute the most important examples within the Val d'Aran context. These systems, with the exception of a few localized studies (Rijckborst 1967), had never been the subject of experimental research (Freixes et al. 1991, 1993).

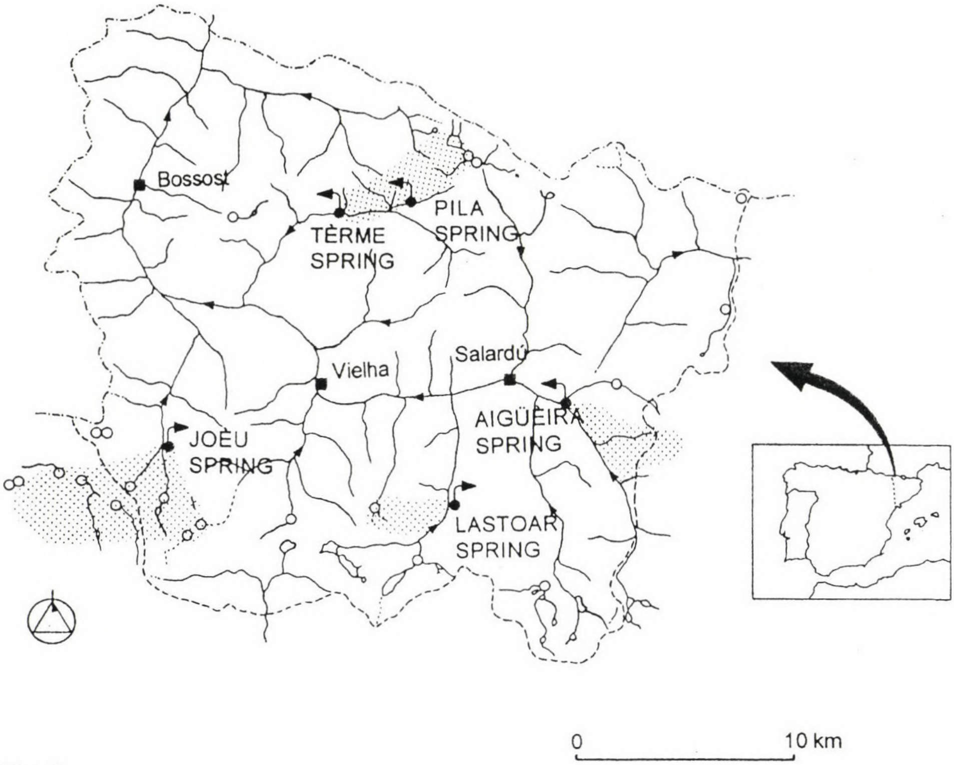

Fig. 1
Geographical location of the major karstic aquifers in the Val d'Aran

The most important physical and hydrodynamic characteristics of these systems are summarized in Table 1.

The Joèu system (Fig. 1, Table 1) is formed by the capture of the high basin of the mediterranean Ésera river by the atlantic Garona. It constitutes the most exceptional example of a karstic capture in the Pyrenees and presents a complex structure of the binary type. The impermeable substrate consists of late Hercynian granodiorites, and Carboniferous metapelites and metapsamites; the aquifer develops in the Devonian metamorphic limestones. The important superficial runoff (Aigualluts, Barrancs, Escaleta and Renclusa rivers) and the existence of many relevant swallow holes (Aigualluts, Renclusa, Hòro, etc.) are remarkable characteristics of this system. The synclinal macro-structure, with the Carboniferous shales being the ceiling of the Devonian limestone, imposes a captive character on the aquifer. There is probably a deep circulation at around 200 m below the system spring altitude (Joèu spring, 1410 m).

The Lastoar system (Fig. 1, Table 1), with a much smaller basin, also presents a complex structure of the binary type. The materials consist of Carboniferous granodiorites, metapelites and metapsamites and Devonian marmoreal limestones. This unit is partially recharged by the swallow hole located in Lake Bargadera (headwaters of the Tarters stream) and discharges in the Valarties valley through the Lastoar spring at 1530 m.

The Tèrme-Pila system (Fig. 1, Table 1) presents a complex structure, with a binary type basin. The materials consist of Cambro-Ordovician metapelites and marmoreal limestones. The system is recharged by many swallow holes located in the carbonate levels which appear to the south of Lake Liat, in the Tor Plain, and in the Palomera range. The Unhola swallow hole which drains Lake Liat is the most important of the system. The discharge takes place through the Pila (1765 m) and Tèrme (1570 m) springs in the Sescorjada and Varradós valleys respectively.

Table 1
Physical and hydrodynamic characteristics of the karstic systems in the Val d'Aran

	Joèu	Lastoar	Tèrme-Pila	Aigüèira
Catchment area (km^2)	27–31	3.5–4.0	9–11	10.5–13.5
Maximum altitude (above msl)	3404	2526	2880	2656
Swallow holes altitude (above msl)	2000	2000	2100	–
Springs altitude (above msl)	1410	1530	1570 / 1765	1475
Discharge (m^3/s)				
mean	1.58	0.17	0.09 / 0.38	0.26
maximum	12–15	0.55	0.36 / 2.20	0.54
minimum	0.25	0.04	0.03 / 0.06	0.13
Dynamic reserves (hm^3)	1–3	≈1	≈4	≈10
Mean annual runoff (hm^3)	49.9	5.3	2.7 / 12.0	8.3

The Aigüèira system (Fig. 1, Table 1) presents characteristics comparable to those of a unary type karst. The aquifer is formed by metamorphic Devonian limestones. The infiltration is of a diffuse character and, unlike the other systems, the existence of any relevant swallow hole has not been proven. The discharge is basically through the Aigüèira spring (1475 m).

Hydrodynamics

The discussion in this chapter will be based on the general hydrodynamic characteristics of the different systems and on its position in the Mangin (1975, 1994) hydrodynamic classification.

The Joèu and the Aigüèira systems present hydraulic characteristics which are completely opposed, as shown in Table 1. The Joèu unit presents the peculiar features of a system with a well-developed karstification and drainage structure. The mean annual runoff, about 50 hm^3 for a basin area between 27 and 31 km^2, is a clear indication of its drainage capability. In Mangin's classification it is located in dominion III (Fig. 2) corresponding to aquifers with higher karstification upstream and with delays in the recharge which in this case are due to the important snow cover and to the high development of the superficial drainage network. During the high-water season the travelling time can be very short, around 12 hours.

The Aigüèira system presents the peculiar features of a fissural aquiferous system: lack of deep karstification and of drainage structure. The mean annual runoff is of relatively small importance: 8.3 hm^3. In any case its dynamic reserves, around 10 hm^3, are very important. In Mangin's classification it is located in dominion V, corresponding to aquifers with a 'k' value above 0.5, i.e. to non-karstic aquifers. In Aigüèira the 'k' value is 1.3 (Fig. 2).

The Tèrme-Pila and Lastoar systems present intermediate hydrodynamic features between those of the Joèu and Aigüèira ones. In Tèrme-Pila, as well as in Lastoar, the karstification degree is significant. A discontinuity exists in the Tèrme-Pila system: the karstification is more important upstream than downstream. Effectively, the spring of Pila is connected to a part of the aquifer which is more karstified and to a drainage structure with a certain degree of development. The Lastoar example, with many particularities (Freixes et al.

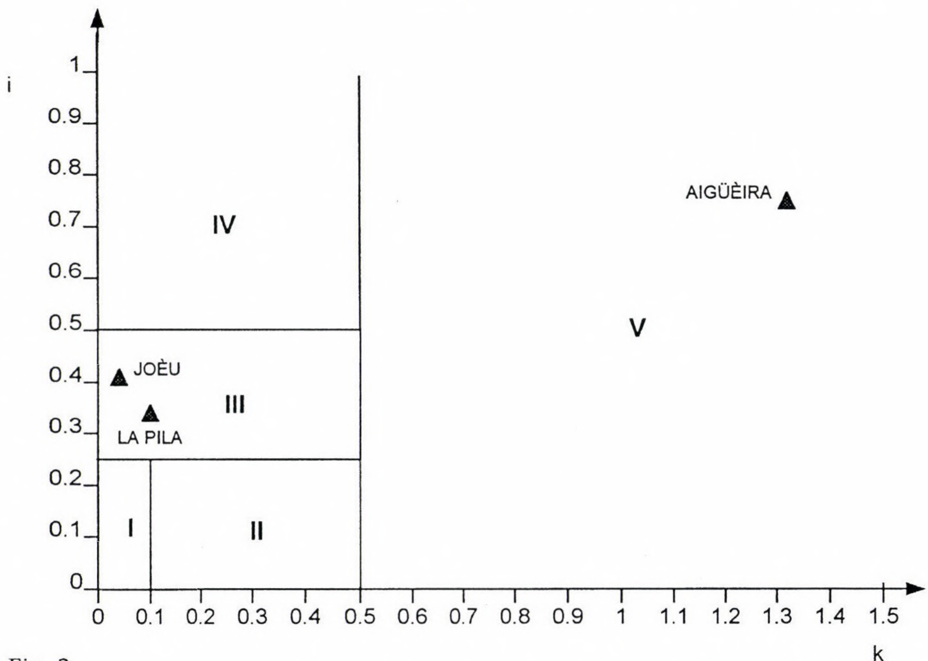

Fig. 2
Characteristics of some of the Val d'Aran aquifers according to the Mangin hydrodynamic classification (1975)

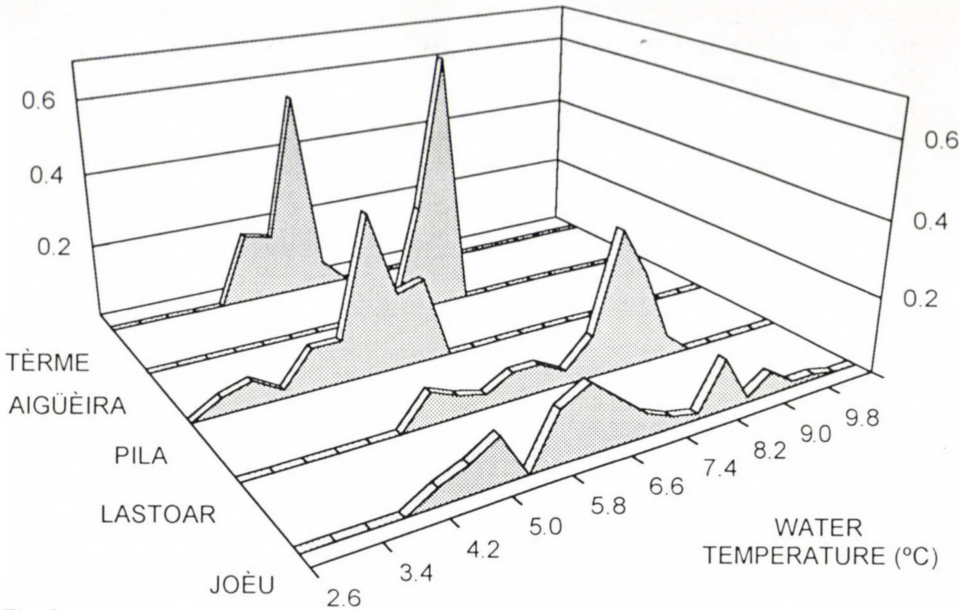

Fig. 3
Frequency distribution of the water temperature in the major karstic springs of the Val d'Aran

1993, in publication; Freixes, in preparation) shows overall relatively well-developed karstification and a certain drainage structure.

Hydrogeothermics

The study of the Aranese aquifers' temperatures has been the subject of a monographic work (Freixes et al. in publication) where the variations in each one of the examples are interpreted. The information obtained from the frequency distributions of the water temperature in the different springs is synthesized below (Fig. 3).

In the Joèu spring, the frequency distribution is markedly polymodal, with four well-defined modes. This distribution indicates that along the hydrologic cycle several waters with completely different temperatures reach the spring (the temperature variation range is 5.6 °C). This shows that the aquifer demonstrates important karstification and a well-developed drainage structure. Thus, the heterogeneity and the complexity characterize the functioning and the structure of the system.

In the Aigüèira spring, the frequency distribution of the temperature is unimodal. This indicates that the water temperature in the spring is quite constant (the variation amplitude is 0.6 °C). The different thermal signals of the superficial waters recharging the aquifer are not observed in the spring, i.e. the input signals are thoroughly homogenized. This result clearly demonstrates the fissural and homogeneous character of the aquifer structure, without deep karstification and without a drainage structure, at least 'sensu strictu'.

The Tèrme and Pila springs present different frequency distributions. In Tèrme (the spring of the system located at a lower altitude) the distribution is bimodal, with one well-marked mode and a second, less prominent one defined by waters of lower temperature (the variation amplitude is 1.3 °C). This result indicates the existence of an incipient karstification, scarcely developed, with certain privileged flow paths which would facilitate the arrival of waters with comparatively faster transit. In the Pila spring (located 200 m above that of Tèrme) the distribution is polymodal (the variation range is 2.8 °C). This fact indicates that the spring drains a dominion of the aquifer with some karstification development and a somewhat significant drainage structure. Therefore, in the Tèrme-Pila system, karstification is more important upstream than downstream and consequently there is a remarkable discontinuity in the spatial development of the karstification.

The Lastoar spring presents a polymodal distribution, with three well-defined modes (the variation amplitude is 3.0 °C). Waters of different temperatures reach the spring indicating the existence of a well-developed drainage structure. Therefore, the karstification has certain importance.

Hydrogeochemistry

Frequency distribution of the mineralization

The data on the evolution of the chemical variables and on the frequency distributions of the mineralization in the Val d'Aran karstic springs have been interpreted in several papers (Freixes et al. 1991, 1993).

The interpretation of the frequency distributions of the mineralization (Fig. 4) is comparable to that carried out for the temperature distributions. Thus, the Joèu spring has a polymodal distribution which indicates a well-developed karstification degree and drainage structure, with a significant complexity. The unimodal distribution of the Aigüèira spring indicates an aquifer with neither deep karstification nor drainage structure. The Lastoar, Tèrme and Pila springs present intermediate characteristics which lead to interpretations equivalent to those achieved from the temperature frequency distributions.

Mineralization and CO_2

The mineralization values of mountain waters are generally speaking quite low. This statement is also valid in the case of subterranean waters, even for those having relatively long travelling times (water-rock contact time) and important temporary storage.

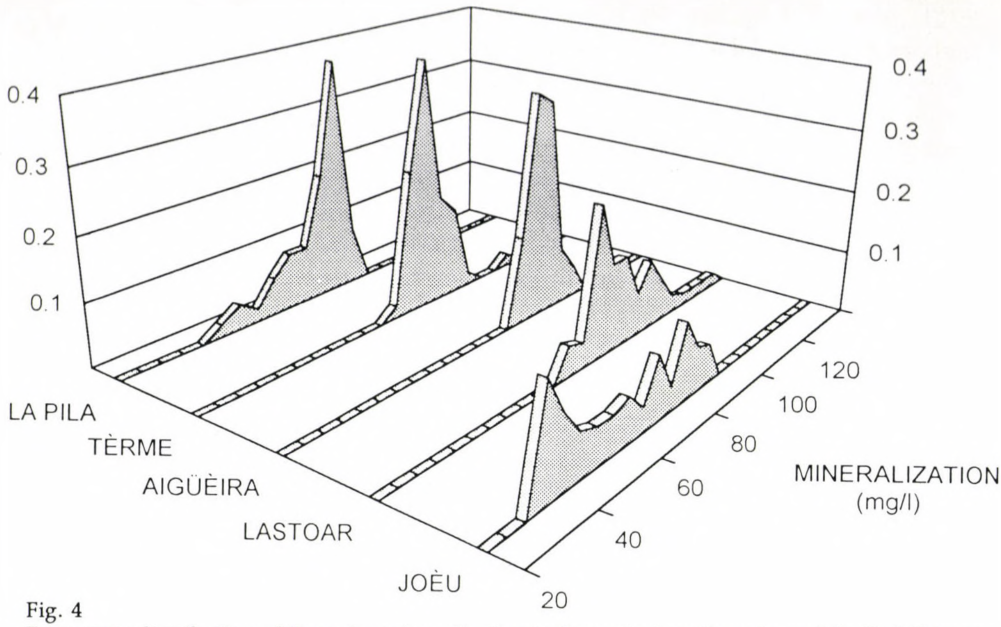

Fig. 4
Frequency distribution of the water mineralization in the major karstic springs of the Val d'Aran

In reality the essential aspect is the limited capability for dissolution in the water as a result of the low CO_2 concentrations. In the case of the Aranese aquifers, these are amongst the lowest mentioned in the bibliography (Table 2). In this sense it is important to point out that the basins feeding the karstic aquifers of the Val d'Aran have average altitudes above 2000 m, being in the alpine and sub-alpine climatic domains. These slopes remain covered by snow over six months of the year. Under these conditions CO_2 production in the soil has a markedly seasonal character with maximum values between the end of spring and the end of summer, and practically nil during the entire period the soil is covered by snow. This type of dynamics leads to low overall CO_2 production and as a result a low availability of CO_2 in the hydric flows.

Another factor contributes to this low availability. Well-developed soils only exist in the lower part of the valleys (areas where the slope is small). Consequently, the area occupied by these soils is relatively small.

CO_2 and the hydrodynamic factor in karstification potential

These last considerations permit the interpretation, with much more accuracy, of the differences observed between the karstic systems in the Val d'Aran as far as water mineralization is concerned. This will also allow specifying the role of the different factors intervening in the karstification potential.

The only system with an essentially autochthonous recharge, of the diffuse type (without a developed superficial drainage network), is the Aigüèira one (with an annual runoff of 8–10 hm^3). This is the system with a comparatively higher mineralization indicating a higher efficiency in CO_2 incorporation. However, this has not led to a higher karstification, but on the contrary to a lower one. This is an aquifer the permeability of which comes mainly from fissuring, without a drainage structure. Its karstification potential has not been important. Under these conditions, the hydrodynamic factor is almost irrelevant and considering the low CO_2 concentrations it is difficult for karstification to progress. Nevertheless, in the entire basin there is a noticeable development of the exokarst (possibly even with inherited forms): karren forms in most of the surfaces, various depressions (dolines, swallow holes of small importance, etc.). We do not know about the existence of important epikarstic aquifers. These observations indicate a higher karstification activity in the superficial part of the massif, which presents the typical features of a karst. That is to say the

Table 2
Mineralization (mg/l) and pCO_2 (%) average values in the springs of the main carbonate aquifers in the Val d'Aran and the high valley of the Noguera Pallaresa

	Mineralization (mg/l)	pCO_2 (%)	Discharge (m^3/s)		n
			minimum	maximum	
Joèu	62.4	0.012	0.200	12.00	85
Lastoar	99.2	0.030	0.040	0.550	54
Tèrme	88.6	0.026	0.035	0.360	46
La Pila	75.7	0.021	0.055	2.200	45
Aigüèira	102.9	0.030	0.180	0.500	71
Artiga 1	169.1	0.059	0.015	0.050	1
Artiga 2	165.5	0.089	0.010	0.060	1
Gresilhon	60.9	0.032	0.030	0.100	10
Pomero	107.9	0.026	0.040	0.800	1
Tarters	115.0	0.115	0.020	0.100	1
Ruda	118.4	0.056			14
Trop. Ruda	118.4	0.052	0.000	0.150	4
Mulleres	94.3	0.028	0.060	1.000	23
Trop. Mulleres	85.8	0.033	0.000	0.060	2
Pto. Salau	199.4	0.063	0.002	0.100	1
Riu Fred	77.4	0.045	0.030	0.300	1
Escala Alta	176.9	0.100	0.020	0.350	1
Horcalh	120.6	0.074	0.020	0.600	1
Freda	42.2	0.006	0.050	0.060	1
Sarraera 1	61.2	0.013	0.030	0.500	1
Sarraera 2	50.3	0.009	0.010	0.600	1

spatial location of the karstification processes, or dissolution, would be mainly concentrated on the surface portion of the karst and probably also in the topmost area (upper part) of the unsaturated zone. This is perfectly coincident with the fact that the recharge is diffuse and the predominant type of infiltration is slow. This slow infiltration controls the dissolution processes in the superficial portion of the karst and consequently the exokarstic morphology (Bakalowicz 1979, 1981; Mangin 1975). The absence of deep karstification could be explained by the limited development of the rapid infiltration.

On the other hand, the Joèu example has a mainly allochthonous recharge (85% of the basin is impermeable), an annual runoff of around 50 hm^3 and is where the lowest mineralizations are measured (limited efficiency of the CO_2 incorporation). However, this is the most karstified aquifer in the Val d'Aran. The contrast between the area occupied by the impermeable and permeable outcrops and the important development of the superficial drainage network (also characterized by a high altitude gradient, from 3404 to 2000 metres) have been of critical importance for the high karstification potential of the system. The experimental hydrological data indicate a karstification and a drainage structure which are well developed. In fact the important concentration of water (concentration of energy) in some points of the basin (swallow holes of Aigualluts, Renclusa, Hòro, etc.) has determined an important karstification and the existence of a deep drainage structure.

Therefore the CO_2 availability and the efficiency of its incorporation to the water are not important factors as far as karstification in the high mountain calcareous aquifers is concerned. In the case of the Aranese examples it seems clear that the degree of karstification of the aquifers is basically conditioned by the predominant type of infiltration and related to the allochthonous (binary systems) or autochthonous (unary systems) recharge of the aquifer. Thus, the hydrodynamic factor (water potential due to the energy concentration in certain sites) becomes fundamental in the karstification potential of the high mountain aquifers in the Val d'Aran.

Even having concentrated the last discussion on the Joèu and Aigüèira systems (which are those presenting more contrasting features) the experimental data obtained in the Lastoar and Tèrme-Pila karsts fully support what has been stated. Thus, in the Lastoar system with an original structure of the binary type karstification is significant even though it is smaller than the Joèu. In the Tèrme-Pila system, we find that the Pila spring is more closely related to a basin domain with impermeable materials which are drained towards various swallow holes (allochthonous recharge) while the Tèrme spring, besides draining a small part of the water of this binary domain, also drains calcareous materials which cause the autochthonous recharge to attain a significant importance. The experimental data show that the aquifer portion more related to Pila is much more karstified than the Tèrme one.

Conclusions

The experimental data acquired in the Joèu and Aigüèira examples are of particular interest for determining the role of the different factors in karstification potential. Thus, it has been shown that the high karstification potential of the Joèu system comes largely from its binary character (85% of the basin is impermeable) and from the existence of an important network of superficial drainage which is totally drained by the Joèu aquifer. Conversely, the weak karstification potential of the Aigüèira aquifer is related to the unary character of its basin, without any organized superficial drainage and with a basically diffuse recharge. This shows a clear predominance of the slow infiltration which has produced a certain development of superficial karstification (presence of karren forms and dolines) but never in deeper layers.

The data on the Lastoar and Tèrme-Pila systems only corroborate the key role played in the karstification potential by the predominant type of infiltration (hydrodynamic factor or water potential), conditioned by the structure of origin (binary in both cases).

The Val d'Aran aquifers are characterized by low CO_2 concentrations but at the same time they present quite contrasting degrees of karstification. This clearly indicates that the role of CO_2 in the karstification potential of the Aranese aquifers is quite relative and clearly remains of secondary importance.

Finally, in view of all that has been presented, an interesting question arises: is it possible to generalize, in meteoric, (with gravity as the motor of karstification) karsts, that the hydrodynamic factor (water potential) is the determinant one in karstification phenomena? An affirmative answer would only confirm the conclusions achieved by Mangin (1975, 1977, 1984, 1986, 1994) stating that he considers the hydrodynamics as the fundamental aspect in the genesis of a karst. Furthermore, it is obvious that the results achieved in the Val d'Aran lead to a relativization of the role of the climate (bioclimatic factor) in karst development, as other authors have shown (Geze and Mangin 1980; Mangin and Bakalowicz 1990; Bakalowicz 1994).

Resumen

Este trabajo se plantea como una reflexión entorno al concepto de potencial de carstificación y los diferentes factores que permiten definirlo.

La discusión está basada en los resultados obtenidos en los principales acuíferos en rocas calcáreas de la Val d'Aran, en cuanto a caudal, temperatura del agua y variables del sistema calco-carbónico. Estos datos experimentales corresponden a medidas quincenales y mensuales realizadas en cinco manantiales (pertenecientes a cuatro sistemas cársticos) entre 1988 y 1991.

Tanto la hidrodinámica, como la hidrogeotermia, como la hidrogeoquímica ponen de manifiesto diferencias importantes en el grado de carstificación de los acuíferos drenados por cada una de las surgencias. Así pues, los sistemas de Joèu, Lastoar y Tèrme-Pila tienen un grado de carstificación significativo, mientras que el sistema de Aigüèira es, esencialmente, fisural, sin carstificación en profundidad (débil potencial de carstificación). Estas diferencias quedan reflejadas en la clasificación hidrodinámica de Mangin y en la variabilidad observada en los datos de temperatura y mineralización (distribuciones de frecuencias). En cualquier caso, todos los puntos de agua

estudiados presentan mineralizaciones y pCO_2 muy bajas, lo cual está en estrecha relación con el hecho que las respectivas cuencas de alimentación están en dominios de clima alpino y subalpino, con vertientes cubiertas de nieve durante la mayor parte del año, con un desarrollo limitado de suelos bien configurados y, por consiguiente, con poco CO_2 disponible para el agua. En cambio, en lo que sí difieren estos acuíferos es en el tipo de recarga principal: así, se observa que el único sistema con recarga esencialmente autóctona de tipo difuso (sistema monádico) es el de Aigüèira. Los demás presentan una clara componente alóctona (afloramientos impermeables drenados por el acuífero) que les confiere un carácter marcadamente binario, y que lleva a que la infiltración localizada a través de pérdidas tenga un papel muy relevante. En este sentido, el caso extremo es el del sistema de Joèu, cuya cuenca tiene un 85 % de superficie impermeable.

Así pues, se pone de manifiesto que la limitada carstificación del acuífero de Aigüèira tiene que estar en relación con un claro predominio de la infiltración lenta, mientras que el elevado potencial observado en Joèu tiene que proceder del importante papel de la infiltración rápida a través de pérdidas.

Por consiguiente, parece claro que, al menos en los acuíferos de montaña de la Val d'Aran, el papel del CO_2 (factor bioclimático) en el potencial de carstificación es muy relativo y que es el factor hidrodinámico (potencial de agua) el que realmente tiene una importancia capital.

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Late Triassic carbonate platform evolution and related early diagenesis and paleokarst phenomena in the Transdanubian Range

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Relationships of the depositional cycles and the early diagenetic and paleokarstic features were studied in the Upper Triassic platform carbonates of the Transdanubian Range. The studies revealed that the nature and intensity of the diagenetic and karstic processes depended mainly upon the climatic conditions and duration of the subaerial episodes punctuating the shallow marine carbonate deposition. During the drier early stage of the platform evolution dolomitization and dolocrete formation prevailed, while in the more humid late stage, microkarstic solution cavities and argillaceous paleosols were formed.

Key words: carbonate sedimentology, depositional cycles, early diagenesis, dolomitization, paleokarst, paleosol, paleoclimate, Upper Triassic, Transdanubian Range

Introduction

During the Middle and Late Triassic large carbonate platforms were formed on the rapidly subsiding passive margin of the Tethys where thick sequences were accumulated. In the wide inner platform area the carbonate sequences are made up of meter-scale peritidal–lagoonal cycles (Lofer cycles – Fischer 1964) as a rule. The cycles are generally bounded by unconformity surfaces, indicating periods of subaerial exposure between deposition of the cycles. During subaerial exposure, in addition to desiccation, initial compaction and cementation of the previously deposited carbonates, under suitable climatic conditions their dissolution may also have commenced, resulting in an early karstification.

Formation of Lofer cycles are generally attributed to orbitally forced sea-level changes (Fischer 1964; Schwarzacher and Haas 1986) although tectonics and other factors may have modified the pattern of the cyclicity. Since in the coastal zone sea-level changes control the position of the groundwater table and through it also processes of dissolution and cementation, a close relationship between cyclicity and early karstification is plausible.

The Upper Triassic Dachstein platform is a classic example of the cyclic platform carbonates providing an excellent opportunity to study the effects of

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sea-level changes both on the carbonate accumulation and on dissolution, i.e. the depositional cycles and the early karstification.

During the very long, approximately 16 Ma history of the platform evolution significant changes took place in the tectonic regime and also in the climate which may have influenced the deposition pattern, the early diagenesis as well as the karstification.

Geologic setting

Location of the study area, and extension and outcrops of the Upper Triassic platform carbonates are presented in Figure 1.

The lithostratigraphic units building up the Transdanubian Range's Upper Triassic carbonate platform are shown in Figure 2. The characterization of the main units of large areal extension are given below. The formations of the smaller intraplatform basins (Feketehegy Fm, Mátyáshegy Fm, Csővár Fm) are not discussed.

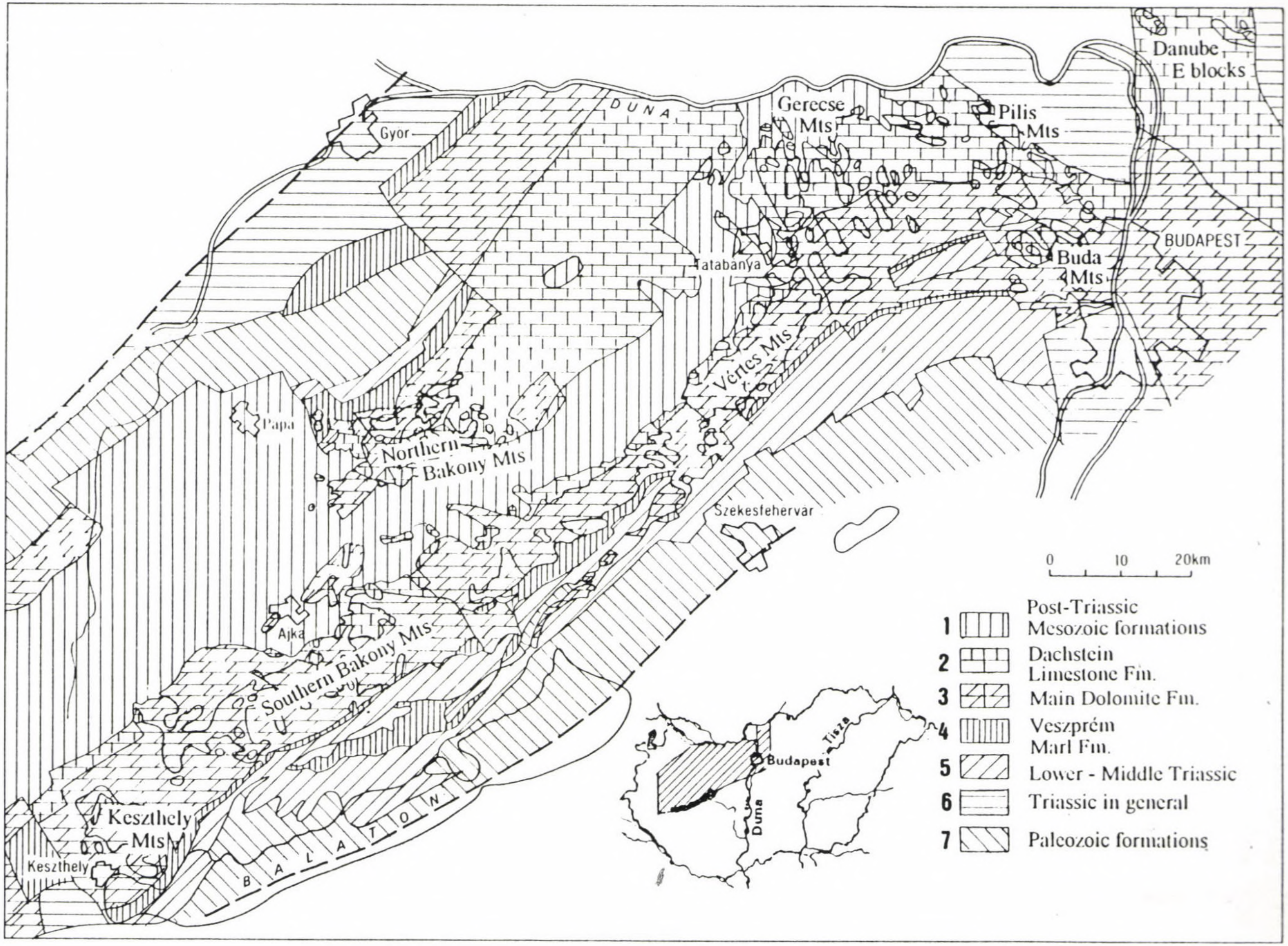
Main Dolomite (Fődolomit) Formation

The Main Dolomite (Fődolomit – Hauptdolomit – Dolomia Principale) consists almost exclusively of dolomite. It is rarely bituminous, except in the westernmost part of the Transdanubian Range. Typical colours of the rocks are pale yellowish, brown, or grey, rarely dark grey.

Meter-scale cyclicity is common (Lofer cycles – Fischer 1964). The cycles generally consist of two rock types, a microlaminated–stromatolitic peritidal facies (member B) and a thick-bedded subtidal facies (member C). Authigenic brecciation is common in both lithofacies. The cycles are rarely bounded by very thin red clayey paleosol or more frequently caliche (dolocrete) horizons (Fig. 3).

The primary, structural and textural characters often disappeared completely due to recrystallization accompanying dolomitization. Thus the most discernible rock texture today is the finely crystallized dolosparite (xeno-morph-A – saccharoidal texture). In several cases some part of the original, textural elements is still recognisable (e.g. bioclast element, the outlines of ooids and pellets). Solution pores commonly mimic their original textural elements (e.g. tube-like pores after dasycladacean algae; pores after intraclasts – Fig. 4). In the original texture of the C-member mud-supported types are characteristic. The dolomicrite (or rather microsparite) and the dolopelmicrite (microsparite)

Fig. 1 →
Extension of the Upper Triassic platform carbonates in the Transdanubian Range. 1. Mesozoic formation younger than Triassic; 2. Dachstein Limestone a) outcrops, b) subsurface; 3. Main Dolomite a) outcrops, b) subsurface; 4. Veszprém Marl Formation (Carnian); 5. Triassic formations older than Carnian; 6. Triassic in general; 7. Paleozoic formations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1 Post-Triassic Mesozoic formations
2 Dachstein Limestone Fm.
3 Main Dolomite Fm.
4 Veszprém Marl Fm.
5 Lower - Middle Triassic
6 Triassic in general
7 Paleozoic formations
Budapest
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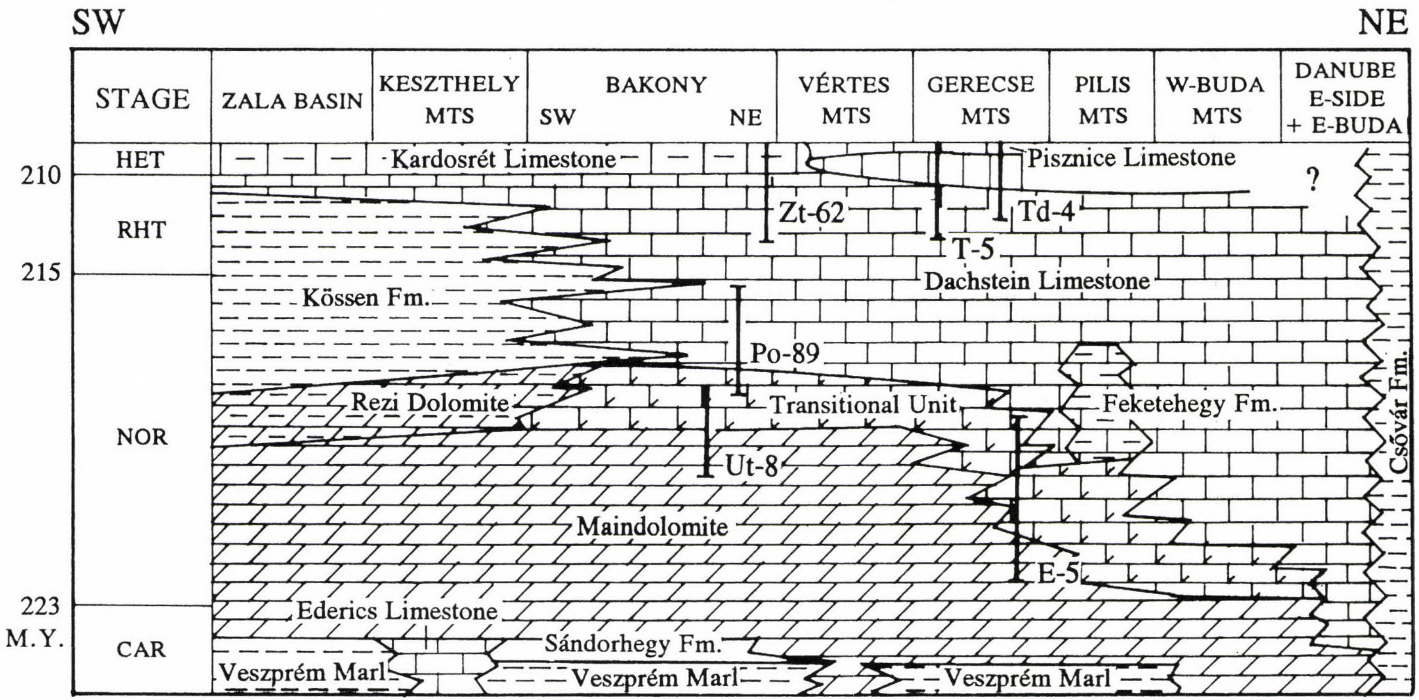

Fig. 2
Stratigraphic scheme showing space-time relationships of the Upper Triassic platform carbonates along a NE–SW profile through the Transdanubian Range. Stratigraphic position of the most important core-sections are also indicated

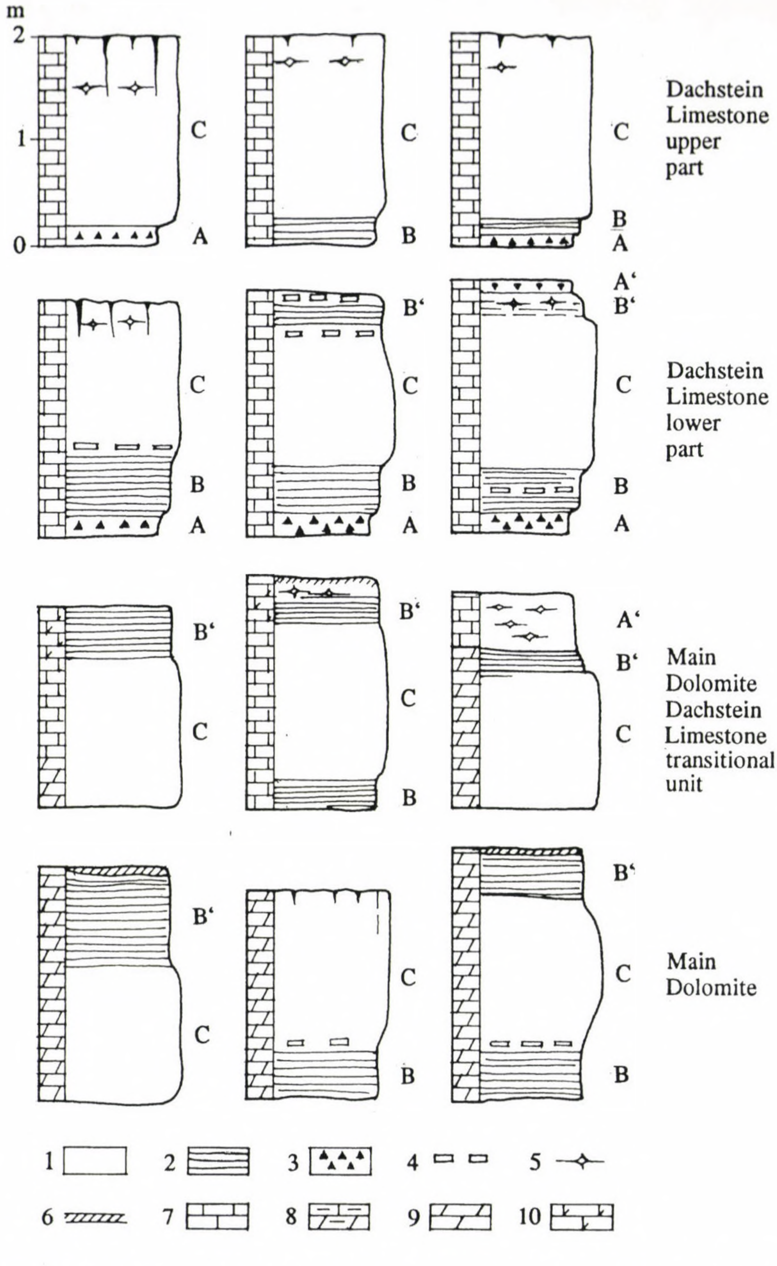

Fig. 3

Typical stacking patterns of the Upper Triassic cyclic carbonates in the Transdanubian Range. 1. subtidal facies; 2. laminitic peritidal facies; 3. black pebbles; 4. rip-ups; 5. sheet crack; 6. caliche; 7. limestone; 8. argillaceous limestone; 9. dolomite; 10. dolomitic limestone often occur and are poorly fossiliferous. Stromatolitic boundstones and caliche crusts are typical in the B-member (Fig. 5).

Among the macrofossils, the megalodont bivalves are the most characteristic (Végh-Neubrandt 1982). They appear in large numbers in the subtidal facies together with other bivalves and gastropods. The most characteristic elements of the microfossil assemblage are ostracods and benthic foraminifers. Crustacean coprolites are common. In certain rock types massive occurrence of dasycladacean algae is characteristic.

As to early diagenetic processes in addition to pervasive dolomitization, brecciation, formation of shrinkage cracks, and mm to cm-size solution pores can be mentioned. The thin paleosol layers are generally dolocretes but kaolinite-rich vertisol- like layers also occur, but rarely. These diagenetic and weathering phenomena are mainly related to high-frequency sea-level changes, causing recurrent subaerial conditions (fourth and fifth order discontinuities – Esteban 1988).

The Main Dolomite is widely extended in the Transdanubian Range. It can be followed from the Buda Mountains as far as the Keszthely Mountains in the westernmost part of the range. The thickness of the formation is 600–1200 m.

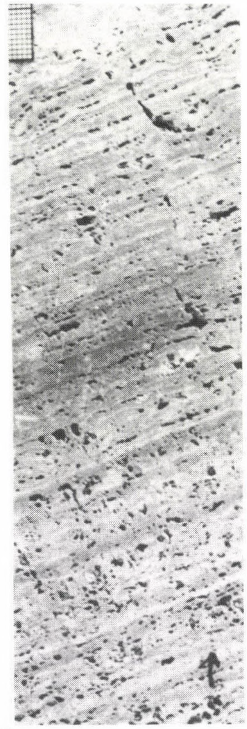

Fig. 5
Member B in the Main Dolomite with fenestral pores. Core Szár-1, 31.1–31.2 m

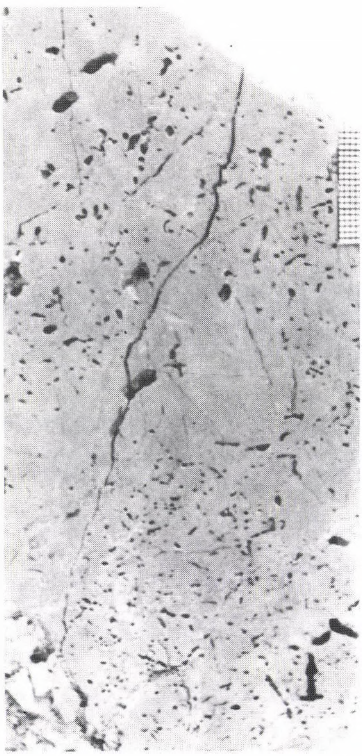

Fig. 4
Member C in the Main Dolomite with biomoldic pores. Core Szár-1, 53.4–53.5 m

Chronostratigraphically the formation begins in the Upper Carnian (Tuvalian) and extends up into different levels of the Norian. According to Oravecz (1963) the overlying formations become progressively younger from NE to SW within the Transdanubian Range. According to our present-day knowledge, however, the upper boundary of the Main Dolomite can be drawn within the Middle Norian in a large part of the Transdanubian Range. In its southwestern part the Main Dolomite is covered by the Rezi Formation of latest Middle Norian age.

Dachstein Limestone Formation

The formation consists mainly of limestones. However, in its lowermost member (transitional between the Main Dolomite and Dachstein Limestone) dolomitized intervals and partially dolomitized rock types are common. Through most of the formation meter-scale (Lofer) cyclicity is characteristic (Figs 6, 7). Thin greyish or pinkish marl or clayey limestone layers appear at the base of the cycles (member A). These in situ or reworked paleosol layers of a few dm thickness frequently contain authigenic breccias or black pebbles (Figs 10, 11). They are followed by mostly microlaminated (stromatolitic) dolomitic limestones (member B). The thick-bedded member C is generally made up by grey limestone. In the lower part of the formation the C beds are partially dolomitized as a rule. This interval was classified as a transitional member between the Main Dolomite and the Dachstein Limestone.

In member A the most common textural types are argillaceous mudstones, (pelmicrite, pelbiomicrite) and skeletal or intraclastic wackestones and packstones (Figs 10, 11, 12). In member B peloidal, fenestral laminites prevail (Figs 13, 14). In member C mudstones, peloidal wackestones and packstones, and skeletal, ooidic, and oncoidal packstones and grainstones are common.

The most characteristic fossils are the megalodonts (Végh-Neubrandt 1982); in certain layers they are buried in life position forming banks composed of large numbers of individuals, in others they appear in storm horizons and elsewhere only fragments are present. Other bivalves and gastropods (*Pleuromya*, *Worthenia*, and *Purporoidea* types) are relatively common and corals forming small patch reefs also occur. Algae, ostracods, foraminifera, and crustacean coprolites play an important role in the microbiofacies. The foraminifer assemblages are good indicators of environmental conditions (Oravecz-Scheffer 1987).

Among the early diagenetic phenomena it is worth mentioning the mm to m-scale solution pores and cavities. They are filled with phreatic isopachous calcite, laminar intrasediments and/or breccias. Shrinkage pores and cracks and multi-phase microfissures with marine and terrigenous infillings are also common (Haas and Balog 1995). Paleosols are generally kaolinite-rich. However, gibbsite and phosphate-bearing weathering crusts also occur locally (Korpás 1980). Thin peloidal, laminated calcrete-crusts, very similar to those described by Wright (1983) as xero-rendzina, were also observed.

The small-scale, bedding plain-parallel early paleokarst features (pores and cavities) were formed as a consequence of high-frequency sea-level oscillation (fourth and fifth order). The major paleosol and paleokarst horizons are probably related to the third order sea-level lowstands.

The 600–900 m-thick Dachstein Formation is widespread in the Middle–Upper Norian and Rhaetian of the Transdanubian Range. Its deposition generally began in the Middle Norian (in the easternmost part of the Transdanubian Range probably in the Early Norian) and continued until the Late Rhaetian. However, in the western part of the Transdanubian Range a large part of the Dachstein Limestone is substituted by the Kössen Formation and only the topmost part appears above the Kössen Formation (Haas 1993).

Fig. 6
Meter-scale cyclicity in the Dachstein Limestone. Dorog, Kis Kőszikla quarry (Gerecse Mts)

Fig. 7
Meter-scale cyclicity in the Dachstein Limestone. Epöl quarry (Gerecse Mts)

Rezi Dolomite Formation

The Rezi Formation is made up predominantly of dark grey or brownish-grey, bituminous dolomites. In the lower and locally in the upper part chert lenses and interbeds are discernible.

The lower part resting upon the Main Dolomite is thinly laminated and platy with microlamination discernible within some layers. In certain horizons authigenic breccias and slumps occur. Thick bedding is characteristic of the middle part with common thin-bedded and laminated interbeds. In the uppermost part the platy, a thin-bedded structure is again characteristic, with slumps, mollusc coquina lenses and limestone beds.

The lower part is poor in macrofossils; fish scales and sponge spicules were obtained from acid residues. Pores characteristic of the middle part of the formation were probably formed by the solution of the calcareous skeletons of unidentified green algae. Bivalve, gastropod and brachiopod moulds occur sporadically. The upper part is very rich in fossils: bivalves (*Modiola, Pteria, Isognomon, Lima, Entolium* and *Cardita* types) and gastropods (*Worthenia, Euomphalus, Coelostylina, Zygopleura* and *Promathildia* types) commonly occur in lenses or in beds as lumachelle.

The Rezi Dolomite Formation is known from the southeastern part of the Transdanubian Range, in the western part of the Southern Bakony, and in the surroundings of the Keszthely Mountains (Budai and Koloszár 1987; Haas 1993). Its thickness is 200–250 m.

Fig. 8
Uneven unconformity surface of a Lofer cycle, and basal part of the overlying cycle. Note intraclast and paleosol accumulation at the very base of the cycle. Tardosbánya, Gorba quarry (Gerecse Mts)

Fig. 9
Irregular unconformity surface and thin intraclastic A-member at the base of the overlying cycle. Lábatlan, Kecskekő quarry (Gerecse Mts)

On the basis of conodonts from the lower member in the Keszthely Mts the base of the formation can be assigned to the upper part of the Middle Norian or to the lower part of the Upper Norian (Budai and Kovács 1986).

Kössen Formation

The Kössen Formation consists of dark grey marl, limy marl, dolomitic marl or silty marl, locally with limestone and dolomite interbeds. Monotonous dark shales rich in organic material prevail in the internal parts of the basin. Toward the basin margins, dolomitic limestone, clayey limestone, marl and limy marl layers alternate in a cyclical fashion while the proportion of argillaceous layers gradually decreases.

Microlamination is common in the marls which are characterized by clayey and silty mudstone, bioclastic wackestone or peloidal wackestone texture. Coquina lenses and thin layers may be interbedded within the marl. The characteristic textural types of the limestones and argillaceous limestones are mudstone and skeletal, peloidal and intraclastic wackestone.

Among the macrofossils bivalves are the most characteristic and are commonly accumulated in lumachelles.

The main development of the Kössen Formation is in the western part of the Transdanubian Range (Zala Basin) where its thickness may exceed 400 m. It thins towards the east where it is interbedded with the Dachstein Limestone (Haas 1993).

On the basis of macrofossils and microfossils the deposition of the Kössen Formation began in the Late Norian and continued until the Middle Rhaetian (Oravecz and Haas 1985; Oravecz-Scheffer 1987; Haas 1993).

Cyclicity

Cyclicity of meter or tens of meter-scale is characteristic for every formation mentioned above. The meter-scale Lofer cycles were first described by Fischer (1964) from the East Alpine Dachstein Limestone. The basic pattern of the Lofer cycles in the Transdanubian Range was discussed by Haas (1991). According to his studies the ideal Lofer cycle is a fairly symmetric transgression-regression cycle (Fig. 13) but in the reality any of the members of the ideal cycle may be missing as consequences of syndepositional or early postdepositional processes. The mean thickness of the cycles is about 2 m; the cyclicity was explained by orbital forcing. The basic cycles should have been controlled by the precession with a periodicity of about 20 ka (Schwarzacher and Haas 1986).

Fig. 10
Typical texture of member A. In reddish argillaceous limestone matrix large light-coloured rip-ups of algal mat origin and small black pebbles. Core Porva Po-89, 488.4–488.5 m (Bakony Mts)

Statistical analysis of the cycles has also revealed that in the different formations (i.e. lithogenic units) cycles of different stacking pattern are characteristic (Haas 1991).

In the Main Dolomite between the fairly flat discontinuity surfaces the intertidal B and the subtidal C beds alternate – BC, BCB′, or CB′ cycle types are characteirstic as shown in Fig. 4. The B beds are relatively thick (0.4–1.2 m) as a rule.

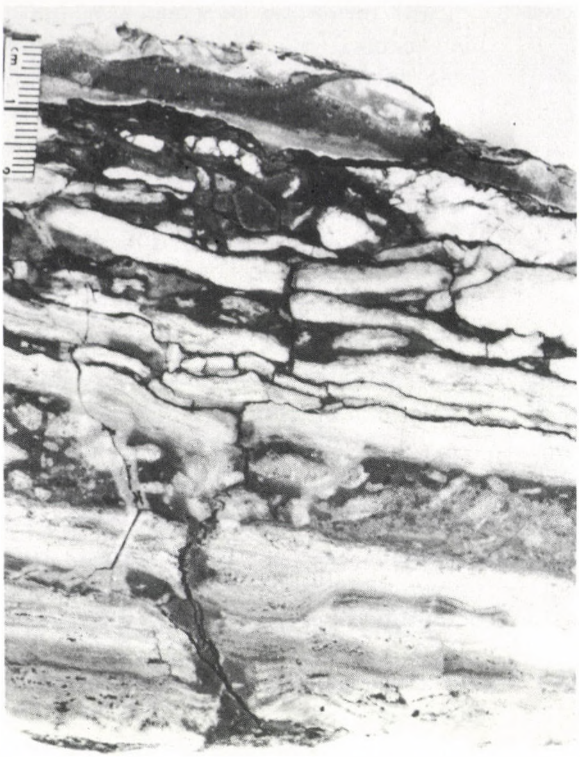

Fig. 11
Member A at the base, covered by algal mat facies of member B. Note large, only slightly reworked flat pebbles in the basal layer. Core Porva Po-89, 254.8–254.9 m (Bakony Mts)

In the lower (Norian) part of the Dachstein Limestone paleosol interlayers (member A) are common but the intertidal B member is frequently missing. Consequently the AC stacking pattern is predominant. Cycles of ABC and ABCB′ and ABCB′A′ stacking pattern are also common (Fig. 4).

In the upper (Rhaetian) part of the Dachstein Limestone the cycles are generally truncated, consequently the AC, BC, and ABC successions are characteristic, and the A, and B members are thin (1–2 dm) as a rule (Fig. 4).

In the transitional unit between the Main Dolomite and the Dachstein Limestone the paleosol layers are generally missing in the dolomitized segments but appear in the less dolomitized intervals.

It is important to note that in the northeastern part of the Transdanubian Range in the paleogeographical zone of the outer platform margin the Dachstein Limestone shows an oncoidal development and in this zone supratidal and intertidal beds are rare. Consequently the cyclicity is faint or practically missing.

Fischer-plots of Dachstein Limestone sequences in the Transdanubian Range suggested significant third order sea-level changes for the Late Norian–Rhaetian interval (Balog et al. 1997). Third order lowstands are also indicated by anomalously thick composite paleosol horizons.

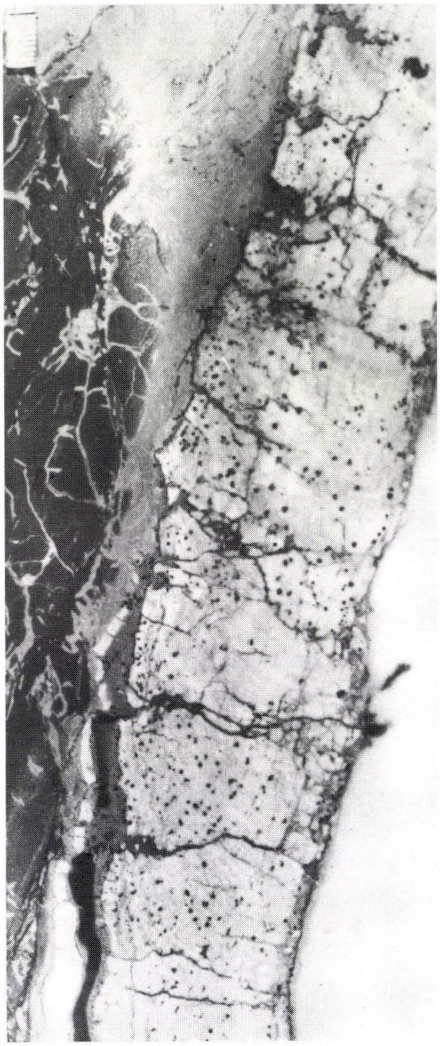

Fig. 12
Large fissure in the Dachstein Limestone, with two generations of fissure-fills. White host-rock is lined by a yellowish limestone layer and the inner part of the fissure is filled by red carbonate silt. Note irregular, dissolved interfaces of the various fissure fills. Core Porva Po-89, 433.6–433.9 m (Bakony Mts)

Third order sea-level changes may have caused the development of the three members distinguished within the Rezi Dolomite (Haas 1993). Orbitally-forced fourth order oscillations may be reflected in the cyclicity of the Kössen Formation in the transition belt between the basinal facies of the Kössen Formation and the platform facies of the Dachstein Limestone (Haas 1993).

Platform evolution

Based on time and space relationships of the Upper Triassic formations the trends of facies changes and regularities in the cyclicity of the sequence evolution of the platform may be outlined.

Infilling of deep intraplatform basins in the Carnian may be considered as a direct precursor of the Late Triassic carbonate platform evolution in the Transdanubian Range. This means that by the Late Carnian the former dissected shelf became levelled, practically flat. On the other hand, the cessation of terrigenous input, probably as a result of the more arid climatic conditions, made possible the massive shallow-marine biogenic carbonate accumulation.

Within the long Late Carnian to Early Jurassic evolution of the platform the following stages can be distinguished.

1. In the Late Carnian a very shallow lagoon was formed on the levelled shelf. The lagoon was probably more or less separated from the open sea by patch reefs and/or ooidic-oncoidal mounds. Facies indicating this setting are known in the Buda Mountains and in a few small basement blocks on the east side of the Danube. Due to high- frequency sea-level oscillations (they

were most probably controlled mainly by 20 ka precession cycles) the larger part of the shelf lagoon periodically entered into the peritidal zone and became subaerially exposed. During the subaerial periods, under the given warm and dry (semiarid) climatic conditions caliche crusts were formed on the surface and the formerly deposited subtidal lime mud was pervasively dolomitized (see Fig. 14).

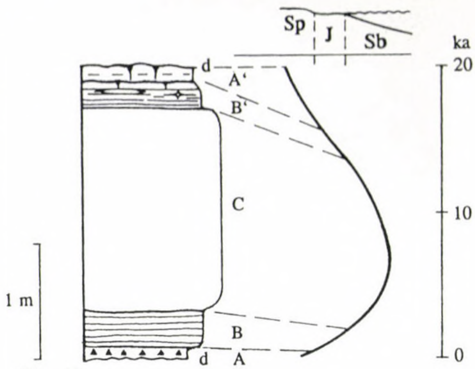

Fig. 13
The ideal Lofer cycle. d – disconformity surface, A – pink or green argillaceous, brecciated limestone or marl (reworked paleosol); B – transgressive peritidal laminites; C – subtidal skeletal carbonates (wackestones–packstones); B' – regressive peritidal laminites, A' – weathering crust (paleosol); Sp – supratidal zone; I – intertidal zone; Sb – subtidal zone

2. At the end of the Middle Norian extensional tectonics led to formation of a new basin (Kössen basin) in the external (landward) part of the platform. In the basin the Rezi Dolomite began to form. Subsequently, in the early Late Norian, most probably as a consequence of increasing humidity, carbonate sedimentation was replaced by the deposition of fine terrigenous sediments of the Kössen Formation (see Fig. 2). This climatic change was also reflected in the early diagenesis of the carbonates of the Dachstein platform. Skeletal wackestones and packstones which were deposited in the inner part of the platform during the highstand intervals of the high-frequency sea-level cycles were not subjected to dolomitization during the lowstand subaerial intervals while the processes of "soilification" (subaerial weathering) became characteristic although the thin soil layers were frequently eroded or redeposited during the subsequent transgressions (Fig. 15).

3. In the latest Rhaetian the tectonic disruption of the carbonate platform was initiated resulting in the disintegration of the platform the northeastern part of the Transdanubian Range, located relatively close to the outer platform margin. Here the topmost Rhaetian and the lowest Jurassic is missing and the platform carbonates (Dachstein Limestone) are covered by pinkish subtidal, open shelf carbonates of Late Hettangian age.

In the inner platform the aggradation of the Dachstein platform continued till the end of the Triassic without any significant facies change.

In the southwestern part of the Transdanubian Range the filling of the basin (Kössen Basin) was completed by the latest Rhaetian making possible the rapid basinward progradation of the Dachstein platform.

In both of the latter areas carbonate platform sedimentation continued even in the earliest Jurassic although the peritidal cycles are missing in this interval. Disintegration of the platform was initiated here only in the Early Sinemurian.

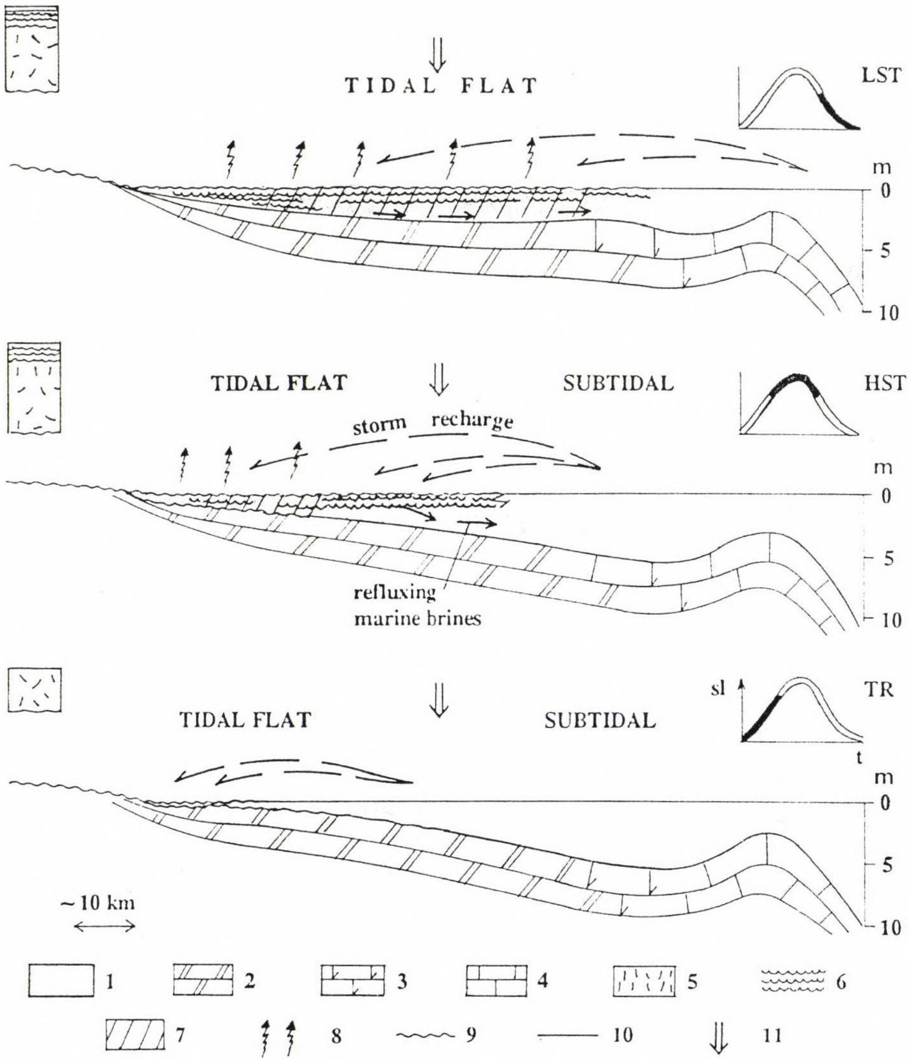

Fig. 14
Effects of high-frequency sea-level oscillation on the sedimentation pattern and early diagenesis during formation of the Main Dolomite. 1. unconsolidated carbonate sediments; 2. dolomite (dolomitized sediments); 3. dolomitic limestone; 4. limestone; 5. skeletal mud; 6. microbial mat; 7. dolomitization; 8. evaporation; 9. unconformity surface; 10. caliche crust; 11. location of the log (left on the figure)

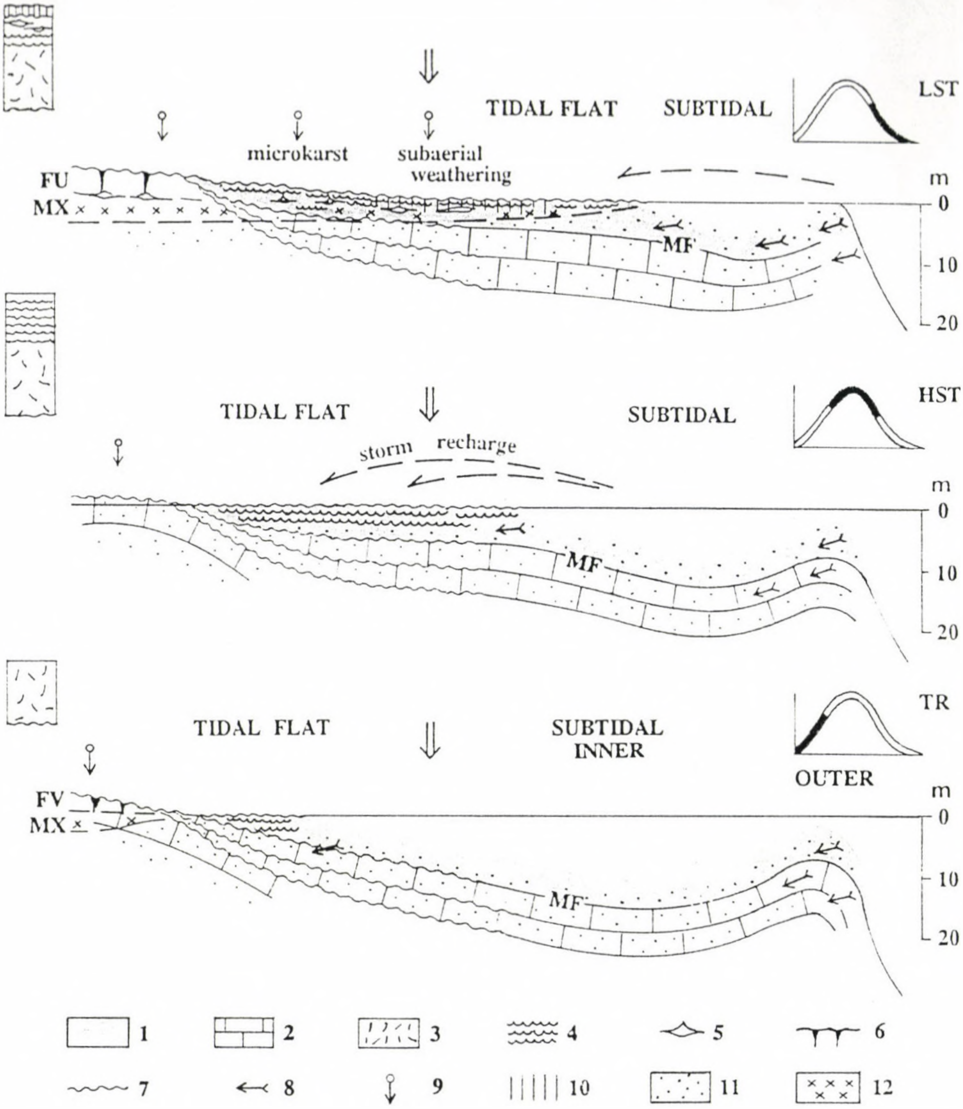

Fig. 15
Effects of high-frequency sea-level oscillation to the sedimentation pattern and early diagenesis during formation of the Dachstein Limestone. 1. unconsolidated carbonate sediments; 2. limestone; 3. skeletal mud; 4. algal mat; 5. solution cavities, sheet cracks; 6. vertical microkarstic solution cavities; 7. erosion surface; 8. infiltration of marine water; 9. meteoric recharge; 10. pedogenesis; 11. pores filled by marine water; 12. pores filled by mixed water

Fig. 16
Large dissolution cavity filled by generations of isopachous calcite and red carbonate mud internal sediment. Core Porva Po-89, 199.5–199.7 m (Bakony Mts)

Early diagenesis and paleokarst phenomena

The early diagenetic features and paleokarst phenomena are genetically related to the cyclicity of the platform carbonates (Main Dolomite, Dachstein Limestone) and probably also to syngenetic tectonic movements. Both of these processes (i.e. early diagenesis and karstification) were governed by some common factors: climatic changes, eustatic sea-level changes and to a certain extent syndepositinal (mainly extensional) tectonics. As a joint effect of these factors submarine sediment accumulation was punctuated by subaerial episodes resulting in fourth and fifth order discontinuity surfaces with the corresponding mm to cm-scale microkarst features. In connection with third order disconformities larger-scale paleokarst horizons and open joint systems may have formed under suitable climatic conditions. Duration of paleo- karst formation periods in the case of the fourth and fifth order discontinuities may have been a few ka to tens of ka, while in the case of the third order disconformities a few 100 ka interval can be assumed.

In close connection with cyclic and subaerially punctuated carbonate accumulation paleokarst development was also cyclic. Development of microkarst phenomena began just after deposition of a Lofer cycle as a rule and may have resumed during the subsequent subaerial intervals, overprinting the earlier karstic forms. A progressive karst evolution occurred which may also have been influenced by the extensional tectonics.

Significant differences in the climate between the early and late stages of the platform evolution are clearly reflected in the early diagenetic and karstic phenomena. In the early stage of the platform evolution (Late Carnian–Early-Middle Norian) under arid climatic conditions during periods of falling sea-level marine or mixed meteoric-marine waters can reflux through the underlying cycle (cycles) causing massive dolomitization (McKenzie et al. 1980; Read and Horbury 1993), and caliche crusts were formed on the subaerially exposed part of the tidal flat.

Fig. 17
Cm-size cavities filled by isopachous calcite. Core Porva Po-89, 120.2–120.4 m (Bakony Mts)

Due to increasing humidity since the early part of the Late Norian a meteoric groundwater system may have formed under the widely-extended tidal flat of the Dachstein platform in the lowstand periods of the short-term cycles. Seaward migration of groundwater and circulation of sea water in the loose skeletal carbonate sediments resulted in the formation of a broad mixing zone (Read and Horbury 1993). Just after deposition of a cycle, during the next emergence, dissolution of aragonitic shells should have occurred causing moldic porosity. It was followed by the formation of larger (cm to dm-size) horizontal cavities (Figs 16, 17). Since maximum dissolution occurs at the groundwater table and the larger cavities are located 10 to 50 cm beneath the unconformity surfaces as a rule, the depth of the groundwater table should not have been more than a few tens of centimeters below the paleo- surface. Dissolution cavities are filled or partially filled by internal sediments of fine

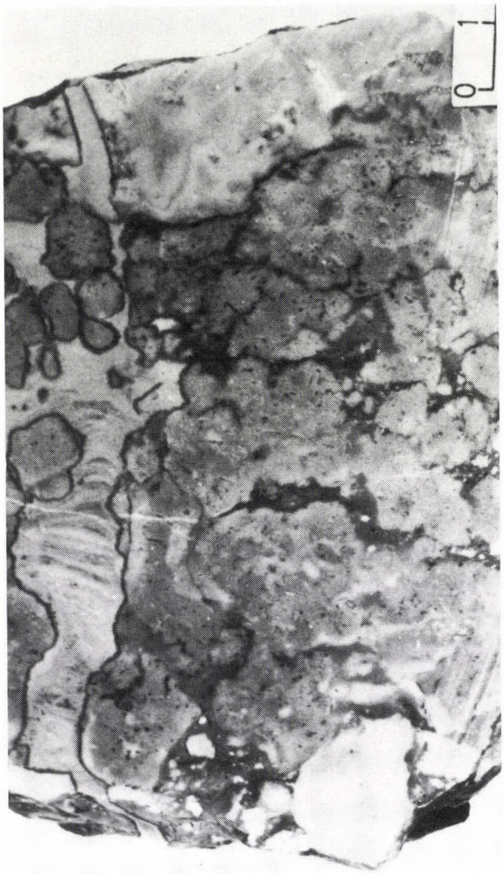

Fig. 18
Internal sediment and collapse-breccia cavity fill. Core Tata T-5, 24.9–25.0 m (Gerecse Mts)

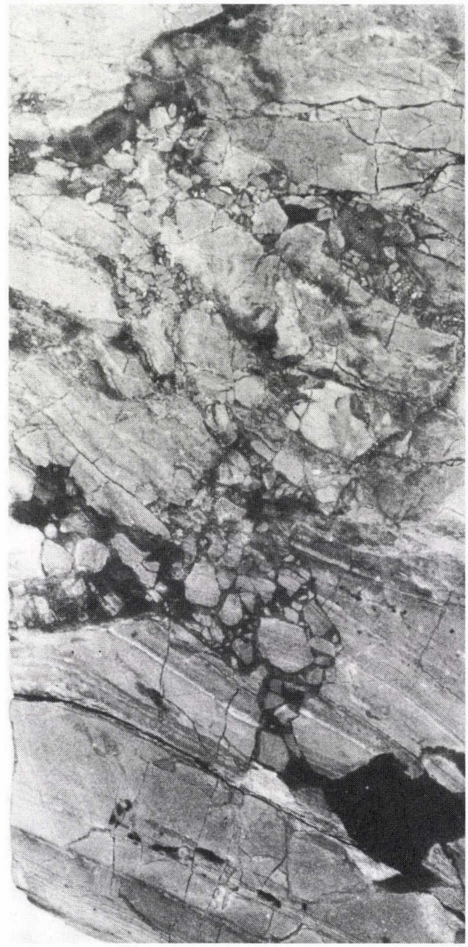

Fig. 19
Collapse-breccia as cavity fill. Core Porva Po-89., 173.0–173.2 m (Bakony Mts)

terrestrial material (paleosol) and collapse breccia (Figs 18, 19) or marine carbonate sediments (storm-transported carbonate mud). Geopetal cavity fills are common. In this case remnant pore spaces are lined with marine cements forming in the subsequent transgressive phases.

Within the Main Dolomite 2–3, and within the Dachstein Limestone 3–5 generally bedding plane- controlled major cave horizons occur which may have formed during the long-lasting subaerial exposure intervals in the convergent stage of the Alpine structural evolution from the Middle Cretaceous onward.

Conclusions

1. The early diagenetic and karstic processes were in close genetic relationship with cyclic evolution of the Late Triassic carbonate platforms in the study area. They were related to the periodically repeating subaerial exposure episodes.

2. Nature and intensity of the processes were controlled mainly by the climatic conditions and the duration of the subaerial intervals. However, they might also have been influenced by the syngenetic tectonic activity.

3. Early dolomitization characterises the early stage of the evolution of the Dachstein platform during the arid Late Carnian to Middle Norian interval. Increasing humidity in the Late Norian led to a basic change in the pattern of the diagenesis. It resulted in the cessation of the intense dolomitization and the formation of argillaceous paleo- sols and microkarstic cavities in the Late Norian–Rhaetian interval.

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Seismic stratigraphy of the Late Miocene sequence beneath Lake Balaton, Pannonian Basin, Hungary

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This study is based on the interpretation of about 190 km of high-resolution, single-channel reflection seismic profiles acquired on Lake Balaton during the GMS93-02 cruise in June of 1993. The seismic survey investigated the lake subsurface at a maximum depth of about 200 msec down to the acoustic basement, which is represented by Sarmatian (Serravallian) strata. Seismic interpretation showed that the Balaton Quaternary deposits rest above a subhorizontal unconformity which dramatically truncates the underlying Pannonian (Late Miocene) sequence. This unconformity marks a major stratigraphic gap that practically encompasses the whole Pliocene and most of the Pleistocene. The Pannonian sequence in turn unconformably overlies Sarmatian limestone and marl at a depth of 50 to 200 m. Sarmatian strata are truncated at the top by a mature (polycyclic) erosional surface that marks a stratigraphic gap spanning a few million years (about 12 to 9 Ma) and includes the amalgamation of two 3rd-order (with 10^6 year periodicities) sequence boundaries. These are the top of the Sarmatian sequence (Sar-1 SB) and the top of the lowermost Pannonian 3rd-order sequence (Pan-1 SB). Incised valleys with erosional terraces and channel fills are also recognized from the seismic record.

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The lowermost part of the Pannonian sequence directly overlying the Sarmatian "basement" is represented by the Szák Formation which is interpreted as a transgressive systems tract (TST). The maximum flooding surface (mfs-2) at the top of this TST corresponds to the top of the *Congeria czjzeki* (top of Lower Pannonian s. Lőrenthey, 1900) and can be dated at 9.0 Ma by calibration with Iharosberény-I well magnetostratigraphy. Overlying them are the Somló Formation and the lower part of the *Tihany* Formation which are interpreted as a highstand systems tract (HST). The architecture of the HST deposits shows typical forestepping strata which downlap onto the underlying TST deposits and are accompanied by local development of small coarse-grained prograding deltas. A subtle 3rd-order sequence boundary (Pan-2 SB) was detected within the Tihany formation, the upper part of which corresponds to the falling stage systems tract (FSST)/ lowstand systems tract (LST) of the next sequence. Pan-2 SB (ca 8.6 Ma) is related to a significant lowering of the base level of erosion in the Pannonian Lake and possibly associated with tectonic and volcanic activity. It locally crops out at the top of the Tihany Peninsula and its occurrence slightly predates the onset of the basaltic eruption of the Tihany Volcano (ca 7.8 Ma). According to our interpretation it can be concluded that Pontian s.str. (younger than 7.0 Ma) strata are practically absent in outcrop in the area of Lake Balaton.

The Pannonian sequence in the study area is affected by intense tectonic deformation. Tectonic activity mostly postdates the Late Miocene and consists of SW–NE strike-slip faulting associated with gentle folding. Evidence of minor tectonic activity exists between mfs-2 and Pan-2 SB (9.0 to 8.6 Ma). This may suggest causal relations between tectonic and coeval volcanic activity in the area of Lake Balaton and offers an alternative explanation (other than purely climatic) to account for 3rd-order changes of base level of erosion within the Pannonian Lake. Significant tilting of the entire Pannonian sequence toward the SE is also seen on seismic profiles that consistently show ESE-dipping strata which are erosionally truncated at the top. This supports earlier interpretation which proposed that the Pannonian Basin underwent significant late-stage tectonic inversion and uplift with consequent subaerial erosion of Latest Neogene strata in Transdanubia and Northern Hungary.

Key words: Lake Balaton, Pannonian Basin, high-resolution seismics, sequence stratigraphy, Late Miocene, Middle Pannonian

1. Introduction

During June of 1993, a geophysical survey was carried out over Lake Balaton, Hungary (cruise GMS93-02). The aim of the cruise was to collect high-resolution single-channel seismic reflection data from the Pannonian (Pannonian s.l.) sequence beneath the Balaton Quaternary deposits.

The idea that classic "marine-type" seismic data could be successfully acquired over Lake Balaton in order to get information on thickness and seismic characteristics of the Quaternary lacustrine deposits is not new (Cserny and Corrada, 1989). What makes the case of Lake Balaton somewhat exceptional, however, is the fairly low acoustic impedance contrast between the lacustrine mud and water-saturated deeper layers which facilitates the propagation of acoustic waves at depth. A further justification for applying high-resolution seismics is provided by the geographic position of the lake itself. The Balaton is located at the southern edge of the Bakony Mountains, a major mountain range in the western intra-Carpathian region (Fig. 1) where the Paleozoic–Mesozoic basement crops out and the Pannonian sequence pinches out toward NNW. Consequently the late Neogene stratigraphic record, which in the deepest

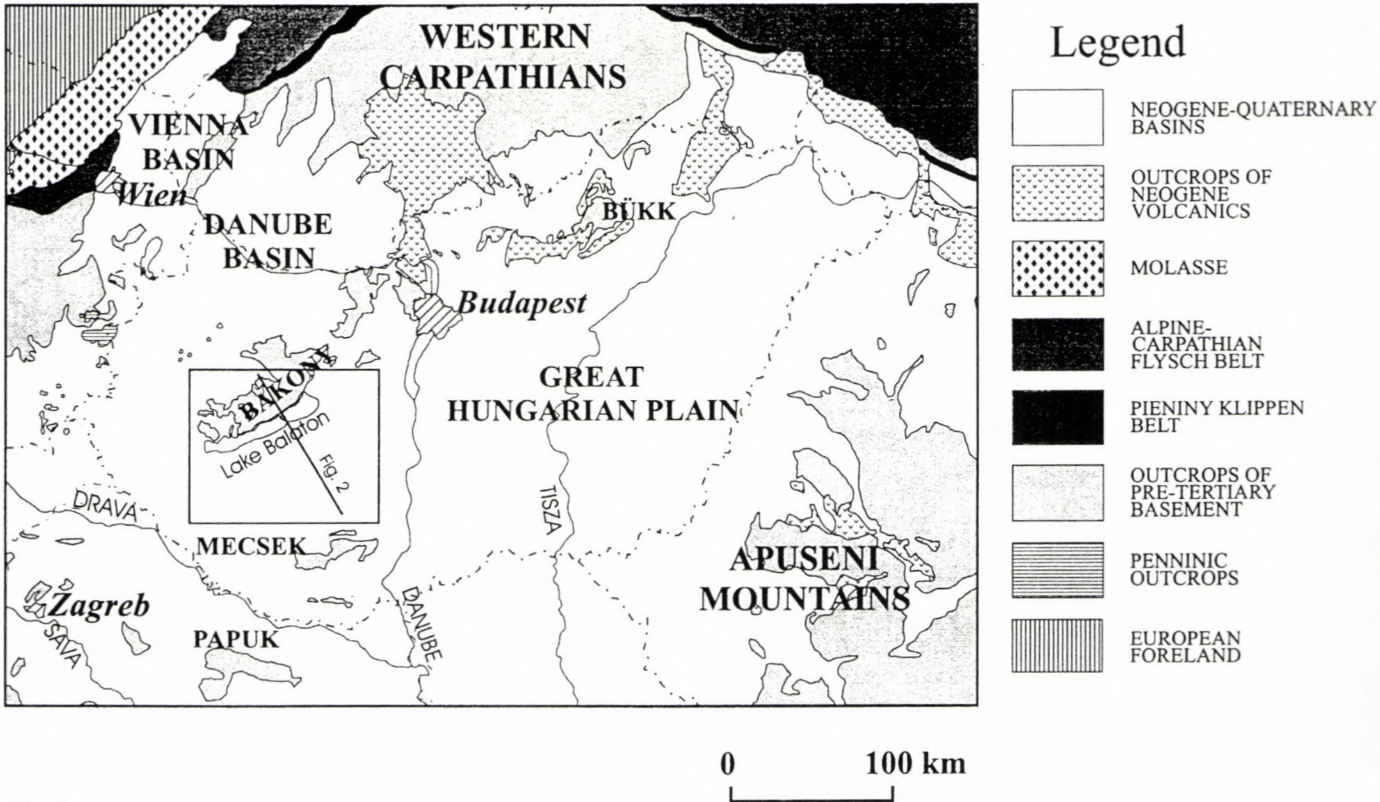

Fig. 1
Tectonic sketch-map of the central-western Pannonian Basin and location of the study area

parts of the Pannonian Basin may exceed several thousand meters, is represented here within a few hundred meters (though affected by major stratigraphic gaps) (Fig. 2). Hence, Lake Balaton can be regarded as a major "water-window", open for geophysical exploration over the Pannonian Basin.

2. *Tectonic setting*

The Pannonian Basin (Fig. 1) is a large intra-montane depression superimposed on the Alpine megasuture (Bally and Snelson 1980; Horváth et al. 1981). It is encircled by the Carpathian Mountains and the Dinarides and includes a complex system of extensional basins which evolved in the Neogene –Quaternary during coeval compression in the outer Carpathian arc (Royden 1988; Horváth 1993). The Pannonian depression is located over a thinned continental crust associated with high heat flow and relatively hot lithosphere (Royden and Dövényi 1988; Horváth 1993). The basinal area consists of a system of several minor, individual basins (Royden and Horváth 1988; Royden 1988) that are separated by relatively shallow basement thresholds and in some place contain more than 7000 m of Neogene to Quaternary basin fill. The basement of the basin system is composed of a Mesozoic nappe pile and remnants of Paleogene basins (Tari et al. 1993; Tari 1994) which locally crop out, thus forming the Hungarian mountains. A relatively continuous chain of mountains from the SW to the NE constitutes the Transdanubian Central Range.

Recent studies on the geodynamic evolution of the Pannonian Basin (Tari et al. 1992; Horváth 1993; Tari 1994; Horváth 1995) have shown fairly complex mechanisms of basin formation and evolution. These include subduction-related extensional collapse of an overthickened orogenic pile and collision-related escape of orogenic terranes. Early to Middle Miocene extensional/strike-slip tectonics caused relatively fast syn-rift subsidence of narrow basinal areas while a significant part of the Pannonian region still remained intact and elevated. Late Middle Miocene marked the onset of the post-rift phase when thermal cooling resulted in a generalized subsidence and broadening of the whole Pannonian area.

Based on the interpretation of seismic profiles and subsidence analysis of boreholes, Horváth et al. (1993), Tari (1994), and Horváth and Cloetingh (1996) showed that during latest Neogene–Quaternary the Pannonian Basin underwent significant tectonic inversion. Reverse faulting associated with uplift occurred locally and the Hungarian mountains began to rise, thus causing extensive erosion of large areas covered by Late Neogene deposits (Fig. 2). A recent study (Horváth 1995) suggested that the tectonic tranquillity of Pannonian Basin was interrupted even earlier in the syn-rift phase at the end of Sarmatian, by compressional (locally transpressional) events. Tectonic activity was accompanied by widespread volcanic activity in the Pannonian Basin. Middle Miocene to Pliocene calc-alkaline volcanism and Pliocene–

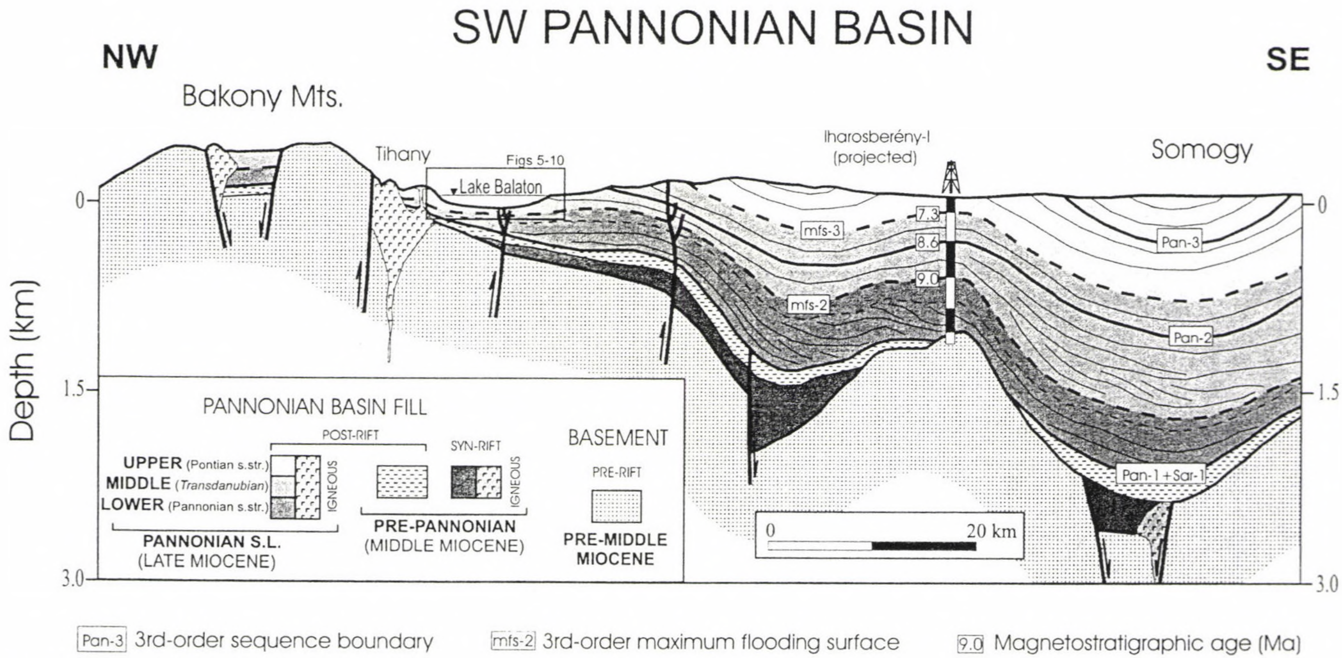

Fig. 2
Sketch-section across the SW Pannonian Basin based on interpreted regional seismic profiles (Sacchi et al. in press) and compiled after Tari (1994). Approximate location is shown in Fig. 1. Note the adopted three-fold subdivision of the Pannonian s.l. sequence (after Sacchi et al. 1997, in press). Erosional truncation of Pannonian strata toward the NW is due to post-Pontian (post-Miocene) uplift of the Bakony Mountains

Pleistocene alkali-basaltic volcanism resulted in a number of volcanic edifices scattered over the basinal area (Balogh et al. 1986).

3. *Stratigraphic setting*

Traditionally, stages and stage systems of the Pannonian Basin (Central Paratethys) have been derived from biostratigraphy complemented with lithologic data. Particularly the classic subdivision of Pannonian strata (Roth 1879; Lőrenthey 1900; Halaváts 1903; Stevanovic 1951) has been mostly based on benthic mollusk assemblages defined within marginal facies of the basin sequence. Due to significant provincialism and facies dependence of Paratethys endemic faunas the stratigraphic correlation within the Pannonian Basin has always been a challenging task and the reliable correlation of Paratethys non-marine stages with standard marine stages is still a long way from reality. To address this problem a regional Paratethys Stage System has been progressively developing since the late sixties (Papp et al. 1968; Cicha and Senes 1968; Steininger and Rögl 1979; Rögl and Steininger 1983; Papp et al. 1985; Steininger et al. 1988; Nagymarosy 1990; Steininger et al. 1988, 1990; Stevanovic 1990; Müller and Magyar 1992, 1995; Sacchi et al. 1997, in press).

It is widely believed that some of the index fossils used to identify the boundaries of the Pannonian stage are distributed according to diachronous biostratigraphic units across the basin. This has been sometimes indicated by the use of "time-transgressive" stage boundaries (Vass et al. 1987, 1988; Steininger et al. 1990; Rögl et al. 1991; Lantos et al. 1992). Current controversies revolve mainly around the absolute ages of Paratethyan stages and their correlation with the standard biostratigraphic time scale.

Additional confusion was created by the use of the term "Pannonian" with different chronostratigraphic meanings (i.e. Pannonian s. Roth 1879; Pannonian s. Lőrenthey 1900; Pannonian s. Stevanovic. 1951). In everyday practice, for instance, many geologists still use the term Pannonian according to Lőrenthey (1900) and use the term "Lower Pannonian" instead of Pannonian s.str., and "Upper Pannonian" instead of Pontian. This also led to the disagreeable practice of assuming an implicit chronostratigraphic correlation between the "local" Upper Pannonian Stage and the Pontian stage of the "official" Stage System, without adequate documentation.

According to our interpretation, none of the stage names in current use in Hungary adequately represents the middle part of Pannonian s.l. stage (ca 9.0–7.3 Ma in the chronology adopted in this study). Based on recent stratigraphic research (Müller and Magyar 1992, 1995; Sacchi et al. 1997), Sacchi et al. (in press) have suggested a three-fold subdivision of the Pannonian s. Lőrenthey (1900) into Lower Pannonian (Pannonian s.str.), Middle Pannonian ("Transdanubian" or "Danubian") and Upper Pannonian (Pontian). The last column to the right in Fig. 3 illustrates such chronostratigraphic subdivision of Pannonian strata which we also adopt in this study.

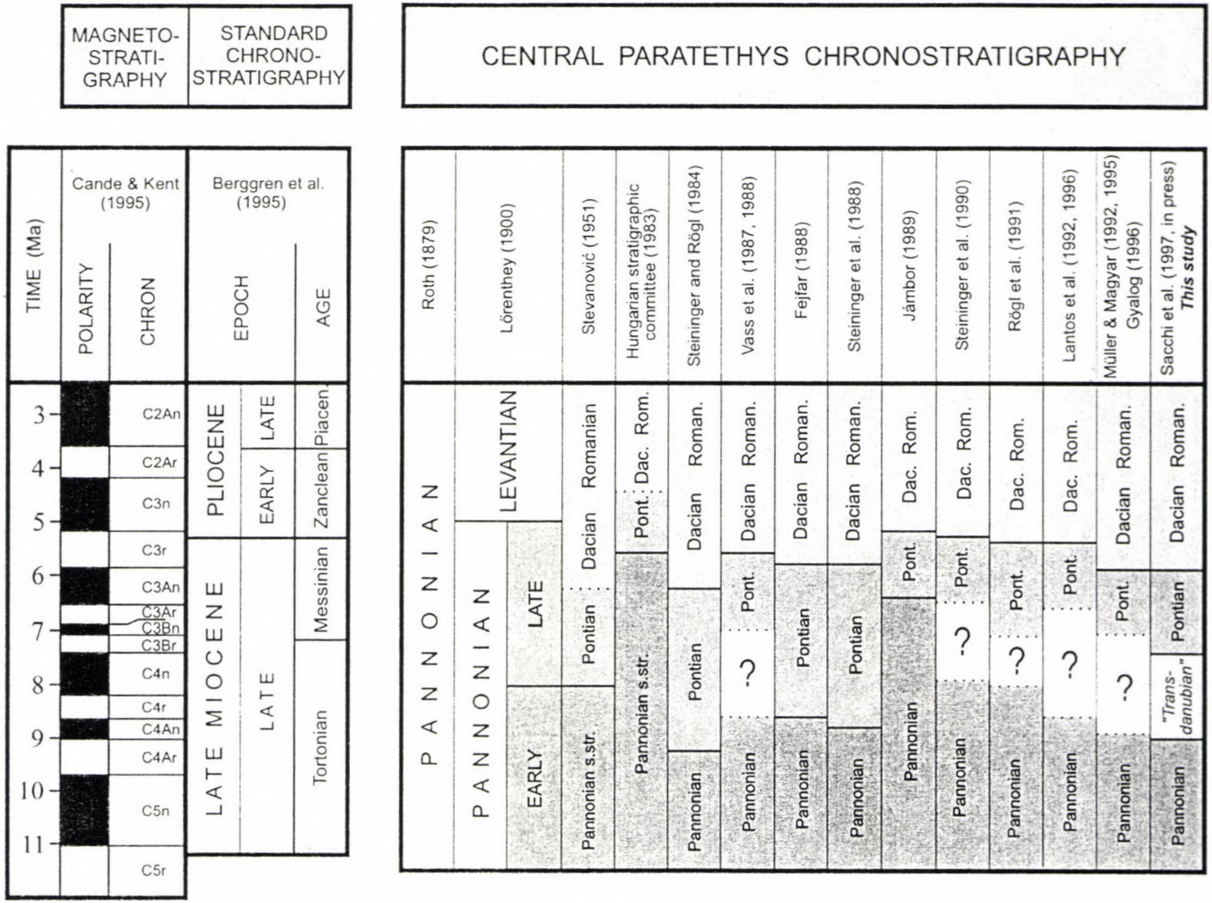

Fig. 3
Synopsis of Late Neogene chronostratigraphic units for the Central Paratethys and the Mediterranean (see Sacchi et al. in press and references therein)

4. Depositional setting of the SW Pannonian Basin

Most recent stratigraphic research has focused on the sequence architecture of shallow marine deposits. Although a consensus seems to be emerging regarding the application of sequence stratigraphic concepts to the continental record (Shanley and McCabe 1994; Miall 1997), very few well-documented examples of non-marine sequences have yet been described. The case history of the Pannonian Basin (Ujszászi and Vakarcs 1993; Csató 1993; Vakarcs et al. 1994) illustrates some of the difficulties in applying sequence concepts to lacustrine and fluvial deposits, suggesting caution when using systems-tract terminology derived from marine processes for the labeling of non-marine events. Sacchi et al. (in press) have recently shown that direct correlation of 3rd-order sequences (cycles with 10^6 year periodicities) of the Pannonian Basin with the Haq et al. (1987) global cycle chart is not possible.

The Pannonian Lake originated about 12 million years ago and survived at least until the Pliocene (Kázmér 1990). Seismic data coupled with well-log interpretation and core sample analysis have shown that the Pannonian Basin was filled up by a fluvial-dominated delta system which prograded into a large lacustrine basin (Pogácsás 1984; Bérczi and Phillips 1985; Mattick et al. 1988; Pogácsás et al. 1988; Horváth and Pogácsás 1988; Juhász 1994). The Late Neogene evolution of the Pannonian basin was accompanied by decreasing rates of subsidence (thermal subsidence) along with an increase of sedimentation rates. This resulted in progradational stratal patterns and late-stage aggradation. Sediment supply mainly followed radial patterns from the Carpathian belt at the margins toward the interior of the basin.

The slowly subsiding area experienced tectonic reactivation during the Late Pliocene and/or Quaternary (Tari 1994; Horváth 1995). This was manifested by accelerated subsidence at the basin center and faulting associated with uplift of the basin flanks (Horváth and Cloetingh 1996). Consequently, extensive erosion occurred over large areas covered by Late Miocene to Pliocene deposits. Erosional truncation of strata due to late-stage uplift of the Bakony Mts is best imaged by high-resolution seismic profiles of Lake Balaton (Figs 5–9).

The Pannonian sequence unconformably overlies Sarmatian strata in the deepest basinal areas and older rocks at the basin margins. Sandy turbidite units interbedded with marl represent the lowermost part of the Pannonian sequence in the deep basins (Danube, Zala and Drava). Sandstone and marl follow toward the top indicating delta slope and delta plain settings. The prograding delta complex is overlain by alluvial deposits. Pannonian strata commonly overlie a major unconformity at the marginal part of the basin. Turbidites and slope deposits are missing and basal transgressive sequences consist of nearshore conglomerates.

A major flooding event occurred in the Pannonian Lake toward the end of the Pannonian s.str. (Sacchi et al. in press). This was manifested by *Congeria czjzeki*-bearing open lacustrine beds (Szák Fm.) which flooded the basin margins. The maximum flooding surface associated with this event (mfs-2), dated ca 9.0 Ma by calibration with magnetostratigraphy at Iharosberény-I well (Fig. 2), represents an isochronous surface at basin scale which can be correlated with the top of Pannonian s.str. stage (Lower Pannonian) (Sacchi et al. 1997). The overlying coarse clastics represent lakeshore progradation. These are usually sand and gravel sorted by waves in beach depositional settings. The offshore-nearshore transition is represented by the Somló Formation. Overlying nearshore deposits are often associated with coastal zones covered by dense vegetation. Sand, variegated silt, huminite-rich silt and calcrete were deposited in lagoons and coastal swamps (Tihany Fm.).

Due to a significant water level drop in the Pannonian Lake at about 8.6 Ma (Pan-2 SB in Fig. 2) extensive subaerial exposure of lake margins occurred. This is widely recorded in the so-called "marginal facies" of western Hungary. Subaerial travertine, evaporitic dolomite beds, calcrete and paleosols (top of the Tihany Fm.) and epigenetic processes typical of the vadose and the upper phreatic zone such as groundwater silcrete (Mindszenty, pers. comm.) developed as a consequence.

A second important flooding event in the basin followed which we tentatively correlate with *Congeria rhomboidea*-bearing beds. The maximum flooding surface associated with this event (mfs-3), dated ca 7.3 Ma (Fig. 2), may be assumed as the top of our Middle Pannonian ("Transdanubian") stage (see also Sacchi et al. 1997 in press). Time-equivalent facies are unfortunately missing at the basin margin as they have been removed by erosion caused by late stage uplift of basin flanks. Consequently it is likely to be the case that Pontian s.str. (younger than 7.0 Ma) strata are practically absent in outcrop in central-western Hungary.

Lacustrine deposits of the Pannonian basin are overlain everywhere by alluvial deposits. Locally basalts and associated volcanoclastics and shallow water limestones cap the Pannonian–Pliocene sedimentary sequence. During the Late Pliocene–Quaternary shallow isolated lakes, wetlands, and mostly continental conditions prevailed throughout the basin.

5. *Seismic survey on Lake Balaton and subsurface data*

The 1993 survey on Lake Balaton (cruise GMS93-02) was carried out on board of the ship *Vízvédelem*. A differential GPS system with slave stations located at the lakeshore was used for accurate ship positioning. Acquisition of seismic signals was obtained with a Uniboom-type single-channel system, 300-Joule energy source, frequency filters and TVG unit. The shot rate was 1 sec. Printout on EPC recorder was coupled with analogue recording on VHS hi-fi magnetic tape which was used for further digital conversion and processing (Magyari in press). The grid of seismic profiles is shown in Fig. 4. Details of navigation courses are summarized in Table 1.

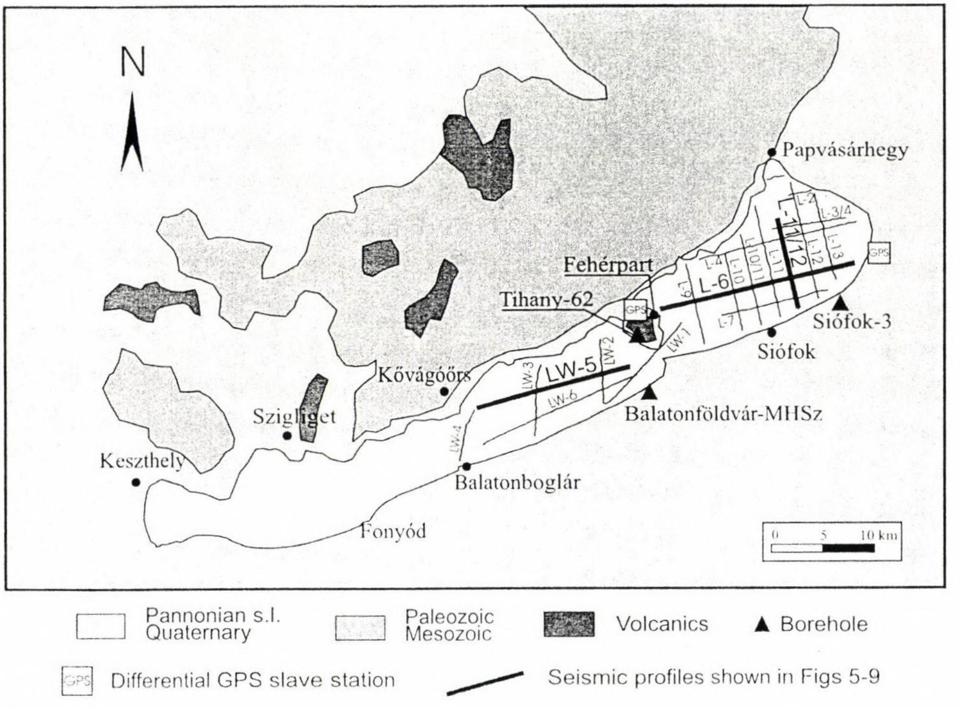

Fig. 4
Location map of high-resolution seismic profiles acquired over Lake Balaton in June of 1993 during the cruise GMS93-02

The survey investigated the Lake Balaton subsurface up to a maximum time-depth of ca 180 msec down to the acoustic basement which is represented by Sarmatian marl and marly limestone. Constraint for seismic stratigraphic interpretation was given by boreholes Tihany-62, Siófok-3, Balatonföldvár-MHSz, and a limited amount of fieldwork on selected areas (Bántapuszta, Papvasárhegy, Tihany, and Kővágóörs) in the surroundings of the lake. Calibration of seismic profiles was obtained by the intersection of seismic section LW-1 with interpolated stratigraphy between boreholes Tihany-62 and Balatonföldvár-MHSz (Fig. 4). Sequence stratigraphic interpretation of seismic profiles was aided by the study of several regional seismic profiles in southern Transdanubia and the Iharosberény-I well-log and magnetostratigraphy (Sacchi et al. in press) (Fig. 2).

5.1. Lake Balaton

Lake Balaton covers an area of about 600 km^2 (Fig. 4). It has a SW–NE elongated shape (ca 78 km in length and up to 14 km in width) and its surface is at 105 m above sea level. It is the largest lake in Central Europe, although the most shallow (water depth is 3–4 meters on average, with a 12 m maximum). The Balaton is part of a relatively young hydrographic system which was formed in the latest Pleistocene–Holocene, during the post-Würm deglaciation of Central Europe. The oldest lacustrine deposits have been dated at approximately 15.000 years b.p. by palynologic analysis (Nagy-Bodor 1988) as well as ^{14}C dating (Cserny et al. 1995).

Lake Balaton presumably formed after the joining of several individual ponds or minor lakes as climatic conditions changed from cold-arid to warm-humid during the *Quercus–Fagus* vegetation phase (Cserny and Nagy-Bodor 1996). The seismic interpretation presented in this study confirmed the suggestion that the shape and possibly the location of Lake Balaton might have been controlled by SW–NE trending strike-slip faults (Fig. 11). Balaton lacustrine deposits show an average thickness on the order of 6–8 m (Figs 5–10) and are represented by silt and subordinately clay and fine sand, with a carbonate content up to 50–70%. Through the latest Pleistocene– Holocene, average sedimentation rates were ca 0.4 mm/year, varying from a minimum of 0.2 mm/year to a maximum of 1.0 mm/year (Cserny et al. 1991). Lake Balaton deposits unconformably overlay Middle Pannonian ("Transdanubian") strata along a relatively smooth erosional surface. This surface can be traced as a regional unconformity and marks a stratigraphic hiatus which practically encompasses the entire Pliocene and most of the Pleistocene.

Table 1
Summary of navigation courses and seismic profiles acquired on Lake Balaton in June of 1993 during the cruise GMS93-02

Date of acquisition	Seismic profile	Heading	Length (km)
17. 06. 93	L-1 (test)	W–E	2.0
17. 06. 93	L-1 (test)	E–W	2.5
17. 06. 93	L-11 (test)	N–S	11.5
17. 06. 93	L-12	S–N	12.0
17. 06. 93	L-13	N–S	9.0
18. 06. 93	L-4	E–W	16.0
18. 06. 93	L-9	N–S	8.0
18. 06. 93	L-6	W–E	19.5
19. 06. 93	L-10	N–S	8.0
19. 06. 93	L-10/11	S–N	9.5
19. 06. 93	L-11	N–S	11.5
19. 06. 93	L-11/12	S–N	9.0
19. 06. 93	L-3/4	E–W	7.5
20. 06. 93	LW-1	NE–SW	11.0
20. 06. 93	LW-2	S–N	6.5
20. 06. 93	LW-3	N–S	7.5
20. 06. 93	LW-4	S–N	5.5
20. 06. 93	LW-5	W–E	15.0
21. 06. 93	LW-6	W–E	16.0

6. Seismic stratigraphy of the Late Neogene succession

Three major seismic stratigraphic units (separated by two major unconformities) have been recognized beneath Lake Balaton. These units, from bottom to top, are (1) the Pre-Pannonian acoustic basement, (2) the Pannonian s.l. sequence and (3) the Upper Pleistocene–Holocene deposits. The interpretation of sections L-6, LW-5 and L-11/12 illustrates the reconstructed

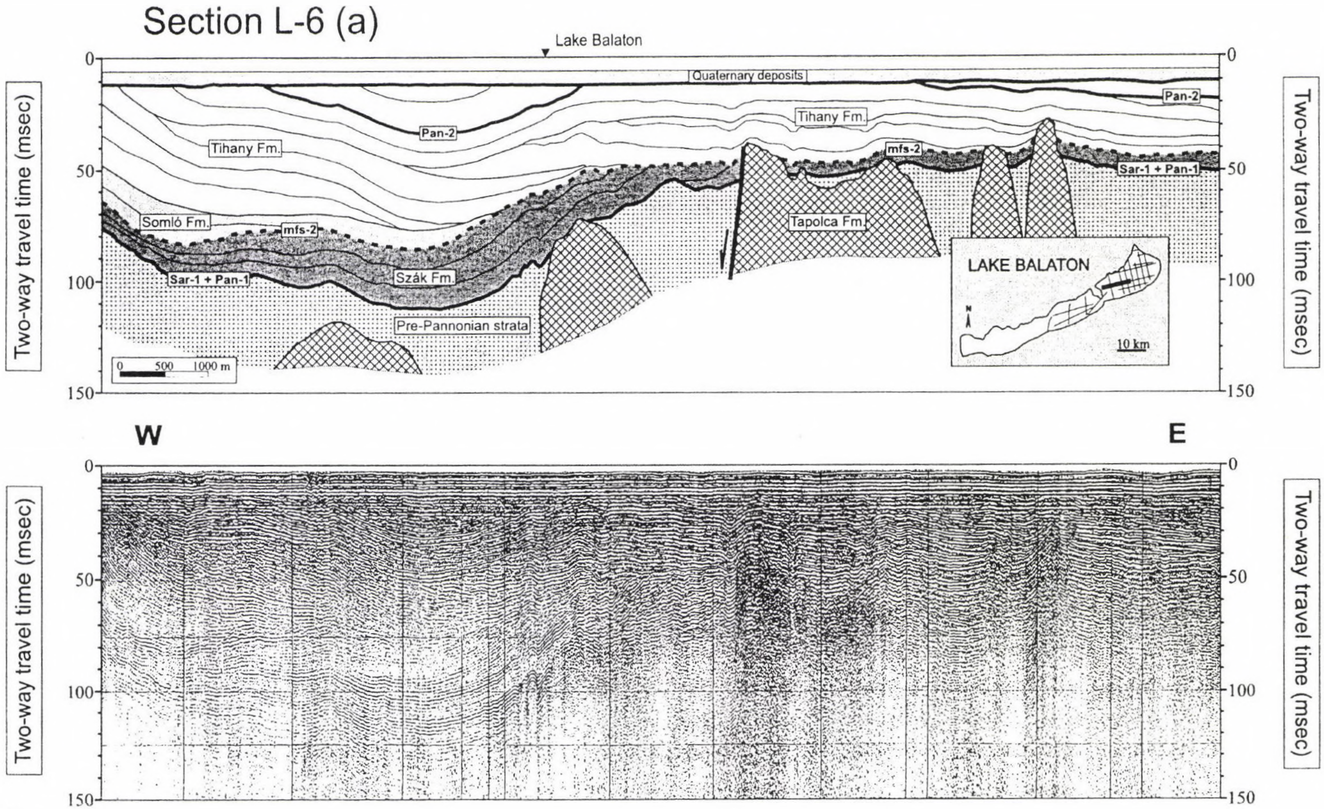

Fig. 5
Interpretation of high-resolution seismic profile L-6 (a), NE Lake Balaton. The Szák Fm. (Lower Pannonian) corresponds to a transgressive systems tract (TST). The Somló Fm. and the lower part of the Tihany Fm. (Middle Pannonian) are interpreted as a highstand systems tract (HST). The upper part of Tihany Fm. (Middle Pannonian) corresponds to the falling stage systems tract (FSST)/ lowstand systems tract (LST) of the next sequence. The volcanic near-surface intrusion of the Tapolca Fm. (Middle–Upper Pannonian) is coeval or slightly postdates the deposition of the Szák Fm. The maximum flooding surface mfs-2 (ca 9.0 Ma) can be used to define the boundary between the Lower Pannoian (Pannonian s.str.) and the Middle Pannonian ("Transdanubian") in the study area

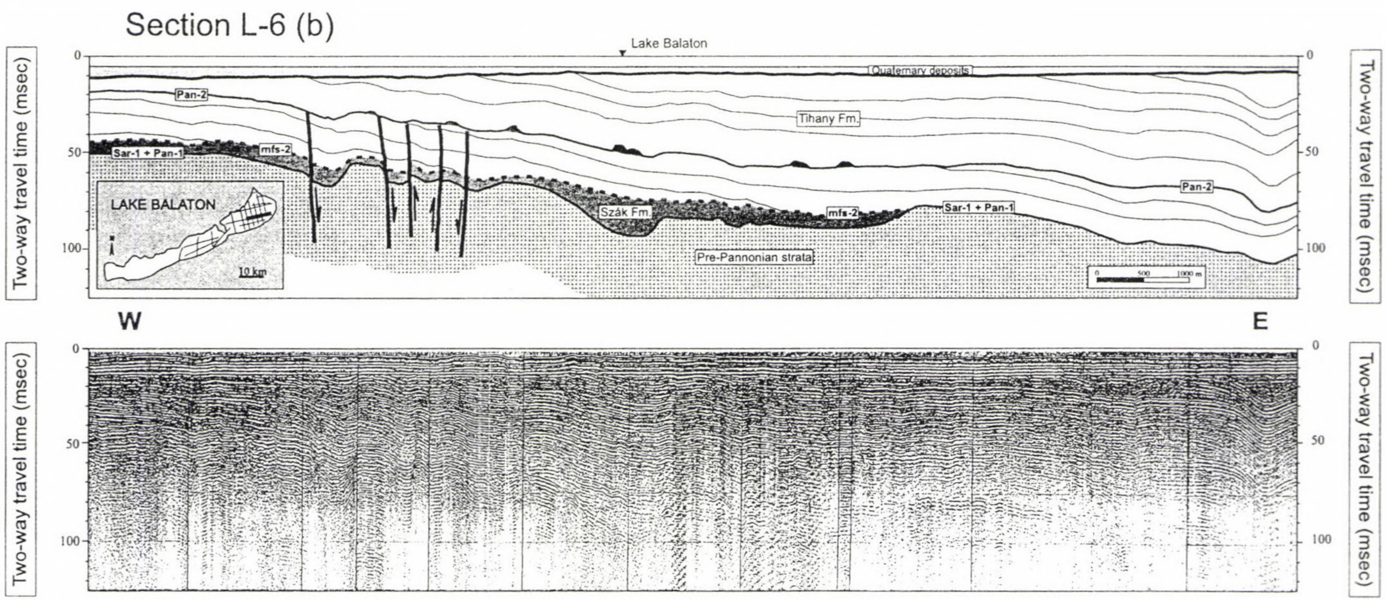

Fig. 6
Interpretation of high-resolution seismic profile L-6 (b), NE Lake Balaton. Reef-like structures visible along sequence boundary Pan-2 (ca 8.6 Ma) are interpreted as travertine mounds (geyserite of the Hungarian literature) which also crop out on top of the Tihany peninsula (see also Fig. 9 and text for discussion). Refer to Fig. 5 for the key to the sequence stratigraphic interpretation

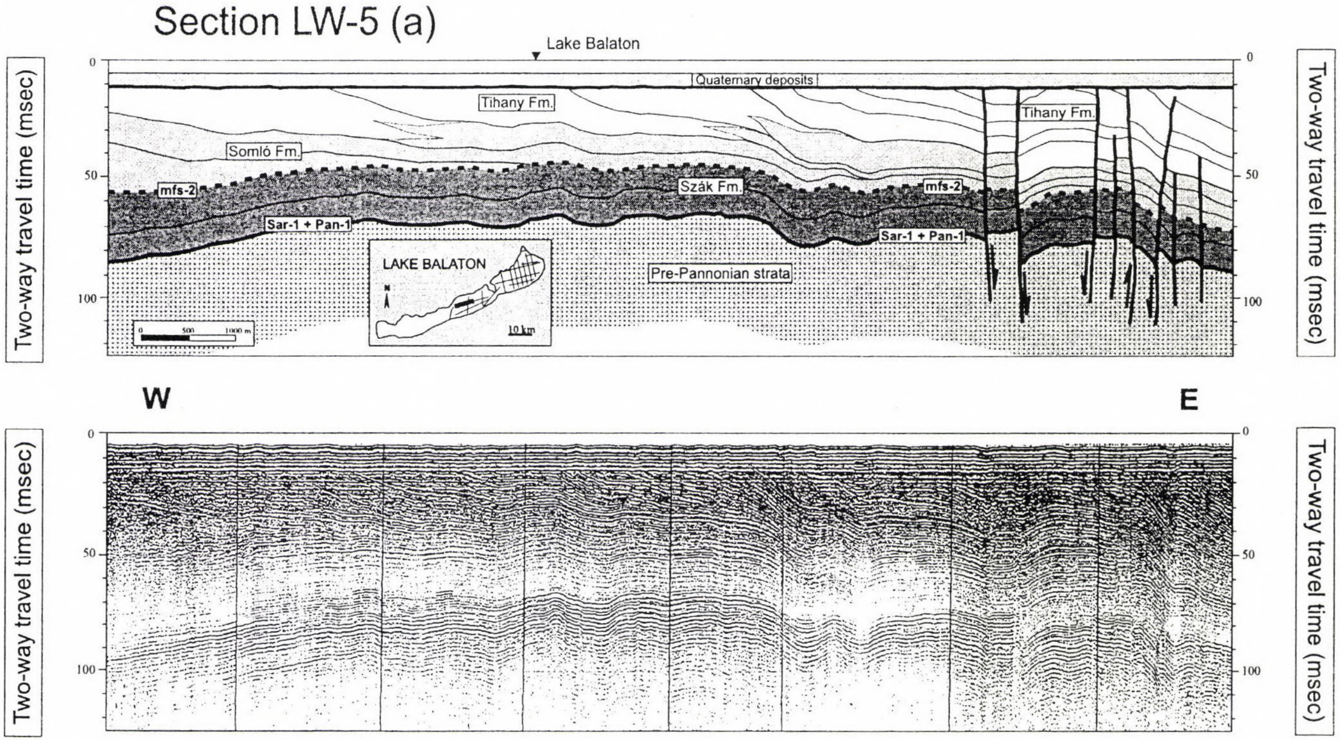

Fig. 7
Interpretation of high-resolution seismic profile LW-5 (a), central Lake Balaton. The tectonic deformation postdates the Late Miocene and is due to strike-slip (left lateral?) SW–NE faulting and associated folding. Refer to Fig. 5 for the key to the sequence stratigraphic interpretation

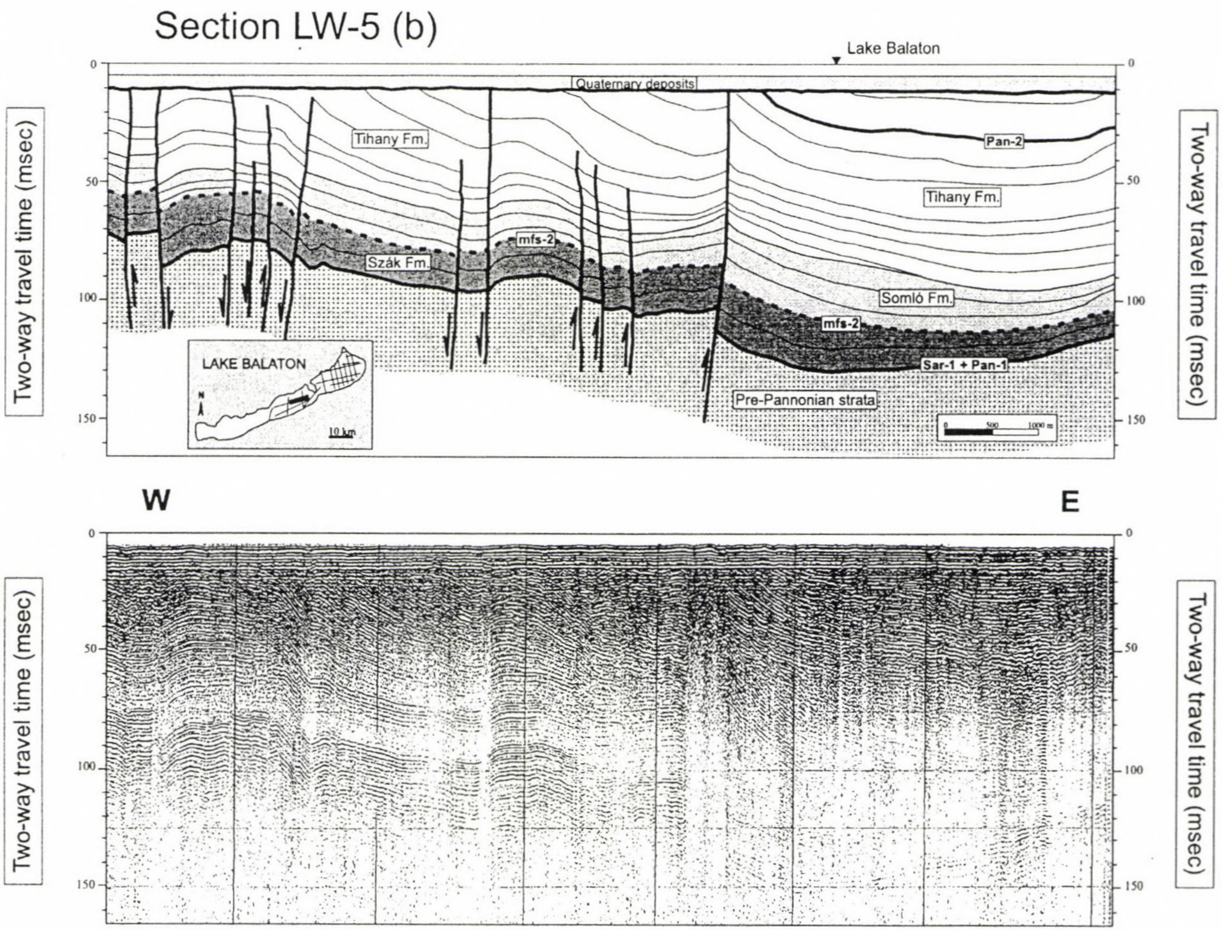

Fig. 8
Interpretation of high-resolution seismic profile LW-5 (b), central Lake Balaton. Note the dramatic tectonic deformation associated with SW–NE strike slip faulting (a fault zone is cut at low-angles by the seismis profile). Refer to Fig. 5 for the key to the sequence stratigraphic interpretation

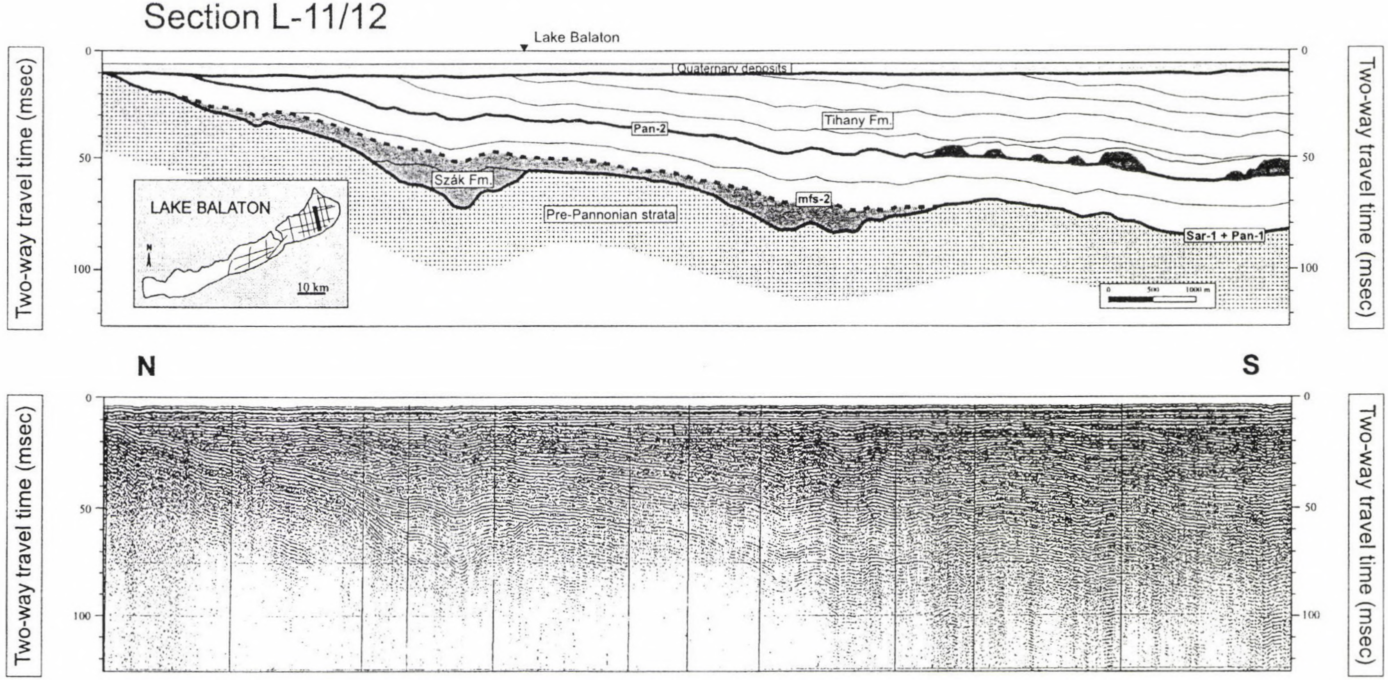

Fig. 9
Interpretation of high-resolution seismic profile L-11/12, NE Lake Balaton. Note the "reef-like" structures along Pan-2 SB and v-shaped valleys and erosional fluvial terraces incised in the pre-Pannonian (Sarmatian) strata. The channel fills within incised valleys are bounded at the top by a transgressive surface and may represent FSST and/or LST deposits. Refer to Fig. 5 for the key to the sequence stratigraphic interpretation

stratigraphy and the outcrop-scale architecture of stacking sequences beneath Lake Balaton (Figs 5–9). Figure 10 shows a composite geologic sketch-section across the study area.

6.1. Pre-Pannonian acoustic basement

The acoustic basement is represented by Sarmatian limestone and marl. Sarmatian strata are dramatically truncated at the top by a mature (polycyclic) erosional surface which includes amalgamation of two 3rd-order sequence boundaries, namely Sar-1 SB (top of the Sarmatian sequence) and Pan-1 SB (top of the first Pannonian 3rd-order sequence). Incised valleys, fluvial terraces and channel fills are also recognized from the seismic record (Fig. 9). Channel fills within incised valleys are bounded at the top by a transgressive surface and may be interpreted as falling stage systems tract (FSST) and/or lowstand systems tract (LST) deposits (Shanley and McCabe 1994).

Correlation with borehole Tihany-62 shows that the unconformity at the top of Sarmatian strata marks a significant stratigraphic gap (about 12 to 9 Ma) ranging from the Upper Sarmatian to the *Congeria czjzeki* beds (upper part of Pannonian s.str.). The duration of this hiatus increases from W to E (Sacchi et al. in press) to include the entire Sarmatian. Good exposures at this stratigraphic level are found at Bántapuszta, where Pannonian s.str. lacustrine strata directly overlie Badenian marine deposits.

6.2 Pannonian sequence

The maximum thickness of this unit is on the order of 200 m. The lowermost part of the Pannonian sequence is represented by the transgressive systems tract (TST) deposits of the Szák Formation which onlap the underlying Sarmatian "basement". The Somló Formation and the lower part of the Tihany Formation follow; they are interpreted in turn as highstand systems tract (HST) deposits. The stratigraphic architecture of these deposits shows typical forestepping strata which downlap onto the underlying TST deposits and are accompanied by local development of small coarse-grained prograding deltas (Figs 5 and 7). A subtle 3rd-order sequence boundary (Pan-2 SB) is also detected within the Tihany Formation, the upper part of which corresponds to the FSST/LST of the next sequence. Pan-2 SB can also be traced as a regional unconformity on seismic profiles throughout SW Hungary (Fig. 2). It is related to a significant lowering of the base level of erosion in the Pannonian Lake and is possibly associated with tectonic and volcanic activity. Pan-2 SB crops out at the top of the Tihany Peninsula. Its occurrence (ca 8.6 Ma) predates the onset of the basaltic eruption of the Tihany Volcano (ca 7.8 Ma, Balogh 1995). Stratal patterns toward volcanic near-surface intrusions suggest that most of the magmatic activity inferred from seismic profiles was coeval or slightly postdated the deposition of the Szák Fm. (ca 9.0 Ma) (Sacchi et al. in press).

Scattered over the top of the Tihany Peninsula, a number of silicified carbonate mounds (geyserite of the Hungarian literature) are found that are currently believed to be Pliocene–Pleistocene in age; they have been interpreted up to now as purely chemical deposits related to post-volcanic activity (Lóczy 1913). According to our interpretation these features are best interpreted as silicified *travertine* mounds, possibly deposited at warm/hot springs. Furthermore, many of these travertine mounds can be traced down into the Balaton subsurface and appear to correlate with Pan-2 SB (Figs 6, 9 and 10). This would suggest that they are Late Miocene in age (Sacchi et al. 1995, in press).

According to our sequence stratigraphic framework we hypothesize that the Kálla Formation, a coarse-grained foreshore deposit that crops out in the Kál Basin north of Lake Balaton, corresponds to some facies heteropy of the lower part of the Tihany Fm., being itself part of a HST progradational–aggradational unit that formed along the Pannonian Lake shore between 9.0 and 8.6 Ma. It is to be noted that a significant part of the Kálla deposits in this area were possibly removed by erosion during the lake level fall which resulted in Pan-2 SB. One puzzling feature within the Kálla Formation is certainly the *kőtenger* (sea of stones). This is a 2 m-thick silicified sandstone and conglomerate bed which may be interpreted as a groundwater silica-cemented regolith (groundwater silcrete) (A. Mindszenty, pers. comm.). According to our interpretation the *kőtenger* may be regarded as an epigenetic feature that developed at the phreatic–vadose interface in response to the drop of the base level of erosion within the Pannonian Lake at 8.6 Ma (Pan-2 SB); in other words it is a "sequence stratigraphic equivalent" of the silicified travertine mounds at Tihany. The occurrence of silcrete would also suggest relatively warm (dry/humid seasonal) climatic conditions during that time.

6.3 *Quaternary deposits of Lake Balaton*

Quaternary deposits of Lake Balaton show an average thickness of 5–6 m. Seismic response of these deposits is characterized by parallel and continuous reflectors. A dramatic erosional surface at the base of these deposits truncates the underlying Pannonian strata (Figs 5–10). This surface can be traced as a major unconformity and marks a stratigraphic hiatus which encompasses practically the entire Pliocene and most of the Pleistocene.

7. Tectonic interpretation

In the area of Lake Balaton the Pannonian sequence is affected by intense tectonic deformation. Tectonic activity mostly postdates the Late Miocene and consists of SW–NE strike-slip faulting, associated with gentle folding (Fig. 11). Evidence of minor tectonic activity exists between mfs-2 and Pan-2 SB (9.0 to 8.6 Ma). This may suggest causal relations between tectonic and coeval volcanic

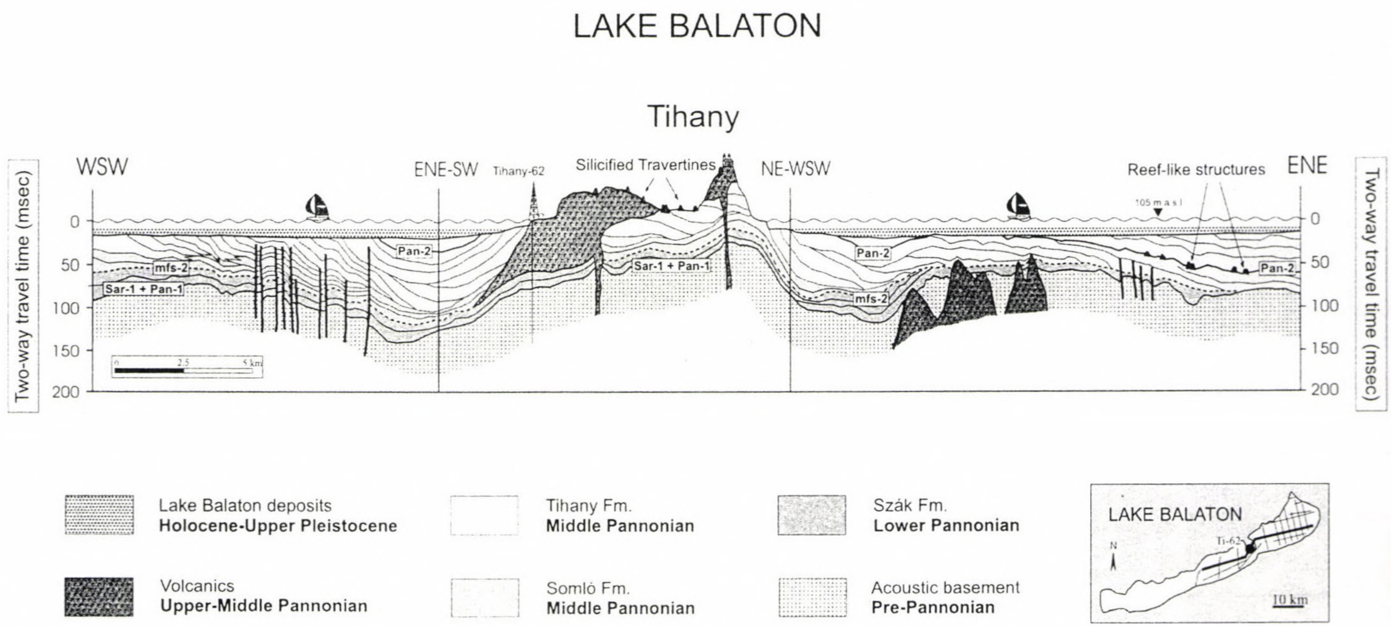

Fig. 10
Sketch-section across Lake Balaton (not to scale). The section is obtained by a combination of interpreted seismic profiles (L-6 and LW-5) and a geologic section of the Tihany peninsula based on outcrop and subsurface data (see the text for discussion)

activity in the study area and may also offer an alternative (other than purely climatic) explanation to account for the 3rd-order changes of base level of erosion within the Pannonian Lake. Significant tilting of the entire Pannonian succession toward the SE is also seen on seismic profiles that consistently show ESE-dipping strata which are erosionally truncated at the top. This supports earlier interpretations which suggested that the Pannonian Basin underwent significant late-stage tectonic inversion and uplift with consequent subaerial erosion of Late Neogene strata in Transdanubia and Northern Hungary (Tari 1994; Horváth 1995).

8. *Summary and conclusion*

The Late Neogene Pannonian Basin is an excellent case study for sequence stratigraphy applied to non-marine settings. The 1993 survey on Lake Balaton has shown that high-resolution seismic may offer a precious tool to link sequence stratigraphic procedure applied to seismic profiles and the fieldwork.

In accordance with the data reported and discussed in this paper, a number of conclusions can be outlined and summarized as follows:

1) Lake Balaton is part of a relatively young hydrographic system which evolved in the latest Pleistocene–Holocene during the post-Würm deglaciation of Central Europe. The unconformity at the base of the Lake Balaton Quaternary deposits is best imaged by high-resolution seismic data as a subaerial erosional surface which truncates the underlying Middle Pannonian ("Transdanubian") strata (Figs 2 and 5–10). This unconformity marks a major stratigraphic hiatus that includes practically the entire Pliocene and most of the Pleistocene.

2) The Pannonian sequence which is found beneath the Quaternary deposits of Lake Balaton (Figs 2 and 5–10) is represented from bottom to top by:

– The Szák Formation: It is interpreted as TST deposits which onlap the underlying Sarmatian "basement". The top of this formation corresponds to a regional maximum flooding surface (mfs-2) which can be dated at 9.0 Ma by calibration with Iharosbéreny-I well magnetostratigraphy. This isochronous stratigraphic surface (mfs-2) can be used to define the base of the Middle Pannonian ("Transdanubian") stage in the type area (Transdanubia).

– The Somló Formation and the lower part of the Tihany Formation: These are interpreted as HST deposits. A 3rd-order sequence boundary (Pan-2 SB) is detected toward the middle part of the Tihany Formation. It is related to a significant lowering of the base level of erosion in the Pannonian Lake and correlates with a magnetostratigraphic age of ca 8.6 Ma (at the Iharosbéreny-I well site). Pan-2 SB is recognized as a regional unconformity on seismic lines. Locally it may be associated with a significant stratigraphic gap and hence cannot be used as a regional time-line.

– The upper part of the Tihany Formation (above Pan-2 SB): It corresponds to the FSST/LST of the next sequence and is mostly developed in the subsurface toward the northeastern part of Lake Balaton. The geometry of seismic reflectors

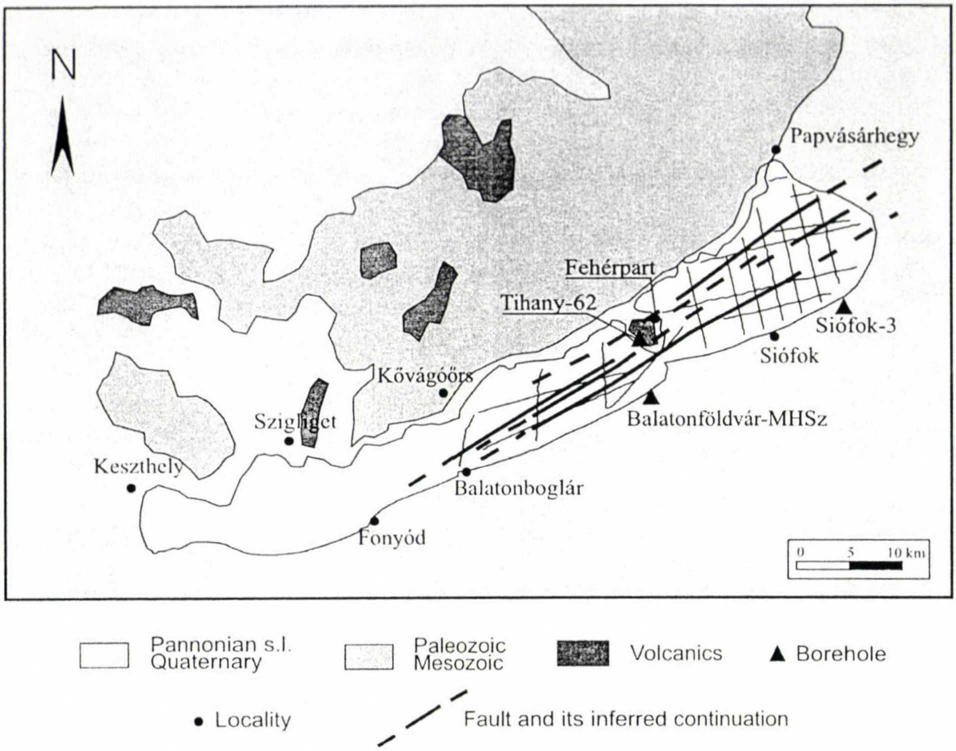

Fig. 11
Main faults detected from the interpretation of high-resolution seismic profiles acquired over Lake Balaton in June of 1993. Tectonic deformation mostly postdates the Late Miocene and consists of WSW–ENE strike-slip (left-lateral?) faulting (see also Figs 5–10) associated with gentle folding. Fault patterns suggest a possible tectonic control over shape and location of the lake itself

suggests aggrading strata, possibly corresponding to marginal lagoon and swamp depositional settings.

3) Silicified travertine mounds (geyserite) occur at the top of the Tihany Peninsula which have long been believed to be Pliocene–Pleistocene in age. However, the observation that these mound-shaped travertines can be traced down to the subsurface of Lake Balaton along Pan-2 SB suggests instead (Figs 6, 9 and 10) that these features might be Late Miocene (ca 8.6 Ma) in age.

4) The *kőtenger* (sea of stones), a 2 m-thick silicified sandstone and conglomerate bed which is found in the Kálla Formation in the surroundings of Lake Balaton, may be interpreted as an epigenetic feature (groundwater silcrete) which developed at the phreatic–vadose interface in response to the lowering of the base level of erosion within the Pannonian Lake, causing the Pan-2 SB (ca 8.6 Ma) to occur.

5) Subsurface-outcrop correlation showed that the basement of the Pannonian s.l. (Late Miocene) sequence beneath Lake Balaton is represented by a major unconformity which erosionally truncates Sarmatian strata. This unconformity includes the amalgamation of two 3rd-order sequence boundaries (Sar-1 SB + Pan-1 SB) (Figs 5–10). It carries evidence of mature subaerial erosion (incised valleys, fluvial terraces and channel fills) and marks a significant stratigraphic gap (about 12 to 9 Ma) ranging from the Upper Sarmatian to the *Congeria czjzeki* beds (upper part of Pannonian s.str.).

6) The Late Pannonian sequence underlying the Quaternary deposits of Lake Balaton is affected by intense tectonic deformation. Tectonic activity postdates the Late Miocene and consists of gentle folding, associated with SW–NE strike-slip faulting (Fig 5–11). Overall tilting of strata toward SSE is detected from the seismic record. This fits earlier interpretations which suggested that the Pannonian Basin underwent significant late-stage tectonic inversion and uplift with consequent subaerial erosion of Late Neogene strata in Transdanubia and Northern Hungary (Tari 1994; Horváth 1995).

7) Pontian s.str. strata (younger than 7.0 Ma) are practically missing in outcrop all across central-western Hungary. Following recent stratigraphic research (Müller and Magyar 1992, 1995; Sacchi et al. 1997) we suggest that none of the stage names in current use in Hungary adequately represent the middle part of Pannonian s.l. stage (ca 9.0–7.4 Ma in the chronology adopted in this study). We also reiterate the point made by Sacchi et al. (in press) who proposed (Fig. 3) *a three-fold subdivision of the Pannonian stage s. Lőrenthey (1900) into Lower Pannonian (Pannonian s.str.), Middle Pannonian ("Transdanubian") and Upper Pannonian (Pontian s.str.).* Deposits of Middle Pannonian age are likely to be the youngest Pannonian s.l. strata in the vicinity of Lake Balaton.

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Application of glauconite to the palaeoenvironmental studies of Tertiary basins in the East-central part of Poland (Lublin Region)

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In the area of geologic studies the mineral glauconite has gained much in importance owing to the great interpretative possibilities of its physical-chemical features. Glauconite, as one of the scarce minerals of sedimentary rocks, is often applied to analyse stratigraphic–palaeogeographical and tectonic problems and to characterise the sedimentological environment. It reacts easily to temperature changes, oxidising-reducing potential and other parameters of the environment. This mineral is interesting not only from a theoretical point of view but is also of practical significance.

On the basis of physical-chemical features of glauconite from Tertiary sediments of the Lublin region the depositional environment of these sediments was characterised confirming their shallow marine origin. Moreover the differentiation of the chemistry of this environment and the association of glauconite with transgressive-regressive cycles is discussed.

Key words: glauconite, sedimentary environment, Tertiary

Introduction

Glauconite has great interpretative potential as an indicator of sedimentary conditions. This is due to the fact that this mineral, being formed in the superficial facies of marine deposits, is sensitive to physical-chemical changes of the sedimentary environment. Among other things it can be used as an indicator of glauconitic sedimentary facies.

In order to draw sedimentological conclusions on the basis of glauconite, complex analytical studies as well as knowledge of its physical and chemical features are required.

The latest theories on glauconite origin assume that the most essential factor of the glauconitisation process is the environment. It is a marine environment of specific physical-chemical parameters. The most important physical factors are water depth, water temperature, rate of sedimentation and climatic conditions. The geochemical environment is characterised chiefly by oxidising-reducing potential (Eh), pH, presence of principal chemical elements (Si, Al, Fe, K) and organic matter. The most favourable environment for glauconite formation is a shallow marine one with limited circulation and distant from the active sedimentation zones.

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A necessary condition for glauconite formation is slow or medium rate of sedimentation. This condition together with other favourable environmental ones occur during marine transgression or regression. Large concentrations of glauconite as a rule accompany transgressive or regressive series. Alpine orogenesis is an example of where such a process caused the deposition of large, broadly extended but temporally limited (the Cretaceous and Tertiary) glauconitic series as a consequence of physical and chemical conditions of environment changes. Large amounts of organic matter which originated from the inundated continental areas during the marine transgression can also create favourable conditions for glauconite formation. Although climatic conditions do not seem to be specific for glauconite formation, a hot climate promotes glauconitisation because the continental area liberates great amounts of Fe. Therefore the most active glauconite formation in modern sediments occurs in the tropical zone (Giresse et al. 1987). The formation rate of glauconite both in older and younger geologic epochs converges with the climatic optimum, transgression evolution and weathering zones (Nikolajeva 1977).

Among geochemical factors of the glauconitisation process negative Eh, adequate pH value and Fe supply play a principal role. The most favourable environment for the glauconitisation process appears to be a weakly alkaline (pH = 7–8) and weakly reductive (Eh = 0–200 mV) environment (Fig. 1).

Many authors (Logvinenko 1976, 1980; Odin and Matter 1981) indicate that glauconite develops in a transitory zone: from reductive, littoral sediments to the oxidised ocean bottom ones. The interface between seawater and sediment with oxidising conditions prevailing above and the reducing ones dominating below is a transitory zone for glauconite formation.

The association of glauconite with organic matter is well known. Organic matter creates a favourable microenvironment for mineralogical transformations and contributes to the formation of an adequate Eh of environment or to the degradation of parent layer silicates. The organic matter creating the reductive and acid conditions is favourable for the formation of the preliminary structure of glauconite whereas crystals of mineral glauconite crystallise in a weakly reducing and alkaline environment (Odin and Létolle 1980).

Iron is a fundamental chemical element in the glauconisation process. It is supplied to the sea in colloidal form by rivers or as an authentic juvenile iron by underwater volcanic exhalations. In the near-shore zone of a sedimentary basin rapid deposition of molecular and colloidal iron by means of gravitation occurs. A great amount of Fe brought from onshore becomes immobilised in the vicinity of estuaries of rivers: due to the high accumulation rate it has no opportunity to react with seawater (Fig. 2). However, because of the presence of organic matter or H_2S in sediment the Fe is reduced to Fe^{2+} and becomes soluble Fe. This explains the phenomenon of a greater concentration of Fe in interstitial waters of marine sediments than in the superficial ones. Moreover, this iron precipitates in sediment as FeS_2 or migrates to marine water. The

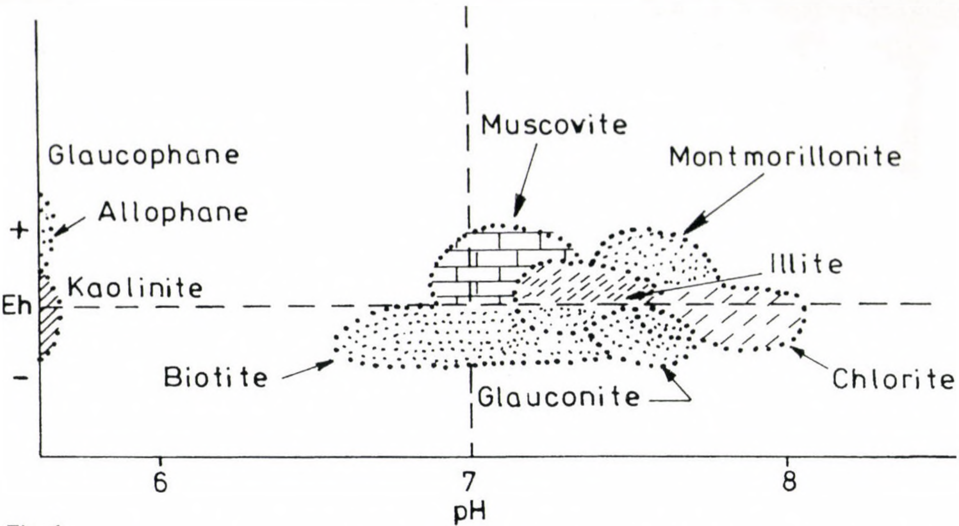

Fig. 1
Stability diagram of clay minerals and mica dependence on EhpH after Dapples (Nikolajeva 1977)

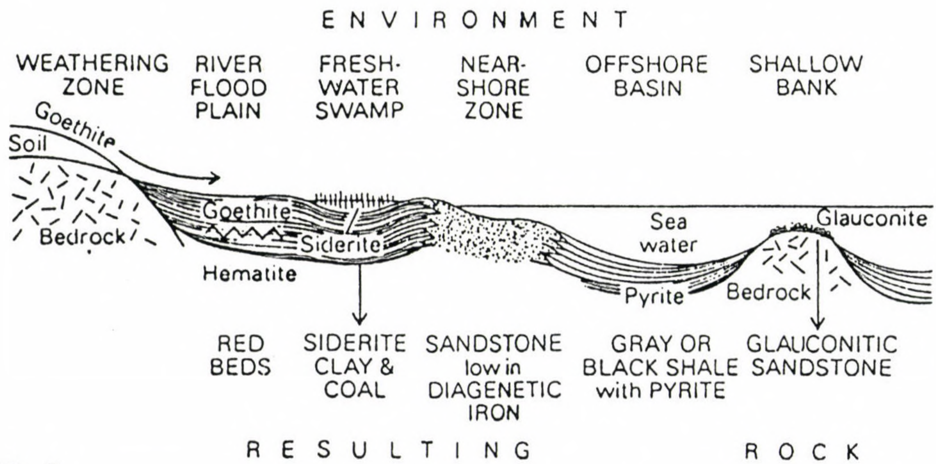

Fig. 2
Idealised cross-section illustrating sedimentary environment where various diagenetic iron minerals are most likely to form. Detrital limonite and hematite are carried mainly as colloids or adsorbed coatings on fine-grained minerals and thereby are separated, by water turbulence in the near-shore zone, from sand grains poorer in reactive iron. This is why shales contain, on the average, more iron than do sandstones, according to Carroll (1975)

glauconite formed on the continental shelf is supplied chiefly by continental Fe, whereas juvenile Fe supplies the glauconite developed on the ocean floor.

Glauconite is formed from diagenetic solutions in the early stage of diagenesis, soon after deposition of the sediment. These conditions must be stable over a longer period of time in order to produce a significant amount of glauconite.

Palaeoenvironmental diagnostics using glauconite

The genesis of glauconite and its physical, mineralogical and chemical features form the source information about the sedimentological environment of glauconiferous sediments. The genesis of authigenic glauconites provides general information about the marine environment of glauconiferous sediments and source areas. On the other hand the variability of the chemical composition of glauconite and the heterogeneity of its chemical character explain the diversity of geochemical environments and postdepositional processes. The pellets of authigenic glauconite and especially the analysis of their abundance and morphological variability can provide valuable information about general conditions of the environment. Environmental factors such as energy conditions, sediment transportation and rate of deposition count among those that influence morphological differentiation of authigenic glauconite grains. Consequently, authigenic glauconite pellets can also be utilised with a high degree of certainty as an important indicator of sedimentological environment.

Glauconite is a sensitive indicator of marine environment conditions because it is formed in exactly determined physical-chemical conditions of this environment (Stoch 1974). Clay minerals and glauconite among them are sensitive to the influence of the environment chemistry because under its influence they undergo alterations; for this reason they can be utilised as environmental indicators (Zchus 1966; Stoch 1969; Wiewióra 1973). The chemical composition of glauconite among other things depends on the facies conditions under which it is formed. Russian scientists, such as Gorbunova (1950), Fienoszina (1961), Nikolajeva et al. (1971) and Kazalov (1982) showed that glauconite can be utilised as a typical indicator of facies. This is proved by the variability of the chemical composition of glauconite and especially of the percentages of Fe_2O_3, Al_2O_3, SiO_2, MgO, K_2O and Na_2O. The more the sedimentation is of deep marine character the poorer glauconite is in Fe and the richer in Al. The Fe_2O_3 content in glauconite diminishes together with the distance from the source area from 22% in inshore facies to 14.4% in pelagic facies. Al demonstrates an opposite tendency than Fe. The content in Al_2O_3 increases together with the distance from the source area from 9.45% in the inshore area to 12.68% in the pelagic zone, while the K content decreases in this direction from 6.80 to 5.16% respectively. The SiO_2 content demonstrates a regular increase from 48.11% in the inshore facies to 55.25% in the pelagic one.

Deep-water glauconite reflects an environment impoverished in Fe as a result of its assimilation in pyrite and marcasite which occurs in clays enriched by organic matter. Decreased contents of Fe in glauconites are compensated by the increase of Al, free SiO_2 admixture and H_2O.

The chemical environment in the sedimentation process and the diagenesis of glauconite are closely associated with the chemical composition of underlying rocks, which in turn depends on the facies of the marine deposits being formed.

The Mg contents in glauconite can be a palaeotemperature indicator of marine waters. The decrease of Mg in glauconite can be explained by low temperature of sedimentation basin water (Nikolajeva 1981) and the stable contents of this element in glauconite in various places of its occurrence can be considered to be an important indicator of marine environment uniformity (Carroll 1975).

The diameter of glauconite grains indicates the depth of the sedimentation environment. The smaller the grains are the farther they are from the continental coast area where they were formed (Nikolajeva 1977).

Apart from being useful in analysing palaeoenvironmental problems glauconite can be also used to explain other geologic phenomena such as general geochemical and oceanographical conditions (Smulikowski 1953), the interpretation of tectonic problems (Wiewióra 1973), orogenic problems (Kohler 1976) and palaeogeographic problems (Burst 1958; Wermund 1961; Logvinenko et al. 1975; Kohler 1980; Nikolajeva 1981; Bornhold and Giresse 1985).

Geologic settings

The study area comprises the upland country consisting of the Lublin Upland, its North Foreland and the Roztocze Region (Fig. 3). Within the borders of this area Jurassic, Cretaceous, Tertiary and Quaternary deposits are present on the surface. The Upper Cretaceous, represented by gaizes[1], opokas[2], marls and marly limestones is of primary significance for the mentioned area because nearly all the Lublin Upland rests upon the deposits of this age.

The Tertiary deposits of the Lublin region are very diversified with regard to facies. They begin with Middle Danian deposits that lie on the corroded, hardened surface of Upper Maastrichtian age (hard-ground). Palaeogene and the Neogene deposits occur on the surface or beneath a very thin cover of Quaternary deposits in the central part of the region, comprising the zone from the Vistula valley in the west to the town Chelm and the neighbouring region

1 gaizes – polish local term denoting organogenic sedimentary rock intermediate between siliceous and terrigenous rocks, containing organogenic silica, detrital quartz, clay minerals, calcium carbonate and frequently glauconite and phosphates

2 opokas – polish local term denoting a siliceous rock with calcium carbonate

Fig. 3
Study area on the background of a map of Poland

in the east. North of this belt the deposits are known mostly from drilling and rarely appear on the surface.

In the Lublin Tertiary only a few stratigraphic levels are represented and in a very incomplete section, mostly marine, practically without any terrestrial sediments. They consist of sands, sandstones, gaizes and limestones. The last-named lithology can be found in the south and they are considered to be more a part of the Roztocze Area.

The Lublin Tertiary glauconitic deposits have interested geologists for many years but only recently has there been a great improvement of the knowledge about glauconite from the Cretaceous–Tertiary boundary in the Chelm Hills region (Gazda et al. 1992), in the ravined Vistula section in the Nasilów area (Krzowski 1993) and in the Roztocze Area (Buraczynski and Krzowski 1994). The development resulted from the attempts to apply glauconite data to the interpretation of the sedimentary environment in the Upper Cretaceous and the Palaeocene (Chelm Hills) and to the analysis of the stratigraphic–palaeogeographical problems of the Eocene (Roztocze Upland). Krzowski (1990, 1991, 1993) and Halas et al. (1990) were among the researchers who promoted the possibility of using glauconite for the geochronological studies of Tertiary deposits in the Lublin region.

Methods applied to the research

Samples of various Tertiary lithological deposits (sands, gaizes, sandstones, glauconitites) of various ages (Palaeocene, Eocene, Oligocene, Miocene) were used in the present study on glauconite. Exposures and cores of boreholes were sampled.

Glauconite was separated electromagnetically by means of a Franz separator, after which the grains of accompanying minerals were removed by hand with the use of a binocular microscope.

In order to minimise mineralogical and chemical heterogeneity of glauconite it was subjected to density separation in a solution of bromoform with methyl alcohol.

The following physical features of glauconite were analysed: colour, grain morphology and specific density.

Glauconite samples representative in respect to grain size (grain fractions from 0.20 to 0.10 mm) and specific density (density fractions from 2.3 to 2.8 g/cm^3) were selected for chemical analysis.

The determination of chemical elements was carried out on dried samples of glauconite at a temperature of 105 ^{o}C.

Each sample was homogenised by rubbing in ethylene alcohol. Along with the tested samples of glauconite the French standard (biotite) of chemical composition was analysed. The SiO_2 content was determined by weight and the Fe_2O_3 and Al_2O_3 content by spectrometer. The K_2O content was measured by flame photometry.

The results of the chemical analysis of the tested glauconites showed their chemical heterogeneity resulting from their mineralogical diversity.

Sedimentological characteristics of the Tertiary basins

In the Tertiary period favourable conditions for the formation of glauconite prevailed. Such conditions as transgressive and regressive cycles of the marine basin, abundant supply of terrigenous material, favourable climatic conditions, adequate bathymetry of sea basins and a slow rate of sedimentation promoted the glauconisation processes. From among the Tertiary glauconitic formations the Palaeogene and especially the Eocene ones are the most extensive in the world. Palaeogene glauconite occurs on practically the all the continents of the world on the platform as well as in the geosynclines. The range of glauconite occurrence in the Neogene sediments, which is distinctly associated with peri-ocean areas near ancient continents, is also wide.

Favourable palaeogeographic–environmental conditions also existed in the Tertiary period in the territory of Poland. In the Lublin region glauconitic sediments from this period are known from the Lower Palaeocene, the Upper Eocene and the Lower Oligocene. The glauconite of the Miocene period is of an allogenic character and was redeposited from older sediments, especially Eocene ones.

Palaeocene basin

Generally the Palaeocene basin is considered to be the final evolutionary stage of the extensive Upper Cretaceous basin. The shallowing of the sea together with the increased inflow of terrigenous material favoured the formation of glauconite the amount of which grows proportionally to the increase of the sand content in the sediment.

On the basis of sedimentary facies formation and studies of glauconite within the Palaeocene basin one can observe its bathymetric differentiation among the western, central and eastern parts of the region. This differentiation is emphasised by facies type and glauconite contents in the sediment. The prevailing glauconitic facies are glauconitic sands and sandstones with phosphorites in the western part, glauconitic gaizes in the central part and glauconitites in the eastern part. Facies changes were accompanied by the differentiation of physical-chemical conditions of sedimentary environment which is proved by the variation of glauconite contents in the profiles: from 13% in the western part, 21–39% in the central part to 63–75% in the eastern part of the basin. The highest contents of glauconite (i.e. where the most favourable conditions for its formation were located) can be found in the eastern platform area of Chelm.

The physical and geochemical features of glauconite indicate the shallow-water character of the Palaeocene basin (Table 1). Among the physical features of glauconite, the predominant coarse grain fraction ($\phi > 0.10$ mm) and the spherical shape of the grains testify to the shallow-water character of the basin, since in shallow-water sediments only a small variety of morphological forms of glauconite can be found. Hein et al. (1974) give three fundamental forms (ovoidal, spheroidal and tabular) of glauconite grains characteristic of a shallow-water marine environment. The greatest number (up to 10) of morphological forms of glauconite grains can be found in the sediments of the continental slope and canyon ridges. Among the chemical features of glauconite high contents of Fe_2O_3 tot. and K_2O indicate the shallow-water character of the Palaeocene basin as well as a low rate of sedimentation in the basin.

The climate of the Palaeocene can be described as subtropical (Liszkowski 1970) which is confirmed by phosphorite facies accompanying glauconite in the western area and palaeotemperatures of the basin waters (+16.8; +17.4; +18.8 °C) determined on the basis of isotopic studies of Palaeocene fossils (Krzowski 1995). On the other hand a low-energy marine environment and

Table 1
The facies features of Palaeocene glauconites

Ordinal	Sample number	Tested area	Physical features of glauconite: Predominant colour: bright green (+) dark green (-)	Physical features of glauconite: Grain size: >0.10 mm (+) < 0.10 mm (-)	Physical features of glauconite: Morphology of grains: simple (+) varied (-)	Geochemical features of glauconite: Fe_2O_3 tot. (weight %)	Geochemical features of glauconite: Al_2O_3 (weight %)	Geochemical features of glauconite: SiO_2 (weight %)	Geochemical features of glauconite: K_2O (weight %)
1.	NA 1	western part	-	+	+	22.30 +	6.12 +	54.30 -	5.98 +
2.	PI 1	central part	-	+	+	23.00 +	4.67 +	48.90 +	8.45 +
3.	GP 1		-	+	+	22.50 +	4.68 +	59.25 -	10.00 +
4.	SG 1	eastern part	+, -	+	+	24.50 +	4.63 +	51.00 +	7.26 +
5.	OC 1		-	+	+	24.00 +	4.51 +	49.60 +	9.08 +

+ inshore facies
- pelagic facies

poly-stages of the glauconisation process are indicated by the lack of density selection of glauconite from the western and the central areas.

The following phenomena show the geochemical differentiation of the sedimentary environment: the occurrence of microcrystalline aggregations of iron sulfides (pyrite) within the cement of glauconitic gaizes and of glauconite grains in the central part thereof, which was not found in the other areas.

Glauconitic sediments were subjected to diagenetic processes which was necessary for the very glauconitisation process.

High contents of SiO_2 in the glauconites indicates an advanced degree of silicification processes in the Palaeocene gaizes.

Secondary alterations of the Palaeocene glauconites suggest the activity of postdepositional processes, chiefly weathering.

The open marine character of the Palaeocene basin is indicated by the results of isotopic studies of the oxygen and carbon of fossils of that period (Krzowski 1995).

Eocene basin

In the Upper Eocene particularly favourable environmental conditions for glauconite formation prevailed. The transgressive basin character, a low sedimentation rate, the supply of terrigenous material from the land and a warm climate contributed to the phenomenon, since in that period a global climatic optimum occurred. The climate in Europe in the Eocene period was warm and wet, of a tropical or subtropical type with a mean annual temperature of about +22 ^{o}C. In the territory of Poland the climate was also tropical. The studies of palaeotemperatures on the basis of stable oxygen and carbon isotopes of fossils (Krzowski 1995) confirmed the Upper Eocene climatic optimum (+23.8 ^{o}C) as well as the open marine character of the sedimentation basin.

The facies distribution of the sediments and the diagnostic features of the glauconite included in them show that the Eocene deposits in the Lublin region are of shallow marine origin (Tab. 2). These sediments are fine-grained, grey-green, frequently muddy and weakly marly quartz-glauconitic sands, locally with well-rounded grains of gravel and lydites, dusty-sandy muds and claystones and sometimes grey-green and olive-coloured clays. The sediments frequently contain numerous, small-sized phosphorite concretions and amber fragments.

Poor sorting of glauconitic sands and the lack of a distinct density separation of the glauconite indicate a low-energy sedimentary environment and weak bottom currents. The varied heavy mineral compositions (Roztocze Area) of the sediments suggest different sediment sources and different depths in the sedimentation basin. The lack of phosphorites in the Roztocze sediments proves the differentiation of sedimentological conditions within Eocene basin.

Epeirogenetic movements of the Pre-Pyrenean and Pyrenean phases had a decisive effect on the type and kind of sedimentation and consequently upon

Table 2
The facies features of Eocene glauconites

Ordinal	Sample number	Tested area	Physical features of glauconite: Predominant colour: bright green (+) dark green (-)	Physical features of glauconite: Grain size: >0.10 mm (+) < 0.10 mm (-)	Physical features of glauconite: Morphology of grains: simple (+) varied (-)	Geochemical features of glauconite: Fe_2O_3 tot. (weight %)	Geochemical features of glauconite: Al_2O_3 (weight %)	Geochemical features of glauconite: SiO_2 (weight %)	Geochemical features of glauconite: K_2O (weight %)
1.	SP 1	central part	+, -	+	-	20.70 +	9.15 +	47.40 +	6.96 +
2.	ŻU 1	central part	+, -	+	-	20.00 +	7.71 +	53.00 -	6.46 +
3.	CH 1	eastern part	+, -	+	-	17.80 -	9.43 +	55.50 -	6.22 +
4.	JA 2	eastern part	+, -	+	-	20.70 +	7.76 +	49.90 +	6.49 +
5.	LE 1	eastern part	+, -	+, -	-	16.00 -	14.40 -	46.60 +	8.99 +
6.	SI 1	northern part	+, -	+	-	19.00 +	6.09 +	51.12 +	8.59 +
7.	ŁA 1	Roztocze area	+, -	+	-	20.40 +	6.96 +	50.90 +	7.38 +
8.	PE 2	Roztocze area	-	+	-	21.90 +	7.94 +	47.00 +	7.92 +

\+ inshore facies
\- pelagic facies

glauconite formation in the Eocene (Henkiel 1984). Secondary alterations of glauconites confirm the existence of a weathering zone in the Eocene which developed in the Danian gaizes as well as in the Maastrichtian marls.

Oligocene basin

Sands, glauconitic muds and rarely clays are characteristic for the Oligocene in the Lublin region. The sands containing considerable amounts of glauconite represent the sedimentation associated with the shoreline changes, i.e. marine transgression or regression (Kosmowska-Ceranowicz 1979).

A climatic minimum falls at the Palaeocene / Neogene transition which was marked particularly distinctly in Central Europe. At the Eocene / Oligocene boundary a violent cooling of the climate occurred and the temperature dropped to +10 °C and in places even to +5 °C (Pozaryska and Odrzywolska-Bienkowa 1978). That temperature minimum can be associated with general changes in the palaeogeography of the European continent.

Glauconite was formed in abundance at the end of the Oligocene period despite not very favourable climatic conditions. Other favourable conditions positively influenced the process, namely a rich supply of terrigenous material from the continent, a shallow sea and a low sedimentation rate. In spite of great weathering of the Oligocene glauconites and the changes in their chemical composition the predominant number of its diagnostic features confirm the shallow-water character of the Oligocene basin (Table 3) and its highly homogeneous density can testify to its having been washed, possibly in a beach environment.

The alterations of glauconites attest to the weathering processes following the uncovering of the Oligocene sediments.

Conclusions

1. Authigenic glauconite grains can be utilised as an important indicator of sedimentary environment. Valuable information on sedimentary environment conditions can be obtained on the basis of the analysis of the abundance and morphological variability of glauconite grains, their chemical composition and their structure. Although the main reason for the morphological differentiation of glauconite grains appears to be the great diversity of the source material (substratum) in the marine environment during glauconite formation this does not exclude the effect of the interaction of bottom currents that facilitated ion exchange between glauconite grains and sea water. Therefore the shape of glauconite grains can be indicative of the energy conditions (hydrodynamics) of the marine basin.

2. On a global scale glauconitic facies always formed in the open marine environment. This environment excludes, among other things, the influence of estuarine zone chemistry where instead of glauconitisation the process of berthieritization occurs. Therefore the very occurrence of authigenic glauconite in the sediments can provide valuable information on the character of this environment.

3. Glauconitisation processes indicate low-energy environments of sedimentation and diagenesis, weathering processes of surrounding onshore areas

Table 3
The facies features of Oligocene glauconites

Ordinal	Sample number	Tested area	Physical features of glauconite: Predominant colour: bright green (+) dark green (-)	Physical features of glauconite: Grain size: >0.10 mm (+) < 0.10 mm (-)	Physical features of glauconite: Morphology of grains: simple (+) varied (-)	Geochemical features of glauconite: Fe_2O_3 tot. (weight %)	Geochemical features of glauconite: Al_2O_3 (weight %)	Geochemical features of glauconite: SiO_2 (weight %)	Geochemical features of glauconite: K_2O (weight %)
1.	DR 1	western part	+,-	+	-	26.00 +	9.37 +	41.70 -	7.36 +
2.	KR 1	eastern part	+,-	-	-	19.20 +	9.10 +	48.90 +	8.78 +

\+ inshore facies
\- pelagic facies

and orogenic movements (marine transgressions or regressions). Consequently glauconitic formations can be associated with eustatic sea level lowering or with the domination of epeirogenetic processes in the continental areas.

4. Precise grain descriptions and density separation of glauconite can indicate the activity of oscillatory currents. Conversely, the lack of distinct glauconite density separation can indicate several stages of the glauconitisation process and the low-energy environment of its formation.

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Calcite mineralization in the Holy Cross Mts., Poland; the present state of knowledge

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This report presents the results of lithologic, petrographic, isotopic and chemical investigation performed on calcites from the Holy Cross Mts. (Fig. 1). In general, 5 phases, i.e., A – Variscan (Visean/Namurian), B – older post-Variscan (post-Namurian through pre-Zechstein), C and D – younger post-Variscan (Permian/Triassic and middle/late Early Triassic), and E and F – Cimmerian-Alpine (Late Jurassic and Late Cretaceous), have been distinguished. This division has been confirmed by the highly diverse $\delta^{13}C$ and $\delta^{18}O$ values of these calcites clustered in 4 populations. The "rose-like" calcite (phases C and D) are most common. The calcites from the Holy Cross Mts formed under different geologic conditions: from a typical marine off-shore zone (phases E+F), through marine relict basins originated at the end of Variscan movements (phase A), to a terrestrial environment (phase B), in the final stage with strongly developed karstic processes (phases C+D). Determinations of oxygen isotope composition and homogenization temperatures of gas-liquid inclusions in the same calcite crystals indicated that the hydrothermal fluids were of marine provenance.

Key words: geology, petrography, chemical and stable isotope analyses, Holy Cross Mts, Poland

Introduction

Various geologic and mineralogical aspects of the calcite mineralization in the Holy Cross Mts have been discussed in many publications and archival reports (including Czarnocki 1953; Fijalkowska and Fijalkowski 1973; Migaszewski et al. 1987; Rubinowski 1970, 1971; Wrzosek and Wróbel 1961). A breakthrough in the problem of calcite mineralization was made by including into the scope of investigation more sophisticated analytical methods, i.e., stable carbon, oxygen, sulfur and strontium isotope determinations, homogenization temperature measurements of gas-liquid inclusions, REE and trace element analyses using inductively coupled plasma-atomic emission spectroscopy and neutron activation methods (Migaszewski 1994; Migaszewski et al. 1994, 1995, 1996a, b, c).

This report summarizes all the relevant data presented in the references cited above, and other unpublished results of investigation performed thus far.

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Geologic setting and local history

The Holy Cross Mts are located in south-central Poland (Fig. 1). They take up an area of about 65 x 35 km. The tallest mountain (£ysica) reaches 612 m. The Holy Cross Mts are made up of a Paleozoic block and the Permo-Mesozoic cover. The latter surrounds an exposed mountain core from the north, west and southwest. The Holy Cross Mts are the only prominent Paleozoic inlier abutting on the Tornquist–Teisseyre Zone between Scandinavia and Romania. The most striking tectonic element is the Great Holy Cross Mountain Fault (WNW–ESE) which divides the entire region into two structural-paleogeographic units, i.e., the £ysogóry (northern) and the Kielce (southern). The former is characterized by a nearly continuous development of sediments spanning the Cambrian through the Upper Devonian (Lower Carboniferous?). On the other hand the Cambrian–Lower Carboniferous profile of the Kielce unit contains numerous stratigraphic gaps.

The lower portion of the Paleozoic profile (Cambrian through Lower Devonian) is developed primarily as terrigenous sediments, whereas the upper (mostly Middle and Upper Devonian) section is composed of carbonate rocks topped by siliceous-clayey-terrigenous sediments (Lower Carboniferous). The Permo-Triassic sequences are generally carbonate-terrigenous, the Lower Jurassic clayey-terrigenous, and the Upper Jurassic and the Upper Cretaceous deposits primarily carbonate. Tertiary sediments composed of claystones, sandstones, limestones and gypsum abut on the southern margin of the Holy Cross Mts.

The Holy Cross Mts were formed as a result of polyphasic tectonic movements that took place in the Caledonian (Cambrian, late Tremadocian, Ludlovian/Přidolian and late Emsian) and Variscan (Frasnian/Famennian?, late Visean–early Namurian) sedimentary–diastrophic epochs (Kowalczewski 1981; Znosko 1984, 1988). In the study area Alpine orogeny was manifested only by block-faulting. Different views prevail on the origin of the Great Holy Cross Mountain Fault. According to Lewandowski (1993), it is a Late Variscan strike-slip fault along which the Kielce terrane shifted from the present Crimea Peninsula toward the northwest, colliding with the nearly autochthonous £ysogóry terrane.

Igneous rocks are scarce and include only diabases and lamprophyres. They pierce Cambrian and Lower and Middle Devonian sediments. It should be emphasized here that the lamprophyres occur only in the Kielce region. Both lithotypes seem to have been derived from the same magma chamber; the lamprophyres presumably represent a contaminated variety of diabases (£abęcki 1970; Ryka 1974).

Exposed Cambrian and Devonian rocks occupy most of the study area. Scattered and veined sulfide and barite mineralization occurs primarily within Devonian, Upper Permian, Lower and Middle Triassic, and in places in Cambrian sediments (Rubinowski 1970, 1971; Migaszewski et al. 1995, 1996a).

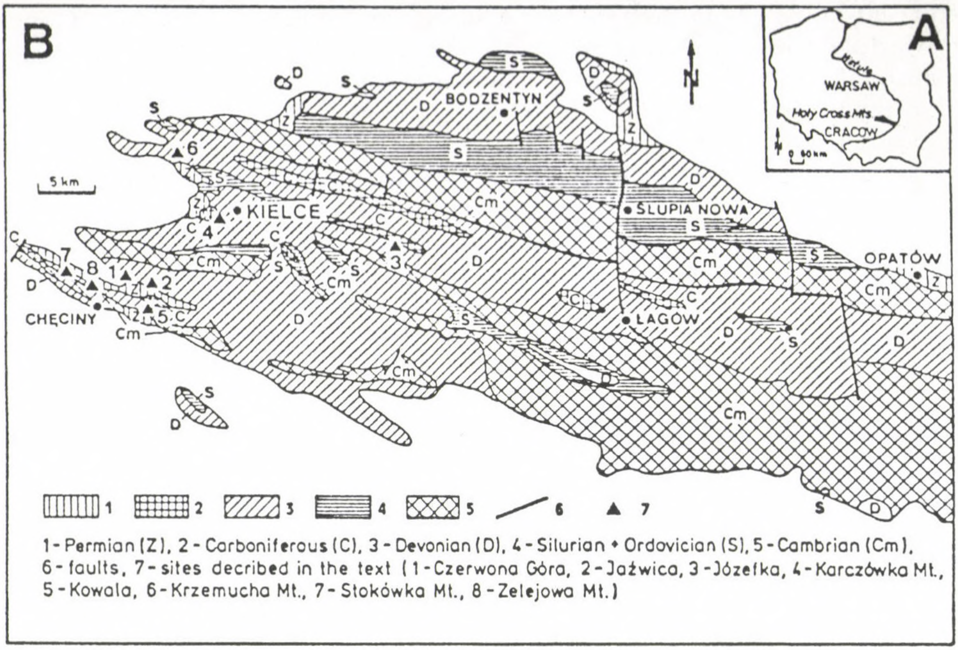


Fig. 1.
Geologic sketch map of the Holy Cross Mountains with sites discussed in the text

Field work indicated only a scarce number of calcite veins in the section spanning the Cambrian to and including the Lower Devonian. In general, they seem to be connected with younger, primarily Permo-Triassic phases. Individual calcite veins and lenses partly linked to synsedimentary hydrothermal dolomitization (Migaszewski 1990, 1991) appear in Middle Devonian carbonate sediments. In the western part of the Holy Cross Mts, some of the Frasnian, Famennian and Tournaisian rocks are strongly cut by calcite veinlets and lenses reaching 4 cm in thickness. In these veinlets quartz occurs in subordinate amount. Calcite is white and somewhat pink making up at least three generations. This calcite mineralization is linked to the Variscan cycle.

The lack of sediments spanning the lower Namurian to and excluding the Zechstein, enables precise dating of early post-Variscan calcite mineralization that forms veinlets only within some pebbles of Zechstein conglomerates, or locally (Krzemucha quarry) dissecting Variscan veinlets that in turn pierce the Givetian/Frasnian rocks.

The principal ("różanka" or "rose-like") calcite mineralization is closely linked to deep-rooted faults generally trending N–S. The dip is steep, varying from 75 to 85^{o}. White to red calcite veins are very thick reaching as much as 42 m (Fijalkowski, unpubl. data). They are concentrated primarily in Devonian and Zechstein carbonate deposits of the western part of the Holy Cross Mts; however, their presence has been confirmed in the entire Paleozoic block and adjacent areas, i.e., in the northern and western Permo-Mesozoic cover, and toward the west and southwest, in the Silesian–Cracovian region. The striking feature of this calcite is a varied content of hematite and colloidal iron oxides and hydroxides reaching 30% Fe (Wróblewski, unpubl. data), as well as inner deposits (fills) of karstic origin.

The sediments younger than the Lower Triassic are poorly mineralized. Calcite fills have been recorded only in the form of veinlets and lenses primarily in the Upper Jurassic (Oxfordian) and Upper Cretaceous (Turonian) sediments (Migaszewski et al. 1987). They are white or yellowish and lack iron oxides and hydroxides, and Cu-, Pb- and Zn-sulfides; in addition, the Cretaceous thermal features are associated with cherts.

As mentioned above, due to the vast extent, abundant reserves and decorative properties, the "różanka" ("rose-like") calcite was widely used as a masonry stone during the Renaissance, Mannerism and Baroque periods. It can be found in many churches, cloisters, castles and palaces of southern Poland being often combined with another decorative stone, the Dębnik black limestone.

Methods

The field work including lithologic and sedimentologic study as well as sampling was carried out in 1993–1995. The structural-textural investigation was performed on hand specimens and polished slabs and microscopic examination was conducted on thin slides and polished sections. Most of these specimens were stained with Evamy's solution which enabled distinguishing calcite and dolomite and also determining the spatial distribution of Fe^{2+} in these two minerals (Migaszewski and Narkiewicz 1983). In addition, to better identify different calcite generation samples were exposed to ultraviolet radiation.

The homogenization temperature measurements of gas-liquid inclusions in calcite crystals were performed on a microscopic heating stage with an accuracy of ±2 °C (Karwowski, Silesian University in Sosnowiec). In order to calibrate two temperature scales, i.e., the oxygen isotope (Tox) and the homogenization (Th), and to determine the $\delta^{18}O_{SMOW}$ of hydrothermal fluid some of the samples were taken for parallel homogenization and isotopic measurements (Fig. 3 in Migaszewski et al. 1996b).

Isotope ratio measurements were performed on over 400 calcite, carbonate rock, sulfide and barite samples. They included carbon, oxygen, sulfur and strontium isotopes. The carbon and oxygen isotopes were determined on CO_2 according to McCrea's procedure (1950) using a modified mass spectrometer MI-1305. The sulfur isotope was determined on SO_2 derived from oxidizing large sulfide crystals with Cu_2O at 950 °C or from barite crystals according to Hałas and Skorzyński's method (1981). The $^{87}Sr/^{86}Sr$ ratio was measured on calcite using a mass spectrometer VG Sector 54 (Burchart, Geochronology Laboratory, Polish Academy of Sciences in Warsaw). The $\delta^{13}C$ and $\delta^{18}O$ on

PDB scale were determined with an accuracy (1σ) of ± 0.07‰, the $\delta^{34}S$ on CDT scale with ± 0.10–0.12‰ and the $^{87}Sr/^{86}Sr$ ratio with ± 0.000009–0.000013.

The carbonate samples were tested for 19 major and trace elements, i.e., Ag, Al, Ba, Ca, Co, Cr, Cu, Fe, K, Mg, Mn, Mo, Na, Ni, Pb, Sn, Sr, V and Zn using flame emission spectrometry (spectrometer PGS 2), inductively coupled plasma-atomic emission spectrometry (spectrometer Jobin-Yvon model JY 70 PLUS), and Flame Atomic Absorption Spectrometry (spectrometer PU 9100 X UNICAM) methods (Szczecińska, Holy Cross Mountain Div. Polish Geological Institute in Kielce, Paslawski, Central Chemical Laboratory, PGI in Warsaw).

In addition 15 calcite samples were analyzed for 8 rare earth elements (Hoffman, ACTLABS in Ancaster, Canada).

Mineralogical, geochemical and isotopic characteristics of the Holy Cross Mountain calcite

Variscan calcite

Most fully developed Variscan calcite mineralization occurs in Krzemucha quarry (Fig. 1). Calcite veinlets and lenses pierce in all directions tightly folded, thin and medium-bedded, marly and bituminous limestones assigned to the upper Givetian/lower Frasnian. Based on the lithologic–petrographic investigation, three Variscan calcite generations have been identified here. They are as follows:

1) fine crystalline calcite, light grey with pyrite and marcasite concretions;

2) micro and fine crystalline calcite, grey ("quartz-like"), in places light grey, semitranslucent. Pyramidal walls of calcite crystals are covered with a black matter resembling asphaltite; locally individual quartz crystals with a hexagonal habit occur (Salwa 1995); copper sulfide inclusions are scarce;

3) vari-crystalline calcite, light grey, locally grey and yellowish, semitranslucent. It is commonly associated with milky, smoky and rocky quartz crystals reaching as much as 3 cm in length; a dark brown color of smoky quartz is derived from asphaltite inclusions (Salwa 1995). In places, the calcite is accompanied by white and pinkish barite, brown saddle dolomite, and pyrite; in addition, small inclusions of iron oxides and hydroxides, and asphaltite also occur.

Individual veinlets contain several calcite generations that indicate many phases of fissuring accompanied by hydrothermal fluid injections.

Another interesting site is Mt. Zelejowa (Fig. 1) widely known for extracting "różanka" calcite on a large scale in the 16–17th Centuries. Variscan calcite veinlets cut by huge post-Variscan "różanka" calcite veins have been recorded here.

Exposed to ultraviolet radiation, the Variscan calcite emits white, blue and violet light. The calcite contains only a small amount of magnesium (0.10 to 6.78%). The content of many trace elements is linked to tiny inclusions of scattered sulfides. The chemical composition of the calcite described has been presented in Tables 1 and 2.

The $\delta^{13}C$ of the Variscan calcite varies from –1.73 to 2.27‰ and the $\delta^{18}O$ from –19.37 to –4.19‰ (Table 1 in Migaszewski et al. 1996b). The $\delta^{18}O$ values indicate respectively a temperature of 168–37 °C assuming that the water in equilibrium with calcite had a $\delta^{18}O$ value of 0‰ (Bowen 1988; Epstein et al. 1953). The Variscan calcite is typified by the prevailing positive $\delta^{13}C$ values similar to those in host rocks, i.e., the Devonian marine limestones (0.31 to 3.23‰). The isotopic composition of strontium ($^{87}Sr/^{86}Sr$) ranges from 0.709325 ± 9 to 0.709480 ± 13 (table 1 in Migaszewski et al. 1996b) which can indicate sea water as the basic constituent of hydrothermal fluids (Hofmann, and Hart 1978; Wadleigh et al. 1981).

Older post-Variscan calcite

In the Holy Cross Mts no dated sediments spanning the Upper Carboniferous to almost the entire Permian (to the bottom of Zechstein) have so far been found. This is the main reason why it is impossible to determine the exact age and the number of the oldest post-Variscan calcite generations. They consist of calcite pebbles or veinlets piercing only limestone pebbles in Zechstein conglomerates as well as scarce veinlets dissecting Upper Givetian/Lower Frasnian limestones in Krzemucha quarry (Fig. 1). Some of the calcite found in pebbles contains hematite laminae and that from Krzemucha scattered Pb-, Zn- and Fe-sulfides. At the present stage of knowledge it is impossible to determine the spatial and secular relationship between these two generations. Nonetheless the "vein" variety from Krzemucha seems to be older than the "pebble" one.

In general, white, fine and medium crystalline calcite is prevalent. The radiated calcite from Krzemucha gives off uniform deep purple light. It should be stressed here that this calcite reaches the highest concentration of REE (Table 2).

The $\delta^{13}C$ varies from –2.32 to 0.23‰ and the $\delta^{18}O$ from –10.43 to –3.89‰. The oxygen isotope thermometer indicates 71–35 °C (assuming that the water in equilibrium with calcite had a $\delta^{18}O$ value of 0‰), respectively, as a most probable temperature of calcite crystallization.

Younger post-Variscan calcite ("różanka" type)

Based on the results of geologic and petrographic investigation, two principal phases of "różanka" calcite mineralization can be distinguished here.

The older phase (C) is composed of two generations (1 and 2):

1) Laminated ("onyx-like") fine and medium crystalline calcite, white-brown-red, is featured by the presence of wavy, in places tooth-like, Fe- and Mn-rich laminae; some of them contain poorly-rounded detritic calcite

Table 1
The content of selected major and trace elements in vein calcite and related deposits from the Holy Cross Mts. (in ppm)

Location	Calcite (phase/generation)	Ag	Ba	Co	Cr	Cu	Mg	Mn	Mo	Ni	Pb	Sn	Sr	V	Zn
Czerwona Góra	calcite (C/1)	0.1	10	–	–	15	90	~100	–	1	620	–	450	3	5000
	calcite (C/1)	0.1	10	–	–	12	170	~500	–	3	180	–	220	4	890
Chelosiowa Jama Cave	calcite (D/7)	0.1	20	6	1	45	1430	~100	–	1	20	–	320	10	20
	flowstone (Permian)	1	140	50	190	89	150	~1000	10	120	500	1	5	100	690
Kowala	calcitic fill (D/7)	–	70	2	200	60	900	10–100	0.1	45	250	–	350	750	140
	calcite (D/7)	–	1	1	1	9	1360	10–100	–	1	20	–	140	10	15
	calcite (D/8)	0.1	6	–	8	75	360	500–1000	–	2	130	–	10	30	120
	calcite (D/10)	–	1	–	–	15	510	10–100	–	–	2400	5	30	20	1050
Krzemucha	calcite (A/3)	0.5	590	8	10	950	990	500–1000	0.1	13	1	5	350	50	14
	calcite (A/3)	0.1	120	–	5	280	67800	500–1000	1	1	5	7	10	10	28
	calcite (D/7)	–	7	–	5	5	1510	100–500	–	–	2	–	1000	10	10
Rzepka	calcite (D/7)	–	1	–	–	28	750	10–100	–	1	12	–	100	10	25
	calcite (D/7)	–	8	–	1	17	5228	10–100	–	12	10	–	740	20	20
	cave fill (Triassic)	1	–	7	10	140	540	10–100	–	54	710	–	–	100	190
Skala	calcite (D/7)	–	30	–	1	10	3020	10–100	–	4	110	–	310	20	35
Skrzelczyce	calcite (C/D)	–	5	–	5	34	1700	10–100	–	–	1300	–	420	10	170
Stokówka	calcite (D/7)	–	1	–	1	17	2180	100–500	–	1	140	–	170	10	50
	calcite (D/8)	1	10	6	12	260	390	1000–5000	–	32	5900	1	130	80	170
	calcite (D/9)	2	4	1	6	320	240	100–500	–	49	3200	10	10	100	260
Tudorów	calcite (D/7)	0.5	30	–	7	12	1580	~1000	–	–	1	–	110	10	70

NOTE: for locations see Fig. 1

Table 2
The content of REE in vein calcite and related deposits from the Holy Cross Mts (in ppm)

Location	Calcite (phase/gener.)	Ce	Eu	La	Lu	Nd	Sm	Tb	Yb
Czerwona Góra	calcite (C/1)	1	0.06	0.9	0.01	< 1	0.20	< 0.1	0.10
	calcite (C/2)	< 1	< 0.05	0.3	< 0.01	< 1	0.08	< 0.1	< 0.05
Kowala	calcitic fill (D/7)	13	0.21	7.3	0.04	6	0.86	0.1	0.33
	calcite (D/7)	3	0.13	2.5	0.03	3	0.55	0.1	0.24
	calcite (D/8)	5	0.19	2.3	0.04	3	0.82	0.2	0.32
	calcite (D/10)	< 1	< 0.05	0.3	< 0.01	< 1	0.07	< 0.1	0.05
Krzemucha	calcite (A/2)	2	0.10	0.9	0.01	1	0.28	< 0.1	0.09
	limestone	27	0.82	16.7	0.23	16	3.25	0.6	1.55
	calcite (B)	36	2.30	7.4	0.27	36	8.64	1.3	1.90
	calcite (D/7)	3	0.08	1.2	0.04	2	0.28	< 0.1	0.24
Skala	calcite (D/7)	1	< 0.05	0.4	< 0.01	< 1	0.09	< 0.1	< 0.05
	calcite (D/8)	13	0.36	5.8	0.08	8	1.58	0.3	0.58
Stokówka	calcite (D/7)	1	0.05	0.8	0.01	1	0.14	< 0.1	0.07
	calcite (D/8)	13	0.54	4.6	0.05	9	1.85	0.3	0.44
	calcite (D/9)	11	0.45	3.3	0.06	7	1.52	0.3	0.56
Tudorów	calcite (D/7)	4	0.06	2.8	0.01	2	0.23	< 0.1	0.07

NOTE: for locations see Fig. 1

grains. This calcite forms veins piercing only matrix (not pebbles) of Zechstein conglomerates. Both the structural position, flowstone appearance and presence of other related morphologic features indicate that the laminated calcite precipitated in the early-diagenetic stage of conglomerate development (apparently late in Zechstein or early in Triassic time) under karstic conditions. This conclusion is also backed up by variations in the $\delta^{13}C$ and $\delta^{18}O$ (Tables 3–5) which can be explained by an influence of various atmospheric factors.

2) Palisade (scalenohedral, "dog tooth-like") coarse and very coarse crystalline calcite (as much as tens of centimeters long), white, in places red-brown. The presence of elongated gas-liquid and liquid inclusions (Karwowski, unpubl. data), as well as iron oxide and hydroxide laminae parallel to fissure walls indicate periodic precipitation of calcite in direct

Table 3
Variation in the $\delta^{13}C_{PDB}$ and $\delta^{18}O_{PDB}$ in different generations of the "różanka" calcite from karstic-tectonic breccia (Museum of the Polish Geological Institute in Kielce)

Generations	$\delta^{13}C_{PDB}$ (‰)	$\delta^{18}O_{PDB}$ (‰)
1	−1.76 to 0.34	−5.52 to −3.61
2	−1.98 to 0.18	−7.47 to −1.14
3	−1.57 to −1.06	−6.33 to −4.46
4	−2.18 to −0.07	−7.89 to −4.61
5	−3.00 to −0.60	−7.34 to −5.29
6	−1.42 to 1.25	−6.95 to −6.81
7	−1.22 to −0.15	−8.08 to −6.96

Table 4
Homogenization temperatures (Th) of gas-liquid inclusions versus "oxygen" temperatures (Tox), the $\delta^{18}O_{PDB}$ and the $\delta^{13}C_{PDB}$ in selected "różanka" calcite

Location	T_h °C	T_{ox} °C*	$\delta^{18}O_{PDB}$ (‰)	$\delta^{13}C_{PDB}$ (‰)
Czerwona Góra	70–53	66–58	-8.87 to -7.73	-2.74 to -0.35
Chelosiova Jama	72–62	64	-7.69	-5.36
cave	62–54	55–42	-6.28 to -5.17	-3.42 to -3.02
	56	57	-6.59	-4.40
Zelejowa	59–55	62–56	-7.29 to -6.42	-6.42– to -6.23
	69–59	57	-6.61	-4.60
	59–52	61–55	-8.16 to -6.22	-6.19 to -1.14

based on assumption that the water in equilibrium with cvalcite had a $\delta^{18}O$ value of 1‰

contact with atmosphere. The impact of this atmospheric factor is also evidenced by variations in the $\delta^{13}C$ and $\delta^{18}O$ even within individual crystals (Table 5); besides, there is no relationship between the temperature and the size of crystals. The temperatures derived from homogenization of inclusions versus the $\delta^{18}O$ are shown in Table 4. In some places calcite crystals are curved or truncated which must have been linked to progressive fissuring during precipitation of calcite.

The younger phase (D) consists of at least eight generations (3 through 10):

3) Ferric ("marble-like I") micro and fine crystalline calcite, brown-red with white specks, in places light brown-grey. This calcite commonly encompasses individual fragments of limestone and calcite assigned to the two older "różanka" generations. Locally, precipitation of the ferric calcite was preceded by crystallization of a pinky-white fine crystalline calcite layer a few millimeter in thickness that veneers the palisade calcite.

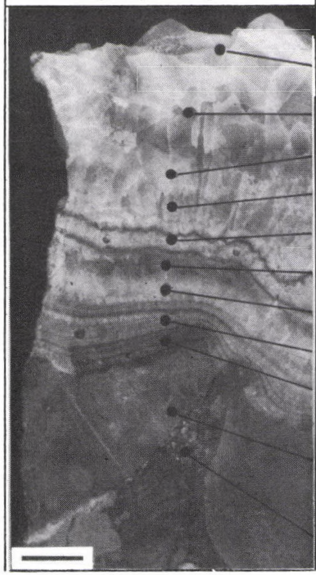
4) Wavy-banded ("ribbon-like") fine and medium crystalline calcite, pink–white or pink–yellow, with delicate parallel-wavy (at bottom, in places at top) or concentric ("roses", "rosettes", etc.) iron lamination. The concentric iron-rich laminae generally envelop limestone and older calcite fragments; this variety gave the name to the entire "różanka" calcite. Aside from flowstones, many isolated features resembling dripstones, i.e., stalactites and stalagmites occur here as well.

5) Ferric ("marble-like I") micro and fine crystalline calcite, brown-red with white specks. In places, it forms structureless fills within the aforementioned dripstones.

6) Eye-banded fine and medium crystalline calcite, white-gray. Most of the eye cores contain fragments of the older "rózanka" calcite. This calcite passes both into the older (5) and the younger (7) generation.

Table 5

The selected isotopic microprofile through the "różanka" calcite vein (phase C, generations 1 and 2) dissecting Permian conglomerates from quarry Czerwona Góra

Petrographic profile	Deposit	$\delta^{13}C_{PDB}$ (‰)	$\delta^{18}O_{PDB}$ (‰)	Tox °C*	$^{87}Sr/^{86}Sr$
	calcite 2	-2.74	-8.33	69	0.709432 (±6)
	calcite 2	-2.42	-8.87	71	0.710047 (±6)
	calcite 1	-4.05	-4.42	44	-
	calcite 1	-3.30	-4.81	46	-
	calcite 1	-2.87	-4.47	44	-
	calcite 1	-4.39	-5.96	53	0.709854 (±6)
	calcite 1	-3.36	-6.20	55	
	calcite 1	-4.98	-5.95	53	0.710012 (±10)
	calcite 1	-4.54	-5.30	49	
	limestone	0.57	-5.50	44	-
	matrix	-2.77	-5.60	51	-

*based on assumption that the water in equilibrium with calcite had a $\delta^{18}O$ value of 0‰

7) Block-palisade (rhombohedral–scalenohedral) micro to very coarse crystalline calcite (up to several centimeters long), pinkish white, sporadically semitranslucent. In some measure it resembles the palisade calcite of generation 2 but is more translucent and additionally interlaminated with hematite. In places, the block-palisade calcite makes up karstic-tectonic breccias cementing fragments of the older "różanka" calcite. It should be emphasized here that this calcite along with the younger calcite 8) intersects both the matrix and the pebbles of Zechstein conglomerates. The block-palisade calcite contains variable amounts of galena and barite.

8) "Honey" fine to coarse crystalline calcite, brown–yellow, in places with scattered galena or copper sulfides. In general, the "honey" calcite conformably overgrows crystals of the preceding calcite.

9) Hematitic micro and fine crystalline calcite, brown-red, in places with scattered galena. This calcite was recorded only in Mt. Stokówka.

10) "Hood" medium to very coarse crystalline calcite, grey, semitranslucent. It forms well-shaped crystals as much as 5 cm in diameter overgrowing calcite primarily of generations 7 and 8.

Most common are generations 4, 7, 8 10 1 and 2. Of the most interesting sites, Czerwona Góra (generations 1, 2, 7, 8 10), Zelejowa (1 through 7), Stokówka (1, 4, 7 through 10) and Kowala (7, 8 10) should be mentioned here. Generations 1 and 2 have not been recorded in the northern (£ysogóry) region of the Holy Cross Mts.

Exposed to radiation of ultraviolet light the "różanka" calcite gives off white, yellow, orange, pale blue, dark blue, as well as deep crimson light characteristic only for generation 8, which seems to be connected with the presence of dispersed humic matter. The results of ultraviolet radiation also indicated that hydrothermal calcite contains irregular intergrowths of authigenic (snow-white) or authigenic-detritic deposits (dark blue) of vadose origin.

The "różanka" calcite is depleted in magnesium (0.01–0.53%). In places, it contains an excessive amount of major and trace elements linked to scattered concentrations of many polymetallic sulfides (Table 1). The REE content is presented in Table 2.

The $\delta^{13}C$ values vary from –7.89 to 1.25‰ and the $\delta^{18}O$ from –14.11 to –0.91‰. The $\delta^{18}O$ corresponds to 122–25 °C, respectively, provided that the water in equilibrium with calcite had a $\delta^{18}O$ value of 1‰ (Migaszewski et al. 1995, 1996a). In general, the "różanka" calcite shows the more negative $\delta^{13}C$ compared to Devonian limestone host rocks. Similar values (–4.70 to 0.67‰) were obtained for the matrix of Zechstein conglomerates (Migaszewski et al. 1995). The negative $\delta^{13}C$ seems to be connected with temperature, pH, total content of carbon, etc. (Ohmoto and Rye 1974). There are no signs indicating an influx of primary fluids from the upper mantle. This conclusion seems to be supported by the $^{87}Sr/^{86}Sr$ ratio (0.708689 ± 9 to 0.711620 ± 10) that differ considerably from that characteristic of carbonates precipitated from primary magmatic waters, as well as by isotopic composition of lead in galena; the latter yields the ratios: $^{206}Pb/^{204}Pb$ = 18.336 to 18.408, $^{207}Pb/^{204}Pb$ = 15.362 to 15.621, $^{208}Pb/^{204}Pb$ = 38.335 to 38.412 (Zartman et al. 1979). As opposed to Permo-Triassic, Tertiary and Quaternary vadose calcite the "różanka" calcite shows no correlation between the $\delta^{13}C$ and the $\delta^{18}O$ (Migaszewski et al. 1994; Durakiewicz et al. 1995). Different generations generally yield the same $\delta^{13}C$ and $\delta^{18}O$. The most negative $\delta^{18}O$ values (and the highest temperatures) are observed in wavy-banded (4), block-palisade (7), hematitic (9), and partly palisade (2) calcite generations. The generation mentioned last shows the largest excursions in oxygen isotope composition (–7.47 to –1.14‰); variation in the $\delta^{13}C$ values are far narrower (–1.88 to 0.18‰). Lower temperatures of crystallization (less negative $\delta^{18}O$) resulted primarily from cooling of hydrothermal fluids in the presence of air. In this case, a greater influence of vadose water would be marked by more negative $\delta^{13}C$ (–10.38 to –9.89‰) which was found to be the case in Permo-Triassic vadose flowstone (Migaszewski et al. 1996a).

An interesting example is the isotopic composition of calcite generations 1 and 2 in Zechstein conglomerates (Table 5). The first generation (1) lining fissure walls yields more negative $\delta^{13}C$ and more positive $\delta^{18}O$ than the second one (2). Moreover these values are close to those encountered in a detritic-clayey-ferric matrix of the conglomerates. The matrix formed in a lagoon with intermittent influxes of seawater which is evidenced by Keith and Weber's ratio (1963) calculated from the equation $Z = 2.048 (\delta^{13}C + 50) + 0.498 (\delta^{18}O + 50)$. In this conglomerate series, it varies from 122.1 (bottom) through 119.0 (central portion) to 123.9 (top) indicating marine (>120) or limnic (<120) provenance of carbonate sediments. A short tectonic pulse which probably occurred late in Permian or early in Triassic time produced faults dissecting poorly lithified conglomerates. These faults also channeled hydrothermal fluids, modified in places by pore water contained in dissected sediments, rainfall, etc. Precipitation of calcite proceeded periodically toward the center of fissures isolating the second generation (2) from the direct influence of the host rock. An increasing fault width enabled formation of large calcite crystals reaching a couple of centimeters in length.

Cimmerian-Alpine calcite

This calcite occurs within Upper Jurassic and Upper Cretaceous marine carbonate rocks in the form of minute lenticular accumulations and veinlets as much as 1 cm thick. They are white, fine and medium crystalline and Mg-depleted. Iron oxides and hydroxides are very scarce; no Cu-, Pb- and Zn-sulfide concentrations have been found. Calcite veinlets piercing Upper Cretaceous carbonate sediments also contain chalcedony.

The above-mentioned calcite exposed to ultraviolet radiation emits white light with a somewhat violet hue.

All the $\delta^{13}C$ values are positive (1.28 to 2.67‰). The $\delta^{18}O$ varying from –8.69 to –5.87‰ is generally less negative than that in the Variscan and post-Variscan calcite. These values correspond to 58–47 °C (assuming that the water in equilibrium with calcite had a $\delta^{18}O$ value of 0‰). The marine provenance of hydrothermal fluids is highlighted by the $^{87}Sr/^{86}Sr$ ratio ranging from 0.707269 ± 11 to 0.708129 ± 12.

Conclusions

In the Holy Cross Mts calcite mineralization was found to have been developed in 5 phases: A – Variscan (Visean/Namurian), B – older post-Variscan (post-Namurian through pre-Zechstein), C and D – younger post-Variscan (Permian/Triassic and middle/late Early Triassic), and E and F – Cimmerian–Alpine (Late Jurassic and Late Cretaceous). This division, based on geologic and petrographic data was confirmed by the $\delta^{13}C/\delta^{18}O$ distribution pattern, i.e. the presence of 4 (A, B, C+D, E+F) genetically different populations (Fig. 2 in Migaszewski et al. 1996b). They formed in various geologic environments varying from offshore zones (phases E+F) through marine relict troughs connected with the waning Variscan movements (phase A) to mixed marine-lagoonal-continental zones (phases B and C+D) affected by strong karstification especially in latest Permian–Early Triassic time (phases C+D). The marine provenance of calcite ascribed to phases E and F has been proved by the presence of fossil formation of sea-floor hot springs in Upper Cretaceous rocks (Migaszewski et al. 1987, 1996a) and the $^{87}Sr/^{86}Sr$ ratio varying from 0.707269 ± 11 to 0.708129 ± 12. In turn, the continental influence on crystallization of calcite assigned to phases C and D has been evidenced by strong variation in the $\delta^{13}C$ and $\delta^{18}O$ values and the $^{87}Sr/^{86}Sr$ ratio as well as by the presence of characteristic elongated inclusions. A large number of different features resembling spaleothems cutting and being cut by hydrothermal calcite as well as karstic authigenic and allogenic intercalations and fills, karstic-tectonic breccias, etc. (Migaszewski et al. 1996a) indicates an impact of karstification on the process of calcite formation. Determinations of oxygen isotope composition and homogenization temperatures of gas-liquid inclusions in the same calcite crystals indicated that the hydrothermal fluids were of marine provenance (Migaszewski et al. 1995, 1996b). The mean $\delta^{18}O_{SMOW}$ was 1‰ being close to that found in mid-ocean hydrothermal fluids (± 1‰); it is completely different from the $\delta^{18}O$ range of meteoric waters (–5 to –22‰ to SMOW). However, in some cases, due to the lack of inclusions, it is hard to say to what extent hydrothermal fluids were modified by brine or marine-brackish porous water, rainfall, etc.

The results of isotopic determinations did not show any spatial variation. The obtained $\delta^{13}C$ and $\delta^{18}O$ values are more or less uniformly distributed in the entire area; in addition they do not change with depth. For instance, in borehole Ostalów PIG 2 (2826.4 m) drilled about 50 km north of Kielce the $\delta^{18}O$ varies from –4.13 to 2.73‰ (42–35 °C).

The most important role is played by the "rózanka" calcite (phases C and D) that volumetrically make up about 80–90% of the entire calcite mineralization in the Holy Cross Mts. These two phases, though genetically connected are separated by a distinct fracture width. As opposed to the remaining phases the "rózanka" calcite occurs within the same fault system trending more or less N–S. These numerous faults dissecting consolidated Variscan synclines and anticlines enhance a unique structural position of the "rózanka" calcite. Its Early Triassic age precludes any direct relationship with the magmatic events in the Holy Cross Mts. The K/Ar dating of biotites in lamprophyres, the youngest igneous rocks of the region, indicates 374.4 Ma (Middle/Late Devonian) through 288.8 Ma (Late Carboniferous/Early Permian) as the most probable time interval (Migaszewski and Halas 1996). Based on all the present data an entirely different model of hydrothermal "pump" should be devised. It could have been linked to a rare geologic event that took place here in Early Triassic time. According to Migaszewski (1996a) the entire Holy Cross Mountain area might have shifted toward NW along the western margin of the East European Platform simultaneously rotating clockwise. These two combined movements caused the forming of parallel faults featured by a close to N–S strike. The produced heat could have warmed up the seawater and triggered its circulation. This conclusion seems to be supported by the lack of any Triassic igneous intrusions in the Holy Cross Mountain region (Migaszewski and Halas (1996).

Petrographic and isotopic investigation of calcite shed a different light onto the age and origin of sulfide and barite mineralization in the Holy Cross Mts. This problem should still be re-examined; nonetheless at the present state of knowledge, Pb-, Cu- and Zn-sulfides seem to have been connected primarily with phases D, in places B, C and A and barites with D, sporadically A. The $\delta^{34}S$ in galena varies from –13.24‰ (Czerwona Góra) to 11.63‰ (Karczówka), and in barite from 16.43‰ (Józefka) to 30.02‰ (Krzemucha). In turn, the $\delta^{18}O_{SMOW}$ in this barite ranges from 20.07‰ (Krzemucha) to 12.08‰ (Jaźwica). These values correspond to a temperature of 52–124 °C, respectively.

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Radiographic investigation of Danube River and Danube Delta sedimentary cores

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Sedimentary cores, collected from different stations located on both the Danube River and Danube Delta were investigate by the X-ray radiography method. All radiographs were obtained by using an industrial type SUPERLILIPUT SL 140 generator. The effects of X-ray beam parallaxes were corrected by means of a specially designed grid onto which all radiographs were projected. A remarkable diversity of features such as sedimentary facies, grain size, laminae thickness and orientation, reworked materials, plant roots and fragments of vegetation as well as traces of benthic animals whose patterns appears on radiographs are presented and discussed.

Key words: x-radiography, sedimentology, stratigraphy, depositional processes, postdepositional processes

Introduction

The radiographic method represents one of the most efficient and nondestructive techniques successfully applied in the investigation of various kinds of sedimentary samples such as slabs, polymetallic nodules, drilling cores, etc. In the last case, the information obtained concerns internal structures of cores such as layer sizes and orientations, degree of mineralization, granularity, animal burrows, recent as well as fossil mollusks shells, etc.

Although the first application of radiography in geology was performed about a hundred years ago (Brühl 1896), the systematic application of this method in sedimentology only began three decades ago (Hamblin 1962; Calvert and Veevers 1962; Zangerl and Richardson 1963). At present, and especially in the last two decades, the investigation of sedimentary formations has been notably diversified. This diversity refers to the technical improvements of the method together with the information that this technique can provide. The data presented in literature deal with the investigation of structure and core composition (Bouma 1964; Clifton 1966; Edmonson and Allison 1970; Huang and Stanley 1972; Keeling et al. 1979; Maldonado et al. 1981; Axelson and El-Daoushy 1989; Butler 1992; Dugmore and Newton 1992) including the activity of various benthic organisms (Howard et al. 1977). At the same time, a number of papers were devoted to the general (Bourn-Bouge 1972) and the specialized procedure of samples collecting and preparing (Aller 1980; Axelson

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and Hakanson 1978; Savadra et al. 1985; Britt et al. 1992; Mehel and Merkt 1992; Allegro et al. 1994) as well as the computer interpretation of densitometrically investigated cores (Axelson 1983). Three monographs have been partially (Bouma 1969; O'Brien and Slatt 1990) or totally (Krinitzki 1970) dedicated to this subject.

During the period between 1975 and 1995, in the frame of a long-term program devoted to the complex investigation of sedimentary formations of the Danube River Basin as well as the Danube Delta some 7800 drilling stations were set up from which more than 4300 cores were collected. About 20% of these cores were radiographically examined before being cut. In this way more than 1700 radiographs were obtained which contain a rich volume of information concerning the properties of sediments from the investigated area.

Some of the most representative radiographs together with a corresponding brief description of the cores are presented in this paper. Almost all described cores were collected from the Danube Delta lakes as well as from the inner channels network. Only two of them are located away from Danube Delta (D 3681 and D 3607), on the Lower Danube Braila Pond. The corresponding locations of coring stations are reproduced in Fig. 1. This paper also represents an attempt to describe several X-ray radiographs (positive prints) obtained by using a "Superliliput SL 140" device and different types of radiographic films to check the efficiency of this kind of tool in core analysis.

Materials and methods

The investigated cores were collected by means of a gravitational Kullenberg corer which leaves them directly embedded in polystyrene tubes with a maximum length of 6 metres with external diameter of 6.5–7.5 cm. The wall thickness of the plastic core tubes was no greater than 0.6 cm. For an optimum X-ray examination, the cores were sectioned in segments no longer than 2 m. To maintain cores wet, the ends of the tubes were sealed by means of textile adhesive band and special plastic stoppers. At the same time, to preserve the internal structure of layers undisturbed the collected cores were stored in a vertical position.

Before being examined the cores, embedded in plastic tubes, were marked every 20–22 cm, by using thin copper wire markers. In this way, a radiographic image of the whole core was obtained by assembling up to 10 radiographs. In special cases, in order to obtain as much details as possible, the same segment of core was examined by sending the X-ray beam in different directions. To reduce image distortion caused by parallaxes to a minimum, no core segments longer than 22 cm were ever examined at one time.

Cores marked in this way were radiographically investigated segment by segment using a SUPERLILIPUT SL 140* industrial portable X-ray generator. This generator had a maximum anodic potential equal to 140 kV at a current up to 5 mA. The experimentally determined optical focal spot had an average

Fig. 1
Map of the lower part of the Danube River and Danube Delta showing the location of coring stations. The positions of coring stations from where the cores described in the present paper were collected are indicated by call outs

diameter equal to 1.35 mm. All radiographs were taken at the same focal distance equal to 85 cm.

Nearly all investigated cores presented the same composition, mainly consisting of peat and other organic materials, sand, silt, loess, etc. Corresponding to this composition, for core diameters between 6.5 cm and 7.5 cm, the optimum anodic potentials were within the 100–130 kV range (Duliu and Mihailescu 1993). Only in a few cases, when peat or other fragments of vegetation were dominant, the most representative radiographs were obtained with a 75–80 kV anodic potential. Generally, the best results were obtained by slowly increasing the acceleration potential from 80–100 kV at the upper part of the core to 100–120 kV at the lower one.

For the focal distance and optical focal spot diameter defined above, geometrical reduced definition or geometrical blurredness (Mehel and Merkt 1992) was equal to 0.15 mm for the tub side closed to X-rays and to 0.015 mm for the opposite side, closer to radiographic film.

The best radiographs were obtained by using industrial, high contrast, radiographic films like STRUCTURIX D7, TEST-X L, and GAMAGRAF G1 and G2. The worst results were acquired by using medical radiographic films. In the all cases standard 10 x 24 cm radiographic films were used, wrapped in 0.02 mm thick intensifier lead screens. A very limited number of radiographs were made directly onto normal photographic paper. Although radiographs obtained in this way have shown good contrast and definition, the exposure time was about ten times greater than in the case of radiographic films, which was prohibitive.

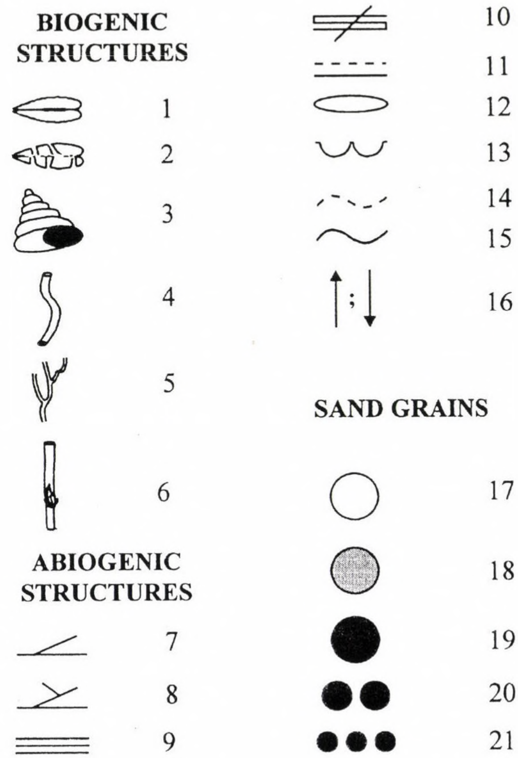

Fig. 2
List of symbols used to indicate the most important features observed on radiographs (after Keeling et al. 1979, modified)

As a function of anodic potential and film type, the exposure time was within the range of 1–2.5 min, its optimum value being experimentally established. The exposed films corresponding to the same core were processed simultaneously. Positive prints were made from the negative films. The best results were obtained by using polietilenated photographic paper.

In order to illustrate the most important details found on the radiographs, a set of symbols, the complete list of which is reproduced in Fig. 2, was used.

A classical X-ray tube emits a divergent beam of radiation which determines an enlargement of the radiographic image with respect to the examined object directly proportional to focal distance of apparatus and sample thickness. This enlargement, which directly increases with the distance from the beam axis, reaches a maximum value of 13% in the case of the present study. As a result, the effective thickness of the sample was greater at sample extremities by 3.57%. Although this fact does not affected the exposure time or the image qualities, some details appear slightly enlarged, especially on the core segment extremities. To correct this effect and to obtain a better accuracy in determining of the size of the various features which appear on radiographs, a special grid was designed onto which all radiographs were projected (see Figs 3–12).

Results and discussion

In Figures 3–12 the most representative features which were observed on the radiographs of the investigated cores are reproduced as positive prints.

Previous investigations (Mihailescu et al. 1982; Winkles et al. 1995; Duliu et al. 1996) have proved that although sedimentation processes occur over the entire Danube Delta, the sedimentation rate varies within this area from 0.6 to 100 mm/year. The highest sedimentation rate corresponds to those lakes and channels which are closed to the main Danube branches.

Hence the total thickness of lacustrine sediments varies between 0.5 m in Matita Lake and 2.5–3 m in Lake Golovita (Noakes and Hertz 1983).

All the investigated sediments are therefore relatively young (present to 2500 y BP) and thus characterized by a great number of common characteristics, better reflected in their lithological composition (see Table 1).

Table 1
Average lithological composition of the Danube Delta sediments as a function of their location

Location	Shells and other mollusk debris	Sand	Silt	Clay
Fluvio-marine area	1.8%	12.8%	59.3%	26.1%
Fluvio-lacustrine area	3.1%	6.3%	63.2%	27.4%
Secondary channels	3.0%	26.7%	48.9%	21.4%

Another peculiarity which must be considered for better characterizing of the morphology of the bottom sediments of the Danube Delta is the fact that water flow (hence solid / liquid discharge) has a double direction: at high level the Danube waters enter the lacustrine depressions, and at low levels they return from the depressions to the major river-bed.

Two main factors have contributed to the actual aspect of the sediments: the bi-directional circulation of the water and the activity of benthic organisms, both revealed with accuracy on radiographs.

The first factor is reflected by the sedimentary facies which in some cases consists of a relatively regular superposition of silty-lutaceous laminae (Figs 3, 4), or by the cross-laminations (Fig. 3). Graded-bed structures are rare and consist of fine sand and silt; in some cases, traces of erosion occur in their upper part (Figs 3, 4 and 5). Lenticular structures are not very common (Fig. 3) but appear accompanying the erosion channels.

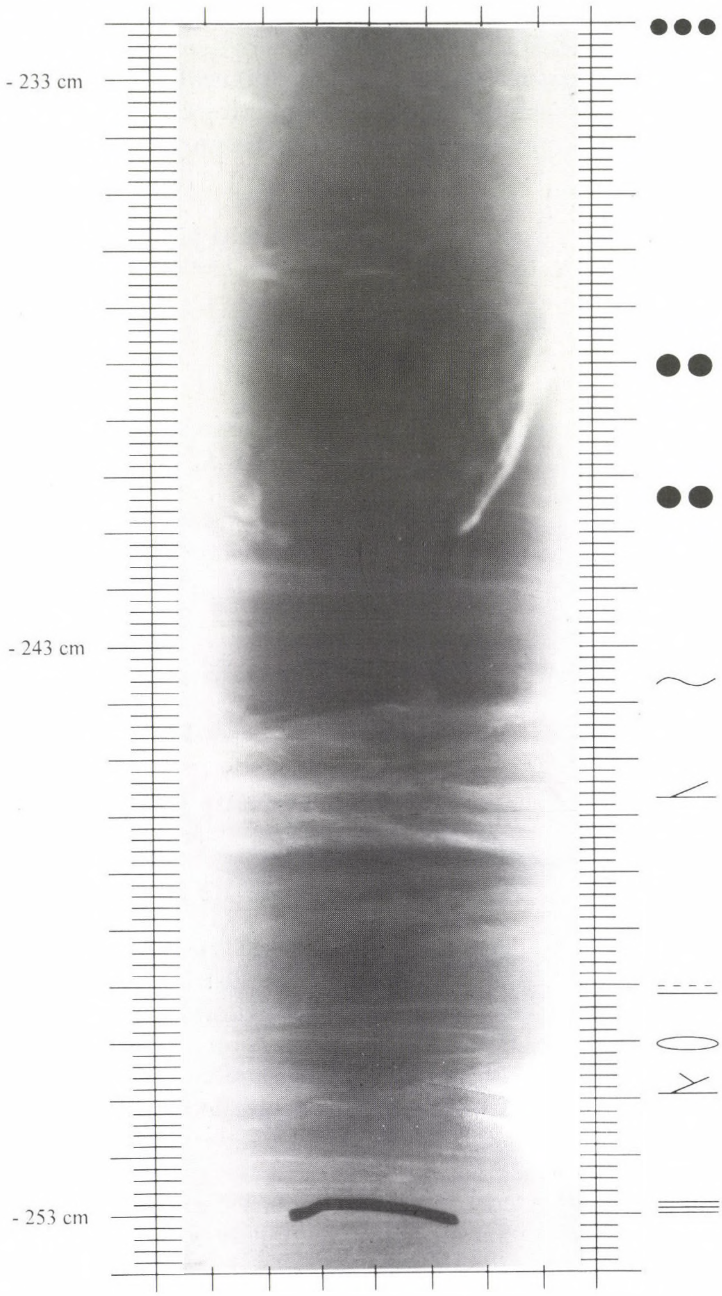

Fig. 3
Core R 697 - Golovita Lake (Danube Delta): Erosion channel. Alternation of sand and silt graded-beds with cross laminations. (TEST X L film; positive print; dimensions in cm)

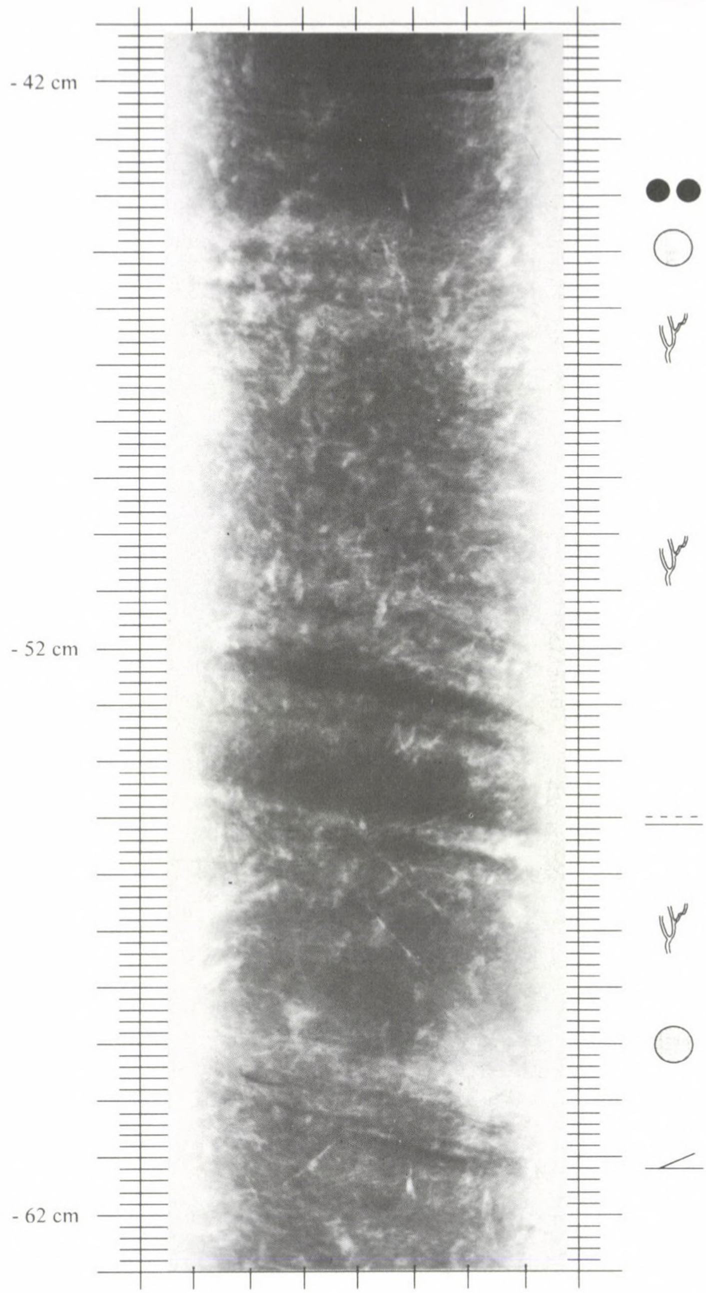

Fig. 4
Core DD 2801 Sulina Channel (1.5 km offshore): Numerous plant debris. Inclined laminae. (TEST X L film; positive print; dimensions in cm)

Fig. 5
Core R 1331 – Razelm Lake, Fundea Bay (Danube Delta): Reworked sediments in the upper part of the core. Complete empty bivalve specimen. (GAMMAGRAF G2 film; positive print; dimensions in cm)

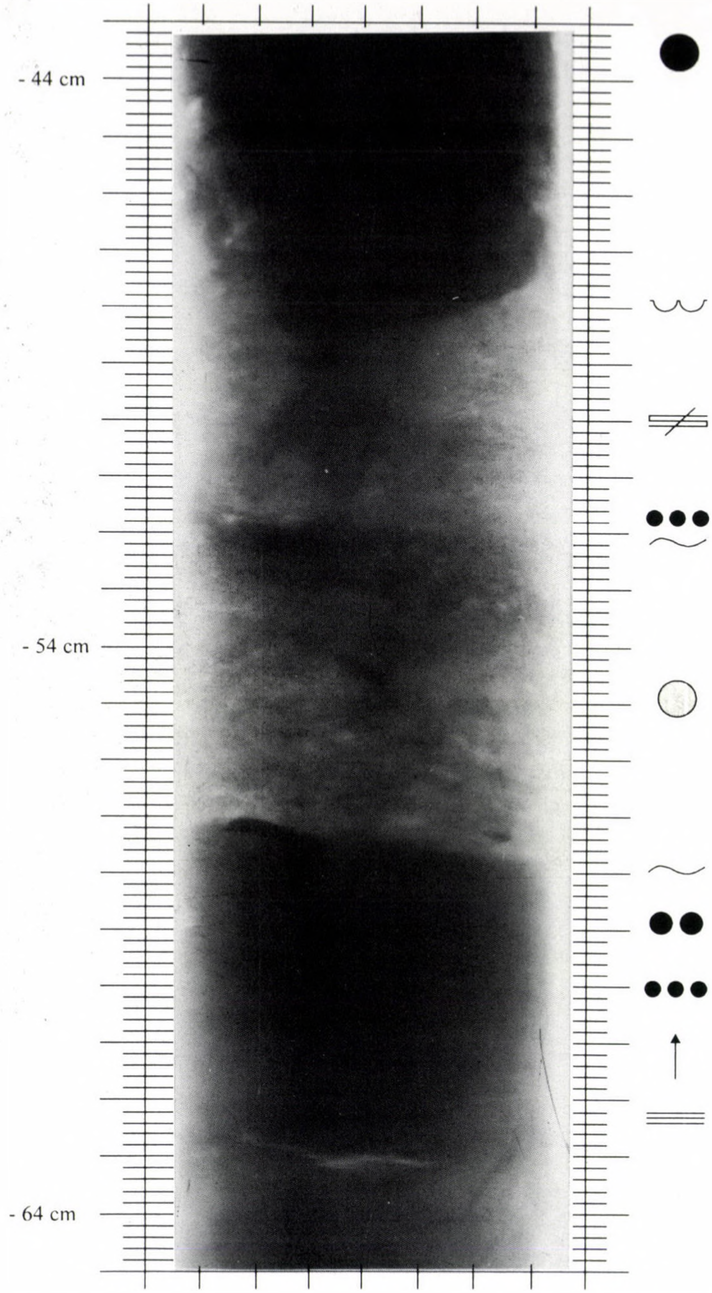

Fig. 6
Core R 361 – Razelm Lake (Danube Delta): Erosion channel. (TEST X-L film; positive print; dimensions in cm)

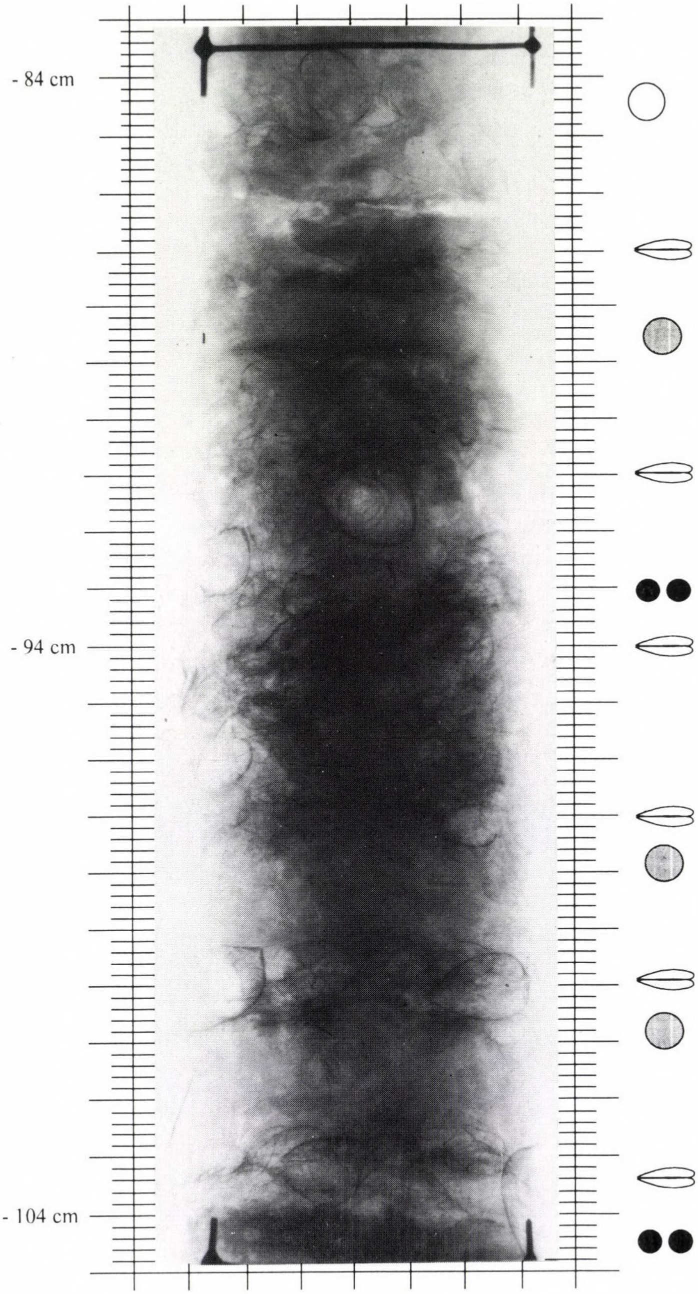

Fig. 7
Core H 22 – Histria Lake (Danube Delta): Graded bedding sediments with bivalve beds – *Cardium edule lamarcki* Reeve. The cross orientation of the mussels suggests a transport process. (GAMMAGRAF G1 film; positive print; dimensions in cm)

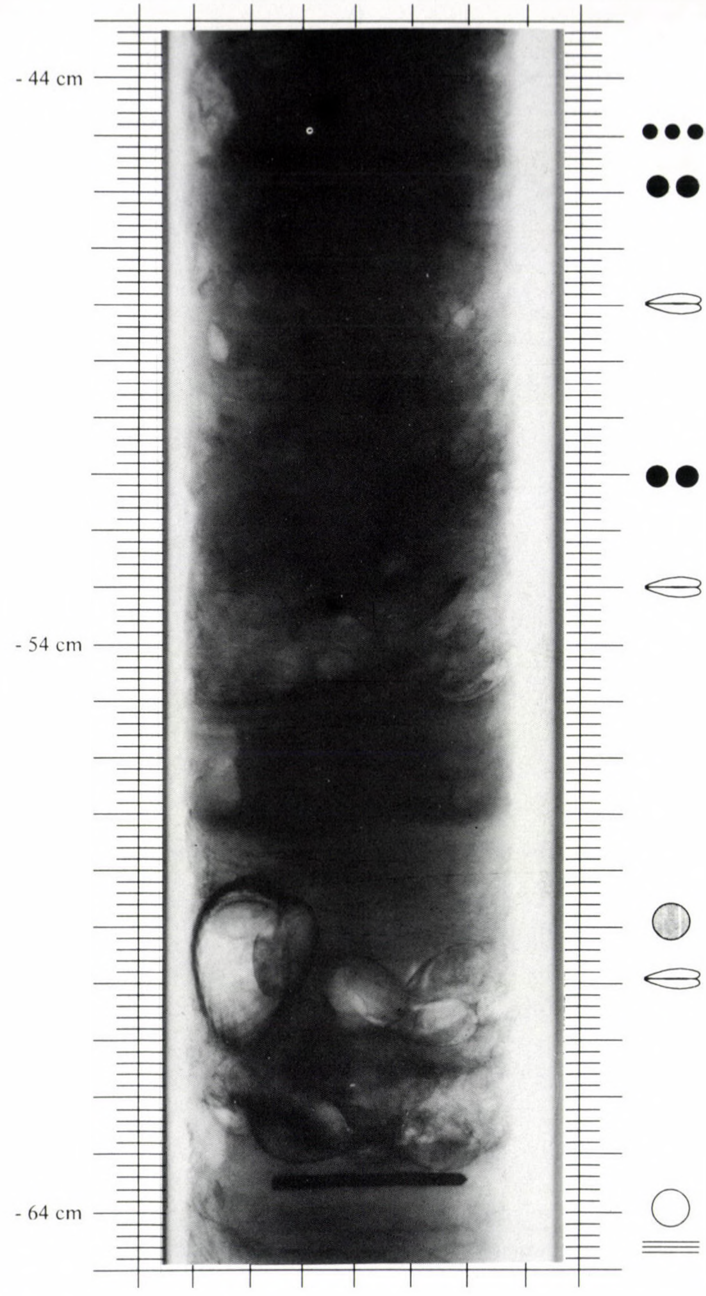

Fig. 8
Core S 98 – Sinoe Lake (Danube Delta): Beds of complete bivalves, either empty or filled with the sediment they were found in. (STRUCTURIX D7 film; positive print; dimensions in cm)

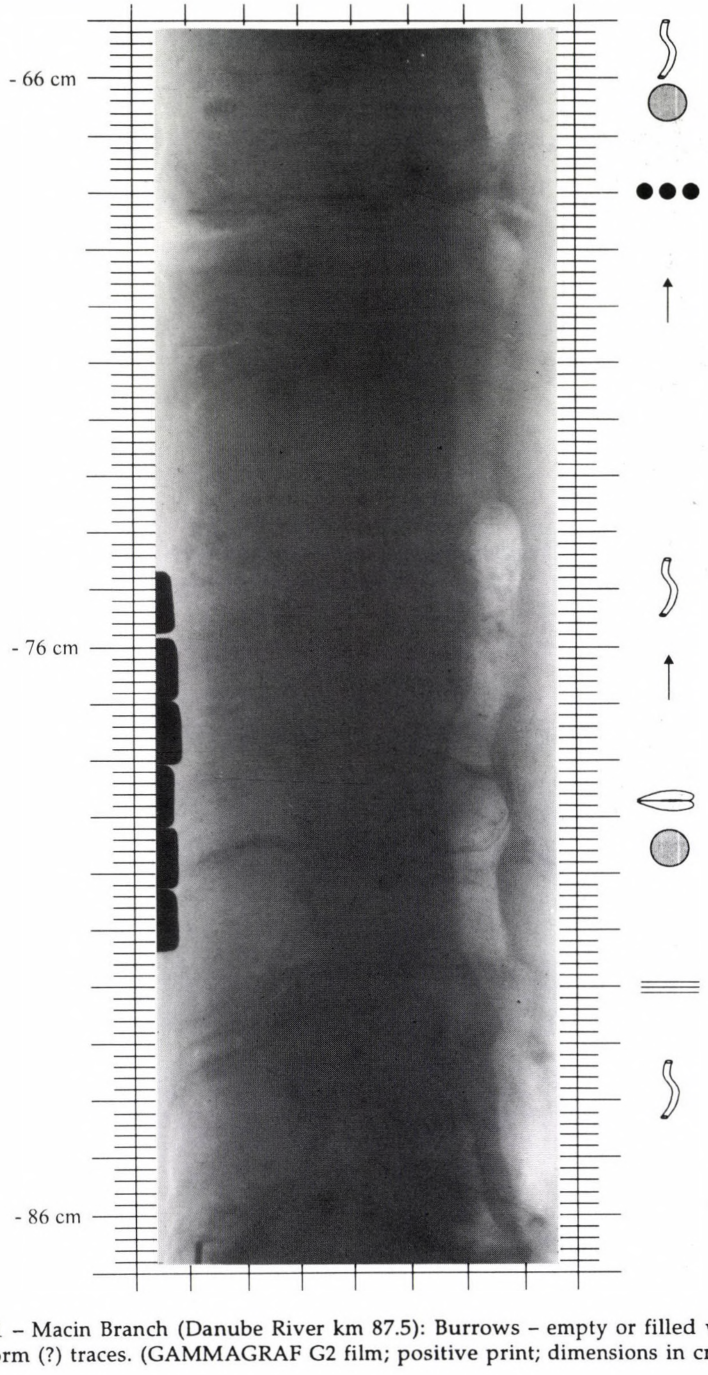

Fig. 9
Core D 3681 – Macin Branch (Danube River km 87.5): Burrows – empty or filled with a finer detritus. Worm (?) traces. (GAMMAGRAF G2 film; positive print; dimensions in cm)

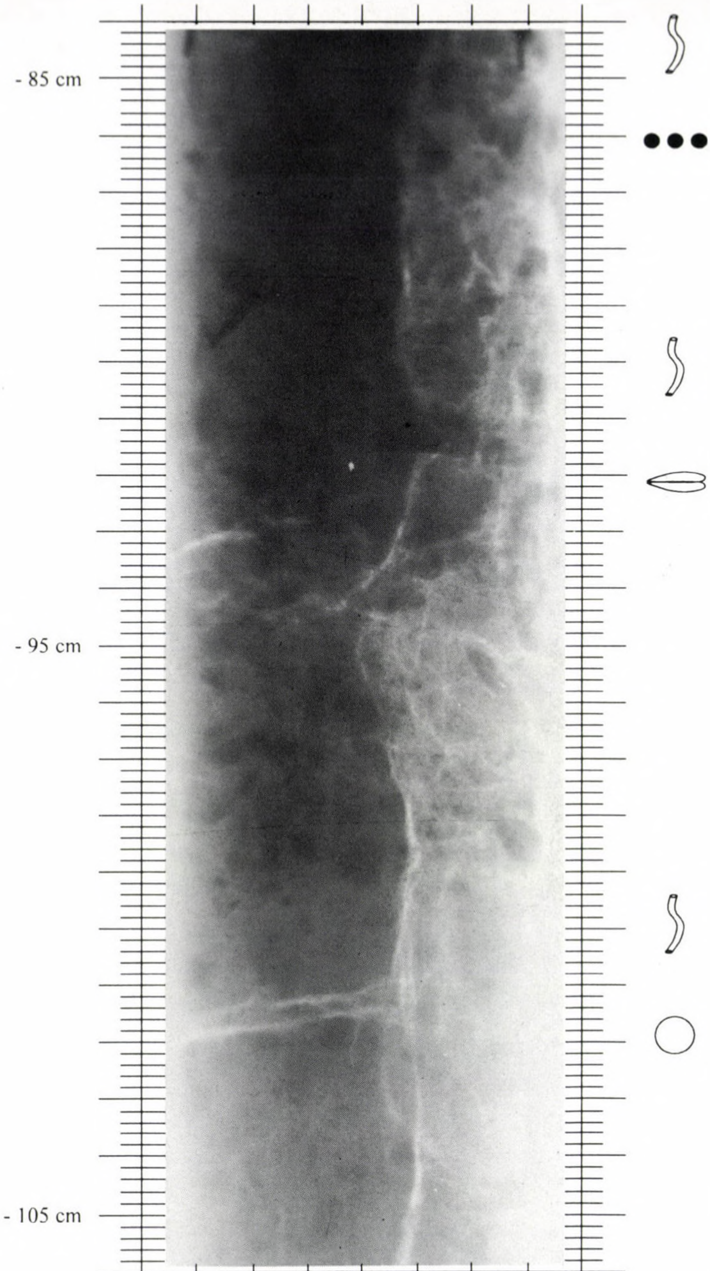

Fig. 10
Core D 3907 – Matusoaia Branch (Danube River) Bioturbation process. (GAMMAGRAF G2 film; positive print; dimensions in cm)

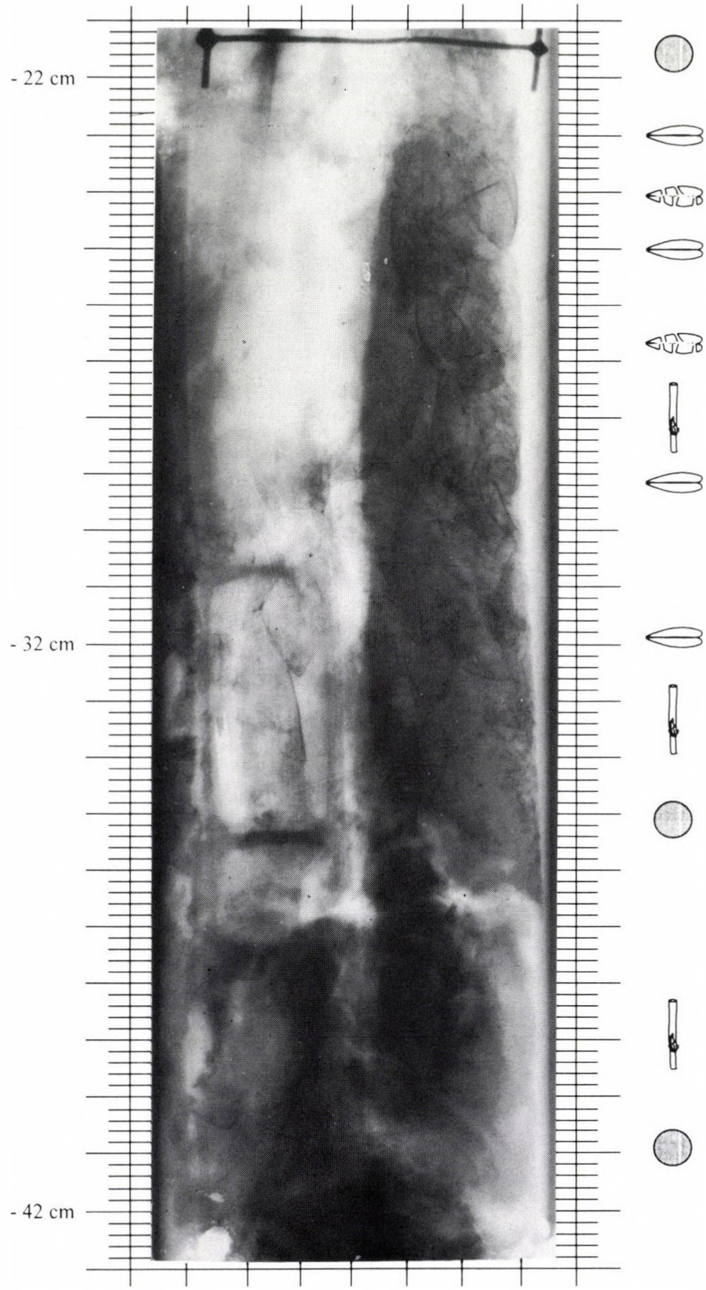

Fig. 11
Core DD 4229 – (20–40 cm core fragment) – Rosu Lake (Danube Delta): Reed stem in a silt bed. Behind the stem, mussel specimens can be seen; the images are superposed. (GAMMAGRAF G2 film; positive print; dimensions in cm)

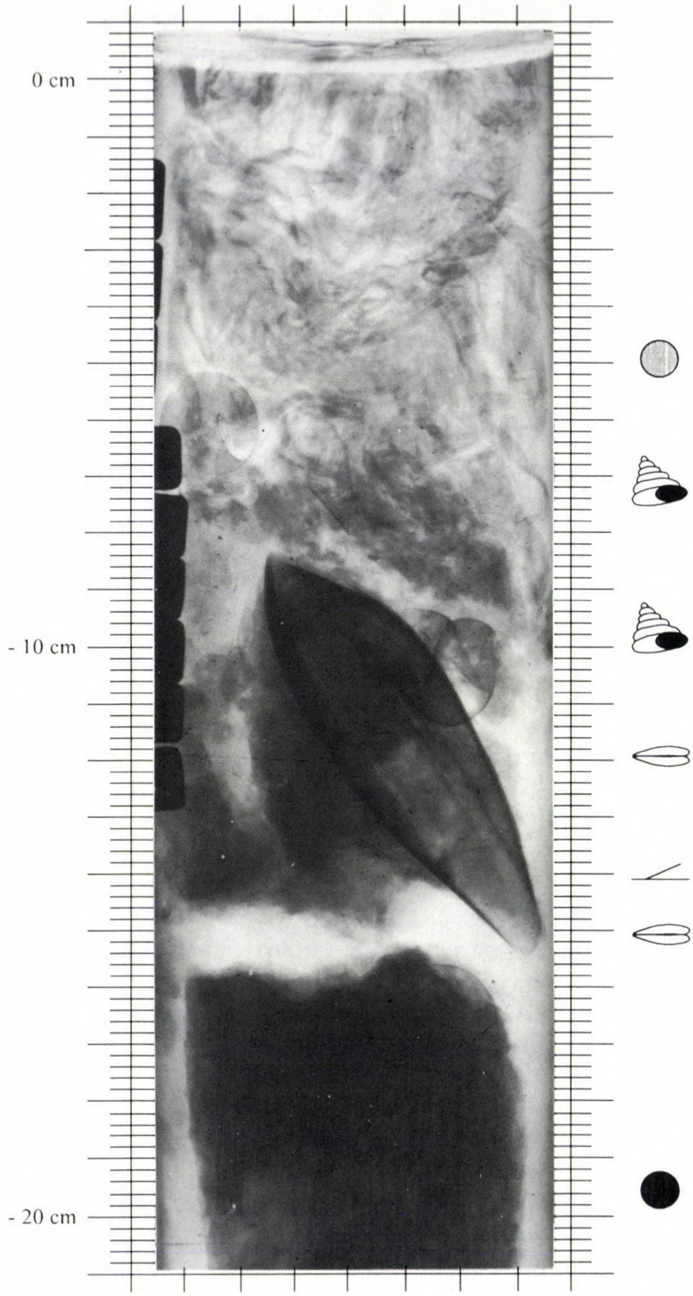

Fig. 12
Core DD 3615 – (0–20 cm core fragment) – Vanova Channel (Danube Delta): Cross-oriented bivalves; the largest specimen is in an obvious position of growth. (GAMMAGRAF G2 film; positive print; dimensions in cm)

At the same time, some radiographs reveal a disturbance of the usual structures, when the quiet processes are interrupted by very fine layers of reworked sediments (Fig. 6)

The occurrence of irregular silts is frequent especially when the mouth of channels perforates the silt and fills it with compact detritus as it can be seen in Figs 3, 5 and 6.

The presence of a great amount of vegetal debris mixed with silt and fine sand is characteristic of the rapid sedimentation process which occurs offshore, in the main streams of Danube branches (Fig. 4).

The second factor, the activity of living organisms, is characteristic of both Danube Delta and Danube River sediments and consists of the presence, in the upper layers (up to 1 m depth) of diverse invertebrates, later identified as chironomides larvae, present-day bivalves (*Dreissena pollimorpha* L., *Cardium edule lamarcki* Reeve, *Limnocardium sp.*) or gastropods (*Viviparus sp.*), which spend their life cycle buried in the mud or on the sediment surface (Mihailescu et al. 1982, Duliu et al. 1996). The selective activity of burrowing organisms living in very fine sand and silt can be observed on numerous radiographs. These animals, with or without shells, characterize the majority of the disturbed sediments (Figs 5, 6), producing a great degree of bioturbation. Very rarely, empty shells occur (Fig. 5).

In many cases, their burrows are filled with fine-grained sand, coarser than the silty particles of the bed they perforate, forming a specific pattern (Figs 7, 8) encountered in many radiographs. The occurrence of burrows in silts and sand with fine laminae is frequent; these burrows are empty or filled with detritus other than the constituents of the bed they perforate (Figs 9, 10). The images are very different and it must be emphasized that the cores have a cylindrical section; hence the images are sometimes superposed (Fig. 10). Sometimes regular shell agglomerations with a marked polarity can be observed (Fig. 8).

Another important component of the superficial sediments consists of plants debris such as roots or even fragments of stems.

Compact silts containing numerous plant debris sometimes occur, mainly in the upper part of the cores (Fig. 4) while the root fragments are rarer and reed stems occur occasionally (Fig. 11).

When mixed with mineral detritus, plant debris participate in the formation of cross laminations (Figs 3, 4).

Beds consisting almost entirely of shell debris are frequent, especially in the sediments of both Razelm and Sinoe Lakes (Figs 7, 8) where the sand matrix is almost absent, the amount of broken shells being close to 100%. Valves of *Limnocardium sp.* can be observed either in living (Fig. 8) or lethal (Figs 7, 12) position. In some cases worm burrows can be seen in the broken shells beds (Fig. 9).

Numerous X-ray photographs are suggestive in regard to the great importance of mollusks in the formation of lacustrine sediments of the Danube

Delta; bivalves rework the sediments they live in (Figs 7, 8), precipitate organic and mineral particles and leave behind shell debris in the sediments (Fig. 5) or make up the sediment surface.

Generally, there are important differences between the aspect of the sediments collected from river beds (Figs 4, 9 and 10) and lacustrine ones (Figs 3, 5–8, 11 and 12). This reflects the differences in sedimentation processes as well as the presence of different mechanisms (mechanic and bioturbation) by which the sediments are reworked.

Concluding remarks

The sedimentary formations collected from the Danube River and Danube Delta represent, by their remarkable diversity, very adequate objects for X-ray investigation. The great volume of information obtained by this method allows an accurate examination of different features which appear on the radiographs. In this way, a careful selection of those core components requiring ulterior detailed investigation could be performed.

The good quality of the radiographic images obtained is mainly due to the SUPERLILIPUT SL 140 portable X-ray generator, which, even in the condition of a summarily equipped laboratory, permitted the performing of good research work. With a reduced weight and multiple possibilities for connection to various power supply sources, including portable electric generators, this device works even on board research vessels, which made it a useful tool for this kind of investigations.

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Vivianite occurrence at the Nistru Mine (Misztbánya), Romania

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Between 1965–1986, we have noticed the occurrence of well-developed prismatic and lamelliform vivianite crystals at Veins 140, 141, 142 and 144 of the Nistru Mine (Misztbánya) located in Transylvania, Romania. These were found between Levels 2 and 5 (230 m respectively 130 m above sea level).

Vivianite crystals were most commonly seen with quartz and pyrite in geodes. They also occurred in kaolinitic groundmass, containing fine-grained marcasite and barite crystals. Crystals range in size from 1–10 cm, averaging 5 cm, and show a striation parallel to the c - axis on facets.

They are commonly light green to bluish-green in colour, rarely dark green, with yellow-green and purple-green reflexes. The indices of refraction are $\alpha = 1.569$; $\beta = 1.599$; $\gamma = 1.635$, the double refraction = 0.066. Crystals show positive optical characteristics.

The crystals show a very good cleavage along 010. They have a specific gravity of 2.76 g/cm^3 and a hardness of 1.8. The chemical composition of the crystal is SiO_2 = 0.66%, FeO = 35.07%, Fe_2O_3 = 9.96%, CaO = 0.55%, MgO = 0.12%, P_2O_5 = 28.04%, H_2O = 25.62%. Trace elements are Ni, Co, and Zn. The main lines of X-ray diffraction are 6.72, 4.91, and 1.678 dÅ. Differential thermal analysis shows endothermic reaction between 170–190 °C, with a mass loss of 24.5%. Formation of the crystals is primary and occurred at low temperature and pressure in the last phase of the hydrothermal process.

Key words: mineralogy, vivianite, Nistru-Romania

Introduction

The Nistru Mine is situated in the Gutin Mountain Range, 8 km northwest of the ancient mining center of Baia Mare (Nagybánya) in the Transylvanian region of Romania (Fig. 1). The Gutin Mountains were formed by Tertiary volcanic activity that produced a sequence of andesite, dacite rhyolite and again andesite rock types. Significant ore mineralization (lead-zinc, silver, and gold) occurs here. This district is the scene of intense mining activity that has been continuous for over nine centuries. A number of famous mineral localities are found here: Baia Mare, Baia Sprie (Felsőbánya), Cavnic (Kapnikbánya), Baiut (Erzsébetbánya), and Nistru (Misztbánya).

As many as fifty hydrothermal veins have been discovered, many of them mined, around the volcanic neck of Handal Rock. Veins are more or less parallel and are hosted in altered pyroxene andesites and interbedded claystones and marls of Badenian and Sarmatian age. Like the well-known gold and

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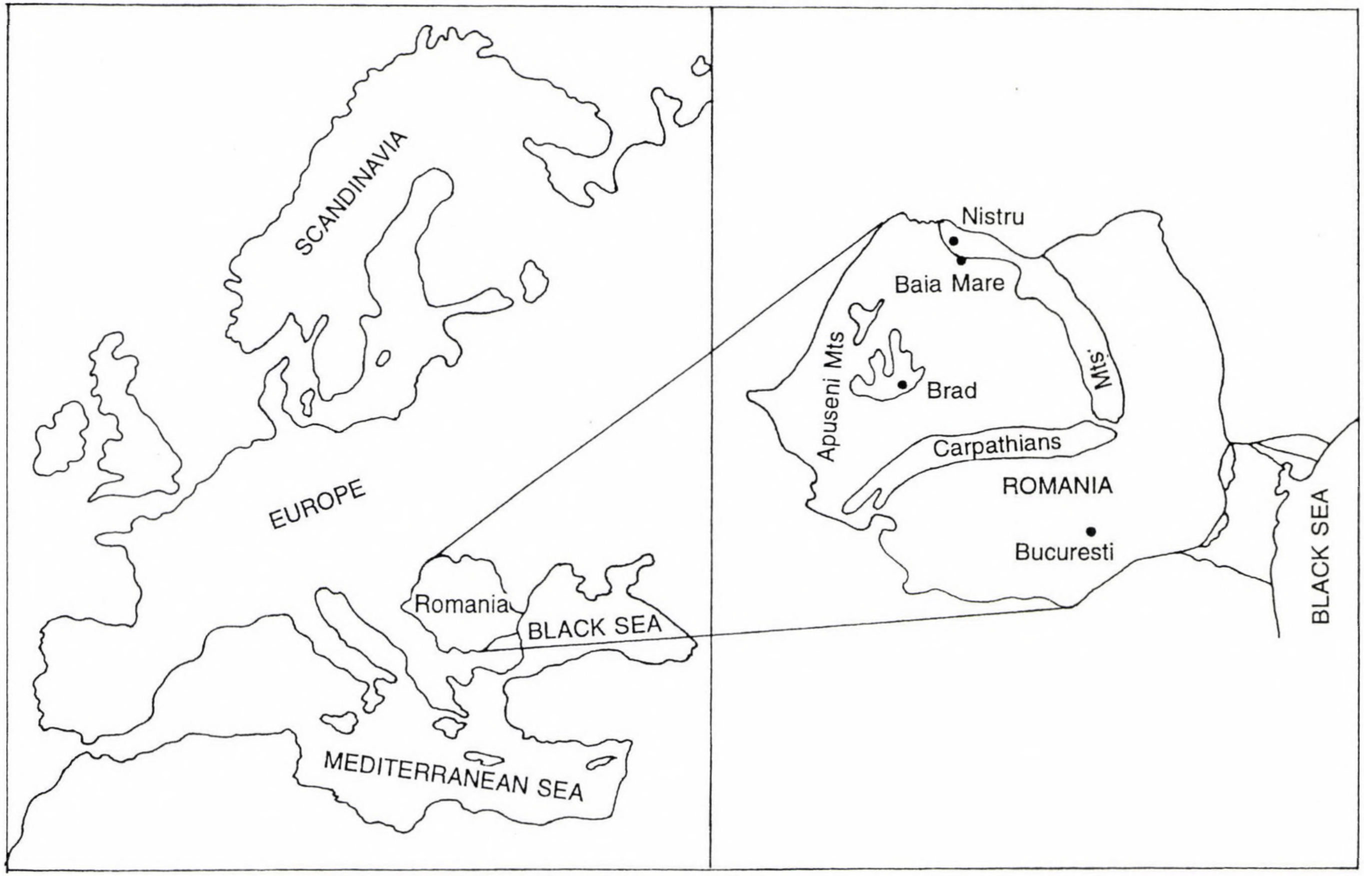

Fig. 1
The location of the study area

silver-bearing quartz veins around Baia Mare, those at Nistru are of the epithermal to mesothermal type. Besides the major fissure-filling minerals of quartz, calcite and barite, the veins contain galena, sphalerite, wurtzite, pyrite, marcasite, and chalcopyrite. Locally abundant but occurring in smaller quantities are proustite, pyrargyrite, stibnite, gold and tellurides hessite, petzite, altaite and sylvanite, various clay minerals and vivianite, the last named of which will be described below. In the general district, the most famous localities of vivianite are the Herja (Kisbánya/Herzsahegy) and Ilba (Iloba) mines.

Vivianite, bobierrite, hörnesite, parasymplesite, erythrite, annabergite and köttigite are members of a series that crystallize in the prismatic class of the monoclinic system with characteristic (except for borbierrite) lamellar habit.

Vivianite is universally considered to be a mineral formed by supergene processes, where phosphate-rich solutions react with the oxidized products of iron minerals such as pyrite, pyrrhotite, marcasite, and siderite in a relatively oxygen-poor environment. The idea that vivianite is secondary stems from the fact that it is often found in strataform iron deposits rich in phosphate, in swamp sediments, and in blue clays. An increasing number of workers in mineralogy now believes that vivianite can also form from hydrothermal solutions, in the company of other minerals, precipitated in various bands within fractures where some of its components have been leached from the surrounding wall rocks.

Its origin in this case appears to be similar to other phases found in the vein filling which formed during the final phases of hydrothermal activity at low temperature and pressure. The deposition of the bulk of some significant elements, such as Fe, Mn, Ba, Ca, and a portion of Au in the accompanying minerals, also occurs in these final stages of hydrothermal activity. Hence by inference, vivianite, under certain conditions, can also form as a hypogene mineral. According to Janovici et al. (1976) this is the mode of formation of vivianite at Rosia Poieni in the Apuseni Mts.

Mode of occurrence

Vivianite was first noted at Nistru in 1965 in Mine Number Two, in the northeastern portion of Vein 144 at 130 meters absolute elevation above sea level. Later, the mineral was also found in more significant quantities in certain portions of Veins 140 and 142, between Levels 230 and 130, 200 to 300 meters below the surface. Well-developed crystals also showed up at Level 230, in an area between Veins 132 and 140, along a 0.4 meter-wide subsidiary structure.

The vivianite crystals occur in the vein material attached to quartz or pyrite in the cavities and vugs of the veins, and more characteristically are imbedded in masses of clay. These vivianite crystals are associated with finely disseminated marcasite and very small barite crystals in the kaolinite mass. The size of vivianite crystals ranges from 10 to 100 mm in length with the average size being 40–50 mm. Their form is either prismatic (110) or tabular

(010) as shown in Fig. 2. Stellate crystal groups, so typical at Herja and Ilba, are not present here. Striations parallel with the c axis can be observed. The largest examples reach lengths of 100 mm, widths of 60 mm, and thicknesses of 15 mm. Crystals of similar size have only been found in Bolivia, Australia, and Japan. The crystals are light to dark green to bluish-green with yellowish to purple-green reflection.

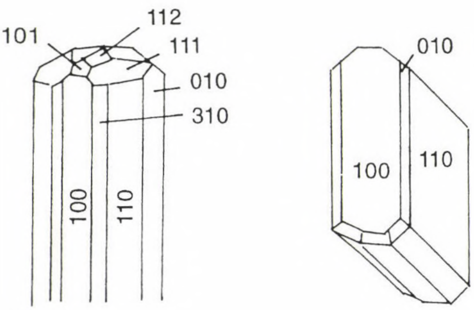

Fig. 2
Morphology of vivianite from Nistru Mine (Misztbánya), Romania

Characteristics of the Nistru material

Analyses have been carried out on crystals from Vein 142, Level 5 (130 m) on a group of crystals.

Indices of refraction are $\alpha = 1.569$, $\beta = 1.599$, and $\gamma = 1.635$; birefringence = 0.066, optically positive (determinations by László Bognár). Density is 2.76 g/cm^3, hardness is 1.8; the (010) cleavage is excellent.

Its chemical composition is as follows (analysis by Tóth Ildikó, S.C. Cuart S.A., Baia Mare, 1992 – Table 1). The chemical analysis for Fe_2O_3, FeO, P_2O_5, and H_2O were determined with classic volumetric and gravimetric methods, and for CaO and MnO with the atomic absorption method.

Table 1
Chemical composition of vivianite from Nistru Mine (Misztbánya), Romania

	Nistru	Plant City, Florida[1]	Bodenmais2	Synthetic[3]
CaO	0.55	0.02	–	–
MgO	0.12	0.25	–	–
FeO	35.07	32.64	35.65	42.96
Fe_2O_3	9.96	9.43	11.60	–
P_2O_5	28.04	29.99	29.01	28.31
SiO_2	0.66	–	–	–
H_2O	25.62	27.70	23.74	28.73
Total	100.02	100.03	100.00	100.00

[1] Janovici V., V. Stiopol, E. Constantinescu: Mineralogie, Bucuresti 1997
[2] Rammelsberger C.F: Pogg Ann. 64. 100, in C. Hintze, Mineralogie, Berlin Leipzig 1933
[3] Barth T.F.: American Mineralogist 22, 1937

Spectrographic analysis gave the following results: Fe 50%, Al < 1%, Ca < 1%, Pmajor, Si < 1%, Mg < 2%, Co 25 ppm, Ni 3 ppm, Zn 3000 ppm (analysis by N. Vladimir, S.C. Cuarţ S.A., Baia Mare).

Differential thermal analysis was carried out on a MOM syst. Paulik device in 1993 at the *Institut de Cercetări si Proiectări pentru Minereuri Neferoase,* Baia Mare. Heating was at 10 °/minute. The DTA curve (Fig. 3) shows an endothermal process at 170–190 °C, losing 24.5% water as indicated on the GT curve, a value which is in accordance with the chemical analysis. The decrease in weight is due to the prolonged elimination of water from the decomposed sample. Weak exothermal effects at 450 °C, 630 °C, 720 °C may be due (according to Todor, 1972) to oxidation processes of iron from bivalent to trivalent, without showing any weight change.

X-ray diffraction analysis was carried out on a Philips device, with Cu Kα radiation, Ni filter, diaphragm 1°, and recorded with goniometer 2 = 1°/minute, at the *Institut de Cercetări si Proiectări pentru Minereuri Neferoase,* Baia Mare in 1993. Values (Table 2) are comparable with those on card JCPDS Nr. 30-662 (1979).

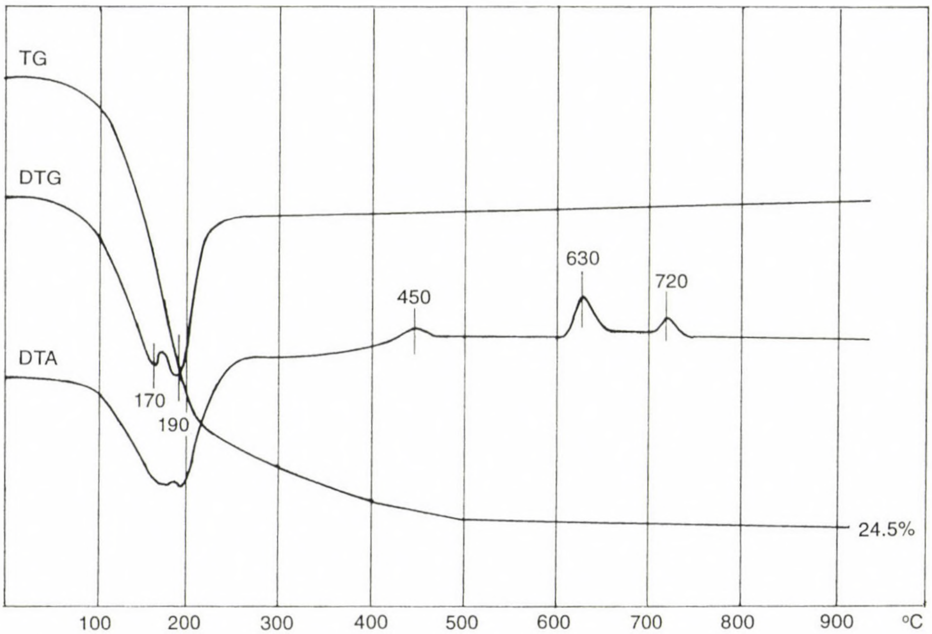

Fig. 3
Results of thermogravimetric analysis of vivianite from the Nistru Mine (Misztbánya), Romania

Table 2
X-ray powder diffraction data of vivianite from Nistru Mine (Misztbánya)

Nistru Mine		JCDP 30–662			Synthetic (Barth 1937)	
dÅ	lnt	dÅ	lnt	hkl	dÅ	lnt
7.92	m	7.93	13	110	8.00	27
6.72	s	6.73	100	020	6.80	100
4.91	m	4.90	12	200	4.91	40
4.71	m	4.558	5	101	4.50	13
4.35	w	4.341	2	011	4.32	4
4.08	m	4.081	12	130	4.09	13
3.850	m	3.849	7	101	3.84	40
3.602	w	3.768	<1	121	3.65	5
3.450	w	3.361	1	040		
		3.343	2	121	3.33	3
3.206	m	3.210	16	0.31	3.20	53
2.976	m	2.985	10	301	2.97	67
		2.960	8	211		
2.769	w	2.770	4	240		
2.720	w	2.728	9	321		
2.692	m	2.706	9	141	2.71	67
2.634	w	2.637	6	330	2.64	8
2.593	m	2.593	4	150		
2.530	w	2.530	8	141	2.52	33
2.512	w	2.514	3	231		
2.467	w	2.448	1	400		
2.421	w	2.421	6	301	2.42	40
2.313	m	2.321	7	051	2.31	27
2.297	w	2.296	1	002		
	w	2.279	1	321		
2.228	w	2.233	5	341	2.23	20
2.189	w	2.194	5	251	2.19	20
		2.173	2	022		
2.108	w	2.108	1	112		
2.069	w	2.075	4	350	2.07	23
		2.039	<1	260		
2.003	w	2.012	2	161	2.01	8
1.948	w	1.964	2	341	1.96	8
1.931	w	1.936	2	161	1.92	33
1.889	w	1.886	2	170	1.89	20
1.814	w	1.816	2	431	1.82	11
1.793	w	1.793	1	361		
1.779	w	1.786	3	451	1.78	13
		1.772	2	071		
1.678	s	1.680	6	080	1.67	40
1.659	w	1.6599	2	352		
		1.5974	3	332		
1.5788	w	1.5834	4	550	1.59	23
1.5498	w	1.5404	1	181	1.55	7
1.4977	w				1.49	12
1.4424	w				1.47	7
1.3405	m					

s = strong, m = medium, w =weak intensity

Discussion

Considering the given facts and the local appearance of marcasite specks on the vivianite, it is suggested that the vivianite found at Nistru formed as a primary mineral during the last stages of hydrothermal activity, under low pressure and temperature. The presence of marcasite and particularly wurtzite, found in abundance in the upper reaches of veins, suggests that vivianite crystals developed from acid solutions which prevailed during the high-level later stage of mineralization. Observations suggest that wurtzite, marcasite, vivianite, and tiny barite crystals found in the masses of clay minerals (illite, dickite, nacrite) formed under similar circumstances at depth.

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Acta Geologica Hungarica, Vol. 41/1, p. 147 (1998)

Forthcoming Event

The Hungarian Geological Society – a scientific association of geologists in Hungary is 150 years old (young)

The Hungarian Geological Society (HGS) together with the German Geological Society is the third oldest geologic society in Europe. It will celebrate the 150th anniversary of its foundation in 1998 under the high patronage of Árpád Göncz, president of the Hungarian Republic.

The first half of the 19th century was an exceptional period for the country when all of society (including those interested in geology) was inspired with the ideas of the reform movements. The original aim, **"to search for the wealth in nature is truth and an obligation"** is still valid but the way to achieve it has changed fundamentally.

The modest festivities will be concentrated around a spring event and an autumn one. The **ceremonial assembly** to be held on **18 March** will be a good opportunity to review the activity of the HGS. For the ceremony the oldest geologic societies of Europe and the societies of the neighbouring countries are invited. The ceremonial opening assembly will be followed by a nation-wide symposium under the title **"How far have we come?"** at the Hungarian Academy of Sciences on **19 March**. Here the results of the work of the different branches of the geology of Hungary will be presented in as compared with those of international geologic research. The spring festivities will end with the **GEO'98** program on **20 and 21 March**, which will be dedicated to the Hungarian geologists living within Hungary and abroad. The subject of the program is geoscientific education and thematic mapping.

The focal point of the autumn festivities will be an **international conference** on the subject of **"The geology of today – for tomorrow"**, from **9 to 11 September**. Four vital topics have been selected for the conference: geological questions of radioactive waste disposal, protection of drinking water resources, integrated and sequence stratigraphy, and GIS in geology. After presentation of the papers the open questions will be discussed with the help of invited keynote lecturers. The aim is to face some of the gravest questions of future society in the next century. A four-day **pre-conference field trip** and a four-day **post-conference one** will be connected to the conference, in Transdanubia and in the Alföld (Great Plain) respectively. The pre-conference field trip in Transdanubia is part of the post-congress field trip of the Carpathian-Balkan Geological Congress in Vienna.

Géza Császár

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GUIDELINES FOR AUTHORS

Acta Geologica Hungarica is an English-language quarterly publishing papers on geological topics. Besides papers on outstanding scientific achievements, on the main lines of geological research in Hungary, and on the geology of the Alpine–Carpathian–Dinaric region, reports on workshops of geological research, on major scientific meetings, and on contributions to international research projects will be accepted.

Only original papers will be published and a copy of the Publishing Agreement will be sent to the authors of papers accepted for publication. Manuscripts will be processed only after receiving the signed copy of the agreement.

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The paper complete with abstract, figures, tables, and bibliography should not exceed 25 pages (25 double-spaced lines with 3 cm margins on both sides).

The first page should include:

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- an abstract of not more than 200 words
- a list of five to ten key words
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The SI (System International) should be used for all units of measurements.

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In text citations the author's name and the year of publication between brackets should be given. The reference list should contain the family name, a comma, the abbreviation of the first name, the year of publication, and a colon. This is followed by the title of the paper. Paper titles are followed – after a long hyphen – by periodical title, volume number, and inclusive page numbers. For books the title (English version), the name of the publisher, the place of publication, and the number of pages should be given.

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Figures and tables should be referred to in the text. Figures are expected in the size of the final type-area of the quarterly (12.6 x 18.6) or proportionally magnified 20–25% camera ready quality. Figures should be clear line drawings or good quality black-and-white photographic prints. Colour photographs will also be accepted, but the extra cost of reproduction in colour must be borne by the authors (in 1998 US$ 280 per page). The author's name and figure number should be indicated on the back of each figure. Tables should be typed on separate sheets with a number.

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Acta Geologica Hungarica, Vol. 41/2, p. 149 (1998)

150th anniversary of the Hungarian Geological Society

1998 is a festive year for Hungarian geology. This year, the 150th anniversary of the foundation of the Hungarian Geological Society is celebrated.

Following the foundation of the English and French professional institutions the Society was the third to be established in Europe, contemporaneously with the German Geological Society, in 1848 – the year of the European revolutions and the Hungarian war of independence.

The Society has been continuously active in support of the development of Hungarian geology since its foundation. It organizes the combined work of the Hungarian geologists, provides a forum for professional debates, organizes international and national conferences and operates eight sections in different branches of science as well as five regional organizations. Its journal (the *Földtani Közlöny*, founded in 1871) offers a regular possibility of publishing research results and gives an overall picture of the life of the Society.

The Society maintains extensive international relationships. It had and has several renowned foreign scientists among its honorary members.

The Society is affiliated with the following international organizations: Association of European Geological Surveys (AEGS), European Federation of Geologists (EFG), European Mineralogical Union (EMU), and the International Mineralogical Association (IMA).

The anniversary will be celebrated by several professional events organized by the Society: on 18 March, a festive General Assembly will be held in the building of the Hungarian Geological Institute; on 19 March, a conference highlighting the results of Hungarian geology in the light of international geologic research will take place at the seat of the Hungarian Academy of Sciences.

In August a conference on the subject of "Geology in Natural Sciences" will be organized, primarily with the object of propagating knowledge within the general public and supporting the teaching of earth sciences in secondary schools. In September an international conference on "Geology of Today – for Tomorrow" will be held, with several days of field trips.

The Editorial Board of the *Acta Geologica Hungarica* extends its congratulations to the Hungarian Geological Society on the occasion of its sesquicentennial year of activity. In connection with this remarkable anniversary and for the benefit of the international scientific community, in the present issue and in the next one the quarterly intends to offer a survey of the history of the Society and briefly introduce the foreign honorary members of the Society who have significantly contributed to the development of Hungarian geology.

János Haas

Akadémiai Kiadó, Budapest

Acta Geologica Hungarica, Vol. 41/2, pp. 151–170 (1998)

150 years of the Hungarian Geological Society Part I: 1847–1971

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Geological Institute of Hungary, Budapest

Introduction

In Europe, the first national (not local) geological societies were founded in the first half of the 19th century. These were the *Geological Society of London* (1807), the *Société Géologique de France* (1830), and the *Hungarian* and the *German Geological Societies* in 1848.

Conception and Birth of the Hungarian Geological Society

A "Society of Hungarian Physicians and Naturalists" was founded in 1840, and held regular meetings every year in different towns of Hungary. The 8th Meeting was held in Sopron (Ödenburg) on 11 August 1847. It was on this occasion that *András K. Zipser,* a teacher of natural history from Besztercebánya (Neusohl, Banská Bystrica) forwarded a proposal to create a private society in order to organize the exploration of mineral resources, in particular of gold placers in the rivers of Hungary. The proposal emphasized that even if no economically significant result would be achieved, "we would obtain useful geological maps".

The proposal was accepted in principle. The Vice-Chairman of the Assembly, *Ágoston Kubinyi,* Director of the Hungarian National Museum in Pest, was charged with the follow-up.

Kubinyi convened five people as the Founding Committee to his family mansion at Videfalva (Vidina), about 4 km North of the town of Losonc (now Lučenec in South Slovakia).

The actual founding meeting took place on 3 January 1848. The aims and tasks of the Society were presented and approved. A Founding General Assembly was foreseen for 18–19 August 1848, in Pest.

However, this was rendered impossible by the adverse political developments. On 11 July *Lajos Kossuth* pronounced in his famous speech: "Our homeland is in danger!" The military conflict between the first independent government of Hungary and the Imperial and Royal Court of Austria was imminent. In fact the (unsuccessful) War of Independence began in late September 1848.

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Meanwhile in Austria, the Court approved the founding of an Imperial Geological Institute (*Geologische Reichsanstalt*). It was formally established on 1 December 1849.

W. von Haidinger, the first Director of the Imperial Geological Institute, wholeheartedly supported Kubinyi's version of the proposal to create a Geological Society in Hungary. He commissioned *M. von Hörnes*, assistant curator of the Museum of Natural History in Vienna, to represent him at the renewed Founding Committee on 24 May 1850.

Close links were foreseen between the new Society with the Imperial Geological Institute in Vienna on the one hand and with the Hungarian National Museum in Pest on the other.

The First General Assembly took place on 6 July 1850 in Pest. Á. Kubinyi was elected President and *Gy. Kováts* Secretary of the Hungarian Geological Society.

The Second General Assembly on 3 September of the same year unanimously adopted the carefully elaborated Statutes. It is worth mentioning that the exploration of mineral resources was no longer the primary aim. This came only after "the detailed studying of areas of geologic interest and the systematic collecting of minerals, rocks and fossils".

A young Hungarian lawyer and graduate of the Selmecbánya (Schemnitz, Banská Štiavnica) Mining Academy, *József Szabó*, was elected Second Secretary.

By the end of the year the Hungarian Geological Society had a total of 76 members. Among these were eight Honorary Members including *W. Haidinger* and *F. Hauer*, who was in charge of the geological mapping of Hungary.

The Heroic or Pioneering Age (1850–1862)

During the first years the main activities of the Society were to organize field trips and meetings with scientific lectures. The first geological excursion went to the famous wine district of the Tokaj Hills. The first scientific meeting took place on 15 July 1851. The very first speaker was W. Haidinger, the Director of the Imperial Geological Institute in Vienna. Another important activity was to collect rocks, minerals and fossils. Most of these were deposited in the Hungarian National Museum; a smaller part went to Vienna.

It was decided to publish the lectures in bilingual (German and Hungarian) issues. However, the first publication was simply an *Activity Report* ("Első jelentés") on the first two years of the Society, edited by the Secretary, Gy. Kováts, and published in 1852. In the summer of 1852 geological fieldwork was carried out on the estates of Prince *P. Eszterházy*, one of the main supporters and sponsors of the Society, at his request. It was led by the dynamic Second Secretary of the Society *József Szabó*.

In 1856, the Statutes were modified. The new version cancelled the formal links with the Imperial Geological Survey in Vienna. It is remarkable, however, that informal links became increasingly effective: scientific spontaneity worked

József Szabó

better than politically determined obligation. This is typical behavior for Hungarians, repeatedly manifested during the history of the Society. The closest links during the entire first decade were with the Hungarian National Museum which benefited from the collected rocks, fossils and minerals while in its turn offering an office to the Society as its first home, Secretary Gy. Kováts being an employee of the Museum.

Also in 1856 appeared the first issue of the "*Proceedings of the Hungarian Geological Society*" (A Magyarhoni Földtani Társulat Munkálatai), including four scientific articles in two languages on 72 pages. This was welcome, but it was not followed by others for years, although a second issue had been announced in the "Introduction" by Gy. Kováts. The members of the Society living in various parts of Hungary, who kept on generously supporting the Society with financial contributions, had virtually no communication with the Secretary sitting in the National Museum in Pest, who was overwhelmed by his duties as one of the curators of the Museum and whose health was continuously deteriorating.

The Impact of a Great Man (1862–1894)

It was in 1862 that J. Szabó took over from Gy. Kováts the entire secretarial work of the Society. He edited the second issue of the "Proceedings" which was published in 1863. It contained 13 articles on 208 pages (three times more than the first one). Four of the papers were written by J. Szabó himself.

The number of (all classes of) members of the Society was 181 in 1863. At the 1866 General Assembly President Á. Kubinyi was succeeded by his brother *F. Kubinyi.*

The historical compromise between the Hungarian nation and the Austrian Court in 1867 opened new vistas. The year 1869 marks an important milestone in the history of Hungarian geology: the founding of the *Royal Hungarian Geological Institute* by King *Franz Joseph I. M. von Hantken*, who was at that time the Secretary of the Society, was appointed its first Director. He was also the editor of the Proceedings, the fifth and last issue of which was printed in 1870.

It was indispensable to redefine the aims and tasks of the Society and to determine its relations with the newly founded Institute. The General Assembly of 9 November 1870 faced this challenge. M. Hantken proposed to organize itinerant ("ambulatory") meetings in different parts of the country, and eventually to create regional branches of the Society. A motion was accepted that the Proceedings should be replaced by a regularly published periodical, the "*Földtani Közlöny*" (Geological Bulletin). The first volume was printed in 1871 in Hungarian. Later it became bilingual, Hungarian and German.

Papers on regional geology were supposed to be published in the *Annals (Yearbook)* of the Geological Institute, to be received by all members of the Society. The library and the collections of the Society were transferred to the Institute (a lengthy process which ended only in 1875).

At the General Assembly of 26 April 1871 an election of officers was held. *Frigyes Reitz* became the President and József Szabó was promoted to Vice President. The first "itinerant" meeting of the Society took place at Selmecbánya (Schemnitz, Banská Štiavnica), 6–12 August 1871, with a geological excursion to its surroundings. The creation of a *Selmec Regional Branch* ("Fiókegylet") was proposed and unanimously approved.

The 100th scientific meeting of the Society was held on 24 January 1872. In the same year *D. Štur,* Director of the Imperial Geological Institute of Vienna and of Slovak nationality, was elected Honorary Member of the Society.

At the *World Exhibition* in Vienna in 1873 the Society was awarded two medals, for the presentation of the coal deposits of Hungary and for its publications. József Szabó, as Vice President of the Society, with F. Reitz as the (rather passive) President, was able to make free use of his extraordinary talent as a versatile and consequent manager in the best sense of the word.

An itinerant meeting was held in the Metalliferous Mountains of Transylvania (2–14 August 1874). By the end of that year the Society had 301 members. It was in 1876 that the Society organized for the first time a geologic excursion for foreign (German) geologists (Budapest–Esztergom).

The main event of the year 1878 was the *First International Geological Congress* convened in Paris (between 29 August and 8 September). J. Szabó, who delivered a lecture entitled "Classification et chronologie des roches éruptives tertiaires de la Hongrie", was elected one of the Vice Presidents of the Congress and a member of the Commission on Geologic Nomenclature. M. Hantken became a member of the Commission on the Unification of Geologic Maps. In the same year, at the World Exhibition in Paris the Hungarian exhibit presenting the state of mining and metallurgy in Hungary was honored with a Gold Medal.

In 1879 the Board decided that in the "Földtani Közlöny" papers could also be published in English or French (the first French paper was J. Szabó's "Classification macrographique des trachytes"). In the following year a motion was adopted to start a second journal called "*Földtani Értesítő*" for science popularization papers, society news etc., the "Földtani Közlöny" being reserved for original scientific communications. Unfortunately the new journal survived only three years.

After the Zagreb earthquake in 1880 the Society became interested in earthquakes. With the financial support of the Hungarian Academy of Sciences a Commission on Earthquakes was established on 9 November 1881 with J. Szabó himself in the chair. Data on Hungarian earthquakes began to be published in 1883.

Beginning with January 1883 the "Földtani Közlöny" became a joint bulletin of the Society and the Institute and started to publish the *Annual Reports* of the Institute. However, this situation lasted no more than three years. At the same General Assembly J. Szabó was elected President of the Society. (It is interesting to mention that *K. Ettingshausen, F. Richthofen* and *K. Zittel* were elected Honorary Members, along with *Andor Semsey*, the great Hungarian Maecenas of the Society and the Institute.)

In 1884 *Gy. Halaváts* compiled the first "*Alphabetic Index of the Publications of the Hungarian Geological Society*", and in 1885 the number of the members reached 400.

J. Szabó paid considerable attention to the production of geological maps. The Selmec Branch produced one for that region. The preparations of a 1 to 1 million scale "*Geologic Map of Hungary*" were begun and a Map Committee was set up, chaired again by J. Szabó himself. The idea was also to contribute to the planned Geologic Map of Europe.

At the General Assembly held on 9 January 1886 a remarkable breakthrough occurred concerning foreign Honorary Members. Apart from one Austrian (*E. Suess*) and two Germans (*H. von Dechen* and *E. Beyrich*), among the new Honorary Members were *A. Daubrée* and *E. Hébert* from France, *W. Blanford* and J. Prestwich from Great Britain, G. Cappellini and G. Meneghini from Italy, and *J. Dana* and *J. Hall* from the United States of America. In this way J. Szabó succeeded in making the Society a truly international one. It was also decided to hold commemorative sessions and to publish obituaries of outstanding deceased members of the Society.

Unfortunately enough, J. Szabó himself was among the first to be honored in this manner. He passed away on 11 April 1894. The grateful members of the Society collected funds and created a József Szabó Memorial Fund, the bylaws of which were adopted in 1897. One part of the interests was to be used to support scientific research and the other to produce a *J. Szabó Commemorative Medal*, to be awarded for some outstanding geological publication, for the first time in 1900.

The Society from the Eve of the Millennium of Hungary until the collapse of the Austro-Hungarian Monarchy (1895–1918)

The General Assembly held on 6 February 1895 elected *János Böckh* to succeed to Szabó as the President of the Society. This was for the first – but certainly not for the last – time that the Director of the Institute was also the President of the Society – a very efficient personal union if the right person was involved.

The *Geologic Map of Hungary* compiled by 1888 was printed in 1896 under the guidance of *Lajos Lóczy*. It was presented to the public on the occasion of the *Millenary Exhibition in Budapest* and was awarded a medal.

It is characteristic of the extremely strong position of J. Böckh that at the International Geological Congress held in Saint Petersburg in 1897 he represented in one person the Ministry of Agriculture, the Hungarian Academy of Sciences, the Royal Hungarian Geological Institute and the Hungarian Geological Society (there were, however, two other participants from Hungary). In 1900 the first J. Szabó Commemorative Medal was awarded to none other than Böckh for two of his regional geological studies, while the Society itself obtained a Gold Medal from the *Paris World Exhibition* for the *Geological Map of Hungary.* The new headquarters (a remarkable Art Nouveau-style palace designed by the architect *Ödön Lechner*) of the Royal Hungarian Geological Institute was inaugurated by the king himself on 7 May 1900, and at the General Assembly of 6 February 1901 *Lajos Telegdi-Roth* was elected President of the Society.

In the 1902 volume of the "Földtani Közlöny" *Antal Koch* published a 23-page paper on *"The History of 50 years of Activity of the Hungarian Geological Society"*. Because of its decreasing membership, the Selmec Branch of the Society was dissolved by the General Assembly on 14 January 1903. The task of observing earthquakes was taken over by the National Institute of Meteorology, and the Commission of Earthquakes of the Society was transformed into a *Seismological Observatory.*

In 1904 Antal Koch became the President of the Society. By that time the membership dropped from 400 to 309, but during the following years began growing again.

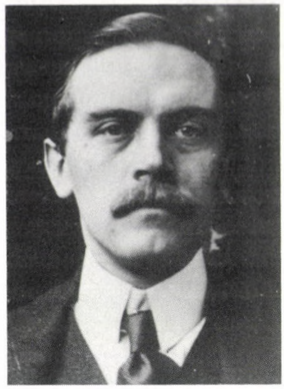
The Society co-organized with the Institute the *First International Conference on Agrogeology* held in Budapest in 1909. After the death of J. Böckh in that year *Ferenc Schafarzik* was elected President on 10 February 1910. In addition, in 1910 a *Speleological Committee* was set up under the chairmanship of *Mihály Lenhossék.* It was soon promoted to a Section of the Society with 133 members and a regular publication entitled *"Barlangkutatás"* (Cave Research). This required a modification of the Statutes which was approved in 1914. Membership climbed to a record number of 741 in 1912. Lajos Lóczy's monumental work on the geology of the Lake Balaton region was honored by the J. Szabó Medal in 1915.

The First World War had a strong negative impact on the Society as well. Funding was reduced, printing costs rocketed, and the membership decreased (a considerable number of the members had to do military service). *Tamás Szontágh* was elected President in 1916. In 1917 the Board of the Society decided to organize a *Hydrological Section,* with 82 members. The first President was *Sebestyén Aladár Kovács.* Membership culminated with 755 in the year 1918.

After the collapse of the Austro-Hungarian Monarchy and the short-lived Hungarian Republic of Count *Mihály Károlyi* (October 1918 – March 1919) the Hungarian Soviet Republic was established on 19 March 1919.

From the Consequences of the lost First World War to Those of the Second (1919–1948)

On 1 April 1919 the Board of the Hungarian Geological Society had to cede its rights to a Directorate. Its members were *Jenő Jablonszky, Lajos Reich* and *Elemér Vadász*. After the 133 days of Communist rule the previous situation was reestablished. Three members left the Society of their own accord and nine others were expelled, among these *Tivadar Kormos* and Elemér Vadász. The General Assembly held on 20 April 1920 elected *Móric Pálfy* as President.



Aladár Vendl

The Society had to make a new start. As a consequence of the Trianon Peace Treaty Hungary lost two-thirds of her territory, including practically all mining districts. The Society lost a considerable number of its members. The financial situation became dramatic. However, interest in geology was soon restored: the truncated country badly needed new mineral resources and a more reasonable utilization of the known ones.

At the General Assembly held on 7 February 1923 *Béla Mauritz* was elected President. The Statutes were once again modified. In 1926, the membership totalled 411, indicating a dramatic drop in comparison with the pre-war period. The Speleological Section was dissolved in May 1926 because an independent *Speleological Society* was created. An offspring of the Hungarian Geological Society, it has operated successfully up to the present day.

Simon Papp

Aladár Vendl took over the Presidency in 1932. A Finnish geologist, *Arne Laitakari*, was elected Honorary Member of the Society. The late thirties were years of relative consolidation. Apart from strictly professional lectures popularizing ones were also held for the general public (until 1943). The "*Földtani Értesítő*" was re-launched in 1937.

A. Vendl resigned at the General Assembly held on 15 February 1940. *Károly Papp* became the new President. However, he was very soon (on 3 March 1941) replaced by *Simon Papp*. Between 1942 and 1945 Hungarian geologists performed field work in Carpathian and Transylvanian areas re-attached to Hungary by the first and the second Vienna Dictates. Some of the results were published in the "Földtani Közlöny".

The military operations passing through Hungary in late 1944 and early 1945 made all scientific activities impossible and left the entire country in ruins. The first step of revival was a Board Meeting held on 15 August 1945, chaired by *Jenő Noszky*. A General Assembly was convened for 19 September 1945. It was unanimously decided to cancel all sanctions taken in August 1919 against those members of the Society who had been protagonists of the Hungarian Soviet Republic. The General Assembly elected *István Vitális* as President. Among others the rehabilitated Elemér Vadász also became a member of the Board.

The year 1946 was marked by galloping inflation which made publication virtually impossible. In 1947 elections were held again. Simon Papp became President with Elemér Vadász and *Tibor Szalai* as Vice Presidents. On 3 January 1948 a modest celebration was held in the Geological Institute commemorating the one hundred years anniversary of the founding of the Society. The two Vice Presidents were the speakers. The annual General Assembly was held on 11 February in the Institute without any solemn centenary celebration. A new umbrella organization, *MTESz* (Federation of Technical and Scientific Societies), created by the State, was inaugurated on 29 June 1948. The affiliation of the Hungarian Geological Society was announced by its Second Secretary *Géza Szurovy*.

This was the year when the Hungarian Communist Party seized political power in Hungary and persecutions began. One of the first victims was the President of the Society: Simon Papp had to resign and to leave the Society. Accused of sabotage he was condemned to death based upon trumped-up charges. The death sentence, however, was commuted to life-long imprisonment. The Board took notice of S. Papp's resignation on 20 October 1948. Elemér Vadász took over as Acting President until the next General Assembly which was to be held in 1949 (see Annex I).

In this dramatic way ended the first century of the Hungarian Geological Society, duly described by its former President Aladár Vendl in his booklet "*History of the One Hundred Year Old Hungarian Geological Society*" (276 p.), which was published in 1958.

Dramatic Transformations Inside and Outside (1949–1958)

The II Centenary General Assembly took place on 16 February 1949. The Hydrological Section announced their intention to found an independent *Hungarian Hydrological Society*. This became a second offspring of the Hungarian Geological Society, also prospering until the present day. The Assembly unanimously accepted the list of the new Board. Accordingly Elemér Vadász became the President assisted by Tibor Szalai and *Elemér Szádeczky-Kardoss* as Vice Presidents and Géza Szurovy as the First Secretary. With this act a fundamentally new era began.

In his presidential address E. Vadász reassured the Assembly that according to Stalin geology could be considered a science. He also quoted a declaration by *Mátyás Rákosi*, Prime Minister of the Communist Government, that the nationalized companies would have to provide the Society with more financial support than the private companies of the past had.

At the Session of the Board on 6 April 1949 two Working Committees were set up: one for *Geological Education* and another for the *Geologic Dictionary* (in fact, *Terminology*). On 7 June 1949, the Paleontological Section of the Society held its inaugural meeting. *K. Telegdi-Roth* was elected President. 4 November 1949 was the centenary of L. Lóczy's birth. On this occasion a commemorative session was held.

The new, "industrial" orientation is well characterized by the fact that the annual Itinerant Meeting was held at Miskolc, the metallurgical center of Northeast Hungary, on 13 November. The political situation made inevitable the holding of a joint session with the Hydrological Society to celebrate – the 70th birthday of *J. V. Stalin*. In 1949, the membership of the Society dropped to 239.

The main steps of the communist-directed "reorganization" having been accomplished a *Centenary General Assembly* could be held. It took place on 30 April 1950. President E. Vadász was very reassuring again. Dividing the 100 years of the Society into (1) a pioneering-heroic, (2) a romantic, (3) a tragic, (4) a declining period (!), he nonetheless declared that "we are not lagging too far behind the vanguard Soviet geology". He was re-elected President, with Elemér Szádeczky-Kardoss and *Ernő Ács* as Vice ("Associate") Presidents. First Secretary Géza Szurovy stepped down and was replaced by *György Kertai* who was to later become one of the most prominent Presidents of the Society.

The Hungarian name of the Society was changed on Vadász's proposal. The original had been "Magyarhoni Földtani Társulat", literally "Geological Society of Hungary", the new version was "Magyar Földtani Társulat", i.e. simply "Hungarian Geological Society". This was an obvious antitraditionalist move.

At the General Assembly of 6 June 1951 E. Vadász admitted that the reorganization "has been not successful so far". He emphasized the importance of (1) criticism and self-criticism, (2) applying dialectical materialism to geology, (3) improving the scientific standard and updating the terminology of the lectures, and (4) dealing seriously (following the Soviet example) with the history of Hungarian geology. After all that he resigned, and with him the entire Board.

In his Secretarial Report Gy. Kertai pointed out the ongoing reform of geological education at both the secondary and the university levels, the increasing demand for linking scientific research to industrial practice, and – last but not least – the need to mercilessly fight the political enemy. The Assembly elected Elemér Szádeczky-Kardoss President, Sándor Vitális and Sándor Koch Associate Presidents, and re-elected György Kertai as Secretary.

In October of the same year Szádeczky-Kardoss delivered a presidential address: "Geochemical Directives in the Exploration of Mineral Resources". He urged a creative, synthesizing, and innovative approach and pointed out some positive achievements (better understanding of geology within the national administration and an accelerated increase in the number of earth science students and in the staff of geological institutions, as well as in the publication of geologic works.)

It is worthy of mention that Szádeczky-Kardoss declared – in a remarkable presidential address delivered at a solemn session of the Month of Hungarian-Soviet Friendship (!) – that "lack of a critical attitude toward Soviet science is a danger. It discredits not only the individual who commits it but also Soviet science itself."

At the General Assembly held on 4 June 1952 Gy. Kertai presented a very detailed Secretarial Report. This was very cleverly introduced by a philosophical exercise applying dialectics according to Engels and Lenin to geological processes, with particular regard to "qualitative jumps". He announced a new form of activity: the "ankets" (in fact, symposia and round table discussions). Four meetings of this type were held: one on geologic education, another to discuss the tasks and duties of geologists, and two on recently published geologic books.

The result of the elections held was as follows: President: Sándor Vitális, Associate Presidents: *Aladár Földvári* and György Kertai, Secretary-General: *Béla Jantsky*. The latter outlined the most important tasks as shown below:

1. Closer cooperation with technical societies
2. Operative links with the geologists of the other "Socialist" countries
3. Training courses at all levels
4. Thematic prizes to be announced
5. More field trips to be organized

B. Jantsky – who had graduated in Czechoslovakia – declared: "Just as there is no Hungarian culture without (the poets) Petőfi and Ady and (the composer) Kodály, Hungarian geology can develop only relying upon the tradition of great Hungarian geologists." He also proposed to re-launch (for the second time) the periodical "*Földtani Értesítő*" which had been stopped in 1946. However, this never happened.

The highlight of the year 1952 was the *establishment of five new Sections* of the Society (beside the already operating Paleontological Section), announced on 3 September 1952. These were:

Section	Chairman	Secretary
Speleology	László Jakucs	Sándor Leél-Össy
Mineralogy & Petrography	Kálmán Sztrókay	Gábor Bidló
Geochemistry	László Tokody	László Soós
Petroleum Geology	László Strausz	János Szabó
Coal Geology	Ferenc Sólyom	Viktor Dank

There was no General Assembly held in the politically troubled year 1953. It was attempted (without success) to create a Geophysical Section.

The 1954 General Assembly took place on 3 May, opened by A. Földvári, who announced "the advent of vertical geology after the horizontal kind", and urged measures to reintroduce geology in the curriculum of secondary schools. First Secretary B. Jantsky admitted that the system of meetings held by thematic sections did not work and that it was inevitable to return to the old system of "central" meetings. He deplored the lack of critical spirit and the preference to details instead of a synthesizing approach.

In fact the second attempt at reorganization had also revealed itself a failure. As a result E. Vadász was elected President. Beside the two Associate Presidents *Ferenc Horusitzky* and Kálmán Sztrókay the new post of a "Managing President" was created for *András Tasnádi-Kubacska. József Fülöp* and *István Pálfalvy* became the Secretaries. Vadász scourged the Assembly: "in most of us, consciously or unconsciously, the damnable spirit of the near past is still surviving".

Volume 85 of the "Földtani Közlöny" was dedicated to the 70-year old E. Vadász. As E. Szádeczky-Kardoss said in his birthday greeting: "Such a personality may have adversaries, but there is nobody who would not admit his great talents and the value of his work." Vadász was awarded the Order of the Red Banner. In one of his papers he wrote: "We are still stuck in unilateral professionalism, without understanding and applying the approach of the Party."

A new type of meeting appears in these years: the "club afternoons" or "club evenings", with slide projections of members who in one way or other succeeded in getting abroad on mission or study tour, some even to the West, beyond the Iron Curtain.

There was no General Assembly convened in 1955 and 1956, but Itinerant Meetings were held (at Pécs and Miskolc, respectively). In 1956 Hungary became member of the *Carpathian–Balkan Geological Association* which was re-established in new form at the International Geological Congress held in Mexico City (Hungary was represented by *Gy. Pantó*).

The political troubles of 1956–57 (demand for political reforms, ("counter")-revolution on 23 October 1956, renewed Soviet military intervention on 4 November 1956, reprisals and repression by *J. Kádár*'s government) seriously hampered the activity of the Society as well. The General Assembly foreseen for 3 November 1956 had to be cancelled (the second, decisive intervention of the Soviet Army came on 4 November). A symbolic act was the deposition of flowers on the graves of M. Hantken and K. Hofman. The "Földtani Közlöny" published reviews on two books edited in Bucharest about the mineral waters, natural gas shows and mineral resources of the Hungarian Autonomous Province of Romania (which was later abolished).

By the autumn of 1957 the situation in the Society was already almost back to "normal". At the commemorative session of 4 December 1957 E. Vadász delivered a speech entitled "Retrospection". He spoke about "indifference" and "stagnation", "reticence", "animosity" and "alienation" of the older generation, and of "formalism", "sterility", and "no connection with Society" of the younger one. He concluded: "My guidance remained unsuccessful". Nevertheless he ended his speech on a more optimistic note: "The disciples stand in a new world, before a life full of hope, where the light of Mind will shine."

It was on 21 March 1958 that the next General Assembly was held. As reported by First Secretary J. Fülöp the Society had 233 paying members in 1954 and 440 in 1958. The elections produced rather odd results. *Ferenc Horusitzky* was elected President, with three Associate Presidents (György Kertai, *János Meisel* and Kálmán Sztrókay), a Managing President (*Endre Lengyel*) and a three-member Presidential Committee (*Kálmán Balogh, László Bogsch* and *Gábor Pantó*). Even more strangely, four secretaries were elected, namely *Jenő Boda, Viktor Dank, Pál Kriván* and *Gusztáv Morvai*. This was obviously a result of a laborious compromise. The disillusioned Vadász was promoted to lifelong Honorary President of the Society and ceased to intervene in the management of the Society.

A successful Itinerant Meeting was held in Szeged (21–23 June) and a substantial Hungarian delegation attended the 4th Congress of the Carpathian Balkan Geological Association in Kiev (15–29 September 1958).

The resignation of President F. Horusitzky in October marked the end of this difficult period.

Unprecedented Boom (1959–1971)

The year 1959 began with an important event: the first new regional branch of the Society was established. This was the *Pécs (or Mecsek) Group* (SE Transdanubia) founded on 5 March 1959.

Between 15–23 September there took place in Budapest a *Conference on the Mesozoic* "of international character", organized by the Hungarian Geological Institute. This was the first large-scale "opening" towards the international community: it was attended by 70 foreign geologists from 10 countries in addition to the 170 Hungarian participants.

The Commission on Geochemistry of the Hungarian Academy of Sciences organized another scientific meeting "of international character": a *Geochemical Conference* (5–10 October). It had 34 attendees from 12 foreign countries and 80 Hungarian participants. A meeting held on 11 November dealing with modern lunar research ("selenology") was the sign of another kind of opening (in this case, a thematic one). The membership of the society attained 511.

On 12 January 1960 a *Section of Clay Mineralogy* was formally established. *Ernő Nemecz* was elected its Chairman. The General Assembly held on 9 March 1960 elected seven foreign Honorary Members. Characteristically enough four of them were from the USSR, two from Czechoslovakia, and only one from the West – from France. It was a turning point in the life of the Society that petroleum geologist György Kertai was elected President and Gusztáv Morvai First Secretary. For the first time mention was made of "Hungarian geologists working abroad, e.g. the successful operation of the expedition in China" (oil exploration). The 1960 Itinerant Meeting was held at Eger and Aggtelek (21–23 October).

Participation in the *International Geological Congress in Copenhagen* resulted in a strong impact of Hedberg' s lithostratigraphic concepts on Hungarian geology (advocated above all by J. Fülöp).

György Kertai

In 1961, a new genre was introduced: the "announcements", for short scientific communications both to be presented at meetings and published in the Földtani Közlöny. The General Assembly of 15 March 1961 was dedicated to József Szabó. The 1961 Itinerant Meeting visited the Zala oil fields (30 June – 2 July). On 11 August, a second Regional Branch of the Society was established at Nyirád: the Central Transdanubian one. The first elected officers were Ernő Nemecz (President) and *Béla Vizy* (Secretary).

Delegations of the Society attended the *Fifth Congress of the CBGA in Bucharest* and a *Volcanological Conference* in Italy. On 12 October, the North Hungarian Regional Branch was set up (Chairman: *János Mónos*, Secretary: *Kálmán Verebélyi*). Another interesting event was the meeting of the Presidents of the Geological Societies of the Socialist Countries in Bratislava, Czechoslovakia.

The Mecsek Group organized a meeting on 13–14 December on the exploration and mining of hard coal in the Mecsek Mountains. Training courses were held for graduate and secondary level geologists ("technicians").

A new thematic Section, that of *Engineering Geology*, held its first meeting in January 1962 (Chairman: *Ferenc Papp*, Secretary: Béla Jantsky). A General Assembly took place on 9 May 1962. The original name of the Society was enthusiastically restored. A M. Hantken Commemorative Medal was proposed and accepted. In July the eminent French volcanologist *Haroun Tazieff* gave lectures for the Society.

On 8–13 October a *Scientific Conference on Oil Mining* was held, co-organized by quite a number of Hungarian, Czechoslovak and Polish institutions. Lectures by visiting foreign experts became regular events (*A. Cailleux* and *E. Roch* from France, *G. Răileanu* from Romania, *V. Semenenko* from the USSR, etc.).

An Electoral General Assembly was held on 27 March 1963. The first *M. Hantken Commemorative Medal* was awarded to *Barnabás Géczy*. According to the very detailed Secretarial Report by G. Morvai, membership was at 941 – a figure higher than ever, indicating a fantastic increase. During the 1960–63 term, 29 percent of the lectures were already being delivered at meetings of the regional branches. The thematic composition was also changing:

- Paleontology-stratigraphy: 26%
- Mineralogy-petrology-geochemistry: 23%

- Applied geology: 16%
- Reports on missions abroad: 12%, etc.

The General Assembly granted the right of vote to the "junior members" (mostly university students and technicians). Gy. Kertai was re-elected President and P. Kriván was promoted to First Secretary.

The number of missions abroad increased considerably (a Committee was created for International Relations). In 1963 the Society was represented at 11 scientific meetings outside Hungary, from Stockholm to Rome and from Warsaw to Lyon. An important development was that the *Section for Petrography and Magmatism of the CBGA* held a meeting in Hungary (27–30 May). An informal "itinerant meeting" was held at Szolnok (17–18 May 1963), and in December, the *Section for Mineralogy and Geochemistry* was re-activated under the chairmanship of Kálmán Sztrókay. Several joint meetings were held with the Hydrological Society, the Mining and Metallurgical Society and even with the Society of Machine Industry.

The International Union of Geological Sciences (IUGS) held its first ordinary Council Meeting in Rome on 14–15 October 1963. Although Hungary was among the 26 countries which were not represented (22 were) the affiliation of Hungary to IUGS was ratified by 1 January 1964. In that year a *Working Committee on Coal Petrography* began operating.

A successful initiative was for the Society to join the annual Reporting Meetings of the Hungarian Geological Institute (MÁFI) and the Hungarian Geophysical Institute (ELGI), comprising a large number of presentations on widely varied topics. In 1963 one member obtained a US fellowship (Woods Hole Oceanographic Institute) and some missions went as far as China.

The Itinerant Meeting of 1964 was held in Western Hungary (Celldömölk–Sopron) on 28-31 May. It was attended by 253 people (!), including – and this has to be pointed out specifically as a most welcome development – 27 from Austria and 38 from Czechoslovakia. Among the foreign lecturers were (along with experts from the GDR, Austria and the USSR) for the first time a US geologist: *Ch. Vitaliano* (26 August 1964) on 24–27 September 1964 the Mecsek Group organized the *First Meeting of Hungarian and Yugoslav geologists*: a significant breakthrough after more than 15 years of lost contact due to Tito's seclusion from the "Soviet Block".

A General Assembly was held – rather unconventionally – on 16 December 1964. President Gy. Kertai gave an objective overview on the "state and tasks of geologic research in Hungary" (by that time he was also the President of the Central Office of Geology, directly responsible to the Council of Ministers, i.e. to the Government). He stressed the importance of the development of international relations in general and of the "brain drain" in particular. He reported that Hungarian geologists and geophysicists were already at work in Cuba, Guinea, Iraq, Mali and Mongolia. First Secretary P. Kriván reported on publications of the Society beside the traditional issues of the "Földtani

Közlöny". These were: a *Register (Index) of the Földtani Közlöny*, a *Special Issue on Clay Mineralogy*, and the almost regularly published *Őslénytani Viták (Discussiones Palaeontologicae)*, started in 1963. Ten (!) new foreign Honorary Members were elected: 4 from Austria, 2 from Poland, 1–1 from Italy, Norway and the USSR. The "accent drift" is conspicuous.

The Bylaws of the *Maria Vendl Memorial Fund* were presented and adopted (M. Vendl was a mineralogist, the first woman in Hungary to become University Professor, at the University of Debrecen. She passed away in 1945, and the Fund was created by her elder brother, a former President of the Society, A. Vendl). 1964 was also the first year of the appearance of the periodical "*Engineering Geological Review*" (Mérnökgeológiai Szemle).

On 1 March 1965 a special session greeted E. Vadász on the occasion of his 80th birthday. The "club afternoons" presented slides from California to Guinea and India (on the occasion of the International Geological Congress held in New Delhi in 1964). A special meeting (with 12 lectures) was devoted to Micropaleontology on 21 April 1965. The annual Itinerant Meeting went to Tokaj on 19–22 June, with the remarkable innovation of the excursion crossing the border to Czechoslovakia. On 18–23 October the *IUGS International Commission on Coal Petrology* held a meeting in Hungary.

The General Assembly took place on 8 December 1965. P. Kriván pointed out a very important development: *bilateral cooperation agreements* were signed with the Geological Societies of *Austria and Yugoslavia* in 1964 and with those of *Czechoslovakia, the GDR and Poland* in 1965. He declared: "We intend to promote the development of links with the USSR and Romania as well."

A *Propaganda Committee* was set up. In 60 Grammar Schools Geological Circles were active for three years. It was proposed to proceed to a modification of the Statutes of the Society: its structure had become more complicated than ever, and "during one three-year term as many lectures are held as were earlier during half a century". Also in that year the first M. Vendl Prize was awarded to *László Tokody*.

In 1966, an electoral General Assembly took place on 23 March. In his presidential address, Gy. Kertai presented the following statistics:

– In 195 in Hungary there were registered 472 graduated geologists and 155 geophysicists, altogether 627 people engaged in geonomy (he considered this term preferable to "geoscience"). Included in this number 501 were males and 126 females. Of these 53 percent were working in industry, 28 percent in research institutions and laboratories, 10 percent in administration (including the museums) and 9 percent in education.

According to the Secretarial Report presented by P. Kriván the membership of the Society totaled 1,002. As a result of the elections, Ernő Nemecz replaced Gy. Kertai as President, while P. Kriván was reelected First Secretary.

The *Second Meeting of Hungarian and Yugoslav geologists* occurred on 10–15 May 1966, this time in Zagreb. The annual Itinerant Meeting was held at Balatonalmádi on 6–8 September, focussing on the problems and achievements of bauxite geology. On 17 November the fourth Regional Branch of the Society was founded at Szeged, chaired by Sándor Koch.

The year 1967 started with the establishment of a *Section for Economic Geology* on 11 January. *Gyula Varjú* became its first Chairman. At the General Assembly held on 16 March 1967, E. Nemecz gave an assessment of the situation of geology in Hungary, pointing out the worldwide tendencies of increasing specialization, the striving for exactitude, and modeling. In Hungary the technical level of laboratory instrumentation was rapidly improving. The emphasis was on petroleum and coal geology. He launched an appeal to be prepared for the International Geological Congress which was to take place in Prague next year, in August 1968.

P. Kriván in his Secretarial Report mentioned that the Section for Mineralogy and Geochemistry became affiliated to the *International Mineralogical Association (IMA)*. He forwarded the prepared modification of the Statutes. After discussion it was approved.

This year was characterized by quite a number of colloquia and workshops. These were: *Paleoecological Colloquium* (17–18 April), *Hydrogeological Conference* (8 May), *Colloquium on Oil and Gas Geology* (15 May), the *Week of Geology* (15–21 May), *Workshop on "Computerized storage and processing of data in geological research and exploration"* (25–26 May), *Workshop on the Ózd Coal Basin* (7–8 September), *Colloquium on Evolution* (4–5 December), *Workshop on the Engineering Geologic Problems of Budapest* (6–7 December).

At the General Assembly of 13 March 1968, First Secretary P. Kriván presented an in-depth analysis of the development of the Society, with particular regard to the thematic sections and the regional branches. Membership had risen to 1187: this was five times as many as in 1948! On 14 March a workshop was held on the possibilities of the regional development of the town of Pécs and Baranya County (SE Transdanubia). On 11 May 1968 Gy. Kertai passed away unexpectedly, aged only 56.

The North Hungarian Regional Branch organized two workshops: on the utilization of the water lifted from coal mines (13 May) and on the long-term exploration prospects of NE Hungary (9 August). The Itinerant Meeting went from the town of Szeged to the Danube–Tisza Interfluve (3–5 September). Mission and study tour reports included exotic regions ranging from Iceland to Kazakhstan and passing through the "Hot May" of student demonstrations in Paris.

The preparations for the International Geological Congress in Prague were frustrated by the military intervention of the troops of the Warsaw Pact countries on 20/21 August 1968. The Congress was interrupted and instead of the planned four only one geologic field trip came to Hungary since most of the participants escaped to the west as soon as they could. (It is typical for the times that in the regular chronicle of events in the "Földtani Közlöny" not a single word has been published about this.)

The General Assembly of 26 March 1969 re-elected E. Nemecz as President and P. Kriván as First Secretary. An *Antal Koch Commemorative Medal* was established. *András Rónai* received the first one. A proposal to create a *Section for General (Dynamic) Geology* was accepted. Its first meeting was held on 14 May. The first Chairman was *Tibor Szalai* and the Secretary *Áron Jámbor.* It began its own periodical "*General Geological Review*" (Általános Földtani Szemle).

However, the undisputed highlight of the year was the *Centenary Celebration of the Hungarian Geological Institute* (MÁFI). This was an unprecedented joint effort of the entire community of Hungarian geologists, directed and coordinated by J. Fülöp, former Director of the Institute and in that year already the successor to Gy. Kertai as the President of the Central Office of Geology. This summer series of events comprised the following:

- Colloquium on the Mediterranean Jurassic
- Conference on Bauxite Geology
- Colloquium on Eocene Stratigraphy
- Neogene Colloquium
- Day of the Geological Institutes (= Geological Surveys) of the World
- Business Meeting of the International Palaeontological Union (IPA)
- Business meeting of the IUGS Commission on Stratigraphy
- International session on Geological Correlation organized by UNESCO.

It was on this occasion that the proposal to organize an International Geological Correlation Program (*IGCP*) was discussed and accepted. A book on the l00-year history of the Hungarian Geological Institute was published in Hungarian and English. For all these events and the ensuing field trips combined, more than 700 foreign geologists came to Hungary from all over the world: ten times more than ten years earlier for the Mesozoic Conference.

Only a month later (11–19 September 1969) the *Carpathian-Balkan Geological Association held its IX. Congress,* for the first time in Budapest. Another month later, the II. Conference of the *International Commission on the Study of Bauxite, Alumina and Aluminum* (*ICSOBA*) was also held in Hungary. It was the right year to create a *Working Committee on the History of Geology.* On 27 November, a *Conference on Paleoclimatology* was co-organized with the Hungarian Meteorological Society.

1970 began with the establishment of a *Section for Mathematical Geology* on 30 January. At a meeting of the Section for Mineralogy and Geochemistry on 11 March, the results of the laboratory study of the lunar samples taken by Apollo-11 were reviewed. The annual General Assembly was held on 20 March. President E. Nemecz outlined the "new role" of geologists in the fields of engineering, construction of highways, space, marine and environmental research, as well as for agriculture. A special *Committee for Youth* was set up; *Andrea Mindszenty* became its secretary. The total of registered members was 1096.

The Itinerant Meeting was dedicated to the 50th anniversary of the death of L. Lóczy Sr. and the 20th anniversary of organized bauxite exploration at

Balatonfüred and Balatonalmádi, on 4–6 May. On 15 June, a *Section for the History of Geology* was established, with *Leontin Fejér* as its Secretary. A second Itinerant Meeting (!) was held in NE Hungary (Rudabánya Mountains, Tokaj Hills, Bükk Mountains) on 21–23 September.

E. Vadász passed away on 30 October 1970. This sorrowful event marked the end of a more than 30-year period which was very much influenced by his personality. It was in this same year that S. Papp also died (set free in 1955, he retired in 1962).

On 5 March 1971, a workshop was held on the issues of "brain drain", organized jointly by the Committee for Youth of the Society, the Central Office of Geology, and GEOMINCO Foreign Trade Company. At the General Assembly of 24 March 1971, in his presidential talk E. Nemecz made the following points:

1. There is increasing demand for different branches of applied geology (engineering geology, marine geology, astrogeology, environmental geology, agrogeology etc.).
2. A global approach based on subcrustal energy flows.
3. Accelerated progress in interdisciplinary research.

At a meeting held on 26 March 1971 Gábor Pantó presented a lecture entitled "Thoughts about geosynclinal and global tectonics". This was *the first presentation of plate tectonics in Hungary.* The 1971 Itinerant Meeting (24–26 June) went to NE Hungary (Debrecen, Kisköre), concentrating on geological problems of subsurface water, oil and gas.

By the end of 1971, the Hungarian Geological Society had reached a remarkable development. It had more than 1,000 members, eight (8) thematic sections, four (4) regional branches, as well as a number of committees and several periodic publications beside the main quarterly journal, the "Földtani Közlöny". In fact it had attained the present-day level in terms of membership and structure.

The events of the period 1972–1997 will be dealt with in Part II in the next issue of this journal.

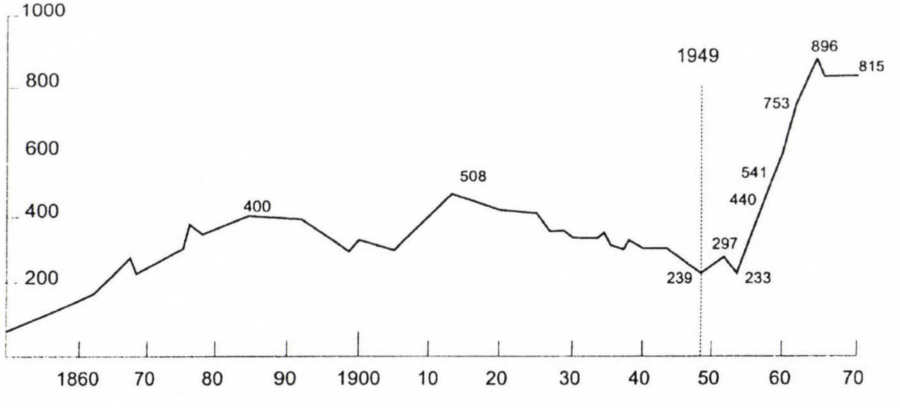


Fig. 1
Dynamics of the membership of the Hungarian Geological Society (1850–1971) (regular individual members only)

Annex I
Presidents and First Secretaries of the Hungarian Geological Society, 1850–1971

Year of election	President	First Secretary
1850	Kubinyi, Á.	Kováts, Gy.
1963		Szabó, J.
1866	Kubinyi, F.	Hantken, M.
1871	Reitz, F.	Bernáth, J.
1874		Sajóhelyi, I.
1877		Inkey, B.
1883	Szabó, J.	Pethő, Gy.
1886		Staub, M.
1899		Lóczy, L.
1901	Telegdi Roth, L.	
1904	Koch, A.	
1907		Lőrenthey, I.
1910	Schafarzik, F.	Papp, K.
1916	Szontagh, T.	
1920	Pálfy, M.	László, G.
1923	Mauritz, B.	Vendl, M.
1924		Zeller, T., Reichert, R.
1932	Vendl, A.	
1934		Papp, F.
1940	Papp, K.	Horusitzky, F.
1941	Papp, S.	Tasnádi-Kubacska, A.
1945	Vitális, I.	Majzon, L.
1947	Papp, S. (forced to resign in 1948) acting: Vadász, E.	Sümeghy, J.
1949	Vadász, E.	Szurovy, G.
1950		Kertai, Gy.
1951	Szádeczky-Kardoss, E.	
1952	Vitális, S.	Jantsky, B.
1954	Vadász, E.	Fülöp J., Pálfalvy I.
1958	Horusitzky, F. (resigned Oct. 1958) acting: Sztrókay, K.	Boda J., Dank V., Kriván P., Morvai G.
1960	Kertai, Gy.	Morvai, G.
1963		Kriván, P.
1966	Nemecz, E.	

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Foreign honorary members of the Hungarian Geological Society

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Every scientific institution honors its ordinary and other members for their outstanding performance and effective work displayed in favor of the institute. One of the most frequent forms of honor is the granting of the laudatory title "honorary member". The Hungarian Geological Society has also been respectful of this tradition.

Already in 1850, following its foundation the Society elected five Austrians persons who took effective part in its establishing or were political leaders in scientific life, as honorary members. Among the former were the first Director of the Geological Institute of Vienna, geologist-mineralogist *W. Haidinger*, and General Superintendent of the Museum of Natural Sciences of Graz (the Johanneum) *F. Thinnfeld*. The latter included the Minister of Education *L. Thun* as well as Ministerial Counselors *K. Geringer* and *J. Hauer*. All of these used their influence in the government in favor of the Society (it should be remembered how badly this was needed, since only one year previously Hungary had taken part in the revolution – which shook Europe in 1848–1849 – against Austrian domination!).

Thereafter the election of further honorary members took place only after the establishment of regular life of the Society and of regular geologic activity (mapping, sampling, publishing). In 1867 *F. Hauer* and *M. Hoernes* of Vienna, in 1869 *K. Peters* of Graz, in 1872 *O. Heer* of Zurich and *D. Stúr* of Vienna became honorary members. F. Hauer, K. Peters and D. Stúr were ranked among the best geologic mappers and stratigraphers of their age. All of them were also active in Hungary and their work is preserved in monographs and several basic studies. K. Peters was also a professor at the University of Budapest between 1855 and 1861. Many Hungarian stratigraphers and paleontologists relied on the outstanding paleobotanical knowledge of O. Heer and the prominent Tertiary malakological experience of M. Hoernes.

The 70s and 80s of the last century were characterized by the expanded development of Hungarian geology. Geologic work became a state-controlled task. In order to promote it, the Hungarian Royal Geological Institute was founded. Regular and systematic geologic mapping and exploration of mineral raw materials were begun; a geologic scientific journal, an annal, reports and

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maps were published. A Geological and then a Palaeontological Department were established at the University of Budapest (the latter was one of the first in Europe with the internationally well-known Miksa Hantken as its first professor). Hungarian geologists began to play a significant role in international relationships, such as the International Geological Congresses; they undertook study tours and visited several countries of Western Europe and even the USA. Relationships and joint studies were formed, as a result of which several foreign authorities became contributed honorary members of the Society.

During this period, numerous famous specialists were elected as honorary members. In 1876 the mineralogist and stratigrapher *B. Cotta* of Freiberg (who visited Hungary), and in 1880 the co-author of the work "Geologie Siebenbergens" (Geology of Transylvania), *G. Stache* of Vienna, were so honored. In 1883 the diligent researcher of the Hungarian fossiliferous flora, *C. Ettingshausen* of Graz, was selected; he elaborated the materials of so many Hungarian localities that the biography of Serjeant, published in 1980, indicates him as "Hungarian". Also chosen in that year were the great geologist, geographer and world-famous explorer (at that time professor of the University of Leipzig) *F. Richthofen*, as well as *K. Zittel*, the author of the first great classifying work of invertebrate and vertebrate fossils, the four-volumes "Handbuch" (these were the years of the first summarizing of the geological knowledge which had only accumulated until then).

Still in this decade, in 1886 11 excellent geologists from six countries increased the roll of honorary members. This time, Vienna was represented only by one geologist, the great tectonist and geomorphologist *E. Suess*. He was the author of the work (ahead of its time) "Das Antlitz der Erde" (The Face of the Earth) which influenced tectonic and geomorphologic approach and research for decades. The two Italians, *G. Capellini* of Bologna and *G. Meneghini* of Pisa, were awarded honorary membership on the basis of successful Italian–Hungarian cooperation (mainly in the field of geologic mapping) begun at the International Geological Congress at Bologna in 1881. The excellent mineralogist and mining engineer *A. Daubrée* (Paris) as well as the stratigrapher who maintained close relations with Hantken, *E. Hébert* (Paris), represent French–Hungarian cooperation on the roll of the honorary members. *E. H. Beyrich* of Berlin and *H. Dechen* of Bonn show the intense geological relationships with the German earth scientists. The activity of E. H. Beyrich helped understand the Hungarian Mesozoic, and especially the Tertiary stratigraphy, while that of H. Dechen provided samples for the exploration of mineral raw materials (today H. Dechen would be called an economic geologist).

Important scientific relationships are revealed by the election of two English geologists as honorary members. *W. T. Blanford* of London, renowned mainly for his research work in India and Southeast Asia, earned the honorary membership of the Society primarily through his vertebrate palaeontological activity, while *J. Prestwich* (also of London) through his Eocene stratigraphic correlation researches. Due to his stratigraphic activity J. Prestwich had the honor of becoming the eponym of a Nummulites species. The species named after him can actually be found among the Hungarian Nummulites fauna.

The Hungarian Geological Survey established contact with two outstanding geologists of the USA in the 80s (J. D. Dana and J. Hall) when József Szabó made a longer study tour of North America. *J. D. Dana* was professor of mineralogy of Yale University (New Haven). He acquired global renown by his systematizing and descriptive activity as well as by his "The System of Mineralogy". His book was referred to as the "Bible" of mineralogy. *J. Hall* (New York) was the first Chairman of the North American sister organization, the "Geological Society of America". In all probability he earned this position with his vast eight-volume work, the "Paleontology of New York". It is a special honor for the Hungarian Geological Society to have both of them among the honorary members.

Thereafter came a long period with no election of honorary members. History and events of Hungarian geology indicate that this was the time of the extensive phase of the development of Hungarian geology. Large-scale research and mapping activities were begun, the number of geologists increased according to the great demands, and regional organizations came into being (e.g. in Selmecbánya [Banská Štiavnica]). All of this furthered the internal buildup of the Society and meant such a burden that only slight attention could be paid to cultivating international relationships and electing honorary members. Furthermore the "great generation" (J. Szabó, M. Hantken, J. Böckh) who had the most extensive international connections died out and the next "great generation" (J. Krenner, L. Lóczy, A. Koch) needed time to renew old relationships and establish new ones. This time arrived in 1913, when the renowned tectonist *A. Heim* of Zurich and the excellent mineralogist and founder of the journal Zeitschrift für Krsytallographie, *P. H. Groth* of Munich, became honorary members. Another election fell in this period in 1916, when during the World War I (it seems that "inter arma non silent musae") the Czech *G. Tschermak* and the German *F. Beyschlag* earned this honor. G. Tschermak, working in Vienna, made his mark as a researcher mainly by the investigation of volcanic rocks. He also founded a scientific journal, Tschermak's Mineralogische und Petrographische Mittelungen, which is one of the highest-standard journals of European mineralogists. F. Beyschlag of Berlin made respectable achievements in the field of stratigraphy and tectonics.

Then another significant pause followed. As a result of World War I and the Paris Peace Treaty the political map and economic life of Central and Eastern Europe were rearranged. As a result Hungary lost about 80% of its mineral raw material assets. The decrease in Hungarian economic potential and the great economic world crisis strongly made their effects felt in geologic research as well. By the time the country would have emerged from the crisis World War II followed, which did not favor this type of science and research either. All these factors explain the lack of any election of honorary members.

During this long period without formally electing honorary members only one new name was put on the tablet of members: in 1942 the chairman of the Finnish Geological Society, *A. Laitakari*, earned the laudatory title. Besides the appreciation of his scientific activity, his election was also a gesture expressing the cooperation and friendship of the related Finnish and Hungarian people in the disastrous years of war.

During the one and a half decades following World War II radical social and economic changes took place in the world. In Central and Eastern Europe a system of planned economy appeared with an intense program of industrialization. Due to a support for the exploration of mineral raw materials in a degree not yet experienced geology became one of the leading branches of research. Members of the Hungarian Geological Society, occupied with their enormous and important tasks, only elected new honorary members in the 60's. However, as if to compensate for the long interruption, in 1960 seven geologists from 3 countries, and in 1964 ten geologists from six countries, were ranked among the honorary members.

In 1960, four Soviet geologists were considered deserving of this title by the Society. Among them *N. S. Satszkiy* (Moscow) and *V. I. Szlavin* (Moscow) dealt mainly with tectonics. N. S. Satszkiy prepared the first tectonic map of the Soviet Union. V. I. Szlavin worked on the problem of the "median mass", which also touched on the "Tisia question". Both of them visited Hungary. V. I. Slavin was involved in the elaboration of the Paleozoic fauna of the Uppony Mts. (Northern Hungary). *O. S. Vialov* (Lvov) came to Hungary several times. He imparted his rich knowledge of stratigraphy, paleontology and paleoecology to his Hungarian colleagues during several lectures given to the Society. *V. S. Saboliev* (Lvov) became world-famous through the geologic research of the diamond beds of Yakutia. Later, during his investigations in the Ukrainian Carpathians, he got in touch with Hungarian geologists, and a successful cooperation was created between them. In that year two noted Czech/Slovakian geologists also became honorary members. *R. Kettner* of Prague excelled in mapping and elaborating the Paleozoic formations of the Barrandian, after which he worked in the Carpathians and Low Tatra; the geological elaboration of the Domica Branch of the Aggtelek cave system is linked to his name. *D. Andrusov* of Bratislava was the greatest expert of the Mesozoic of the Klippen belt of the NW Carpathians. Both of them maintained extensive and fruitful professional relations with Hungarian scientific circles. The excellent paleontologist (dealing mainly with Neogene pectinids) and stratigrapher, *J. Roger* (Paris), got in touch with Hungarian colleagues in the course of editorial work for the Lexique Stratigraphique International. During his visit to Hungary in 1958 he gave several lectures which made a great impression. He supported the Paleontological Collection of the Hungarian Museum of Natural Sciences, which suffered a loss during the fire of 1956, with comparative materials and extensive specialized literature.

In 1964 four Austrian geologists joined the ranks of the honorary members. Their common feature is that all of them are famous in their special fields, honorary doctors of several universities and honorary members of scientific societies. All of them, although they worked in Vienna, visited Hungary and maintained excellent connections with their Hungarian colleagues. *E. Clar* was a prominent expert of economic geology and mineralization. *H. Küpper* was a petroleum geologist, stratigrapher and tectonist who emphasized the structural relations between the Alps and the Transdanubian Range. *O. Kühn* was a paleontologist, stratigrapher, paleoecologist and paleobiologist and a well-known researcher of hydrozoans and rudists. *F. Machatski* was a mineralogist studying feldspars and classifying silicate minerals. In the person of *D. Nalivkin* of Moscow, an international authority of geologic map editing was honored by the Society. In the professional spectrum covered by the honorary members a new research area is represented by *E. Tongiorgi* (Pisa): Quaternary geology. This was the first time that names of Polish, Yugoslav and Norwegian geologists made the list of honorary members. *R. Kozlowski* (Warsaw) threw light on the taxonomic status of the graptolitinae and *M. Ksiazkiewicz* (Krakow) was a researcher of the genesis of the Carpathian flysch formations as well as of the evolution of flysch basins. *V. K. Petkovic* (Belgrade) was the principal promoter of the business of the Carpatho-Balkan Geological Association, and *I. T. Rosenquist* (Oslo) first applied isotope technology in investigating clay minerals. In 1969 a Hungarian living in the United States, the paleobotanist *J. Jablonszky*, obtained this delayed honor.

In the 70s a significant number of geologists of significant merit was again added to the list of honorary members of the Society, and in 1971 the list of names was enriched by eight members. Among them three were Soviets: the tectonist *A. A. Bogdanov*, the petrographer-economic geologist *F. V. Chukrov* and the paleontologist-paleoecologist *R. F. Hecker*. The principal work of *R. F. Hecker* (who made a lecture tour in Hungary), the "Introduction to palaeoecology", was also published by a western publishing firm in English. Germans were represented by the stratigrapher *H. R. Gartner* and the "modern-day Dana" (*H. Strunz*), who elaborated a new, modern mineral classification system. A French, a Czech, and a Yugoslav geologist completed the list of those elected in 1971. The excellent Jurassic ammonite specialist *M. Collignon* (Paris), the mineralogist-petrographer *V. Zoubek* (Prague), as well as the Neogene paleontologist-stratigrapher *P. Stefanovic* (Belgrade), who maintained a regular work connection with Hungarian Pannonian specialists (he came frequently to Hungary).

At the elections of 1972 the Austrian *H. Wieseneder* (Vienna), dealing mainly with petrology and rock mechanics, was chosen. In 1975 the excellent oil geologist (also Austrian) *R. Janoschek* of Vienna (he maintained good work connections with several Hungarian geologists and promoted the Society's efforts) and the Slovakian *J. Senes* (Bratislava) were admitted by the Hungarian Geological Society. A great Neogene stratigrapher-paleontologist, J. Senes maintained extensive and intensive work relations with the Hungarian geologists. This is confirmed by several of his papers written together with Hungarian colleagues.

The Society next conferred the title of the honorary membership in 1978. All of the newly elected earth scientists had been in Hungary and were in continuous professional contact with their Hungarian colleagues. The author of "Geology of Bulgaria", the tectonist *E. Boncev* (Sofia), spared no effort to cultivate Bulgarian–Hungarian geologic relationships. He was the first Bulgarian honorary member of the Society. The election of *M. Mahel* (Bratislava) and *J. Vachtl* (Prague) marks the vitality of the Czechoslovakian–Hungarian geologic connections. Hungarian geology benefited from the stratigraphic and even more from the tectonic work, of M. Mahel and from the mineralogic–petrographic activity of *J. Vachtl*. The honorary membership of the Pole *K. Smulikowsky* indicates the dynamically developing Polish–Hungarian geologic cooperation. The mineralogist and economic geologist *A. Watznauer* (Freiberg) was closely associated with several Hungarian geologists. During joint work participated in special investigation of Hungarian rock samples. The prominent paleontologist and stratigrapher *H. Zapfe* (Vienna) dealt mainly with the elaboration of Triassic formations. He maintained excellent and close contacts with the Hungarian Triassic specialists and effectively promoted institutional Austrian–Hungarian geologic relations. With his death in 1996 Hungarian geologists lost a true friend.

Honorary members have been elected more recently. In 1991 the Austrian *T. Gattinger* (Vienna), the Pole *A. Slaczka* (Krakow) and *P. Teleki* living in Reston (USA) came into possession of this laudatory title. T. Gattinger made his mark as a technical geologist and stratigrapher. A. Slaczka obtained global recognition for his results in investigating the sedimentation of Mesozoic formations. P. Teleki became known by his hydrogeologic and geologic managing activity. All of them visited Hungary frequently and during these times forged close contacts and cooperations with their Hungarian colleagues.

In 1994, three earth scientists were honored with the title. The Hungarian *Z. Cserna* lives in Mexico. He was a field geologist, then university teacher, and finally a professor at the University of Mexico City. In the last years he has been the mentor of Hungarian geologists working in Mexico. With his deep knowledge, wise advice, and wide-ranging contacts he always backs his fellow-countrymen. Another Hungarian, *G. Földváry*, lives at the other end of the world, in Sydney. On the basis of numerous field trips in Hungary and by means of 20 years' persistent literary activity, he wrote his work "Geology of the Carpathian Region" (including a detailed geology of Hungary) which was honored by the laudatory title. The French academician *J. Dercourt* deservedly became an honorary member of the Society for his great paleogeographical and paleoecological work based on plate tectonics, and for the establishment of institutional French–Hungarian geologic connections (exchange programs, scholarships, etc.). With his profound knowledge and professional influence he has always supported Hungarian geology.

In 1997 the Hungarian *L. Trunkó*, living in Karlsruhe, was deservedly elected as an honorary member of the Society. With his work "Geology of Hungary", published in Stuttgart, he rendered significant service to Hungarian geology. This work offers a possibility to make foreign scientists acquainted in detail with the geology of Hungary.

On the basis of the present summary it can be seen that the Hungarian geologists, when selecting foreign honorary members of the Hungarian Geological Society, always based their decisions on the scientific value, work results and the impact on Hungarian science and political interests. They were always in close touch with the most prominent geologists of their age and tried to honor the latest achievements in geology.

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Alpine prograde and retrograde metamorphisms in an overthrusted part of the basement, Great Plain, Pannonian Basin, Eastern Hungary

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As proved by a hydrocarbon exploratory borehole, a pre-Alpine–Alpine polymetamorphic complex was thrusted over a low-temperature prograde metamorphic Mesozoic sequence in a Neogene depression of the Pannonian Basin, eastern Hungary. The detectable first, prograde, medium thermal gradient amphibolite facies event was followed by a retrograde metamorphism connected to a mylonite formation in the overthrusted complex. The polymetamorphic history of this gneiss–mica schist–amphibolite complex has also been recorded by the changes in chemistry of amphibole, white K-mica, and plagioclase. The footwall Mesozoic parautochthon is built up by an upper sub-unit (dolomitic–fine clastic), a middle (calcareous–fine clastic) and a lower one (calcareous–basic to intermediate volcanoclastic). On the basis of phyllosilicate crystallinity and vitrinite reflectance data, the metamorphic grade of the upper part corresponds to the low-temperature part of the anchizone and that of the middle and lower parts to the epizone, having been very close to the boundary between the epizone and anchizone. The maturity parameters of the basin-filling Neogene formations refer only to diagenetic conditions, indicating non-equilibrium conditions. The Neogene (<5 Ma) burial has had no detectable effects on the metamorphic conditions of the basement units, which had already occupied their present tectonic position during their post-metamorphic uplift (cooling) in the Upper Cretaceous. The retrograde metamorphism in the polymetamorphic nappe and the low-temperature prograde metamorphism in the Mesozoic parautochthon proved to be of Eo-Alpine (Cretaceous) age, and was older than the overthrusting. The tectonic (shearing) strain connected to the overthrusting influenced the structural state of phyllosilicates in a rather irregular way, both in the hanging wall and the footwall units.

Key words: Pannonian Basin, Great Plain, Hungary, metamorphic basement, Alpine metamorphism, illite crystallinity, chlorite crystallinity, geothermometry, geobarometry, polymetamorphism, low-grade metamorphism, epizone, anchizone

Introduction

A hydrocarbon exploratory borehole cutting across a peculiar section of basement indicates inverse metamorphic zonation and post-metamorphic thrusting ("transported metamorphism") in the eastern part of the Pannonian Basin (Fig. 1a). A polymetamorphic complex was thrusted over a weakly metamorphosed or unmetamorphosed Mesozoic sequence called "Sáránd

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Parautochthon", as proved by the discontinuous core materials from the Sáránd-I borehole at the village of Sáránd, near the town of Debrecen, Eastern Hungary. Similar weakly metamorphosed rocks were also described from the close vicinity of Sáránd, namely from wells Bam-1 and 2 at the village of Bagamér (Fig. 1b).

The aims of the present study are first to reconstruct the polymetamorphic history of the hanging wall (overthrusted) complex and secondly to determine the diagenetic-metamorphic grade (zone) of the Sáránd Parautochthon, considering also the present burial conditions and the diagenesis of the Neogene basin-filling sedimentary rocks. For these purposes the present paper completes already available mesoscopic and microscopic observations with new results of microstructural, mineral paragenetic, X-ray diffractometric illite and chlorite crystallinity, crystallite size and lattice distortion, white mica geobarometric, K-Ar isotope geochronologic, electron microprobe mineral chemical and vitrinite reflectance data, in order to perform a correct characterization of the metamorphic processes. The following questions were to be answered from the new petrographic data set:

(i) Did the Neogene burial cause any mineralogical or petrographic changes in the basement complexes?

(ii) Did the overthrusting (and the overthrusted complex) play an effective role in the incipient metamorphism of the footwall complex?

(iii) Are there any relations between the prograde metamorphism of the footwall and the retrograde metamorphism of the hanging wall?

Geologic framework and previously available data

The study area is located in the eastern part of the Tisia Terrane (Kovács et al. in press) or Tisia or Tisza microplate called also megaunit (Fig. 1a). The Tisia Terrane represents a fragment of the northern (stable European) border of Tethys detached from the continent and moved to its present position by horizontal microplate displacements mainly during the meso-Alpine tectonophases (Géczy 1973; Kovács 1982; Kázmér and Kovács 1985; Kovács et al. in press). The investigated core material presents geologic information on the northwestern flank (borehole Sáránd-I) and the northern and northeastern surroundings (boreholes Bagamér Bam-1 and 2) of the Derecske Depression, one of the major depressions in the pre-Tertiary basement of the Pannonian Basin (Fig. 1b). The metamorphic basement of the Derecske Depression belongs to the Bihor Autochthon (more precisely, parautochthon; see Szederkényi 1984; Balázs et al. 1986 and Szederkényi et al. 1991). Its metamorphic evolution is characterized by pre-Variscan (?) or early Variscan (?) medium thermal gradient (Barrovian) amphibolite facies metamorphism overprinted locally by a Variscan high thermal gradient, mainly amphibolite facies event and late Variscan and/or Alpine retrogression, mylonite and cataclasite formation (Szederkényi 1984; Árkai et al. 1985; Szili-Gyémánt 1986; Árkai 1987; Szederkényi et al. 1991).

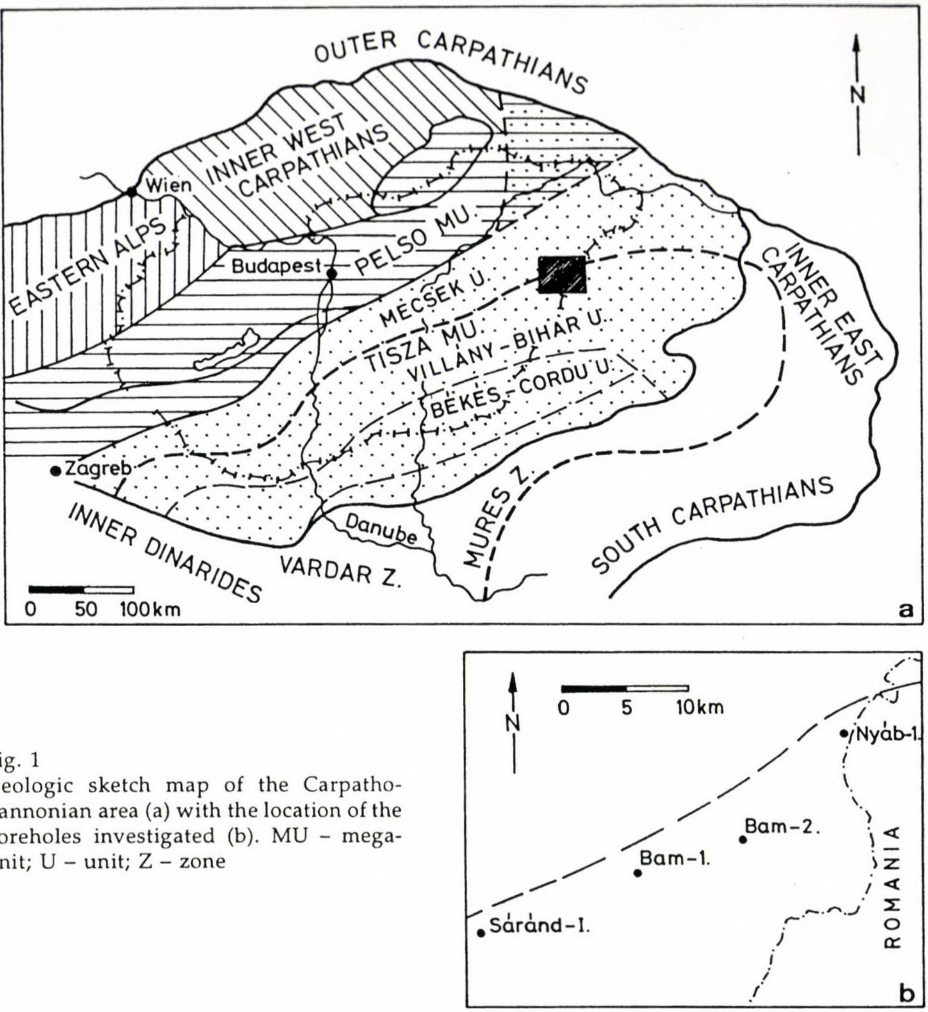
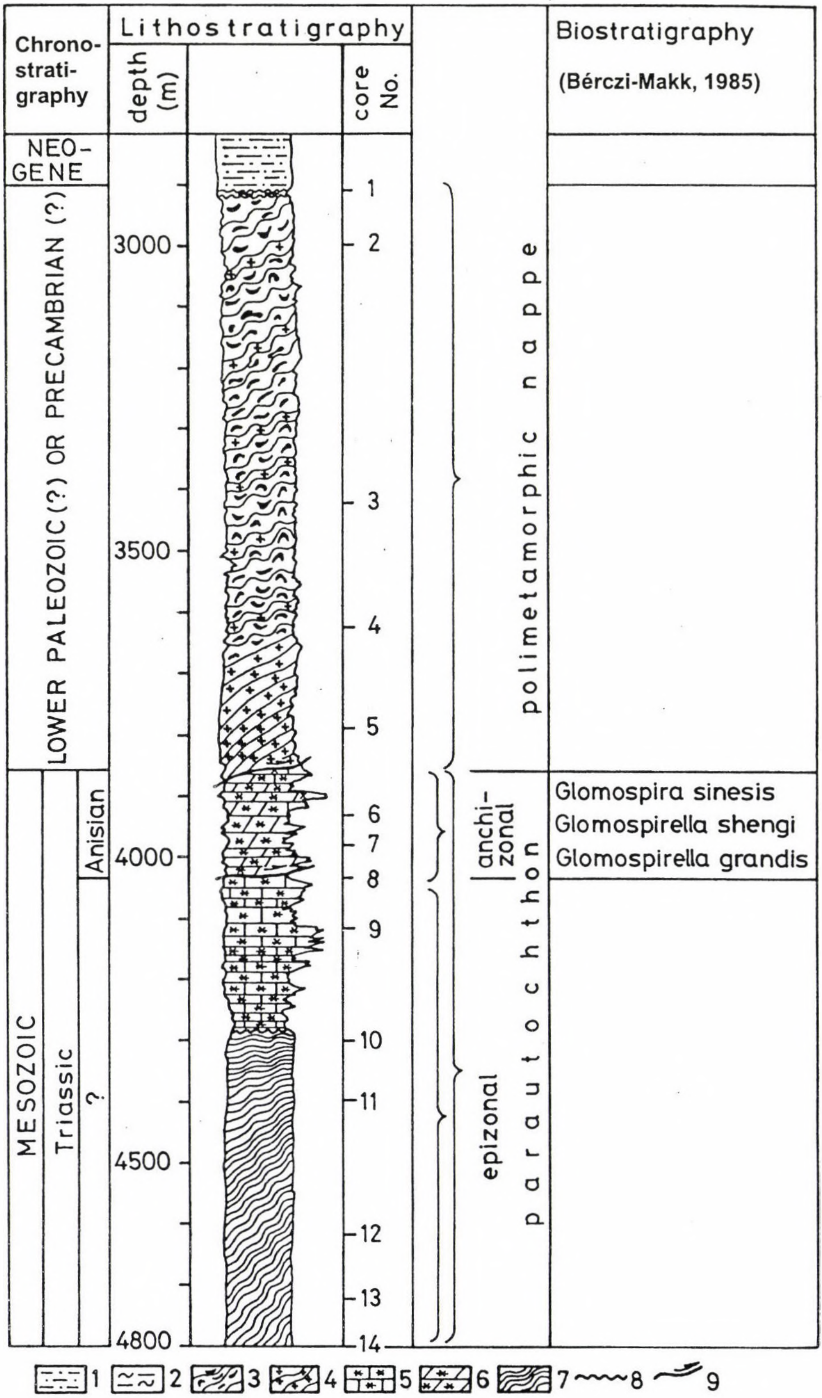


Fig. 1
Geologic sketch map of the Carpatho-Pannonian area (a) with the location of the boreholes investigated (b). MU – mega-unit; U – unit; Z – zone

Pap (1990) summarized the geologic and petrographic data obtained from the core material of borehole Sáránd-I by Nusszer (1984) and Jámbor and Ravasz (1984). Beneath thick Pannonian and thin Miocene basin-filling sedimentary sequences, this borehole intersected a 924 m-thick polymetamorphic complex that had been overthrusted onto a presumably younger, lower-grade parautochthonous unit (Fig. 2). According to the authors mentioned above the polymetamorphic complex consists of an upper mica schist–gneiss part and a lower, amphibolitic one. The amphibolite facies metamorphism of the pelitic and psammitic sedimentary and basic magmatic rocks was followed by retrograde metamorphism of prehnite-pumpellyite and subordinately greenschist facies. The metamorphic evolution ended with mylonitization and


Chrono-stratigraphy
Lithostratigraphy
depth (m)
core No.
Biostratigraphy
(Bérczi-Makk, 1985)
NEOGENE
LOWER PALEOZOIC (?) OR PRECAMBRIAN (?)
MESOZOIC
Triassic
Anisian
?
3000
3500
4000
4500
4800
1
2
3
4
5
6
7
8
9
10
11
12
13
14
polimetamorphic nappe
anchizonal
epizonal
parautochthon
Glomospira sinesis
Glomospirella shengi
Glomospirella grandis
1 2 3 4 5 6 7 8 9

formation of cataclasites, presumably at the end of the Cretaceous. In the upper part of the underlying parautochthon, carbonatic metasedimentary rocks predominate over slates. The Middle Triassic (Anisian) age of this part of the parautochthon has been proved by micropaleontological data (Bérczi-Makk 1985). Jámbor and Ravasz (1984) described greenschist facies albite-chlorite schists of volcanoclastic origin in the lower part of the unit. According to Nusszer (1984) pelitic-marly-carbonatic and subordinately psammitic sedimentary material alternates with metavolcanites and volcanoclastic rocks, forming various slates and schists. The stratigraphic age of this sequence is unknown. Supposing continuous stratigraphic sequence of layers in normal tectonic position a Lower Triassic age was assumed by Bérczi-Makk (1985). Considering the lithofacies conditions of the Middle Triassic in the various tectonic units of Hungary a Middle Triassic (Ladinian or Carnian) age is also possible, thus implying an inverse stratigraphic sequence. In contrast, Jámbor and Ravasz (1984) argued for a possible lower Paleozoic age using lithological analogies. According to the latter authors the upper, carbonatic part of the parautochthon suffered prehnite-pumpellyite facies (anchizonal) metamorphism, although neither indicative mineral assemblages were found nor phyllosilicate crystallinity measurements performed on the samples of the unit.

The overthrusted polymetamorphic complex

Rock types, pre-metamorphic lithologies

On the basis of 18 rock samples collected from 5 cores of the Sáránd-I borehole, the following rock types were distinguished (from the top to the bottom): mylonite [originated from biotite-muscovite schist (±garnet)]: core No. 1; mylonitized quartzite: core No. 2; mylonitized biotite-muscovite gneiss and mylonite: core No. 2; biotite-muscovite-bearing quartzite with cataclastic deformation (core No. 2); mylonite of garnet-bearing biotite-muscovite schist origin: core No. 2; mylonite [from biotite-muscovite gneiss (±garnet)]: core No. 3; mylonitized amphibolite with younger cataclastic deformation: core No. 4; mylonitized, amphibole-bearing muscovite-biotite gneiss with younger cataclastic deformation: core No. 5 and mylonite of amphibolite origin with cataclastic alteration: core No. 5.

The gradational transitions between gneisses, mica schists and amphibolites suggest that the polymetamorphic pile might have been formed from carbonate-poor or -free terrigeneous clastic (pelitic-psammitic) sediments intercalated with basic volcanoclastic material.

← Fig. 2
Geologic section of the pre-Tertiary part of borehole Sáránd-I. 1. clay marl with streaks of siltstone; 2. clay marl; 3. mylonite; 4. mylonitized amphibolite; 5. banded calcschist, marble; 6. dolomitic marble; 7. metatuffite; 8. unconformity; 9. overthrust

Conditions of metamorphic events

Table 1
Average chemical compositions of plagioclases from polymetamorphic rocks of the borehole Sáránd-I

Core/depth	3/3435.4 m	4/3656.7 m
generation	I	II
n	3	4
SiO_2	62.72	68.10
Al_2O_3	22.69	19.80
CaO	4.04	0.44
Na_2O	9.36	11.70
K_2O	0.12	0.13
Total	99.05	100.27
Number of cations per 8 oxygens		
Si	8.402	8.922
Al	3.582	3.057
Ca	0.580	0.061
Na	2.430	2.973
K	0.021	0.022
total	15.028	15.046
An	19.1	2.0
Ab	80.2	97.3
Or	0.7	0.7

n = number of measurements

The assemblages of the gneisses and mica schists consisting of muscovite, biotite, quartz, and plagioclase ± garnet represent the first observable prograde event indicating amphibolite facies conditions. The biotite was completely, the garnet partially chloritized, and the plagioclase (in average: $An_{19.1}Ab_{80.2}Or_{0.7}$, see Table 1) altered locally into fine-grained white mica ("sericite"). Rare white mica pseudomorphs may indicate totally destroyed staurolite and/or an Al_2SiO_5 mineral. The distribution of chemical elements in the almandine-rich garnet grains is rather homogeneous (Table 2). The muscovite remained intact or changed its composition by isomorphic substitutions. The oldest generation of white K-micas (large, frequently deformed flakes) are of muscovitic composition characterized by relatively low (< 6.3 per formula unit) Si, high Al^{iv}, Na and low Fm [$\Sigma(Mg+Fe_{total})$] contents (1st generation in Table 3 and Fig. 3). The mole fractions of muscovite (X_{Ms}) in the muscovite-paragonite solid solutions vary between 0.80 and 0.82. As no coexisting paragonite was found these values can only be applied to the rough estimation of the possible minimum temperature of metamorphism. Thus, using the corrected paragonite-muscovite solvus geothermometer of Guidotti et al. (1994), the minimum temperature of formation of muscovite varies between 470 and 515 °C, giving an average of 505 °C for 4 samples. Due to the alteration of minerals caused by the intense retrogression, the various thermobarometers available in the literature could not be applied for the precise determination of the physical conditions of the amphibolite facies event[1].

The tschermakitic hornblende composition of amphibole porphyroclasts, measured in the inner parts of the grains (1st generation in Table 4), indicates

1 Files containing the chemical compositions of metamorphic minerals can be obtained either in printed or in electronic forms from the first author upon request (E-mail address: arkai@sparc.core.hu). The instrumental parameters and measuring conditions are given in detail by Árkai et al. (1995b).

Table 2
Average chemical compositions of garnets from the polymetamorphic rocks of borehole Sáránd-I

Core/depth	2/3017.4 m		2/3018.6 m	3/3435.4 m	
n	7		3	15	
	x	s	x	x	s
SiO_2	38.05	0.35	36.88	37.62	0.30
TiO_2	0.03	0.01	0.03	0.01	0.02
Al_2O_3	21.19	0.22	21.53	21.37	0.23
FeO	33.44	0.56	35.14	33.50	0.61
MgO	2.43	0.13	2.89	3.61	0.47
MnO	3.29	0.26	1.70	2.61	0.86
CaO	2.21	0.28	1.91	1.87	0.44
Na_2O	0.01	0.01	0.01	0.01	0.02
K_2O	0.01	0.02	0.00	0.00	0.00
Total	100.66	0.68	100.10	100.60	0.43
Number of cations per 24 oxygens					
Si	6.073	0.050	5.933	5.994	0.020
Ti	0.004	0.002	0.003	0.001	0.002
Al	3.986	0.038	4.082	4.012	0.037
Fe^{2+}	4.463	0.064	4.727	4.463	0.061
Mg	0.579	0.032	0.693	0.856	0.107
Mn	0.444	0.035	0.231	0.353	0.118
Ca	0.377	0.048	0.330	0.319	0.077
Na	0.003	0.002	0.004	0.004	0.005
K	0.002	0.004	0.000	0.000	0.001
Total	15.931	0.032	16.003	16.002	0.009

x – average; s – standard deviation; n – number of measurements

amphibolite facies metamorphism of the metabasites. As the plagioclase of the amphibolites was albitized and altered into fine-grained white K-mica, only a very rough approximation can be made on the P–T conditions. Compared with the facies and isograd schemes of Winkler (1979) and Bucher and Frey (1994) the chemistry of amphiboles indicates temperatures >490–500 °C. ΣAl values of the amphiboles indicate ca. 8 kbar at 500 °C, 5 kbar at 550 °C and 4 kbar at 600 °C using the amphibole-plagioclase thermobarometer of Plyusnina (1982). (The uncertainties derive from the unknown composition of the first generation of plagioclase.) These estimates, however, imply medium thermal gradient amphibolite facies conditions, agreeing with the earlier results from the neighboring areas of the basement obtained by Árkai et al. (1985) and Árkai (1987).

The gneiss–mica schist–amphibolite complex was mylonitized to various degrees: pervasively mylonitized layers alternate with layers where signs of initial mylonitization are shown in thin shear zones, whereas the original (gneissous or schistose) structures have still been preserved. The mylonitization

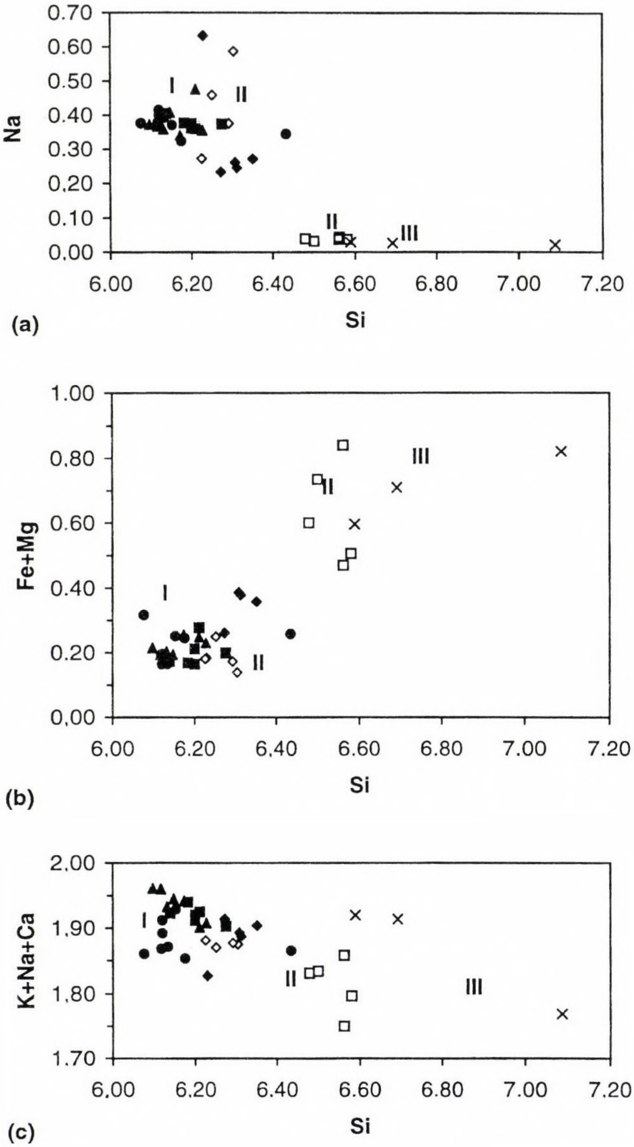


Fig. 3
Chemical compositions of white K-micas in the polymetamorphic complex encountered in borehole Sáránd-I. I – large porphyroclasts crystallized during the amphibolite facies event (filled symbols); II – small flakes of the mylonitic matrix (open symbols); III – small flakes from cataclastic fissure fillings. ◆ – mylonite from biotite-muscovite schist core No. 1, 2933.0 m; ■ – cataclastic, mylonitized, garnetiferous biotite-muscovite gneiss, core No. 2, 3017.4m; x = cataclastic filling from the previous sample; ● – mylonite from garnetiferous biotite-muscovite schist, core No. 2, 3018.6 m; ▲ – mylonite with younger cataclastic deformation (origin: garnetiferous biotite-muscovite gneiss)

was accompanied by retrograde recrystallization. In gneisses and mica schists Ca-containing plagioclase was partially altered into white K-mica (sericite) and albite; biotite and garnet completely or partly into chlorite. In addition to the crushed, more or less altered clasts of the amphibolite facies minerals the fine-grained mylonitic matrix also contains newly-formed chlorite, quartz, albite, calcite, ankerite and a fine-grained second generation of white K-micas.

Table 3
Average chemical compositions of white K micas from the polymetamorphic rocks of borehole Sáránd-I

Core/depth	1/2933.0 m			2/3017.4 m					2/3018.6 m		2/3435.4 m	
generation	I		II	I		II		III	I		I	
n	5		4	6		5		3	8		7	
	x	s	x	x	s	x	s	x	x	s	x	s
SiO_2	47.40	0.53	47.87	46.31	0.51	48.48	0.58	49.82	46.93	0.56	45.76	0.67
TiO_2	0.48	0.24	0.08	0.54	0.07	0.45	0.23	0.01	0.56	0.07	0.53	0.06
Al_2O_3	34.29	1.12	35.88	35.18	0.47	30.59	1.08	28.12	36.12	1.38	35.25	0.40
FeO	1.06	0.26	0.77	0.78	0.10	2.69	0.77	4.24	0.91	0.16	0.92	0.11
MgO	0.99	0.30	0.52	0.56	0.17	1.62	0.33	1.11	0.63	0.19	0.58	0.09
MnO	0.02	0.02	0.04	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.02	0.01
CaO	0.04	0.03	0.07	0.01	0.02	0.06	0.03	0.00	0.01	0.01	0.04	0.02
Na_2O	1.28	0.67	1.67	1.45	0.06	0.14	0.02	0.10	1.47	0.12	1.47	0.18
K_2O	9.15	1.17	8.63	9.03	0.11	10.28	0.20	10.59	9.02	0.23	9.01	0.32
Total	94.71	0.80	95.56	93.86	0.50	94.31	0.81	94.00	95.68	1.08	93.57	0.72
Number of cations per 22 oxygens												
Si	6.295	0.046	6.269	6.201	0.045	6.537	0.044	6.789	6.165	0.113	6.157	0.049
Al^{IV}	1.705	0.046	1.731	1.799	0.045	1.463	0.044	1.211	1.835	0.113	1.843	0.049
Al^{VI}	3.663	0.109	3.809	3.753	0.041	3.399	0.141	3.307	3.757	0.063	3.748	0.028
Ti	0.049	0.024	0.009	0.054	0.007	0.046	0.024	0.001	0.056	0.008	0.054	0.006
Fe^{2+}	0.118	0.028	0.085	0.087	0.010	0.303	0.088	0.483	0.100	0.018	0.103	0.012
Mg	0.196	0.060	0.102	0.112	0.034	0.327	0.068	0.225	0.124	0.036	0.117	0.018
Mn	0.003	0.002	0.005	0.000	0.000	0.000	0.000	0.001	0.001	0.001	0.002	0.002
Ca	0.005	0.004	0.009	0.002	0.002	0.008	0.004	0.000	0.002	0.002	0.005	0.002
Na	0.329	0.171	0.424	0.376	0.016	0.038	0.004	0.025	0.375	0.029	0.383	0.046
K	1.551	0.205	1.443	1.543	0.020	1.768	0.042	1.842	1.512	0.043	1.548	0.061
Total	13.914	0.027	13.886	13.927	0.023	13.889	0.075	13.884	13.927	0.045	13.960	0.028

x – average; s – standard deviation; n – number of measurements

Table 4
Average chemical compositons of amphiboles from the polymetamorphic complex (borehole Sáránd-I)

Core/depth		5/3656.3 m			5/3656.7 m			
generation		I	transitional I–II	II	I		transitional I–II	II
n		1	2	4	10		1	3
		x	x	x	x	s	x	x
SiO_2		44.87	48.77	50.07	44.52	1.09	46.28	47.48
TiO_2		0.48	0.39	0.18	0.52	0.15	0.52	0.40
Al_2O_3		11.80	6.84	4.80	11.71	1.03	9.77	7.59
FeO		15.47	14.71	15.76	15.60	1.48	14.31	15.31
MgO		11.90	13.49	13.61	11.59	0.98	12.48	12.80
MnO		0.27	0.22	0.21	0.27	0.03	0.24	0.28
CaO		12.74	12.43	12.56	12.06	0.32	12.34	12.43
Na_2O		1.34	0.68	0.43	1.36	0.14	1.15	0.87
K_2O		0.41	0.18	0.15	0.48	0.08	0.45	0.33
Total		99.28	97.70	97.75	98.11	0.80	97.55	98.48
Number of cations per 23 oxygens								
T	Si	6.478	7.066	7.272	6.483	0.123	6.763	7.016
T	Al^{IV}	1.522	0.934	0.728	1.517	0.123	1.137	0.984
M_1, M_2, M_3	Al^{VI}	0.486	0.236	0.094	0.495	0.120	0.446	0.310
M_1, M_2, M_3	Ti	0.052	0.042	0.020	0.057	0.017	0.057	0.043
M_1, M_2, M_3	Fe^{3+}	0.538	0.530	0.536	0.669	0.168	0.402	0.429
M_1, M_2, M_3	Fe^{2+}	1.330	1.256	1.379	1.232	0.225	1.347	1.423
M_1, M_2, M_3	Mg	2.561	2.909	2.945	2.514	0.193	2.718	2.761
M_1, M_2, M_3	Mn	0.033	0.027	0.026	0.033	0.004	0.030	0.034
M_4	Ca	1.972	1.930	1.957	1.883	0.058	1.933	1.927
M_4	Na	0.028	0.070	0.043	0.117	0.058	0.067	0.073
A	Na	0.347	0.121	0.079	0.279	0.071	0.259	0.171
A	K	0.076	0.034	0.027	0.090	0.026	0.084	0.060
Total		15.423	15.155	15.106	15.369	0.068	15.343	15.231
$Mg/(Mg+Fe^{2+})$		0.658	0.698	0.681	0.671		0.669	0.660

x – average; s – standard deviation; n – number of measurements

"Spiny-like" quartz-mica overgrowths are also present around quartz porphyroclasts.

The period of intense mylonitization was followed by cataclastic deformation, the intensity of which proved to be increasing downwards to the main shear zone separating the overthrusted part from the parautochthon. The fractures are filled with fine-grained material that is identical with the matrix of the mylonites but displays no preferred orientation. Thus, the matrix of the cataclasites might form merely by simple mechanical re-distribution of the mylonitic matrix caused by fracturing.

Both mineral chemical and structural data confirm the polygenetic nature of the matrix-forming, fine-grained white K-micas (Table 3 and Fig. 3). At 2933.0 m of core No. 1 the chemistry of this second generation is largely identical with that of the first generation, suggesting that the small flakes have formed by mechanical crushing of the large mica flakes. In contrast at 3017.4 m of core No. 2, the fine-grained mica that forms the mylonitic matrix differs from the first generation in its higher Si, $\Sigma(Mg+Fe_{total})$ and lower Al^{iv}, and Na contents (Fig. 3). These differences are also reflected in the distribution of the $d_{33\bar{1},060}$ spacing determined on disorientated whole rock powder mounts using the modified method of Sassi (1972) as described by Árkai et al. (1991). The histrogram of the $6x(d_{33\bar{1},060}) \approx b$ values (Fig. 4) displays two distinct maxima. The maximum of smaller intensity at 8.990–8.995 Å represents the large mica flakes crystallized during the prograde amphibolite facies event, for the *b* values converge to that of ideal muscovite at around 9 Å with increasing temperature, disregarding the pressure conditions (Guidotti and Sassi 1976). The main maximum between 9.025–9.030 Å corresponds to the second (and third) generations of white K-micas, formed during and after the mylonitization and cataclastic deformation respectively. The average *b* values calculated from the chemical data of the white K-micas using the regression equations of Guidotti et al. (1992) are in agreement with the maxima determined by XRD measurements: 1st generation: 8.997 Å (large flakes formed in amphibolite facies); 2nd generation: 9.027 Å (small flakes in the mylonitic matrix with preferred orientation) and 3rd generation: 9.029 Å (small flakes in the fissure-filling matrix of the cataclasites, without any preferred orientation).

The *b* values of the 2nd and 3rd populations may suggest an intermediate thermal gradient (pressure) metamorphic event, using the geobarometer of Guidotti and Sassi (1986), supposing that this method can also be applied for retrogressive systems. As no relevant data are available in the literature in this respect further evidence is needed to justify this conclusion.

Figure 5 displays the chemical differences between the prograde and retrograde amphiboles represented by porphyroclasts and fine-grained matrix respectively. Note the quasi-continuous changes from tschermakitic hornblende through hornblende to actinolitic hornblende that indicate the non-equilibrium character of the retrograde event.

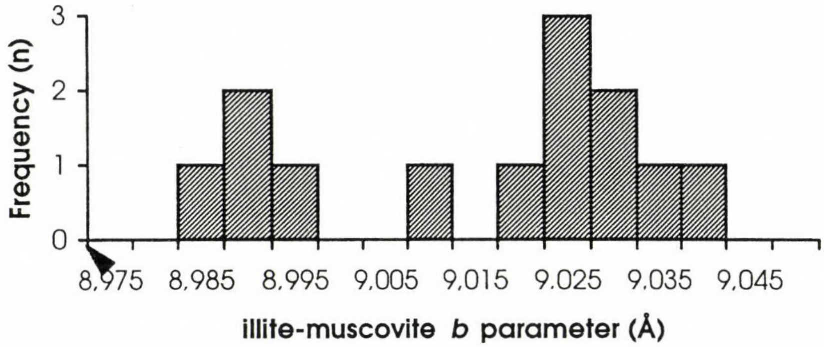


Fig. 4
Distribution of the *b* (=6xd$_{33\bar{1},060}$) values of the white K-micas from the polymetamorphic complex

Applying Plyusnina's (1982) thermobarometer to the second generations of amphibole (actinolitic hornblende and magnesio-hornblende with ΣAl=0.822–1.294 per formula unit) and of plagioclase (An=0.9–3.2%) determined in the strongly mylonitized amphibolites, rough estimates of $T \leq 450^{o}C$ and $P \approx 3–6$ kbar were obtained. This pressure range confirms the medium pressure character of the mylonitic recrystallization deduced from the chemistry and *b* value of the matrix forming white K-mica. As the amphibole was chloritized during the further steps of retrogression the second generation of amphibole and plagioclase might represent the higher-grade (initial) stage of this process.

The temperature of the further steps of the retrograde event can be estimated from the chemical composition of chlorites (Table 5). As with the white K-micas there are no differences between the chlorites of the mylonitic matrix displaying preferred orientation and chlorites connected to cataclasis. Using the classification of Foster (1962) the chlorites are mostly brunsvigite, rarely ripidolite. The chlorite-Al^{iv} geothermometer of Carthelineau (1988), elaborated originally for meta-magmatic rocks of intermediate composition, yields 320 ^{o}C for the mylonitized amphibolite of core No. 5. The somewhat higher temperatures of 331–351 ^{o}C calculated for the mica schist and gneiss samples of cores Nos 2 and 3 may be sufficiently explained by the disturbing effect of differing bulk rock chemistries (primarily higher Al contents) as was demonstrated by Árkai and Sadek Ghabrial (1997).

Microstructural features show that the <2 µm grain-size fractions prepared from the polymetamorphic samples are built up predominantly by the products formed during the retrograde metamorphism. To characterize the conditions of mylonitic recrystallization XRD phyllosilicate crystallinity measurements were also carried out. The sample preparation technique, the measuring conditions and the calibration of diagenetic and metamorphic illite crystallinity

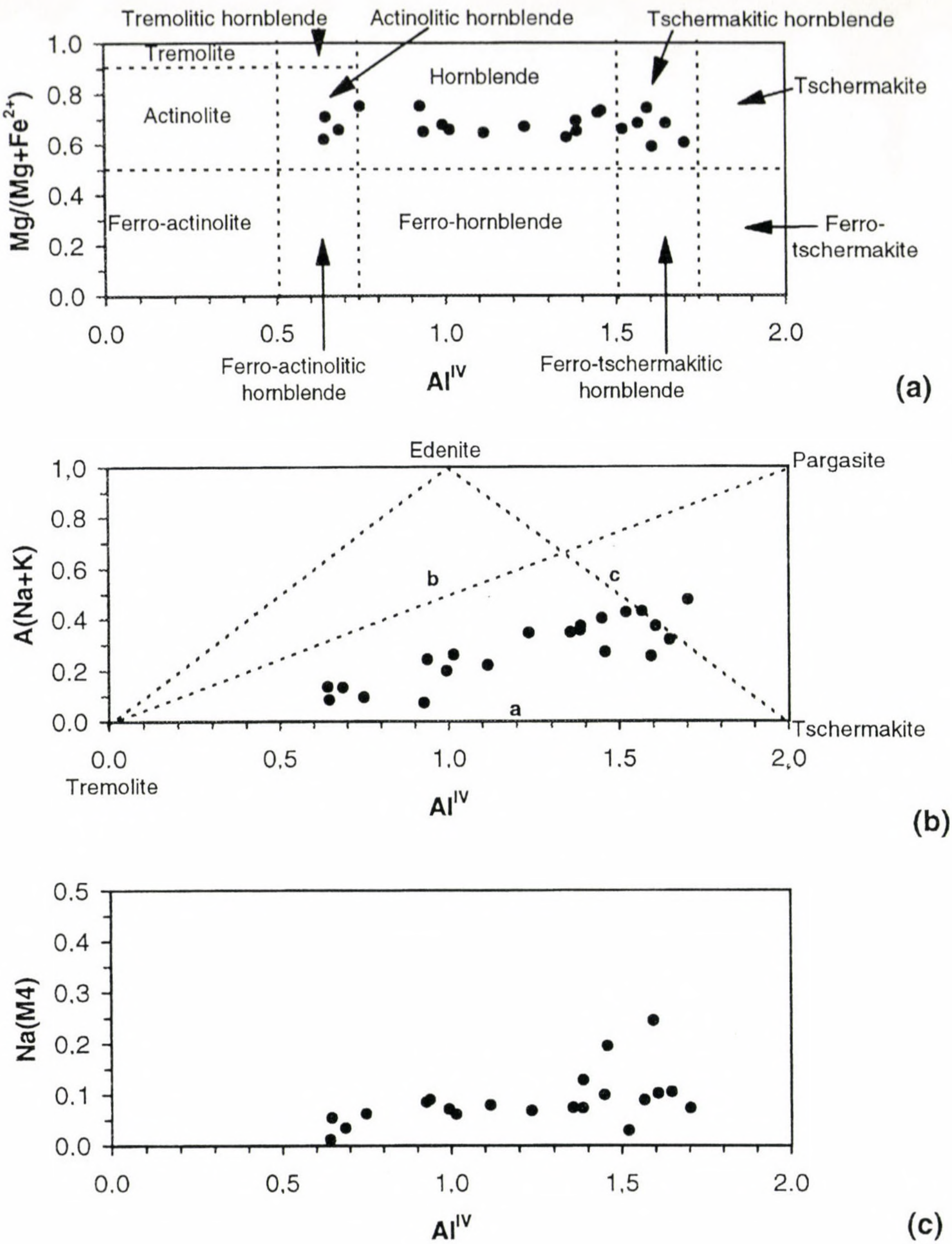


Fig. 5
Chemical compositions of amphiboles from polymetamorphic rocks

Table 5
Average chemical compositions of chlorites from the polymetamorphic complex (borehole Sáránd-I)

Core/depth	2/3017.4 m		2/3018.6 m		3/3435.4 m		4/3656.3 m		4/3656.7 m	
n	6		7		9		5		7	
	x	s	x	s	x	s	x	s	x	s
SiO_2	25.25	0.50	26.31	0.96	26.06	1.10	27.26	0.81	27.05	0.71
TiO_2	0.14	0.08	0.12	0.04	0.23	0.14	0.06	0.09	0.03	0.02
Al_2O_3	21.26	0.44	21.13	1.14	20.29	1.56	20.01	1.37	19.55	0.89
FeO	31.60	1.31	30.18	1.32	29.43	1.89	24.28	1.45	22.71	1.28
MgO	10.06	1.03	10.80	0.61	11.49	0.82	16.44	0.93	17.80	1.02
MnO	0.41	0.06	0.25	0.04	0.57	0.14	0.37	0.11	0.41	0.04
CaO	0.04	0.01	0.10	0.03	0.23	0.41	0.38	0.44	0.06	0.06
Na_2O	0.01	0.01	0.00	0.01	0.00	0.00	0.06	0.14	0.00	0.00
K_2O	0.02	0.01	0.07	0.03	0.66	0.75	0.05	0.08	0.00	0.00
Total	88.79	0.48	88.96	1.03	88.96	0.66	88.90	1.97	87.61	0.90
Number of cations per 28 oxygens										
Si	5.433	0.093	5.587	0.192	5.561	0.220	5.629	0.151	5.625	0.126
Al^{IV}	2.567	0.093	2.413	0.192	2.439	0.220	2.371	0.151	2.375	0.126
Al^{VI}	2.825	0.125	2.876	0.118	2.661	0.194	2.496	0.123	2.416	0.110
Ti	0.023	0.012	0.019	0.006	0.037	0.022	0.009	0.013	0.005	0.003
Fe^{2+}	5.687	0.249	5.358	0.205	5.255	0.399	4.197	0.293	3.952	0.259
Mg	3.225	0.326	3.420	0.211	3.652	0.232	5.062	0.276	5.515	0.254
Mn	0.075	0.012	0.044	0.007	0.103	0.026	0.064	0.017	0.072	0.007
Ca	0.009	0.003	0.022	0.006	0.054	0.096	0.084	0.097	0.014	0.014
Na	0.005	0.004	0.002	0.003	0.000	0.000	0.024	0.053	0.000	0.000
K	0.004	0.003	0.018	0.008	0.178	0.204	0.012	0.021	0.000	0.001
Total	19.853	0.093	19.759	0.068	19.940	0.079	19.948	0.030	19.974	0.060

x – average; s – standard deviation; n – number of measurements

(IC) and chlorite crystallinity [ChC(001) and ChC(002)] zones correspond to those described by Árkai (1991a) and Árkai et al. (1995c). All of the IC and ChC values are characterized by distributions displaying one well-defined maximum. The mean of the IC values corresponds to the boundary between the anchizone and epizone (Fig. 6), while the mean of the ChC(001) falls into the epizone and the mean of the ChC(002) to the high-T part of the anchizone (Table 6). On the basis of the rough temperature calibration of the zone boundaries (Frey 1987; Árkai 1991) an approximate temperature of ca. 300 °C can be deduced for the retrograde recrystallization connected to mylonitization and cataclastic deformation.

There are no systematic differences in IC and ChC values either between the various rock types or in function of the intensity of cataclastic deformation, confirming that cataclasis might cause mainly mechanical re-distribution and crushing only without any significant changes in chemistry and structure of these phyllosilicates.

Traces of very weak subaerial or near-surface chemical weathering modified by subsequent burial diagenesis are found only in 4 samples of cores Nos 2 and 5, represented by subordinate amounts of illite/smectite and rarely, smectite/chlorite.

The Sáránd Parautochthon unit

Rock types and modal composition

In the Sáránd-I borehole the upper part of the parautochthon between 3941 and 3991.6 m is made up of cipollino-like calcite-bearing dolomite marble, dolomite slate with fine-grained white K-mica-rich bands and networks, and dolomite-bearing pelitic-silty slates (cores Nos 6 and 7). In the acid-insoluble residues of the carbonatic rocks and in the pelitic-silty slates quartz and white K-mica predominates, pyrite is a characteristic constituent, while plagioclase and chlorite occur only sporadically, in small amounts.

Below this dolomitic upper part, rocks formed from various mixtures of pelitic and calcareous sediments are known from the cores Nos 8 and 9 between 4047.8 and 4134 m. The rock-forming minerals of the banded calcschist and marble (often with cipollino-like white mica and chlorite bands and networks) are calcite (dominant), quartz (considerable), white K-mica, chlorite and dolomite (subordinate). Plagioclase is practically absent. Thus, chemically strongly weathered ("mature") terrigenous detrital material was mixed with the dolomitic (upper part) and calcareous sediments (middle part of the parautochthon).

In the footwall of the calcareous-pelitic middle part (with certain overlap) a thick lower part characterized by volcano-sedimentary rocks was distinguished (cores Nos 9–14 in the depth interval 4132.3–4799.2 m). In this sub-unit fine-grained volcanoclastic rocks of intermediate-basic chemistries, terrigenous

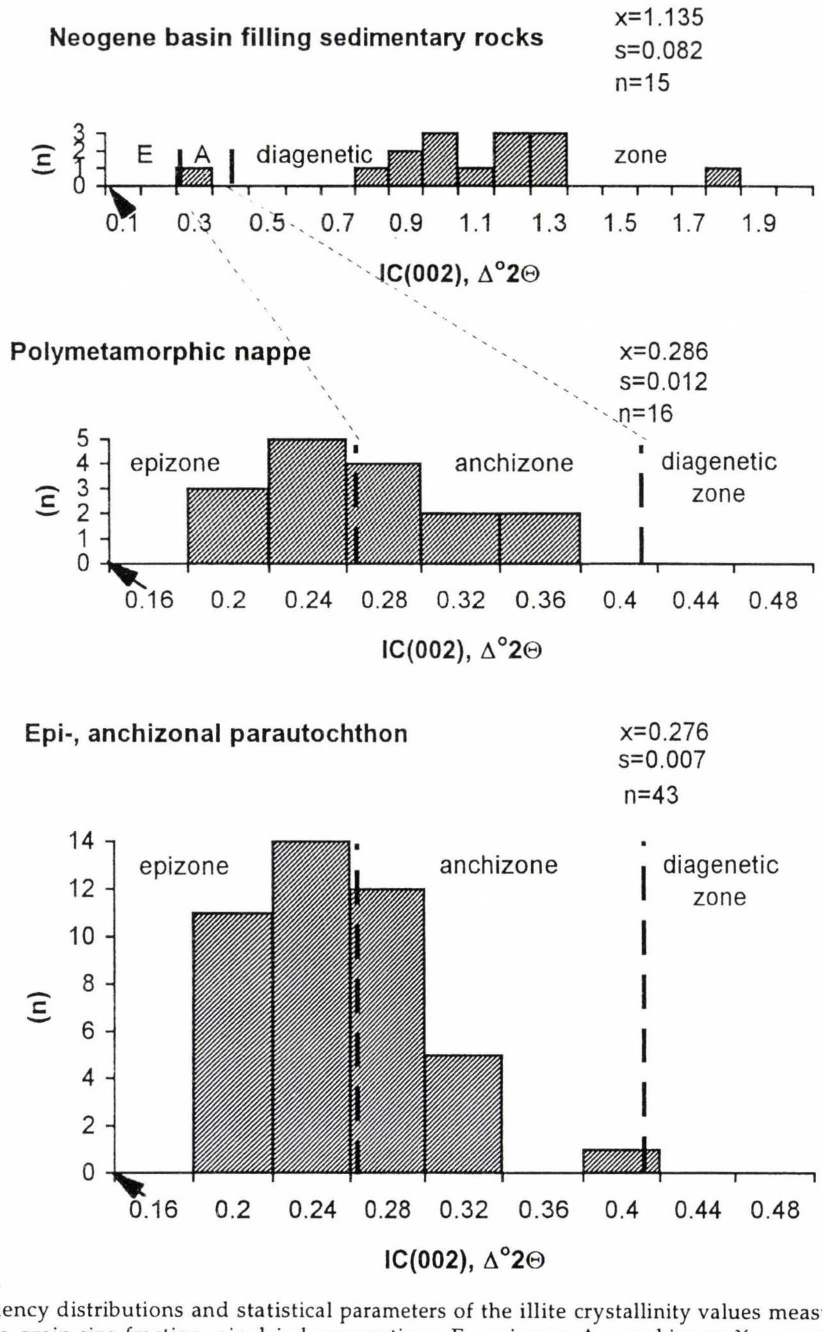


Fig. 6
Frequency distributions and statistical parameters of the illite crystallinity values measured on <2 μm grain-size fraction, air-dried preparations. E– epizone; A – anchizone; X – average; s – standard deviation; n – number of measurements

fine clastic and chemical calcareous sediments were mixed in strongly varying proportions, usually characterized by the predominance of the volcanoclastic material (banded, calcareous metatuffites with slaty cleavage). The considerable amounts of albite as compared to the quartz, the chlorite>white mica ratio, the common occurrence of hematite and the microstructure with lenses and bands typical of the volcanoclastics all indicate the volcanogenic-sedimentary origin of these rocks. Their intermediate–basic chemistry is deduced from the lack of quartz and feldspar clasts and from the dominance of chlorite in addition to the major element bulk chemical compositions.

Considering the lithologies, modal compositions and microstructural features the rock types described from cores Nos 5–10 between 2862.1 and 3410.7 m of the Bagamér Bam-1 borehole [brecciated, recrystallized limestone, banded, calcareous–pelitic slate, pelitic-silty slate and subordinately: psammitic calcschist, shistose marble, and banded slate (pelitic–carbonatic–psammitic)] are very similar to the middle (pelitic–calcareous) sub-unit of the parautochthon. The banded, slaty metatuffites from cores No. 11 of borehole Bam-1 and No. 3 of borehole Bam-2 are correlated to the lower (metatuffitic) sub-unit of the parautochthon (Fig. 7). The considerable amounts of albite and chlorite and the practical lack of white K-mica indicate the basic origin of the volcanoclastic material. Core No. 12 of borehole Bam-1 consists of a recrystallized brecciated limestone. In borehole Bam-2, beneath the metatuffite a hypabissal, porphyritic metadiorite body was reached. Clynopyroxene and plagioclase represent the relic magmatic minerals of the metadiorite which was pervasively altered producing an assemblage of quartz, albite, epidote, chlorite, white K-mica, hematite and rutile.

Microstructural features

Most of the rock types that build up the Sáránd Parautochthon display various kinds of foliation. They are commonly banded as a consequence of rhythmically alternating proportions of terrigenous clastic, carbonatic and volcanoclastic sediments. Foliation originating from the preferred orientation of grains (re)crystallized in an inhomogeneous, anisotropic stress field is also a common phenomenon. The rocks generally display slaty and crenulation cleavage, where the plane of the first foliation (slaty cleavage) is usually parallel with the sedimentary layering. Crenulation of the first foliation planes and shearing of the asymmetric microfolds, producing secondary (transversal or crenulation) cleavage planes are also frequent. The formation of crenulation cleavage is often accompanied by small-scale material transport: phyllosilicates and opaque minerals accumulate in the cleavage planes while the inter-cleavage metadomains become enriched in quartz, albite and carbonate minerals.

"Spiny-like" quartz-mica overgrowths around detrital quartz grains and quartzite structures, characteristic of anchizonal and epizonal transformations (see Frey 1970) are equally frequent. The carbonate minerals strongly

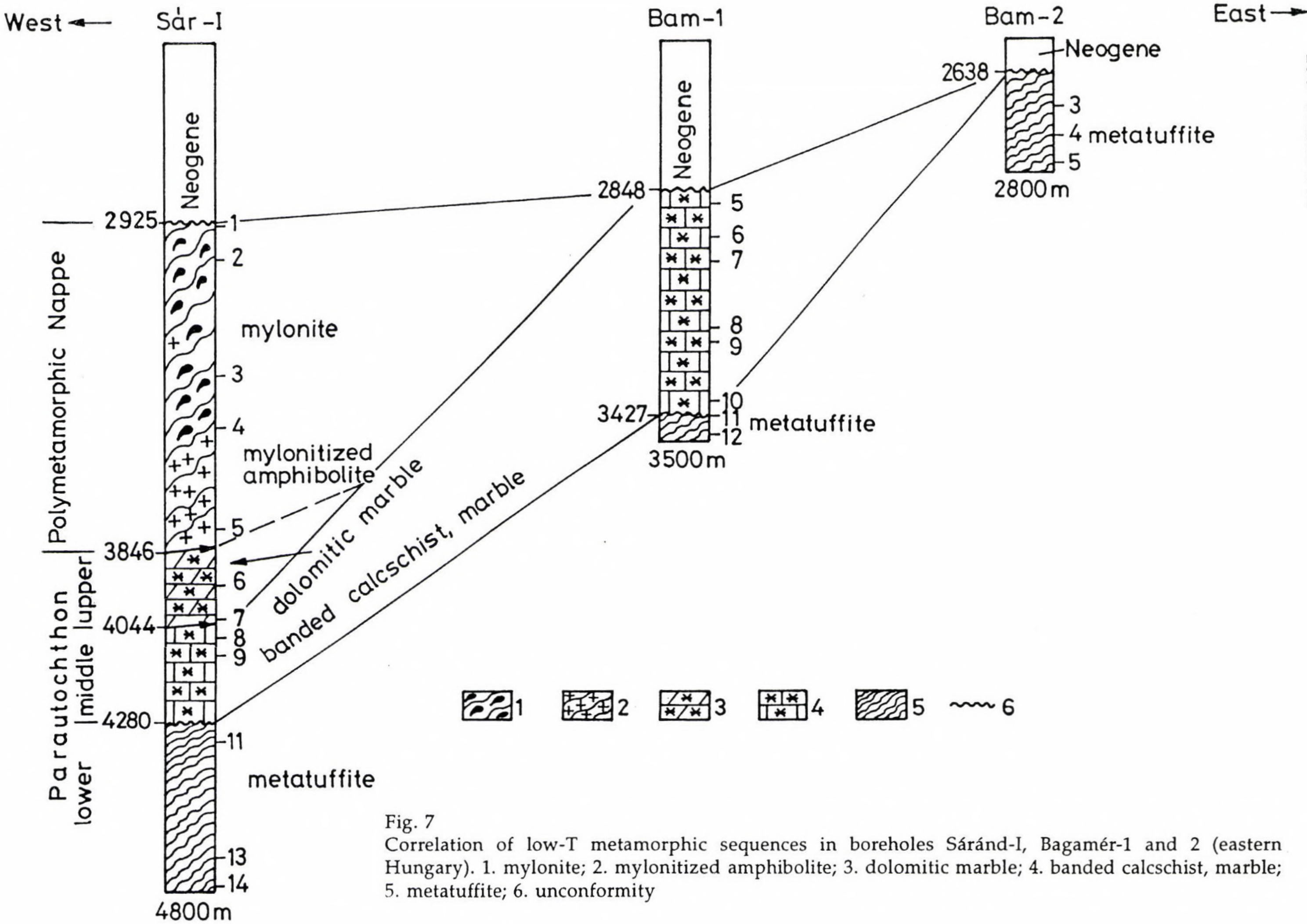


Fig. 7
Correlation of low-T metamorphic sequences in boreholes Sáránd-I, Bagamér-1 and 2 (eastern Hungary). 1. mylonite; 2. mylonitized amphibolite; 3. dolomitic marble; 4. banded calcschist, marble; 5. metatuffite; 6. unconformity

recrystallized in the anisotropic stress field, displaying a shape-preferred orientation of grains.

All these microstructural features suggest very low- to low-grade orogenic (dynamothermal) metamorphism of the Sáránd Parautochthon.

Metamorphic mineral assemblages

The mineral assemblages commonly consisting of quartz, albite, white K-mica, chlorite, carbonate minerals (dolomite, calcite, siderite), hematite, and pyrite are not indicative of the physical conditions of metamorphism. The secondary mineral assemblage of the metadiorite (sample Bam-2.4) that contains quartz, albite, chlorite, muscovite, epidote, calcite and hematite may be stable both in the chlorite zone of the greenschist facies (ca. 300–450 °C) and in the high-grade anchizone (ca. 250–300 °C). In one metatuffite sample stilpnomelane was also determined, in addition to quartz, albite, chlorite, calcite and hematite. Stilpnomelane is usually formed in Fe-rich, anchizonal metasedimentary rocks as well as in metavolcanoclastics also containing pumpellyite or glaucophane (Frey 1987). Breitschmid (1982) reported the lowest-grade occurrence of stilpnomelane from a late diagenetic environment (210 °C/1.6 kbar).

Illite and chlorite crystallinity, vitrinite reflectance

Frequency distributions and statistical parameters of illite crystallinity [IC(002)] are shown in Fig. 6. The crystallinity values display quasi-normal distributions with one maximum. The means of the IC(002) and ChC(002) values correspond to the boundary between the anchizone and epizone (in the sense of Kübler 1990), while that of the ChC(001) values falls into the epizone.

The IC and ChC mean values of the various rock types are summarized in Table 6. The upper (dolomitic – fine clastic) part of the parautochthon is characterized by IC values corresponding to the low (ca. 200–250 °C) temperature part of the anchizone. The average vitrinite reflectance values measured in this sub-unit confirm this statement deduced from the IC data (Table 7).

The middle (calcareous – pelitic) and the lower (metatuffitic) parts of the parautochthon show epizonal IC and ChC averages. The relatively large difference in IC between the upper and middle sub-units cannot sufficiently be explained either by lithologic effects or by the differences in present burial depths. As the difference in depth of burial between the low-T anchizonal core No. 7 and the epizonal core No. 8 of the borehole Sáránd-I is only 50–60 m, the authors must assume that the two sub-units came into close spatial relation by post-metamorphic tectonic movements, providing an example of "transported metamorphism" which is well known in the Alpine and Pannonian–Carpathian realms (see Frey 1988; Árkai 1983; Árkai and Kovács 1986). It is worth mentioning that the metatuffitic rocks generally show somewhat smaller IC and ChC values (implying somewhat higher apparent

Table 6
Illite and chlorite crystallinity averages measured on <2 μm grain-size fraction, air-dried samples

Complex	Lithology	Borehole	IC(002)			ChC(001)			ChC(002)		
			x	s	n	x	s	n	x	s	n
Neogene basin-filling formations	clastic and clastic-carbonatic sediments, ignimbritic rhyolite and tuff	Bam-1, Bam-2 and Nyáb-1	1.135	0.082	15	–	–	–	0.443	0.012	8
Polymetamorphic nappe	mylonite, mylonitized gneiss, mica schist, amphibolite	Sáránd-I	0.286	0.012	16	0.287	0.012	12	0.298	0.011	17
Epi-, anchizonal metamorphic Mesozoic	dolomitic – fine-clastic (upper) group	Sáránd-I	0.353	–	4	–	–	–	–	–	–
	calcareous–fine-clastic (middle) group	Sáránd-I	0.278	0.034	25	0.256	0.032	10	0.280	0.055	10
	basic-intermediate, carbonatic meta-tuffitic (lower) group	Sáránd-I Bam-1	0.249	0.029	14	0.255	0.026	20	0.260	0.019	24

x – mean; s – standard deviation; n – number of measurements;
The intervals of the anchizone are 0.284–0.435 for IC(002), 0.310–0.390 for ChC(001) and 0.284–0.348 $\Delta\ ^{\circ}2\Theta$ for ChC(002), respectively (for details see Árkai et al. 1995c)

Table 7
Vitrinite reflectance data

Borehole	Core	Depth (m)	R_{random}(%)	s	R_{max}(%)	s	R_{min}(%)	s	n
			Neogene basin-filling sedimentary rocks						
Nyáb-1	7	1754.0	0.63	0.10					50
Bam-1	3	2386.4	0.93	0.30					50
Bam-1	3	2388.2	1.01	0.29					50
			Epi-, anchizonal metamorphic parautochthon						
Sáránd-I	6	3942.8	2.82	0.56	3.71	0.46	1.02	0.34	50
Sáránd-I	6	3942.2	3.59	0.71	4.30	0.68	1.10	0.32	50

s – standard deviation; n – number of measurements

metamorphic grades) than the pelitic-carbonatic rocks. This relatively small, but systematic difference may be explained by lithologic effects, namely by the higher porosity and permeability of the tuffitic rocks.

White K-mica geobarometry

The method elaborated by Sassi (1972) for metapelitic rocks of lower greenschist facies (chlorite zone) and also extended by Padan et al. (1982) to the high-T part of the anchizone could not be applied to the Sáránd Parautochthon in a statistical way, because the lithological features and modal compositions in most cases do not fit the requirements given by Guidotti and Sassi (1976, 1986). There are only three samples with adequate (pelitic-silty) lithology and modal composition (quartz>white K mica>>albite, chlorite±calcite) that give *b* values of 8.976–8.978 Å for the low-T anchizonal and 9.005 Å for the transitional anchizonal/epizonal samples, implying low or transitional, low-intermediate pressure (i.e., high, transitional high-intermediate metamorphic thermal gradient) range.

All of the other rock types of the parautochthon proved to be inadequate for pressure estimation. In the upper sub-unit the albite-bearing dolomite marble provided 8.990 Å, while the albite-free dolomitic rock types gave *b* values scattering between 9.003 and 9.028 Å. The calcareous – fine clastic rocks of the middle sub-unit are devoid of albite. Their *b* values scatter between 9.008 and 9.050 Å, giving a mean of 9.019 Å (s=0.016 Å, n=6). Due to their higher Fe, Mg and lower Al contents as compared with the pelitic-silty rocks, the metatuffites provided high *b* values (in average: 9.044 Å, s=0.010 Å, n=6). The albite-free, calcite-bearing pelitic-silty slates and the calcschist and marble rock types from borehole Bam-1 are characterized by moderately lower *b* values than the corresponding middle sub-unit of the parautochthon in borehole Sáránd-I (b=8.998–9.011 Å, in average: 9.005 Å, s=0.005 Å, n=12 in the pelitic-silty and b=8.998–9.025 Å, in average: 9.006 Å, s=0.015 Å, n=6 in the calcareous rocks).

Burial diagenesis of the Neogene basin-filling formations

Terrigenous pelitic-silty, subordinately psammitic sediments with varying carbonate contents form the younger, thick, dominant (Pannonian) part of the Neogene section, while the thin, lower (Miocene) part consists of acidic volcanoclastics (ignimbritic rhyolite, rhyolite tuff, rhyodacite tuff) and small amounts of limestone. The terrigenous rocks contain considerable amounts of detrital constituents such as quartz, plagioclase, muscovite and chlorite, implying chemically not or only weakly weathered detritus. Quartz, plagioclase, K-feldspar and partially chloritized, opacitized biotite represent the magmatic mineral phases of the volcanoclastic rocks.

The <2 μm grain-size fractions of the terrigenous clastic rocks are built up by illite, illite/smectite irregular mixed layer clay mineral, kaolinite, chlorite and subordinately by smectite. The IC(002) and ChC(002) mean values of the Neogene rocks unequivocally indicate diagenetic conditions (Fig. 6 and Table 6). The large IC and ChC values together with the R_{random} averages (Table 7) suggest that the diagenesis of the Neogene sedimentary rocks may be correlated to the intervals of the upper wet gas and heavy oil zones as classified by Kübler et al. (1979). The occurrences of analcime (Bam-2.2 and Nyáb-1.2) and clinoptilolite (Nyáb-1.1., 1369 m) in the carbonate-free acidic volcanoclastics and their clay mineral assemblages (illite/smectite mixed-layers, smectite, kaolinite, subordinate amounts of chlorite) is in accordance with the conclusions obtained from the IC and R data of the terrigenous sedimentary rocks.

Geochronologic data and considerations

On the basis of the analogies in lithology and metamorphic evolution with the neighboring metamorphic formations, an early Variscan or pre-Variscan age is supposed for the medium thermal gradient, amphibolite facies prograde event (see Árkai et al. 1985; Szederkényi et al. 1991). The age of the retrogression connected to the mylonitization and cataclastic deformation has been uncertain so far: late Variscan and/or Alpine ages seem to be possible as well. In order to determine the age(s) of retrogression in the overthrusted polymetamorphic pile and the age of the progressive low-T metamorphism of the Sáránd Parautochthon, K-Ar dates of the white K-mica-rich <2 μm grain-size fractions were determined (Table 8). For the description of sample preparation, analytical procedure, quality of measurements and a detailed, critical review of geochronological interpretation of the white K-mica K-Ar dates the authors refer to Árkai et al. (1995a).

The K–Ar dates of the Sáránd Parautochthon scatter between 68.3 and 91.7 Ma, with a mean value of 81.1 Ma (s=8.1 Ma, n=11). There are no systematic differences in ages between the upper (dolomitic) and the middle (calcareous) and lower (metatuffitic) sub-units. The date of the metadiorite (Bam-2.4) falls within the interval given above. On the basis of the measured well temperatures,

Table 8
K-Ar isotopic dates of the illite-muscovite-rich, <2 μm grain-size fractions

Borehole	Core	Depth (m)	K (weight %)	$^{40}Ar_{(rad/g)}$, ncm^3/g	$^{40}Ar_{rad}$(%)	K–Ar date (Ma)
Polymetamorphic nappe						
Sáránd-I.	1.	2929.0	3.82	$1.2118x10^{-5}$	24.6	79.8±5.0
Sáránd-I.	2.	3017.4	4.74	$1.8727x10^{-5}$	59.3	99.0±4.0
Sáránd-I.	3.	3434.2/b	3.46	$1.3363x10^{-5}$	47.1	96.8±4.2
Sáránd-I.	5.	3796.5	2.60	$8.7668x10^{-6}$	39.0	84.8±4.0
Sáránd-I.	5.	3796.85	1.13	$4.3836x10^{-6}$	36.8	97.2±4.7
Epi-, anchizonal parautochthon						
Sáránd-I.	6.	3941.0	6.43	$2.0692x10^{-5}$	59.7	81.0±3.3
Sáránd-I.	6.	3942.2	5.32	$1.7184x10^{-5}$	62.4	81.3±3.3
Sáránd-I.	7.	3991.6	7.76	$2.0990x10^{-5}$	67.0	68.4±2.7
Sáránd-I.	9.	4132.3	4.31	$1.1649x10^{-5}$	43.7	68.3±3.0
Sáránd-I.	9.	4134.0	5.38	$1.6898x10^{-5}$	32.1	79.0±4.1
Sáránd-I.	11.	4401.1/b	5.36	$1.60155x10^{-5}$	51.4	75.4±4.2
Sáránd-I.	13.	4728.5/a	1.60	$5.8141x10^{-6}$	33.5	91.2±4.6
Sáránd-I.	14.	4795.6	4.34	$1.3933x10^{-5}$	55.7	80.8±3.5
Bam-1.	6.	2888.0	5.49	$1.9558x10^{-5}$	60.5	89.4±3.6
Bam-2.	3.	2644.7	5.99	$2.0320x10^{-5}$	80.0	85.3±3.3
Bam-2.	4.	2702.45	2.00	$7.3058x10^{-6}$	34.9	91.7±4.6

a non-corrected thermal gradient of 54 $^{o}Ckm^{-1}$ is calculated by linear approximation, from which – taking into consideration the thermal disturbances around the boreholes, a real present-day gradient of ca. 60 $^{o}Ckm^{-1}$ can be deduced. Thus, a present temperature of ca. 290 °C is expected at the TD of borehole Sáránd-I, namely at 4800 m from KB. Surprisingly, there is no significant correlation between the present burial depth ranging from ca. 2650 to 4800 m and the K–Ar dates of the samples from the parautochthon, implying that the differences in present-day temperatures connected to the differences in present burial depth of the investigated samples have had no appreciable differential effects on the K–Ar dates. This statement is in apparent contradiction with the results of Hunziker et al. (1986), who have found that the resetting of the <2 μm illite-muscovite K–Ar dates (i.e., the opening of the K-Ar system) becomes complete at a temperature of 260±20 °C during 10±5 Ma. Considering that the rapid, intense burial (deepening) of the given part of the Pannonian Basin started only at the beginning of the Pannonian stage (ca. 5 Ma ago), the effective heating time in this case might have been much smaller than in the case reported by Hunziker et al. (1986). However, the possibility of certain rejuvenation of the K-Ar dates caused by the Neogene burial and heating could not be ruled out, although its measure should not have exceeded 10–20%, considering the data of Hunziker et al. (1986) and taking also into account the results of Balogh et al. (1990). Thus, the K–Ar dates presented here indicate an Upper Cretaceous (post-Austrian or -Subhercynian) uplift (cooling below 260±20°C) of the low-T metamorphic parautochthon. Consequently, taking into account the Triassic stratigraphic age of the Sáránd Parautochthon (or at least its upper part; see Bérczi-Makk 1985) and the Alpine metamorphic evolution paths of the tectonic domains in Hungary (Árkai 1991b), the age of the prograde metamorphism of the parautochthon is most probably Cretaceous, connected to the Austrian (ca. 100–105 Ma) phase.

Cretaceous dates (80–99 Ma, in average: 91.5 Ma, s=8.6 Ma, n=5) were also obtained from the white K-micas of <2 μm grain-size fractions separated from the overthrusted polymetamorphic complex (Table 8). These values and their homogeneous distribution suggest that the retrograde metamorphism connected to the mylonitization might be most probably of Eo-Alpine age (Cretaceous, Austrian phase). If this retrograde metamorphism would have been much older (e.g., Variscan), and the rocks would have been subjected only to partial "reworking" (incomplete recrystallization) during the Alpine tectonocycle, the K–Ar dates should scatter in a large time interval, as has been documented by Árkai and Balogh (1989) in an other part of the Pannonian Basin.

The K–Ar dates of the polymetamorphic complex are slightly but systematically higher than those of the low-T metamorphic parautochthon. To this difference the following explanations can be given:

(i) during the post-metamorphic uplift (cooling) the polymetamorphic pile was in a higher tectonic position, and consequently, it reached the closure temperature of the white K-mica K-Ar system earlier than the parautochthon;

(ii) the crushed fragments of the 1st (amphibolite facies) muscovite generation are also present (even in small proportion) in the <2 μm fractions of the polymetamorphic rocks, which may cause an increase of K–Ar dates.

Because the disturbing effect of detrital (older) white micas cannot be ruled out either in the rocks of the parautochthon (disregarding the volcanogenic ones), explanation (i) seems to be more reliable at present.

Conclusions, relations of metamorphism and tectonics

The overthrusted part of the polymetamorphic basement is built up by alternations of mica schist, gneiss, quartzite and amphibolite formed from terrigenous clastic and basic magmatic (volcanoclastic) rocks. The observable first, prograde metamorphism of presumed early-Variscan(?) or pre-Variscan(?) age produced medium thermal gradient amphibolite facies assemblages. This event was followed – after a considerable time span – by a retrograde metamorphism connected to mylonite formation. The temperature of the retrograde event decreased from ca. 450 °C, as proved by the amphibolites, to ca. 320–300 °C, as deduced from the chlorite geothermometric and illite and chlorite crystallinity data of the fine-grained mylonitic matrix. The subsequent cataclastic deformation was not accompanied by any considerable metamorphic recrystallization.

The footwall Sáránd Parautochthon consists of three sub-units. The upper, paleontologically proved Middle Triassic part is built up by dolomitic and fine-clastic rocks, while the middle and lower (most probably also Mesozoic, Triassic or Jurassic?) sub-units are characterized by calcareous and terrigenous finely clastic and calcareous – basic-intermediate volcanoclastic lithologies. The whole parautochthon was exposed to an orogenic (regional dynamothermal) metamorphism. The metamorphic grade of the upper sub-unit corresponds to the low-temperature (ca. 200–250 °C) part of the anchizone, while that of the middle and lower sub-units mainly to the epizone, the temperature of which might be very close to that of the boundary between the anchizone and epizone (ca. 300 °C).

The maturity parameters of the Neogene terrigenous and volcanogenic clastic rocks indicate only diagenetic alterations. Considering the actual geothermal conditions, both clay mineral and vitrinite characteristics refer to non-equilibrium conditions: the coalification and the clay mineral aggradation have been retarded, presumably because of the rapid burial.

Comparing the diagenetic–metamorphic grade indicators, sharp changes in IC, ChC and clay mineral associations have been demonstrated between the Neogene sedimentary cover and the basement complexes. There are no present-day depth-related significant, systematic, or gradual changes in the

K–Ar dates of fine-grained white K-micas and in the IC and ChC distributions either in the polymetamorphic or in the prograde, low-T metamorphic complexes. All these facts suggest that the increase of temperature (up to ca. 290 °C at 4800 m) and pressure due to the Neogene burial has had no detectable effects on the mineralogical and petrographic indicators of metamorphism, most probably because of the short (<5, presumably <3 Ma) effective heating time during the Neogene.

The Mesozoic Sáránd Parautochthon and the overthrusted polymetamorphic complex were uplifted in the Upper Cretaceous, being already in their present tectonic position: the overthrusted polymetamorphic complex cooled down and reached the closure temperature of the <2 μm white K-mica (ca. 260 °C) earlier than the underlying parautochthon.

Disregarding the upper (dolomitic – fine clastic) sub-unit which as a whole shows a definitely lower metamorphic grade than the underlying sub-units of the Sáránd Parautochthon, there are no distinct gradual IC and ChC changes in the overthrusted and underlying complexes (Fig. 8) proving that the retrograde metamorphism in the overthrusted part and the prograde metamorphism in the footwall were older than the overthrusting.

Considering the K–Ar age distributions and the (at least partly proved) Triassic stratigraphic age of the Sáránd Parautochthon, the retrograde metamorphism of the polymetamorphic complex and the prograde regional metamorphism of the parautochthon might be connected to the eo-Alpine, Cretaceous Austrian phase. Judging from the sporadic geobarometric data, the polymetamorphic complex should have been in a lower crustal position than the Mesozoic Sáránd Parautochthon during this event.

Figures 9 and 10 demonstrate the changes of apparent mean crystallite size [i.e., the mean size of crystallites (in the crystallographic c axis direction) that coherently scatter the X-rays] and lattice strain values determined from the XRD line profiles of the first basal reflections using the modified Voigt method (for details of the calculations see Árkai et al. 1996). The relatively large scatter in size and strain of white K-mica and chlorite – not only in the close vicinity of the main overthrust plane but also in distant samples – show that the tectonic shearing strain connected to the overthrusting process strongly influenced the structural state of the phyllosilicates in a rather irregular manner, acting mostly along thrust-parallel shearing plane swarms both in the hanging wall and the footwall units.

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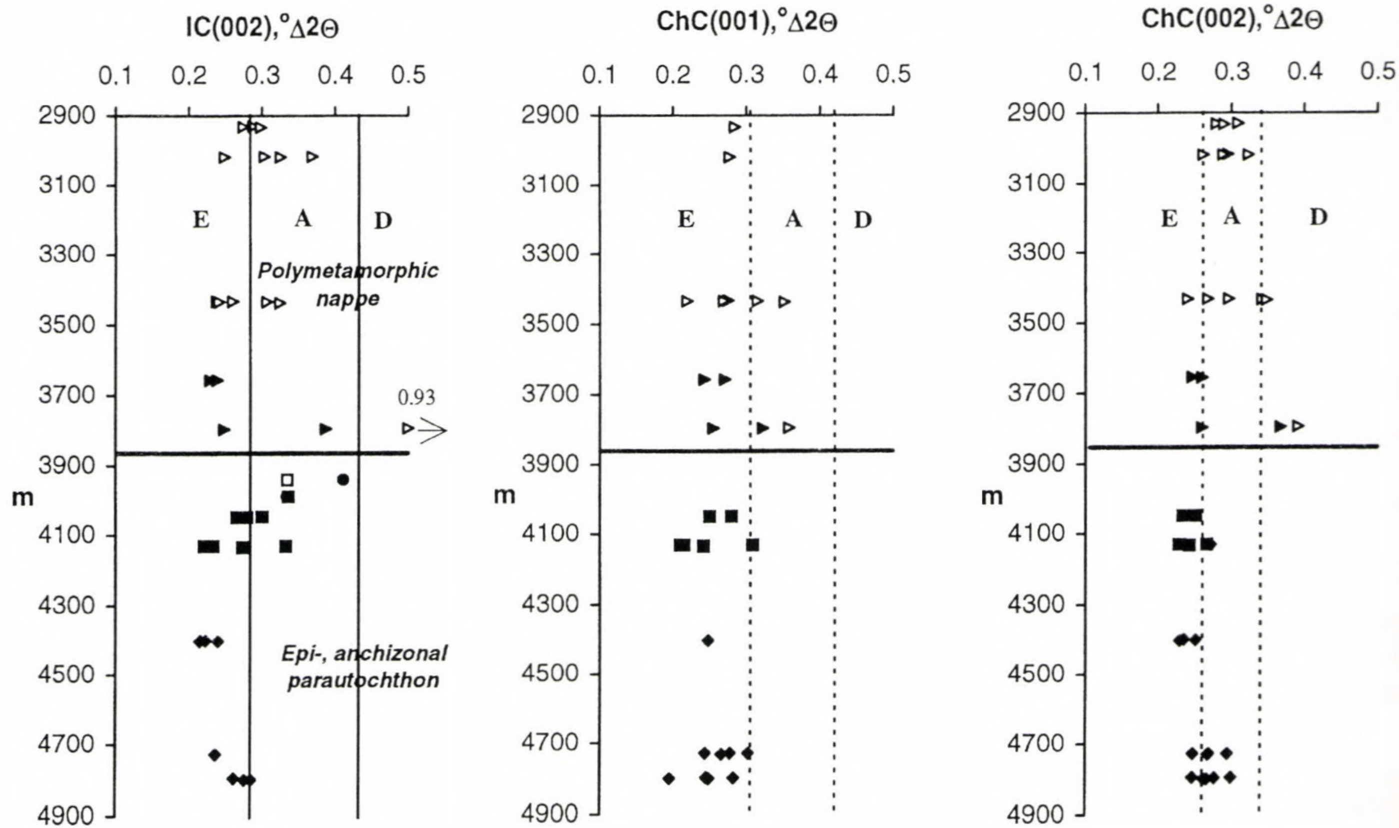

Fig. 8
Variations of illite and chlorite crystallinity indices as a function of present burial depth in the basement units. – mylonite (gneiss, mica schist); ▲ – mylonite (amphibolite); ❑ – dolomitic rocks (upper sub-unit); ● – slate (upper sub-unit); ■ – calcareous, fine clastic rocks (middle sub-unit); ◆ – metatuffitic (lower) sub-unit

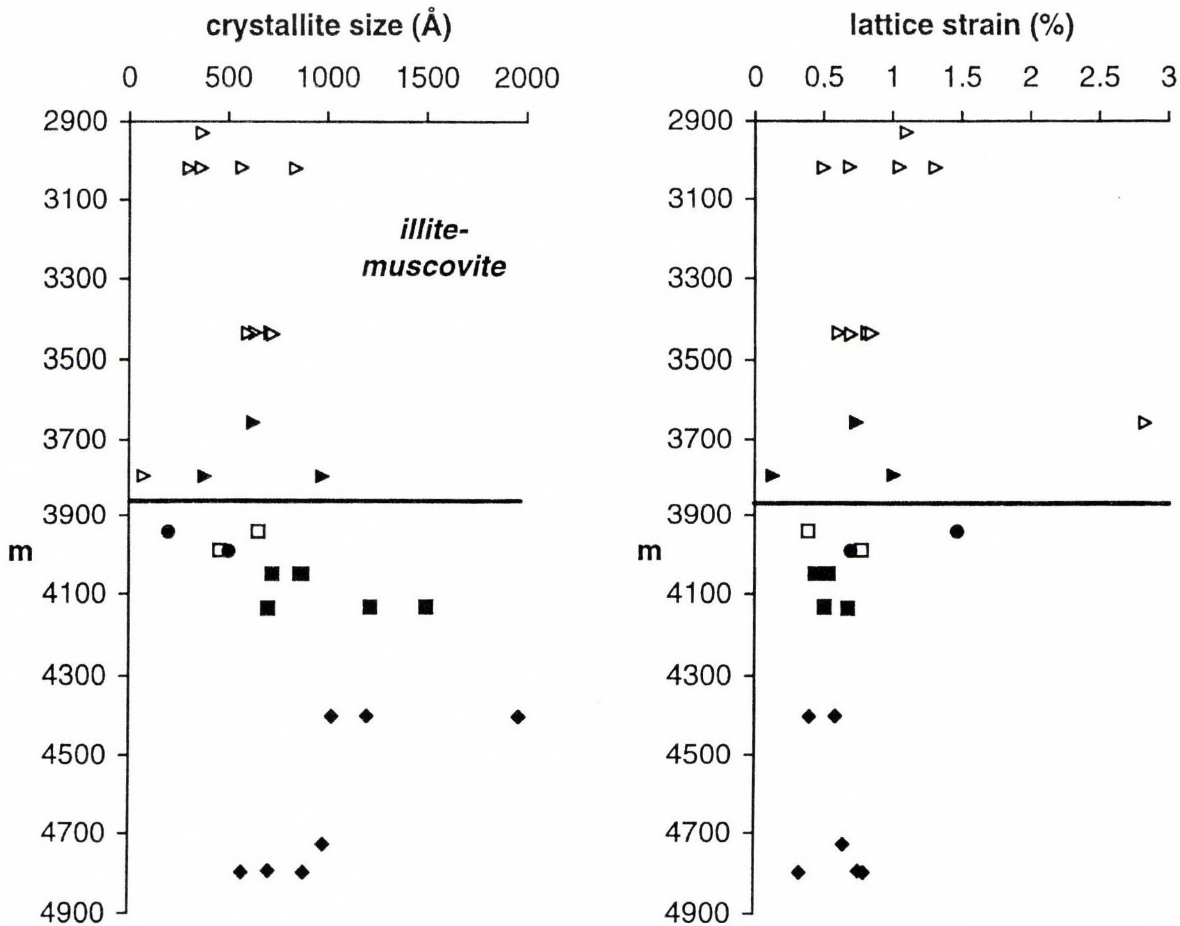

Fig. 9
Changes in mean crystallite size and lattice strain values of illite-muscovite calculated for the <2 μm grain-size fraction samples. For legend see Fig. 8

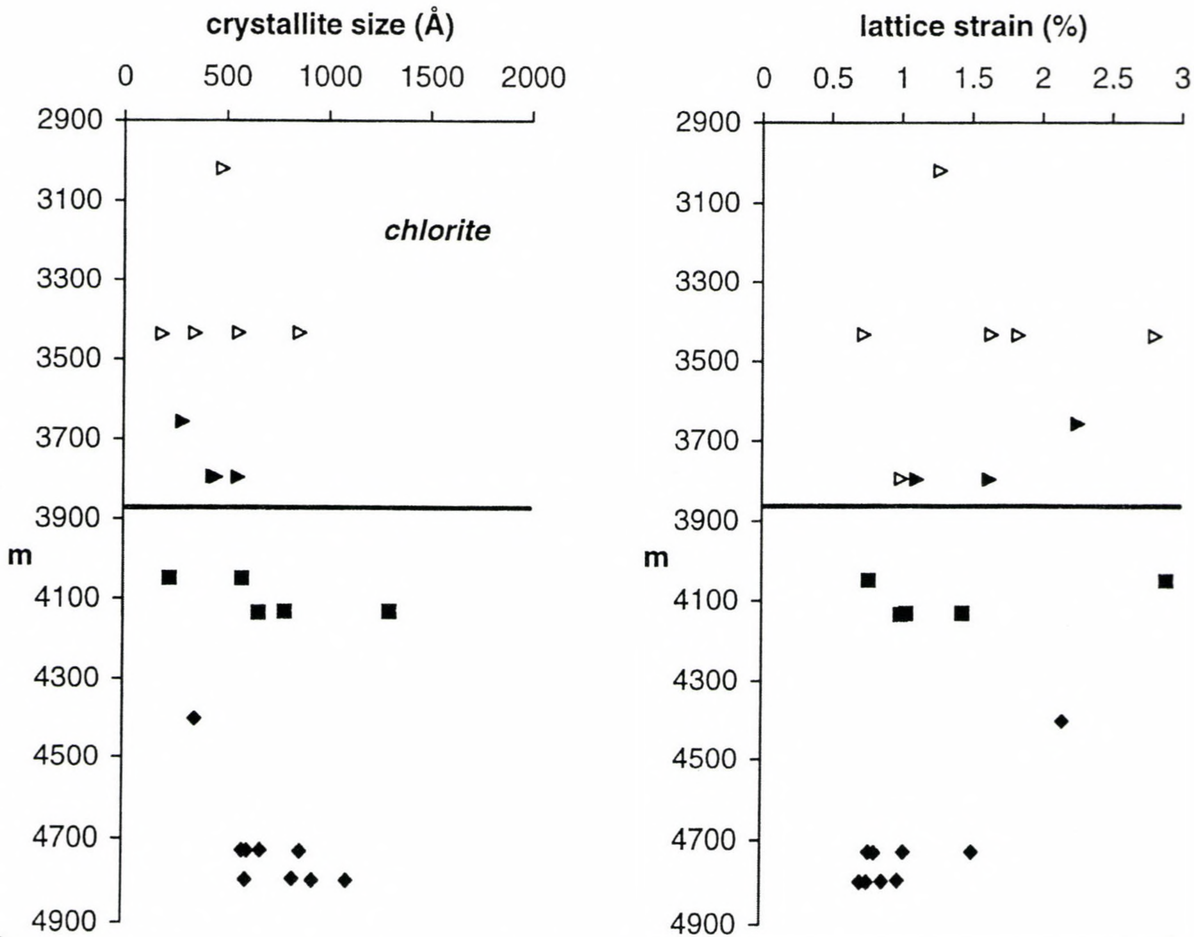

Fig. 10
Variations in mean crystallite size and lattice strain values of chlorite calculated for the <2 μm grain-size fraction samples. For legend see Fig. 8

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Fluid mixing in the Mátra and the Börzsöny ore deposits: A stable C–O–H isotope study

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Stable carbon, oxygen, and hydrogen isotope compositions of carbonates and inclusion fluids trapped in calcites related to the ore deposits of the Mátra and the Börzsöny Mountains have been measured in order to assess the origin and evolution of ore-forming fluids. The complex evaluation of these data revealed that fluid mixing was characteristic for both ore deposits. In the Mátra deposit meteoric water, formation water of meteoric origin, and magmatic water were mixed during ore formation. In the Börzsöny deposit magmatic water and formation water were also present in significant amounts, whereas no direct addition of meteoric water is observed. An additional component with very negative $\delta^{13}C$ and δD values also appears that can be attributed to oxidation of organic matter.

Key words: Mátra, Börzsöny, ore deposit, calcalkaline volcanic rocks, stable isotopes

Introduction and geological background

The ore deposits studied in this paper are situated in North Hungary (Fig. 1) and belong to the Cenozoic Inner Carpathian volcanic arc. Recent petrogenetic studies based on isotope geochemical investigations of these predominantly andesite complexes have been performed by Salters et al. (1988) and Downes et al. (1995). Comprehensive reviews of the geology of the volcanic arc have been published in special volumes of Acta Volcanologica (vol. 7(2), 1995) and AAPG Memoir 45. The volcanic arc consists mainly of subduction-related andesitic rocks with subordinate amounts of dacite and rhyolite.

Many studies have been made on the geology of the *Börzsöny Mts*; thus, the following summary is based mainly on the most important review articles (Liffa and Vígh 1937; Pantó and Mikó 1964; Szádeczky-Kardoss et. al 1967; Kubovics and Pantó 1970; Nagy 1978; Balla and Korpás 1980). The basement of the Börzsöny Mts is composed of metamorphic rocks in the central and northern part and of Mesozoic carbonate rocks in the southeastern part. The contact of these basement complexes has a jigsaw pattern. The volcanic activity of Carpathian–Badenian age began with volcanosedimentary rocks followed by a stratovolcanic series of pyroxene and amphibole andesites. The next stage

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was the collapse of the caldera with contemporaneous emplacements of subvolcanic bodies and radial dikes.

The ore formation is related to the caldera collapse and took place in two stages. The first stage is massive breccia-pipe type ("Rózsabánya-type") characterized by pyrrhotite, Fe-rich sphalerite and calcopyrite with sphalerite inclusions. The fluids responsible for the formation of the second stage ores contained significant amounts of Au and Ag. The ore paragenesis of this stage is characterized by arsenopyrite, galenite, and Bi-minerals. The mineral paragenesis of the Nagybörzsöny deposit is very complex, the description of which can be found in the comprehensive book of Koch (1985).

A synthesis of the geology of the *Mátra Mts* has been published by Varga et al. (1975). The geology and the ore deposits of the Mátra Mts have attracted the attention of many researchers, among whom the most important reviews are given by Vidacs (1957) and Kubovics and Pantó (1970).

Little is known about the basement of the Mátra Mts. Rare xenolith occurrences suggest the presence of metamorphic and granitic rocks below the central and the western part of the mountains, whereas bore-holes revealed Bükk-type Mesozoic carbonate rocks below the eastern part. The main part of the mountains consists of a stratovolcanic series of Lower Badenian and Carpathian. The Badenian "middle stratovolcanic series" (called also "variable andesite" due to its varying alteration) contains significant amounts of lava rocks beside the volcanic debris. The volcanic activity was ended by a basic andesite lava series (called "covering andesite") after a short calm period.

There are two main areas of ore formation in the central and western parts of the Mátra Mts: Gyöngyösoroszi–Mátraszentimre and Parádsasvár. Both deposits are situated in the Lower Badenian "varying andesite". The covering andesite was not affected by the hydrothermal activity. Among the 25 ore dikes known in the Gyöngyösoroszi area, 10 dikes with 500–1000 m lengths and 0.8–1.2 m widths were mined. Most of the dikes are striped with quartz as the dominant mineral. The main ore minerals (Koch 1985; Nagy 1986) are pyrite, gel pyrite, melnikovite, galenite, sphalerite, calcopyrite, marcasite, wurtzite and antimonite. The average metal contents are Pb 0.6%, Zn 4.0%, Au 0.8 ppm, and Ag 75 ppm.

Gatter (1987) has suggested mixing of fluids during ore formation stages on the base of fluid inclusion microthermometric studies of the Mátra Pb–Zn deposit. Such fluid mixing can cause local variations of pH, Eh, and T conditions, giving rise to mineral precipitations. Since ore-forming fluids contain predominantly H–C–O–S compounds, stable isotope investigations can provide clues to the understanding of the origin and evolution of these fluids. This paper presents a stable C, O and H isotope study of carbonates and inclusion fluids trapped in them in the Mátra and the Börzsöny ore deposits. Using these data we will discuss the origin of fluids responsible for the precipitation of ore and gangue minerals in these systems.

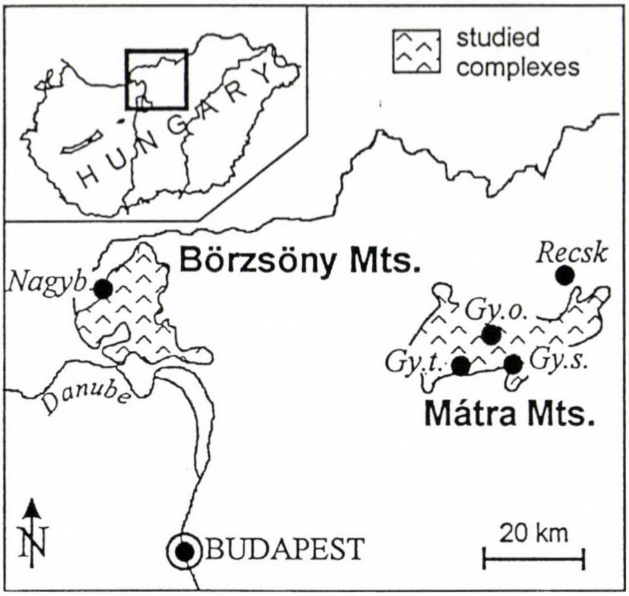


Fig. 1
Location map for the studied ore deposits and host volcanic systems. Nagyb. – Nagybörzsöny; Gy.o. – Gyöngyösoroszi; Gy.t. – Gyöngyöstarján; Gy.s. – Gyöngyössolymos

Analytical methods

Preparation of CO_2 from calcites was performed according to the H_3PO_4 reaction method (25 °C overnight) established by McCrea (1950). Dolomites, rhodochrosites, and siderites underwent reaction at 50 °C for 3 days to 1 week depending on the mineral (Al-Aasm et al. 1990). Inclusion waters in selected calcite samples were released by heating in vacuum at 500 °C, then the H_2O was reacted with zinc (produced by J.M. Hayes, Bloomington, U.S.A.) at 500 °C using the teflon-sealed tube method of Coleman et al. (1982). The hydrogen was transferred into the mass spectrometer with the zinc kept at the reaction temperature, so that the H isotope fractionation due to hydrogen–zinc reaction (Demény 1995) does not affect the δD values measured. $^{13}C/^{12}C$, $^{18}O/^{16}O$ and δD ratios were determined with a Finnigan MAT Delta S mass spectrometer at the Laboratory for Geochemical Research and the results are expressed in the δ notation as ‰ relative to PDB ($\delta^{13}C$) and V-SMOW ($\delta^{18}O$, dD). Reproducibilities of the $\delta^{13}C$ and $\delta^{18}O$ values are better than ±0.2 ‰ whereas replicate measurements of δD values of inclusion waters gave reproducibilities of ±2‰.

Results and discussions

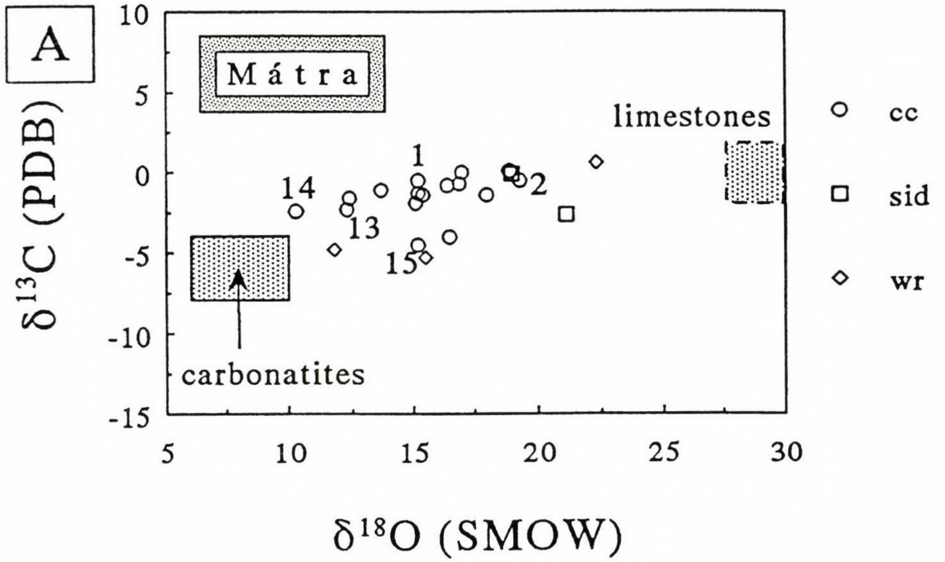
$\delta^{13}C$ and $\delta^{18}O$ compositions of carbonates and dD values of inclusion waters trapped in calcites are listed in Table 1 and plotted in Fig. 2. Carbon isotope compositions obtained from carbonates collected in the *Mátra Mountains* (–5.3 to 0.1‰) scatter between the ranges of mantle-derived carbonatites (–8 to –4‰, Taylor et al. 1967) and marine limestones (–2 to 2‰; Hoefs 1973; see Fig. 2), indicating mixing of fluids derived from magmatic and sedimentary sources. Oxygen isotope compositions of carbonates are determined by fluid compositions and carbonate precipitation temperatures; thus, although the $\delta^{18}O$ values (10.3 to 22.3‰) also scatter between the ranges of mantle-derived carbonatites (6 to 10‰ Taylor et al. 1967) and marine limestones (about 30‰; Hoefs 1973) these data cannot be interpreted mechanically with mixing of magmatic and sedimentary components (see below). The C and O isotope compositions of different carbonate species – calcite veins, siderite crystals and groundmass carbonate of carbonitized andesites – fall in the same trend proving that precipitations of these forms of carbonate were induced by the same process.

δD values of inclusion waters can provide valuable information on the origin of solutions from which the carbonates were precipitated. H_2O trapped in fluid inclusions in minerals that do not contain H in their structure can preserve the original D/H ratio of the water from which the mineral in question was precipitated, provided that no later loss of hydrogen occurred (Kreulen 1987; Mavrogenes and Bodnar 1994). δD values of inclusion waters of calcite veins of the Mátra Mountains selected to represent the ends of the $\delta^{13}C$–$\delta^{18}O$ trend shown in Fig. 2, range from –81 to –47‰ and can be assigned to many kinds of natural waters (meteoric water, magmatic water, formation waters, etc.; see Sheppard 1986). Thus, the interpretation of these data requires the calculation of $\delta^{18}O$ compositions of waters in equilibrium with the calcites (see below).

Carbonates collected in the *Börzsöny Mountains* show a larger $\delta^{13}C$ scatter than the Mátra carbonates, ranging from –11.3 to 4.9‰, but their $\delta^{18}O$ ranges are practically equal (Fig. 2). The presence of a sedimentary component can be assumed in the Börzsöny samples as well but the positive $\delta^{13}C$ data indicate a slightly different sedimentary – maybe evaporitic – environment from which a part of the vein-forming fluids originated. Unlike the Mátra carbonates, $\delta^{13}C$ values down to –11.3 are also found. Such negative $\delta^{13}C$ values can be attributed to two processes: admixing of CO_2 derived from oxidation of organic matter with the circulating fluids (see Ohmoto 1986, for a compilation of literature data), or CO_2 degassing that leads to ^{13}C-depletion in the remaining fraction due to the carbon isotope fractionation between the CO_2 released and the CO_2 dissolved in the fluid (Mook et al. 1974; Lesniak and Sakai 1989). Similarly to the Mátra carbonates, the C and O isotopic compositions of different carbonate species (calcite, siderite, rhodochrosite, and groundmass carbonate of andesite)

Table 1
Stable carbon, oxygen and hydrogen isotope compositions of carbonates and inclusion waters in the Mátra and Börzsöny Mountains. All data are expressed as ‰ relative to PDB ($\delta^{13}C$) and V-SMOW ($\delta^{18}O$ and δD). The samples are calcites except where stated otherwise

samples	$\delta^{13}C$	$\delta^{18}O$	δD
Mátra			
1	-0.5	15.2	-76
2	-0.5	19.3	-81
3	-1.3	15.2	
4	-4.0	16.5	
5	-1.9	15.1	
6 (white)	-1.1	13.7	
6 (black)	-1.4	15.4	
7 (white)	-1.4	18.0	
7 (yellow)	-0.8	16.4	
8	-0.7	16.9	
9	0.1	18.9	
10 (siderite)	-2.6	21.1	
11	-1.6	12.4	
12	0.0	17.0	
13	-2.3	12.3	-47
14	-2.4	10.3	-57
15	-4.5	15.2	-62
16 (siderite)	-0.1	19.0	
17 (andesite)	-5.3	15.5	
18 (andesite)	0.6	22.3	
19 (andesite)	-4.8	11.8	
Börzsöny			
20 (siderite)	-1.5	22.1	
21	-10.1	21.0	-116
22	3.7	20.7	
23	3.0	19.3	-72
24 (siderite)	-5.5	20.0	
25	-10.7	21.3	
26 (rodochrosite)	-4.1	17.0	
27 (dolomite)	4.9	21.9	
28	0.7	16.3	
29	-6.4	14.7	-56
30	-10.3	20.2	-83
31	-5.5	20.8	
32	1.3	28.8	
33 (calcite)	2.1	12.6	
33 (limestone)	-0.7	27.0	
34 (andesite)	-11.3	22.0	
35 (andesite)	-3.5	16.0	



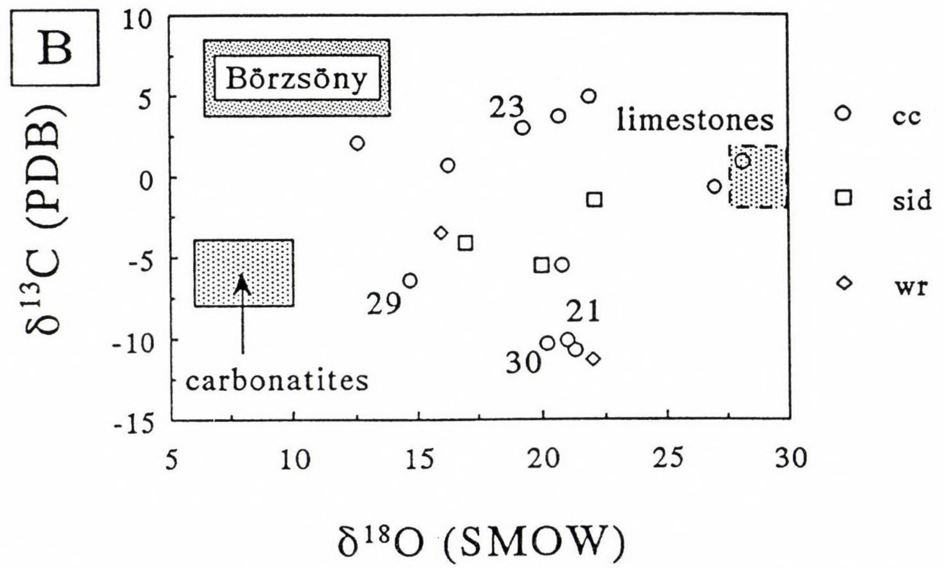


Fig. 2
Stable carbon vs. oxygen isotope compositions (in ‰) of calcites related to the Mátra and the Börzsöny ore deposits. cc: calcite, sid: siderite, wr: whole rock. Compositional fields for carbonatites and limestones are from Taylor et al. (1967) and Hoefs (1973), respectively. Numbers are of samples (see also Table 1)

do not differ from each other indicating a common process that led to carbonate precipitation.

Three of the δD values obtained from inclusion waters of calcite veins selected as previously described overlap the δD range of the Mátra calcites, whereas sample 21 gave an unusually low δD value of –116‰. It is worth noting that this sample belongs to the group characterized by low $\delta^{13}C$ values (see Fig. 2). The interpretation of these δD values will be given later together with discussions of $\delta^{18}O$ values calculated for waters in equilibrium with the calcites.

Plotting together the C and O isotopic compositions of the Mátra and Börzsöny carbonates three main processes can be distinguished (Fig. 3):

1. Mobilization and reprecipitation of sedimentary carbonate $^{13}C/^{12}C$ ratios of which are characteristic for normal marine limestones (I and III) or evaporitic carbonate rocks (II). Depending on fluid compositions and carbonate precipitation temperatures the $\delta^{18}O$ values of carbonates scatter widely (I to III).
2. Presence of magmatic CO_2 (IV).
3. Admixing of ^{13}C-depleted CO_2 derived either from oxidation of organic matter or from a magmatic fluid that had undergone significant CO_2 degassing that brought about a negative $\delta^{13}C$ shift in the remaining fluid fraction (V).

The isotopic compositions of waters from which the carbonates were precipitated can provide essential information on these processes. Thus, $\delta^{18}O$ values of waters in equilibrium with calcites selected for δD determinations were calculated using the equation given by O'Neil et al. (1969). The formation temperatures of the calcite veins of the Mátra Mountains were first estimated at 150–300 °C by Cornides et al. (1966), then at 140 °C by Gatter (1987) based on fluid inclusion microthermometry. Later investigations of fluid inclusion microthermometry of the Mátra Pb–Zn ore deposit suggest, however, higher temperatures up to 200–250 °C (Vető 1988; Gatter 1994, pers. communication). Therefore a precipitation temperature of 200 °C was used in the calculations. Deviations from this temperature by about 50 °C would result in $\delta^{18}O$ uncertainties of about 2–3%, which do not affect the interpretations given below. The formation temperatures of the calcite veins of the Börzsöny Mountains have been estimated at >260 °C by Vető-Ákos (1982); thus the $\delta^{18}O$ values of waters were calculated at 270 °C for the Börzsöny samples. The δD values measured directly and the calculated $\delta^{18}O$ values are plotted in Fig. 4. The fields defined for the fluids of the Mátra and Börzsöny Mountains are slightly different, but the $\delta^{18}O$ differences might arise from the uncertainties of temperature estimations, while most of the dD values fall in the same range.

Based on the position of the compositional field of the Mátra fluids relative to the main fluid types of interest shown in Fig. 4 these fluids might have contained three components: meteoric water, formation water of meteoric origin that had undergone significant O isotope exchange with ^{18}O-rich sedimentary rocks, and magmatic water (that is H_2O in equilibrium with magma at high

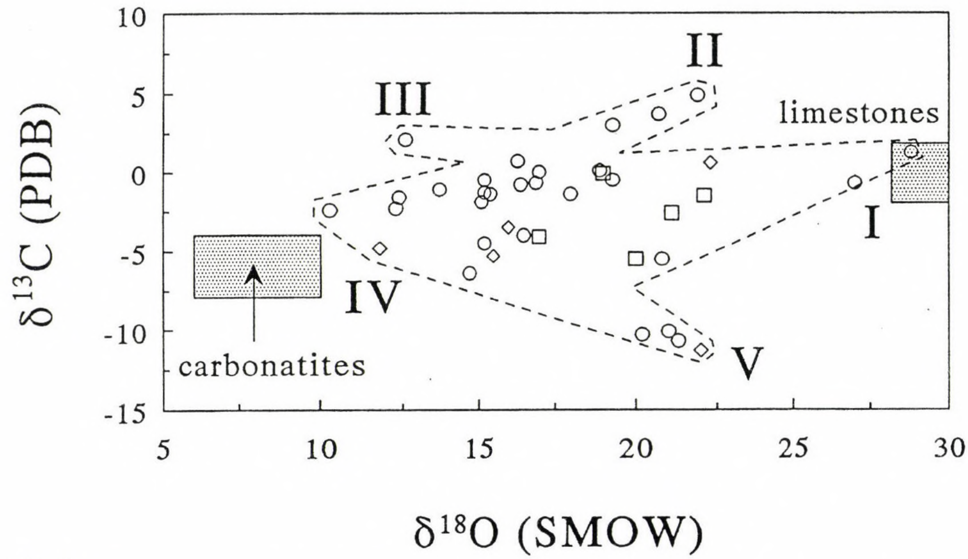


Fig. 3.
A compilation of stable carbon and oxygen isotope compositions (in‰) of calcites related to the Mátra and the Börzsöny ore deposits and possible processes responsible for the isotope distributions. See legend in Fig. 2. I: admixing of sedimentary carbonate; II: admixing of evaporitic sedimentary carbonate; III: hydrothermal mobilization of sedimentary carbonate; IV: admixing of magmatic fluids, V: influence of organic derived CO_2 or degassed magmatic fluid (see also Demény and Harangi 1996)

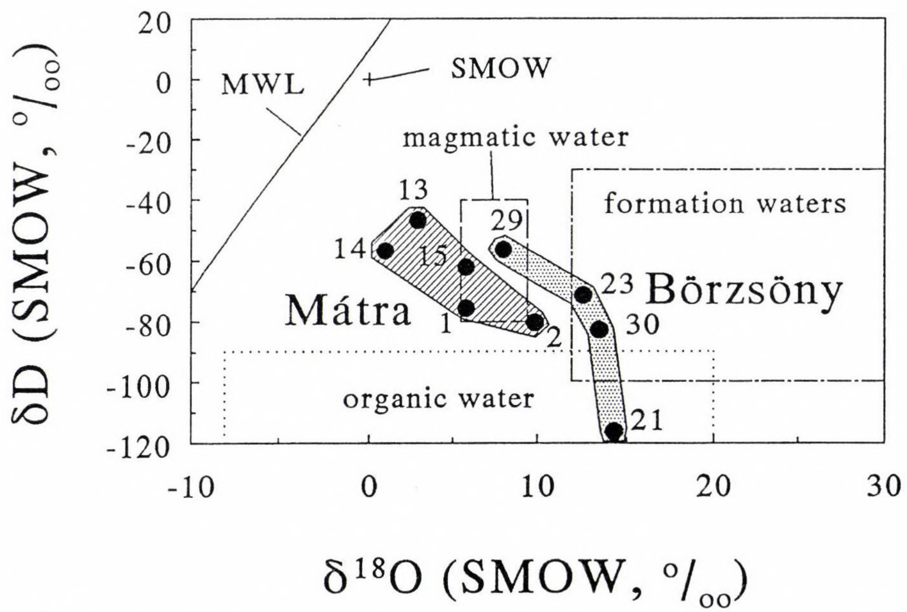


Fig. 4.
δD (measured) and $\delta^{18}O$ (calculated) values of ore-forming fluids in the Mátra and the Börzsöny ore deposits. Numbers mark samples (see Table I). Fields of water types are from Sheppard (1986)

temperatures, Sheppard 1986). Sample 15 which has a magmatic $\delta^{13}C$ value (–4.5 %) plotting in the magmatic water field in the δD–$\delta^{18}O$ diagram suggests the existence of magmatic fluids. The calcite veins that show sedimentary $\delta^{13}C$ signatures (samples 2 and 14) indicate the presence of formation waters (with a meteoric water origin as shown by the δD values) and meteoric waters, respectively. Based on the comparison of these $\delta^{13}C$, δD and $\delta^{18}O$ values, the studied carbonate precipitations of the Mátra Mountains were brought about by mixed fluids that originated from three sources.

Among the calcites from the Börzsöny Mountains, the location of sample 29 which has a magmatic $\delta^{13}C$ value (–6.4‰) plots within the magmatic water field suggesting the presence of magmatic fluid component in the Börzsöny fluids as well. The calcite sample showing sedimentary $\delta^{3}C$ signature (sample 23, 3.0‰) gave δD and $\delta^{18}O$ compositions falling into the field of formation waters, whereas the δD and $\delta^{18}O$ values determined for those veins that are relatively ^{13}C-depleted plot within the fields of formation waters (sample 30) and an unusual type of fluid, organic water (sample 21). The latter fluid type means water formed by oxidation of organic material, that might explain the negative $\delta^{13}C$ value found in this vein and argues against the CO_2 degassing hypothesis.

The similarities in the C, O and H isotope compositions of carbonates and inclusion waters in the Mátra and Börzsöny Mountains suggest that similar processes took part in their formations, namely the mixing of meteoric, formation and magmatic waters. These solutions contained C of sedimentary and magmatic origin due to mobilization of sedimentary carbonate and admixing of ascending magmatic fluids. An additional fluid component derived from oxidation of organic matter appears markedly in the Börzsöny Mountains. This scenario differs from that proposed by Cornides and Kiss (1976) who attributed the negative and positive $\delta^{13}C$–$\delta^{18}O$ correlations they observed in calcite veins of the Mátra Mountains to inequilibrium precipitation of carbonate from a fluid with limited amount of dissolved C and local variations in precipitation temperatures, respectively. The results of combined carbonate-fluid investigations presented in this paper, however, would rather favor the influence of fluid mixing that have also resulted in $\delta^{13}C$–$\delta^{18}O$ correlations.

Such mixed fluids are common in vein-type polymetallic ore formations (Ohmoto 1986) but the existence of organic water in such fluids has not been observed frequently (Sheppard and Charef 1986; Munoz et al. 1994). A sulfur isotope study is being undertaken to determine the effects of these fluids on the compositions of the ore minerals.

Conclusions

Carbon and oxygen isotope compositions of carbonates and hydrogen isotope compositions of inclusion fluids in the Mátra and Börzsöny ore deposits have been determined in order to assess the origin of ore-forming fluids. The $\delta^{13}C$

and $\delta^{18}O$ values of the carbonates of the Mátra Mountains scatter between the ranges of normal marine limestones and magmatic carbonates, whereas the carbonates of the Börzsöny Mountains show a large $\delta^{13}C$ scatter between –11.3 and 4.9‰. The positive $\delta^{13}C$ data might reflect interaction of fluids with evaporitic carbonate rocks, while the carbonates that show ^{13}C-depletion might have been precipitated from fluids that contained organic-derived CO_2 or from fluids that had suffered extensive CO_2 degassing.

D/H ratios measured in inclusion waters trapped in calcite veins and $\delta^{18}O$ values calculated from temperature estimates and $\delta^{18}O$ values of calcites suggest circulation of mixed fluids in both the Mátra and the Börzsöny ore formations. Based on comparisons of C, O and H isotope compositions the fluids responsible for carbonate precipitation in the Mátra Mountains contained three components: meteoric water, formation water of meteoric origin, and magmatic water. Direct admixing of meteoric water appears to be negligible in the fluids of the Börzsöny Mountains, but an additional component derived from oxidation of organic material took also part in the carbonate formation.

Acknowledgement

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Appendix

Localities of the studied samples

Mátra

1. Gyöngyösoroszi, Károly dike, 200 m level.
2. Gyöngyösoroszi, Károly dike, 250 m level.
3. Gyöngyösoroszi, Aranybányabérc, II. dike, 150 m level.
4. Gyöngyösoroszi, Aranybányabérc, 150 m level.
5. Gyöngyösoroszi, Aranybányabérc, 250 m level.
6. Gyöngyösoroszi, Arany Péter dike, 460 m level.
7. Mátraszentimre, dump.
8. Gyöngyösoroszi, Jávoroskút dike, 350 m level.
9. Gyöngyösoroszi, Kiskút dike, 400 m level.
10. Gyöngyöstarján, Hársashegy dike, 4. bore hole, 90–93 m level.
11. Gyöngyöstarján, Hársashegy dike, 4. bore hole, 93–94 m level.
12. Gyöngyöstarján, Hársashegy dike, 4. bore hole, 94–95 m, 2. level.
13. Parádsasvár, Béke shaft, No. 550 dike, lower shaft.
14. Parádsasvár, Béke shaft, No. 550 dike, transit shaft.
15. Parádsasvár, Béke shaft, No. 550 dike.
16. Recsk, Csákánykő quarry.
17. Recsk, Csákánykő quarry.
18. Jobbágyi, Nagyhársas.
19. Nagybátony, shallow level drill hole, Hajnács Hill sill.

Börzsöny

20. Nagybörzsöny, Rózsa shaft.
21. Nagybörzsöny, Rózsa shaft.
22. Nagybörzsöny, lower Rózsa shaft, dump.
23. Nagybörzsöny, lower Rózsa shaft, dump.
24. Nagybörzsöny, lower shaft, dump.
25. Nagybörzsöny, Fagyosasszony shaft.
26. Nagyirtáspuszta, Bezina I shaft.
27. Nagyirtáspuszta, Bezina I shaft.
28. Nagyirtáspuszta, No. 7 bore hole, 1040.6 m.
29. Nagybörzsöny, Rózsabánya, No. 23 borehole, 156.4 m.
30. Nagybörzsöny, Rózsabánya, No. 23 borehole, 192.2 m.
31. Nagybörzsöny, Rózsabánya, No. 23 borehole, 165.5 m.
32. Szendehely, No. 5 bore hole, 45.4–45.6 m.
33. Berkenye, No. 4 bore hole, 308.4 m.
34. Perőcsény, Bedéktűz quarry.
35. Csarnó creek, workers' house.

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Paleobathymetrical analysis of Upper Eocene–Lower Miocene Foraminifera of the Hungarian Paleogene Basin

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The foraminiferal faunas of Upper Eocene–Lower Miocene formations in Hungary can be subdivided into 23 benthic foraminiferal assemblages from marine lagoon to middle bathyal realm. Based on the bathymetrical analysis of benthic foraminifera a sequence of events can be determined in this time span. The main results are as follows:

– there is no significant paleoecological change in the small foraminiferal faunas at the Eocene–Oligocene boundary

– there are two benthic foraminifera blooms in the Oligocene (the first is near the Eocene–Oligocene boundary, the second is in the upper part of Lower Oligocene)

– the high percentage of planktic foraminifera shows extended water stratification during the deposition of the Tard Formation in the Lower Oligocene

– near the Kiscellian/Egerian (Lower/Upper Oligocene) boundary a significant decrease in water depth can be detected based upon the first occurrence of uppermost bathyal and sublittoral foraminifera assemblages

– dominant *Heterolepa–Melonis* assemblages point to deepening and high sedimentation rate in the Upper Oligocene "schlier" basins

– the allochtonous assemblages and specimens indicate intensive tectonic movements during the deposition of Lower Kiscell Clay and Eger Formation.

Key words: foraminiferal assemblages, Upper Eocene to Lower Miocene, paleobathymetry

1. *Introduction*

Our knowledge of the biostratigraphical development of the Hungarian Paleogene Basin derives from investigations of molluscs, nannoplankton and foraminifera assemblages (Báldi 1983; Báldi-Beke 1977; Nagymarosy 1990, 1992; Sztrákos 1974, 1982; Horváth 1980). Detailed investigations of foraminiferal biostratigraphy and facies analysis have also been carried out (Horváth 1980, 1981, 1985). The results of bathymetrical analysis have been used for a long time to describe the subsidence history of the individual basins (Horváth and Nagy-Gellai 1989; Báldi and Nagy-Gellai 1990; Nagymarosy et al. 1995). This paper attempts to define the depositional depth of the typical Upper Eocene–Lower Miocene formations of the Hungarian Paleogene Basin, and to describe the faciological and tectonic events linked to the development of these formations.

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2. *Geologic setting*

The Hungarian Paleogene Basin was a retroarc flexural basin along a SW–NE strike according to Báldi and Báldi-Beke 1985; Báldi 1983, 1986; Tari et al. 1993 (Fig. 1).

The distribution of lithostratigraphic units of the Hungarian Paleogene Basin was determined by tectonic lineaments such as first-level Balaton and Mid-Hungarian Lines and the second-level Buda and Darnó Lines (Nagymarosy 1990). This paper focuses on the NW and SE Buda, the East Cserhát, and the Bükk Paleogene sub-basins (Fig. 2).

Small, deep, elongated basins (mostly transtensional depressions) formed along the tectonic lines (Báldi and Báldi-Beke 1985). The boundaries of the sub-basins can be characterized by sharp facies changes; the sedimentary cycles show rapid subsidence followed by rapid uplift. These different sub-basins developed from Early Lutetian to Early Miocene.

3. *Preparation of samples*

Samples were prepared by mechanical and chemical washing methods. After drying, the samples were weighed (100–500 g), then desegregated in a 10% H_2O_2 solution. They were then wet-sieved with a 63 and a 125 μm-mesh screens (Schrőder et al. 1987). The dried residue was examined for foraminifera.

4. *Paleoecological analysis methodology*

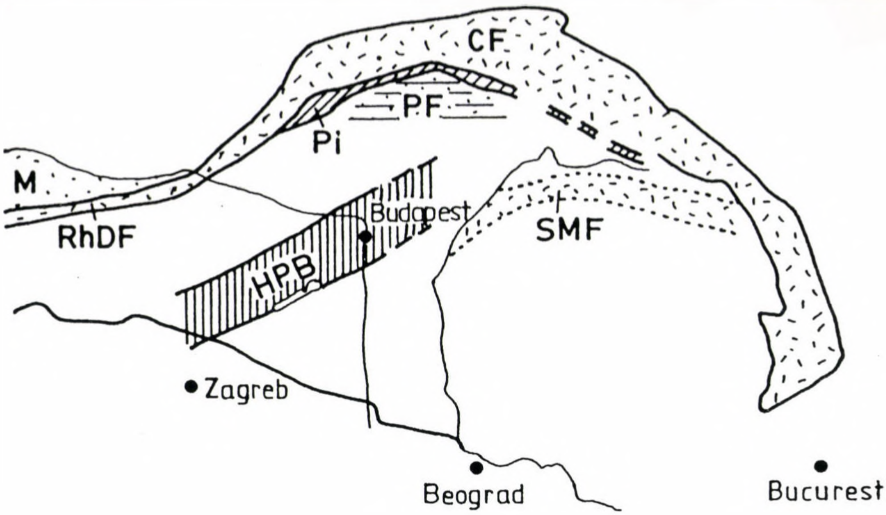
All benthic and planktic foraminifera specimens were separated from the washing residue. The following values were calculated for each sample:

- ratio between planktic and benthic foraminifera
- ratio between agglutinated and calcareous benthic foraminifera
- abundance of foraminifera in total microfauna.

The α index by Murray (1973, 1991) was also calculated and the planktic–benthic index percentage after Wright (1977), modified Van Merle et al. (1987), was used.

Based on the above-mentioned ratios, indexes, and the selected taxa depth ranges were determined for each assemblage. All data on recent distribution of benthic foraminifera were taken from the literature.

In the material studied, the Oligocene assemblages are characterized by long-ranging and bathymetrically widespread taxa. Many of them live in the deep marine environment today. In such causes the estimates of paleobathymetry are based on the analogy of the ecological preferences of recent taxa which have a similar morphology to the extinct ones. This seems to be valid in assemblages not older than Oligocene (Douglas 1979). The Oligocene represents a transition between the typically Paleogene (mainly Paleocene and Eocene) and typically Neogene faunas (Berggren 1984). According to Boltovskoy



M	East-Alpine fore-deep Molasse
RhDF	Rhenodanubian Flysch
CF	Carpathian Flysch
Pi	Pieninides
PF	Podhale Flysch
SMF	Szolnok Flysch
HPB	Hungarian Paleogene Basin

Fig. 1
The general setting of the Hungarian Paleogene Basin in the Alpine–Carpathian System (after Báldi 1986)

(1980a, b) the Oligocene and Quaternary bathyal foraminifera assemblages consist almost entirely of the same species, showing that the Recent benthic fauna developed no later than the Oligocene.

5. *The foraminiferal assemblages*

5.1. *Buda Marl Formation*

The Buda Marl ("Ofener Mergel" in Hofmann 1871; Hantken 1873) is dominantly a light gray, calcareous marl. Allodapic graded limestones with small specimens of *Nummulites fabianii* are interbedded in the lower-middle part of this formation (Varga 1983). The Buda Marl grades upward into the

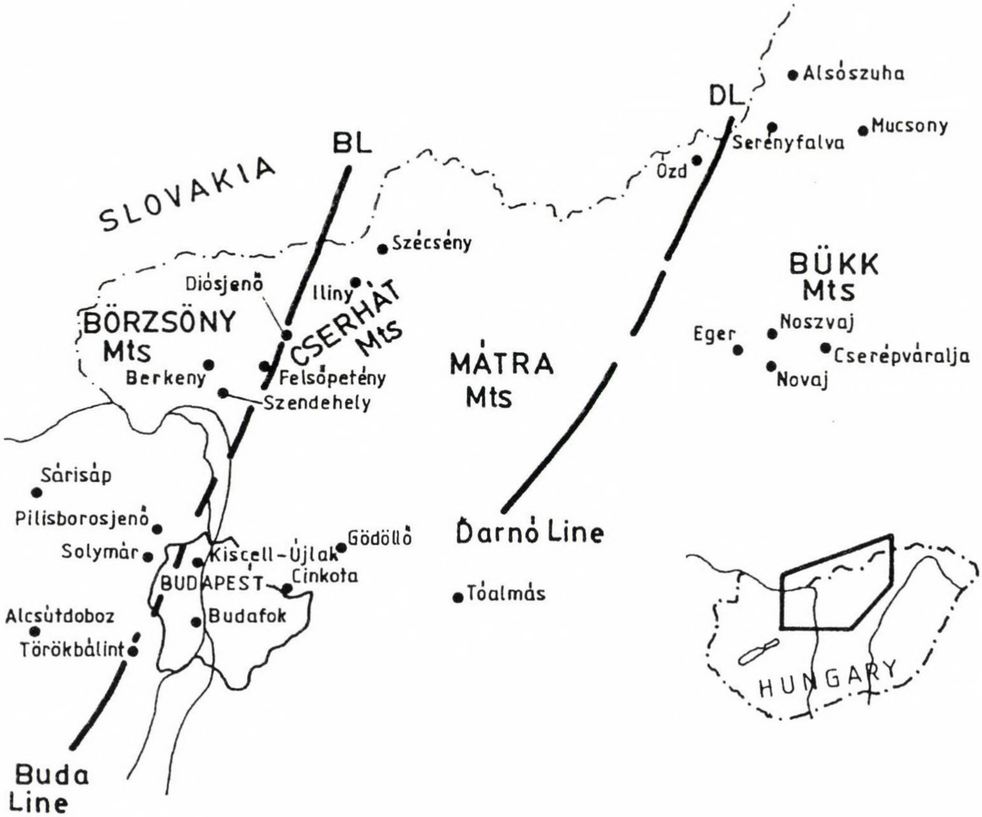


Fig. 2
Topographic locations of geographic names used in the text

Tard Clay Formation without interruption. The Eocene–Oligocene boundary has been fixed within the upper part of the Buda Marl (Báldi et al. 1984) (Fig. 3).

Two types of foraminiferal assemblages can be separated. These are:

5.1.1. *Globigerina–Gemellides assemblage*

This foraminiferal assemblage is very rich both in species and in specimens. The diversity index value is 12–18, indicating a stable open marine environment. The planktic/benthic ratio generally varies between 0.5 and 0.8. Among the benthic forms the calcareous species are more abundant than the agglutinated ones. Frequent agglutinated genera are *Cyclammina, Tritaxia, Vulvulina,* and *Dorothia*. The abundant calcareous benthic taxa are *Gemellides eocaenus, G. costatus, Uvigerina eocaena, U. spinicostata, U. cocoaensis, Gavellinella asterians, Korobkovella grosserugosa, Almaena, Asterigerinata, Queraltina. Hantkeninas* are absent; specimens of the *Turborotalia cerroazulensis*-group are very rare, with

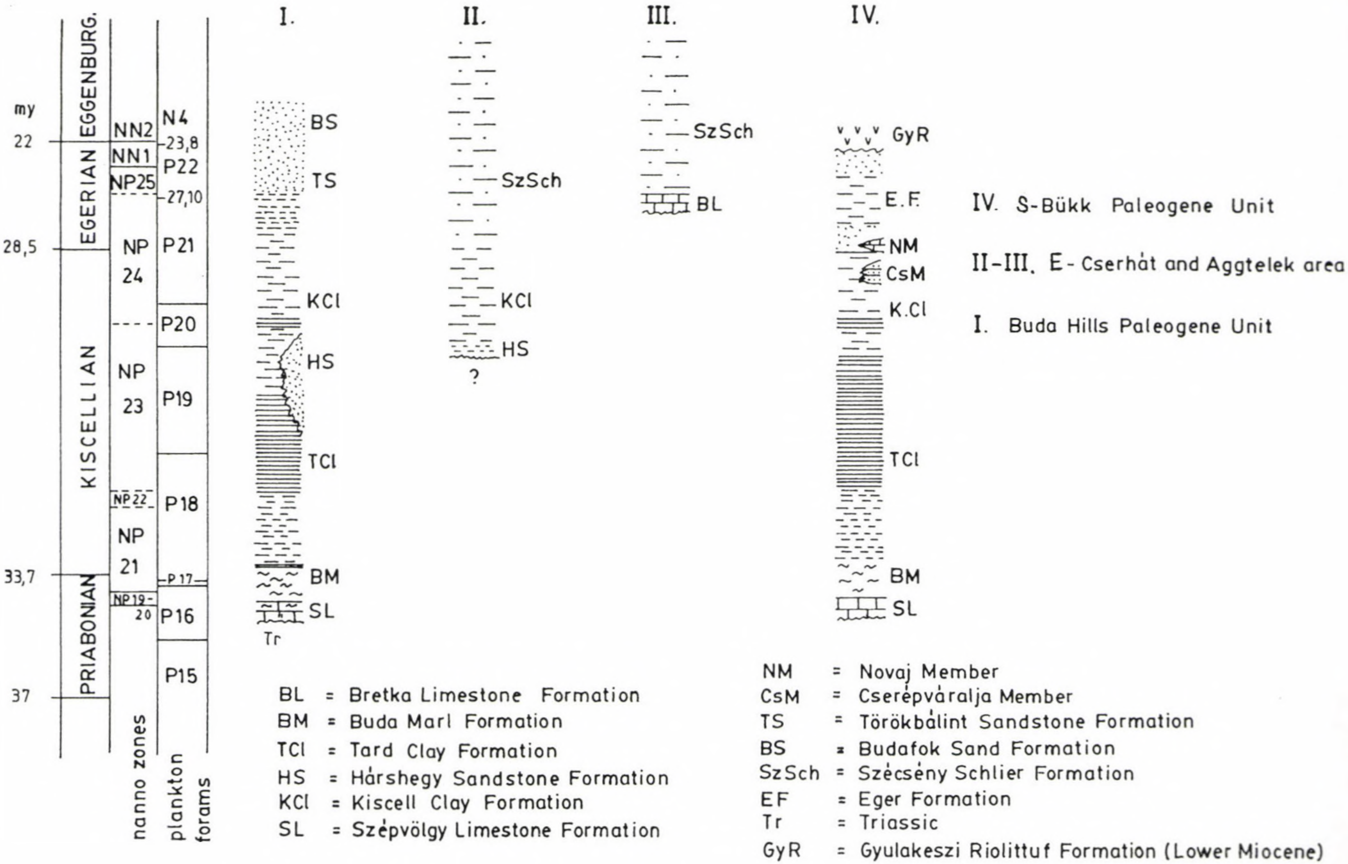


Fig. 3
Stratigraphic columns of the Upper Eocene–Lower Miocene formations in the units of the Hungarian Paleogene Basin. Correlation of biostratigraphy, geochronology and lithostratigraphy (after Nagymarosy 1992)

only one or two specimens found (mainly *T. c. pomeroli* = *T. centralis*). The large planktonic forms (such as *Turborotalia ampliapertura, Subbotina tripartita, S. linaperta, S. eocaena*) and *Turborotalia increbescens* are very frequent. The small *Tenuitellinids* and *Chiloguembelinas* are still rare.

On the basis of the bathymetrical distribution of the above-mentioned taxa the paleobathymetry of the *Globigerina–Gemellides* assemblage can be estimated as follows: minimum depth of 100 m and maximum depth of 600–1000 m. It can be assigned to the middle bathyal domain (after Morkhoven et al. 1986). This estimation corresponds to Berggren's (pers. comm., 1988).

Occurrence: Borehole Kiscell-1 (Fig. 4), Budapest underground construction boreholes, Budapest–Pusztaszeri street, Síkfőkút, Budapest R-6/3. Age: Late Priabonian, nannoplankton zones NP 19–20 (Nagymarosy 1992; Báldi-Beke 1977), plankton foraminifera zone P 16 (Horváth, in prep).

5.1.2. *Globigerina–Bulimina–Bolivina* assemblage

In the upper part of the Buda Marl the foraminifera fauna shows a continuous transition to the *Globigerina–Bulimina–Bolivina* assemblage.

The most abundant taxa are *Bulimina sculptilis, B. alazanensis, B. alsatica, Bolivina aenariensiformis, B. budensis, B. elongata, B. nobilis, B. reticulata, B. semistriata, B. vaceki* s.l., and large globigerinids. *Gemellides costatus, G. eocaenus, Uvigerinids* (with striae and spines, such as *U. eocaena, U. chirana, U. spinicostata, U. rippensis, U. gracilis*), *Chilostomella oolina, Globobulimina ovata, Subbotina eocaena, S. corpulenta, S. angiporoides, T. ampliapertura, D. prasaepis, Gg. officinalis, Gg. praebulloides* are also abundant. The α index value is 10–15. The planktonic foraminifera are more abundant than benthic specimens, the planktic/benthic ratio is varying between 0.4 and 0.7. Small *Tenuitellinids* and others such as *"Gg". postcretacea, Tenuitella gemma, Tenuitellinata angustiumbilicata* (Li Qianyu 1987), *Globanomalina barbadoensis, Chiloguembelina gracillima,* and *Ch. cubensis* become frequent. The agglutinated taxa are less significant; mainly *Cyclammina* and *Karreriella* occur. Miliolina are very rare or absent.

On the basis of the above-mentioned selected taxa and planktic/benthic ratio the estimation for the biotope of the *Globigerina–Bulimina–Bolivina assemblage* is a minimum depth of 100 m (which is the upper limit the entry of the *Bolivina–Bulimina–Uvigerina* fauna) and a maximum depth of 600–1000 m (middle bathyal zone).

Occurrence: Borehole Kiscell-1 (Fig. 4), Budapest-Rózsadomb 6/3, Budapest underground construction boreholes. Age: Uppermost Priabonian–Lowermost Kiscellian, nannoplankton zones NP 21–22 (Nagymarosy 1992; Báldi-Beke 1977), P 17 and the base of P 18 plankton foraminifera zones (Horváth, in prep.).

5.2. *Tard Clay Formation*

The Tard Clay Formation was separated as an independent unit between the Buda Marl and the Kiscell Clay by Majzon (1940, 1942, and 1966). The formation is a dark gray silt with low carbonate and high organic content (Vető 1987). Its thickness is 60–100 m. The deeper part of this formation is slightly laminated, and the upper part consists of an alternation of black and white laminae with monospecific nannofloras in the white laminae (Báldi-Beke 1977; Nagymarosy 1983). Turbiditic intercalations with nummulites occur in the lower parts (Varga 1983). The most common macrofossils of the Tard Clay Formation are plant remains and fish scales. Until the sixties, the Tard Clay was thought to be the so-called "foraminifera-free zone" of Majzon (1966). The investigations of Báldi et al. resulted a relatively rich fauna (molluscs, large sized ostracodes, and foraminifera) and microflora in the formation (Báldi 1984, 1986; Báldi et al. 1984; Nagymarosy 1983, 1992; Monostori 1983, 1986; Horváth 1980, 1983).

5.2.1. Assemblages with small *Tenuitellinids* and *Chiloguembelinas*

In the lower, slightly laminated member the foraminifera assemblages consist mainly of small *Tenuitellinids* and *Chiloguembelinas*.

The percentage of planktic forms in the foraminifera assemblages is between 80 and 100%. *Globigerina officinalis, T. angustiumbilicata, T. gemma, "Gg". postcretacea, Chiloguembelina cubensis, Ch. gracillima,* and more rarely *Globanomalina barbadoensis*, are abundant. Benthic specimens are rare. *Bolivina (B. nobilis, B. elongata, B. beyrichi, B. aenariensiformis), Caucasina oligocaenica, Chilostomella oolina, Allomorphina trigona, Trifarina budensis, Uvigerina gracilis, Asterigerinata falcilocularis, Cibicidoides ungerianus, C. pseudoungerianus* occur sporadically. The diversity value is low indicating a change of the environmental conditions during the deposition of the Lower Tard Clay. Ostracods, pteropods, and fish scales are frequent.

Based upon the low diversity value of the benthic species (*Bolivina, Caucasina, Trifarina,* etc.) the bottom water was oxygen-deficient. The water mass was stratified as is shown by dominating surface dwelling planktic forms and by the absence of typical intermediate and deep planktic species.

Occurrence: Boreholes Kiscell-1 (Fig. 4), Alcsútdoboz-3, Cserépváralja-1 (Fig. 5), Budapest-Rózsadomb 6/3, Budapest underground construction boreholes, Felsőpetény-395. Age: Early Kiscellian, P 18 plankton foraminifera zone.

5.2.2. The upper part of the Tard Clay Formation is dark, microlaminated silt. Fossils are usually absent; only some macroflora elements (leaves, grains) as well as fish remains occur. In this part of the formation the foraminiferal fauna is generally absent. There are a few levels with very special, varying microfaunas such as:

– *Ammomarginulina* assemblages (Alcsútdoboz-3, Csillaghegy and Üröm clay pits). These assemblages consist of *Ammomarginulina, Eratidus, Ammobaculites, Rhabdammina, Trochammina, Reophax, Hyperammina* and more fish remains. Neither calcareous nor planktic specimens are present. The low diversity value indicates a special environment;

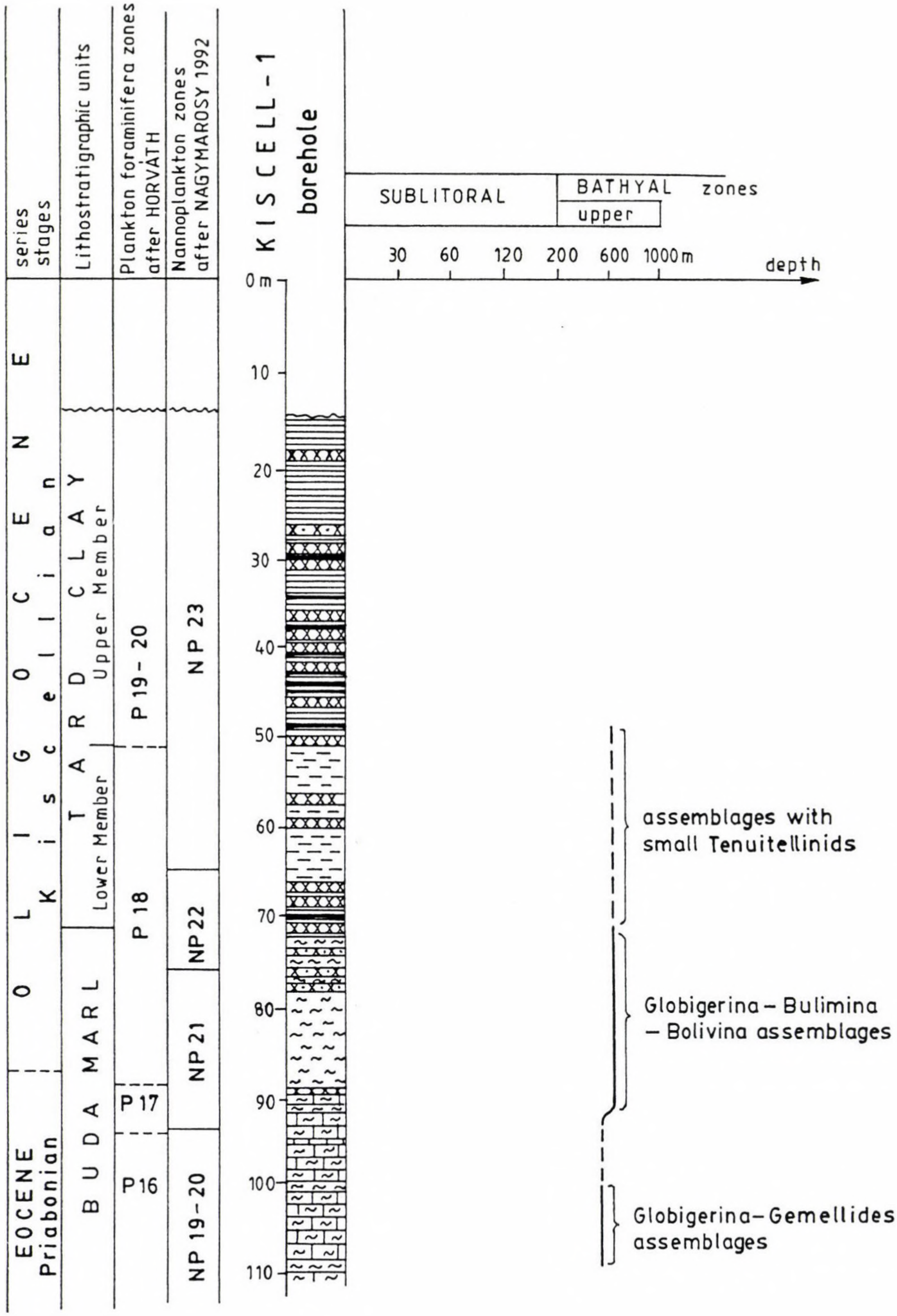


Fig. 4
Profile and foraminiferal assemblages of the Kiscell-1 borehole

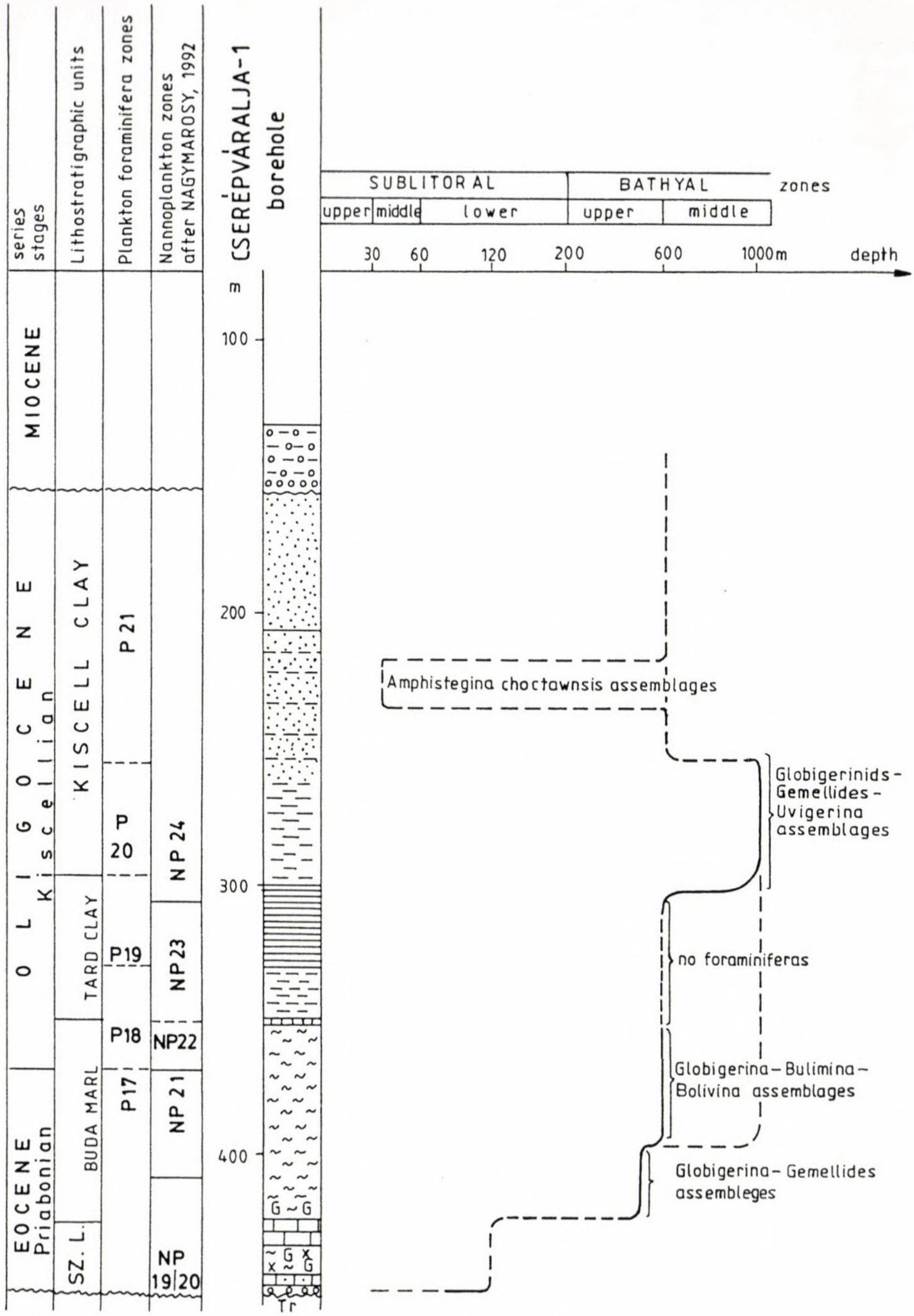


Fig. 5
Profile and foraminiferal assemblages of the Cserépváralja-1 borehole

– monospecific *Globotextularia* (?) assemblages (Budapest underground construction boreholes);

– *Rhabdammina* assemblage (Felsőpetény-361) consisting only of *Rhabdammina, Trochammina, Bigenerina* and *Cyclogyra*.

Similar agglutinated foraminifera assemblages are known in laminites from Jurassic to Recent (Nagy and Løfaldi 1981; Bernhard 1986, Koutsoukos and Hart 1990; Phleger and Soutar 1973).

In the topmost part of the Tard Clay Formation and in the lowermost part of the Kiscell Clay "marine impacts" can be found with monospecific foraminiferal faunas: *Caucasina* assemblages (mainly *C. oligocaenica*), *Uvigerina* assemblages (*U. moravia, U. cocoaensis*), and *Globigerina* assemblages (*Gg. officinalis, Gg. ouachitaensis* group, *Gg. praebulloides* group), and an *Oridorsalis umbonatus* assemblage. Such marine impacts can be observed in the Budapest underground construction boreholes and in the Demjén K-6 borehole as well.

5.3. *Kiscell Clay Formation*

The Kiscell Clay ("Klein-Zeller Tegel" by Peters 1859) consists of calcareous clay and clayey marl. It is non-stratified, bioturbated, with a $CaCO_3$ content between 10 and 35%. K–Ar dating of glauconite from the formation gives an age of 33,5 Ma. (Báldi et al. 1975). The occurrence of the Kiscell Clay in the Buda Mountains was the subject Hantken's classic foraminifera investigations (Hantken 1868, 1875). The nannoflora belongs to the lower part of NP 24 zone (Nagymarosy and Báldi-Beke 1988; Nagymarosy 1992).

5.3.1. *Globigerinids–Gemellides–Uvigerina* assemblage

The lower stratigraphical level can be characterized by *Cassidulina vitálisi* of the *Globigerinids–Gemellides–Uvigerina* assemblage

In this assemblage *Globigerinidae* (mainly *Gg. praebulloides* group, *Gg. officinalis, Gg. ouachitaensis* group, and *S. eocaena, S. angiporoides*), *Gemellides* (mainly *G. costatus* and *G. eocaenus*), *Uvigerina hantkeni*, and *U. gallowayi* are abundant. *Cassidulina vitálisi, Hansenisca soldanii*, and *Nodosariidae* are common. The *Bolivina* and *Trifarina* genera are represented by a few species only. *Fontbotia wuellerstorfi* is sometimes very frequent (50 specimens/sample), e.g. in the Egerszalk-2 and Cserépváralja-1 boreholes). The α index value is generally a very high 16–18 indicating a normal, open marine bathyal environment, as does the percentage of planktics (between 30 and 60%). Calcareous and hyaline forms are dominant. The calcareous/agglutinated ratio is variable depending on the quantity of the sandy sediment influx. The most frequent agglutinated species are *Tritaxia szabói, Triplasia hungarica, Vulvulina haeringensis, Dorothia textilaroides, Cyclammina acutidorsata*, and *Karreriella hantkeniana*.

Occurrence: Boreholes Cserépváralja-1 (Fig. 5), Egerszalók-2, and Cinkota-1, Budapest underground construction boreholes, Budapest-Péterhegy, Felsőpetény-395. Age: Late Kiscellian; lower part of NP 24 nannoplankton zone

(Nagymarosy 1992), and probably the topmost part of the P 20 and the lower part of the P 21 plankton foraminifera zones.

Majzon (1942, 1961, and 1966) divided the Kiscell Clay into four foraminifera coenozones. According to Sztrákos (1975) and Horváth (1980) only the *Cassidulina vitálisi* coenozone can be considered as a stratigraphically permanent level in the lower part of the Kiscell Clay (Fig. 6).

5.3.2. Upper stratigraphical level with *agglutinated foraminifera assemblages*

In these assemblages the diversity is high. *Gemellides costatus, G. eocaenus, Uvigerina hantkeni, U. gallowayi* and the agglutinated taxa are dominant. The agglutinated specimens are large (1–5 mm), as in the case of one assemblage with *Cyclammina acutidorsata, Tritaxia szabói,* and another with *Tritaxias, Karreriella, Dorothia, Haplophragmoides, Gaudryina, Vulvulina, Rhabdammina, Rhizammina, Saccammina, Ammodiscus,* and *Martinottiella.* The agglutinated specimens often amount up to 50% of the total foraminiferal fauna. *Percultazonaria, Lenticulina, Dentalina, Laevidentalina, Stilostomella,* huge *Pyramidulina* (e.g. *P. latejugata*), *Gyroidina, Hansenisca, Korobkovella grosserugosa* and *Neoeponides* are also abundant. The tests of *Gemellides, Praeglobobulimina* and *Pteropods* are sometimes filled up with pyrite. The latter can also be found in form of pyrite molds. Planktic forms occur rarely; sometimes they are even absent. The occurring planktic specimens are large and possess bullae (e.g. different *Catapsydrax*).

Based upon bathymetric data of the foraminifera the minimum depositional depth of the upper member of the Kiscell Clay might have been 200 m and the maximum depth can be estimated at 600–1000 m (the middle bathyal domain).

Occurrence: Boreholes Cinkota-1 and Alcsútdoboz-3, brickyard clay pits in Pilisborosjenő, Ujlak, Törökbálint, Kiscell and Eger-Wind (Fig. 7), borehole Mucsony-136, Novaj-Nyárjas, Budapest underground construction boreholes, Felsőpetény-395. Age: Late Kiscellian; NP 24 nannoplankton zone (Nagymarosy 1992), and probably the P21a plankton foraminifera zone.

5.3.3. Uppermost stratigraphical level with *Spiroloculina–Cibicidoides* assemblages (transition to the Egerian)

The α index value (around 10) corresponds to an average value for the outer shelf seas (Murray 1973). The percentage of Textulariina is 12–15%. Mainly *Textularia gramen, T. pala, T. bronniana, Spirotextularia sagittula, Spiroplectinella carinata* occur. The presence of Miliolina is significant at 20–25%, and *Spiroloculina canaliculata* is more frequent still. Even more abundant are *Quinqueloculina, Triloculina, Pyrgo* and *Sigmoilina. Cibicidoides ungerianus, Hansenisca soldanii,* and *Ceratobulimina contraria* are abundant together with *Hoeglundina eocaenica, Heterolepa dutemplei, Amphicoryna spinicostata, Lenticulina inornata,* and *Hanzawaia americana.* Planktic taxa are rare or absent.

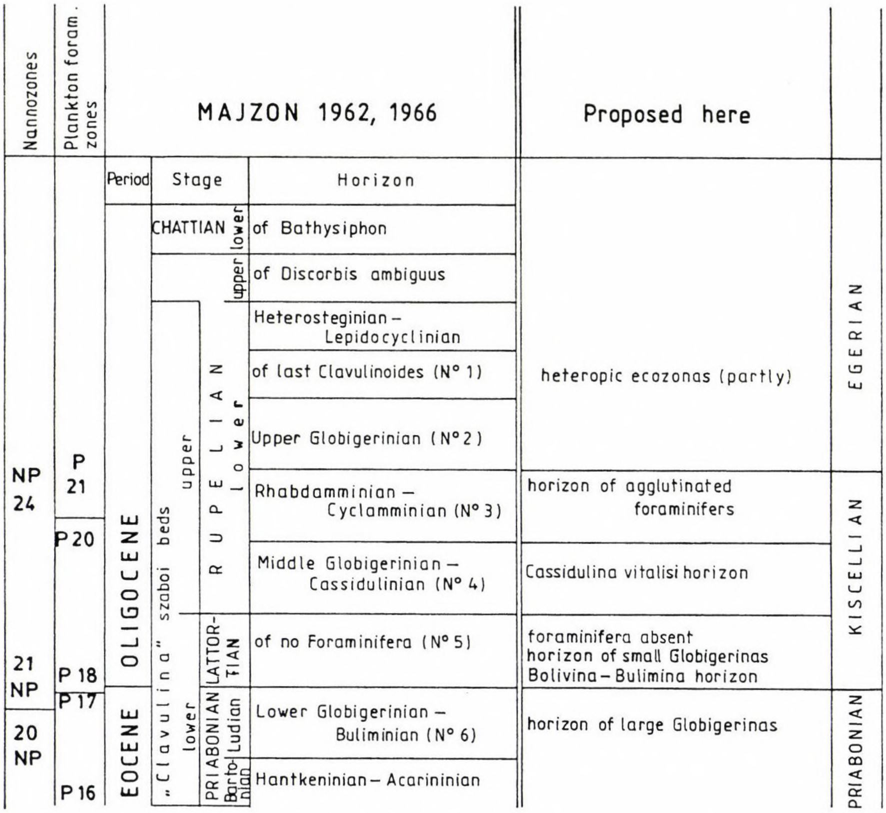


Fig. 6
The Majzon's coenozones in the view of recent biostratigraphical knowledge

The depth of the biotope of the *Spiroloculina–Cibicidoides* assemblage might have been between 100 m as a minimum and 200–250 m as a maximum (in the outer neritic zone). These assemblages can be compared – among others – to the outer shelf biofacies in present-day eastern Indonesia (Van Merle 1988).

Occurrence: Boreholes Budafok-2 (Fig. 8) and Alcsútdoboz-3. Age: Late Kiscellian; upper part of NP 24 nannoplankton zone (Nagymarosy 1992), P 21 plankton foraminifera zone.

5.3.4. Cserépváralja Member of the Kiscell Clay Formation

This member consists of sandstone, rarely pebbly sandstone, intercalated into the Kiscell Clay and is interpreted as a fluxoturbidite (Báldi 1976).

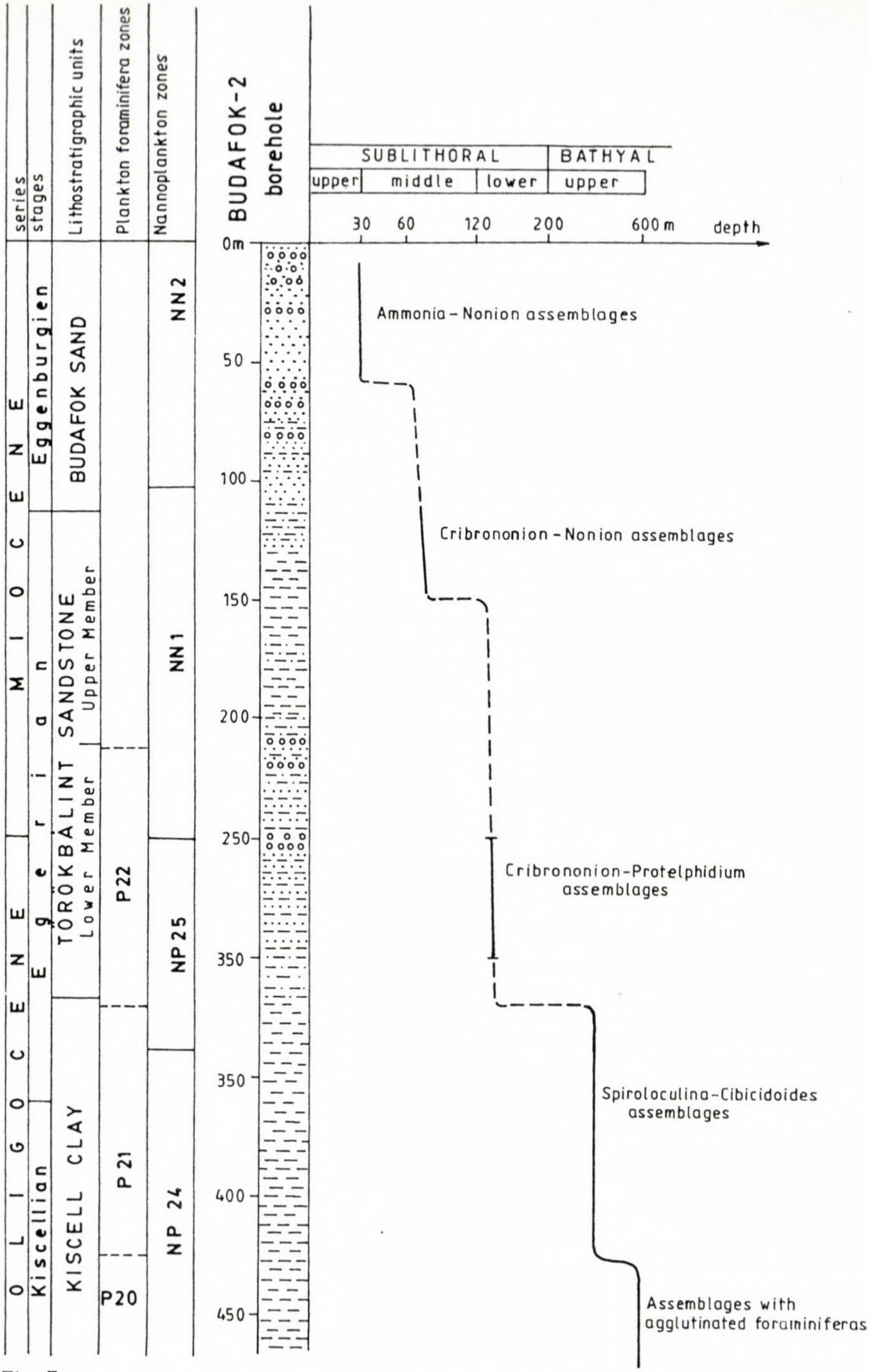


Fig. 7
Profile and foraminiferal assemblages of the Budafok-2 borehole

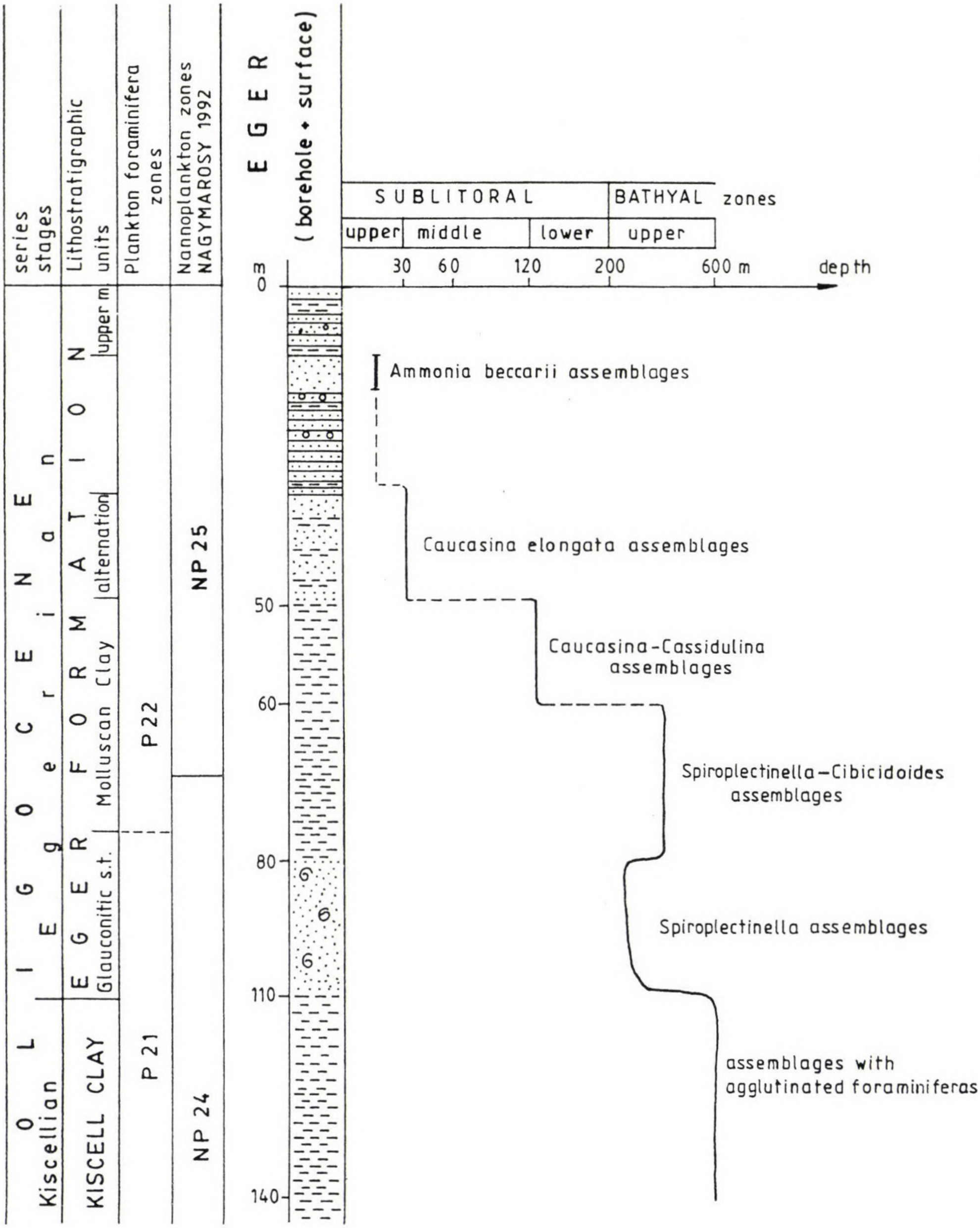


Fig. 8
Profile and foraminiferal assemblages of the Eger type section

Amphistegina choctawensis assemblage

In this rich foraminiferal assemblage the diversity is high, calcareous benthic taxa are dominant, *Miliolina* are absent, and *Textulariina* are rare. The percentage of planktic taxa varies. Dominant species are *Amphistegina choctawensis* and *Asterigerinata falcilocularis. Elphidium crispum, Neoeponides schreibersii, Cibicidoides ungerianus, Lenticulina spp.* are frequent. Species of *Neoconorbina, Almaena* are common. In some samples *Lobatula lobatula, Hanzawaia boueana, Globobulimina ovata* and *Bolivinas* are also common.

The *Amphistegina assemblage* occurs as reworked material mixed into the bathyal *Globigerina–Gemellides–Uvigerina assemblage* of the Kiscell Clay.

Occurrence: Novaj–Nagyimány, borehole Cserépváralja-1 (Fig. 5). Age: Late Kiscellian, NP 24 nannofossil zone (Nagymarosy 1992), P 21 plankton foraminifera zone.

5.4. Hárshegy Sandstone Formation

The Hárshegy Sandstone was deposited in the belt extending west of the Buda Line along the submarine barrier cliff zone, where currents transported clastic sediments from NE to SW (Fodor et al. 1994). Along the Buda Line, the grains of the Hárshegy Sandstone are cemented by allochtonous SiO_2 (Báldi and Nagymarosy 1976). Northwest of the Buda Line (Sárisáp, Alcsútdoboz) the Hárshegy Sandstone is not silicified. It grades upwards without hiatus into the Kiscell Clay Formation. The greatest part of the formation can be grouped into the nannoplankton zone NP 24.

The typical, partly silicified Hárshegy Sandstone contains a foraminiferal fauna in kaolinitic clay interbeddings or fine sandy clay intercalations. These intercalations can be found around Sárisáp and Pilisvörösvár (Horváth in Báldi et al. 1976), west of the Buda Line, where the Hárshegy Sandstone lies transgressively on Triassic Limestone. The foraminifera fauna is represented by several assemblages, such as:

– *Ammonia propinqua – Protelphidium subgranosum* assemblages, in which the diversity index value is very low. *Textulariina* and *Miliolina* are absent. The dominant taxa are *A. propinqua, P. subgranosum* and *Cribrononions*. There are also allochtonous planktic foraminifera (mainly Middle Eocene taxa) as well as ostracods and small gastropods in the fauna. Occurrence: Sárisáp-111.

– *Ammonia propinqua* assemblage, with a diversity value is between 1 and 3 indicating a non-marine environment. *Miliolina* are absent. *Ammonia propinqua* is abundant, *Cribrononion* and *Elphidium* are frequent. Sometimes some *Quinqueloculina* molds can also be found. Occurrence: Sárisáp-111, p-112, p-115, p-121, and p-128 boreholes.

– *Ammomarginulina* assemblages, with a diversity index no higher than 3. *Ammomarginulina, Eratidus, Ammobaculites,* and *Miliammina* are frequent. Occurrence: Sárisáp-112, p-122 boreholes.

The Textulariina–Miliolina–Rotaliina triangular plot after Murray (1973, 1991) indicates hyposaline lagoon or hyposaline shelf environments for these above-mentioned foraminiferal assemblages. Similar recent foraminiferal faunas exist for instance in Blizzard Bay of North America (Murray 1968) and in Kiel Bay in North Europe (Lutze 1965).

These assemblages were first recognized by Nagy-Gellai (1964a, b 1973) and Brestenska and Lehotayova (1960) in the Dorog Basin NW of Budapest. From the Hárshegy Sandstone ostracods were also described, indicating oscillating salinity (Monostori in Báldi et al. 1976).

Age: Late Kiscellian, NP 24 nannoplankton zone (Nagymarosy 1992).

5.5. Törökbálint Sandstone Formation

The Törökbálint Sandstone ("Pectunculus Sande" by Hofmann 1871; Koch 1871) is conformably underlain by the Kiscell Clay and overlain by marine formations of Eggenburgian age. The lower part of the Törökbálint formation is made up of sandy, silty sediments with cross-stratified coarse sand intercalations (T. Makk in Báldi et al. 1973). The middle and upper part of this formation consist of more silty and clayey beds. The nannoplankton (NP 24–25 and NN 1 zones after Báldi-Beke in Báldi and Senes 1975; Nagymarosy and Báldi-Beke 1988), the foraminiferal fauna (Horváth in Báldi et al. 1973; Horváth 1980), and the mollusks (Báldi 1986) indicate a Late Oligocene-Egerian age.

Two types of foraminifera assemblages can be recognized in the Törökbálint Sandstone:

5.5.1. *Cribrononion–Protelphidium assemblage*

The α diversity value varies between 5 and 7. The Miliolina represent 10 to 40% of the fauna, dominated mainly by *Quinqueloculinas*. Planktic foraminifera are rare. The abundant calcareous taxa are *Cribrononion hiltermanni, Cr. minutum, Protelphidium subgranosum, Ammonia propinqua,* and *A. beccarii. Quinqueloculinas* and some species of *Pyrulina, Globulina, Pseudopolymorphina, Sigmomorphina* are common.

East of the Buda Line the ratio of allochtonous elements in the foraminifera faunas is very high (Cretaceous and Eocene planktic forms, such as *Globotruncana, Globigerina, Acarinina, Morozovella,* Horváth in Horváth and T. Makk 1974), but they are absent in the Alcsútdoboz-3 borehole.

The biotope of *Criborononion–Protelphidium* assemblage was the shelf zone, no deeper than 100 m. The minimum depth of this assemblage might have been 0–30 m.

Occurrence: Boreholes Budafok-2 (Fig. 7), Alcsútdoboz-3, and Solymár-72.
Age: Early Egerian; NP 25 nannoplankton zone (Báldi-Beke and Báldi 1974; Nagymarosy 1992).

5.5.2. *Cribrononion–Nonion assemblage*

In this assemblage the percentage of *Textulariina* is variable; *Textularia gramen* occurs most frequently. The calcareous benthic taxa are abundant, mainly *Cribrononion hiltermanni, Cr. minutum, Cribroelphidium dollfusi, Nonion boueana, Protelphidium subgranosum, Pararotalia spp.*, and *Ammonia beccarii*. Planktic forms are rare or absent (Horváth 1980, 1981).

The diversity diagram and the triangular plot together with the recent distribution of the abundant taxa indicate that the biotope of the *Cribrononion–Nonion assemblage* was in the shelf zone. The minimum depth could have been 0 m and the maximum depth between 60 and 100 m, probably 0–30 m on average.

Occurrence: Borehole Budafok-2 (Fig. 7). Age is Egerian, NN 1 nannoplankton zone (Nagymarosy 1992).

5.6. Budafok Sand Formation

The Budafok Sand Formation ("Amonien Sande" by Lőrenthey 1911) consists of cross-stratified, coarse sand with gravelly sand, calcareous clay and sandy silt intercalations. It is rich in the Loibersdorf-type mollusc fauna of Eggenburgian age (Báldi 1973, 1986).

Ammonia-Nonion assemblage

Ammonia beccarii and *Nonion boueana* are abundant. Some *Cribrononions, Elphidiums, Protelphidium, Hanzawaia, Rotalia, Globulina*, and *Almaena* are common. The α index is low: 4–5. The planktic specimens are extremely rare.

The minimum depth of the biotope of this assemblage was 0 m, the maximum depth possibly 30–60 m. This assemblage is analogous to some recent ones which live in sandy, high energy nearshore environments (Murray 1991).

Occurrence: Budafok-2 borehole (Fig. 7). Age: Lower Miocene, Eggenburgian, NN 2 nannofossil zone (Nagymarosy and Báldi-Beke 1988; Nagymarosy 1992).

5.7. Eger Formation

The Eger Formation extends east of the Darnó tectonic belt over the southern foreland of the Bükk Mountains (Eger, Novaj, and Noszvaj) and occurs in the Sajó river valley (Mucsony). It lays conformably over the Kiscell Clay at the southern and eastern margin of the Bükk Mts. It has a varying lithologic composition and can be subdivided into four members in the type section of Eger (Báldi 1973). The Eger Formation belongs to the upper part of NP 24 and the lower part of NP 25 nannozones after Báldi-Beke in Báldi and Senes 1975; B.-Beke and Báldi 1974).

5.7.1. Eger Formation, glauconitic sandstone member

Spiroplectinella assemblage

Spiroplectinella carinata is very abundant, and *Heterolepa dutemplei, Lenticulinas, Planulina costata, Bolivina antiqua, B. semistriata,* and *B. nobilis* are common. *Textularia gramen, Ceratobulimina contraria, Neoeponides screibersii,* some *Cibicidoides,* and *Almaena osnabrugensis* may also occur frequently. *Miogypsinas* can be found in these layers as well (Drooger in Báldi et al. 1961; Papp in Báldi and Senes 1975). The diversity strongly depends on the volume of glauconite. Where the glauconite volume is high, the number of foraminifera is low. The level with the *Spiroplectinella* assemblages corresponds to Majzon's "*Discorbis ambiquus*" horizon or coenozone (1966).

Based on the presence of *Spiroplectinella, Textularia* and *Bolivina,* the *Spiroplectinella* association might have lived between 100 to 600 m, probably between 200–400 m.

Occurrence: Eger–Wind (Fig. 8), Novaj–Nyárjas. Age: Lower Egerian, NP 24 nannofossil zone (Nagymarosy 1992).

5.7.2. Eger Formation, Mollusc-bearing Clay member

Two types of foraminifera assemblages characterize this member of the Eger Formation: the *Spiroplectinella-Heterolepa* and the *Caucasina-Cassidulina* assemblages.

Spiroplectinella–Heterolepa assemblage

Spiroplectinella carinata, Heterolepa dutemplei and *Lenticulina inornata* are abundant and characteristic. *Charltonina budensis, Hansenisca soldanii, Cibicidoides ungerianus,* and a few subspecies of *Almaena osnabrugensis* are frequent. The diversity index value is between 12–16. The amount of Rotaliina is particularly significant; their percentage is 65–87%.

This assemblages might lived at a depth between 0 m (the appearance of *Miliolina*) and 200–250 m (based on *Miliolina* frequency).

Occurrence: Eger–Wind brickyard clay pit (Fig. 8), Novaj–Nyárjas, Mucsony-136. Age: Egerian, the uppermost part of NP 24 and NP 25 nannoplankton zones (Nagymarosy 1992).

Caucasina-Cassidulina assemblage

Caucasina elongata and *Cassidulina crassa* are abundant. *Textularia gramen, Haplophragmoides canariensiformis, Fursenkoina acuta, Nonion boueana,* and *Nonionella libeusi* are frequent. The diversity index value is between 5–10 in average.

The diversity index, the triangular plot and recent data on the distribution of some dominant foraminifera of the *Caucasina elongata–Cassidulina crassa* assemblages suggest that the biotope of these assemblages was on the middle or outer shelf region, between 50 to 100 m depth, at a normal salinity.

Occurrence: Eger–Wind (Fig. 8). Age: Egerian, NP 25 nannoplankton zone (Nagymarosy 1992).

5.7.3. Eger Formation, member with clay and sand alternation

Caucasina elongata assemblage

The assemblage is characterized by *Caucasina elongata* with *Haplophragmoides canariensiformis*, *Textularia gramen* and *Lobatula lobatula*. The low diversity value indicates a change in the normal open marine environment. Based on the high percentage of *Textulariina* and *Haplophragmoides canariensiformis* the biotope of the *Caucasina elongata assemblage* might have been in shallow to middle sublittoral zone, between 0 and 60 m of depth.

The foraminifera assemblages of the Eger Formation indicate continuous decrease of water depth during the deposition of the formation, from the outer shelf–uppermost bathyal biotope of the *Spiroplectinella–Heterolepa* assemblages to the inner shelf biotope of *Caucasina elongata* assemblages.

Occurrence: Eger–Wind brickyard (Fig. 8). Age: Egerian, NP 25 (Nagymarosy 1992).

5.7.4. Eger Formation, so-called "uppermost member" (Báldi 1966)

Ammonia beccarii assemblage

The abundant species is *Ammonia beccarii*, the frequent ones are *A. propinqua* and *Cribrononion minutum*. The diversity index value is very low. *Miliolina* and the planktic specimens are absent in this foraminifera fauna. This assemblage lived in a nearshore environment. As agglutinated taxa, like *Ammobaculites*, *Ammotium*, *Trochammina*, *Ammobaculites*, *Ammoscalaria* are absent, the lagoon environment is not probable (Murray 1973, 1991).

Depth: minimum 0 m, maximum 30 m; probably not deeper than 5–10 m.

Occurrence: Eger–Wind type-section (Fig. 8). Age: Late Egerian.

5.7.5. Eger Formation, Novaj Member

Lithology: diversified. It consists of coarse glauconitic sandstone, Lepidocyclina limestone, Corallinacea limestone, Miogypsina marl, and fine glauconitic-clayey sandstone. It is conformably underlain by the Kiscell Clay but their boundary is very sharp.

In the coarse glauconitic sandstone there is a *Spiroplectinella–Planulina assemblage* which corresponds to the *Spiroplectinella assemblage* in the glauconitic sandstone of the Eger section (Horváth 1985). The *Lepidocyclina* fauna was determined by Drooger (1961, in Báldi et al. 1961) and Kecskeméti (in Báldi et al. 1961), the *Miogypsina, Heterostegina* and *Operculina* were determined by Papp (in Báldi and Senes 1975). In the marl and the fine sandstone the *Amphistegina assemblage* is dominant.

Amphistegina lessonii assemblage

The Rotaliina are dominant in this assemblage (91–93%); the rate of the planktonic specimens is variable. The planktic specimens are large with a thick, porous wall. *Amphistegina lessoni, Reusella, Neoconorbina, Bolivina liebusi, B. antiqua, Asterigerinata, Discorbis, Cassidulina, Trifarina, Elphidium, Neoeponides* and *Rosalina* are abundant. *Miogypsina, Heterostegina* and *Operulina* occur frequently. In this assemblage the rate of the sessile benthic taxa (*Rosalina globularis, Planorbulina mediterranensis, Patellina corrugata, Asterigerinata, Amphistegina, Neoconorbina terquemi, Escornebovina cuvillieri, Cycloloculina annulata,* etc.) is very high.

Minimum depth: 0–30 m. Maximum depth: 100–120 m, probably 30–60 m.

Occurrence: Novaj–Nyárjas, Sály-1. Age: Egerian, the uppermost part of NP 24 and NP 25 nannoplankton zones (B.-Beke and Báldi 1974).

5.7.6. Szécsény Formation

The Szécsény "Schlier" Formation ("Chattischer Schlier" by Noszky 1912; Horusitzky 1940; Schréter 1940) is a non-stratified, monotonous, fine sandy and clayey silt. It is found at the northern margin of the Cserhát, Mátra and Bükk Mountains. Muscovite debris, fish scales, and compressed Bathysiphon tubes occur frequently. Molluscs are common in some layers (*Lentipecten corneum denudatum*; Báldi 1973, 1986). The nannoplankton indicates zone NN 2–3 for the upper part of this formation, and the lower part probably belongs to the NP 24–NN1 zones (Nagymarosy and Báldi-Beke 1988; Nagymarosy 1992). The thickness of the Szécsény Formation may reach 800 meters (Ózd, Alsószuha). In the northeast, the formation lies transgressively over Triassic and Devonian limestone. The transgressive basal beds are the Bretka Limestone at Imola or the glauconitic sandstone with *Miogypsina formosensis* (Putnok).

In the west (in the Cserhát hills and Mátra Mountains) the Szécsény Formation lies conformably over the Kiscell Clay (borehole Szécsény-1) and is covered by the Lower Miocene Budafok Sand or Pétervására Sandstone (Fig. 3).

Heterolepa dutemplei–Melonis pompilioides assemblage

The foraminifera fauna of the Szécsény "Schlier" Formation is very variable (Horváth 1972). The diversity index value is generally low (between 5–10); planktic forms are rare or absent. In this assemblage *Bathysiphon, Martinottiella, Cyclogyra, Cyclammina, Pullenia sphaeroides, Sphaeroidina bulloides, Marginulina behmi, Heterolepa,* and *Lenticulina* are frequent.

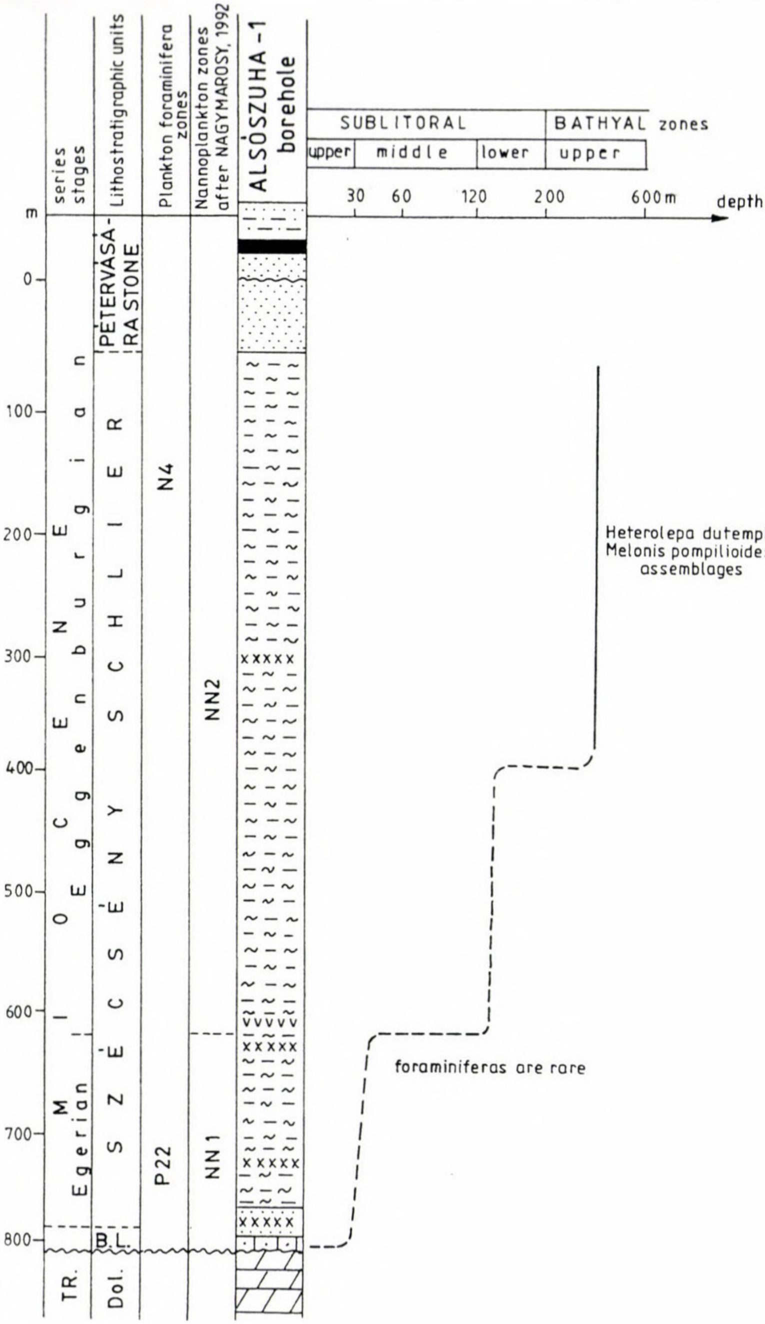


Fig. 9
Profile and foraminiferal assemblages of the Alsószuha-1 borehole

Occurrence: Boreholes Szécsény-1, Szécsény-Öreghegy, Iliny-1, Ózd-2, Putnok-brickyard clay pit, Serényfalva-2, Csízfürdő-2, Alsószuha-1 (Fig. 9). Age: Uppermost Egerian–Eggenburgian.

6. *Stratigraphical and paleoecological changes in the Oligocene foraminifera assemblages*

On the basis of the occurrence and distribution of foraminifera assemblages the following sequence of events can be observed in North Hungary for the Upper Eocene–Lower Miocene time span:

1. there is no dramatic change in the small foraminifera fauna at the Eocene–Oligocene boundary; some tropical, warm-temperate taxa disappear however, such as *Queraltina, Almaena taurica,* and *Subbotina linaperta;*

2. appearance of frequent Bulimina–Bolivina indicates the beginning of the paleoecological change near the Eocene–Oligocene boundary. This change might have be a multi-component one:

– change in the O-content. Oxygen-deficient bottom water began to develop. The *Bolivinidae* prefer an environment with low oxygen content, therefore their bloom can be related to the beginning of the isolation of the basin, the decrease of water circulation, the extension of the oxygen minimum zone (Nocchi et al. 1988). At the time of the *Bulimina–Bolivina bloom* planktic fauna was also abundant, thus indicating that the oxygen-deficient zone had not reached the surface of the seawater. The *Bulimina–Bolivina acme* is in agreement with the beginning of the isolation of the Paratethys – which took place at 36–35 Ma. according to Báldi (1980);

– decrease of the bottom water temperature. Shackleton (1986) showed that around 35 Ma. the temperature of the oceanic bottom water suddenly decreased several degrees centigrade during 10,000–100,000 years. This intensive cooling can also be observed in epicratonic seas, because the cool-temperature planktonic forms (small *Tenuitellinids* and *Globigerinids*) became very frequent.

3. Planktic foraminifera show that water stratification was very significant at the time of the deposition of the lower part of the Tard Clay (NP 23 zone). Two types of the planktic assemblages can be distinguished:

– planktic fauna consisting of "normal" forms with average size, strongly porous tests, such as *S. angiporoides, D. galavisi, Gg. officinalis,* and *Gg. praebulloides.* These were the dwellers of intermediate water depth;

– the "small"-sized forms, smooth, with finely porous tests, mainly *G. liverovskae, "Gg". postcretacea, Tenuitella gemma, Tenuitellinata angustumbilicata, Chiloguembelina gracillima,* and *Ch. cubensis;* rarely with *Globanomalina barbadoensis.* This planktic foraminifera level is roughly coeval with the *Cardium lipoldi–Ergenica cimlanica* mollusc zone (Báldi 1980, 1984) and of the NP 22–23 nannoplankton zones (acme of *Reticulofenestra ornata* and *Lanternitus minutus;* Nagymarosy 1992).

4. In the upper, laminitic part of the Tard Clay (NP 23 zone) the *Ammomarginulina* level and the general absence of foraminiferal fauna indicate that

– the shallow water was brackish

– an oxygen-deficient water-layer existed at the bottom of the sea and significant water-mass stratification developed.

5. The *Uvigerina* bloom at the base of the Kiscell Clay. This bloom indicates that the bottom water of the sea was slightly dysaerobic and the O_2–H_2S interface was the near the sediment surface (the base of NP 24).

6. The appearance of the abundant planktic and benthic foraminifera (*Globigerina–Gemellides–Uvigerina assemblage*) after the *Uvigerina bloom* indicates that the water mass became rich in oxygen. The O_2–H_2S interface remained within the sediment.

7. Occurrence of *Caucasina*-rich levels in the zone of the laminitic intercalations of the Kiscell Clay (within the NP 24 nannofossil zone) which can indicate transportation of foraminifera tests from the shallow shelf region.

8. At the boundary of the Kiscellian/Egerian (NP 24 zone) a significant decrease of water depth can be detected, indicated by

– the appearance of the *Spiroloculina canaliculata* assemblage in the non-typical Kiscell Clay. The biotope of these assemblages could not have been deeper than the middle or lower sublittoral (outer neritic) zone, in contrast to the upper and middle bathyal assemblages of the typical Kiscell Clay;

– the appearance of monospecific neritic assemblages (*Sphaeroidina bulloides, Caucasina elongata,* and *Asterigerinoides gürichi* assemblages) at the lower part of the Törökbálint Sandstone;

– the appearance of the *Spiroplectinella* assemblages (lower sublittoral or outer shelf environment) in the glauconitic sandstones of the Eger Formation; appearance of large foraminifera (*Lepidocyclinids*) and thin limestone intercalations with calcareous algae (Kriván-Hutter 1961) in the Novaj Member.

9. The deepening and the rate of the sedimentation of the Upper Oligocene schlier basins (Serényfalva, Alsószuha) was very rapid (NN1–3 zones) based on the very quick change between the Miogypsina and the *Heterolepa–Melonis* assemblages.

7. Conclusions

The results of the paleoecological study have shown, that:

– the Hungarian Paleogene Basin can be characterized by bathyal conditions from the Late Priabonian to the beginning of the Egerian. The basin reached upper and middle bathyal depths (600–1000 m) during the deposition of the upper part of the Buda Marl (*Globigerina–Bulimina–Bolivina assemblages*) and during the deposition of the Kiscell Clay (*Globigerina–Uvigerina–Gemellides* and *agglutinated assemblages*);

– during the isolation of the Eoparatethys (Early Oligocene) and the deposition of the Tard Clay the water depth could have not decreased greatly. It is at this time that water-mass stratification must have begun (based upon the simultaneous occurrence of the "normal"- and "small"-sized planktic assemblages, and special benthic assemblages with *Bolivina*) in the lower part of the Tard Clay);

– the *Ammomarginulina* assemblages of the upper Tard Clay can be correlated with the *Ammonia*- and *Ammomarginulina* assemblages of the Hárshegy Sandstone. These assemblages indicate the same level in time;

– the so-called Kiscell-type foraminifera fauna with very high diversity indicates a water depth of 500 to 1000 m (upper and middle bathyal zone) during the deposition of the Kiscell Clay;

– rare laminite beds with *Caucasina* associations indicate that an oxygen-deficient environment developed periodically;

– near the Kiscellian/Egerian boundary (in the "atypical" Kiscell Clay and at the boundary of the Kiscell and Eger Formations) the sea level fall can be estimated between 100 and 300 m. The bathyal foraminifera assemblages are followed here by uppermost bathyal and neritic foraminifera assemblages in the region of the deposition of the Törökbálint Sandstone and the Eger Formation (in the Buda and the Bükk Mountains). In these regions the decrease of the water depth continued in the upper part of the Egerian and in the Eggenburgian (*Cribrononion–Nonion* and *Ammonia* assemblages in the Törökbálint Sandstone and the Budafok Sand);

– at the same time a new basin formed in North Hungary (at Serényfalva, Alsószuha) in which Szécsény Schlier was deposited. The foraminifera fauna indicates rapid deepening of these basins;

– the allochthonous specimens and assemblages in Oligocene beds, such as allodapic limestone layers in the Buda Marl and Tard Clay (Varga 1982, 1983), Eocene planktic and benthic specimens in the Tard Clay (Sztrákos 1974), Cretaceous and Eocene planktic specimens at the lower part of the Törökbálint Sandstone (Horváth and T-Makk 1974, Horváth 1981), and *Amphistegina* assemblages (Cserépváralja Member) in the Kiscell Clay around Eger indicate intensive reworking, perhaps due to tectonic movements.

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Taxonomy

Order: *FORAMINIFERIDA* Eichwald, 1830
Suborder: Textulariina Delage and Herouard, 1896
Bathysiphon eocaenicus Cushman and Hanna, 1927 (fide Ellis and Messina Catalogue)
Bathysiphon filiformis M.Sars, 1870 (p. 251)
Bathysiphon taurinense Sacco, 1893 (p. 166, textfig. 2)
Rhabdammina eocanica Cushman and Hanna, 1927 (fide Ellis and Messina Catalogue)
Rhizammina sp.
Saccammina sphaerica Brady, 1871 (fide Ellis and Messina Catalogue)
Saccamina variabilis Bogdanovicz, 1960 (p. 260, pl. 1, figs 4–5)
Hyperammina nodata Grzybowski, 1896 (p. 275, pl. 8, fig. 16)
Hyperammina sp.
Ammodiscus incertus (d'Orbigny, 1839 – Operculina incerta, p. 49, pl. 6, figs 16–17)
Ammodiscus grzybowskii Emiliani, 1954 (p. 106, pl. 21, fig. 3)
Ammodiscus miocaenicus Karrer, 1877 (p. 372)
Glomospira charoides (Parker and Jones, 1860 – Trochammina squammata var. charoides, p. 304) (pl. I, fig. 4)
Glomospira gordialis (Parker and Jones, 1860 – Trochammina squammata var. gordialis, p. 304)
Miliammina sp.
Reophax grandis Grzybowski, 1898 (fide Ellis and Messina Catalogue)
Reophax pilulifera Brady, 1884 (p. 292, pl. 30, figs 18–20)
Haplophragmoides canariensiformis Sztrákos, 1979 (p. 57, pl. 2, figs 2–3, pl. 33, figs 1–2)

Haplophragmoides fragile Hoeglund, 1947 (p. 137, pl. 10, figs 115–116)
Haplophragmoides deforme (Adreae, 1884 – Haplophragmium deforme p. 197, pl. 8, fig. 1)
Haplophragmoides latidorsatus (Bornemann, 1855 – Nonionides latidorsata p. 339, pl. 16, fig. 4)
Haplophragmoides obliquecameratus Marks, 1951 (p. 35, pl. 5, fig. 1)
Haplophragmodes quinquelocularis Subbotina, 1960 (p. 182, pl. 2, figs 1–2)
Haplophragmoides vasiceki vasiceki Cicha and Zapletalova, 1963 (p. 79, text.fig. 1)
Trochamminoides sp.
Ammobaculites sp.
Ammomarginulina expansa (Plummer, 1933 – Ammobaculites expansus, fide Ellis and Messina Catalogue)
Eratidus foliaceus (Brady, 1884 – Haplophragmium foliaceum, p. 250, pl. 33, figs 23–25)
Triplasia hungarica (Majzon, 1948 – Centenaria hungarica, p. 24, text.fig. 1)
Budashevaella wilsoni (Smith, 1948 – Haplophragmoides wilsoni, p. 49, pl. 9, figs 11–12)
Alveolophragmium sp.
Reticulophragmium venezuelanum (Maync, 1955 – Alveolophragmium venezuelanum, p. 141, pl. 26, text.figs 1–8)
Sabellovoluta humboldti (Reuss, 1851 – Spirolina humboldti, p. 65, pl. 3, figs 17–18)
Cyclammina acutidorsata (Hantken, 1868 – Haplophragmium acutidorsatum, p. 82, pl. 1, fig. 1)
Cyclammina amplcetens Grzybowski, 1897 (p. 292, pl. 12, figs 1–3)
Cyclammina karpatica Cicha and Zapletalova, 1963 (p. 108, text. fig. 17)
Cyclammina placenta (Reuss, 1851 – Nonionina placenta, p. 72, pl. 5, fig. 33)
Cyclammina rotundidorsata (Hantken, 1875 – Haplophragmium rotundidorsatum, p. 12, pl. 1, fig. 2)
Ammobaculoides agglutinans (d'Orbigny, 1846 – Spirolina agglutinans, p. 137, pl. 7, figs 10–12)
Spiroplectammina agglutinans (d'Orbigny, 1839 – Textularia agglutinans, p. 144, pl. 1, figs 17–18)
Spiroplectammina deperdita (d'Orbigny, 1846 – Textularia deperdita, p. 244, pl. 14, figs 23–25)
Spiroplectinella carinata (d'Orbigny, 1846 – Textularia carinata, p. 247, pl. 14, figs 32–34)
Spiroplectinella pecinata (Reuss, 1850 – Textularia pectinata, p. 381, pl. 49, figs 2–3)
Vulvulina haeringensis (Gümbel, 1868 – Venilina haeringensis, p. 648, pl. 2, fig. 84. = Textilaria subflabelliformis Hantken, 1875)
Vulvulina pectinata Hantken, 1875 (p. 68, pl. 7, fig. 10)
Spirotextularia sagittula (Defrance, 1824 – Textularia sagittula, p. 177, pl. 13, fig. 5)
Ammoglobigerina globigeriniformis Parker and Jones, 1865 (fide Ellis and Messina Catalogue)
Trochammina nobensis Asano, 1951 (p. 8, figs 3–4)
Gaudryina difformis (Halkyard, 1918 – Gaudryina rugosa d'Orbigny, var. difformis, p. 42, pl. 2, figs 7–9)
Gaudryina fortiuscula Bermúdez, 1949 (p. 74, pl. 3, figs 57–58)
Gaudryina rugosa d'Orbigny, 1840 (p. 44, pl. 4, figs 20–21)
Tritaxia haeringensis (Cushman, 1936 – Clavulinoides haeringensis, p. 22, pl. 3, fig. 17)
Tritaxia havanensis (Cushman and Bermúdez, 1937 – Clavulinoides havanensis, p. 3, pl. 1, figs 12–13)
Tritaxia szabói (Hantken, 1868 – Clavulina szabói, p. 83, pl. 1, figs 4, 6–7)
Globotextularia sp.
Dorothia textilaroides (Hantken, 1875 – Gaudryina textilaroides, p. 15, pl. 1, fig. 6)
Eggerella irregularis (Hantken, 1875 – Gaudryina irregularis, p. ? pl. ? fig. ?
Karreriella chilostoma (Reuss, 1852 – Textilaria chilostoma, p. 18)
Karreriella hantkeniana Cushman, 1936 (p. 36, pl. 5, fig. 19)
Karreriella siphonella (Reuss, 1851 – Gaudryina siphonella, p. 78, pl. 5, figs 40–42)
Martinottiella communis (d'Orbigny, 1826 – Clavulina communis, p. 268)
Martinottiella rhumbleri (Cushman, 1936 – Listerella rhumbleri, p. 38, pl. 6, fig. 4)
Bigenerina nodosaria d'Orbigny, 1826 (p. 261)
Sahulia trochus (d'Orbigny, 1840 – Textularia trochus, p. 45, pl. 4, figs 25–26)
Sahulia turris (d'Orbigny, 1840 – Textularia turris, p. 46, pl. 4, figs. 27–28)
Textularia bronniana d'Orbigny, 1846 (p. 244, pl. 14, figs 20–22)
Textularia elongata (Hantken, 1875 – Textilaria elongata, p. 67, pl. 15, fig. 3)
Textularia gramen d'Orbigny, 1846 (p. 248, pl. 15, figs 4–6)

Textularia pala Czjzek, 1848 (p. 148, pl. 13, figs 25–27)
Textularia speyeri (Reuss, 1865 – Plecanium speyeri, p. 449, pl. 1, fig. 3)
Siphotextularia concava (Karrer, 1868 – Plecanium concavum, p. 129, pl. 1, fig. 3)
Planctostoma eocaena Sztrákos, 1979 (p. 60, pl. 4, figs 6–7, pl. 33, fig. 5)
Planctostoma oligocaena Sztrákos, 1979 (p. 61, pl. 4, figs 8–10, pl. 33, figs 6–7)
Cylindroclavulina rudislosta (Hantken, 1875 – Clavulina cylindrica, p. 18, pl. 1, fig. 8)
Tritaxilina reussi (Hantken, 1868 – Gaudryina reussi, p. 83, pl. 1, fig. 2)

Suborder: Spirillinina Hohenegger and Piller, 1975
Spirillina sp.
Patellina corrugata Williamson, 1858 (p. 46, pl. 3, figs. 86–89)
Patellina sp.

Suborder: Miliolina Delage and Hérouard, 1896
Cornuspira involens (Reuss, 1850 – Operculina involens, p. 370, pl. 46, fig. 20)
Cornuspira olygyra Hantken, 1875 (p. 20, pl. 1, fig. 10)
Spiroloculina canaliculata d'Orbigny, 1846 (p. 269, pl. 16, figs 10–12)
Spiroloculina dorsata Reuss, 1875 (p. 123, = Spiroloculna limbata, Bornemann, 1855)
Spiroloculina tenuissima Reuss, 1867 (p. 55, pl. 1, fig. 11)
Quinqueloculina akneriana d'Orbigny, 1846 (p. 290, pl. 18, figs 16–21)
Quinqueloculina badenensis d'Orbigny, 1846 (p. 290, pl. 20, figs 10–12)
Quinqueloculina contorta d'Orbigny, 1846 (p. 298, pl. 20, figs 4–6)
Quinqueloculina ermanni Bornemann, 1855 (p. 351, pl. 19, fig. 6)
Quinqueloculina hauerina d'Orbigny, 1846 (p. 286, pl. 17, figs 25–27)
Quinqueloculina impressa Reuss, 1851 (p. 87, pl. 7, fig. 59)
Quinqueloculina juleana d'Orbigny, 1846 (p. 298, pl. 20, figs 1–3)
Quinqueloculina longirostra d'Orbigny, 1826 (p. 137)
Quinqueloculina ludwigi Reuss, 1865 (p. 126, pl. 1, fig. 12)
Quinqueloculina mayeriana d'Orbigny, 1846 (p. 287, pl. 18, figs 1–3)
Quinqueloculina seminula (Linné, 1758 – Serpula seminulum, p. 786)
Pyrgo bulloides (d'Orbigny, 1826 – Biloculina bulloides, p. 131)
Pyrgo simplex (d'Orbigny, 1846 – Biloculina simplex, p. 264, pl. 15, figs 25–27)
Triloculina consobrina d'Orbigny, 1846 (p. 277, pl. 17, figs 10–12)
Triloculina gibba d'Orbigny, 1826 (p. 299)
Triloculina oblonga (Montagu, 1803 – Vermiculum oblongum, fide Ellis and Messina Catalogue)
Triloculina tricarinata d'Orbigny, 1826 (p. 133)
Triloculina trigonula (Lamarck, 1804 – Miliolites trigonula, fide Ellis and Messina Catalogue)
Triloculina valvularis Reuss, 1851 (p. 85, pl. 7, fig. 58)
Sigmoilina celata (Costa, 1850 – Spiroloculina celata, p. 126, pl. 1, fig. 14)
Sigmoilinita tenuis (Czjzek, 1848 – Quinqueloculina tenuis, p. 149, pl. 13, figs 31–34)
Articulina sulcata Reuss, 1850 (p. 383, pl. 49, figs 13–17)

Suborder: Lagenina Delage et Hérouard, 1896
Chrysalogonium longiscatum (d'Orbigny, 1846 – Nodosaria longiscata, p. 32, pl. 1, figs 10–12)
Dentalina acuminata (Hantken, 1875 – Nodosaria acuminata, p. 32)
Dentalina capitata (Boll, 1846 – Nodosaria capitata, fide Ellis and Messina Catalogue)
Dentalina fissicostata Gümbel, 1870 (p. 48, pl. 1, fig. 46)
Dentalina semilaevis Hantken, 1875 (p. 32, pl. 12, fig. 3)
Grigelis pyrulus (d'Orbigny, 1826 – Nodosaria pyrula, p. 253 = Nodosaria stipitata Reuss, 1870)
Laevidentalina elegans (d'Orbigny, 1846 – Dentalina elegans, p. pl. 1, figs 52–56)
Laevidentalina intermedia (Hantken, 1875 – Dentalina intermedia, p. 25, pl. 3, figs 4, 8)
Laevidentlian sp.
Nodosaria soluta (Reuss, 1851 – Dentalina soluta, p. 64, pl. 3, fig. 4)

Pseudonodosaria acuta (LeRoy, 1941 – Pseudoglandulina acuta, p. 27, pl. 2, figs 1–2)
Pseudonodosaria aequalis (Reuss, 1863 – Glandulina aequalis, p. 48, pl. 3, fig. 28)
Pseudonodosaria disrecta (Reuss, 1850 – Glandulina disrecta, p. 366, pl. 40, fig. 3)
Pyramidulina bactridium (Reuss, 1865 – Nodosaria bactridium, p. 130, pl. 1, figs 24–25)
Pyramidulina latejugata (Gümbel, 1868 – p. 619, pl. 1, fig. 32)
Pyramidulina minor (Hantken, 1875 – Nodosaria bacillum Defrance var. minor, p. 26, pl. 2, fig. 7)
Svenia pauperata (d'Orbigny, 1846 – Dentalina pauperata, p. 46, pl. 1, figs 57–58)
Tollmannia sp.
Frondicularia sp.
Amphimorphina hauerina Neugeboren, 1850 (p. 1217, pl. 4, figs 13–14)
Parafrondicularia striata (Hantken, 1875 – Flabellina striata, p. 43, pl. 13, fig. 13)
Plectofrondicularia digitalis (Neugeboren, 1850 – Frondicularia digitalis, p. 120, pl. 3, fig. 3)
Plectofrondicularia raricosta (Karrer, 1877 – Frondicularia raricosta, p. 381, pl. 164, fig. 28)
Dimorphina sp.
Lenticulina arcuatostriata (Hantken, 1868 – Cristellaria (Robulina) arcuatostriata, p. 93, pl. 2, fig. 30)
Lenticulina calcar (Linne, 1758 – Nautilus calcar, p. 709)
Lenticulina cultrata (Monfort, 1808 – Robolus cultratus, p. 214, no. 54)
Lenticulina depauperata (Reuss, 1851 – Robulina depauperata, p. 70, pl. 4, fig. 29)
Lenticulina inornata (d'Orbigny, 1846 – Robulina inornata, p. 102, pl. 4, figs 25–26)
Lenticulina limbosa (Reuss, 1863 – Robulina limbosa, p. 55, pl. 16, fig. 69)
Lenticulina princeps (Reuss, 1865 – Robulina princeps, p. 32, pl. 5, fig. 3)
Lenticulina vortex (Fichtel and Moll, 1798 – fide Ellis and Messina catalogue) pl. IV. fig. 20
Percultazanaria fragaria (Gümbel, 1868 – Marginulina fragaria, p. 57, pl. 1, fig. 58)
Percultazanaria pseudodecorata (Hagn, 1952 – Vaginulinopsis pseudodecorata, p. 49, pl. 5, fig. 11)
Saracenaria hantkeni Cushman, 1933 (Saracenaria arcuata d'Orbigny var. hantkeni p. 4, pl. 1, figs 11–12)
Saracenaria propinqua (Hantken, 1875 – Cristellaria propinqua, p. 52, pl. 5, fig. 4)
Frondovaginulina tenuissima (Hantken, 1875 – Frondicularia tenussima, p. 43, pl. 13, fig. 11)
Palmula budensis (Hantken, 1875 – Flabellina budensis, p. 44, pl. 4, fig. 17)
Amphicoryna badenensis
Amphicoryna scalaris (Batsch, 1791 – Nautilus (Ortoceras) scalaris, p. 14, pl. 2, fig. 4)
Amphicoryna tunicata (Hantken, 1868 – Cristellaria (Marginulina) tunicata, p. 91, pl. 2, fig. 24; syn: A. marginuliformis Nyirő, 1961)
Amphicoryna spinicosta (d'Orbigny, 1846 – Nodosaria spinicosta d'Orbigny, p. 37, pl. 1, figs 32–33)
Astacolus eocaenus (Gümbel, 1870 – Vaginulina eocaena, p. 632, pl. 1, fig. 49)
Marginulina behmi (Reuss, 1865 – Cristellaria (Marginulina) behmi, p. 138, pl. 2, fig. 37)
Planularia budensis (Hantken, 1875 – Robulina budensis, p. 58, pl. 7, fig. 1)
Planularia dubia Kleinpell, 1938 – (p. 207, pl. 13, fig. 4)
Planularia kubinyii (Hantken, 1868 – Cristellaria (Robulina) kubinyii, p. 92, pl. 2, fig. 29)
Lagena acuticosta Reuss, 1861 (p. 305, pl. 1, fig. 4)
Lagena gracilicosta Reuss, 1858 (p. 434) (pl. V, fig. 11)
Lagena striata (d'Orbigny, 1839 – Oolina striata, p. 21, pl. 5, fig. 12)
Lagena sulcata (Walker and Jacob, 1798 – Serpula (Lagena) sulcata, p. 634, pl. 14, fig. 5)
Lagena tenuis (Bornemann, 1855 – Ovulina tenuis, p. 317, pl. 12, fig. 3)
Pygmaeoseistron hispida (Reuss, 1859 – Lagena hispida, p. 335, pl. 6, figs 77–79)
Pygmaeoseistron vulgaris (Williamson, 1858 – Lagena vulgaris, p. 3, pl. 1, fig. 5)
Glandulopleurostomella subcylindrica (Hantken, 1875 – Polymorphina subcylindrica, p. 51, pl. 14, fig. 14)
Globulina gibba gibba (d'Orbigny, 1826 – Polymorphina (Globulina) gibba, p. 268, no. 20)
Globulina gibba tuberculata d'Orbigny, 1846 (p. 230, pl. 13, figs 17–18)
Globulina granulosa (Egger, 1857 – Polymorphina (Globulina) granulosa, p. 290, pl. 14, figs 1–2)
Guttulina austriaca d'Orbigny, 1846 (p. 223, pl. 12, figs 23–25)
Guttulina communis (d'Orbigny, 1846 – Polymorphina (Guttulina) communis, p. 224, pl. 12, figs 26–28)
Guttulina hantkeni Cushman and Ozawa, 1930 (p. 33, pl. 5, figs 4–6)

Polymorphina decora Reuss, 1863 (p. 152, pl. 3, fig. 41)
Polymorphina incerta Egger, 1857 (p. 186, pl. 13, figs 19–21)
Pseudopolymorphina dollfusi Cushman and Ozawa, 1930 (p. 106, pl. 27, figs 6–7)
Pyrulina fusiformis (Roemer, 1838 – Polymorphina fusiformis, p. 386, pl. 3, fig. 37)
Pyrulinoides sp.
Sigmomorphina regularis (Roemer, 1838 – Polymorphina regularis, p. 385, pl. 3, fig. 21)
Ramulina sp.
Cushmanina sp.
Bucherina marginata (Walker and Jacob, 1784 – Serpula (Lagena) marginata, p. 2, pl. 1, fig. 7)
Bucherina orbignyana (Seguenza, 1862 – Fissurina (Fissurine) orbignyana, p. 66, pl. 2, figs 25–26)
Favulina hexagona (Williamson, 1848 – Entosolenia squamosa Montagu var. hexagona, p. 20, pl. 2, fig. 23)
Oolina globosa (Montagu, 1803, – Vermiculum globosum, p. 523)
Fissurina laevigata Reuss, 1850 (p. 366, pl. 46, fig. 1)
Pseudofissurina scarenaensis (Hantken, 1883 – Lagena scarenaensis, p. 24, pl. 1, fig. 9)
Glandulina dimorpha (Bornemann, 1855 – Guttulina dimorpha, p. 345, pl. 17, fig. 5)
Glandulina laevigata (d'Orbigny, 1826 – Nodosaria (Glandulina) laevigata, p. 253, no. 1)

Suborder: Robertinina Loeblich and Tappan, 1984
Ceratobulimina contraria (Reuss, 1851 – Rotalina contraria, p. 76, pl. 5, fig. 37)
Ceratocancris eximus Rzehak, 1883 (fide Ellis and Messina Catalogue)
Lamarckina sp.
Hoeglundina elegans (d'Orbigny, 1826 – Rotalina (Truncatulina) elegans, p. 276)
Hoeglundina eocaenica (Cushman and Hanna, 1927 – Epistomina eocaenica, p. 53, pl. 5, figs 4–5)

Suborder: Globigerinina Delage and Hérouard, 1896
Chiloguembelina gracillima (Andraea, 1884 – Textilaria gracillima, p. 235, pl. 8, fig. 9)
Globorotalia clemenciae (Bermúdez, 1961 – Turborotalia clemenciae, p. 1321, pl. 17, figs 12–13)
Globorotalia siakensis (Leroy, 1939 – Globigerina siakensis, p. 262, pl. 4, figs 20–22)
Globorotalia obesa Bolli, 1957 (p. 119, pl. 29, figs 2–3)
Paragloborotalia opima nana (Bolli, 1957 – Globorotalia opima nana, p. 118, pl. 28, fig. 3)
Paragloborotalia opima opima (Bolli, 1957 – Globorotalia opima opima, p. 117, pl. 28, figs 1–2)
Turborotalia ampliapertura (Bolli, 1957 – Globigerina ampliapertura, p. 108, pl. 22, figs 5a–7b)
Turborotalia ceroazulensis cerroazulensis (Cole, 1928 – Globigerina cerro-azulensis, p. 217, pl. 32, figs 11–13)
Turborotalia increbescens (Bandy, 1949 – Globigerina increbescens, p. 120, pl. 23, figs 3a–c)
Tenuitella gemma (Jenkins, 1966 – Globorotalia gemma, p. 1115, pl. 11, figs 97–103)
Tenuitellinata angustiumbilicata (Bolli, 1957 – Globigerina ciperoensis angustiumbilicata, p. 109, pl. 22, figs 12–13)
Catapsydrax martini scandretti (Blow and Banner, 1962 – Globigerinata martini scandretti, p. 111, pl. 14, figs V–X)
Catapsydrax unicavus Bolli, Loeblich, and Tappan, 1957 (p. 37, pl. 7, fig. 9)
Globigerinita sp.
Dentoglobigerina galavisi (Bermúdez, 1961 – Globigerina galavisi, p. 1183, pl. 4, fig. 3)
Globoquadrina sp.
Globorotaloides suteri Bolli, 1957 (p. 117, pl. 27, figs 9–13)
Subbotina angiporoides (Hornibrook, 1965 – Globigerina angiporoides, p. 835, text.figs 1–2)
Subbotina eocaena eocaena (Gümbel, 1870 – Globigerina eocaena, p. 84, pl. 5, figs 109a–b)
Subbotina eocaena praeturritilina (Blow and Banner, 1962 – Globigerina turritilina praeturritilina, p. 99, pl. 13, figs A–C)
Subbotina linaperta linaperta (Finlay, 1939 – Globigerina linaperta, p. 125, pl. 13, text.figs 54–57)
Subbotina utilisindex (Jenkins and Orr, 1973 – Globigerina utilisindex, p. 133, pl. 1, figs 1–6, pl. 2, figs 1–9, pl. 3, figs 1–3)

Subbotina tapuriensis (Blow and Banner, 1962 – Globigerina tripartita tapuriensis, p. 99, pl. 101, figs H–K)
Subbotina tripartita (Koch, 1926 – Globigerina bulloides d'Orbigny var. tripartita, p. 746, pl. 4, figs 4a–c)
Globanomalina barbadoensis (Blow, 1969 – p. 409, pl. 53, figs 5–7)
Globanomalina micra (Cole, 1927 – Nonion micrus, p. 22, pl. 5, fig. 12)
Globigerina angulioffícinalis Blow, 1969 (p. 379, pl. 11, figs 1–5)
Globigerina angulisuturalis Bolli, 1957 (p. 109, pl. 22, figs 11a–c)
Globigerina ciperoensis ciperoensis Bolli, 1954 (p. 1, text.figs 3–6)
Globigerina euapertura Jenkins, 1960 (p. 51, pl. 1, fig. 8)
Globigerina liverovskae (Bykova, 1960 – Globorotalia liverovskae, p. 322, pl. 7, figs 1–3)
Globigerina ouachitaensis gnaucki Blow and Banner, 1962 (p. 91, pl. 9, figs L–N)
Globigerina ouachitaensis ouachitaensis Howe and Wallace, 1932 (Globigerina ouachitaensis, p. 74, pl. 10, fig. 7)
Globigerina praebulloides leroyi Blow and Banner, 1962 (p. 93, pl. 9, figs R–T)
Globigerina prebulloides occlusa Blow and Banner, 1962 (p. 93, pl. 9, figs U–W)
Globigerina prebulloides praebulloides Blow, 1959 (p. 180, pl. 8, fig. 47; pl. 9, fig. 48)
Globigerinoides primordius Banner and Blow, 1962 (p. 115, pl. 9, figs D–F)
Globoturborotalita woodi (Jenkins, 1960 – Globigerina woodi, p. 352, pl. 2, fig. 2)

Suborder: Rotaliina Delage and Hérouard, 1896
Bolivina antiqua d'Orbigny, 1846 (p. 240, pl. 14, figs 11–13)
Bolivina beyrichi beyrichi Reuss, 1851 (p. 83, pl. 6, fig. 51)
Bolivina beyrichi carinata Hantken, 1875 (p. 55, pl. 7, fig. 12)
Bolivina beyrichi morhinvegi Cushman, 1935 (Bolivina morhinvegi, p. 32, pl. 5, fig. 1)
Bolivina budensis (Hantken, 1875 – Textilaria budensis, p. 57, pl. 15, fig. 1)(pl. VI, fig. 3
Bolivina concinna Knipscher and Martin, 1955 (p. 261, text.fig. 1)
Bolivina crenulata crenulata Cushman, 1936 (p. 50, pl. 7, fig. 13; syn: Bolivina plicatella Cushman)
Bolivina crenulata trunensis Hofmann, 1967 (p. 147, pl. 5, figs 1–4)
Bolivina dilatata dilatata Reuss, 1850 (p. 381, pl. 48, fig. 15)
Bolivina elongata Hantken, 1875 (p. 55, pl. 15, fig. 3)
Bolivina floridana Cushman, 1918 (p. 49, pl. 10, fig. 4)
Bolivina hebes Macfadyen, 1930 (p. 59, pl. 2, fig. 5)
Bolivina liebusi Hofmann, 1967 (p. 176, pl. 2, figs 6–8)
Bolivina molassica Hofmann, 1967 (p. 158, pl. 4, figs 9–11)
Bolivina nobilis Hantken, 1875 (p. 56, pl. 15, fig. 4)
Bolivina oligocaenica oligocaenica Spandel, 1909 (p. 208, pl. 1, fig. 16)
Bolivina oligocaenica varica Hofmann, 1967 (p. 167, pl. 4, fig. 14) pl. VI, fig. 10
Bolivina semistriata Hantken, 1875 (p. 95, pl. 2, fig. 34)
Brizalina aenariensiformis (Myatlyuk 1960 – in Subbotina, Bolivina aenariensiformis, p. 223, pl. 5, figs 1–6)
Latibolivina fastigia (Cushman, 1936 – Bolivina fastigia, p. 51, pl. 7, fig. 17)
Latibolivina fastigia droogeri (Cicha and Zapletalova, 1963 – Bolivina fastigia droogeri, p. 122, text.fig. 4)
Latibolivina reticulata (Hantken, 1875 – Bolivina reticulata, p. 65, pl. 15, fig. 6)
Bolivinella rugosa Howe, 1930 (p. 267, pl. 212, fig. 4)
Cassidulina crassa d'Orbigny, 1839 (p. 56, pl. 7, figs 18–20)
Cassidulina laevigata d'Orbigny, 1826 (p. 281)
Cassidulina vitólisi Majzon, 1948 (p. 24, fig. 2)
Cassidulinoides oblongus (Reuss, 1850 – Cassidulina oblonga, p. 376, pl. 48, figs 5–6)
Globocassidulina globosa (Hantken, 1875 – Cassidulina globosa, p. 64. pl. 16, fig. 2)
Burseliana inexculta (Franzenau, 1889 – Cassidulina inexculta, fide Ellis and Messina Catalogue))
Burseliana margareta (Karrer, 1877 – Cassidulina margareta, fide Ellis and Messina Catalogue)

Ehrenbergina serrata Reuss, 1850 (fide Ellis and Messina Catalogue)
Turrilina alsatica Andreae, 1884 (fide Ellis and Messina Catalogue)
Hopkinsina bononiensis primiformis – Uvigerina bononiensis primiformis Papp and Turnovsky, 1953 (p. 121, pl. 5/1, figs 1–2)
Virgulopsis pupoides (Nyirő, 1954 – Corrosina pupoides, p. 69, text. fig.)
Siphogenerina sp.
Tubuligenerina tubulifera Parker and Jones, 1863 (fide Ellis and Messina Catalogue)
Bulimina alazanensis Cushman, 1927 (fide Ellis and Messina Catalogue)
Bulimina alsatica Cushman and Parker, 1937 (p. 39, pl. 4, figs 6–7)
Bulimina sculptilis Cushman, 1927 (fide Ellis and Messina Catalogue)
Bulimina striata d'Orbigny, 1826 (p. 269, no. 2.)
Globobulimina pyrula (d'Orbigny, 1846 – Bulimina pyrula, p. 184, pl. 11, figs 9–10)
Globobulimina ovata (d'Orbigny, 1846 – Bulimina ovata, p. 185. pl. 11. figs 13–14
Protoglobobulimina pupoides (d'Orbigny, 1846 – Bulimina pupoides, p. 185, pl. 11, figs 11–12)
Uvigerina cocoaensis Cushman and Edwards, 1937 (p. 68, pl. 10, fig. 12)
Uvigerina eocaena Gümbel, 1870 (p. 645, pl. 2, fig. 78)
Uvigerina gallowayi Cushman, 1929 (p. 94, pl. 13, figs 33–34)
Uvigerina gracilis Reuss, 1851 (p. 77, pl. 5, fig. 39)
Uvigerina hantkeni Cushman and Edwards, 1937 (p. 60, pl. 8, figs 15–16)
Uvigerina mantensis Cushman and Edwards, 1938 (p. 84, pl. 14, fig. 8)
Uvigerina moravia Borsma, 1984 (p. 114, pl. 1, figs 1–4)
Uvigerina multistriata Hantken, 1871 (p. 129)
Uvigerina parviformis Papp, 1953 (p. 305, pl. 1, figs 1–3)
Uvigerina rippensis Cole, 1927 (p. 11, pl. 2, fig. 16)
Uvigerina rudligensis Papp, 1975 (p. 283, pl. 1, figs 1–4)
Uvigerina semiornata semiornata d'Orbigny, 1846 (p. 189, pl. 11, figs 2–24)
Angulogerina angulosa (Williamson, 1858) – Uvigerina angulosa, p. 67, pl. 5, fig. 140
Angulogerina globosa (Stoltz, 1925 – Uvigerina tenuistriata Reuss var. globosa Stoltz, fide Ellis and Messina Catalogue)
Kolesnikovella tubulifera (Kaaschieter, 1961 – Angulogerina abbreviata (Terquem) var. tubulifera, p. 198, pl. 10, figs 1–2)
Trifarina bradyi Cushman, 1923 (p. 99, pl. 22, figs 3–9)
Trifarina budensis (Hantken, 1868 – Rhabdogonium budensis, p. 90, pl. 1, fig. 19)
Trifarina halkyardi (Cushman and Edwards, 1937 – Angulogerina halkyardi, p. 60, pl. 8, fig. 14)
Reusella spinulosa (Reuss, 1850 – Verneuilina spinulosa, p. 3374, pl. 47, fig. 12)
Reusella triquetra (Franzenau, 1842 – Bulimina triquetra, pp. 108–109)
Fursenkoina acuta (d'Orbigny, 1846 – Polymorphina acuta, p. 147, pl. 13, figs 18–21 = Virgulina schreibersiana Czjzek, 1848)
Fursenkoina halkyardi (Cushman, 1936 – Virgulina halkyardi, p. 47, pl. 7, fig. 5)
Coryphostoma digitale (d'Orbigny, 1846 – Polymorphina digitalis, p. 235, pl. 14, figs 1–4)
Coryphostoma minutissima (Spandel, 1909 – Bolivina minutissima, p. 209, pl. 1, fig. 11)
Coryphostoma sinuosa (Cushman, 1936 – Loxostomum sinuosum, p. 60, pl. 8, fig. 16)
Caucasina elongata (d'Orbigny, 1826 – Bulimina elongata, p. 269)
Caucasina oligocaenica Khalilov, 1952 (fide Ellis and Messina Catalogue)
Ellipsoglandulina vásárhelyii (Hantken, 1868 – Nodosaria (Dentalina) vásárhelyi, p. 89, pl. 2, fig. 35)
Pleurostomella eocaena (Gümbel, 1868 – p. 630, pl. 1, fig. 53)
Nodogenerina adolphina (d'Orbigny, 1846 – Dentalina adolphina, p. 51, pl. 2, figs 18–20)
Nodogenerina rohri Cushman and Stainforth, 1945 (fide Ellis and Messina catalogue)
Siphonodosaria consobrina (d'Orbigny, 1846 – Dentalina consobrina, p. 46, pl. 2, figs 1–3)
Siphonodosaria verneuili (d'Orbigny, 1846 – Dentalina verneuili, p. 48, pl. 2, figs 7–8)
Stilostomella hoernesi (Hantken, 1868 – Nodosaria (Dentalina) hoernesi, p. 89, pl. 1, fig. 14)
Baggina philippinensis (Cushman, 1921 – Pulvinulina philippinensis, p. 31, pl. 58, fig. 2)
Cancris auriculus de Montfort, 1808 (= Nautilus auricula Fichtel et Moll, 1798, p. 108)

Cancris turgidus Cushman and Todd, 1942 (p. 92, pl. 24, figs 3–4)
Valvulineria complanata (d'Orbigny, 1846 - Rosalina complanata, p. 175, pl. 10, figs 13–15)
Eponides repandus (Fichtel and Moll, 1798 - Nautilus repandus, fide Ellis and Messina Catalogue)
Eponides polygonus Le Calvez, 1949 (p. 28, pl. 5, figs 90–92)
Schlosserina asterites (Gümbel, 1870 – (Rosalina asterites, fide Ellis and Messina Catalogue)
Stomatorbina concentrica (Jones and Parker, 1864 - Pulvinulina concentrica, p. 470, pl. 48, fig. 14)
Stomatorbina torrei (Cushman and Bermúdez, 1937 - Lamarckina torrei Cushman and Bermúdez
Mississippina monsouri Howe, 1930 (fide Ellis and Messina Catalogue)
Discorbis squamula (Reuss, 1875 – Discorbina squamula, p. 85, pl. 5, fig. 2)
Neoeponides eximius (Hantken, 1875 – Discorbina eximia, p. 76, pl. 15, fig. 8)
Neoeponides schreibersii (d'Orbigny, 1826 – Rotalina schreibersii, p. 154, pl. 8, figs 4–6)
Neoconorbina terquemi (Rzehak, 1888 – fide Ellis and Messina Catalogue)
Rosalina globularis d'Orbigny, 1826 (p. 271, no. 69)
Sphaeroidina bulloides d'Orbigny, 1826 (p. 267, no. 1)
Glabratella loerentheyi Sztrákos, 1982 (p. 25, pl. 25, figs 1–12, pl. 31, figs 3, 6)
Siphonina reticulata (Czjzek, 1848 – Rotalina reticulata, p. 145, pl. 13, figs 7–9 = Siphonina fimbrata Reuss, 1850)
Siphonina subreticulata Myatlyuk, 1953 (p. 234, pl. 8, fig. 7)
Cibicidoides budayi (Cicha and Zapletalova, 1958 - Cibicides budayi, p. 14, pl. 4, fig. 2)
Cibicidoides pseudoungerianus (Cushman, 1922 - Truncatulina pseudoungeriana, p. 97, pl. 20, fig. 9)
Cibicidoides pygmeus (Hantken, 1875 – Pulvinulina pygmea, p. 78, pl. 10, fig. 8)
Cibicidoides roemeri (Reuss, 1856 – Rotalia roemeri, p. 240, pl. 4, fig. 52)
Cibicidoides ungerianus (d'Orbigny, 1846 - Rotalina ungeriana, p. 157, pl. 8, figs 16–18)
Biapertorbis biaperturata Pokorny, 1956 (p. 265, figs 4–6)
Planulina ambiqua (Franzenau, 1888 - Rotalia ambiqua, p. 106, pl. 2, figs 9–11)
Planulina compressa (Hantken, 1875 - Truncatulina compressa, p. 72, pl. 8, fig. 8)
Planulina costata (Hantken, 1875 – Truncatulina costata, p. 73, pl. 9, fig. 2)
Halkyardia minima (Liebus, 1911 – Cymbalopora radiata Hagenow var. minima, p. 952, pl. 3, fig. 7)
Cibicides sp.
Falsoplanulina ammophila (Gümbel, 1870 - Rotalia ammophila, fide Ellis and Messina catalogue)
Fontbotia wuellerstorfi (Schwager, 1866 - Anomalina wuellerstorfi, p. 258, pl. 7, figs 105, 107)
Lobatula lobatula (Walker and Jacob, 1798 - Nautilus lobatulus in Kanmacher, syn: Lobatula vulgaris Fleming, 1828, fide Ellis and Messina Catalogue)
Dyocibicides biserialis Cushman and Valentine, 1938 (p. 31, figs 1–2)
Stichocibicides cubensis Cushman and Bemudez, 1936 (fide Ellis and Messina Catalogue)
Cycloloculina annulata Heron-Allen and Earland, 1908 (p. 533, pl. 12, figs 1–7, syn: Planorbulina eocaenica Terquem, 1882)
Planorbulina mediterranensis d'Orbigny, 1826 (p. 280, no. 2; syn.: Planorbulina vulgaris d'Orbigny, 1839)
Korobkovella grosserugosa (Gümbel, 1868 - Truncatulina grosserugosa, p. 660, pl. 2, fig. 104)
Asterigerinata falcilocularis (Subbotina, 1960 - Asterigerina falcilocularis, p. 194, pl. 6, figs 12–13)
Asterigerinoides gürichi (Franke, 1912 - Discorbina grichi, p. 29, text.fig. 8)
Biasterigerina planorbis (d'Orbigny, 1846 – Asterigerina planorbis, p. 205, pl. 11, figs 1–3)
Asterigerina bimammata Gümbel, 1870 (fide Ellis and Messina Catalogue)
Asterigerina rotula (Kaufmann, 1867 - Hemistegina rotula, fide Ellis and Messina Catalogue)
Amphistegina choctawensis (Cushman and McGlamery, 1938 – Asterigerina choctawensis, fide Ellis and Messina Catalogue)
Amphistegina lessoni d'Orbigny, 1826 (p. 304, no. 3)
Nonion boueana (d'Orbigny, 1826 – Nonionina boueana, p. 108, pl. 15, figs 11–12)
Nonionella liebusi Hagn, 1952 (p. 161, pl. 2, fig. 10)
Protelphidium subgranosum (Egger, 1857 - Nonionina subgranosa, p. 299, pl. 14, figs 16–18)
Astrononion sp.
Melonis affinis (Reuss, 1851 – Nonionina affinis, p. 72, pl. 5, fig. 32)

Melonis pompilioides (Fichtel and Moll, 1798 – p. 31, pl. 2, figs a–c; syn: Nonionina soldanii d'Orbigny, 1846)
Pullenia sphaeroides (d'Orbigny, 1825 – Nonionina sphaeroides; syn: Nonionina bulloides d'Orbigny, 1826, p. 293)
Pullenia quinqueloba (Reuss, 1851 – Nonionina quinqueloba, p. 71, pl. 5, fig. 31)
Alamaena taurica Samoylova, 1940 (fide Ellis and Messina Catalogue)
Almaena osnabrugensis (Munster, 1838 – Planulina osnabrugensis, fide Ellis and Messina Catalogue) (pl. XIII. fig. 5)
Pseudoplanulinella hieroglyphica (Sigal, 1950 – Planulinella hieroglyphica, fide Ellis and Messina Catalogue)
Pseudoplanulinella siphoninaeformis (Sigal, 1950 – Kelyphistoma siphoninaeformis, fide Ellis and Messina Catalogue)
Queraltina epistominoides Marie, 1950 – (p. 74, figs: p. 75, tfs. 1–3, p. 76, tf. 8, p. 77, tf. 9)
Allomorphina trigona Reuss, 1850 (p. 380, pl. 48, fig. 14)
Chilostomella ovoidea Reuss, 1850 (p. 380, pl. 48, fig. 12)
Quadrimorphina petrolei (Andraea, 1884 – Pulvinulina petrolei, p. 217, pl. 8, fig. 15)
Alabamina wolterstorffi (Franke, 1925 – Rotalia wolterstorffi, p. 186, pl. 6, fig. 66)
Svratkina perlata (Andreae, 1884 – Pulvinulina perlata, p. 216, pl. 8, fig. 12)
Charltonina budensis (Hantken, 1875 – Pulvinulina budensis, p. 65, pl. 8, fig. 6; syn: Pulvinulina tangentialis Clodius, 1922)
Oridorsalis umbonatus (Reuss, 1851 – Rotalina umbonata, p. 75, pl. 5, fig. 35)
Anomalinoides affinis (Hantken, 1875 – Pulvinulina affinis, p. 78, pl. 10, fig. 6)
Anomalinoides crassiseptatus (Cushman and Siegfus, 1935 – Anomalina crassiseptata, p. 95, pl. 14, fig. 18)
Anomalinoides similis (Hantken, 1875 – Pulvinulina similis, p. 78, pl. 10, fig. 5)
Gavellinella asterians (Fichtel and Moll, 1798 – Nautilus asterians, fide Ellis and Messina Catalogue; syn: Rotalina cryptomphala Reuss, 1850, p. 371, pl. 47, fig. 2, and Anomalina alazanensis Nuttall, 1932, p. 31, pl. 8, figs 5–7)
Heterolepa dutemplei (d'Orbigny, 1846 – Rotalina dutemplei, p. 157, pl. 8, figs 19–21)
Gemellides costatus (Franzenau, 1884 – Heterolepa costata, p. 183, pl. 5, fig. 2)
Gemellides cubensis (van Bellen, 1938 – Cibicides cubensis, fide Ellis and Messina Catalogue)
Gemellides eocaenaus (Gümbel, 1868 – Rotalina eocaena, p. 650, pl. 2, fig. 87)
Escornebovina cuvillieri (Poignant, 1965 – Rotalia cuvillieri, p. 103, pl. 1, figs 1–2, 5–6; syn: Pseudopatellina plana Kenawy and Nyirő, 1967)
Escornebovina legányii (Kenawy and Nyirő, 1967 – Patellina legányii, p. 103, pl. 1, figs 1–2)
Gyroidinoides girardanus Reuss, 1851 (Rotalina girardana, p. 73, pl. 5, fig. 34)
Gyroidinoides mamillatus (Andreae, 1884 – Gyroidina girardana Reuss var. mamillata, p. 234, pl. 9, fig. 4)
Hansenisca soldanii (d'Orbigny, 1826 – Gyrodina soldanii, p. 278)
Hanzawaia americana (Cushman, 1918 – Truncatulina americana, p. 63, pl. 20, figs 2–3; pl. 21, fig. 1)
Hanzawaia boueana (d'Orbigny, 1846 – Truncatulina boueana, p. 169, pl. 9, figs 24–26)
Hanzawaia horcici (Cicha and Zapletalova, 1958 – Cibicides horcici, p. 23, pl. 8, figs 4–9)
Linaresia granosa (Hantken, 1875 – Truncatulina granosa, p. 65, pl. 10, fig. 2)
Pararotalia audouini (d'Orbigny, 1850 – Rotalia audouini, p. 407, fig. 9)
Pararotalia canui Cushman, 1928 (p. 55, pl. 3, fig. 2)
Pararotalia lithothamnica (Uhlig, 1886 – Rotalia lithothamnica, p. 195, pl. 5, figs 8, 10, 11, text.fig. 6)
Rotalia trochus Romer, 1838 (p. 388, pl. 3, fig. 47)
Ammonia beccarii (Linné, 1758 – Nautilus beccarii, fide Ellis and Messina Catalogue)
Ammonia kiliani (Andreae, 1884 – Pulvinulina kiliani, p. 256, pl. 11, fig. 1)
Ammonia propinqua Reuss, 1856 (p. 241, pl. 4, fig. 53)
Cribroelphidium dollfusi (Cushman, 1936 – Nonion dollfusi, p. 65, pl. 12, fig. 4)
Cribroelphidium subnodosum (Roemer, 1838 – Robulina subnodosa, p. 391, pl. 3, fig. 61)
Cribrononion hiltermanni (Hagn, 1952 – Elphidium hiltermanni, p. 163, pl. 1, fig. 6; pl. 2, fig. 14)
Cribrononion minutum (Reuss, 1865 – Polystomella minuta, p. 478, pl. 4, fig. 6)

Elphidium crispum (Linné, 1758, fide Ellis and Messina Catalogue)
Elphidium flexuosum (d'Orbigny, 1846 – Polystomella flexuosa, p. 127, pl. 6, figs 15–16)
Elphidium latidorsatum (Reuss, 1864 – Polystomella latidorsata, p. 10, pl. 1, fig. 16)
Elphidium ortenburgense (Egger, 1867 – Polystomella ortenburgensis, p. 302, pl. 15, figs 7–9)
Elphidium semiinvolutum Myatlyuk (p. 788, figs 1–5)
Miogypsina (Miogypsina) septentrionalis Drooger, 1960 (p. 41, pl. 1, figs 1–2; pl. 2, figs 1–8)
Miogypsinoides (Miogypsina) formosensis Yabe and Hanzawa, 1928 (Miogypsina (Miogypsinoides) dehaarti (Van der Vlerk) var. formosensis, p. 535, fig. 1
Heterostegina costata d'Orbigny, 1846 (p. 212, pl. 12, figs 15–17)
Operculina complanata (Defrance, 1822 – Lenticulites complanatus, p. 453)

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Revision of "*Pachydiscus neubergicus* Hauer 1858", Sümeg, Transdanubian Cental Range, Hungary

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Herbert Summesberger
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An ammonite from Sümeg (Hungary) formerly misidentified as *Pachydiscus neubergicus* Hauer (1858) was introduced into Hungarian literature as an index fossil of the Maastrchtian. It turned out to be in fact Eupachydiscus levyi (De Grossouvre 1894) indicating Lower Campanian.

Key words: revision, *Eupachydiscus levyi*, Lower Campanian, parastratigraphy, Sümeg

Introduction

Several ammonites were collected in the past in the Cretaceous of the Sümeg area (Városi quarry of Haraszt near the city of Sümeg (Fig. 1 Transdanubian Central Range; Hungary). Five of them are preserved in the Geological Institute of Hungary (MÁFI). One of them (MÁFI/K-8645) played under the wrong name of "*Pachydiscus neubergicus* (Hauer 1858)" a misleading role, when the Cretaceous stratigraphy of Hungary was established. It was originally identified by its collector J. Noszky (1944; on labels only) and it is revised and described below.

The present paper summarizes the most important data collected during the revision of the specimen which had been exhibited for 25 years in the ceremonial hall (Cretaceous display) of the Geological Institute of Hungary.

The appearance of this false index fossil in the literature (Lóczy 1913; collection of Károly Papp) and the subsequent age classification of specimens by Noszky (1944, 1957), Vadász (1960) and Benkő-Czabalay (1961) led to the later biostratigraphic confusion (Benkő-Czabalay 1964a, b, Czabalay 1983; Góczán 1978; Góczán et al. 1989; Góczán and Siegl-Farkas 1989; Siegl-Farkas 1993).

The occurrence of *Pachydiscus neubergicus* in the limestone quarries of Haraszt near Sümeg was first mentioned by Lóczy (1913): "Károly Papp determined an ammonite specimen of the size of a barrow wheel as the species *Pachydiscus neubergicus* Hauer". The occurrence of the ammonite is described by Lóczy as in the "yellowish-grey marly limestone" representing the upper 100 m of the Cretaceous layers at Sümeg. Lóczy also lists the co-occurring species "*Inoceramus cripsii* Mant." and "*Rhynchonella clandicans* Coqu.". At the beginning of the 1910s

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these large ammonites were placed in the Darnay Museum, and their further fate cannot be reconstructed any more.

In the course of mapping in the surroundings of Sümeg-Csabrendek in 1957, Noszky (1958) characterized the *Inoceramus*-bearing formations, besides their "characteristic *Globotruncana*-bearing microfauna", by "usually large-sized *Pachydiscus*-like fossils". Unfortunately he gave neither an exact description nor picture of the fossils nor of their localities.

Emőke Jocha-Edelényi (pers. comm.), who took part in the geological mapping of the area, Noszky (1958) also assigned the quarries at the foot of the Rendeki Hill to Haraszt, and the specimen in question was presumably also found here (Fig. 1).

Obviously the identification of the species *Pachydiscus neubergicus* might have motivated the statement of Noszky according to which "chronologically, even the presence of the Maastrichtian Stage would not be surprising in the case of the beds closing the Upper Cretaceous of Sümeg".

Though it was Noszky who first mentioned a Maastrichtian age in the area, never brought it into connection with the presence of *Pachydiscus neubergicus*, and nowhere in the literature is there any indication that he mentions the species *neubergicus*.

In accordance with "*Pachydiscus neubergicus*" the Maastrichtian Stage was mentioned first by Vadász (1960), with reference to Noszky: "Noszky's still not published division mentions *Parapachydiscus neubergicus* from *Inoceramus*-bearing marl layers. This corresponds to the Maastrichtian Stage."

From this time onwards references to the "*Pachydiscus neubergicus*" and its assignment to the Maastrichtian Stage are continuously present in the chronostratigraphic classification of the Upper Cretaceous formations of the Transdanubian Central Range.

According to Benkő-Czabalay (1961): "in his field report, on the basis of the presence of *Parapachydiscus neubergicus* (Schlotth.) (sic!) Noszky (1957) conveys the suggestion that the *Inoceramus*-bearing limestone and marl group is possibly already of Maastrichtian age." Also here Benkő-Czabalay is the first to mention that this stratigraphic classification corresponds to the results of the pollen and foraminifer investigations. She was also first to correlate the exploratory Sp-boreholes drilled north of Sümeg with the layers exposed in the Városi Quarry of Haraszt (Benkő-Czabalay 1964). "In the Városi Quarry of Haraszt, in the layer corresponding to the horizon with *Inoceramus balticus* of the exploration boreholes at Sümeg, Noszky collected the species *Parapachydiscus neubergicus* Schlotth."

Góczán quotes this discovery in the Lexique Stratigraphique International published in 1978.

In her later correlation work Czabalay stated (1983) "With respect to the Inoceramidae and the ammonite species *Parapachydiscus neubergicus* (Schlotth.) the Polány Marl Formation can be placed into the Lower Maastrichtian. On the basis of the microfauna, even Upper Maastrichtian could be determined."

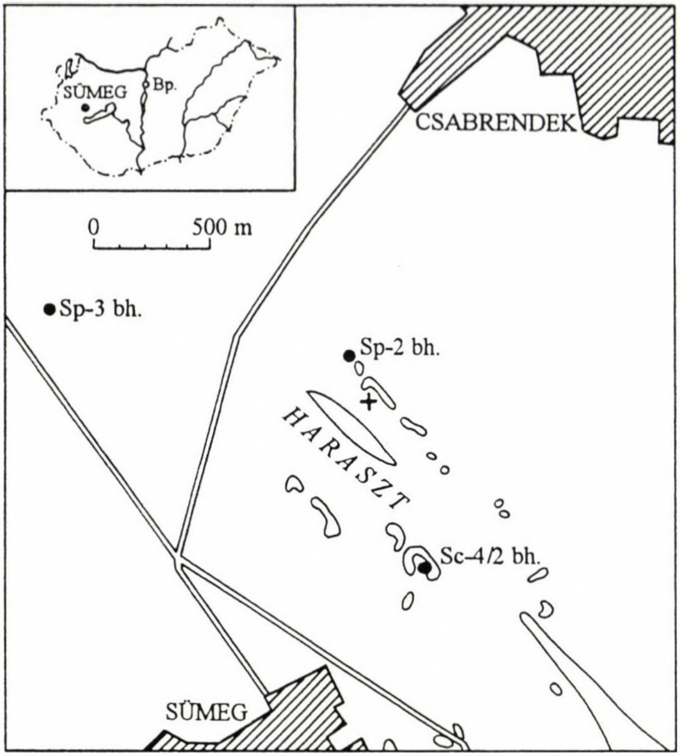

Fig. 1
Area of Haraszt near Sümeg. + – quarry (Polány Marl Formation), where "*Pachydiscus neubergicus*" was presumably found by Noszky, (after Haas et al. 1985)

She also mentioned that the "*Parapachydiscus neubergicus*"-bearing layers can be correlated with palynozone "G" as well as with the *Globotruncana arca* – *Globotruncana stuartiformis* zone.

In the geological map of the monograph on Sümeg Haas et al. (1985) indicate the exact location of the quarries of Haraszt at Sümeg (Fig. 1). Concerning the macrofossils of Haraszt they make the following statement: "Ammonites are scarce. Beside a few *Scaphites*, one specimen of *Pachydiscus neubergicus* Schlotth. sampled by J. Noszky Jr. from the Városi Quarry in the Haraszt area is known. In terms of the ammonite chronozonation this species is the zonal index fossil of the Maastrichtian Stage. Since borehole Sc-4/2 was spudded in the yard of this quarry an exact interregional correlation is possible. A comparison of the sequences suggests that the recovered specimen may have derived from the upper part of the lower interval of the Polány Formation".

In the section representing the total Upper Cretaceous sequence of the Transdanubian Range, first Góczán et al. (1989) and then Góczán and Siegl-Farkas (1989, 1990) indicate the location of the most probable occurrence of *Pachydiscus neubergicus* of Noszky, pointing out its index and correlation value: "The palynomorph assemblages from the ammonite-bearing *Parapachydiscus neubergicus* beds have been taken as the standard for correlation of the Early Maastrichtian deposits."

The specimen was figured by Siegl-Farkas (1997) under *Pachydiscus levyi* De Grossouvre 1894 (Summesberger, pers. comm.).

Thereafter, Yazykova revised the same specimen as *Eupachydiscus levyi* Grossouvre 1894 (in Bodrogi et al. 1997).

In the parastratigraphic zonations of the Upper Cretaceous sequences of the Transdanubian Range the first step toward omitting the Maastrichtian Stage was the designation of the most probable place of the Santonian-Campanian boundary (Lantos et al. 1996). It was followed by the establishing and correlation with each other of the dinoflagellata and nannoplankton zones (Siegl-Farkas and Wagreich 1996; Siegl-Farkas 1997) allowing the completely developed sequence of the Transdanubian Range to be assigned into Upper Santonian–Upper Campanian.

This result is also confirmed by the ammonite species *Eupachydiscus levyi* De Grossouvre 1894 revised here, now assigned into the Lower Campanian.

The Subcommission on Cretaceous Stratigraphy proposed at the meeting in Brussels 1995 (Odin 1996) the first occurrence of *Pachydiscus neubergicus* (Hauer 1858) as the index fossil for definition of the base of the Maastrichtian.

Systematic palaeontology

Family Pachydiscidae Spath, 1922

Genus Eupachydiscus Spath, 1922

Eupachydiscus levyi De Grossouvre 1894

Plate I, figs 1, 2

Material: The specimen from the Noszky collection is stored at the Museum of the Geological Institute of Hungary (K-8645) from Sümeg, Haraszt, Város Quarry, lower third of the Polány Marl Formation.

Synonymy

1894 Pachydiscus Levyi n. sp., – De Grossouvre. p. 178, pl. 21., pl. 30, figs 1, 2.
1908 Pachydiscus Levyi (De Grossouvre) – De Grossouvre, p. 312.
1922 Eupachydiscus levyi (De Grossouvre)– Spath, p. 124.
1932 Pachydiscus aff. Levyi (De Grossouvre), – Collignon, p. 23, text.fig. 2. pl. 6, figs 3, 3a.
1938 Eupachydiscus Levyi (De Grossouvre), – Collignon, p. 14, text. fig. A, pl. 3, figs 3, 3a, p. 27.
1955 Pachydiscus Levyi (De Grossouvre) – Collignon, p. 34, text. fig. 5.
1986 Eupachydiscus cf. levyi (De Grossouvre) – Kennedy, p.163 pl. 5, figs. 1, 2.
1997 Pachydiscus levyi (De Grossouvre) – Summesberger (in Siegl-Farkas), p. 85, pl. 8, figs 1, 2, 3.

1997 Eupachydiscus levyi (De Grossouvre) - Yazykova (in Bodrogi et al). p. 693

Description

The specimen is a large fragment of an inner mould of a phragmocone. The surface is heavily corroded. No shell remains are preserved. The shell proportion is only slightly flattened by post mortem compaction. The shape is discoidal, coiling modestly involute, approximately 50 percent of the preceding whorl being covered, decreasing with growth to about 40 percent, the umbilicus comprising 23% of the diameter. The whorl section is high oval. Umbilical wall and umbilical shoulder are gently rounded. The flanks are slightly inflated convergent at the outer half without a marked ventrolateral shoulder. Greatest breadth is below mid flank. The venter is narrowly rounded without a marked keel. About 20 primary ribs arise with a more or less prominent bulla, being rursiradiate at the umbilical shoulder. Over the flanks they are slightly concave. At the outer third of the flank they flex forward crossing the venter in a projected loop. Few intercalatory ribs arise above mid-flank crossing the venter in the above-mentioned manner. At the last part visible, possibly the adult stage, the ribbing changes to a more distant and slightly irregular style. The suture is visible only in small fragments.

Dimension

D max 237 mm, Wh 92.4 mm, Wb 58.5 mm, U 53.6 mm, U% 23.

Stratigraphic occurrence

Eupachydiscus levyi is a typical Campanian species occurring in the lower part of the stage. It was originally described by De Grossouvre (1894) from basal Campanian units of Contes-les-Pins in the Alpes-Maritimes of France. Collignon (1938, p. 65) described Lower and Middle Campanian occurrences of Madagascar. A fragment of *Eupachydiscus* cf. *levyi* was mentioned by Kennedy (1986, pl. 5, fig. 1) from the Lower Campanian Vaals Formation of The Netherlands. The Hungarian occurrence is the first described in Central Europe. Taking the described stratigraphical distribution into consideration a Campanian age must be assumed for the specimen described, Early Campanian age seems likely.

Acknowledgements

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Plate I
1, 2. Eupachydiscus levyi (De Grossouvre 1894)
Inv. Nr.: K-8645, Sümeg, Haraszt. M = 0.5x
Photo: Ildikó Laky, MÁFI

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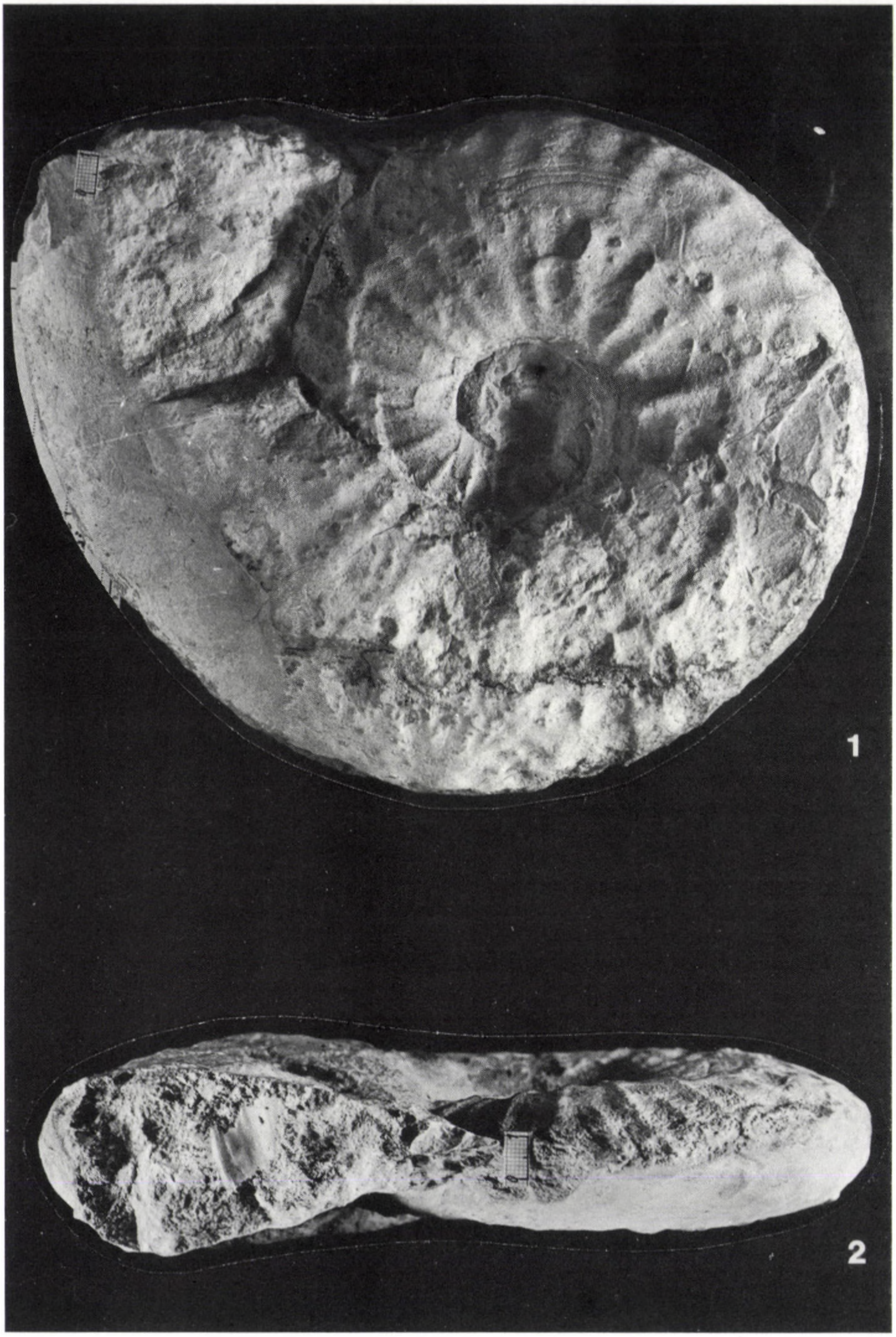
Plate I



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150 years of the Hungarian Geological Society Part II: 1972–1997

Irma Dobos Endre Dudich Vilma Széky-Fux

Introduction

The present paper is the continuation of the one published in the preceding issue of this journal, covering the years 1847–1971. The events of the years 1972–1985 have been compiled by V. Széky-Fux, those of 1986–1995 by I. Dobos; E. Dudich has added the rest.

Consolidated Prosperity (1972–1990)

The year 1972 started with the formal establishment of a *Club of Mineral Collectors*. On 15 March, new officers were elected. *Viktor Dank* became the President and *Géza Hámor* the new Secretary General. An important conference was organized jointly with the Hydrological Society: "Hydrogeological problems in the coal, bauxite and manganese ore bearing areas in Transdanubia" (23 May). The first issue of the *Annals of the History of Hungarian Geology* (Földtani Tudománytörténeti Évkönyv) was printed.

Viktor Dank

Two "international" events of the year were of particular importance for Hungarian geology in general and for the Society in particular. At the 24th International Geological Congress in Montreal, Canada, *Gyula Grasselly* was elected one of the Vice Presidents of IUGS, and the proposal (drafted in Budapest in 1969) to launch an *International Geological Correlation Programme* was accepted (it was ratified in November of the same year by the General Conference of UNESCO in Paris).

1973 saw the 125th anniversary of the Society. At the General Assembly held on 14 March, Secretary General Géza Hámor set two major goals: "to enhance democracy in our scientific life and to pay special attention to interdisciplinary

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research". Occasional papers were published: the proceedings of several specialization courses (construction materials, mathematical geology, sedimentology, and geologic cartography). By that time there were so many foreign visitors that it had to be decided to limit the annual number of presentations by foreigners to five.

The 125th anniversary was celebrated on 25–27 April at the Hungarian Academy of Sciences. In his presidential address, Viktor Dank reviewed the last 25 years of the Society. The full text was published in Hungarian and German (see References). On this occasion three memorial tablets were unveiled (to *J. Szabó, F. Schafarzik* and *A. Koch).* The Itinerant Meeting went to the town of Esztergom which was celebrating its millennium.

1974 was the *Year of Copper:* the Itinerant Meeting was held at *Recsk* in the Mátra Mountains, presenting in much detail the various aspects of the recently discovered deep-seated porphyry copper deposit of Recsk (bound to Eocene andesitic subvolcanism). Other major meetings of the year dealt with environmental protection (at Pécs) and the application of mathematical methods and computer techniques in mineral exploration.

In 1975 V. Dank and G. Hámor were re-elected as President and Secretary General, respectively. This was the *Year of Coal:* the Itinerant Meeting was held in the *Tatabánya* Eocene brown coal basin. It was also in this year that the *Association of European Geological Societies* (AEGS) was founded at *Reading,* U.K. The Hungarian Geological Society was represented by Vice President *Vilma Széky-Fux* and *István Bérczi.* Thus the Society has been a member of the Association from the very beginning.

1976 was the *Year of Water.* The Itinerant Meeting was held at *Kecskemét* in the Danube–Tisza Interfluve, focussing on the hydrogeologic problems of the Great Hungarian Plain (14–15 October). A month later, a National Conference on Tectonics aroused much interest. In the *Földtani Közlöny,* lively discussion went on about how to apply the ideas of plate tectonics to the Pannonian Basin.

1977 began with the first *Day of the History of Hungarian Geology,* dealing with the history of exploration of mineral resources in Hungary prior to 1945. Its full Proceedings were published in English (a remarkable innovation!) by the Földtani Közlöny in 1980. The Itinerant Meeting of the year was held – another important innovation – jointly with the Society of Hungarian Geophysicists, in the Börzsöny Mountains (dealing with the possibilities and methods of polymetallic ore exploration). The *fifth Regional Branch* of the Society was established, for the *Budapest Area.* A *Revision Committee* and a *Committee for Drilling Technology* were set up.

In 1978 the Electoral General Assembly re-elected V. Dank and G. Hámor for a second time. It was commented rather favorably that since 1970 the Society has been accepting and performing consulting work which contributed considerably to its finances. The system of regional thematic conferences was deemed a success. The accent was obviously shifted towards basic problems of economic viability: an inter-society conference was held on the *"Economic Problems of Mineral Resources Management in Hungary"* (20 Oct.) The Itinerant Meeting was held in the south of the Trans-Tisza area. A Paleogeographic Conference was held (9–10 Nov.) The Board of the Society made serious efforts to elaborate a long-term program.

In March 1979 the Geological Society of Cuba was founded, upon the initiative and with the active assistance of Hungarian experts working in Cuba, on the model of the Hungarian Geological Society. At the General Assembly (of Delegates, according to the new regulations) Secretary General G. Hámor stated that in the thematic sections special attention was being paid to the problems of environmental protection. The *IGCP* held an international *Kaolin Symposium* in Budapest. The Itinerant Meeting went to the Mecsek Mountains in southern Transdanubia, reviewing the problems of hard coal exploration. Towards the end of the year, *Sz. Bérczi* held a lecture on the stratigraphy of the Moon (!).

It was in 1980 that the overall membership grew to over 1500 (1594). The highlight of the year was a series of five *Regional Plan Symposia* discussing the medium and long-term development plans of various regions of Hungary.

At the Electoral General Assembly in 1981 V. Dank was re-elected as President, and *I. Bérczi* was elected Secretary General. A *Commission on Ethics* was established. A (Second) National Conference took place on the results and tasks of mineral exploration and a *Register Volume of Földtani Közlöny* was printed (comprising the years 1961–1975).

The centenary of *Gy. Princz*'s birth was celebrated in 1982, jointly with the Hungarian Geographic Society, in Szeged. The annual General Assembly approved the formal cooperation agreements signed with the Society of Hungarian Geophysicists, the National Society of Mining and Metallurgy, and with the Turkish Geological Society. Gy. Grasselly reported to the Assembly about the connection of Hungarian geology with the IUGS. A novelty was the public presentation and judging of nine films dealing with geologic themes.

INHIGEO held its Xth International Symposium in Budapest, on the history of geologic mapping and geologic cartography (16–22 August). The full Proceedings (in English) were published by the Hungarian Academy of Sciences in 1984. The Itinerant Meeting visited SW Hungary (Somogy and Zala counties).

The lectures held at Kaposvár focussed on various problems of environmental geology, engineering (construction) geology and agrogeology. For the first time the Section of Engineering Geology organized a study tour by bus abroad. It was a great success. (Gy. Vitális successfully organized 12 more until 1997.)

In his presidential address delivered at the 1983 General Assembly, V. Dank pointed out the difficult position of Hungarian geology "in the rather unfavorable context of the world economy". The Club of Mineral Collectors became a special group of the Mineralogical-Geochemical Section. Three conferences held are worthy of mention: on the geologic aspects of environmental and nature protection, on the use of computers in geologic research and exploration, and on the utilization of geothermal energy in agriculture. A specialization course on structural geology took place in the town of Miskolc, and it attracted a high number of participants. The Itinerant Meeting was held in Transdanubia, and an ad hoc committee was set up to review the activity of the Society, headed by Gy. Vitális, to recommend purposeful modifications in order to cope with new challenges.

The year 1984 began with the 25th anniversary meeting of the Southern Transdanubian Regional Branch at Pécs. At the General Assembly Secretary General István Bérczi qualified the previous year (1983) as "the beginning of cautious rearrangement". Accent was shifted to the larger-scale meetings of "conference" type and to the specialization courses. In harmony with this statement, in September a Course of Sedimentology was organized at Sümeg. The Itinerant Meeting visited Miskolc.

In January 1985, the ad hoc committee presented a proposal to modernize the activity of the Society. At a commemorative meeting held on the centenary of E. Vadász's birth, J. Fülöp stated that (in Hungary) "the fields of research progressing faster than average are hydrogeology, environmental geology, agrogeology, automation of material testing, computerized data processing, modeling and quantification. Due to the scarcity of funds the importance of economic geology is increasing." President V. Dank declared that *"we have arrived at a phase of development that is market-oriented and market-controlled"*.

It was a special honor for Hungarian geology that *Gy. Bárdossy* was elected Foreign Vice President of the Geological Society of France. Another important international event was the *VIIIth Conference on the Mediterranean Neogene of the RCMNS of IUGS,* held in *Budapest,* with an excursion to the Lower Miocene fossil footprint area of *Ipolytarnóc.*

On 12 March 1986 the Electoral Meeting elected Géza Hámor President and István Bérczi Secretary General. A *"Pro Geologia Applicata"* Memorial Medal and a *Pál Kriván Memorial Fund* were created. On 7–8 May at Szeged a conference was held "On the role and application possibilities of mathematics in the earth sciences". This was the first of a still ongoing series of annual conferences on geomathematics.

Géza Hámor

At the 1987 General Assembly of delegates President Géza Hámor announced a change of orientation, pointing out that the Society has also the task of defining and protecting the interests of its members. The Itinerant Meeting was organized jointly with the Society of Hungarian Geophysicists at Balatonszemes (from this year on this joint venture was to be repeated at least every second year). A field trip went to Yugoslavia. In the field of publications, an innovation was the *first English-language Special Issue of the Annals of the History of Hungarian Geology* (followed successively by five more; see References).

In 1988 a revision of the membership was carried out which resulted in a (not too dramatic) drop of 100. In September, a new *Section of Geomathematics and Computer Science* was established, with György Bárdossy as its President. The Itinerant Meeting was held in the karst & cave area of Aggtelek–Jósvafő.

On 23 January 1989 a special meeting presented and discussed *"The engineering-geologic and hydrogeologic investigation of the Bős–Nagymaros Danube dam system"*. In February, Transylvanian geologists who had emigrated from Romania presented various aspects of the geology of Transylvania. In the same month a meeting dealt with the application of remote sensing to the exploration of metallic and non-metallic mineral resources.

At the 1989 General Assembly (22 March) President G. Hámor made two important points. One of these is the decreasing demand for mineral raw materials, resulting – unduly – in a decrease of geologic research; the other is the enormously increasing demand for a better knowledge of the geologic environment, which in turn requires the multilateral intensification of geologic research. In his Secretarial Report, I. Bérczi pointed out with regret that the Society faced serious financial problems as far as its publications were concerned.

It was in the year 1989 that for the first time a *Day of Geologic Environmental Protection* was held. A remarkable international event was the *21st International Micropalaeontological Colloquium* in Hungary. The (joint) Itinerant Meeting was devoted to problems of the Little Hungarian Plain, with its center in the town of Sopron.

1990 began with a conference on the problems of storage of hazardous waste (27 February). A particularly noteworthy event was the beginning of the posthumous rehabilitation process of *Simon Papp*. He was awarded the highest scientific distinction of Hungary, the *Széchenyi Prize* (on 13 March). This was followed in 1991 by the *V. Zsigmondy Medal* of the National Society of Mining and Metallurgy. Finally, the mortal remains of S. Papp and his wife were transferred to the statue park of the Museum of the Oil Industry at Zalaegerszeg.

The 1990 General Assembly was combined with a national meeting of geologist technicians. The Itinerant Meeting took place at Pécs (20–22 September). – An interesting new event was the Symposium on Basalt Bentonite in October. The year was ended with an extraordinary General Assembly held on 12 December, in order to modify the Statutes of the Society.

Facing the Challenges of the New, Democratic Hungary (1991–)

The "starting shot" for the new era was a talk given by *Paul G. Teleki* of the US Geological Survey on *"Earth Sciences and Market Economy"* on 21 January 1991. (Paul Teleki comes from a distinguished family of the Hungarian–Transylvanian aristocracy. His grandfather, *Pál Teleki*, was a University Professor of Geography and Prime Minister of Hungary in 1940–41. Small wonder that his grandson was invited by the first post-communist government headed by *J. Antall* as an adviser on problems of earth sciences management.) It attracted a record number of participants (270).

Tibor Kecskeméti

The Electoral General Assembly elected *Tibor Kecskeméti* as the new President of the Hungarian Geological Society and *János Halmai* as the new Secretary General. As the new President said: "After the extensive phase of development, an intensive one has to follow. A lack of funds must be complemented by the grey substance (brain)". The new Statutes were accepted, and it was decided to reform the editing procedure of the Földtani Közlöny which had built up a considerable delay of being published. The office of the Society was moved to its present-day site (in Buda, Fő utca 68, I. 102).

The Itinerant Meeting was held at Szeged (jointly with the Society of Geophysicists), with oil and gas exploration as the central topic. Officers of the Society took active part in the work of the *Hungarian Parliament's Commission for Environmental Protection* and in the discussions about the *new Mining Law* which was under preparation. As far as the international scene is concerned, the *First World Meeting of Young Geologists (New Wave)* was held in *Budapest*, 22–24 August. The Central Office of Geology ceased the publication of its journal, the *Földtani Kutatás*, which used to be supplied to all members of the

Hungarian Geological Society. On the other hand a *Foundation for Hungarian Geology* was established (it became operational in early 1992). The membership of the Society reached a relative minimum of 948 in 1991.

On 9 January 1992 an extended Board Meeting was held on a crucial topic: *"Proposal to the Government about the establishment of a Hungarian Institute of Earth Sciences and about the tasks there-of"*. This would have meant the merging of the Central Office of Geology (KFH), the Hungarian Geological Institute (MFI) and the Hungarian Geophysical Institute (ELGI). An animated discussion developed between Paul G. Teleki, Adviser to the Government, György Komlóssy, President of the Central Office of Geology since December 1990, and Gábor Gaál, Director of the Hungarian Geological Institute (Geological Survey of Hungary), which revealed their widely different views on how to restructure Hungarian geology. (Finally the idea of merging was dropped.)

In March the environmental damages caused by the former military objects (airports, barracks, etc) of the Soviet Army and the methods of their restoration were presented and discussed. In this year the General Assembly was held not in Budapest but in Szeged, with particular emphasis on the challenges, possibilities and dangers of the future. In May, the President and the Secretary General in the (earlier monthly, from now on bimonthly) *HÍR-LEVÉL (Newsletter)* of the Society summoned all members to contribute financially to the funds of the Society, in order to avoid bankruptcy in 1993 (the call had a positive and efficient echo).

The Itinerant Meeting was held in the mining town of Salgótarján, reviewing the interrelations of geologic exploration, mining, environmental and nature protection. The ensuing excursion passed the state border to Slovakia. The Mineralogical–Geochemical Section became a member of the *European Mineralogical Union (EMU)*. On 4 November a *Conference on Geoarchaeology* opened new vistas (up-to-date petrologic study of architectural monuments etc.)

The 1993 General Assembly (17 March) was attended and addressed by *László Gyurkó*, the Minister of Environmental and Regional Development. This was a reflection of the fact that an attempt was being made to transfer geology from the competence of the Ministry of Industry to the Ministry of Environment and Regional Development (in the end this did not happen). President T. Kecskeméti outlined a triple strategy: (1) Beside the traditional values, to integrate new ones, (2) to strengthen the publications of the Society, and (3) to promote the advancement of geologic education at all levels. Secretary General J. Halmai pointed out that only the Földtani Közlöny could be published with the funds of the Foundation for Hungarian Geology. The other Hungarian-language publications of the Society would be successively stopped (in fact the last one – the Általános Földtani Szemle, General Geological Review – was stopped definitively only in 1995, the last issue having been financed from a special research fund (OTKA). A lecture presented by János Földessy was more optimistic: "A way out of the crisis: private enterprises in geology". Another positive event was the publication by Gábor Csiky of a Special Issue of the Annals of the History of Hungarian Geology: *"Chapters from the History of the Hungarian Geological Society"* (in English). This work deals with the story until 1949.

The Boards of the Hungarian Geological Society and of the Society of Hungarian Geophysicists held a joint, extended meeting with *István Farkas,* the newly appointed Director General of the Hungarian Geological Service (Survey), created to replace the Central Office of Geology which had been abolished. The President of the Hungarian Geological Society became a member of its advisory Geologic Council, just as he has already been an ex-officio member of the Commission on Geology of the Hungarian Academy of Sciences. Between 19 and 26 September 1993, the *Association of European Geological Societies* held its 8th meeting *(MAEGS-8)* in *Budapest,* followed by three field trips. *Endre Dudich,* the Chairman of the local Organising Committee, became President of the AEGS for two years (until the MAEGS-9 held in Saint Petersburg, Russia, in September 1995). Furthermore, the Hungarian Geological Society became an affiliate of the *European Association of Petroleum Geologists (EUPG).* The possible ways of survival and further development of the Society were discussed at a brain-storming held in November. A month earlier (13–14 October) an international and interdisciplinary conference was held on *"Ferenc Nopcsa and Albania"* in Budapest (Nopcsa was a renowned expert in fossil reptiles, in particular dinosaurs, Director of the Royal Hungarian Geological Institute, and an all-round explorer of Albania in the first quarter of this century).

On 3 March 1994 a bilingual (Hungarian and Slovak) memorial tablet was unveiled on the wall of the former Mansion of the Kubinyi brothers at Videfalva (Vidina), near Lucenec (Losonc) in Southern Slovakia, where the founding meeting of the Hungarian Geological Society was held in January 1848. Speeches were delivered by O. Samuel, President of the Slovak Geological Society, T. Kecskeméti, President of the Hungarian Geological Society, and E. Dudich, President of the Association of European Geological Societies. It is worth mentioning that O. Samuel read his text in Hungarian, E. Dudich his in Slovak, and that of T. Kecskeméti was interpreted into Slovak, in a spirit of friendship and mutual understanding independent of the strained political atmosphere.

An Electoral General Assembly took place on 18 March 1994. The new President was *István Bérczi. J. Halmai* was re-elected Secretary General. The new president stated: "Our Society must be capable of renewing itself... this is the beginning of a genuine revolution." On 26–30 September, the *Paläontologische Tagung* of the German Palaeontological Society was held in Hungary. The Itinerant Meeting, again jointly with the Society of Hungarian Geophysicists, was held at Sárospatak in the Tokaj area. A commemorative meeting was held on the centenary of the death of József Szabó, who was the most outstanding geologist (and President of the Society) in the past century (see in Part I). The Hungarian Geological Society obtained Observer's status in the *European Federation of Geologists (EFG).*

István Bérczi

1995 began with the public presentation of the countrywide screening performed in order to identify geologic sites suitable for the disposal of low and medium level radioactive waste (from the nuclear plant operating at Paks on the Danube). The General Assembly unanimously accepted a statement forwarded by the Commission of Education concerning the National Basic Plan of Education (Nemzeti Alaptanterv, NAT). It was in this year that for the first time a cooperation was established with the Deva Branch of the Rumanian Geological Society: mutual excursions were organized on the topic of ore geology. The Itinerant Meeting held at Alsóörs focussed on the environmental status and problems of Lake Balaton, followed later in the year by a national conference held at Siófok: "Geology for the protection of our natural and man-made environment."

At the 1996 General Assembly Secretary General J. Halmai stated: "the success of our meetings suggests that the situation of our profession is becoming stabilized". Several foreign companies displayed interest in the gold potential of Northern Hungary: a meeting was held summarizing the known possibilities. In the summer of 1996 two important international meetings were held in Hungary. From 9 to 13 June, the 3rd Conference *"Mineralogy and Museums"(M & M),* and from 15 to 22 August *HUNGEO'96:* the First World Meeting of Hungarian Earth Scientists (geologists, geophysicists, geographers and cartographers). The latter was particularly important for the colleagues living in the neighboring countries as members of the Hungarian ethnic minority. The (joint) Itinerant Meeting was held at Kerekegyháza on the Great Hungarian Plain. On 17 October the twenty years of Cuban–Hungarian joint geologic research activities were reviewed. Centenary commemorations were held at Szeged for Sándor Koch, at Sopron for Miklós Vendel, at Tápióság for Károly Papp and – last but not least – with ten years delay – for Aladár Vendl at his birthplace, the village of Ditró (Ditrau) in Transylvania (Rumania).

At the General Assembly of 1997 (19 March) I. Bérczi was re-elected President, and *Géza Császár* was elected as the new Secretary General. A full-day session was devoted on 15 April to the 25 years of geologic mapping activity of the Hungarian Geological Institute in Mongolia (1966–1991) and to the bauxite exploration expedition to Vietnam (1985–1987). On 18 April, a new thematic

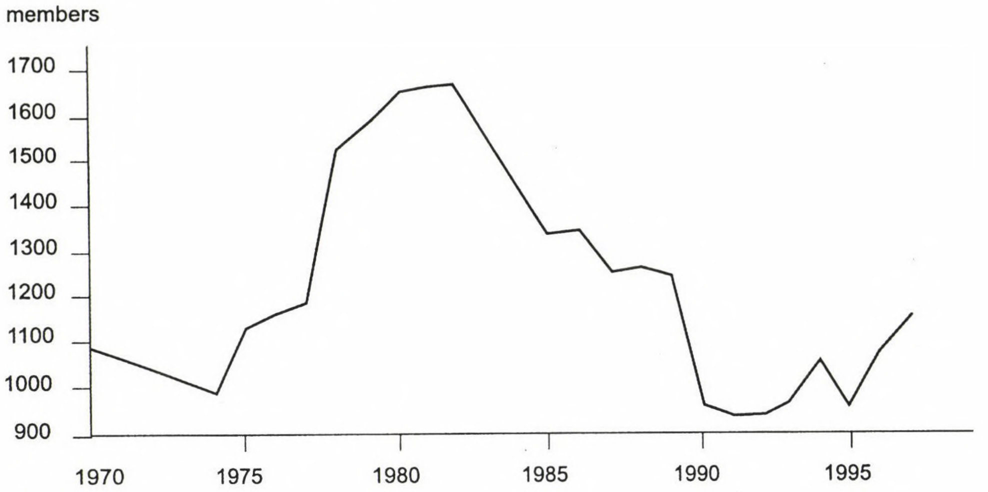

Fig. 1
Dynamics of the membership of the Hungarian Geological Society, 1972–1997

section was established: the *Section of Teaching and Public Education*, with *Miklós Kozák* as its President. In June, a National Conference on the Stability of Abrupt Slopes was held at Paks. It was a significant breakthrough that *GEO 97*, the second meeting of Hungarian earth scientists, was organized by the Transylvanian colleagues at Csíksomlyó, Rumania (21–25 August). On 25–26 September a widely attended conference was organized at Szeged on the role of the geologic environment on regional and urban development. This was the first year in which individual members of the Society had the right to offer one percent of their annual taxes to the benefit of the Society.

On 1 January 1998, the year of its 150th anniversary, the Hungarian Geological Society became a full member of the *European Federation of Geologists*. This would provide its members with the possibility of acquiring the *"Eurogeologist"* title.

The *Sesquicentennial Festive General Assembly* was held in the Geological Institute of Hungary on 18 March 1998. Greetings were presented by the representatives of the Geological Societies of Austria, the Czech Republic, Croatia, Slovakia, Slovenia and Yugoslavia. The Geological Society of London, the German Geological Society and the Rumanian Geological Society have sent greetings in writing. President I. Bérczi summarized the moral conclusion of the one and a half centuries struggle of the Society, and E. Dudich, President of the Section for the History of Hungarian Geology, recalled its main milestones. The Secretarial Report was presented by G. Császár. On this special occasion, silver ammonites were awarded to several distinguished members. Beside several Hungarians, two foreigners were also elected Honorary Members of the Society: Werner Janoschek (Vienna, Austria, Secretary General of the Austrian Geological Society, and Ondrej Samuel, Bratislava, Slovakia, Past

President of the Slovak Geological Society (for earlier foreign Honorary members, see the paper by T. Kecskeméti in the previous issue of this journal). On the 19 March a scientific session was held at the Hungarian Academy of Sciences, presenting the "state of art" in the various disciplines and research fields of the geosciences in Hungary. On 20 March, the GEO 98 meeting of Hungarian Earth Scientists took place at the Geographical Research Institute of the Hungarian Academy of Sciences in Budapest.

The present-day structure of the Hungarian Geological Society is shown in Annex 1.

Acknowledgements

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Annex 1

Organisational Units of the Hungarian Geological Society
(indicating the year of foundation and the chairperson in 1988)

Thematic Sections

1949	Palaeontology (& Stratigraphy)	Hably, Lilla
1960	Clay Mineralogy	Földvári, Mária
1962	Engineering and environmental Geology	Greschik, György
1963	Mineralogy and Geochemistry	Viczián, István
1967	Economic Geology (no more active)	
1969	General Geology	Kovács, Sándor
1970	History of Geology	Dudich, Endre
1970	Mathematical Geology (Reorganised as Geomathematics and Informatics in 1988)	Füst, Antal
1997	Training and Public Education	Kozák, Miklós

Regional Branches

1959	Southern Transdanubia (Pécs)	Benkovics, István
1961	Northern Hungary (Miskolc)	Egerer, Frigyes
1961	Central and Northern Transdanubia (Veszprém)	Kneifel, Ferenc
1966	Great Plain (Szeged)	Pap, Sándor
1977	Budapest	Vörös, Attila

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Evolution trends in basinal deposits of Jurassic–Early Cretaceous age – Examples from the Western Carpathians and the Northern Calcareous Alps

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A common trend occurred in the evolution of Jurassic–Early Cretaceous basinal sequences in the Western Carpathians and the Northern Calcareous Alps. This was expressed by a gradual deepening of the basins since the beginning of the Jurassic until the Middle/Late Jurassic transition when radiolarites were formed at great depths, below or near the CCD. Jurassic radiolarite formation in the Western Carpathians and the Northern Calcareous Alps was unrelated to oceanic rifting. A shallowing trend in the examined basins, from the Upper Oxfordian to the Jurassic/Cretaceous boundary, ended with the Neocimmerian phase of uplift and gravity faulting, best expressed in submarine ridges that separated particular basins.

Key words: Jurassic–Early Cretaceous, basinal development, correlation, West Carpathians, Northern Calcareous Alps

Introduction

This paper discusses evolutionary trends in selected examples of the Jurassic–Early Cretaceous basinal sedimentary successions of the Northern Tethyan realm, in the Inner and Outer Carpathians and the Northern Calcareous Alps (Fig. 1). A characteristic feature of all of these basins, irrespective of their palaeogeographic–paleotectonic position, is the presence of a key radiolarite horizon dated to near the Middle/Upper Jurassic boundary (mainly Oxfordian). This was the time when the North Tethyan oceanic branch reached its greatest depths.

Examples of well-dated Jurassic–Lower Cretaceous rock sequences of the Pieniny Klippen Belt (Klippen and Magura successions) and the Tatra Mts. (High-Tatric and Sub-Tatric successions) will be compared with less well-dated sequences of the Northern Calcareous Alps, at their southern (Baumeckkogel Scale and Göller Nappe – Tyrolicum Tectonic group) and the northern (Reichraming and Ternberg nappes – Bajuvaricum Tectonic Group) margins, respectively.

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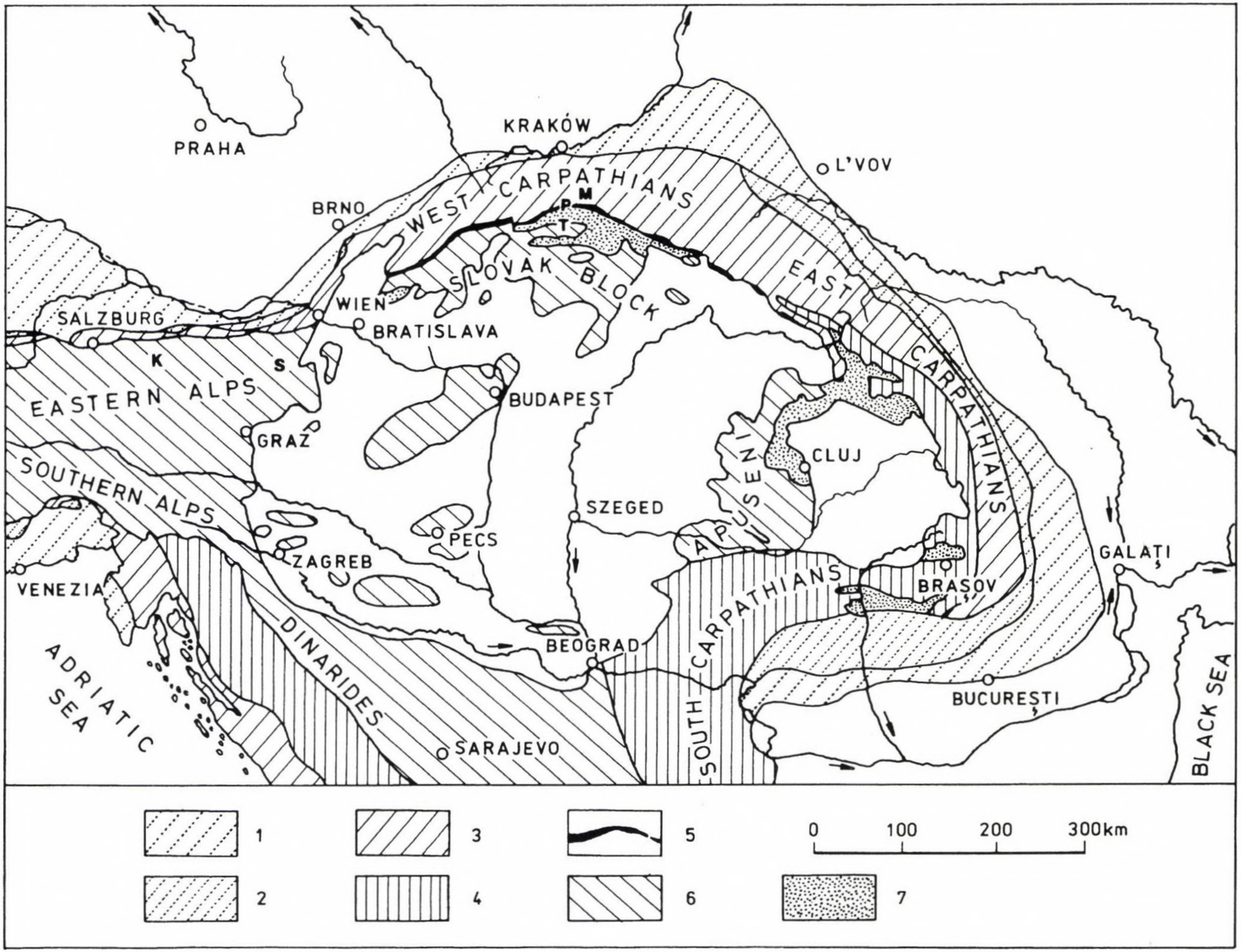
PRAHA
KRAKÓW
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BRATISLAVA
SALZBURG
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EASTERN ALPS
BUDAPEST
GRAZ
CLUJ
APUSENI
SOUTHERN ALPS
SZEGED
PECS
ZAGREB
GALAȚI
BRAȘOV
VENEZIA
SOUTH CARPATHIANS
BEOGRAD
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BUCUREȘTI
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300km

Radiolarites as a key tool for regional correlation of Jurassic oceanic deposits

Deep-water facies and fauna. Radiolarites were already recognized as an important correlative lithostratigraphic unit within the Middle–Upper Jurassic basinal sequences of the Pieniny Klippen Belt (Western Carpathians) in the middle of the present century (Andrusov 1953; Birkenmajer 1953, 1957, 1960). Radiolarites are also widespread in the Alpine–Mediterranean Jurassic (e.g., De Wever and Baudin 1996).

The deep-water character of radiolarites in the Pieniny Klippen Belt is indicated by their microfossil content: mainly siliceous radiolarian plankton, with an admixture of pelagic calcareous organisms such as – probably – pteropods (rather frequent), planktonic crinoid ossicles such as *Saccocoma* (rather frequent), and globigerinid tests (sporadic). The megafossils are represented mainly, or exclusively, by frequent aptychi, rhyncholiths and belemnite rostra – the hardest, calcitic parts of pelagic cephalopods' shells most resistant to dissolution. Thin-shelled, possibly pelagic bivalves (very rare), terebratulids (rare), and echinoids (sporadic), are a subordinate element, mainly in calcareous radiolarites. Poorly preserved ammonite shells are very rare (Sujkowski 1932; Andrusov 1953; Birkenmajer 1958a; Gasiorowski 1958; Birkenmajer and Gasiorowski 1960, 1961; Kwiatkowski 1981).

In the West Carpathians, there is a characteristic association of Jurassic radiolarites with red nodular limestone of the *ammonitico rosso* type (Birkenmajer 1953, 1958a, 1963b, 1977; Lefeld 1981; Lefeld et al. 1985). This condensed pelagic facies is also widely distributed in the Alpine–Mediterranean Jurassic (e.g., Aubouin 1965; Jenkyns 1974).

Mode of origin. A characteristic banding of the radiolarites, expressed in alternations of thicker siliceous and thinner clay or marl bands, is considered to be a primary sedimentary feature. Together with fine-graded bedding, horizontal and cross-lamination, it was an effect of deposition from turbidity currents which sporadically introduced clouds of clay and fine calcium carbonate suspension material that spread out over the deep ocean bottom, interrupting the quiet deposition of siliceous radiolarian ooze (Birkenmajer and

← Fig. 1
Location of the discussed Jurassic–Lower Cretaceous basins in the Western Carpathians and Eastern Alps (Northern Calcareous Alps). 1. Autochthonous Neogene molasse (Foredeep, outer zone); 2. Allochthonous Neogene molasse (Foredeep, inner zone); 3. Tertiary flysch nappes; 4. Mid-Cretaceous nappes in the East and South Carpathians, and comparable folded/thrust outer zones elsewhere; 5. Pieniny Klippen Belt; 6. Mid-Late Cretaceous nappes and comparable folded/thrust and unfolded inner zones elsewhere; 7. Inter-arc post-nappe cover in the Carpathians and the Apuseni Mts. (Paleogene, partly Late Cretaceous); blank – Foreland, intramontane molasse and Tertiary volcanics. Selected Jurassic-Lower Cretaceous basins: K – Kirchdorf a.d. Krems (Ternberg and Reichraming nappes); M – Magura Nappe (Grajcarek Succession); P - Pieniny Mts. (Klippen successions); S – Schwarzau im Gebirge (Göller Nappe, Schwarzau Scales); T – Tatra Mts. (High-Tatric and Sub-Tatric units)

Gasiorowski 1961; Kwiatkowski 1981). For the Oxfordian radiolarites in the Pieniny Klippen Belt the frequency of turbidity currents was calculated at one suspension cloud for every 29,000 years (Birkenmajer and Gasiorowski 1961), or even at one cloud per 100,000 years (Kwiatkowski 1981).

Relation to oceanic rifting and/or volcanism. In many areas of the Alpine–Mediterranean realm the deposition of the *ammonitico rosso* marks zones of simultaneous oceanic rifting (e.g., Cecca et al. 1992). However, the appearance of the *ammonitico rosso* (red nodular limestone) and radiolarite facies in the West Carpathian basins during the Jurassic was unrelated to oceanic rifting. The Mesozoic rifting and oceanic crust formation occurred there much earlier (during the Triassic in the Pieniny Klippen Basin; probably during the Liassic in the Magura Basin) or was totally absent (in the Tatric–Fatric basinal system) – see Birkenmajer (1986). Generally, in these basins, there is also no association of deep-water deposition with synchronous volcanic activity.

An exception to this rule has been found in a variety of the Grajcarek Succession (Magura Basin) at Poiana Botizii in the Romanian Carpathians (Eastern Carpathians). There the deposition of the ?Callovian/Oxfordian radiolarites was simultaneous with an island arc-type volcanism (Bombita and Savu 1986).

Relation to the CCD. Only a part of the Jurassic radiolarites in the Pieniny Klippen Belt are purely siliceous rocks that had formed below the CCD, probably below 3000 m water depth (Birkenmajer 1977). In this case, they represent bathymetrically controlled abyssal radiolarian oozes (cf. Aubouin 1965; Garrison 1974). The calcareous radiolarites (consisting of alternating more siliceous and more calcareous bands) are more frequent. They probably formed at, or slightly above, the CCD of the Jurassic Tethyan ocean (Birkenmajer 1977).

Symmetric vertical sequence. A characteristic symmetry in vertical sequence of pelagic deep-water facies is best shown by the condensed sedimentary columns of the Niedzica Succession (Pieniny Klippen Belt) in Poland (Birkenmajer 1953, 1957, 1977) – Fig. 2. The lower red nodular limestone (Niedzica Limestone Fm.: Upper Bajocian–Callovian ammonite faunas) there is followed successively by: lower red radiolarites, green radiolarites, upper red radiolarites (Czajakowa Radiolarite Fm.), and finally by an upper nodular limestone (Czorsztyn Limestone Fm.: Kimmeridgian–Tithonian, sometimes Kimmeridgian–Valanginian). It seems that the green, usually carbonate-free, radiolarites are the deepest, probably deposited at or below the CCD, when the oceanic trough reached its greatest depths (Birkenmajer 1958a, 1977; Birkenmajer and Gasiorowski 1961).

Biostratigraphic age. Biostratigraphic dating of the radiolarites in the Pieniny Klippen Belt of Poland was based on numerous aptychi (Gasiorowski 1959, 1962; Birkenmajer, Gasiorowski 1960; Durand-Delga and Gasiorowski 1970), and on ammonites derived from underlying and overlying strata (Andrusov 1953; Birkenmajer 1963b). This indicated that the bulk of red and green radiolarites (Czajakowa Radiolarite Fm.) was formed during the Late

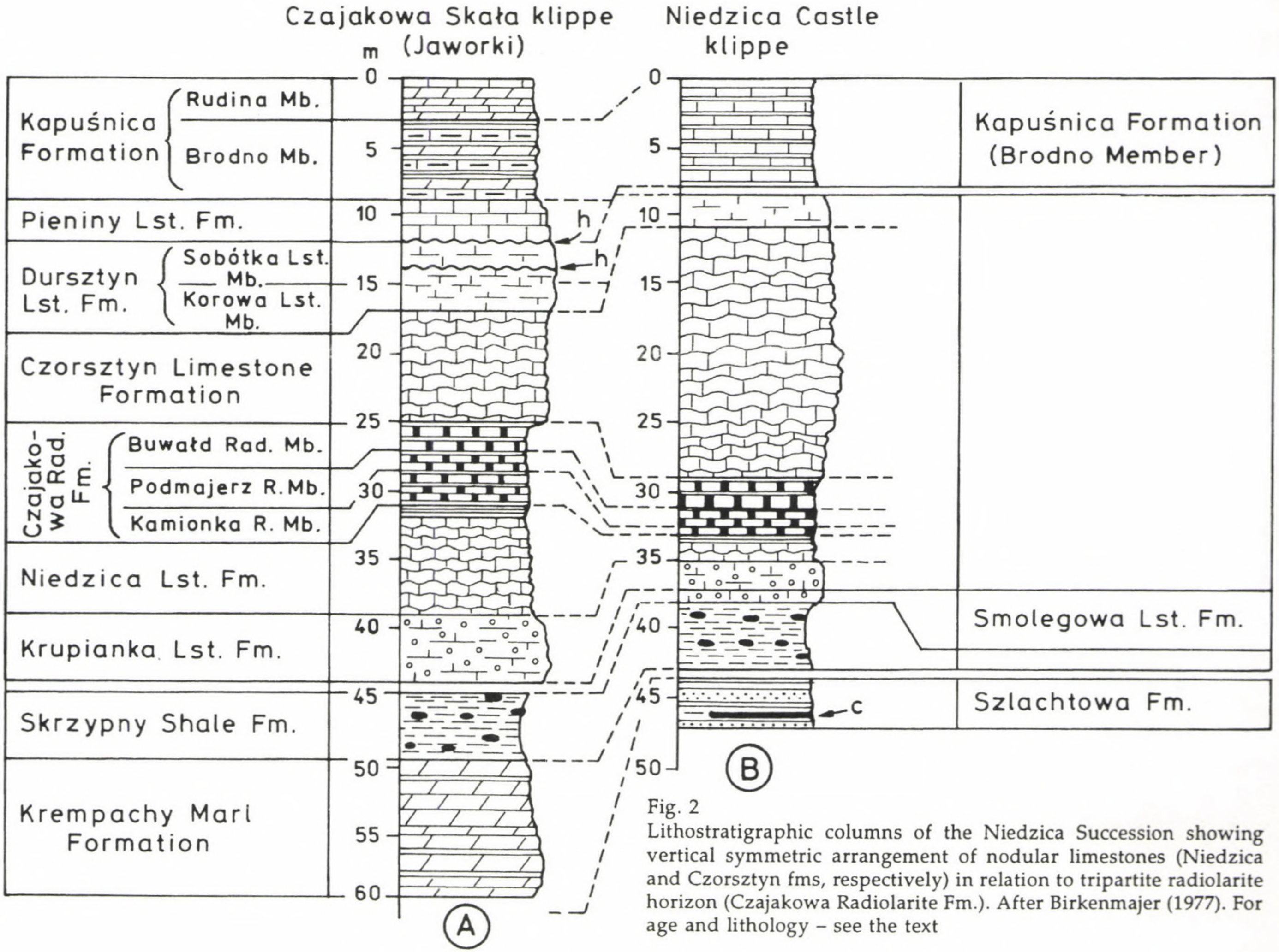

Fig. 2
Lithostratigraphic columns of the Niedzica Succession showing vertical symmetric arrangement of nodular limestones (Niedzica and Czorsztyn fms, respectively) in relation to tripartite radiolarite horizon (Czajakowa Radiolarite Fm.). After Birkenmajer (1977). For age and lithology – see the text

Oxfordian, their total age-range being ?Callovian through Early Kimmeridgian (Birkenmajer 1977).

The age of the radiolarites was refined by a recent study of radiolarian tests. Several biozones correlatable with the standard Late Jurassic Unitary Associations: U.A.7–8 (Oxfordian), U.A.8 (Late Oxfordian), U.A.8–9 (Late Oxfordian–Kimmeridgian), and U.A.9 (Kimmeridgian), have been distinguished in the basinal Klippen successions and in the Grajcarek Succession (Widz 1991, 1992; Widz and De Wever 1993; Birkenmajer and Widz 1995).

In the deepest part of the Pieniny Klippen Basin (Pieniny and Branisko successions) and in the Magura Basin (Grajcarek Succession), banded, spotty, manganese-coated radiolarites (Sokolica Radiolarite Fm.) underlie green and red radiolarites (Czajakowa Radiolarite Fm.). Their age is poorly known so far; it probably corresponds to the Late Bajocian through Early Oxfordian (Birkenmajer 1977; Widz 1991).

Plate-tectonic position. The Jurassic radiolarites in the West Carpathians had formed in several, rather narrow, oceanic basins/troughs separated from each other by intraoceanic ridges. These ridges were splinters of Triassic terrestrial and marine platform underlain by crystalline (mainly Variscan) substratum that were cut off from the North European Platform by the successively northward-relocated oceanic rifts (Fig. 3). Probably only two of the basins, the Pieniny Klippen Basin and the Magura Basin, had narrow strips of oceanic crust, of Early–Middle Triassic and Early Jurassic ages, respectively (Birkenmajer 1986, 1988).

The Pieniny Klippen Basin as a standard stratigraphic reference for the Jurassic radiolarite correlation

In spite of great tectonic deformations recognized in the Pieniny Klippen Belt, its Jurassic through Cretaceous successions, both of basinal and submarine ridge types, are known in considerable detail thanks to biostratigraphic studies of rich megafossil and microfossil assemblages by several generations of Earth scientists since the mid-19th century (see reviews in Andrusov 1931, 1938, 1953; Birkenmajer 1958a, 1963a, b, 1977; Scheibner 1968). The Pieniny Klippen Belt may thus serve as a standard stratigraphic reference for Jurassic–Early Cretaceous correlations with the neighboring basins and ridges of the Central Carpathians (Tatra Mountains: High-Tatric and Sub-Tatric successions) and the Outer Carpathians (Grajcarek Succession), and even with such distant basins and ridges as those of the Northern Calcareous Alps.

Czorsztyn Ridge, southern margin. This was the area of deposition of the Czorsztyn Succession (Fig. 4A) which begins with basinal-type spotty marls and limestones (Krempachy Marl Fm.: Pliensbachian–Aalenian) followed by dark spherosiderite shales (Skrzypny Shale Fm.: Aalenian–Bajocian). No direct,

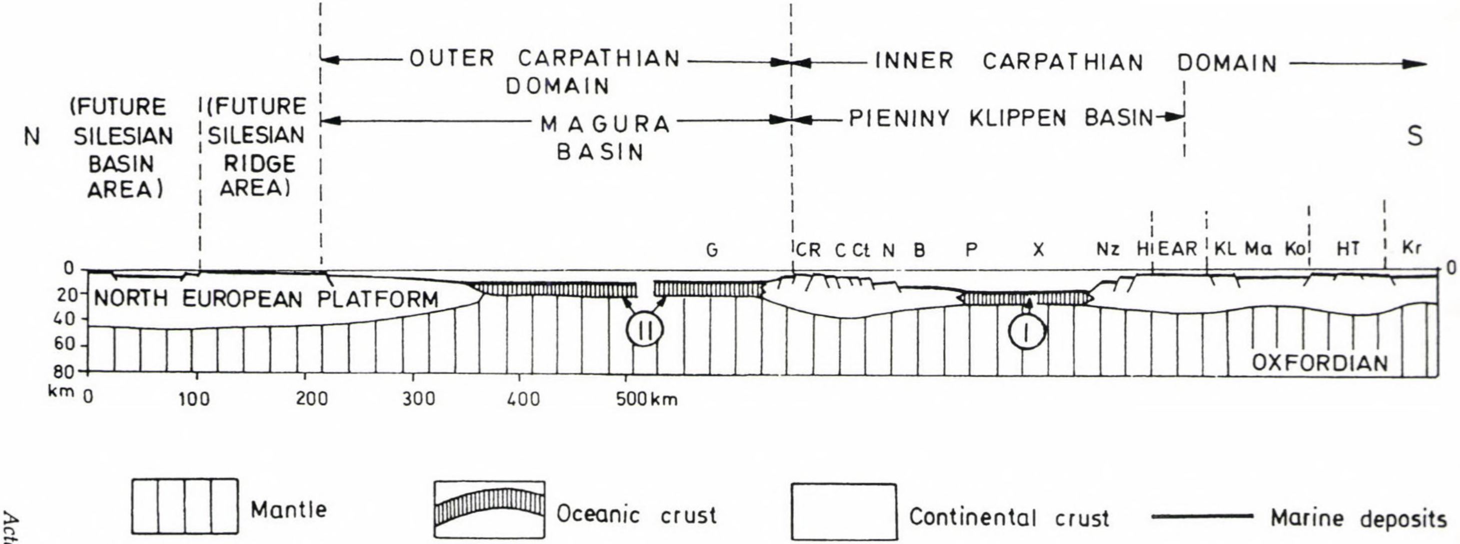

Fig. 3
Palinspastic reconstruction of the Outer and Inner Carpathian basins during the Oxfordian (after Birkenmajer 1988). I – Triassic oceanic crust in the Pieniny Klippen Basin; II – Lower Jurassic oceanic crust in the Magura Basin; G – Grajcarek Succession; CR – Czorsztyn Ridge; C – Czorsztyn Succ.; N – Niedzica Succ.; Ct – Czertezik Succ.; N – Niedzica Succ.; B – Branisko Succ.; P – Pieniny Succ.; X – Hypothetical Ultra-Pieniny Succ.; Nz – Nižná Succ.; H – Haligovce Succ.; EAR – Exotic Andrusov Ridge; KL – Klape Succ.; Ma – Manín Succ.; Ko – Kostelec Succ.; HT – High-Tatric Succ.; Kr – Križna (Sub-Tatric) Succ.

pre-Liassic stratigraphic basement is known; the whole succession was uprooted from its probably Triassic base during the Tertiary folding.

A short-lived but thick carbonate wedge consisting of massive white crinoid limestone (Smolegowa Limestone Fm.) was laid down upon spherosiderite shales during the Bajocian. This wedge soon became drowned to shelfbreak/slope depths, when hematite-rich, bedded crinoid limestone (Krupianka Limestone Fm.: Bajocian–Bathonian), followed by red, nodular, *ammonitico rosso*-type limestone (Czorsztyn Limestone Fm.: Bathonian or Callovian–Lower Tithonian) had formed.

In a more northern zone of the Czorsztyn Succession, the appearance of deep-water sedimentary breaks, hematite hardgrounds and limestone breccias during the Bajocian–Bathonian may be an expression of submarine faulting during the Mesocimmerian phase (Birkenmajer 1963b). Synsedimentary faulting in this zone locally continued through Callovian and Oxfordian, and was repeated during Kimmeridgian–Tithonian times (Birkenmajer 1958b; Mišík 1979; Mišík and Sýkora 1993).

A gradual shallowing which began during the Kimmeridgian continued through the Jurassic/Cretaceous transition. A complex horst-and-graben structure was formed as a result of sea-bottom uplift associated with submarine gravity faulting (Neocimmerian phase). Omission surfaces, deep-water hardgrounds and neptunian dykes developed on the horsts, while organogenic limestones and resettled fault breccias accumulated in the grabens (Birkenmajer 1958b, 1963a). A strongly diversified succession of organogenic calcareous deposits (red and white subpelitic calpionellid limestones, ammonite and brachiopod coquinas, crinoid limestones) and limestone sedimentary breccias (Dursztyn Limestone and Lysa Limestone Fms: Tithonian–Berriasian), were followed by hematite-red, bedded crinoid limestones (Spisz Limestone Fm.: Valanginian–Hauterivian).

A submarine hiatus of variable duration (Aptian, or Aptian through Middle Albian) separates the Spisz Limestone from red Albian foraminiferal marls and limestones (Chmielowa Fm.). The succeeding, Upper Albian, spotty, banded, foraminiferal limestones and marls (Pomiedznik Fm.) begin the pelagic marl deposition stage which continued into the Campanian (and even until Early Maastrichtian in the Czorsztyn Succession).

In the northern part of the succession, this hiatus may be even longer, representing the Aptian through times the Middle Cenomanian (Birkenmajer 1963a, 1977; Birkenmajer and Jednorowska 1987).

Basinal successions. There are a number of basinal successions characterized by the presence of key radiolarite horizons at the Middle/Late Jurassic boundary. Some of these, like the Kýsuca and the Branisko successions, are common to the entire Pieniny Klippen Belt in the Carpathians. The others, like the Czertezik Succession, are common to the Polish and West Slovak sectors of the belt only. Still others have an even more restricted distribution: the Niedzica Succession occurs only in the central, Polish/Slovak sector of the belt, while the Pruské and the Podbiel successions are encountered in West Slovakia only. All the above basinal successions are treated as transitional ones with respect to the deepest one (the Pieniny Succession).

In the Polish sector of the Pieniny Klippen Belt, the Czertezik and Niedzica successions occupied deeper and deeper parts of the southern slope of the Czorsztyn Ridge. Originally they were probably underlain by a thin Triassic platform cover and attenuated continental crust. Still farther south, the Branisko-Kýsuca and the Pieniny successions which occupied an axial part of the basin (see Fig. 3) probably had only a very thin continental-crust base with or without Triassic sedimentary cover. The Jurassic radiolarite horizon increases in thickness in the same basinward direction (Figs 4B, 5).

The Jurassic–Cretaceous basinal Klippen successions were uprooted as nappes from their basement during Late Cretaceous folding. In the axial part of the basin ("Ultra-Pieniny Succession" – "X" in Fig. 3), this basement (as reconstructed from exotic pebbles found in Upper Cretaceous conglomerates) might include the Upper Triassic pelagic limestones and a Lower-Middle Triassic oceanic crust. During the Cretaceous, a part of the basement was obducted and formed an accretionary prism in front of the Andrusov Cordillera. The oceanic crust was consumed in the subduction zone during the latest Cretaceous (Birkenmajer 1986, 1988; Birkenmajer et al. 1990).

In the Czertezik and Niedzica successions, which were deposited on a wedge-shaped southern slope of the Czorsztyn Ridge (see Fig. 3), the sedimentary sequence begins with thin black flysch (Szlachtowa Fm.: Toarcian–Aalenian) that is unknown from the Czorsztyn Succession. Further up, from the Aalenian through Bajocian–Bathonian, the succession generally follows the model of the Czorsztyn Succession: Krempachy Marl Fm., Skrzypny Shale Fm., Smolegowa Limestone Fm. and Krupianka Limestone Fm.

Differentiation between these successions started during the Bajocian. In the Czertezik Succession, radiolarites of the Czajakowa Formation rapidly replace grey crinoid limestones with cherts (Flaki Limestone Fm.) and are followed in turn by red nodular limestone (Czorsztyn Limestone Fm.: Kimmeridgian) and by white, banded cherty limestones (maiolica/biancone facies, Pieniny Limestone Fm.: Tithonian–Barremian). In the Niedzica Succession, which probably was laid down on a subordinate submarine horst (formed during Mesocimmerian submarine bottom uplift and faulting), condensed pelagic deposition took place: thin red crinoid limestone (Krupianka Limestone Fm.: Bajocian) is followed by thin red nodular limestone (Niedzica Limestone Fm.: Bajocian–Callovian), and then by the Czajakowa Radiolarite Formation (Oxfordian) showing typically developed symmetric tripartite subdivision (see Fig. 2). Red nodular limestone (Czorsztyn Limestone Fm.: Kimmeridgian, sometimes Kimmeridgian–Valanginian) appears again above the radiolarites.

In the Branisko and Pieniny successions, the deep basinal sequence begins with Liassic (Pliensbachian) spotty limestones and marls (Fleckenkalk–Fleckenmergel facies).

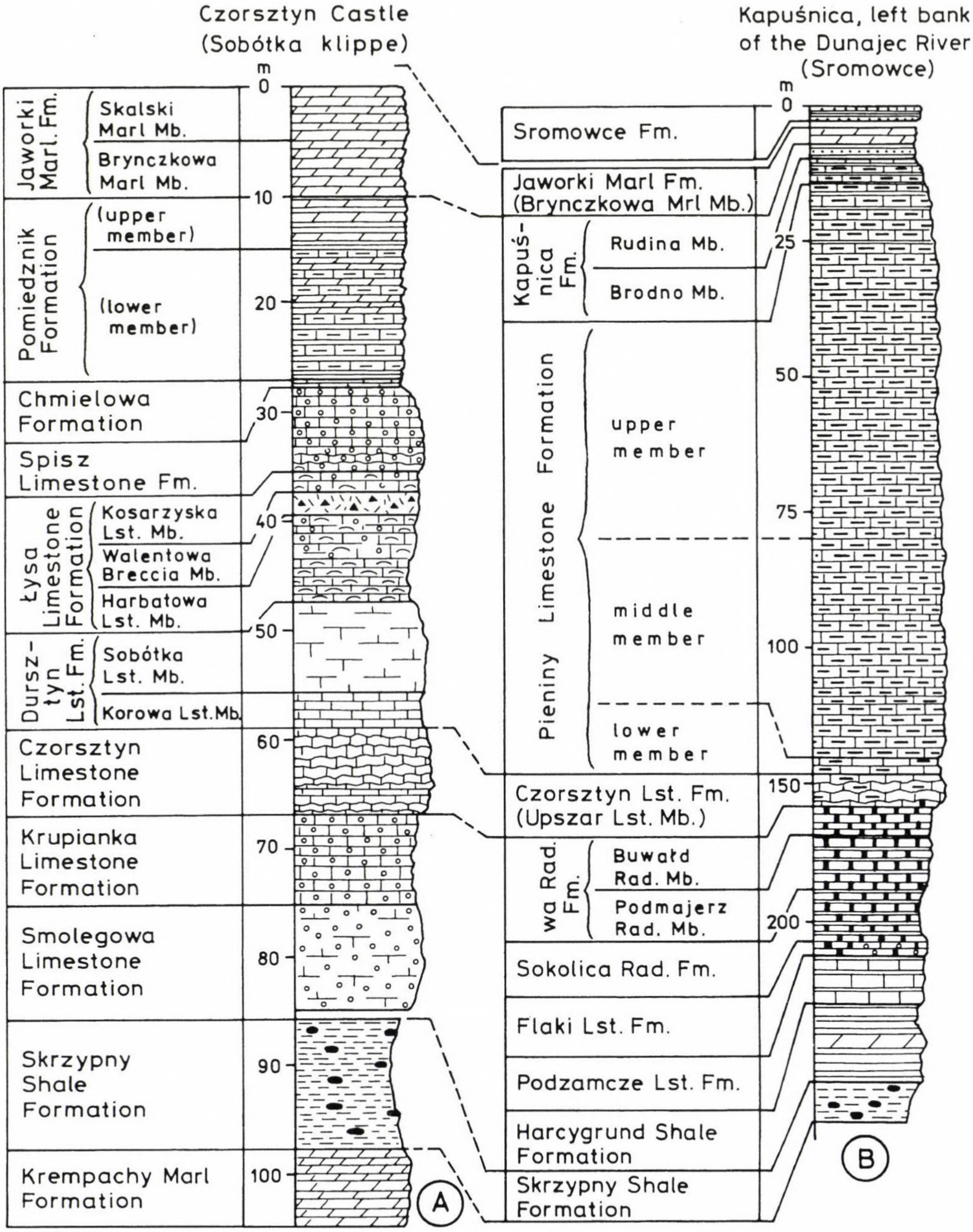

Fig. 4
Lithostratigraphic columns of the Czorsztyn Succession (A) which was deposited on an intraoceanic ridge, and the Branisko Succession (B) – deposited in an oceanic trough (after Birkenmajer 1977). For age and lithology – see the text

Thin black flysch deposits (Szlachtowa Fm.: Toarcian–Aalenian) appear locally at the base of the Branisko Succession in its northern part. They underlie thin spotty marls (Krempachy Marl Fm.: Lower–Middle Aalenian) which are followed by spherosiderite shales (Skrzypny Shale Fm.: Upper Aalenian), black *Bositra* shales (Harcygrund Shale Fm.: Bajocian), and again by spotty limestones and marls (Fleckenkalk–Fleckenmergel facies, Podzamcze Limestone Fm.: Bajocian).

Grey crinoid limestone with spongolite cherts, locally with chamosite and phosphate nodules (Flaki Limestone Fm.: ?Upper Bajocian–Callovian), appears at the transition of the Branisko to the Pieniny succession. Its deposition reflects a short-lived event of positive movements of the sea bottom possibly associated

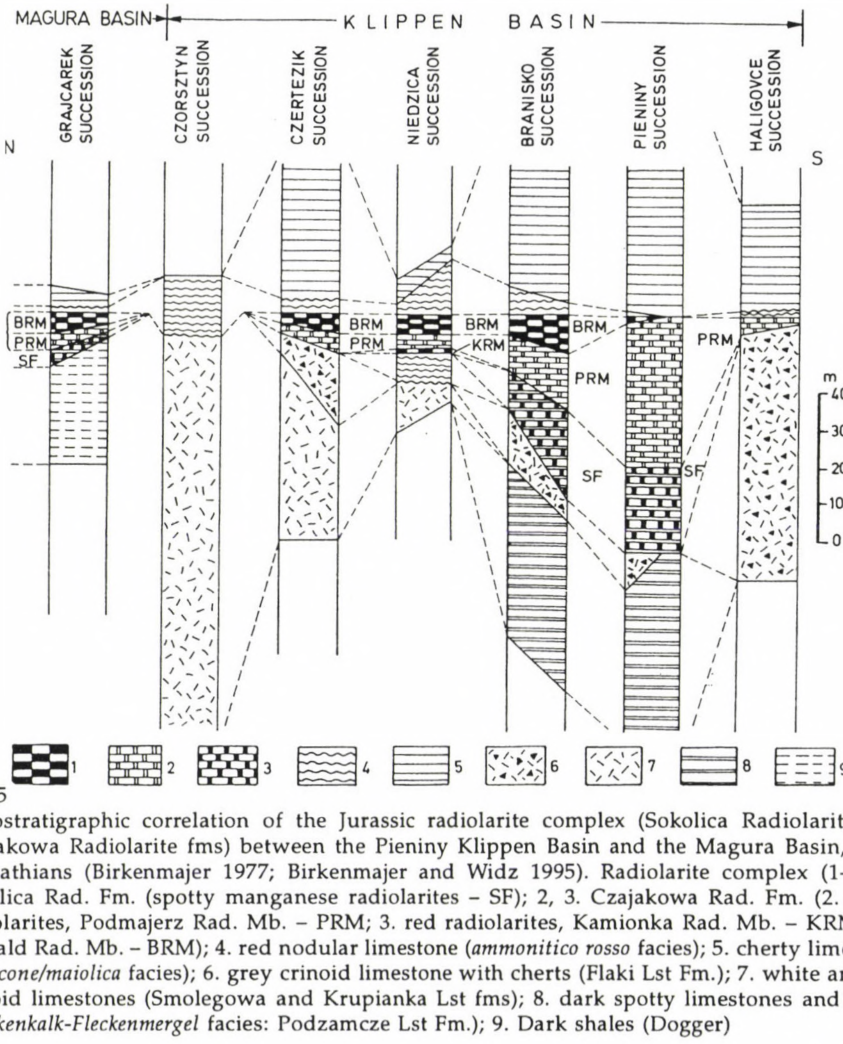

Fig. 5

Lithostratigraphic correlation of the Jurassic radiolarite complex (Sokolica Radiolarite and Czajakowa Radiolarite fms) between the Pieniny Klippen Basin and the Magura Basin, West Carpathians (Birkenmajer 1977; Birkenmajer and Widz 1995). Radiolarite complex (1–3): 1. Sokolica Rad. Fm. (spotty manganese radiolarites – SF); 2, 3. Czajakowa Rad. Fm. (2. green radiolarites, Podmajerz Rad. Mb. – PRM; 3. red radiolarites, Kamionka Rad. Mb. – KRM and Buwald Rad. Mb. – BRM); 4. red nodular limestone (*ammonitico rosso* facies); 5. cherty limestone (*biancone/maiolica* facies); 6. grey crinoid limestone with cherts (Flaki Lst Fm.); 7. white and red crinoid limestones (Smolegowa and Krupianka Lst fms); 8. dark spotty limestones and marls (*Fleckenkalk-Fleckenmergel* facies: Podzamcze Lst Fm.); 9. Dark shales (Dogger) with submarine faulting (Mesocimmerian phase). Abrupt replacement of these limestones by banded, manganese-coated, spotty radiolarites (Sokolica Radiolarite Fm.) indicates a rapid drop of the basin bottom, possibly along a submarine fault scarp. Thereafter the basin entered its deepest stage comparable with abyssal radiolarian-ooze plains of the modern oceans. Deposition of a thick radiolarite complex (Czajakowa Radiolarite Fm.) followed: green, carbonate-free radiolarites (Oxfordian–Kimmeridgian in a southern variety of the Pieniny Succession; Oxfordian in the Branisko Succession) are overlain (in the Branisko Succession, and in a northern variety of the Pieniny Succession) by red, usually calcareous, radiolarites (Upper Oxfordian).

A shallowing trend is shown by the appearance of a thin nodular limestone (Kimmeridgian) in the Branisko–Kýsuca successions, and in a northern variety of the Pieniny Succession. There follow white, massive, subpelitic calpionellid limestones, being soon replaced by a thick complex of well-bedded, white pelagic limestones with black cherts (biancone/maiolica facies, Pieniny Limestone Fm.: Tithonian–Barremian). Dark shale interlayers and dark spots appear in the upper part of the limestone complex which passes upwards into black and green, spotty foraminiferal marls and shales (Aptian–Albian).

This stage of temporarily poorer aeration in the bottom waters (anoxic event) could be a result of basin reorganization and shortening caused by subduction-related contraction. A uniform pelagic deposition of green, variegated and red foraminiferal marls (scaglia/couches rouges facies, Jaworki Formation: mainly Upper Albian–Santonian) that followed was interrupted from time to time by siliciclastic turbidite deposition (flysch) filling erosional submarine canyons. During the Campanian the marls were totally replaced by the flysch (Sromowce Fm.); the clastics were supplied from southern sources (Andrusov Cordillera) save for the northern part of the Czorsztyn Ridge where red pelagic marl deposition continued until Early Maastrichtian (Birkenmajer and Jednorowska 1987).

The subduction of the Triassic oceanic crust under the Andrusov Cordillera continued through the Late Cretaceous, eventually resulting in total consumption of the crust and in basin closure by nappe-folding at about the Cretaceous/Paleogene transition (Birkenmajer 1986, 1988).

Southern margin of the basin. In the present author's opinion the southern margin of the Pieniny Klippen Basin is represented by the Haligovce Succession (Birkenmajer 1959, 1977, 1986, 1988) and its equivalent – the Nižná Succession (Scheibner 1967). The Jurassic of the Haligovce Succession, lying unconformably upon Lower Triassic carbonates, begins with Liassic crinoid and massive limestones, with subordinate sandstone and shale. They are followed by massive crinoid-cherty limestone (Flaki Limestone Fm.: Bajocian–Callovian) resembling that of the Czertezik Succession. The radiolarites (Czajakowa Radiolarite Fm.) are 2–4 m thick only, and are underlain and overlain by equally thin variegated limestones (Czorsztyn Limestone Fm. – see Fig. 5). This resembles the symmetric sequence of strata in the Niedzica Succession (see Fig.

2) which developed on the opposite, northern margin of the Klippen Basin (see Fig. 3). Higher up follows a rather thick, white, cherty limestone (Pieniny Limestone Fm.: Tithonian–Barremian) passing upward into massive, organogenic, Urgonian-type limestone (Haligovce Limestone Fm.: Barremian–Aptian). The latter limestone, known also from the Nižná Succession, indicates proximity of an Urgonian carbonate platform that once separated the deep Pieniny Klippen Basin from the shallower Manín Basin (see Fig. 3). There follow Upper Cretaceous marls and flysch deposits comparable with those of the remaining Klippen successions.

It should be mentioned that the Haligovce and the Nižná successions are treated differently by some Slovak geologists (e.g., Michalík 1994) who attribute these successions to the Manín Unit.

Correlation with Inner Carpathian units: High-Tatric and Sub-Tatric Successions in the Tatra Mts

High-Tatric Succession. A number of Jurassic facies and their sequence recognized in the High-Tatric Succession closely resemble those known from the Pieniny Klippen Belt, particularly of the Czorsztyn and Czertezik successions. This was the reason that many formal names for the Middle and Upper Jurassic lithostratigraphic units derived from the Pieniny Klippen Belt (Birkenmajer 1977) were introduced to the High-Tatric Succession in the Tatra Mts. by Lefeld et al. (1985).

The type High-Tatric Succession in the Tatra Mts. lacks Jurassic radiolarites; its Kimmeridgian limestone (Czorsztyn Limestone Fm.) is less typical *ammonitico rosso* than that of the Czorsztyn Succession. A well-developed organogenic Urgonian limestone (Wysoka Turnia Limestone Fm.: Barremian–Aptian) occurs in the High-Tatric Succession; this limestone did not develop in the Czorsztyn Succession and is less typically developed in the Haligovce and Nižná Successions of the Klippen Belt.

Sub-Tatric Succession (Križna Nappe). A closer analogy to the Jurassic facies development and sequence may be found among basinal developments of the Pieniny Klippen Belt and the Križna Nappe (Fatric unit) since the Middle Jurassic. This is best shown by the occurrence in the Krizna Nappe of the Sokolica Radiolarite Formation followed by a symmetric succession of: red nodular limestone (Niedzica Limestone Fm.), red and green radiolarites (Czajakowa Radiolarite Fm.), and reddish nodular and platy limestones (Czorsztyn Limestone Fm.) – see Lefeld (1981), Lefeld et al. (1985). The maiolica/biancone facies (Pieniny Limestone Fm.: Tithonian–Berriasian) is usually more marly than in the basinal successions of the Klippen Belt. There follows a thick sequence of marly beds with allodapic limestone and thin sandstone intercalations (Koscieliska Marl Formation: Lefeld et al. 1985) with calpionellids indicating Berriasian through Valanginian ages (Pszczólkowski

1996). This development differs considerably from the age-equivalent rocks in the Pieniny Klippen Belt.

Correlation with the Outer Carpathians: Magura Basin

Grajcarek Succession. The Grajcarek Succession was originally deposited in a southern part of the Jurassic–Cretaceous Magura Basin of the Outer Carpathians, along the northern margin of the Czorsztyn Ridge. It was folded and accreted to the Laramian orogen of the Pieniny Klippen Belt during the Late Cretaceous as a result of partial subduction of its supposedly Lower Jurassic (pre-Toarcian) oceanic crust. The Grajcarek Unit extends along the Pieniny Klippen Belt from Poland as far east as Poiana Botizii in Romania (Birkenmajer 1977, 1986, 1988; Bombita and Savu 1986).

Thick black shaly-marly flysch (Szlachtowa Formation: Toarcian–Aalenian) forms the base of the unit, a common feature of some basinal Klippen successions. Locally, originally closer to the northern margin of the Czorsztyn Ridge, shallow-marine ostreid and oolitic ferruginous limestones probably form an intercalation in the upper part of the flysch (Birkenmajer and Tyszka 1997). Marly shales with ferruginous concretions (Opaleniec Fm.: Aalenian–Bajocian) follow upon the flysch beds, while spotty limestones (Fleckenkalk–Fleckenmergel facies, Harcygrund Limestone Fm.: Bajocian) and the *Bositra* shales occur only exceptionally.

Starting from the Middle/Upper Jurassic boundary there is a full analogy in facies development and their sequence between the Grajcarek and the Branisko successions. Spotty manganese radiolarites (Sokolica Radiolarite Fm.) are well developed in both successions. They are followed by equally well-developed green and red radiolarites (Czajakowa Radiolarite Fm.) dated by radiolarians at the Upper Oxfordian–Lower Kimmeridgian (Widz 1992; Birkenmajer and Widz 1995). A thin red marl band with aptychi (Palenica Marl Mbr. of the Czorsztyn Limestone Fm.: Kimmeridgian–Lower Tithonian) separates the red radiolarites from thin Pieniny Limestone (Tithonian–?Barremian). The latter is overlain by thin spotty marls (Kapusnica Fm.: Aptian), similarly to the basinal successions of the Klippen Belt.

The Kapusnica Formation is the last unit common to the Grajcarek and the Klippen successions. The remaining Lower and Upper Cretaceous lithostratigraphic units of the Grajcarek Succession are typically Outer Carpathian in their facies development and sequence (Birkenmajer 1963b, 1965, 1977).

An interesting variety of the Grajcarek Succession occurs at Poiana Botizii, Romania (East Carpathians). Beneath nodular limestones and aptychus shales (Kimmeridgian–Lower Tithonian, equivalent to the Czorsztyn Limestone Fm.) there occurs a thin band of detrital limestone (probably Oxfordian) with ophiolite fragments. It is underlain by red and green radiolarites (Oxfordian–?Callovian, Czajakowa Radiolarite Fm.) with an intercalation of limestone containing volcanic detritus. The radiolarites rest upon basaltic or basaltic-andesite tuffs 2–9 m thick which are underlain by acidic tuffaceous shales up to 3 m thick. These volcanics were attributed by Bombita and Savu (1986) to the Callovian, however without palaeontological evidence.

This is the only locality in the Grajcarek Succession where the formation of Jurassic radiolarites seems to be directly associated with simultaneous volcanic activity. It might indicate an early (Callovian–Oxfordian) activation of the subduction zone at the Magura Basin/Czorsztyn Ridge border and appearance of subductional calc-alkaline (andesitic) island arc-type volcanism within the eastern termination of the ridge. Ophiolite fragments which occur in detrital limestone overlying the radiolarites could derive from obducted older (Early Jurassic) oceanic crust of the Magura Basin.

Correlation with Northern Calcareous Alps

Tyrolicum Tectonic Unit in Schwarzau im Gebirge (Lower Austria)

Tectonic units. The area to the north of the Schneeberg–Rax massif (map sheet ÖK 74 Hohenberg, Lower Austria – Figs 1, 6) represents a contact zone between the Juvavic (Juvavicum) and the Tyrolic (Tyrolicum) group of nappes of the Northern Calcareous Alps (see, e.g., Tollmann 1976, 1987; Janoschek and Matura 1980; Oberhauser 1980). The following tectonic units have been recognized (see Cornelius 1951; Lobitzer et al. 1990; Birkenmajer 1993, 1996a) – Fig. 6.

(4) The Schneeberg Nappe (higher Juvavicum) occupies the highest tectonic position. In the area south of Schwarzau im Gebirge it was thrust from the south directly upon the Schwarzau Scales which represent the dismembered southern part of the Göller Nappe (higher Tyrolicum);

(3) The Mürzalpen Nappe (lower Juvavicum) occupies an intermediate position between (4) and (2). It emerges from below the Schneeberg Nappe to the west and east of the Schneeberg-Rax massif but wedges out in the area of Schwarzau im Gebirge;

(2) The Schwarzau Scales (highest Tyrolicum) represent the innermost (southernmost) part of the Göller Nappe dismembered in front of the Schneeberg Nappe into three large tectonic scales (from north to south): (2.1.) the Baumeckkogel Scale; (2.2.) the Mitterriegel Scale-2; (2.1.) the Mitterriegel Scale-1;

(1) The Göller Nappe (Ötscher Nappe Group, higher Tyrolicum) is the lowest unit in the area of Schwarzau im Gebirge.

The Jurassic–Lower Cretaceous strata are present in the Baumeckkogel Scale (2.1.), and in the Göller Nappe (1), but are missing from the Mitterriegel scales (2.2. and 2.1.).

Baumeckkogel Scale. In the Baumeckkogel Scale the Jurassic is represented by a very condensed, strongly tectonized and often brecciated basinal sequence, maximally 80 m thick (Fig. 7A).

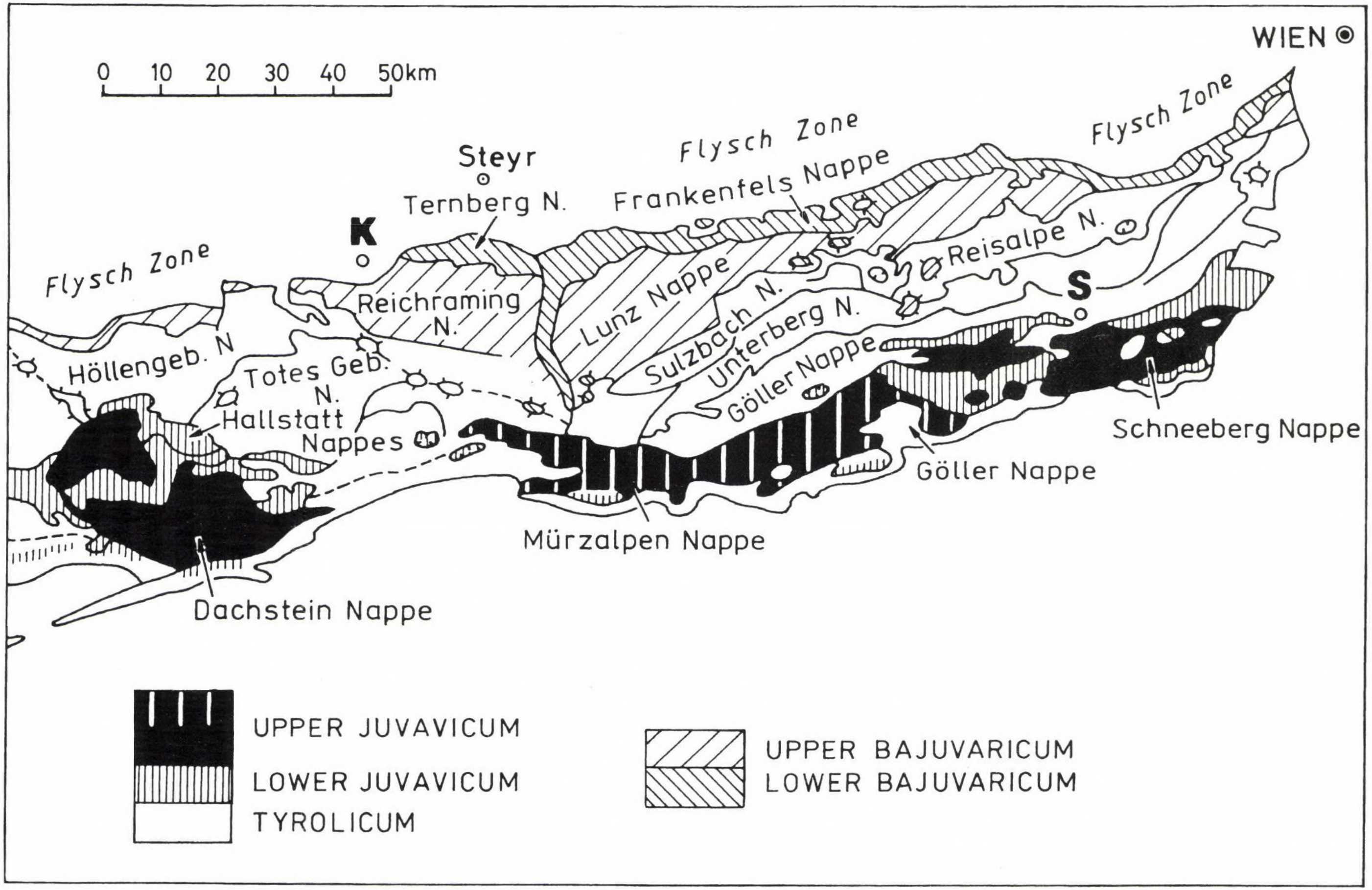

Fig. 6
Distribution of mid-Cretaceous nappes of the Northern Calcareous Alps (adapted from Tollmann 1987). K – Kirchdorf an der Krems; S – Schwarzau im Gebirge

Directly above the Dachstein Limestone (Dachsteinkalk: Norian) platform or the thin dark limestone of the Kössen Formation (Kössenerschichten: Rhaetian) there rests without interruption (sedimentary hiatus) deep-water, condensed, Fe-Mn-enriched, massive limestone (Hierlatzkalk, probably Lower Liassic) about 10 m thick at most. The following sequence of strata was recognized: deep-red, hematite- and manganese oxide-rich limestone at the base, passing upward into hematite-manganese ore (2 m); irregularly bedded, sometimes crinoidal, limestone with intercalations of hematite-manganese ores 1–2 cm thick (3 m); grey massive limestone with black manganese spots (1 m); red, irregularly bedded, slightly nodular hematitic limestone (at least 2–3 m thick). There follow grey to black spotty limestone and marl (Allgäuschichten: Fleckenkalk–Fleckenmergel facies, higher Liassic) associated with grey to black crinoid limestone (Liassic or Dogger) known exclusively from fragments in the scree. The thickness of the unit varies between 10 and 30 m.

The next unit is represented by a characteristic but rarely well exposed radiolarite complex (Kieselkalk) consisting of two units. Green banded radiolarites and siliceous limestones 1–5 m thick occur at the base and are overlain by red radiolarites and siliceous limestones 1–2 m thick. These radiolarites are a close equivalent to the Czajakowa Radiolarite Formation (Oxfordian) in the Pieniny Klippen Belt (Klippen and Grajcarek successions) and the Križna Basin in the West Carpathians (see Fig. 4B).

The radiolarites are followed by red, slightly nodular or crinoidal limestone (Klauskalk), up to 2 m thick (equivalent to the Kimmeridgian Czorsztyn Limestone Fm.), and that by light-greenish to grey-whitish to white, massive to poorly bedded, slightly siliceous limestone, maximally 30 m thick. The latter corresponds to the Pieniny Limestone Formation (Tithonian–Neocomian) of the West Carpathians. No younger strata are known from the Baumeckkogel Scale.

Göller Nappe. In the Göller Nappe the Liassic through Kimmeridgian succession of marine strata (Fig. 7B) is similar to that of the Baumeckkogel Scale. The Jurassic also begins there with deep-red to pink, often crinoid-bearing limestone (Hierlatzkalk, Lower Liassic) up to 10 m thick, resting with a sedimentary hiatus (lack of the Rhaetian) upon the Triassic carbonate platform limestones (Dachsteinkalk: Norian). The succeeding dark-grey spotty limestones (Allgäuschichten: Fleckenkalk–Fleckenmergel facies, higher Liassic), 10–20 m thick, are sometimes associated with greenish spongolite cherts (?Upper Liassic–Dogger?). There follows a symmetric sequence of radiolarites and red, often nodular limestones: lower red limestone (about 3 m) with lenses of red chert in the uppermost part (1 m); lower red radiolarites alternating with red cherty limestone (2 m); green radiolarites (1 m); upper red radiolarites (0.3 m); red siliceous limestone with thin intercalations and lenses of red radiolarite (0.2 m). This sequence corresponds in facies, and probably also in age, to that of the Niedzica (see Fig. 2) and Križna successions in the West Carpathians: Niedzica Limestone Fm. (Dogger) – Czajakowa Radiolarite Fm. (Oxfordian) – Czorsztyn Limestone Fm. (Kimmeridgian).

Thereafter appears a complex of limestones at most about 320 m thick (Plassenkalk *vel* Falkensteinkalk: Tithonian and Lower Neocomian) which forms a gently folded plateau in the southern part of the Göller Nappe at Schwarzau

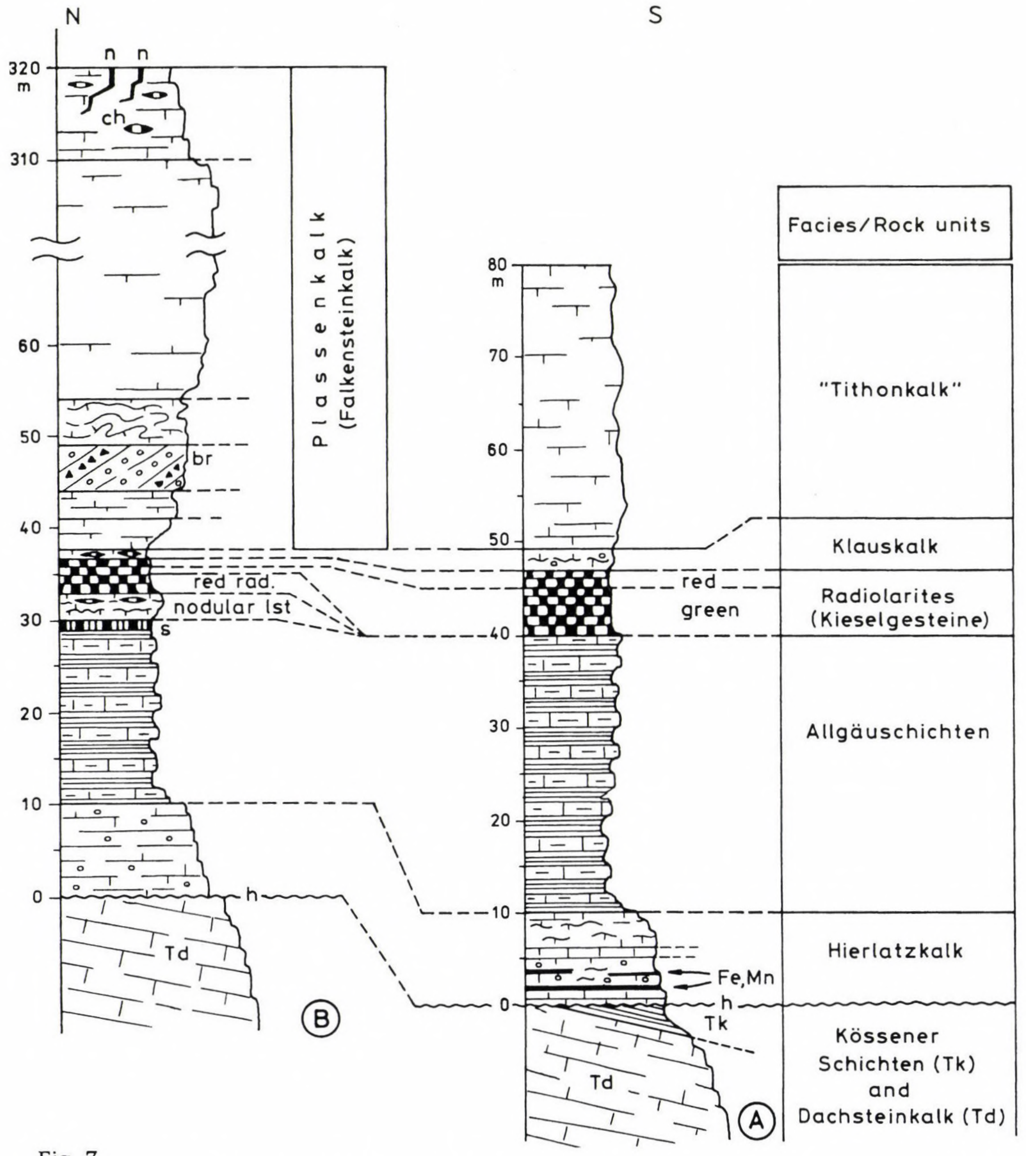

Fig. 7
Jurassic-Lower Cretaceous succession in the Göller Nappe (B) and in the Baumeckkogel Scale (A). Area of Schwarzau im Gebirge, Lower Austria. For lithology see the text and Fig. 8. br – sedimentary limestone breccia; ch – chert nodules; Fe-Mn – hematite-manganese ore bands; h – hiatus; n – neptunian dykes; s – spongolite

im Gebirge. Its complete sequence includes the following strata: pink, massive limestone at the base (2–3 m); yellowish, massive to bedded limestone (2–3 m); banded, light-grey to beige and pink crinoid limestone arranged in cross-bedded sets, with intercalations of limestone sedimentary breccias (5 m); thin-bedded, often slightly nodular, pink to yellowish, to light-beige, limestones with scarce crinoid trochites, echinoid spines and mollusk shell fragments, sometimes strongly disturbed by submarine slumping (5 m); massive to poorly bedded, cliff-forming, white to light-beige, subpelitic to fine-crystalline limestone (about 300 m); massive or thin-bedded cream-yellow to pink-yellow limestone with irregular grey chert nodules, rich in fossil remains such as brachiopod shells and echinoid spines (10 m). Clastic veins (neptunian dykes) often cross the limestone; they are filled with limestone fragments from the surrounding rock or by red Upper Cretaceous foraminiferal marls (upper Gosau Group).

Comparisons. At Schwarzau im Gebirge the Lower to Middle Jurassic development and vertical sequence of marine facies are similar both in the main body of the Göller Nappe and in its southernmost unit represented by the Baumeckkogel Scale. At the Middle/Upper Jurassic boundary there appears in both units a radiolarite horizon that marks the deepest basinal stage in the Jurassic Göller Basin.

An asymmetric sequence of green radiolarites – red radiolarites – red limestone (Klauskalk) which occurs in the Baumeckkogel Scale resembles those of the Branisko/Kýsuca and Pieniny successions in the Carpathians. This development could correspond to a deeper part of the Göller Basin.

A symmetric sequence (lower red limestone – lower red radiolarites – green radiolarites – upper red radiolarites – upper red limestone (Klauskalk)) in the Göller Nappe analogous to that of the Niedzica Succession in the Carpathians suggests deposition in a shallower part of the basin, probably on the southern slope of a submarine ridge. The appearance at the Jurassic/Cretaceous boundary of a thick, shallow-marine, intraoceanic organogenic limestone platform (Plassenkalk *vel* Falkensteinkalk) supports this assumption.

Correlation with Northern Calcareous Alps: Bajuvaricum Tectonic Units at Kirchdorf an der Krems (Upper Austria)

In the area between Kirchdorf an der Krems and the Steyr River Valley (Sheet ÖK-68 Kirchdorf a. d. Krems, Upper Austria – Figs 1, 6) which represents outer margin of the Northern Calcareous Alps, the Ternberg and Reichraming nappes (Bajuvaricum) are thrust over the Rhenodanubian Flysch Nappe (e.g., Bauer 1953; Tollmann 1976, 1987; Del Negro 1977; Janoschek and Matura 1980; Oberhauser 1980; Prey 1992). The Jurassic–Lower Cretaceous strata occur both in the Ternberg Nappe which occupies a lower tectonic position, and in the Reichraming Nappe with forms higher tectonic unit.

Ternberg Nappe. Two tectonic units are distinguished in the Ternberg Nappe: the Lower Ternberg Nappe (LTN) and the Upper Ternberg Nappe (UTN); the latter is further subdivided in three subunits (Birkenmajer 1995, 1996b, in press).

(1) In the *LTN* there is a condensed sequence of Jurassic through Lower Cretaceous, predominantly calcareous deposits 30–70 m thick, devoid of any radiolarite horizon (Fig. 8). The basement of the Jurassic sequence is formed by the Triassic carbonate platform deposits which were block-faulted and strongly eroded at the Triassic/Jurassic boundary (Paleocimmerian movements). A peculiar feature of the Liassic transgression is the lack of any basal conglomerate. Subpelitic to crinoid-enriched light-colored massive Liassic limestones were laid down directly upon tilted and eroded Triassic carbonates (Hauptdolomit in the north, Dachsteinkalk in the south). The Liassic limestones are 10 m thick in the north (Fig. 8A) growing in thickness to more than 20 m southwards (Fig. 8B). The appearance of crinoid limestone (Hierlatzkalk), of red, partly nodular limestone, and of spotty limestone (Fleckenkalk) in the same direction indicate a transition to basinal conditions.

The second stage, was the deposition of red *ammonitico rosso*-type limestone (Adneth/Klauskalk) locally replaced by crinoid limestone (Vilserkalk). They represent intraoceanic shelf platform and shelfbreak deposits formed above the CCD, probably spanning the Late Liassic through Kimmeridgian time.

The third stage, in continuity with the second one, was the deposition of thin-bedded pelagic limestone alternating with thin marls (Aptychenkalk: Tithonian–Neocomian). It is overlain by Neocomian marls which terminate the Jurassic-Lower Cretaceous sedimentary succession in the LTN.

(2) In the *UTN*, the Jurassic–Lower Cretaceous marine sequence consists of calcareous and siliceous rocks 140–180 m thick (Fig. 9A, B). They were laid down in a basinal setting, probably directly upon a thinned Triassic carbonate platform slope. There are only small tectonic blocks of Triassic carbonate rocks (Dachsteinkalk and Hauptdolomit) present at the base of the unit.

Poorly aerated basinal conditions already appeared during the Rhaetian (Kössenerschichten: black shales) and continued through the Liassic with spotty limestones and marls (Allgäuschichten) becoming siliceous (spongolites) higher up. In the highest subunit of the UTN, originally occupying the southernmost position, a thin grey crinoid limestone band appears at the base of the Jurassic succession beneath the spotty limestones (Allgäuschichten). Light-colored, fine-grained crinoid limestone (Vilserkalk) represents a sedimentary wedge of well-oxygenated organoclastic material redeposited from a shallower zone during the Middle Jurassic stage of oceanic expansion. This development resembles that of the Bajocian Smolegowa Limestone Formation in the Pieniny Klippen Belt, which was deposited on the southern slope of the Czorsztyn Ridge (Czertezik and Niedzica Successions).

Green (older) and red (younger), banded, siliceous and calcareous radiolarites (Kieselgesteine/Radiolarite) mark the deepest stage of basinal development. They correlate with the Oxfordian radiolarites (Czajakowa Radiolarite Fm.) in

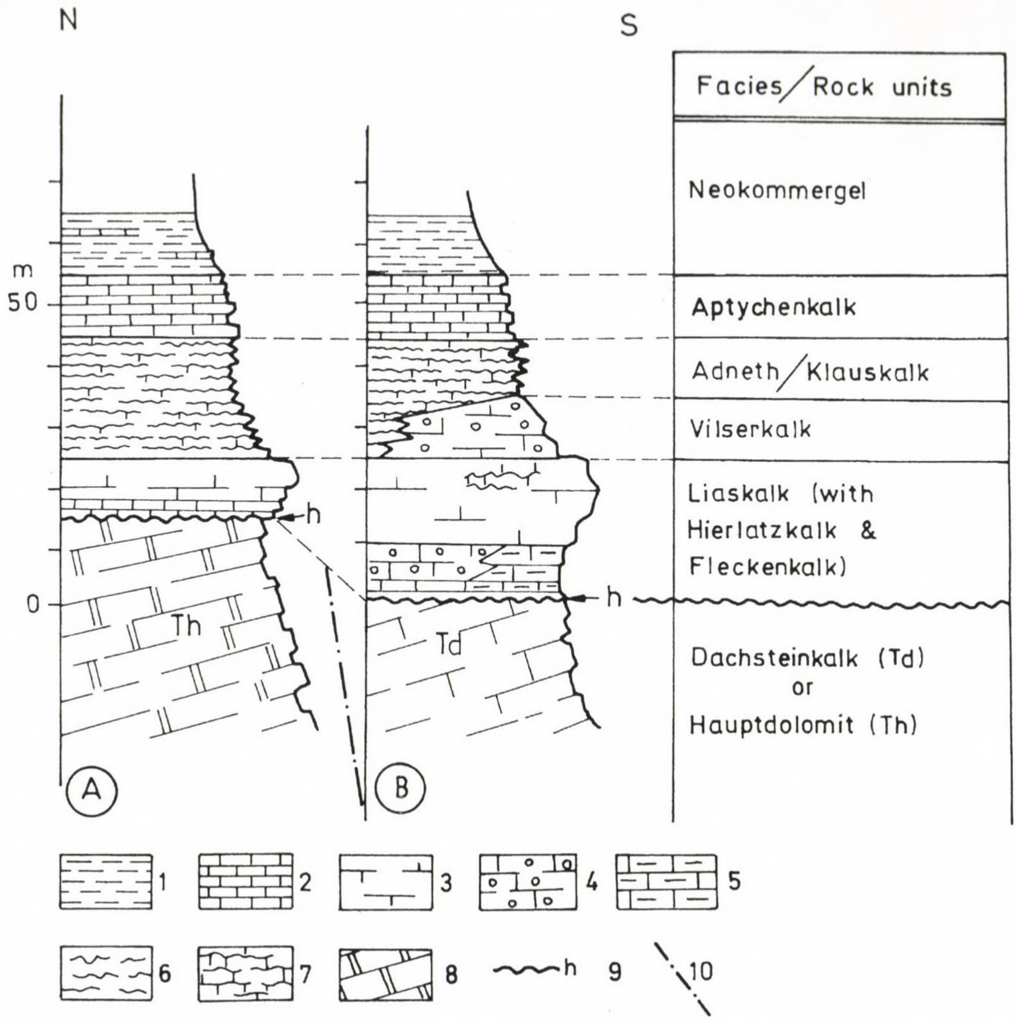

Fig. 8
Jurassic–Lower Cretaceous succession in the Lower Ternberg Nappe. Area of Kirchdorf an der Krems, Upper Austria. A – northern part of the nappe; B – southern part of the nappe; 1. marl; 2. bedded limestone; 3. massive limestone; 4. crinoid limestone; 5. spotty limestone/marl; 6. nodular structure; 7. red nodular limestone (*ammonitico rosso*); 8. dolostone; 9. unconformity (h – hiatus); 10. fault

the West Carpathians. The radiolarites wedge out to the south (they are missing in the upper subunit of the UTN). There follows red *ammonitico rosso*-type limestone (Klauskalk: Kimmeridgian) correlating with the Czorsztyn Limestone Formation in the West Carpathians. In a northward direction it totally replaces the radiolarites (Fig. 9B). The upward transition of this limestone into white,

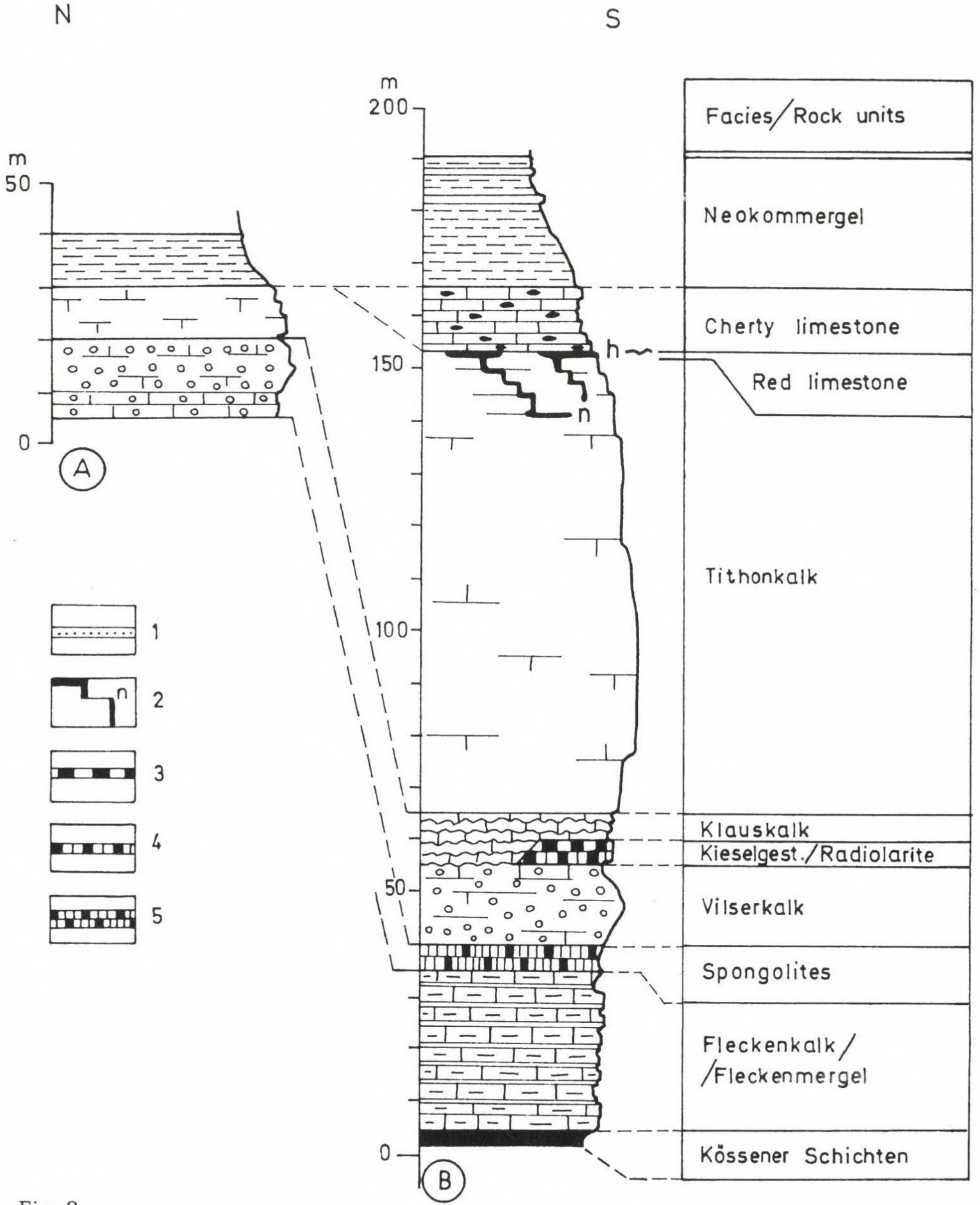

Fig. 9
Jurassic–Lower Cretaceous succession in the Upper Ternberg Nappe, lower subunit. Area of Kirchdorf an der Krems, Upper Austria. A – northern part of the lower subunit; B – southern part of the lower subunit; 1. sandstone intercalation; 2. neptunian dyke (n) and hiatus (h); 3. red radiolarite; 4. green radiolarite; 5. spongolite. For other symbols – see Fig. 8

subpelitic calpionellid limestone ("Tithonkalk": Tithonian–Berriasian, about 100 m thick) marks continuation of pelagic sedimentation.

A break in sedimentation of calpionellid limestone during the Berriasian is evidence for positive movements of the sea-bottom (Neocimmerian phase). Tension cracks that opened in the already diagenesized limestone, were subsequently filled with, and covered by, a thin layer of red, often crinoid-enriched limestone. Pelagic sedimentation conditions reappeared during the Neocomian resulting in the deposition of banded cherty limestone (biancone/maiolica facies) of reduced thickness resembling the Pieniny Limestone Formation of the West Carpathians. Deposition of marls with fine-grained sandstone (turbidite) intercalations (Neokommergel) was the final stage of basin sedimentation.

Reichraming Nappe (RN). The Jurassic rocks were laid down upon the Triassic carbonate platform (Dachsteinkalk). Basinal conditions already appeared there during the Rhaetian when black shale and limestone (Kössenerschichten, 2 m thick) were deposited.

In the northern part of the RN (Fig. 10A) a condensed Lower–Upper Jurassic sequence of red crinoid limestone (Hierlatzkalk) and red nodular limestone (Adnethkalk) with hematite-manganese oxide ore bands and concretions could have been deposited on a submarine swell. The appearance of a thin intraformational limestone conglomerate might indicate simultaneous submarine faulting and reworking of the already consolidated deposits. The red calcareous radiolarite/siliceous limestone which follows might be an equivalent to the Oxfordian Czajakowa Radiolarite Fm. The relationship between the above deposits to basinal spotty limestones and marls (Allgäuschichten) which were found at a more southerly location (Fig. 10B) has not been elucidated.

The next unit is a red *ammonitico rosso*-type limestone (Klauskalk, probably Kimmeridgian). It is followed by white massive limestones ("Tithonkalk": probably Tithonian-Berriasian) strongly resembling those of the UTN. Locally these limestones become thin-bedded. Their minimum thickness amounts to 20–30 m but may grow southward to 80–100 m (?).

Light, thin-bedded limestones, sometimes siliceous (Aptychenkalk: Neocomian), usually represent the youngest element of the RN. Locally (along the trail below Kremsmauer–Falkenmauer), above the latter unit there appears brownish to grey marly limestone (Neocomian/Albian transition?) followed by reddish marl which could represent the youngest, mid-Cretaceous unit of the RN (Fig. 10B).

Comparisons. The development and sequence of the Jurassic facies in the Ternberg and Reichraming nappes suggest a simple basin model: a submarine ridge in the north (LTN) gradually giving way southward to a deep basin with Jurassic radiolarite deposition (UTN) and rimmed in the south by another submarine ridge or swell (RN, northern part).

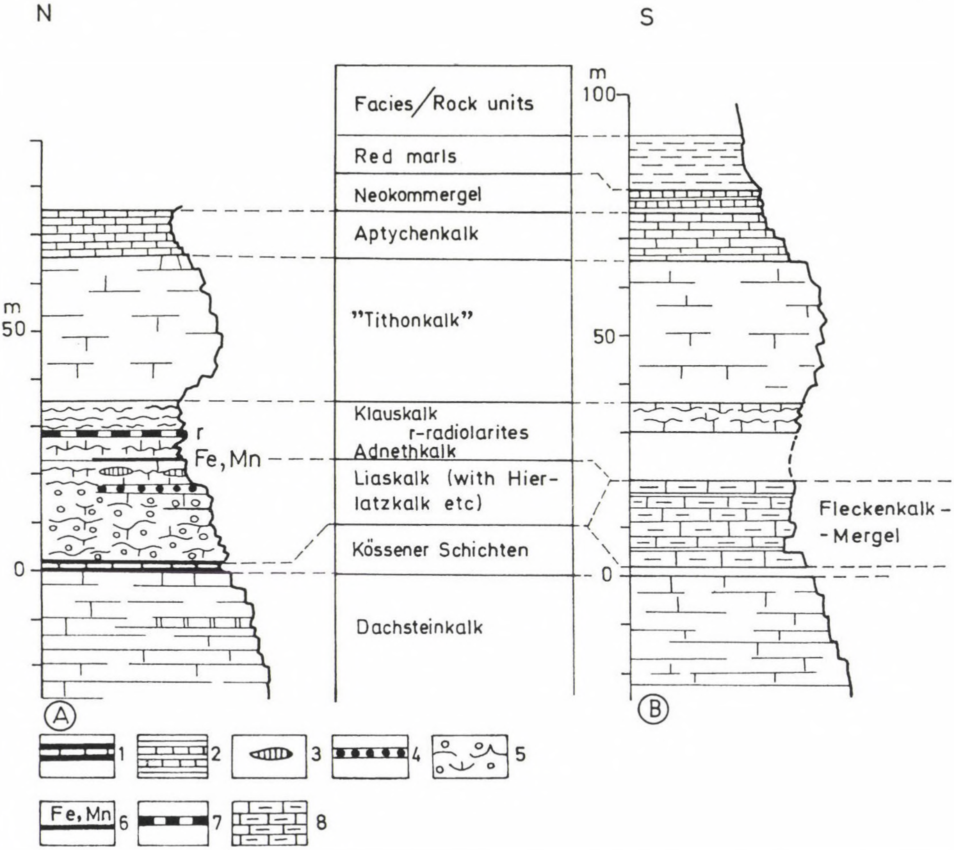

Fig. 10
Jurassic–Lower Cretaceous succession in the Reichraming Nappe. Area of Kirchdorf an der Krems, Upper Austria. A – northern part of the nappe; B – southern part of the nappe; 1. black shale and limestone; 2. bedded limestone and marl; 3. chert nodules; 4. intraformational calcareous conglomerate; 5. crinoid/nodular limestone; 6. Fe-Mn ore bands; 7. red radiolarite/siliceous limestone; 8. spotty limestone (Fleckenkalk). For other symbols – see Figs 8, 9

The appearance of basinal facies in the southern part of the Reichraming Nappe as expressed by the appearance of Liassic spotty limestones and marls (Allgäuschichten), and of pelagic limestones and marls at the end of the Lower Cretaceous, might indicate the existence of another sub-basin still further south.

The development, and sequence of Jurassic facies in the UTN very much resemble, in the Middle Jurassic through Tithonian/Lower Neocomian strata, those of the Czertezik Succession (Pieniny Klippen Basin). In its Liassic through Neocomian sequence of strata it is also close to the Križna Succession of the West Carpathians. In both cases we are dealing with successions laid down on slopes of submarine ridges which had an attenuated Triassic platform/ pre-Triassic continental crustal base.

Conclusions

(1) A common trend in the evolution of the Jurassic–Early Cretaceous basinal marine sequences is recognizable in both the Western Carpathians and the Northern Calcareous Alps. It is expressed by (a) gradual deepening of the basins from the Liassic through the Oxfordian, with radiolarite formation at the Middle/Late Jurassic transition; (b) a shallowing of the basins from the Kimmeridgian through the Jurassic/Cretaceous transition.

(2) Both in the Western Carpathians and in the Northern Calcareous Alps the radiolarites are pelagic, deep-water deposits. Purely siliceous radiolarites were laid down most probably below CCD, at depths corresponding to abyssal plains and oceanic troughs. Calcareous radiolarites were, for the most part formed at, or slightly above, the CCD.

(3) A characteristic banding of the radiolarites, accentuated by thin clay-shale or marly shale alternations, is a primary sedimentary feature related to deposition from diluted turbidity currents that brought clay and carbonate detritus in suspension from distant coastal areas, interrupting the quiet deposition of radiolarian ooze.

(4) The radiolarites are usually associated with deep-water deposits such as red *ammonitico rosso*-type nodular limestones considered an intraoceanic shelfbreak/slope deposit. In marginal parts of the basins, both in the Western Carpathians and Northern Calcareous Alps, a characteristic symmetric succession of facies had often developed: the radiolarites are underlain and overlain by red nodular limestones and within the radiolarite horizon the red radiolarites form its lower and upper parts, with green radiolarites present in the middle. Such symmetric sequences had originated at depths close to the CCD, the green radiolarites probably representing the deepest stage of basinal deposition.

(5) The red and green radiolarites span the period from ?Callovian, mainly through Early to Late Oxfordian, sometimes to Early Kimmeridgian. The deepest stage, believed to be represented by the green radiolarites, has been biostratigraphically dated at Early Oxfordian.

(6) The synchroneity of radiolarite formation at the Middle/Late Jurassic transition in different, widely spaced basins of the Outer and Inner West Carpathians, and in more distant basins of the Northern Calcareous Alps corresponds to a maximum extension of the Jurassic basins in the entire North Tethyan oceanic realm.

(7) In the Jurassic basinal sequences of the West Carpathians discussed here (the Klippen Belt successions, the Križna Succession (Tatra Mts.), and the Grajcarek Succession (inner part of the Magura Nappe, Outer Carpathians)) the formation of radiolarites was unrelated to oceanic rifting. The rifting,

associated with oceanic crust formation, occurred much earlier there, namely during the Early-Middle Triassic in the Pieniny Klippen Basin and probably during the Early Jurassic in the Magura Basin. There was no Mesozoic oceanic rifting and oceanic crust formation in the Križna basinal succession (see Birkenmajer 1986, 1988; Birkenmajer et al. 1990).

(8) The present author sees no field evidence for the presence in the Pieniny Klippen Basin, during the Oxfordian through Barremian, of an active post-Hettangian ("Vahic") oceanic crust undergoing "passive rifting" between two continental-type margins, namely the "Oravic" (i.e. the Czorsztyn Ridge) in the north and the Tatric in the south, as assumed by Plašienka (1995, Fig. 10).

(9) In the Northern Calcareous Alps – the Göller Nappe and Schwarzau Scales (Tyrolicum) in the south, and the Ternberg and Reichraming nappes (Bajuvaricum) in the north – the formation of Jurassic radiolarites was also unrelated to oceanic rifting. The Jurassic radiolarite-bearing sequences described here were laid down in narrow, possibly fault-bounded basins, separated from each other by radiolarite-free submarine ridges/swells.

(10) The shallowing of the basins, from the Kimmeridgian towards the Tithonian/Early Cretaceous transition as recorded in both the West Carpathians and the Northern Calcareous Alps, correlates in time with sea-bottom uplift and gravity faulting (Neocimmerian movements) as recognized in sedimentary sequences of submarine ridges and swells that separated the basins.

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From 1992 to 1996, the present author had an opportunity to investigate Jurassic successions of the Northern Calcareous Alps in the course of detailed geologic mapping at 1:10,000 scale of Sheets ÖK-74 Hohenberg and ÖK-68 Kirchdorf an der Krems. The fieldwork was carried out on behalf of the Austrian Geological Survey (Geologische Bundesanstalt, GBA) in Vienna. This was part of GBA projects coordinated by Drs W. Mandl and H. Egger, respectively. GBA's Deputy Director Dr W. Janoschek, and Director Prof. Dr. H.-P. Schönlaub, provided financial means for the mapping. Drs R. Oberhauser and W. Schnabel most kindly helped the author with regional geologic literature. Dr J. Michalík (Geological Institute, Slovak Academy of Sciences, Bratislava) and an anonymous referee are thanked for their helpful remarks on the manuscript.

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Geodynamics of the Middle Adriatic offshore area, Croatia, based on stratigraphic and seismic analysis of Paleogene beds

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Stratigraphic interpretation of the Paleogene deposits clearly displays two different sedimentary realms, which are recognized as belonging to different tectonostratigraphic units. The western unit belongs to the Adriatic indenter and the eastern one to the Dinaridic thrust belt. In the western realm the stratigraphic succession of the deep marine sediments ranges from the Upper Cretaceous to the Middle/Upper Eocene. Carbonate turbidites deposited on the deeper part of a platform slope are characterized by a mixture of wackestone/mudstone rich in planktonic foraminifers and packstone/rudstone with shallow-water, platform-derived detritus. Based on the data from the wells Koraljka-1 (Kok-1), Ksenija-1 (Kse-1), Jadran-1 (J-1), Jadran-2 (J-2), Jadran-21/1 (J-21/1) and Jadran-10 (J-10), at the end of the Upper Eocene (Priabonian) period the area was drowned and covered by siliciclastic turbidites. This complete succession of Cretaceous to Paleocene sediments has no onshore equivalent. In the eastern realm Paleogene shallow-water carbonates are identical to the same deposits of the External Dinarides. A long hiatus between the Upper Cretaceous and the Cuisian is evident in wells Jadran-9 (J-9), Jadran-3 (J-3), Kornati More-4 (KM-4) and Susak More-1 (SM-1). Only in the area of the Kate-1 well is a certain continuity of the sedimentation present in the environment of shallow restricted lagoons, repeatedly interrupted by short phases of exposure at the time of overall regional emersion in the Early Paleogene. In the remainder of the platform the stratigraphic succession begins with a transgression of the uppermost Paleocene and/or Cuisian limestone that overlies karstified Cretaceous deposits. Transgressive succession started with fenestral and Charophita limestone (facies X) which was followed by foraminiferal wackestone/packstone, marl with ostracodes and fine-grained breccia conglomerate (facies A), foraminiferal packstone/grainstone limestone and miliolid mudstone/wackestone limestone (facies B and C), porous limestone?, grainstone? (Facies Y), fossiliferous mudstone/packstone limestone (facies D), algal boundstone/bindstone and wackestone (facies E), foraminiferal mudstone/packstone (facies F) and Nummulite-Discocyclinae floatstone/packstone and grainstone/rudstone (facies G). These sediments were deposited close to the carbonate platform margin. Beds of the X-facies are younger than Cretaceous and older than Cuisian. A transgressive sequence from A- to Y-facies was formed during the Cuisian whereas the others belong to the Lutetian and partly to the Biarritzian. At the end of the Middle Eocene they were tectonically compressed and covered by flysch deposits. At Paleogene time the deep marine sediments were deposited at a considerable distance from the Dinaridic carbonate platform (DCP). A comprehensive study of the flysch deposits suggests that the drainage area was located to the west and not related to the DCP. Underthrusting of A to the northeast created the imbrications of the Dinaridic frontal thrust which is clearly expressed in the tectonic repetition of Paleogene sequences in the J-9 and SM-1 wells. The continuing compression caused the inversion of the Adriatic indenter during the Miocene.

Key words: Paleogene, offshore wells, tectonostratigraphy, Adriatic, Dinarides

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Introduction

The geology of the Central Adriatic offshore area (Fig. 1) presented here is the result of twenty years of exploration for oil and gas by offshore drilling and seismic prospecting. About twenty wells have penetrated the Paleogene deposits. This study undertakes their sedimentologic and tectonostratigraphic interpretation. Several recent papers based on well data from this area deal with stratigraphy and biostratigraphy: Kalac and Tari-Kovacic (1986), Kalac et al. (1990, 1991), Lucic et al. (1991, 1993), Veseli et al. (1991), Lucic (1993). Facies analyses and structure of the Eocene carbonate platform were presented by Tari-Kovacic (1992). The age and sedimentary facies interpretation is based on the detailed study of the cored intervals (lithostratigraphic and biostratigraphic analyses) and sedimentological core descriptions. The wireline log facies interpretation (Selley 1985) was carried out on the logs that primarily indicate the lithological characteristics of the sediments; natural radioactivity (gamma rays and spectralog) and the porosity logs (neutron, density and sonic logs). This technique is primarily qualitative and relies heavily on the visual identification of the coherent units. The criteria used for separating the units depend on the geologic data acquired by drilling. There is no standard procedure for separating certain wireline log facies although it would be logical to separate the obvious coherent rocks units first, and then in correlation with those of neighboring wells gradually perform the detailed separation. For the wireline log facies which were not cored sedimentary facies was determined by comparison with similar wireline log facies that were cored in the same or in the neighboring wells, providing relevant time–space correlation of the entire area.

Stratigraphy

The area under consideration can be divided into two different tectonostratigraphic units which are separated by a significant tectonic break (Fig. 1). The western unit corresponds to the Adriatic indenter/Promontory (Channel et al. 1979; Ratschbacher 1991). Sedimentation was continuous from the Maastrichtian through the Paleogene under the deep-water conditions where carbonate turbidites predominate. Deep wells J-1 (Fig. 2), J-2, J-21/1, J-22/1, J-0, Kse-1 (Fig. 3) and Kok-1 present the best record of that unit's history. In the eastern unit the deposits consist of shallow marine carbonates of Upper Cretaceous and Eocene age. They are part of the Dinaridic carbonate platform. Paleocene deposits occur sporadically, as in the Kate-1 offshore well area and in several restricted onshore outcrops. During the Late Eocene and Miocene these carbonates were compressed and formed a particular imbricated zone detached from the *Dinaridic thrust belt.* The best records of the stratigraphic development of this unit are from the wells J-3, J-9, KM-4, Kate-1 and SM-1.

Western realm

Sediments encountered in the wells of this area display similar sedimentary features, i.e. a rhythmical alternation of mudstone/wackestone and allochtonous bioclastic limestones (packstone/grainstone to rudstone), to the underlying Cretaceous deposits. According to their content of planktonic and benthic foraminifers as well as to the plankton-benthos ratio and to the lithological features, it is presumed that their deposition during the Paleocene and the Eocene took place on the deeper part of a continental slope. The sporadic occurrence of redeposited benthic remains indicates gravity-induced depositional mechanisms, i.e. turbidity currents. The Maastrichtian age of the succession underlying the Paleocene deposits is well dated by planktonic foraminifers and the nannofossil assemblage. The oldest Paleogene deposits are related to the Danian stage (Fig. 6) which is proved by the presence of *Morozovella pseudobulloides* (Plummer), *Planorotalites compressa* (Plummer) and *Morozovella trinidadensis* (Bolli). Planktonic foraminifers dominate the well records of Thanetian to Lutetian and Bartonian age. Priabonian carbonate turbidites are well dated only in the J-1 and J-2 wells by the presence of planktonic foraminifers *Turbirotalia cerroazulensis* (Cole), *Globigerina yeguaensis* (Weinzierl et Applin), *Globigerinatheca mexicana kugleri* (Bolli, Loeblich et Tappan) (Lucic et al. 1993). The main lithological change from carbonate to siliciclastic turbidites occurred during the Priabonian. Those siliciclastic turbidites were partly or almost completely eroded due to tectonic inversion during the Miocene.

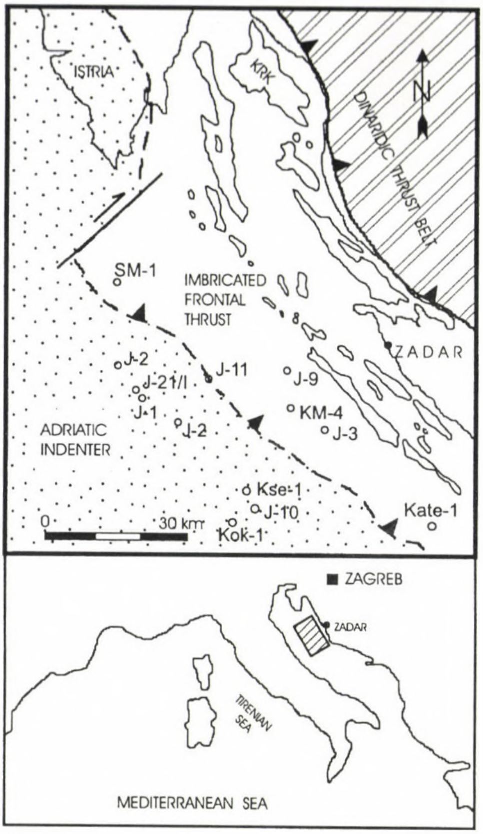

Fig. 1
Location map

Eastern realm

The eastern tectonostratigraphic unit constitutes an area in which the Paleogene beds were tectonically imbricated during the Late Eocene in front of the Dinaridic thrust belt (Fig. 1). The best record of the stratigraphic

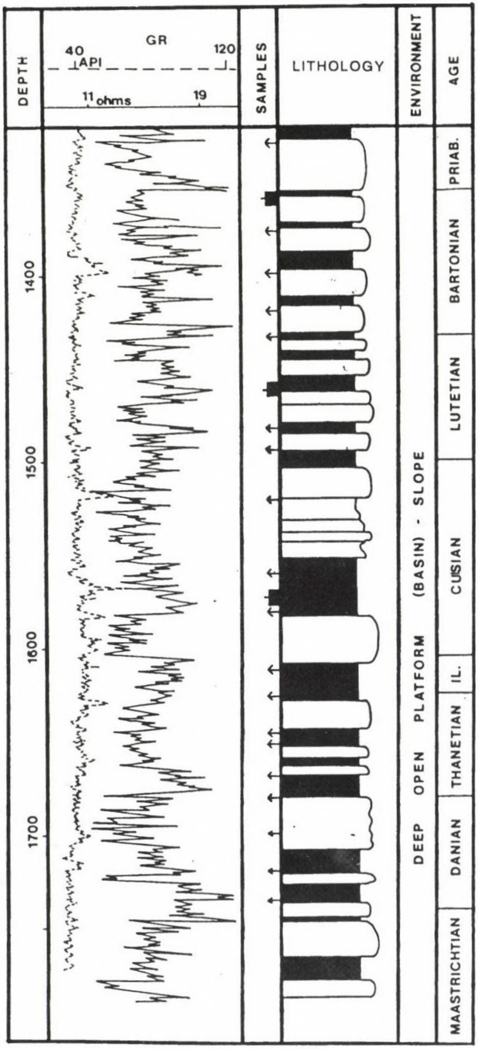

Fig. 2
Sequence of Paleogene strata in the deep well J-1. For legend see Fig. 4

development in this unit was produced by the J-3, J-9 (Figs 4 and 5), KM-4, Kate-1 and SM-1 wells which penetrated a transgressive sequence of Eocene carbonate platform limestones (as clearly demonstrated by cores, cuttings and wireline logs). In the SM-1 well total loss of drilling mud in circulation occurred at the depth of 755 m and the interpretation was made by log correlation. The section is composed only of tectonically repeated Cuisian facies pattern (X, A, B+C and Y). Lutetian part of the sequence was probably eroded or represents a hiatus. The succession of the wireline log facies is almost the same in all the wells (Figs 4 and 5). It begins with facies consisting of mudstone and mudstone/wackestone limestone that overlies eroded and karstified Late Cretaceous (Senonian and Turonian) carbonates. Facies A is similar to facies X but its natural radioactivity is very high (70 API on the gamma ray curve) due to the unusually high uranium content. The presence of uranium could be explained by high contents of organic matter. The next facies in the sequence is B+C reflecting an alteration of tidal flat (Fig. 8) packstone/grainstone (B) and mudstone/wackestone (C) limestones as is well expressed on porosity, density and sonic logs. Facies Y was identified only on the wire-line logs of the SM-1 well. There is an assumption that the low gamma radiation and high acoustic value correspond to porous limestone (? grainstone) deposited in a high-energy shoal environment of the beach as the final member of the regressive sequence. Facies D corresponds to clean and tight fossiliferous mudstone and packstone limestones deposited in restricted carbonate environments. This facies is characterized by low radioactivity/ porosity and high density/sonic log responses. Facies E is composed of algal boundstone (bindstone) and wackestone. The wireline response of the facies is characterized by a serrated log shape of all the curves and low gamma ray values referring to clean, washout limestone. Facies F (foraminiferal mudstone and packstone) and facies G (floatstone/packstone and grainstone/rudstone packed with nummulites and Discocyclina) represent respectively the reef front, fore-reef and shoal deposits on a platform edge. Facies F is expressed on the logs as the intervals of low porosity and high density/sonic values, while facies G has higher porosity, lower density and lower sonic values. According to the paleontological data the age of the complete carbonate sequence can be dated as Cuisian (facies A and B+C) and Lutetian-Bartonian (facies D, E, F and G). The oldest ?Paleogene deposits were penetrated only by the Kate-1 well. In this well, below facies x peritidal carbonate deposits similar to the younger Eocene B+C facies were found. The scarce flora and fauna are not representative either for Cretaceous or for Paleocene deposits. The small rotaliid foraminifers (*Protelphidium* sp. and discorbids) occur in the contemporaneous sections studied on the mainland. They were deposited in a very shallow intertidal and supratidal regime with numerous emersions, representing the final member of the Cretaceous sedimentary cycle and the beginning of the Early Paleogene Thanetian or Ilerdian transgression. The age of facies X is also undocumented because of its scarce fauna content consisting of miliolids, discorbids and ostracods, representing the first unit of the transgressive cycle that ends at the end of the Middle Eocene. They are genetically related to the overlying deposits of the Cuisian facies B+C, which age is dated by the presence of *Coskinolina perpera* Hottinger et Drobne, *Chrysalidina makarskae* (van Soest), *Periloculina dalmatina* Drobne and *Coskinolina liburnica* Stache. The age boundary between Cuisian and Lutetian coincides with the facies change from facies B+C (coastal environment) to facies D (shallow platform environment). A Lutetian age is also attributed to facies E, F and G. They are dated by the presence of the benthic foraminifers *Asterodiscus stellatus* Gümbel, *Discocyclina sella* D'Archiac, *Sphaerogypsina globula* (Reuss), *Asterigerina rotula* (Kaufmann) *Nummulites* sp. and *Alveolina* sp. The pelagic influence on facies G and F is well expressed by

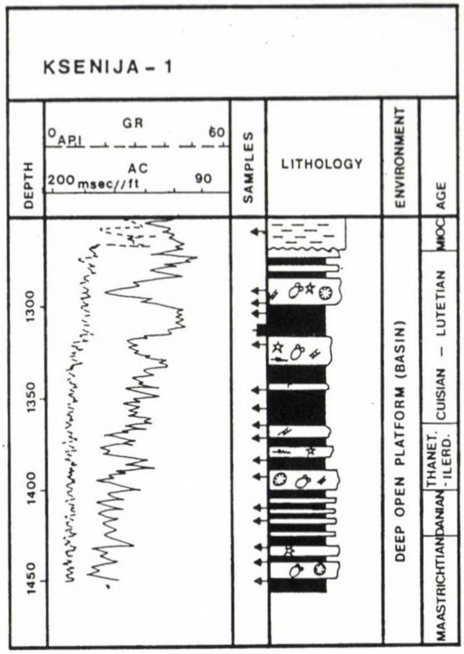

Fig. 3
Sequence of Paleogene strata in the deep well Ksenija-1. For legend see Fig. 4

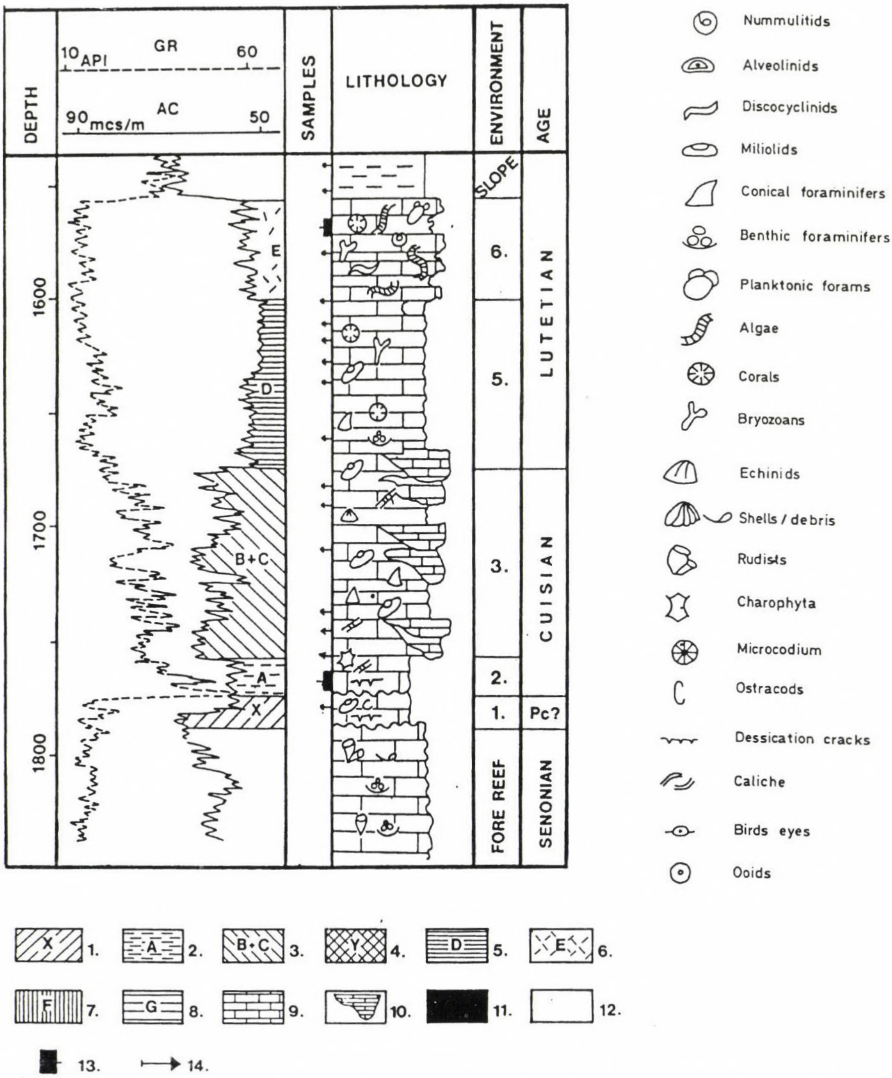

Fig. 4
Sequence of Paleogene strata in the deep well J-3. 2–5. Facies of fenestral and Charophyta limestone (X); salt and brackish marshes, tidal flat environment; 2. facies of foraminiferal wackestone–packstone, marl with ostracods and fine-grained breccia conglomerate (A); coastal environment with repeated shallowing and emersions; 3. coastal environment with repeated shallowing and emersions; 4. facies of porous limestone? – grainstone? (Y), beaches?; 5. facies of fossiliferous mudstone/packstone limestone (D); shallow platform and mud mounds environment; 6. facies of algal boundstone (bindstone) and wackestone (E); platform margin build-ups; 7. facies of foraminiferal mudstone/packstone (F); foreslope environment, 8. facies of Nummulites–Discocyclina floatstone/packstone and grainstone/rudstone (G); fore slope environment; 9. shallow-marine mudstone/wackestone, 10. shallow-marine packstone/ grainstone; 11. deep marine mudstone/wackestone; 12. deep marine wackestone/packstone; 13. core samples; 14. mud samples

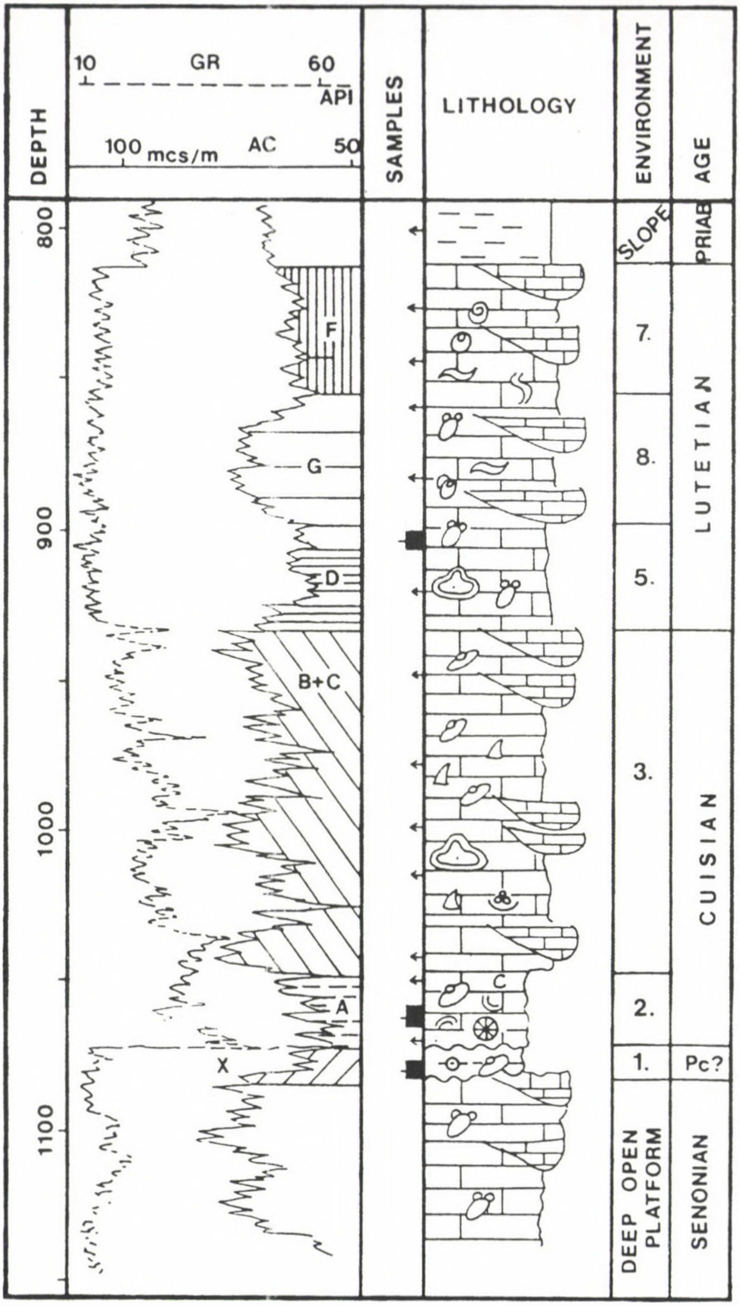

Fig. 5
Sequence of Paleogene strata in the deep well J-9. For legend see Fig. 4

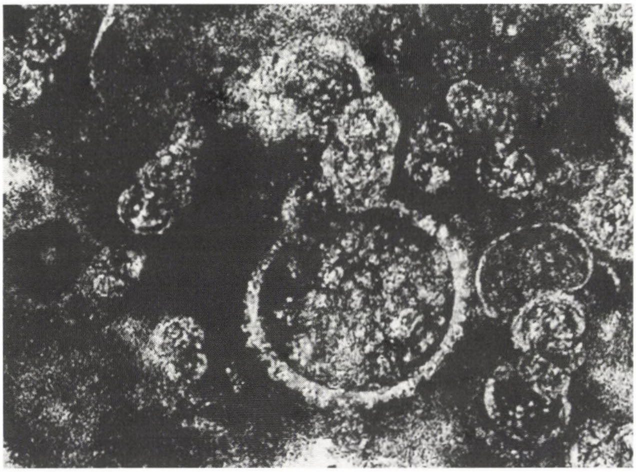

Fig. 6
Pelagic microfacies, Kok-1, 2450 m, Lowermost Danian, x150

the presence of the planktonic foraminifers *Globigerina eocena* (Gümbel), *Globigerina tripartita* Koch, *Catapsidrax dissimilis* (Cush. et Berm.), *Globigerina boweri* (Bolli), *Turboratalia cerroazulensis* (Cole) and *Turborotalia centralis* (Cushman). During the Lutetian the carbonate platform around the J-3 well was drowned by siliciclastic turbidites; this is well documented by the presence of benthic and planktonic foraminifers. The carbonate platform around the Kata-1 well was drowned some time later in the Biarritzian, documented by the presence of *Operculina bericensis* Oppenheim, *Nummulites millecaput* Boubee and *Nummulites dufrenoy* D'Arch. et Haime (Kalac et al. 1986, 1990).

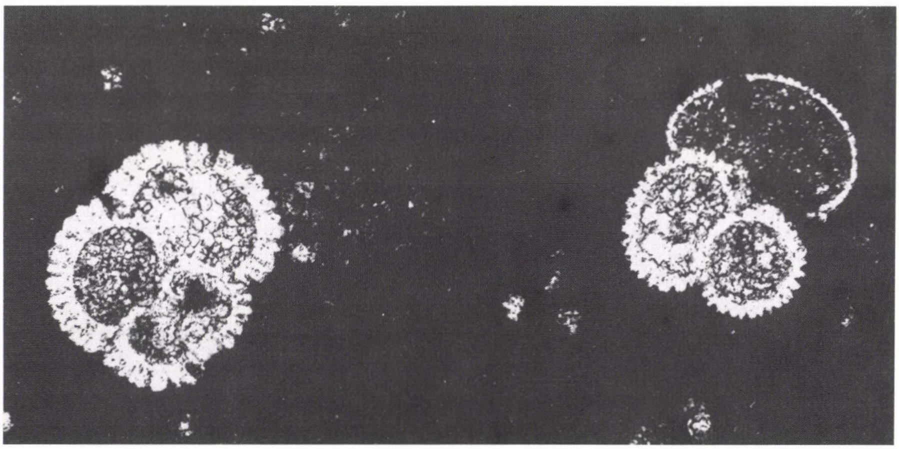

Fig. 7
Pelagic microfacies, Kok-1, 2254-2258 m, Lutetian, x100

Fig. 8
Facies of foraminiferal packstone–grainstone (B); J-9, 1029–1032 m, Cuisian, x30

Structural interpretation

The slope and basin-related Maastrichtian to Priabonian, predominantly carbonate turbidites of the Western unit are embraced in complex structural unit known among Croatian geologists as the Central Adriatic Uplift. The age and the sedimentary environment are well defined by planktonic foraminifers and nannofossils which are predominant in the pelagic or hemipelagic part of the turbiditic sequences. The well-log correlation along the slope is favorable and suggests the presence of broad and thick fan aprons. The movement of turbiditic currents was synchronous, mostly initiated by tectonic uplifts. The seismic-stratigraphic interpretation (Fig. 11) shows the generally E–NE direction of the sediment transportation. It is out of the question that the western unit represents the immediate continuation of the platform as its westward slope.

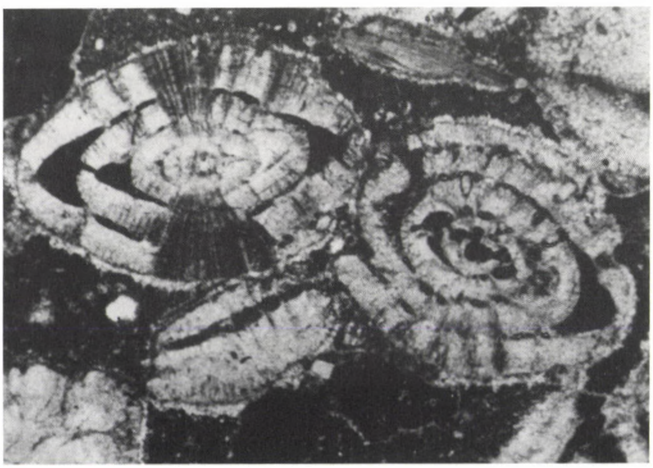

Fig. 9
Facies of Nummulites – Discocyclina packstone–grainstone (G), 915–929 m, Lutetian, x30

Comparing Paleocene sedimentary sequences in the wells from the western unit with the eastern ones, the lack of a corresponding facies belt to platform edge and upper slope is evident. Moreover, the calcareous turbidites of the western unit are too well sorted, fine-grained and have too uniform an assemblage to represent truly proximal slope facies types. During the Late Eocene and Oligocene that area was part of the Dinaridic foreland on which the flexural foreland basin (Fig. 11) began to develop. Advanced compression is related to the movement of the Adriatic indenter northward and northeastward and its indenting beneath the Southern Alps and Dinarides resulted in partial inversion of its eastern parts during the Miocene. The main characteristic of that large inverted structure is a complex system of shallow horst and graben-like structures. The north-oriented steeply dipping reverse faulting (Matiec 1994) in some parts of the system is the result of the approximately north-south related post-Miocene transpression of dextral lateral movements. Eastward of the major tectonic break (Fig. 1) a schematic correlation (Fig. 10) shows the spreading of Eocene carbonate deposits along the Dinaridic structural trend, from the SM-1 well in the northwest to the Kate-1 well in the southeast. It represents the older (carbonate) part of the Paleogene and Neogene transgressive megasequence. Facies X, A, B and C are almost identical in all the wells. They were deposited in coastal environment (facies X and A), and in the tidal flat zone (facies B and C). In the SM-1 well area the sedimentation of the Eocene carbonates was interrupted at the beginning of the Lutetian by the uplifting of the carbonate platform. This caused emersion which lasted until the end of the Upper Eocene when the deposition of carbonate sediments started again. Sediments of Upper Cretaceous, Eocene and Lower Oligocene age are repeated twice in the well due to a reverse fault at the depth of approximately 945 m. The total thickness of the Eocene sequence is 140 m in the thrusted and 142 m in the autochthonous part of the section. Toward the south and southeast the rest of the carbonate platform was oscillating and slowly drowning, with a tendency toward increasing thickness. Transgressive sediments continued to accumulate through the Lutetian with facies D, E, F and G characterized by a shallow restricted platform and platform-edge depositional environment. The Eocene carbonates are 288 m thick in the J-3 well, 369 m in the KM-4 well, and 395 m in the Kate-1 well. The thicknesses of the folded and reverse-faulted Eocene carbonate section in the J-9 well are 132, 157 and 256 m. In the J-3 well area reefal limestone (facies E) is the end member of the calcareous part of the transgressive megasequence which has been overlaid by flysch sediments. Facies G and F in the J-9, KM-4 and Kate-1 wells are cyclically repeating. Shoals of open shallow waters interfinger with sediments of the fore reef environment. Such an aggradation of facies is the result of multiple, relative changes of the sea level that do not correspond to the eustatic curve of sea level change (Haq et al. 1978). Rather it reflects the pulsing of the platform before its final drowning in front of the overthrusting structures of the Dinarides in the East. Because of folding (recumbent fold) and

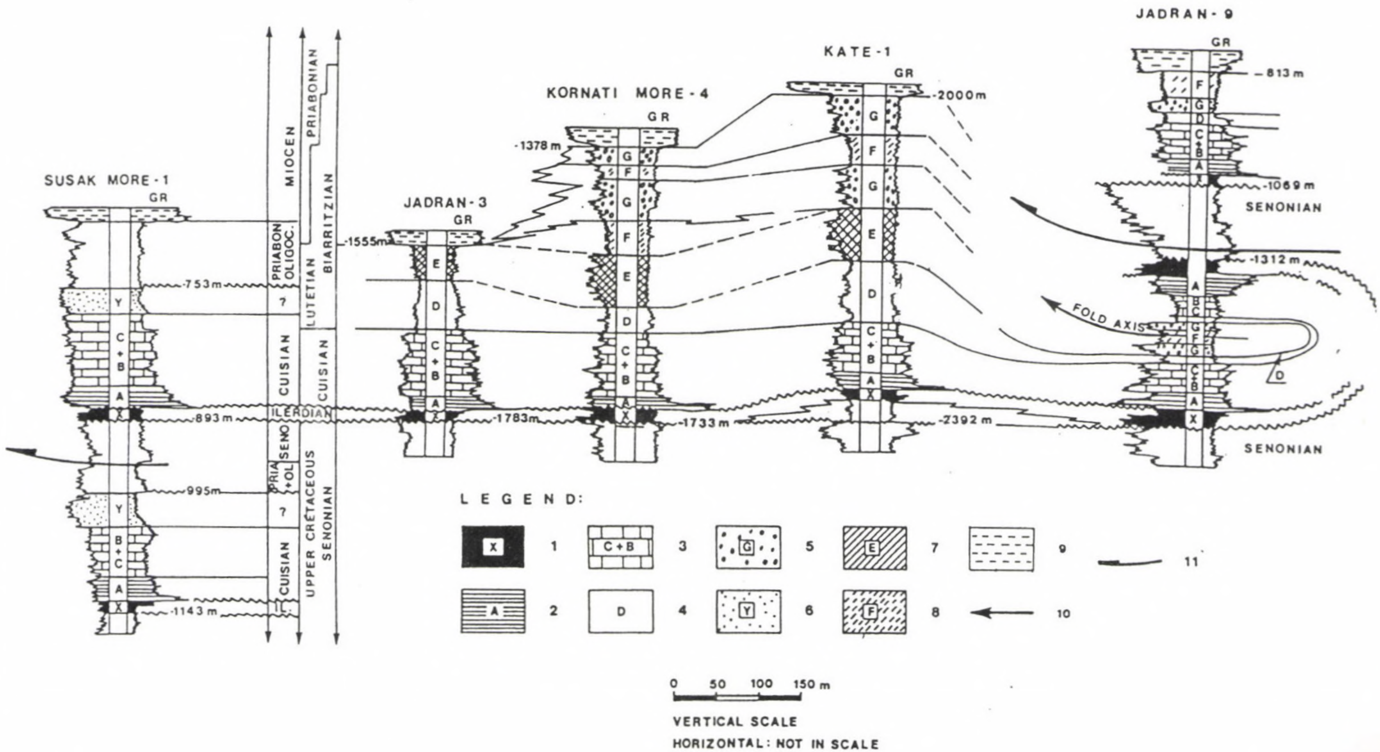

Fig. 10

Facies display and structural position. 1. peritidal deposits; 2. coastal deposits; 3. tidal flat deposits; 4. restricted platform deposits; 5. open platform shoals; 6. ?beach deposits; 7. reef; 8. forereef deposits; 9. flysch; 10. fold axis; 10. thrust fault

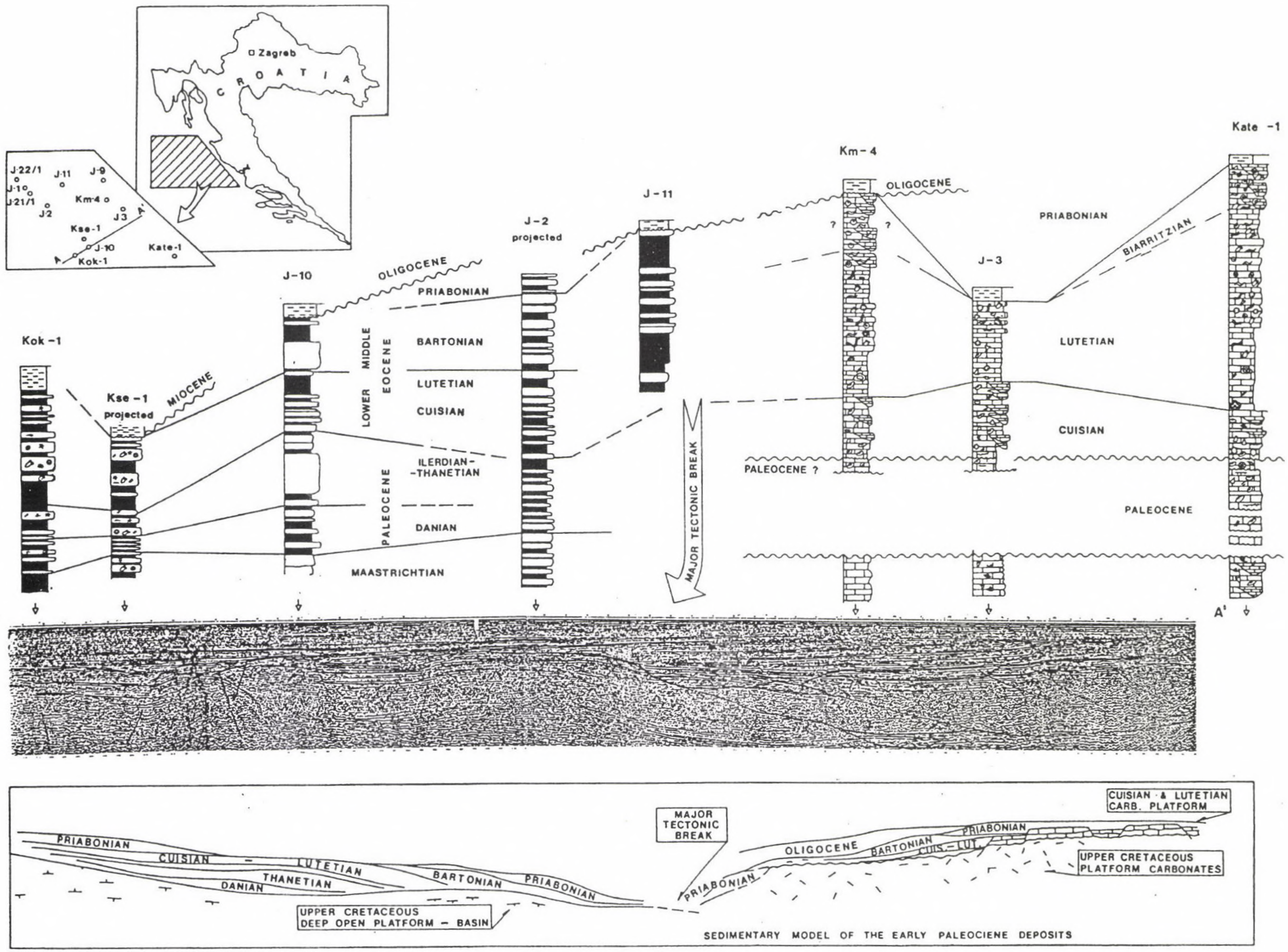

Fig. 11
Facies distribution and tectonic display

thrusting, the same Paleogene carbonate sediments were found in three intervals in the J-9 well, with the middle interval being in inverse position (older sediments above the younger ones). In all the three intervals the sequence of facies from X to G are identical. The thickness of the overturned interval (132 m) is less than the other two intervals probably because of compaction under the overlying hanging wall.

Tectonic considerations

The studied Adriatic offshore belongs to the Apulian microplate of the Alpine Orogenic System (Yilmaz et al. 1996). Cuisian and Lutetian platform carbonates of the eastern unit, in the offshore wells and in the outcrops on the mainland (as shown by Tari-Kovacic 1992) are affected by the imbrications of the frontal thrust of the Dinaridic thrust belt, detached along a Late Jurassic – Early Cretaceous anhydrite level (Tari-Kovacic 1994). From the point of view of plate tectonics the imbricated zone of the Cretaceous and Eocene platform carbonates represents an accretionary wedge detached from the Dinaridic thrust belt, according to its stratigraphy and, even more revealingly, to its folded and reverse faulted structural style. It was underthrusted by the Adriatic indenter (Ratschbacher 1991), as documented by deep refraction seismic (Aljinovic ad Blaškovic 1981), as more or less steeply dipping fault zone in front of the thrust belt. It was considered by Herak (1986) to be the main subduction zone. In the studied area the western unit, as a part of the Adriatic indenter, occupies the zone of the Central Adriatic Uplift that plunges from the Istrian peninsula southward through the middle part of the Adriatic Sea. It was formed during Oligocene and Miocene indenting of the Adriatic toward the Dinarides and Alps, causing tectonic inversion of Cretaceous and Paleogene slope and basinal deposits. Those sediments were deposited on the western margin of the basin which existed at that time between the Adriatic and Dinaridic carbonate platforms, probably as the continuation of Budva basin. Considering the 'tectonic environment' of Alpine movements in that area, the missing eastern margin of the basin was probably immerged by underthrusting (or even subduction?) the Adriatic indenter beneath the Dinaridic thrust belt at the end of the Late Eocene.

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Synsedimentary tectonic events in the Middle Triassic evolution of the SE Transdanubian part of the Tisza Unit

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On the carbonate ramp developed over Lower and Middle Triassic continental and then shallow marine sediments of SE Transdanubia (Tisza Unit – Jakabhegy Sandstone, Patacs Siltstone) and the Lower Anisian evaporitic series of coastal sabkha origin (Magyarürög Anhydrite, Hetvehely Dolomite), formations of Muschelkalk character were deposited. In the process of development of the ramp, two sharp changes can be demonstrated. Features indicating synsedimentary tectonics first appear in the Bithynian Rókahegy Dolomite and then in the Pelsonian Zuhánya Limestone Formation. These are contemporaneous with the early Alpine tectonic events detected in German and Alpine areas. At the end of the Pelsonian varied sedimentary environments came into being due to the eustatic decrease in water level and tectonic activities. Evaporitic middle Muschelkalk can be demonstrated followed by platform-type carbonates.

Key words: SE Transdanubian, Middle Triassic evolution, Muschelkalk, facies interpretation, ramp evolution

Introduction

Middle Triassic formations of the area were deposited on the northern shelf of the early Tethys, on an epicontinental regular ramp (Török 1986; Bleahu et al. 1994). Based on geologic mapping and borehole data it is inferred that changes in the process of development of the ramp were caused by structural movements. Their effects are studied and efforts are made to correlate them with events known in the broader environment.

Formations indicating dilatation tectonics

In SE Transdanubia, the Lower and Middle Triassic processes of development of the Mecsek and Villány Mountains (Fig. 1) show no significant differences. The Mecsek facies area can be divided into two terranes (Chikán and Konrád 1982):

– The western Mecsek, which together with borehole Gálosfa-1 (Gf-1) and the "northern sliver" shows a closer relationship with the Villány Mountains, sequence;

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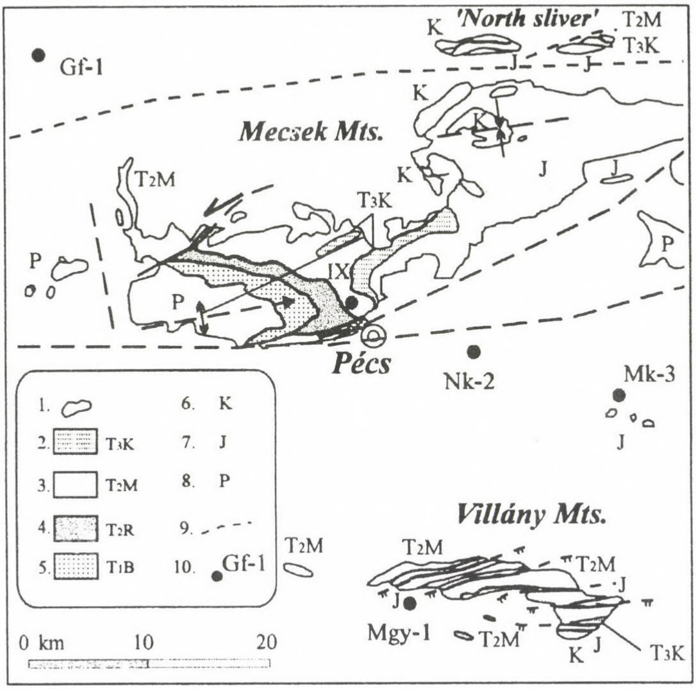


Fig. 1
Surficial extension of the Triassic formations in SE Transdanubia with the indication of the important boreholes. 1. boundary of outcrop of Paleozoic and Mesozoic formations; 2. Upper Triassic detrital sediments; 3. Middle Triassic "Muschelkalk"-type carbonates; 4. Middle Triassic "Röt"-type formations; 5. Lower triassic Buntsandstein-type Jakabhegy Sandstone; 6. Cretaceous formations; 7. Jurassic formations; 8. Paleozoic formations; 9. important structural line, 10. deep drilling

– The central and eastern Mecsek sequence where the units overlying the Zuhánya Limestone are very different from those of the other area (Figs 2, 3).

Laterally the most significant change is shown by the Rókahegy Dolomite and the Zuhánya Limestone. Determination of the facies of both formations is rendered more difficult by contradictory information, since phenomena referring to different water depths can be observed. The thickness conditions also change as in the case of the Villány Mountains the facies area is thicker than the average.

In general the lower boundary of the Rókahegy Dolomite cannot be drawn precisely because of the partial dolomization of the footwall. In the Villány Mountains, its thickness – as against the 20–35 m in the western Mecsek – increases from the northern foreland towards the southern slivers from 40 m to 270 m (Nagy and Nagy 1976). In borehole Máriagyüd-1 (Mgy-1) its thickness is 64 m and 86 m respectively by tectonic repetition, with graded breccia beds in five horizons. Breccias indicate intertidal erosion (Fig. 4/a), but some of

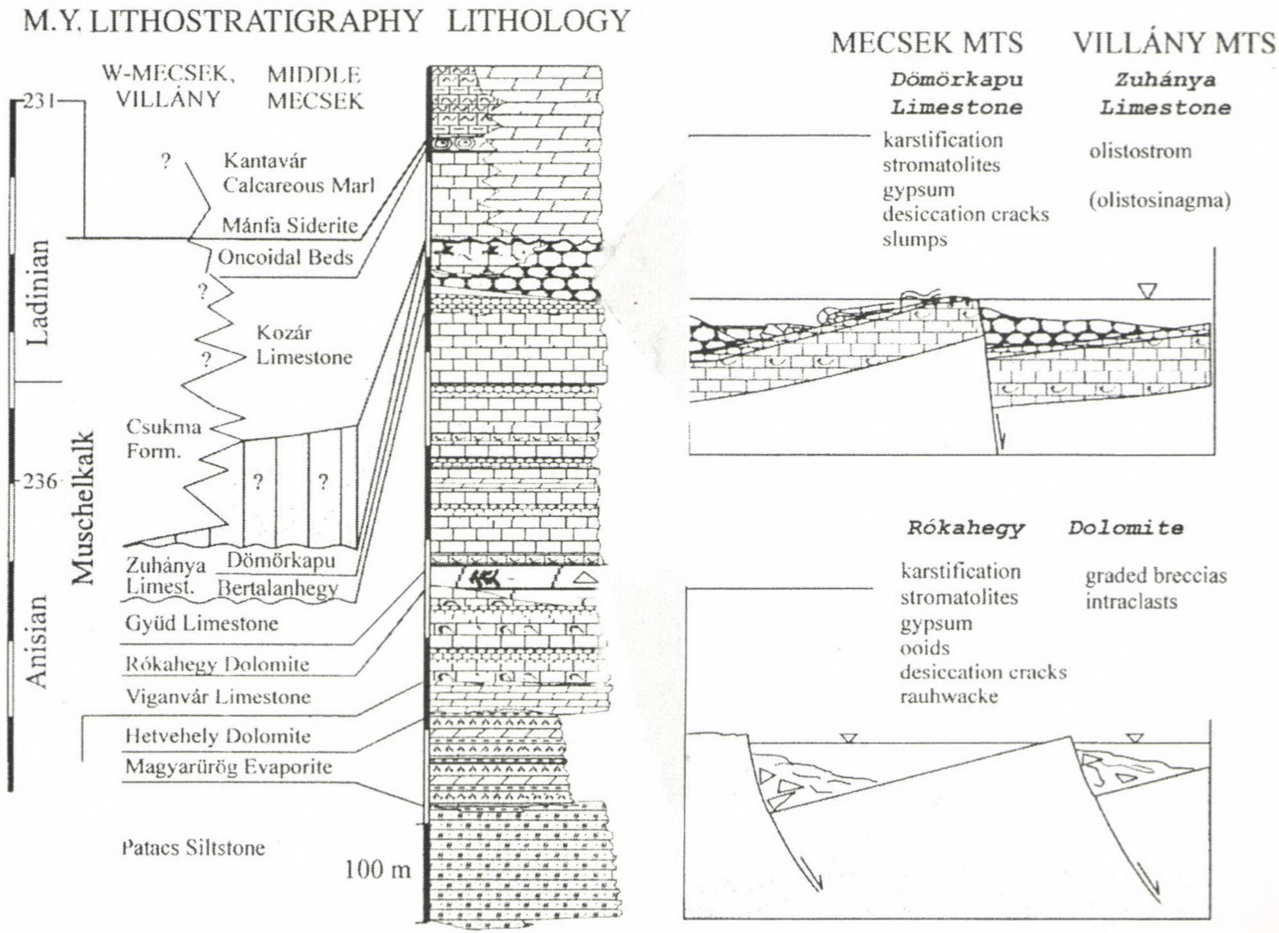


Fig. 2
Stratigraphic column of the Muschelkalk-type formations of SE Transdanubia with emphasis on the formations indicating extensional tectonics (with strong vertical exaggeration)

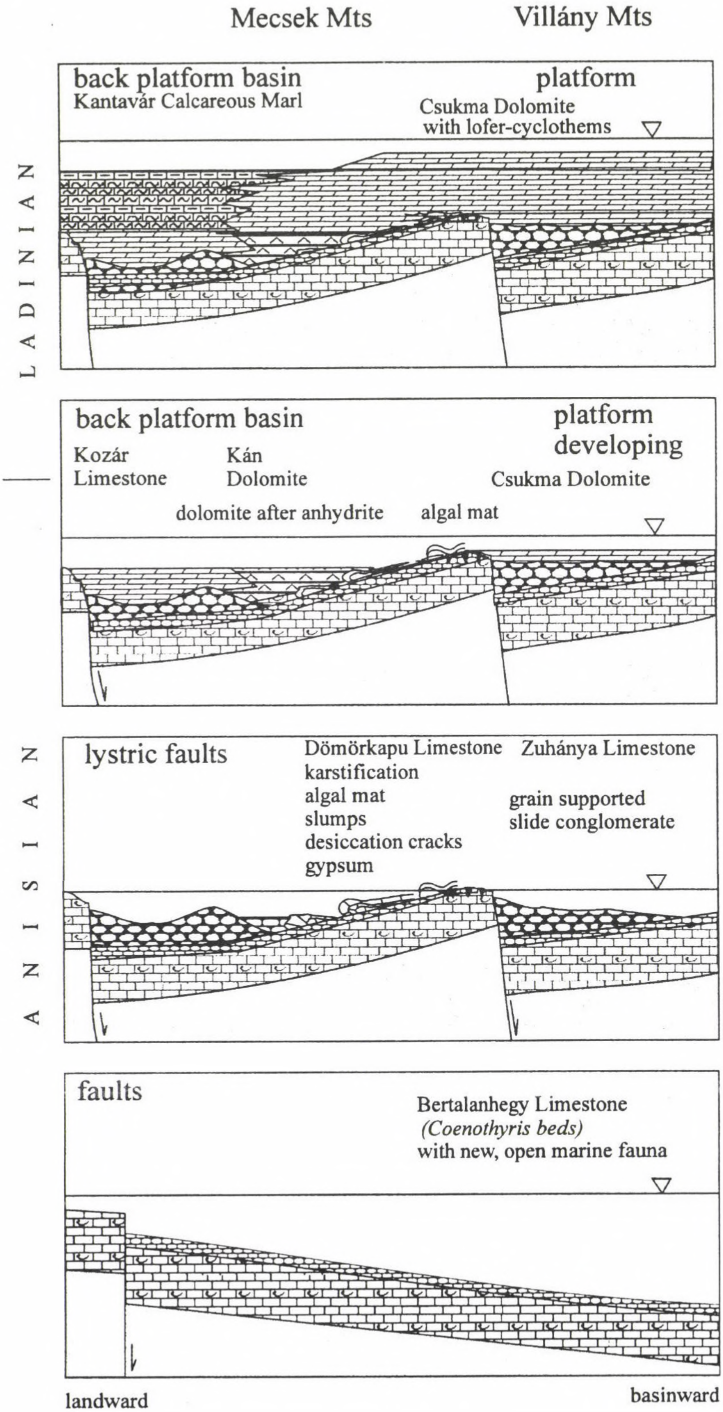


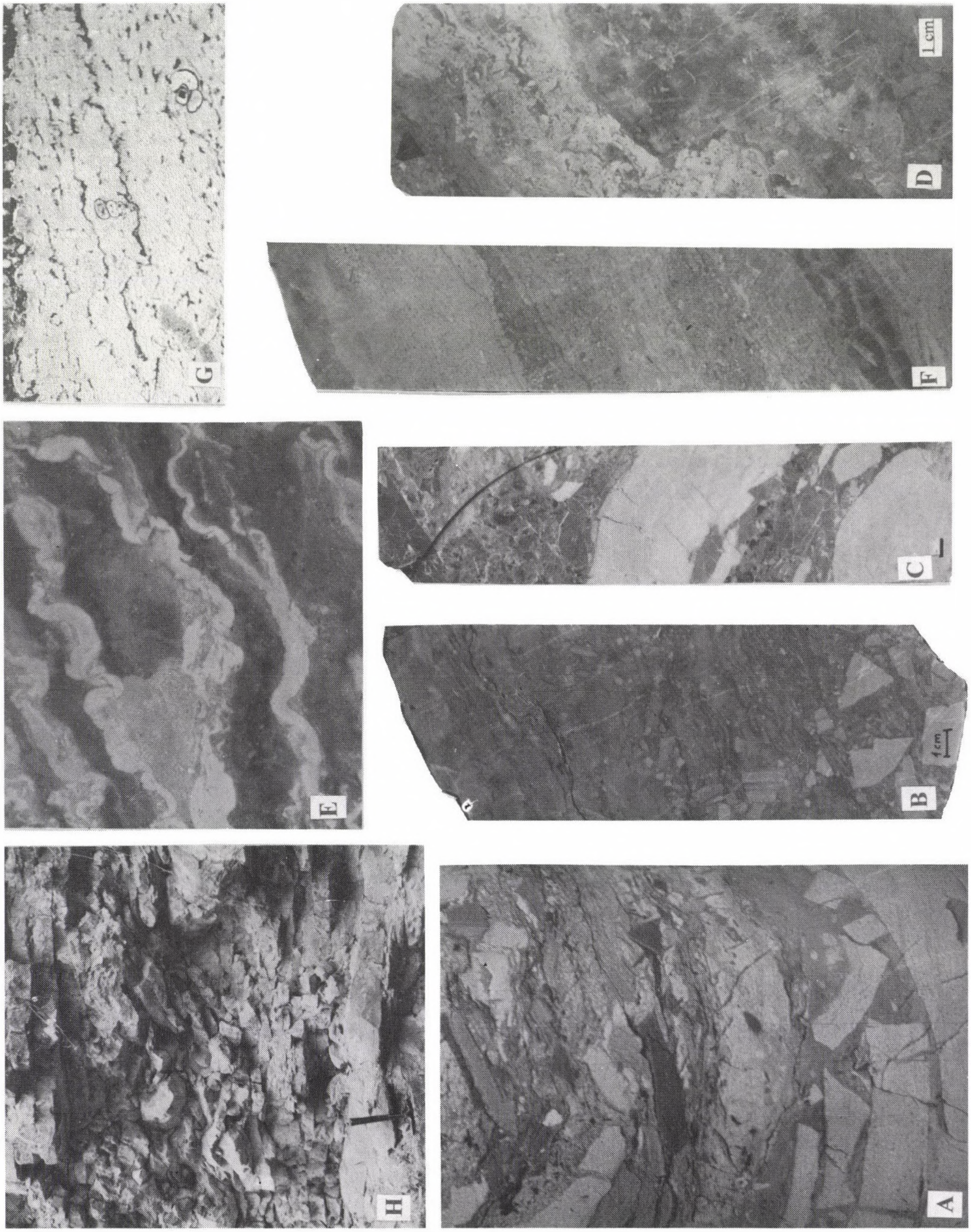
Fig. 3
Effect of the Pelsonian listric faults on the course of sedimentation in SE Transdanubia. Conceptual section with strong vertical exaggeration

several m thickness, graded, of scarp breccia character also occur (Fig. 4/b). Displacements along slopes, mudstreams and slumps can be observed. Above the breccias, oolites (oodolosparites) and small-sized oncoids appear. Pelsparites are frequent. In its poorish fauna, echinoderm shell fragments, a few bivalve shell fragments, radiolarians, rare sponge remnants, and foraminifera can be found. The Mecsek facies area consists of an alternation of fauna-barren, yellow, clayey and red, very thin crystalline layers. In this horizon, Thecosmilia coral reefs had been reported earlier (Böckh 1876; Kolosváry 1958), which are – according to our investigations – transformed remnants of gypsum precipitations, formed in supratidal conditions (Konrád 1997, in print). In its cover, calcareous dolomite rauhwacke with weathered out anhydrite pebbles can be found. In boreholes Gálosfa-1 and Somberek-1 claystones and even fine sandy siltstones occur in the formation.

Nagy E. (1968) describes the Rókahegy Dolomite ("boundary dolomite") of the Mecsek Mountains as epicontinental shallow marine sediments of decreasing salinity. On the basis of stromatolite structures, desiccation cracks, bioturbation and "reefal biostromes", they "were deposited in a peritidal and restricted environment (lagoon) periodically invaded by normal sea waters" according to Freytet and Cros (1984). According to Nagy and Nagy (1976) the intercalated, more clayey, laminated layers indicate shallow marine conditions, while the ooidal layers with shell fragments point to nearshore water with currents in the area of the Villány Mountains. Based on the microfacies investigations of Welsch (1992) it was formed in a characteristically peritidal environment. The alternation of supratidal–intertidal (desiccation structures, authigenic breccias) and subtidal (ooidal–peloidal sparites, crinoidal layers, and vermicular dolomites) formations indicates varying water depth. According to Rálisch-Felgenhauer and Török (1993) it was deposited in a shelf-lagoon zone of open water circulation.

The environment of formation of the Zuhánya Limestone Formation is also problematical (Nagy E. 1968; Nagy and Nagy 1976; Konrád 1990; Török 1993;). It was deposited on the footwall with a sharp boundary. Conodonts found in the rock material of *Paraceratites binodosus* reported by Detre (1973) are characteristic of the uppermost part of the Pelsonian (Kovács and Papsova 1986). In the Mecsek Mountains the unit can be divided into members (the Bertalanhegy Limestone and the Dömörkapu Limestone). Intercalations of the Bertalanhegy Limestone type also occur subordinately in the Villány facies area.

According to the description of Nagy and Nagy (1976), the unit is 130 m thick in the Villány Mountains (the Szava Hill exposure shows nodules, marl intervals and thin bedding, already reminiscent of the Bertalanhegy Limestone of the Mecsek Mountains). It is thick-bedded with bed thickness decreasing northward. A 13 m-thick bed in the Zuhánya Mine was described by Nagy and Nagy (1976) as an intraclastite. On the basis of the rock structure, the thick-bedded development, the subordinate character of the matrix and the varying lithofacies formed in different environments the unit can be regarded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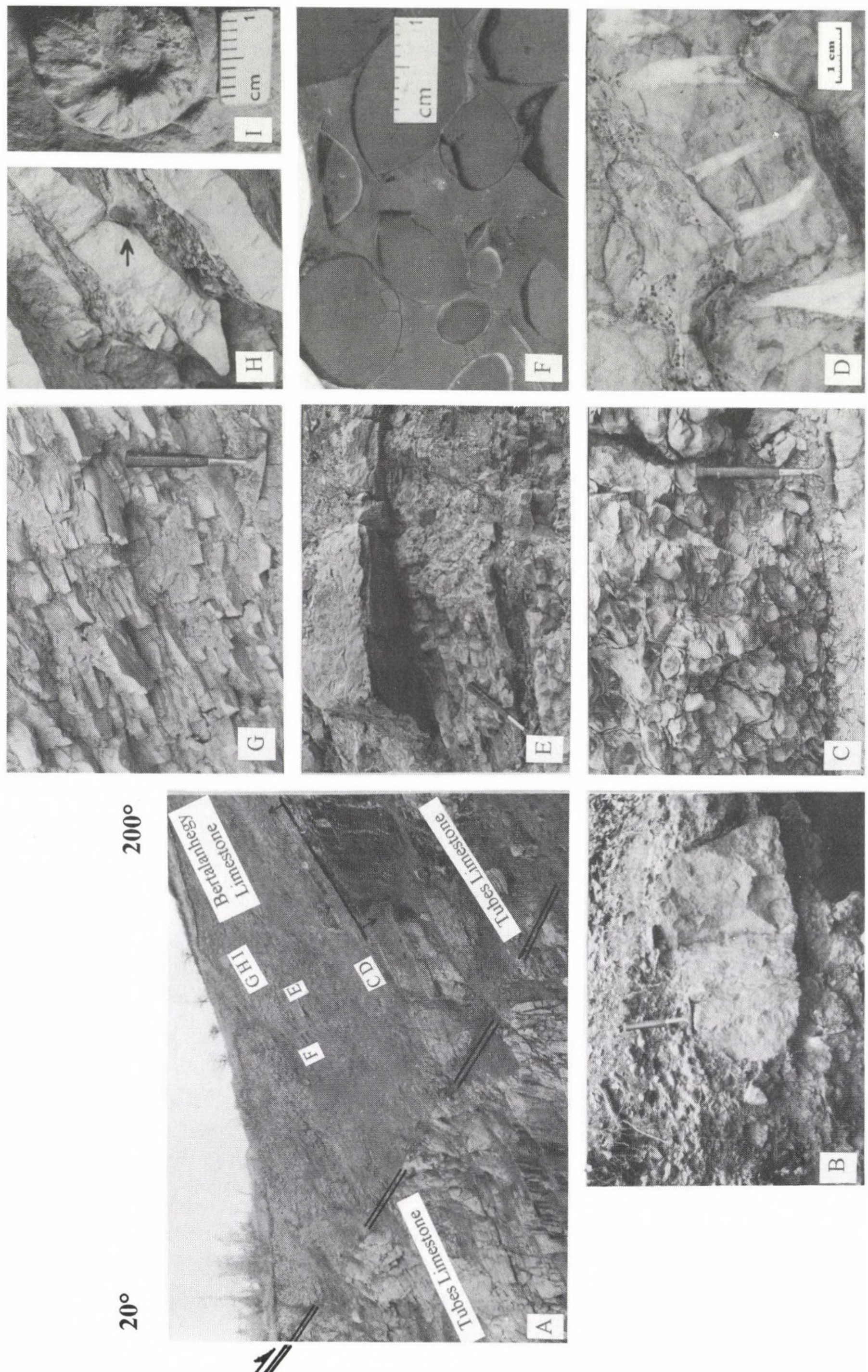
as a grain-supported slide conglomerate (Fig. 4/c). The matrix is a yellow, more rarely grey, clayey, equigranular microsparite. The material of the clasts varies; in some places it is of an extraclastite character. Within the clasts radiolarians occur and thin bivalve shells can be found in great numbers. Gastropod, ostracode and sponge remnants are rare. In the matrix only reworked fauna can be found, which is proved by the roundness and the tilted geopetal structure of the frequent brachiopods (*Coenothyris vulgaris, Tetractinella trigonella*). In the course of the diagenesis stilolites were formed on the contact surfaces of the intraclasts, and calcisparites in the cavities, which lack matrix. It is probable that the thick, pinching out fissures (Fig. 5/d) came into being during the reworking; they are assumed to be late diagenetic due to the lack of the matrix material and are filled up by calcite as clear as water (this can be observed very frequently in the clasts of the conglomerate). In the Mecsek Mountains the position and lithofacies of the Bertalanhegy Limestone Member of the Zuhánya Limestone Formation (Fig. 5) can be best studied in the exposure of the Bükkösd Quarry. Displaced along a fault, it is deposited on the Tubes Limestone Member of the Lapis Limestone Formation (Figs 5/a and 5/c) with a sharp boundary in the upper part of the quarry. Fig. 5/b shows the same formation in an exposure at Orfu. Over the "nodular" limestones clay marls and then a limestone bed are deposited (Fig. 5/e) with load casts on the lower bedding surface. Above it follow laminated limestones with clay and marl intercalations (Fig. 5/g). Brachiopod-bearing layers can be found in it. The brachiopods occur either as allochthonous tempestites or in parautochthonous position (Török 1993; Fig. 5/f). Sometimes the frequently varying infilling material, different from the host sediment, indicates allochthonous embedding and taphocoenosis. In the clay marls between the laminated limestones bivalves and brachiopods can be found in allochthonous position (in Fig. 5/h a *Plagiostoma lineatum* can be seen) and on the bedding planes of the limestone lamellas *Ceratites*-like organisms occur (Fig. 5/i). The pelagic fauna appears suddenly above the intraclastite and then disappears gradually. According to Bóna (1976) conodonts also appear suddenly, then after a decrease in number of specimens gradually disappear from the sequence. According to Nagy (1968), dominance conditions of fossils change both horizontally and vertically.

← Fig. 4

Lithofacies from the Rókahegy Dolomite, the Zuhánya Limestone and the Csukma Formation. A – Authigenic breccia in the Rókahegy Dolomite. Borehole Mgy-1, 795.7 m, x1.5; B – Graded breccia in the Rókahegy Dolomite. Borehole Mgy-1, 791.2 m; C – Grain-supported intraclastite in the Zuhánya Limestone. Borehole Mgy-1, 411.5 m, x0.4; D – One m-thick slump folds in the Zuhánya Limestone. Orfu, W of the Sárkány Well; E – Dolomitized enterolithic anhydrite in the Kán Dolomite. Borehole Gf-1, 1,221 m, x1.2; F – Member B of the Lofer cycle in the Csukma Dolomite. Borehole Mgy-1, 352.3 m, x0.5; G – Algal-stromatolite in member B of one of the Lofer cycles of the Csukma Dolomite (part of Fig. 3/f). Borehole Mgy-1, 352.4 m, x8; H – Member A of the Lofer cycle with black pebble from the Csukma Dolomite. Borehole Mgy-1, 351.2 m


20°
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Bertalanhegy Limestone
GHI
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The Dömörkapu Limestone of the Mecsek Mts. is of similar facies as the Zuhánya Limestone of the Villány facies area. On the basis of the borehole data and exposures, its thickness is 30–80 m. It is also a grain-supported intraclastite (Konrád 1990), the poorish matrix of which was compared by Nagy (1968) with the infilling of tropical karst cavities. Probable subaerial dissolution cavities can really be found which are infilled with yellow microsparite corresponding to the matrix of the intraclastite, and which abound in calcite pseudomorphs after gypsum and anhydrite, determined wrongly by Kolosváry (1958) as corals. Algal stromatolites with desiccation cracks and limestone layers with desiccation cracks also occur. Near the Sárkány Well of Orfű, one m-thick slump folds can be found (Fig. 4/d). In contrast with the general SW direction of the other slumps known from the Anisian this one had moved northward.

According to Nagy and Nagy (1976) during the formation of the Zuhánya Limestone significant differences in bottom relief had taken shape and material reworked from the higher regions created the "mottled" intraclastites. Radiolarians and thin-shelled bivalves indicate deeper water conditions. According to Török (1993), it is a deeper/outer ramp facies. According to Rálisch-Felgenhauer and Török (1993), based on the appearance of pelagic fauna elements and the frequency of *Glomospira densa* its nodular structure is caused by the open, deeper water shelf slope sediments and submarine mud slumps.

Tectonic interpretation

In general the Rókahegy Dolomite is regarded as the closing member of the 'Rot facies' in the lithostratigraphical classification of SE Transdanubia. However, the development of the carbonate sequence of Muschelkalk character had already begun with the footwall Viganvár Limestone. Thus, the Rókahegy Dolomite Formation indicates a break in the evolution of the carbonate ramp.

In the formation of the Rókahegy Dolomite and the Zuhánya Limestone the role of displacements along listric faults seems to be the crucial point. This is

← Fig. 5
Characteristics of the Bertalanhegy Limestone Member of the Zuhánya Limestone Formation in the Mecsek Mts. A – Occurrence of the Bertalanhegy Limestone in the upper horizon of the Bükkösd Quarry. B – Basal conglomerate of the Bertalanhegy Limestone W of the Sárkány well at Orfű. C – The Bertalanhegy Limestone is deposited with a sharp boundary and grain-supported conglomerate (olistosynagma) over the Tubes Limestone in the Bükkösd Quarry. D – Part of Fig. 5/c. Wedging crevices of the limestone clasts of the conglomerate formed during the resedimentation which were then were filled in by calcite as clear as water. E – On the eight m-thick conglomerate clay marl followed by a limestone bank, are deposited. F – Higher up in the sequence brachiopods appear, sometimes in rock-forming quantity. Autochthonous and allochthonous embeddings occur in equal amounts. G – In the upper part of the exposure, laminated limestone with clay marl and marl intercalations can be found with the richest fauna of the SE Transdanubian Triassic. H – *Plagiostoma lineatum* in clay marl. Part of the right edge of Fig. 5/g. I – *Ceratites* sp. on the bed plane of the laminated limestone

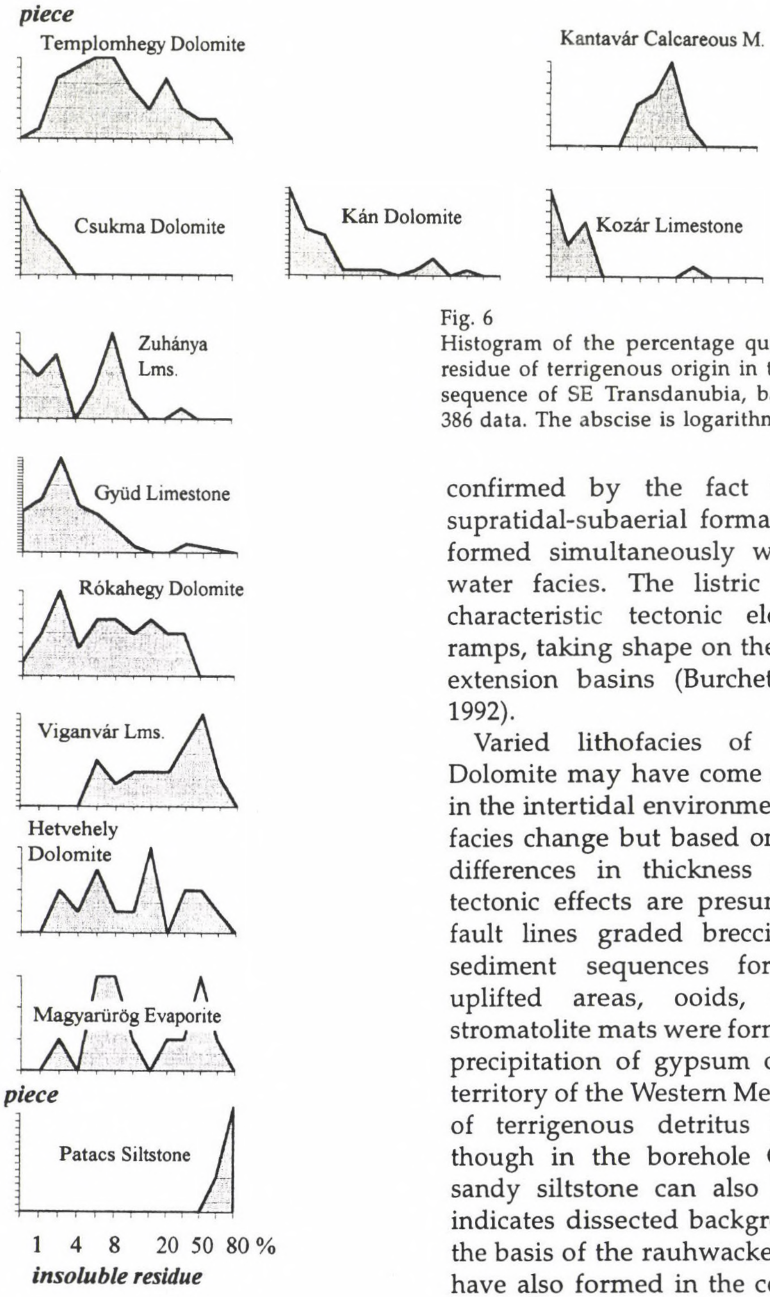

Fig. 6
Histogram of the percentage quantity of insoluble residue of terrigenous origin in the Middle Triassic sequence of SE Transdanubia, based on altogether 386 data. The abscise is logarithmic!

confirmed by the fact that intertidal-supratidal-subaerial formations were also formed simultaneously with the deeper water facies. The listric faults are the characteristic tectonic elements of the ramps, taking shape on the margins of the extension basins (Burchette and Wright 1992).

Varied lithofacies of the Rókahegy Dolomite may have come into being even in the intertidal environment due to lateral facies change but based on the significant differences in thickness synsedimentary tectonic effects are presumed. Along the fault lines graded breccias and thicker sediment sequences formed. On the uplifted areas, ooids, oncoids, alga-stromatolite mats were formed. In the latter precipitation of gypsum occurred. In the territory of the Western Mecsek the amount of terrigenous detritus is subordinate, though in the borehole Gálosfa-1 finely sandy siltstone can also be found. This indicates dissected background basins; on the basis of the rauhwacke evaporites may have also formed in the coastal sabkha. It also explains the early diagenetic dolomization resulting from the infiltration

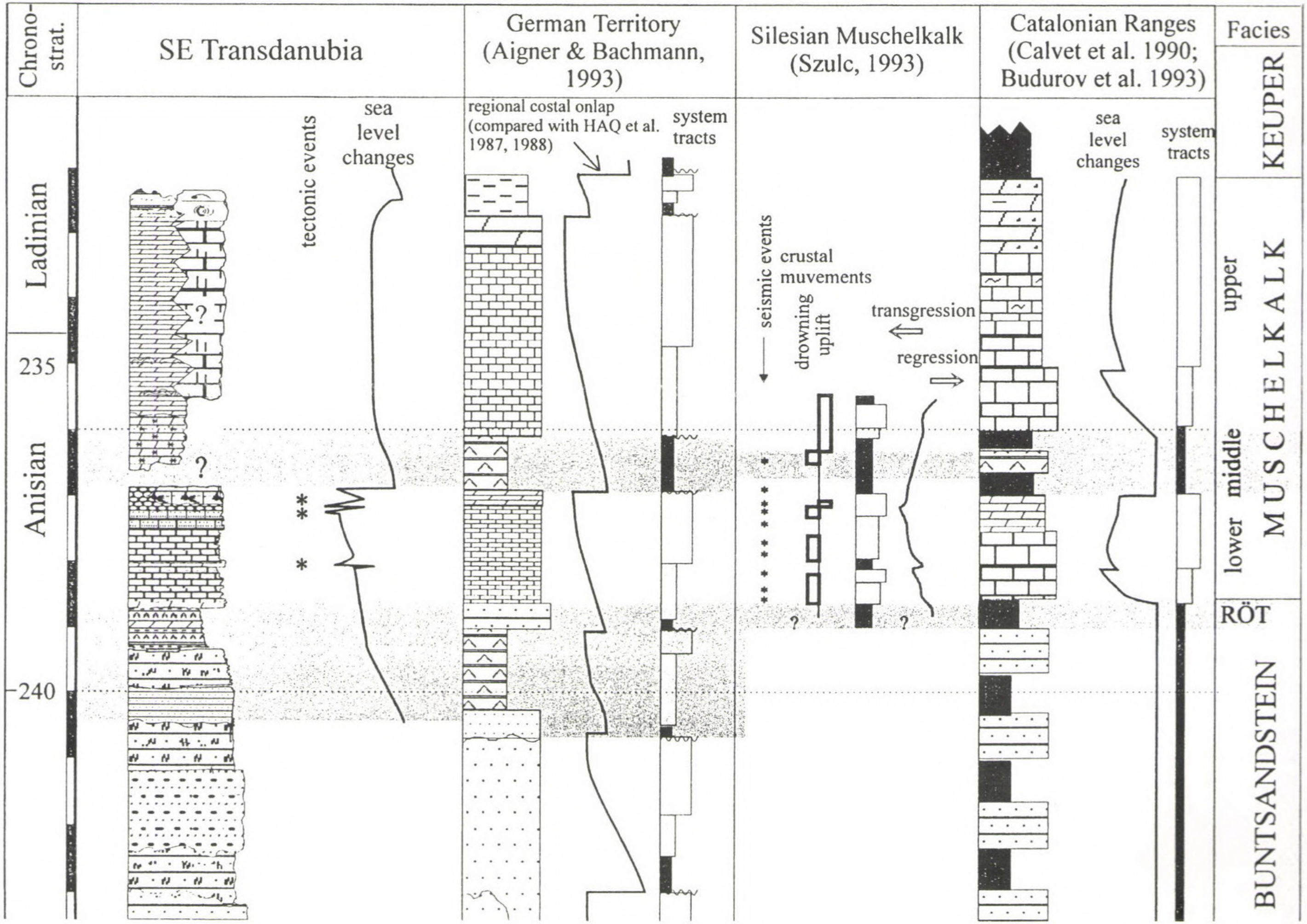

Fig. 7
Comparison of the Middle Triassic of SE Transdanubia with the sequence of the important German facies areas

of Mg-rich water which is associated with the formation and also affects the upper part of the footwall.

The lateral lithofacies change of the Zuhánya Limestone, the dissecting basement, the sudden appearance of the pelagic fauna, and the formation of the grain-supported intraclastite (with extraclastites) indicate significant synsedimentary tectonics.

Effect of the Pelsonian extension tectonics and the global water-level subsidence on sedimentation

Following the synsedimentary tectonic events the formations correlated with the middle Muschelkalk evaporite complex of the German facies areas were probably formed due to eustatic decrease in water level. In the Western Mecsek and borehole Gf-1, in the lower, sucrose, mottled part of the Kán Dolomite deposited over the Zuhánya Limestone, dolomitized relics of enterolithic, knotted anhydrite (Fig. 4/e) and algal-stromatolite desiccation structures can be recognized; the unit is characterized by the occurrence of late diagenetic Baroque dolomite. The dolomization may have also affected the upper part of the Zuhánya Limestone, especially its matrix of higher porosity. Sulfate reduction is marked by the occurrence of pyrite. This also explains the appearance of the pyritous dolomite horizon described in borehole Vágottpuszta-2 (Wéber 1978).

Between the open and relatively deep-water Zuhánya Limestone and the overlying Csukma Formation, and the Kozár Limestone respectively a hiatus can also be presumed from the lack of facies transition (see Fig. 2).

In the Villány facies area the Zuhánya Limestone is overlain by the Csukma Dolomite, the "purest" carbonate of the SE Transdanubian Triassic (Nagy and Nagy 1976). A change in the insoluble residue of the Anisian-Lower Ladinian formations can be seen in Fig. 6. The Csukma Dolomite was formed in isolation from the influx of terrigenous material. In accordance with this Lofer cycles of inner carbonate platform facies have been detected in borehole Máriagyüd-1: member A deposited on the erosion surface is composed of dark green, dark grey, frequently mottled clayey carbonate with slightly rounded carbonate pebbles, in some places with black pebbles (Fig. 4/h), followed by the loferite of member B (parallel, horizontal, fenestral algal-stromatolite with gastropod remnants and desiccation cracks – Figs. 4/f and 4/g). Member C is homogenous, dolomitized limestone, the original fauna and texture of which cannot be recognized.

On the basis of the above it can be concluded that the tectonic processes leading to the deposition of the Zuhánya Limestone created background basins of limited water circulation which were partly restricted by blocks tilted along listric faults. Eustatic subsidence of water level also contributed to its formation. The regularly deepening carbonate ramp ceased to exist. Varied, heteropic formations came into being (Fig. 3): the later dolomitized evaporite (part of the Kán Dolomite), the Kozár Limestone, the paludal Mánfa Siderite, the Lower Ladinian lacustrine coaly formation (Wéber 1978), the Kantavár Calcareous

Marl and the "wengen schist". In the foreland (over the Villány facies area) the formation of a carbonate platform began. Toward the open ocean it was bordered by the Wetterstein limestone reef facies of the Bácska area (Bleahu et al. 1994).

Correlation of the extensional tectonic events

On the basis of the ammonite, brachiopod and conodont fauna of the Bertalanhegy Limestone (Detre 1974; Kovács and Papsova 1986; Pálfy and Török 1992), of the effects of the eustatic water-level fluctuations, and of the lithologic correlation the series of events leading to the formation of the Zuhánya Limestone is coeval with the Pelsonian-Illyrian synsedimentary tectonics (Fig. 7). The latter have been described in the Northwestern Calcareous Alps and in Germanic facies areas (Szulc 1993), in the Southern Alps and the Balaton Highland (Budai and Vörös 1992, 1993). Wéber (1965, 1978) first mentions the green clay intercalations of the Kán Dolomite, which were described by him as of volcanic tuff origin (characterized by illite and kaolinite with high Ti-, Cu-, Pb-, Co-, Ni-content) on the basis of the results of material investigations. Based on the sequences of boreholes Gf-1 and Mgy-1 the green, clayey dolomite intercalations proved to be the A member of the Lofer cycles. If further investigations demonstrate their volcanic origin they could be identical with the Lower Ladinian tuff occurrences known in the Southern Alps and the Pelsonian Unit (Szabó and Ravasz 1970; Wéber 1978; Budai and Vörös 1993).

Summary

Two formations of the Middle Triassic regularly deepening carbonate ramp facies of SE Transdanubia show features indicating synsedimentary tectonic events: the Rókahegy Dolomite and the Zuhánya Limestone.

In the case of the Rókahegy Dolomite Formation, tectonic effects are presumed based on the significant differences in thickness and the lateral facies changes of lithofacies referring to an intertidal environment.

The Zuhánya Limestone Formation formed as a result of movements which can be emplaced at the end of the Pelsonian. Due to the dismembering of the basin and the eustatic decrease in water level the basin became strongly differentiated and the regularly deepening ramp character ceased to exist. Varied, heteropic lithofacies took shape. A hiatus can also be presumed. In the sabkha environment which was later dolomitized, evaporitic material (lower part of the Kán Dolomite) correlatable with the middle Muschelkalk came into being. Carbonate platform facies formed (Csukma Dolomite) and in the background basins bituminous limestone (Kantavár Calcareous Marl) and terrigenous detritus ("wengen silt") deposited.

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A new thalamid sponge from the Upper Triassic (Norian) reef limestones of the Antalya region (Turkey)

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The small thalamid sponge *Discosiphonella minima* n. sp. is described from four reef block samples at three different localities near Antalya (southern Turkey). The relationship between Paleozoic and Upper Triassic representatives of *Discosiphonella* are discussed.

Key words: Systematic Paleontology, Sponges, "Sphinctozoa", *Discosiphonella*, Upper Triassic, Norian, Antalya, Turkey

Introduction

The majority of Permian thalamid sponges, which are one of the main reef-builder organism groups in the Permian, disappear by the end of the Paleozoic owing to Late Permian mass extinction. The Paleozoic sphinctozoid genera disappear at the end of the Permian and do not occur in Lower and Middle Triassic deposits. Like all other reef-builders the number of sphinctozoid sponge species increases during the Triassic reaching their highest diversity in the Norian/Rhaetian (Senowbari-Daryan 1990; Riedel and Senowbari-Daryan 1991; Flügel 1994).

In the Carnian only a few, but in the Norian/Rhaetian many more sphinctozoid sponge taxa show similarities or have identical features to those which disappeared in the Late Permian. However, sponges with the same morphological features may be totally different in microstructure and chemical composition of the rigid skeleton, as well as of the spicular skeleton if it was secreted primarily. An example of this is the Paleozoic genus *Cystothalamia* Girty, 1908, with multi-layered, cyst-like chambers around an axial spongocoel. Sponges with the same morphological features and chamber construction also occur in the Triassic (e.g., *Cystothalamia bavarica* Ott 1967, and *Cystothlamia slovenica* Senowbari-Daryan 1982) and were also assigned to *Cystothalamia* by other authors before 1990. The Triassic representatives differ from Permian *Cystothalamia,* which have an aragonitic skeleton with a spherulitic

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microstructure, by their high-Mg calcite skeletal mineralogy and granular microstructure (see Senowbari-Daryan 1990: p. 54). For this reason Senowbari-Daryan (1990) proposed the genus *Alpinothalamia* for the Triassic *Cystothalmia*-like thalamid sponges with a skeletal mineralogy composed of high-Mg calcite.

Discosiphonella Inai 1936, (synonyms: *Ascosymplegma,* Rauff 1938; *Cystauletes,* King 1943; *Lichauanospongia,* Zhang 1983; see Senowbari-Daryan 1990) is another sponge genus which disappears in the Late Permian and reappears in Norian/Rhaetian reefs. The morphologically similar Triassic genera (*Prethalamopora,* Russo 1981; *Diecithalamia,* Senowbari-Daryan 1990) and Jurassic/Cretaceous genus (*Thalamopora,* Römer 1841) as well as the Paleozoic/Triassic genus *Discosiphonella* Inai and their morphological and structural differences are discussed by Senowbari-Daryan (1990).

The abundant Permian genus *Discosiphonella* is a rare sponge in Upper Triassic reefs. The occurrence and geographic distribution of Triassic species is discussed later.

A variety of thalamid and non-thalamid sponges occurs in Triassic reef limestones near Antalya in southern Turkey. However, only limited reports of this group are known from this area. Cuif and Fischer (1974), Cuif et al. (1972), Senowbari-Daryan (1990, 1994), Riedel (1990) and Cremer (1993, 1994) described some taxa from the Norian reef boulders of this region.

A monoglomerate sponge, morphologically identical with representatives of the Paleozoic genus *Discosiphonella,* was found in some reef blocks near Antalya in south Turkey. This sponge is described here as *Discosiphonella minima* n. sp. The material investigated is deposited in the Staatssammlung für Paläontologie und Historische Geologie (SPHG), München [Invertar-Nr.: SPHG-SDL (Senowbari-Daryan and Link) 1997].

Geologic setting and localities

Upper Triassic sediments in the Taurides in south Turkey were deposited in local shallow basins (Flügel and Link 1996). These sediments exhibit small-sized lithological units preponderantly formed by the input of terrigenous detritus and carbonate production (Gutnic et al. 1979; Flügel and Link 1996).

In this dynamic environment the activity of frame-builders only occasionally produced differentiated reef buildups characterized by patch reefs of special living communities and dominated by sponges. Owing to a high mobility of the sea bottom many patch reefs were reworked and transported as boulders ("Cipits") into the deeper area of the basin. Here the transported reef boulders and consequently the reef organisms were imbedded in marly sediments and are well preserved. Sponges are the most abundant organisms in these boulders.

The new thalamid species described here was found in four samples from three different localities belonging to two different tectonic units (Fig. 1). Sample CS 1 was found at the base and sample CS 22 at the top of a small hill called

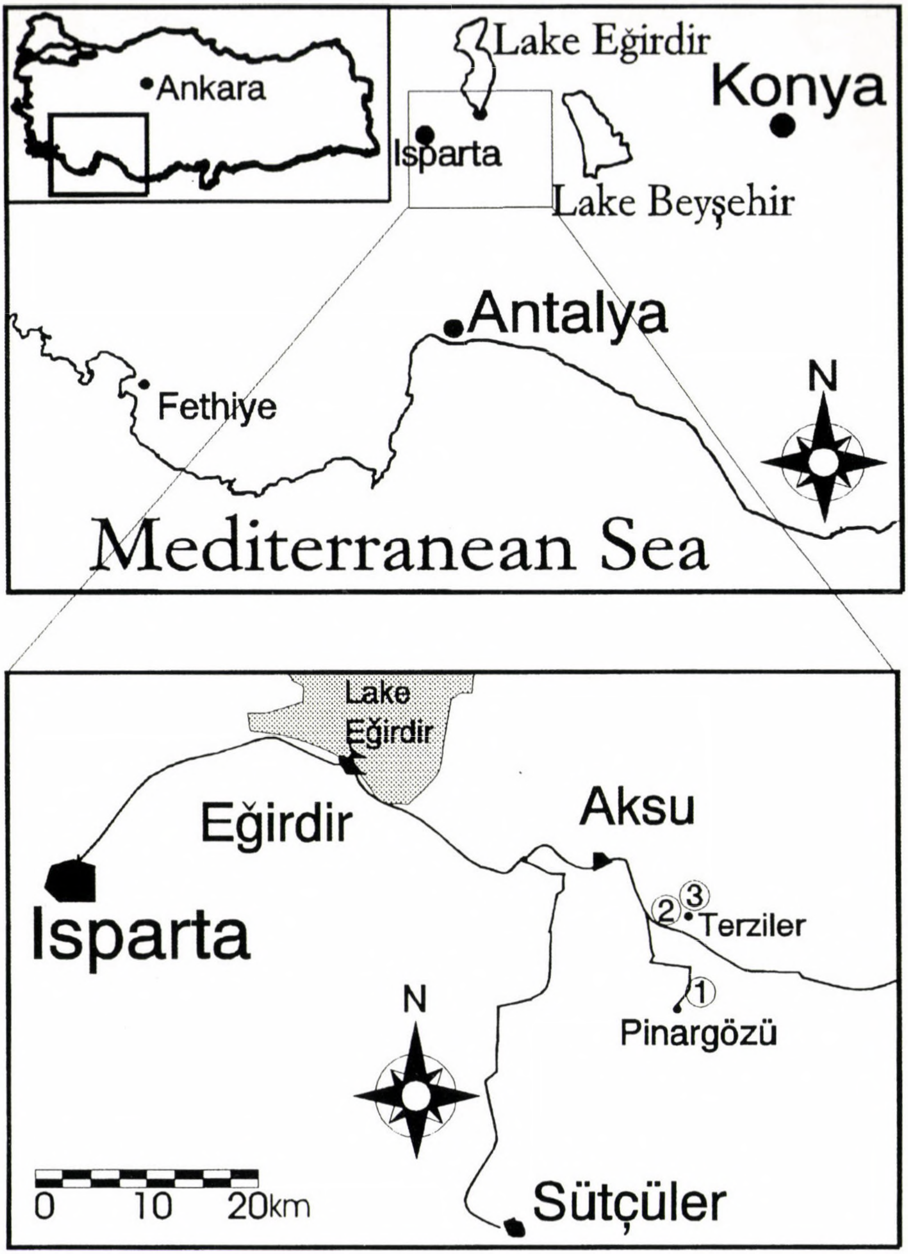


Fig. 1
Location map of the study area. Outcrops referred to in the text: 1. Caglarseki Tepe; 2. Road Aksu-Terziler; 3. Terziler

Caglarseki Tepe, about 200 m north of the village of Pinargözü (Senel et al. 1992). This locality is situated at the northern border of the parautochthonous unit of Beydaglari–Karacahisar. Sample TE 6 was found as talus north of the village of Terziler at the margin of the Carnian locality reported by Riedel (1990). However, this sample actually belongs to the Norian coming from a small local tectonic unit. Riedels locality belongs to the Anamas–Akseki autochthonous unit while our sample comes from the northern Antalya nappe. Both are representatives of the same depositional environment. The outcrop of sample AK 4 is located on the road from Aksu to Terziler just about 1.5 km south of Terziler. This locality is also part of the northern Antalya nappe.

Systematic Paleontology

Class Demospongea Sollas, 1875

Subclass Ceractinomorpha Levi, 1973

Order Permosphincta Termier and Termier, 1974

Suborder Porata Seilacher, 1962

Family Sebargasiidae Laubenfels, 1955

(pro Sphaerosiphniidae Steinmann, 1882)

Subfamily Cystothalamiinae Girty, 1908

Remarks

Belyaeva (1991: in Boiko et al.) and Wu (1991) independently established the family Cystauletidae, as discussed by Senowbari-Daryan and Ingarat-Helmcke (1994: p. 16): "First we want to give the name Cystauletidae Belyaeva (non Cystauletidae Wu) priority because she named the type genus and defined the family Cystauletidae. However, the genus *Cystauletes* King 1943, was synonymized with *Discosiphonella* Inai 1936, by Senowbari-Daryan (1990)".

Wu (1991) also introduced the family Imbricatocoeliidae with the type genus *Imbricatocoelia* Rigby, Fan and Zhang, 1989. The species *Cystothalamia ramosa* Senowbari-Daryan and Rigby, described from the Upper Permian of Djebel Tebaga in southern Tunisia, was placed in the genus *Imbricatocoelia* by Wu. We agree with this placement, but *Imbricatocoelia paucipora* Rigby, Fan and Zhang, 1989 is the type species of this genus and not *C. ramosa* as designated by Wu. Boiko (1991: in Boiko et al.) established the family Ascosymplegmatidae with the type genus *Ascosymplegma* Rauff, 1938 and an additional genus *Subascosymplegma* Deng 1981.

Belyaeva (1991: in Boiko et al.) assigned the following genera to the family Cystauletidae: *Cystauletes* King 1943, *Polycystocoelia* Zhang 1983, *Lichuanospongia* Zhang 1983, and *Squamella* Belyaeva 1991.

The type species of *Squamella* is *S. lichatchevi* Belyaeva (1991: in Boiko et al,: pp. 106–107, pl. 22, figs 4–5). *S. lichatchevi*, however, is representative of the genus *Imbricatocoelia* Rigby, Fan and Zhang, 1989, as demonstrated by the perforation of the chamber walls. Therefore the genera *Imbricatocoelia* and *Squamella* are synonymous and *Imbricatocoelia* has a priority.

The genera *Discosiphonella* Inai 1936, *Cystauletes* King 1943, *Ascosymplegma* Rauff 1938, and *Lichuanospobngia* Zhang 1983, were synonymized by Senowbari-Daryan (1990). The unification of the mentioned genera and the problem with the range of the family Cystauletidae Belyaeva (1991: in Boiko et al.) and Wu (1991) is discussed by Senowbari-Daryan and Ingavat-Helmcke (1994: 16). We think the following families are synonymous with Cystothalamiidae Girty: Cystauletidae Belyaeva, 1991 (in Boiko et al.), Cystauletidae Wu 1991, and Imbricatocoeliidae Wu 1991. The genera assigned to the family Ascosymplegmatidae Boiko should be placed in the family Cystothalamiidae Girty (e.g., *Ascosymplegma* Rauff) or in the family Colospongiidae Senowbari-Daryan 1990 (*Subascosymplegma* Deng 1981).

Following the systematic classification of Senowbari-Daryan (1990), Rigby, Fan and Han (1995) have described a new species – *Cystauletes crassa* – from the Upper Permian of South China. The authors introduced the generic name *Cystauletes* because "these sponges (holotypes and paratypes of *Discosiphonella manchuriensis* Inai, which are missing: see Rigby et al. 1995) have a relatively simple perforate endowall, in a structure less complex than the canalled wall in *Cystauletes mammilousus* King 1943, from the Upper Carboniferous of Oklahoma" (Rigby et al. 1995: p. 238). Therefore the question of whether or not *Cystauletes* King in an independent genus remains uncertain.

Genus *Discosiphonella* Inai, 1936

Synonyms: Ascosymplegma Rauff, 1938; *Cystauletes* King, 1943; *Lichuanospongia* Zhang, 1983

Type species: Discosiphonella manchuriensis Inai, 1936

Additional species (up to 1990): see Senowbari-Daryan (1990: p. 58, tab. 8)

Additional species (after 1990): Described as *Cystauletes* or *Lichuanospongia* after 1990: ?*Cystauletes squamilis* Belyaeva (1991: in Boiko et al.) – Permian of the Far East; *Cystauletes primoriensis* Belyaeva (1991: in Boiko et al.) – Permian of the Far East; *Cystauletes bzhebsi* Belyaeva (1991: in Boiko et al.) – Upper Triassic of the Northern Caucasus; *Cystauletes grossa* Rigby, Fan and Han (1995) – Upper Permian of southern China; *Lichuanospongia primorica* Belyaeva (1991: in Boiko et al.) – Permian of the Far East.

Cystothalamia crassa Bylaeva (1991: in Boiko et al.) – Upper Permian of the Far East. This sponge seems to have single-layered chambers and therefore could perhaps be a species of *Discosiphonella*.

Discosiphonella minima n. sp.

(Figs 2A–D, 3A–D)

Derivation of name: *minimus* (Lat. = small), because of the small size of the sponge compared with other *Discosiphonella* species.

Holotype: The axial longitudinal section figured in Fig. 2A (SPHG: Senowbari-Daryan and Link, 1997)

Paratypes: All specimens figured in Fig. 2B–C, Fig. 3A–D.

Type horizon and type locality: Norian, floated carbonate boulder found north of the village of Terziler (Fig. 1).

Diagnosis: Small species of the genus *Discosiphonella* with single-layered cyst-like chambers arranged around a relatively wide axial spongocoel. The diameter of the spongocoel is about 50% of the entire sponge diameter. In the longitudinal sections the chambers are oval. 12 to 17 chambers are arranged on one whorl. The spherulitic microstructure of the rigid skeleton indicates the aragonitic mineralogy of the rigid skeleton.

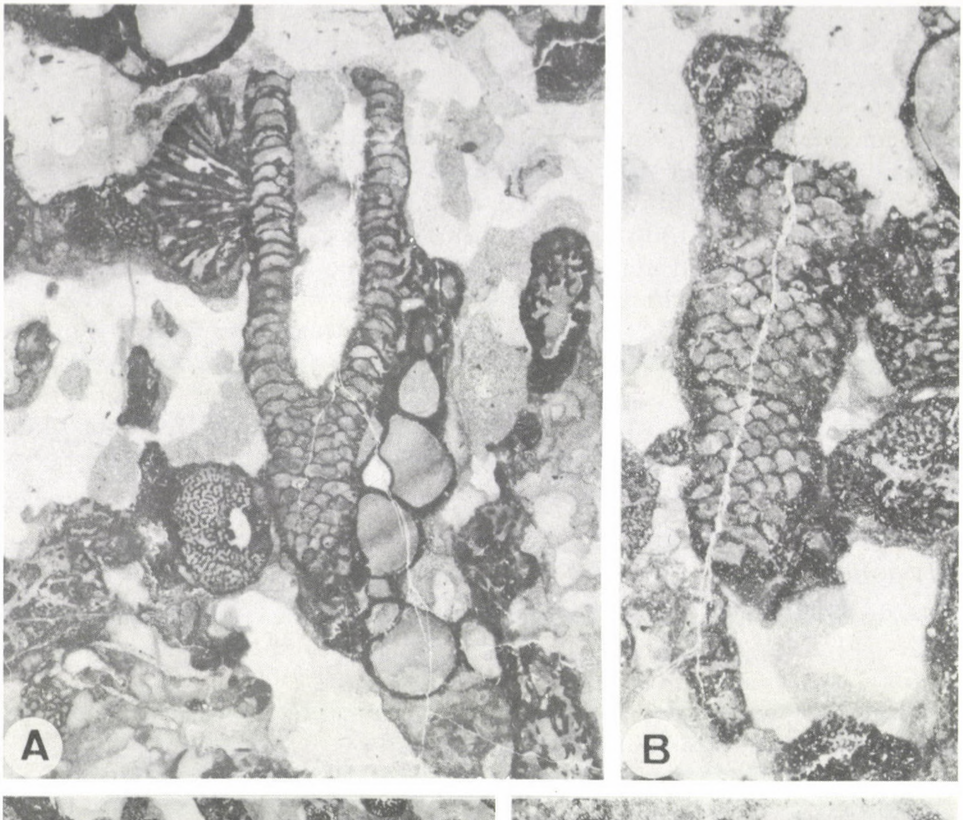
Differential diagnosis: See discussion following the species description.

Material: Several specimens in natural weathered rock surface, polished slabs and thin sections from sample TE 6 (Terziler: holotype), CS 1 and CS 22 (Caglarseki Tepe), and two specimens in thin section AK 4 (road from Aksu to Terziler, Fig. 1).

Description

The single and unbranched stems of this sponge are composed of numerous single-layered cyst-like chambers arranged around an axial spongocoel. The individual specimens generally reach diameters of approximately 5 mm. The holotype (Fig. 2A) is one of the largest specimens with a diameter of 7 mm

Fig. 2 →
Discosiphonella minima n. sp. from the Upper Triassic (Norian) reefal limestones near Antalya, Southern Turkey. All the specimens are in polished slabs. **A)** Holotype. Longitudinal axial section through a relatively large sponge exhibiting the wide axial spongocoel and the single-layered cyst-like chambers arranged around the spongocoel. The sponge is encrusted by another sphinctozoid (lower right) and chaetetid sponges (upper left). *Discosiphonella minima* n. sp. and other associated fauna are typically preserved in boundstone. SPHG-SDL-TE6, x3. **B)** Marginal section through the chambers of the sponge wall shows the honeycombed chambers. SPHG-SDL-TE6, x3.5. **C)** Oblique cross-section (1) and marginal longitudinal section (2). Different sphinctozoid, inozoid and chaetetid sponges which are associated with *Discosiphonella minima* n. sp. SPHG-SDL-TE 6, x1.5. **D)** Oblique section through a fragment of a specimen shows single-layered honeycombed chambers (lower part). SPHG-SDL-TE 6, x6


A
B
C
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and a length of 25 mm. It is composed of numerous single-layered cyst-like chambers arranged around the axial spongocoel of approximately 3.5 mm in diameter. In the holotype and in all other specimens the relatively wide spongocoel reaches almost 50% of the whole sponge diameter.

In longitudinal sections the chambers are oval to crescent-like in shape, attaining a maximum height of 0.8 mm and a diameter of about 2 mm. In oblique sections through the sponge wall, the chambers appear honeycombed to hemispherical (Fig. 2B). The cross-sections and oblique sections (Figs 2D, 3B, 3D) show single-layered chambers arranged around the axial spongocoel. The number of oval chambers in each whorl varies between 12 and 17.

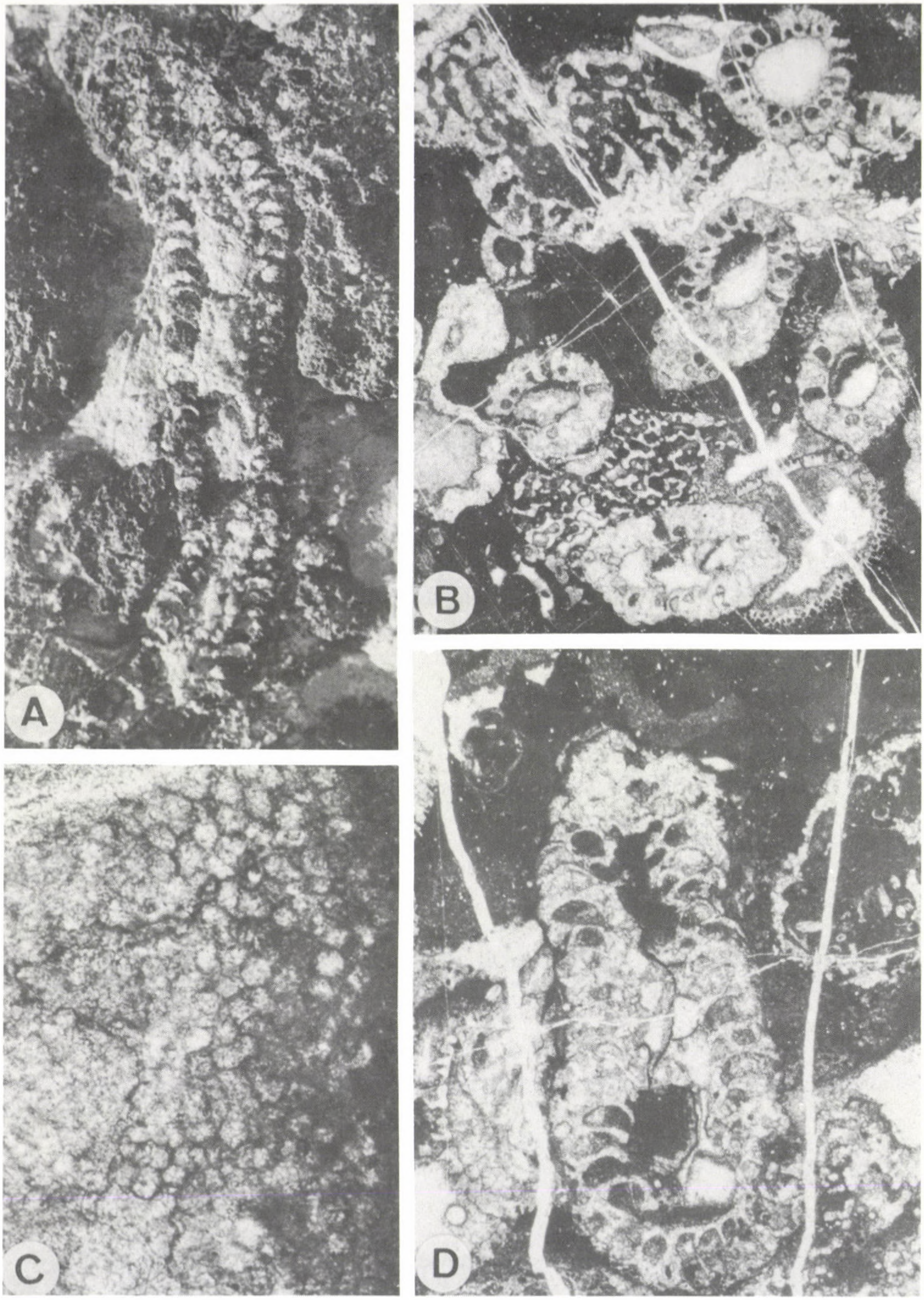
The chamber exowall, interwall and endowall are pierced by pores of the same size as the thickness of the chamber walls. In some specimens (Fig. 2D) the outer wall is thicker than the interwalls or endowalls. The chamber interiors are completely intact, with no vesiculae or filling structure. The spherulitic microstructure of the rigid skeleton indicates the aragonitic mineralogy (Fig. 3C). A spicular skeleton is not known.

Discussion

Sphinctozoid sponges with a rigid aragonitic skeleton such as the genus *Discosiphonella* Inai are rarely known from the Norian/Rhaetian. *Discosiphonella* is not known from Middle Triassic or Carnian reefs.

Rauff (1938) was the first author to describe the species *Discosiphonella torosum* (= *Ascosymplegma torosum*) from the Triassic of the Peruvian Andes. Zankl (1969) reported the occurrence of another species of *Discosiphonella* (described as *Cystauletes* sp. by Zankl) from the Norian/Rhaetian Dachsteinkalk reef of the Hohe Göll (Austria). Boiko (1991: in Boiko et al. p. 169, pl. 63, figs 6–7) described another species of the genus *Ascosymplegma* as *A. caucasicum* from the Upper Triassic (Norian) of the Northern Caucasus. However, the poorly documented species *A. caucasicum* is composed of tube-like chambers (pl. 63, fig. 5 in Boiko et al., 1991) or egg-shaped chambers arranged moniliform or glomerate (pl. 63, fig. 6 – Boiko et al., 1991). *A. caucasicum* is excluded from *Discosiphonella* and is not discussed below. This sponge should be placed into the family

Fig. 3 →
Discosiphonella minima n. sp. from Upper Triassic (Norian) reefal limestones near Antalya in southern Turkey. Specimens are in weathered rock surfaces or thin sections. **A)** Longitudinal section of the weathered rock surface shows the wide spongocoel and the single-layered oval chambers arranged around the axial spongocoel. Natural weathered rock surface. SPHG-SDL-CS 1, x4. **B)** Cross- and oblique sections of several specimens exhibiting the single-layered cyst-like chambers around an axial spongocoel. Some specimens exhibit geopetal structures. Thin section. SPHG-SDL-CS 22/1, x4. **C)** Wall structure of *Discosiphonella minima* showing well-preserved small spherulites. Thin section. SPHG-SDL-CS 22/1, x7. D) Oblique to longitudinal section exhibits the single-layered and oval chambers arranged around the axial spongocoel. Thin section. SPHG-SDL-CS 22/1, x7


A
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Colospongiidae and attributed to *Cinnabaria* Senowbari-Daryan, 1990, *Platythalamiella* Senowbari-Daryan & Rigby, 1988, or *Neoguadalupia* Zhang, 1987. The third Triassic species of *Discosiphonella* is described by Belyaeva (1991: in Boiko et al. p. 165, pl. 59, figs 8–9) as "*Cystauletes*" *bzhebsi* from the Northern Caucasus.

Discosiphonella minima n. sp. differs from the all Paleozoic species listed by Senowbari-Daryan (1990: p. 58, tab. 8) and from *Cystauletes grossa* Rigby, Fan and Han (1995) by its small dimensions. A small species of *Discosiphonella* with a sponge diameter of 3–4 mm, described by Senowbari-Daryan and Ingavat-Helmcke (1994) from the Upper Permian of the Phrae province in Northern Thailand, has almost the same size but differs from *D. minima* in the size of its spherical chambers.

All the species described as *Lichuanospongia* or *Cystauletes* by Belyaeva (1991: in Boiko et al.) are extremely poorly documented. As discussed earlier, the species ?*Cystauletes squamilis* Belyaeva is a representative of the genus *Imbricatocoelia* Rigby, Fan and Zhang, 1989, and should be excluded from *Discosiphonella*. *Lichuanospongia primorica* Belyaeva (1991: in Boiko et al. p. 103, pls. 2–3) is described only by a marginal section which does not permit an affiliation to *Discosiphonella*. This could be a species of *Cystothalamia* or another genus. It would be better to limit this "species" just to this section until a detailed description and more documentation are available. The outer size of *Cystauletes primoriensis* Belyaeva described from the Upper Permian of Primorije is almost identical to *D. minima* n. sp. but the last species has a very wide spongocoel (approximately 50% in *D. minima*, about 30% in *D. primoriensis*). The following Triassic species of *Discosiphonella* (including *Ascosymplegma* Rauff) have been described in the literature:

Discosiphonella torosum (= *Ascosymplegma toruosum* Rauff, 1938)
Discosiphonella sp. 1 (= *Cystauletes* sp. 1, Zankl, 1969)
Cystauletes bzhebsi Belyaeva (1991: in Boiko et al.)

Discosiphonella minima differs from both Triassic species described by Rauff and Zankl by its small size and the sponge elements.

Discosiphonella bzhebsi (= *Cystauletes bzhebsi*) described as a new species by Belyaeva (1991: in Boiko et al., pl. 59, figs 8–9) is documented by two sections. The species of fig. 8 is an *Amblysiphonella* and should be excluded from the genus *Discosiphonella* or *Cystauletes*. The specimen in fig. 9 cannot be assigned with certainty to any genus. The shape of the chamber and the perforation pattern of the chamber walls of this species are totally different from those in our species from southern Turkey.

Organism Association

The following organisms are associated with *Discosiphonella minima* n. sp. within the same rocks: Sphinctozoid (*Antalythalamia riedeli* Senowbari-Daryan 1994, "*Stylothalamia*" sp., *Colospongia* sp., *Paradeningeria* sp., thalamid sponge gen. et sp. indet.), inozoid (*Peronidella* sp. 1, *Peronidella* sp. 2, inozoid sponges gen. et sp. indet. 1 and 2) and chaetetid sponges gen. et sp. indet. In thin section AK 4 (Aksu–Terziler road locality) *Discosiphonella minima* is also associated with *Colospongia* sp., *Uvanella rhaetica* (Senowbari-Daryan and Schäfer), *Microtubus communis* Flügel, *Ophthalmidium* sp., and typical Norian/Rhaetian crusts of "Spongiostromata". The organism association points to a Norian age of the rocks.

Geographic and stratigraphic distribution of the genus Discosiphonella

Discosiphonella is an abundant sponge known from different Carboniferous and Permian localities in the USA, Europe, Japan and China (for more information see Senowbari-Daryan 1990; Rigby et al. 1998).

Discosiphonella is a rare sponge genus in Triassic deposits. It was described from Norian/Rhaetian reefs in Peru (Rauff 1938), the Caucasus (Belyaeva 1991: in Boiko et al.), the Alps (Zankl 1969), Greece (Senowbari-Daryan, unpublished material; see Senowbari-Daryan et al. 1996), and now from Turkey.

Discosiphonella is one of the numerous sponge genera which disappeared in the Late- Permian and reappeared in the Norian.

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Stratigraphic and facial correlation of the Szendrő–Uppony Paleozoic (NE Hungary) with the Carnic Alps–South Karawanken Mts and Graz Paleozoic (Southern Alps and Central Eastern Alps); some paleogeographic implications

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The authors present a stratigraphic and facial correlation of the Variscan sequences of the Szendrő–Uppony Paleozoic of the Gemer–Bükk units (NE Hungary) with those of the South Alpine Carnic Alps–South Karawanken Mts Paleozoic and of the Upper Austroalpine Graz Paleozoic. Following a review of lithostratigraphic units of each area, a correlation is made between the Szendrő–Uppony and the two Alpine structural units in terms of second order cycles within the Variscan megacycle. The close similarity between these Variscan sequences suggests their original proximity on the southern (Apulian) carbonate shelf of the Alpine–Dinaric arm of the Prototethys. From this position the Szendrő–Uppony Paleozoic units (as part of the Gemer–Bükk domain or "composite terrane") were displaced during Tertiary tectonic escape processes into the Pannonian region. The overlapping of Carnic Alps, resp. Graz Paleozoic facial characteristics in the Szendrő–Uppony Paleozoic indicate that the Periadriatic Lineament was not a paleogeographic boundary during the Paleozoic (and the Mesozoic).

Key words: Paleozoic, Austria, Hungary, lithostratigraphy, cycle stratigraphy, sedimentary facies, paleogeography

1. *Introduction*

Detailed conodont-biostratigraphic and sedimentological studies since the late sixties and mapping carried out contemporaneously have greatly refined the knowledge of Paleozoic stratigraphy in the Carnic Alps – South Karawanken Alps (for summary see Schönlaub 1979, 1980a, b, 1985; Kreutzer 1992) and in

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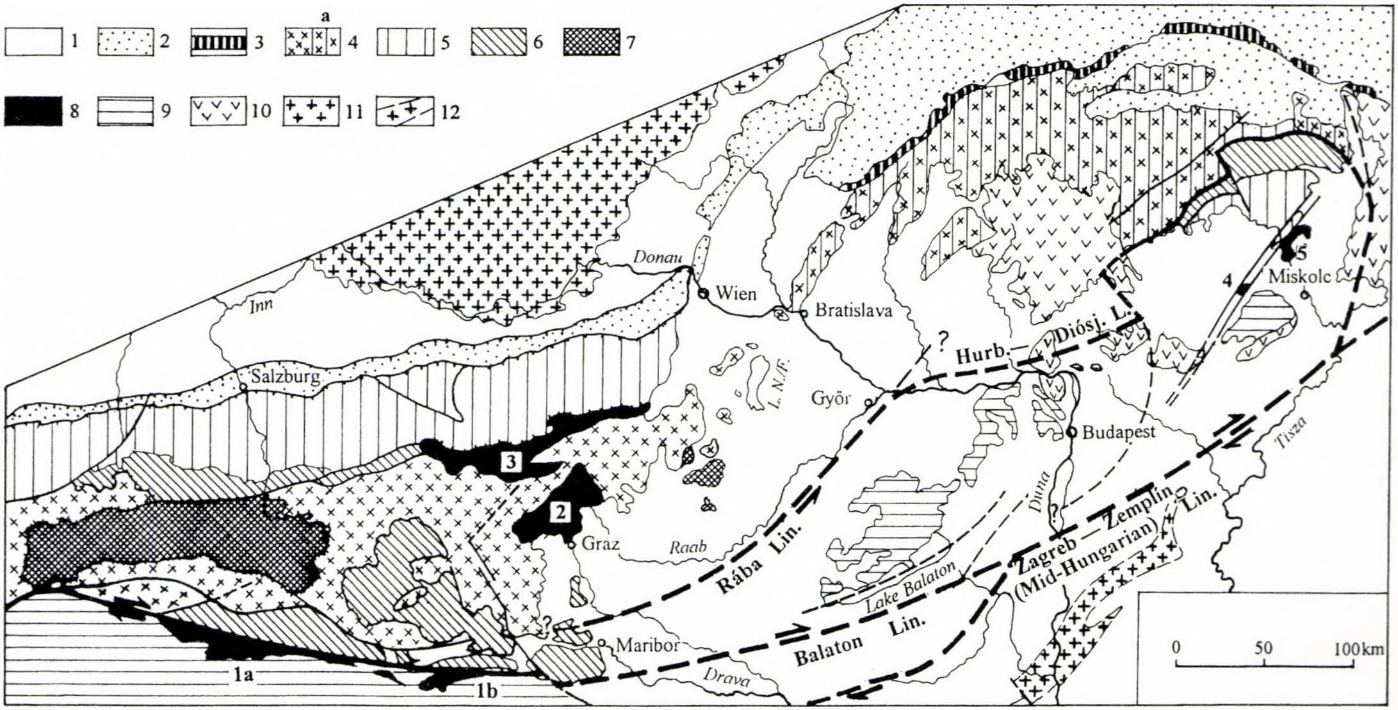
the surroundings of Graz (for summary see Ebner 1980; Ebner and Becker 1983; Flügel 1975; Flügel and Neubauer 1984b).

Similar studies were carried out mainly during the eighties in the alpidically-metamorphosed Paleozoic of the Szendrő and Uppony Hills in northeastern Hungary, along with mapping of the two regions (Kovács and Péró; Tertiary framework by Szentpétery). During these studies the close similarity to the Paleozoic of the above-mentioned two Alpine units became increasingly obvious (see also Flügel 1980; Ebner et al. 1991; Fülöp 1994). Beside the considerable similarity recognized between the formations of fossil-proven age, the comparison of the Variscan sequences of the two regions was also beneficial for establishing the stratigraphic position of those formations of the Szendrő–Uppony Paleozoic which lack determinable fossils. A preliminary formation chart can be found in the publication by Kovács and Péró (1983), whereas the latest reviews on the lithostratigraphic subdivision are presented in Kovács (1989b, 1992) and Fülöp (1994). A detailed, monographic description is planned for publication together with the geologic maps.

In addition to establishing the facies similarities, a joint evaluation of the Variscan sequences of the two regions contributed toward answering several important paleogeographic and major tectonic problems. Thus, following the recognition of the tectonic escape of the Pelso Megaunit (=composite terrane) (Kázmér and Kovács 1985)[1], correlation of the Variscan basement provides a significant contribution toward restoring the original position of the megaunit, above all to that of the Gemer–Bükk units constituting its "head". Recognition of the original proximal

Fig. 1 →
Tectonic/terrane sketch map of the East Alpine – North Pannonian – West Carpathian ("ALCAPA") and adjacent regions, showing the present setting of correlated units. Base maps: Beck-Mannagetta and Matura, in Oberhauser, 1980; Fuchs 1985; Plašienka 1991; Vozárová and Vozár 1992, 1996; Kovács et al. 1995. 1. Alpine molasse areas; 2. Alpine flysch units (incl. the Podhale flysch of the Inner West Carpathians); 3. Pieniny Klippen Belt; 4. Lower and Middle Austroalpine units; 4a. "Tatro-Veporic Composite Terrane" (according to Vozárová and Vozár 1992, 1996) or "Slovakia Terrane" (according to Plašienka 1991); 5. Mesozoic of the Northern Calcareous Alps (Upper Austroalpine) and of the Silicicum of the Gemer-Bükk domain; 6. Upper Austroalpine Paleozoic of the Northern Graywacke Zone (western part) and of the Gurktal Nappe System (incl. its Mesozoic cover series); Drauzug – Pohorje units; Gemer Paleozoic of the Gemer – Bükk domain; 7. Penninic units; 8. Paleozoic series involved in the present correlation: 1a: Carnic Alps, 1b: South Karawanken Mts, 2. Graz Paleozoic, 3. Northern Graywacke Zone (eastern part, Noric Nappe; the Veitsch Nappe is omitted), 4. Uppony Unit; 5. Szendrő Unit; 9. Southern Alps, Transdanubian Range, Bükk Mts; 10. Neogene volcanics; 11. Bohemian Massif; 12. Variscan anatectic granite zone of the Mecsek–Northern Great Plain Zone (according to Buda 1995). Lin. – lineament; Hurb.–Diósj. L. – Hurbanovo–Diósjenő Lineament; L.N./F. – Lake Neusiedl/Fertő. *Note*: 1. The Balaton and Buzsák lines are not distinguished, the track of dextral displacement is shown according to Balla et al. 1987. 2. The Kainach Gosau overlying the Graz Paleozoic is omitted

1 More recently this "escape" is understood within the context of that of the entire "North Pannonian–Inner West Carpathian Composite Terrane"; cf. Balla, 1988; Csontos et al. 1992; Neubauer et al. 1995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Graz
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(Mid-Hungarian)
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Balaton Lin.
Maribor
Drava
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paleogeographic position of the Variscan units compared with each other in this paper and that of the tectonic escape helped to understand the origin of certain "Dinaric" elements (above all the Graz Paleozoic) in the Upper Austroalpine Nappe System (c.f. Flügel 1980; Tollmann 1987; Ebner et al. 1991).

A paper by the present authors, which contains the correlation of the Carboniferous formations and a few paleogeographic consequences, has already been published (Ebner et al. 1991). In this contribution we give the correlation of the entire Variscan sequences of the two regions (ending with the flysch stage) and its most important paleogeographic implications. Its first draft was prepared in 1992 and presented at the ALCAPA Symposium in Graz (Ebner et al. 1992).

2. *Outline of the Variscan sequences*

We begin our outline with the better-known Carnic Alps–South Karawanken Alps and Graz Paleozoic sequences. A third Austroalpine unit, the Noric Nappe of Northern Graywacke zone showing a Lower Paleozoic similar in many respects to that of the Carnic Alps, is not dealt with in detail; however, at the end of the paper it is also involved in the correlation to a certain extent, as far as the sediments of the basal coarse clastic cycle are concerned.

2.1. Carnic Alps–South Karawanken Alps

The two mountain ranges along the southern side of the Gailtal Lineament are characterized by largely comparable Variscan sequences; therefore, they are usually discussed together in the literature. There are only local differences within their stratigraphic range; the formation charts (for detailed descriptions see Schönlaub 1979, 1980a, b, and 1985; Tollmann 1985; Kreutzer 1992b) are shown in Figs 2 and 3 (thickness data also given there).

Late Ordovician

The basal cycle of the Variscan megacycle is represented mainly by coarser grained sediments (quartzites, graywackes). Within the Carnic Alps four facies units can be distinguished:

Uggwa Facies, including sandy shale and quartzite, with limestone in its top part, rich in fossils represented mainly by bryozoans and brachiopods, as well as in the limestone by conodonts;

Himmelberg Facies, including graywacke, quartzite and sandy shale (Himmelberg Sandstone), overlain by limestone containing cystoids and conodonts (Wolayer Limestone);

Bischofalm Facies, including non-fossiliferous grey quartzite (Bischofalm Quartzite); this quartzite constitutes the basal part of this facies unit persistent throughout the Silurian–Devonian to the Tournaisian;

Fleons Facies: similar to the Uggwa Facies, but the bryozoan shale alternates with graywacke and quartzite containing acidic and basic volcanoclastics, partly even intercalations; sometimes conglomerate horizons do also occur. The rich fossil content is indicative of late Ordovician (Caradocian and Ashgillian).

In the Seeberg Paleozoic of the South Karawanken Alps the uppermost Ordovician shallow-water limestone overlies a 60 m thick acidic–intermediate pyroclastic sequence (Loeschke and Rolser 1971).

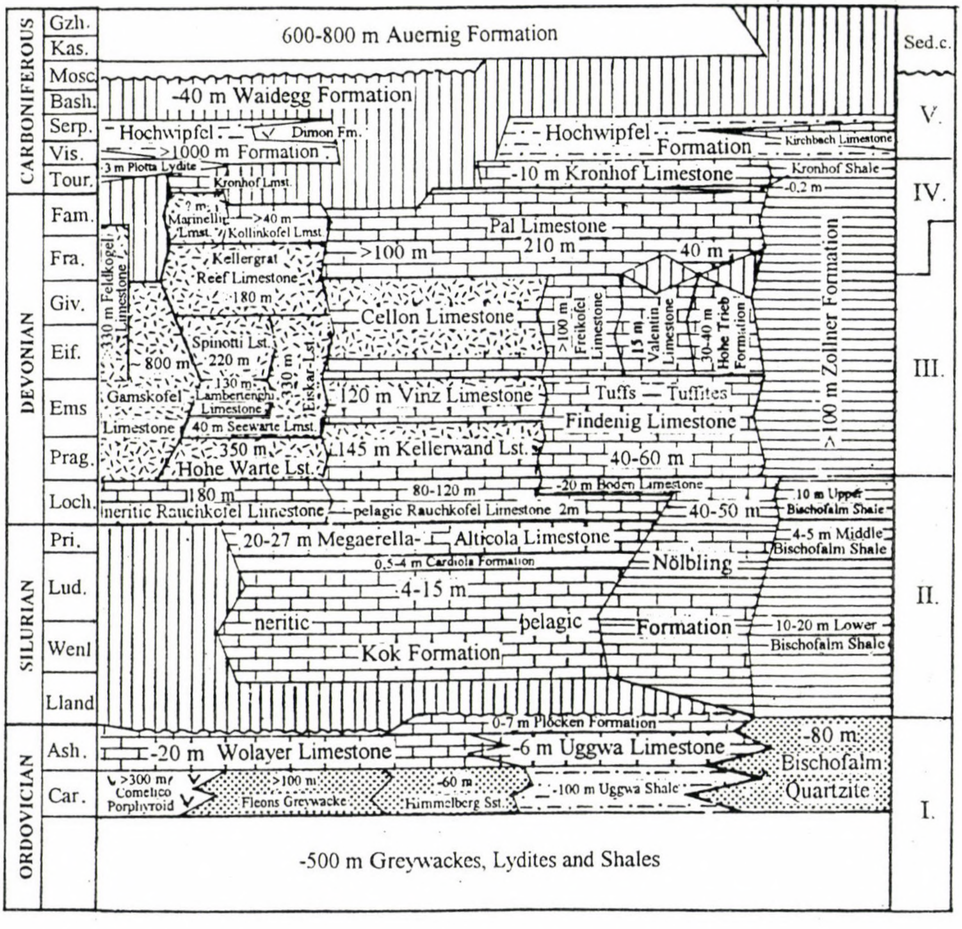


Fig. 2

Variscan formations of the Carnic Alps (after Schönlaub 1985 and Kreutzer 1992b). 1. Early Variscan coarse-grained siliciclastic rocks (greywackes, quartzites); 2. Variscan flysch; 3. euxinic deep-water, carbonate-free sediments; 4. pelagic carbonates; 5. platform carbonates; 6: volcanics. *Sedimentary cycles*: I. Late Ordovician coarse detrital cycle; II. Silurian–earliest Devonian (Lochkovian) pelagic carbonate cycle; III. Early Devonian–Late Devonian (Pragian–Frasnian) carbonate platform cycle; IV. Late Devonian–earliest Carboniferous (Frasnian–Tournaisian) pelagic carbonate cycle; V. late Early Carboniferous–Middle Carboniferous (Visean–Early Moscovian?) flysch cycle

CARBONIFEROUS	U.	STEFANIAN		Auernig Beds		
		WESTFALIAN		600 m Hochwipfel Flysch		
		NAMURIAN			~~ SUDETIC PHASE	
	L.	VISÉAN		?	Banded lmst. cu I-III χ	HIATUS
		TOURNAI- SIAN			~~ BRETON. PHASE	
DEVONIAN	U.	FAMENNIAN		?	Flaserl. to III-VI	
		FRASNIAN	20 m ?	Lydite, siliceous shale	to III	200 m Crinoidal lmst. "Crinoid breccia of seeland"
	M.	GIVETIAN	40 m ?	Lydite, shale	Red flaser lmst.	200 m Coral reef lmst.
		EIFELIAN		Siliceous shale,		
	L.	EMSIAN	30 m	"Flaser-Netz" lmst.		
			20 m	Shale, siliceous shale		
		SIEGENIAN		Dark platy lmst.	30 m Bronteus	
		GEDINNIAN			40 m Black platy lmst.	
SILURIAN	U.	PRIDOLIAN		Shale, Alticolalmst. (Flaserl.)	40 m Red nodular lmst. with orthoceratids	
		LUDLOWIAN		Cardiolalmst (Black l.) Koklmst. (Orthoceras lmst)	> 4 m Cardiola beds (lmst., shale) ?	
	M.	WENLOCKIAN		Shale, Lydite	? Orthoceras lmst.	
	L.	LLANDOV	10 m	Crinoidal lmst.	Sandy Series	
			10 m	Lower beds (lmst.-sst.)		
ORDOVICIAN	U.	ASHGILLIAN	6 m	Clayey-flaser lmst.	Flaserlimestone	
		CARADOCIAN			> 60 m Acidic to intermedier tuff.	
	M.			Shale	(BASIS NOT EXPOSED)	
	L.			?		

Fig. 3
Variscan formations of the South Karawanken Mts. (after Tollmann 1985)

Silurian

During the next phase of the Variscan sedimentary cycle – with the exception of the continuous Bischofalm Facies – pelagic carbonate sedimentation became predominant, with a transgressive character. The facies units formed in basin and swell environments produced by vertical movements are differentiated according to the time-range of hiatuses.

Plöcken Facies: Basin environment with sedimentation beginning at the base of the Silurian, but the graptolitic shale – limestone sequence of the Llandoverian Stage is dissected by hiatuses. The Wenlockian to Pridolian stages are represented by reddish or greyish, orthoceratic limestones rich in conodonts ("Kok-Kalk", "Alticola-Kalk").

Wolayer Facies: Ridge environment, with a hiatus up to the Wenlockian, in places even extending into the Ludlovian. The higher part of the Silurian is represented by a pelagic limestone sequence corresponding to that of the Plöcken Facies, but is more condensed.

Findenig Facies: Transitional development between the above two pelagic carbonate facies and the Bischofalm facies: alternation of black shales and dark limestones, with siliceous shale and lydite intercalations (Nölbling Formation).

Bischofalm Facies: Deep water euxinic facies: alternation of usually black-colored graptolitic shale, siliceous shale and lydite. It develops in transition from the Upper Ordovician coarse clastics.

Early–Middle Devonian

During the Lochkovian essentially the same facies pattern prevailed as in the Silurian, with continuation of the pelagic carbonate sedimentation (Rauchkofel-Kalk: dark-grey, platy limestone and grey crinoidal limestone). As carbonate platforms began to build up in the Pragian, three main sedimentary environments were differentiated, with transitions between them:

Carbonate platform facies: About 1100 m thick stromatoporoid and coral-bearing reef, reef-slope and back-reef lagoonal sequence, ranging in time from early Pragian to late Frasnian. The platforms were built up predominantly on the areas of the Silurian ridge-type Wolayer facies. Platform building ceased at latest at the end of the Frasnian due to eustatic sea-level rise.

Pelagic basinal facies (carbonate): Red or reddish, argillaceous micritic flaser limestone with orthoceratids (Findenig Limestone), then grey, banked or flaser limestone with goniatites (Valentin Limestone). Its sedimentary environment was formed mainly in the area of the Silurian Plöcken Facies.

Bischofalm Facies: The deep-water euxinic sedimentation, below the level of carbonate compensation, continued from the Silurian throughout the Devonian. It is still characterized until the end of the Lochkovian (until the extinction of graptolites) by black, graptolitic shales, and then a light grey-colored lydite – siliceous shale sequence follows named Zollner Formation.

Late Devonian–Earliest Carboniferous (Frasnian–Tournaisian)

Following a global trend, due to eustatic sea-level rise, carbonate platforms were drowned by the end of the Frasnian and only two sedimentary domains existed thereafter: the pelagic carbonate (with hiatuses of different time interval from place to place) and that of the continuing Bischofalm Facies.

Pelagic basinal facies (carbonate): In the former basinal areas it continued from the Middle Devonian, whereas on the former carbonate platforms it began at latest from the end of the Frasnian and lasted to the end of the Tournaisian. During the Late Devonian max. 40 m-thick varicolored flaser limestone (Pal-Kalk) was deposited, in places with ammonoids. It was followed in the Tournaisian by similar deposition (Kronhof-Kalk), but in strongly reduced thickness (usually 1–2 m, exceptionally 10 m). Continuous stratigraphic sequences have been preserved only in a few places: the intense vertical movements preceding the flysch stage led to local uplifts and karstic erosion.

For this reason a considerable part of the sequence is missing: at the Devonian–Carboniferous boundary a hiatus comprising various time-intervals can be recognized in most areas. It is not excluded, however, that in part of the hiatuses were caused by bottom-currents, which swept away the deposited lime mud. In extreme cases the entire Upper Devonian–Tournaisian pelagic limestone sequence is missing and the Hochwipfel Flysch lies immediately over the Frasnian carbonate platform or Frasnian pelagic limestone. The basinal carbonate deposition lasted until the topmost conodont zone of the Tournaisian, to the *Scaliognathus anchoralis* zone.

In the South Karawanken Mts. fissure fillings can frequently be found, which contain mixed conodont faunas representing nearly all zones of the Late Devonian - Early Carboniferous (Tessensohn 1974). These are explained by karstic erosion and resedimentation, similarly to the limestone breccias which also occur frequently. The continuous deposition is represented here by the "Bänderkalk", according to conodonts also up to the *Scaliognathus anchoralis* zone. On the other hand *Gnathodus bilineatus bilineatus* was found in the mixed faunas, resp. in the breccias as well, indicating the continuation of basinal carbonate sedimentation at least until the upper Visean.

Bischofalm Facies: Sedimentation below CCD continued unchanged until the end of the Tournaisian. The top part of the sequence is constituted by black lydites of a few meters thickness, with conodonts up to the *Scaliognathus anchoralis* zone (Herzog 1988). With this conodont zone quiet, deep-water euxinic sedimentation ended and flysch sedimentation began.

Flysch stage (Visean–Lower Moscovian)

After the *Scaliognathus anchoralis* zone a turbiditic, argillaceous–silty–sandy sequence was deposited (Hochwipfel Flysch), the thickness of which exceeds 1000 m in the Carnic Alps. In the lower part of the sequence limestone olistostromes, and in the upper part lydite olistostromes, resp. "parabreccias" (Spalletta et al. 1980; Spalletta 1982) are intercalated. As single slide-blocks limestone olistothrymmata also occur. Typical flysch sedimentary structures are rarer in the Carnic Alps, while they are characteristic for the South Karawanken Mts.; also, limestone olistothrymmata are larger there (Tessensohn 1971).

The Hochwipfel Flysch is divided, in parts of the Carnic Alps, into two parts by a limestone horizon (called "Kirchbach-Kalk") occurring at the Visean/Serpukhovian boundary interval. It is made up of grey, partly nodular, pelitic, micritic limestone with conodonts belonging to the *Paragnathodus nodosus* zone. The "Kirchbach-Kalk", resp. the limestone olistostromes contain Visean–Serpukhovian exotic limestone clasts of carbonate platform facies (with corals, calcareous algae, foraminifera; Flügel and Schönlaub 1990); this facies is nowhere known in situ in the Carnic Alps.

2.2. *Paleozoic of Graz*

The thrust complex of the Graz Paleozoic is formed by three major nappe systems which differ in sedimentary facies, stratigraphic range of sedimentary units and metamorphic overprint (Fritz et al. 1992).

Low grade to amphibolite facies metamorphics are represented in the lower nappe system. The sequence of this metamorphosed unit is best known from the Schöckel Nappe, the lithologies of which are summarized as Schöckel and Passail Groups.

The Laufnitzdorf Nappe and Kalkschiefer Nappe, defined by the sedimentary sequences of the Laufnitzdorf Group and the Kalkschiefer Group, respectively occur either in the basal parts of the Graz thrust complex or, due to two-step stacking, in an intermediate structural level.

The upper nappe system, the Rannach and Hochlantsch Nappe, comprises the Rannach and Hochlantsch Groups. These very low grade to low grade metamorphic units are similar in stratigraphic range (Silurian–Carboniferous), but their sedimentary/facial evolution is different.

Because of facial variability, tectonics and metamorphism, especially in the lower nappe units, biostratigraphic control of pre-Middle Devonian is poor.

This review is based mainly on the works of Ebner 1980b; Ebner et al. 1980; Ebner and Becker 1983; Flügel 1975; Flügel and Neubauer 1984b; and Fritz et al. 1992.

Rannach Group

Kehr Formation: The Lower Kehr Fm. (~100 m) is dominated by tuffs/tuffites with rare intercalations of diabases (mandelstein) and grey to brownish flaser limestones. The Upper Kehr Fm. (50–70 m) is composed of grey, sometimes tuffitic shales and colored flaser limestones. Age: The volcanism is dated by conodonts within the Late Silurian (Ludlovian/Pridolian; Neubauer 1991).

"Crinoidal Beds" (=Parmasegg Fm.; Fritz 1991): A laterally and vertically rapidly changing environment of dark shales, platy dark limestones, grey to ocher siltstones and sometimes sandstones, dolomites and crinoidal limestones.

Thickness: ~ 150 m Age: Pridolian–Late Siegenian/Early Emsian

"Dolomite–Sandstone Formation": This 500–1000 m thick sequence is characterized by a basal member of yellowish dolomitic quartz sandstones (50–100 m, barrier sands) followed by various types of dolomites (Dolomite Member). Locally these two members are separated by basic tuffites (~50 m, Diabase Member). The dolomites represent supratidal to intratidal facies. Dark dolomites belong to a restricted lagoonal environment (Fenninger and Holzer 1978; Ebner et al. 1980a). Age: Lower Devonian (Emsian).

Barrandei Limestone Formation: Dark bluish-grey, thick-bedded, highly fossiliferous (corals, stromatoporids, brachiopods) limestones with intercalations of low-energy siliciclastic shales. (Hubmann 1990). Age: Upper Emsian–Lower Givetian. Thickness: 80–100 m.

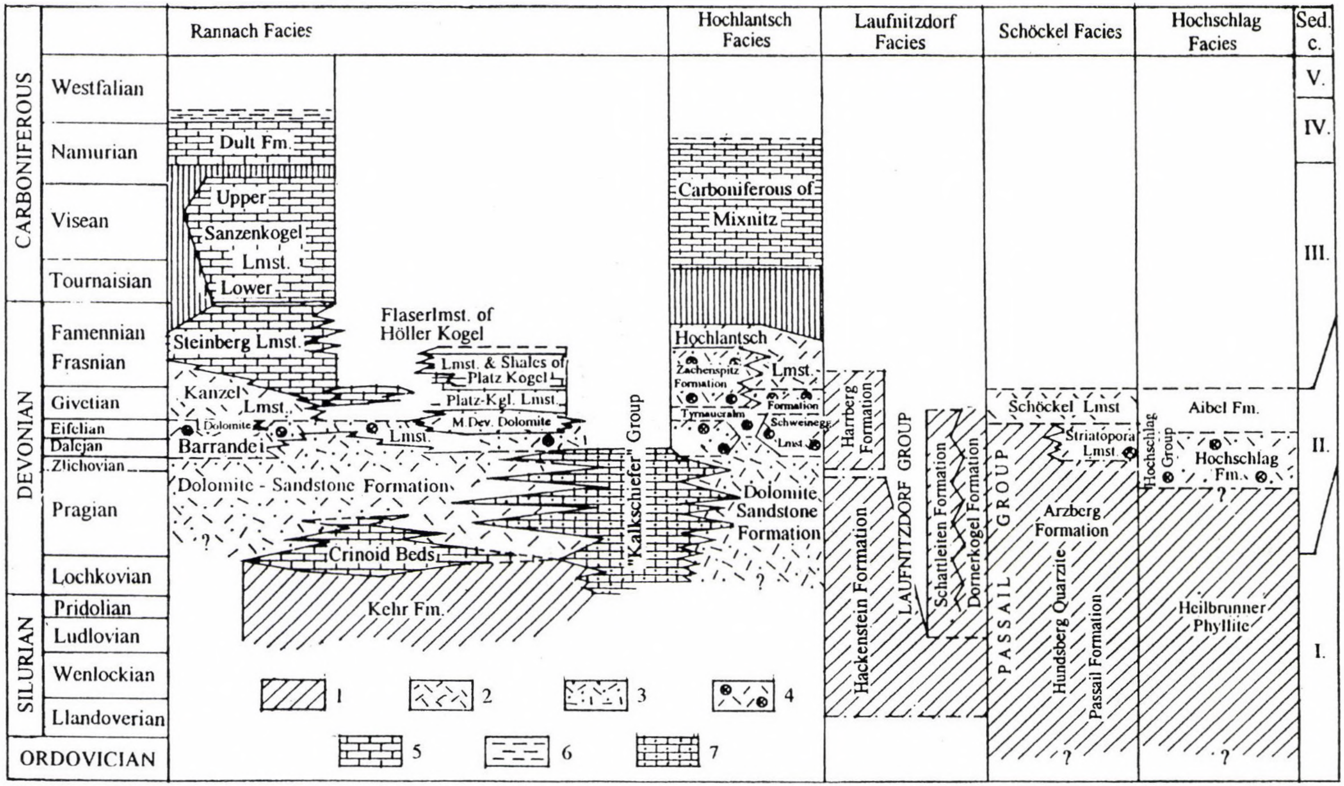


Fig. 4

Stratigraphic chart of the Graz Paleozoic (after Ebner and Becker 1983). 1. volcano-sedimentary sequence (predominantly phyllites and metavolcanics); 2. platform carbonates; 3. carbonate-siliciclastic shelf deposits; 4. the same as 3, with corals; 5. pelagic carbonates; 6. Carboniferous, mainly pelitic deposits (corresponding in age to the Variscan flysch stage, but not of turbiditic origin); 7. calcareous slate (= "Kalkschiefer") sequence. *Sedimentary* cycles: I. Late Ordovician(?)–Early Devonian volcano-sedimentary cycle; II. late Early Devonian–early Late Devonian (Pragian–Frasnian) carbonate platform cycle; III. Late Devonian–Early Carboniferous (Frasnian–Visean) pelagic carbonate cycle; IV. Middle Carboniferous (Namurian B–?early Westfalian or early Bashkirian) "flysch" cycle (not in flysch development, however)

Kanzel Limestone Formation: Light grey, thick-bedded to massive, sometimes coral-bearing limestones of sheltered lagoon facies. Thickness: ~100m. Age: Givetian, locally extending to the lowermost Frasnian. An equivalent to the Kanzel Limestone Formation is the Platzlkogel Limestone Formation which represents a more open and more energetic carbonate shelf environment (Ebner et al. 1980b).

Steinberg Limestone Formation: Reddish brown–yellowish, violet or grey, pelagic (cephalopod) limestones of flaser type. Age: Frasnian–Famennian, but sometimes it may begin in the uppermost part of the Givetian. Thickness: Usually between 20–80 m depending on a hiatus often cutting the upper part of the Steinberg Limestone in some sections.

Sanzenkogel Formation: This is the continuation of the pelagic sequence of the Steinberg Limestone. The only lithological difference is that it contains lydite, respectively shale intercalations in a max. of a few meters thickness. Age: Tournaisian–Namurian A. In some sections the Devonian/Carboniferous boundary is covered by a stratigraphic gap up to Latest Tournaisian/Early Visean caused by karstification and non-sedimentation. Thickness: Depending on the gap at its base as well as its hanging wall, from a few meters to max. 30 m. Where sedimentation of the Lower Sanzenkogel Formation (Tournaisian) occurred, this time span is only represented by 2–3 m of micritic limestones. In these areas the boundary between Lower and Upper Sanzenkogel Formation is marked by a band of shales with phosphoritic nodules.

Dult Formation: Dark grey, thick-bedded limestones (10–15 m), followed by black shales (50–70 m) containing limestone and calcschist intercalations in its lower part. Age: Namurian B–Namurian C/? Westfalian A.

The Dult Formation overlies the footwall rocks with an erosional unconformity covering the *Homoceras* Stage of the uppermost Namurian A. In one locality the entire Sanzenkogel Formation and parts of the Steinberg Limestone Formation were eroded before deposition of the Dult Formation. Features of flysch sedimentation are missing within the Dult Formation. However, recently some evidence of reworked limestones and graded allodapic limestones were found within the Dult Formation, and several erosional horizons were recognized as well: one at the base (within or above the *Gnathodus bilineatus bollandensis* zone) and another one between the Dult limestone and the Dult shales above the *Idiognathoides noduliferus* zone (Ebner 1977a, b, 1978).

Haigger Group

To the northwest the Lower Devonian of the Rannach Group interfingers with the Haigger Group. This sequence is composed of an alternation of variously colored platy and massive limestone with (silty) shales. This environment represents a more open marine equivalent of the Crinoidal Beds and Dolomite Sandstone Formation on a shelf sporadically influenced by a clastic input. Age: (based on conodonts) Lower Devonian (Gedinnian–Emsian; Buchroithner 1978).

Hochlantsch Group

The base of the Hochlansch Group is formed by the Dolomite Sandstone Formation in similar lithologies as in the Rannach Group. Facies differentiation of both units occurred in the hanging wall sequence starting with the Tyrnaueralm Formation in the Early Givetian (Gollner and Zier 1985).

Tyrnaueralm Formation (="Calceola Schichten" auct.): This consists of dolomites, sandstones, rauhwackes, basic volcanics and limestones. In its lower part it is dominated by lithologies similar to the "Dolomite–Sandstone Formation". It is followed by limestones locally rich in coral and stromatoporid bioherms. Up to a few tens of meter-thick volcanics (mainly pyroclastics, mandelstones and spilites) are intercalated within the dolomite as well as the limestone-dominated parts. The facies is interpreted as a littoral environment in the lower and as a sublittoral higher-energy water facies in its higher parts. Thickness: 140–500 m. Age: (?Early) Givetian.

Zachenspitz Formation (="Quadrigeminum-Kalk" auct.): Alternation of thin-bedded, grey to light-brown micritic limestones, massive to thick bedded, light-grey limestones and fossil-rich dark micritic limestones. Coral-stromatopora-bioherms are also found. This environment is interpreted as a shallow subtidal deposit with variable energy conditions and a high rate of sedimentation. Within the latest Givetian and the Frasnian the environment was changing to an open marine one. Tuffs, lapilli-tuffs and tuffites in a thickness of a few meters are bound to the latest Givetian. Age: Late Givetian–Frasnian. Thickness: 400 m.

Hochlantsch Limestone Formation: A massive and, in the middle part, sometimes bedded, light grey-reddish limestone of 800 m thickness. Fossils are rare (only a few corals and stromatoporids). 600 m of the overall thickness belongs to the Upper Devonian. The environment is interpreted as a low subtidal, open marine platform with low-energy conditions. The top of the formation is formed by a pre-Upper Tournaisian erosional surface overlain by a few dm-thick limestone breccia with mixed conodont faunas of Early Upper Devonian and Upper Tournaisian (Zier 1981; Gollner and Zier 1985). Age: Late Givetian–Famennian (*Palmatolepis marginifera* zone). Thickness: 800 m.

Carboniferous of Mixnitz. The base of the Carboniferous of Mixnitz is represented by a breccia of reworked Hochlantsch Limestone in a red calcareous matrix with mixed conodont faunas (Zier 1981; 1983). The breccia is overlain by massive to poorly-bedded reddish-brown micritic, pelagic limestones (with conodonts, cephalopods, and radiolarians). Lydite intercalations and chert nodules are common. The gap at the base was formed by a pre-Upper Tournaisian and post-*Palmatolepis marginifera* zone uplift and erosion. The formation of mixed faunas occurred during the Upper Tournaisian marine transgression. Age: Upper Tournaisian–Namurian B Thickness: 90–100 m.

Schöckel and Passail Groups

The integration of the below-mentioned formations into one tectono-stratigraphic unit is rather problematic. According to the interpretation of Flügel and Neubauer (1984b), the Passail Formation represents a distinct stratigraphic as well as tectonic unit.

Passail Phyllite: Volcano-sedimentary sequence, consisting of alternation of tuffs, tuffites, small bodies of diabase and dark grey phyllite. Sericite phyllites are locally connected with sericite quartzites (Hundsberg Quartzite). Age: Early Paleozoic (? Ordovician–Silurian). Thickness: A few 100 m.

Arzberg Formation: Black slate, carbonate phyllite, carbonate graphitic slate, banded and dark limestones. Stratiform Pb/Zn and barite mineralizations are bound to this formation. Dark fossiliferous coral-bearing limestones (Striatopora Limestone) occur in close relation to the Schöckel Limestone in the hanging wall. Another local intercalation at the boundary to the Schöckel Limestone is the Lammkogel Quartzite. Age (based on conodonts): Upper Silurian–Lower Devonian. Thickness: 200–300 m.

Schöckel Limestone: Bluish-grey and white banded marble. Locally the basal Schöckel Limestone is represented by light yellow dolomites, quartzites, rauhwackes, shales and calcschists (Raasberg Formation). Age: According to its position and sporadic fossil finds Lower Devonian–? Middle Devonian. Thickness: several 100 m.

Laufnitzdorf Group

Hackenstein Formation: A volcano-sedimentary sequence consisting of an alternation of basic volcanics, dolomite and limestone (tuffaceous crinoidal limestone, nodular limestone). The magnesite of Breitenau is also intercalated into this sequence. Age: the limestone intercalations contain conodonts of Lower Silurian–higher Lower Devonian. Thickness: 150–200 m.

Schattleiten Formation: At the base a six m-thick dark bluish-grey, thick-bedded limestone, then alternation of sericitic sandstone, slaty siltstone and slate, finally light-grey to bluish-grey stilolitic limestone, resp. marble. Sometimes metatuff and magnesite intercalations also occur. Age: the limestones contain Upper Silurian–higher Givetian conodonts. Thickness: > 100 m.

Dornerkogel Formation: In its lower part greenish sandstone, dark grey shale and conglomerate composed of volcanites alternate. The upper part is constituted by dark bluish-grey and greenish-grey, thick-bedded limestone. Age: unclear because of its tectonic boundary and lack of fossils. However, the abundance of phytoclasts in the clastic rocks may reflect a Carboniferous age. Thickness: > 100 m.

Harrberger Formation: Grey to black shale sequence, with intercalations of grey pelagic limestone, dolomite, greenish-grey sandstone, lydite and metatuff. Age: the limestone intercalations contain Emsian–Frasnian conodonts. Thickness:

70 m. The contact with other units is not clear. Probably the boundary to the underlying Hackstein Fm. is a tectonic one.

Kalkschiefer Group:

Alternation of dark grey, thin to well-bedded micritic and biodetrital limestones with (silt) sandstones and calcareous shales. Age: Biostratigraphic evidence is poor. The stratigraphic range is at least Lower Devonian (Zlichovian)–Middle Devonian (Givetian). Thickness: A few 100 m.

2.3. *Szendrő Mts.*

The Paleozoic of the Szendrő Mts. (or Szendrő Unit), with a general NNW-vergent structure, is made up of two subunits, the northerly Rakaca Subunit and the southerly Abod Subunit. The latter thrusts over the former from the SSE. Mesozoic formations are unknown. The Paleozoic rocks underwent a Cretaceous greenschist facies metamorphism; Variscan metamorphism cannot be proven (Árkai 1983; Árkai et al. 1995). The present review is based mostly on Kovács 1989b, 1992 as well as on Fülöp 1994.

Abod Subunit

Irota Formation (="Tapolcsány Formation" auct.; Kovács 1989b). Alternation of black, graphitic phyllite, black siliceous slate and in its lower part grey metasandstone. In the middle part white, calcareous phyllite (Árkai, unpubl. report) intercalations also occur. In the top part (as exposed by the borehole Felsővadász-1 which is the type section) corals occur in marly–silty rocks marking the transition towards the Szendrőlád Limestone Formation. It is a euxinic basinal facies with high pyrite content. Age: based on its stratigraphic position Silurian (?)–Lower Devonian. Thickness: min. 300–400 m.

Szendrőlád Limestone Formation: Predominantly bluish-grey limestone, alternating with a lesser amount of light grey phyllite and metasandstone. Four subfacies can be distinguished within it: biohermal limestone, basinal limestone, sandy limestone, and phyllite–metasandstone. Corals (Mihály 1978) and in places crinoids are characteristic for the formation. Age (based on corals and conodonts): Eifelian–Lower Frasnian. Thickness: several hundred (min. 400) m.

Abod Limestone Formation: Brownish yellow, finer crystalline, or white, coarser crystalline, sericitic–chloritic, metatuffitic limestone, with thin (max. a few dm thick) intercalations of green metatuff. The sericite–chlorite content usually appears in a characteristic network due to two-phase schistosity ("cipollino"). The facies is basinal, with traces of contemporaneous volcanic activity. It develops gradually, with bed-by-bed alternation, from the Szendrőlád Limestone. Age (based on conodonts): Frasnian–Famennian. Thickness: min. 200 m.

Bükkhegy Marble Formation: Brownish yellow–yellowish white or brownish white, sometimes pinkish, massive or thick-bedded marble of carbonate platform facies. It overlies the Szendrőlád Limestone Formation. Age: based on its stratigraphic position and Alpine analogies (drowning of platforms): Frasnian. Thickness: min. 200 m.

Rakaca Subunit

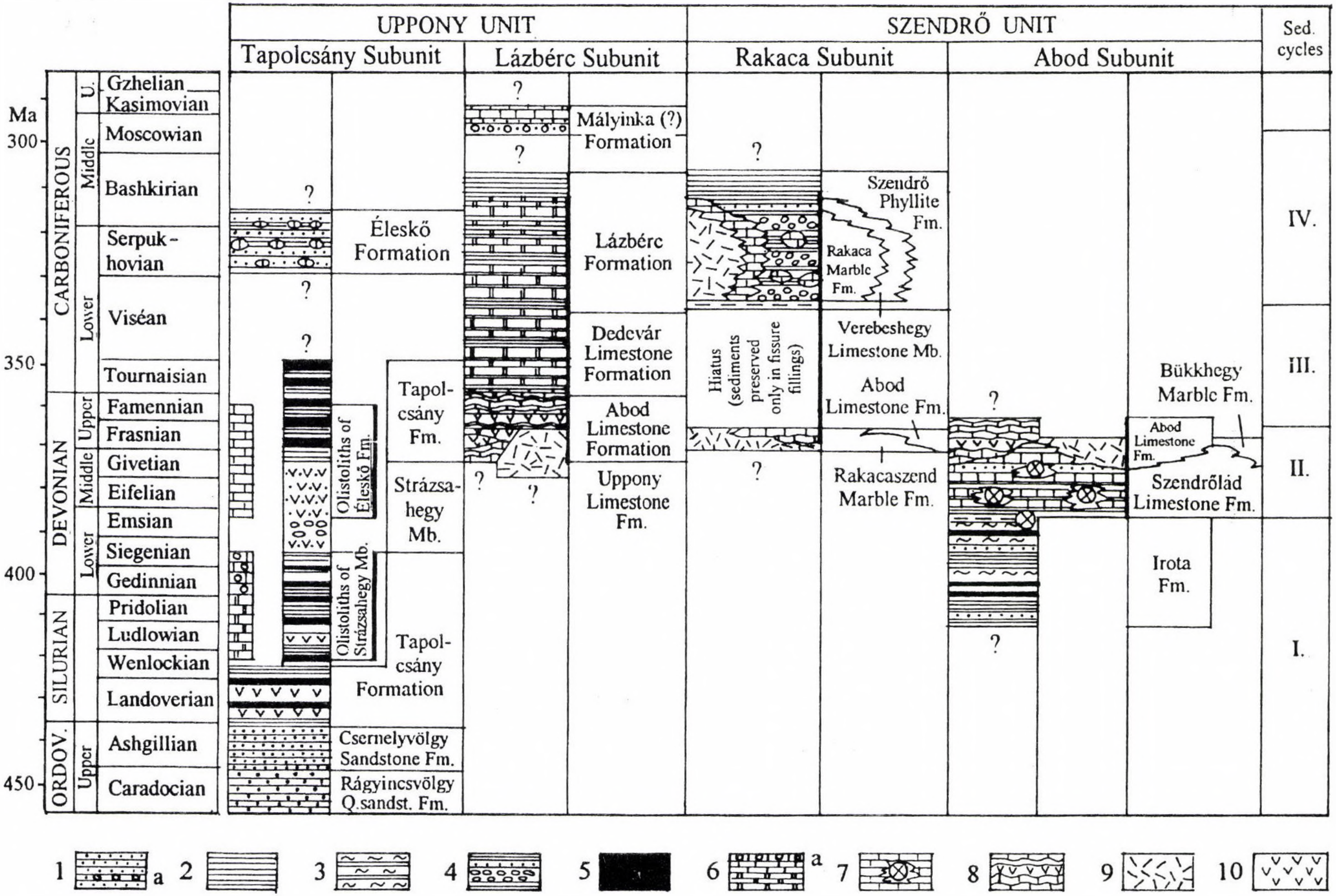
Rakacaszend Marble Formation: Bluish-grey, white-banded, thick-bedded, coarse crystalline marble of carbonate platform facies. Age (based on its relationship to the pelagic limestones described below): Frasnian. Thickness: about 200 m.

Pelagic limestones (*Abod Limestone* s.l.) in connection with the *Rakacaszend Marble*. These rocks may occur in two situations: 1) in continuous successions, and 2) as fissure infillings.

1) In continuous development: in the northernmost marble range the Rakacaszend Marble is overlain in a max. few tens of meters thickness by yellowish-brown to brownish-yellow, sometimes bluish or brownish-grey, sericitic, finer crystalline marble, which is also chloritic in some places (typical Abod Limestone).

2) As fissure fillings: in the middle (the main) marble range often yellowish-brown–purplish red, finer crystalline, sericitic marble fissure fillings occur, containing Upper Frasnian to Upper Visean mixed conodont faunas (from the *Palmatolepis gigas* zone to the *Paragnathodus nodosus* zone). In this range the Rakacaszend Marble is overlain either by Upper Visean pelagic limestone (only with 1–2 m thickness) or immediately by the Szendrő Phyllite representing the flysch stage. The underlier of the Rakacaszend Marble along the northwestern foot of Nagy-Somos Hill is constituted by Lower Frasnian bluish-grey, crinoidal limestone or brown pelagic limestone (*Polygnathus asymmetricus* zone), over which the platform was prograded.

Rakaca Marble Formation: In the southernmost marble range, separated from the middle range by the zone of the so-called "median slate" (forming the mostly tectonically sheared overlier of the latter) again bluish-grey, white-banded, coarse crystalline marble occurs, lithologically identical to the Rakacaszend Marble. It interfingers here, however, with a basinal facies. The platform/basin transition is represented by a marbleized sedimentary breccia (called locally as "foamy marble"). South of Rakacaszend the lower part of the marble alternates with brownish-grey crinoidal limestone (the entire unit is called "*Kopaszhegy Limestone Member*" in Fülöp 1994), which contains Lower Visean (*Gnathodus texanus* zone) conodonts. The alternation of the two types of carbonates bears witness to a prograding platform slope. In other sections the basinal facies both underlies and overlies the typical marble, or even interfingers with it. This basinal facies (called *Verebeshegy Limestone Member*) is constituted of dark bluish-grey, sometimes reddish-brown, thin or thick-banked, finer crystalline marble. Fissure fillings corresponding to the Upper Devonian Abod Limestone are missing in this range. Age: the Verebeshegy Limestone ranges in age, according to conodonts, from the Upper Visean (*Paragnathodus nodosus* zone)



UPPONY UNIT
SZENDRŐ UNIT
Sed. cycles
Tapolcsány Subunit
Lázbérc Subunit
Rakaca Subunit
Abod Subunit
Ma
300
350
400
450
CARBONIFEROUS
DEVONIAN
SILURIAN
ORDOV.
U.
Middle
Lower
Upper
Middle
Lower
Upper
Gzhelian
Kasimovian
Moscowian
Bashkirian
Serpuk-hovian
Viséan
Tournaisian
Famennian
Frasnian
Givetian
Eifelian
Emsian
Siegenian
Gedinnian
Pridolian
Ludlowian
Wenlockian
Landoverian
Ashgillian
Caradocian
Éleskő Formation
Tapol-csány Fm.
Olistoliths of Éleskő Fm.
Strázsa-hegy Mb.
Olistoliths of Strázsahegy Mb.
Tapol-csány Formation
Csernelyvölgy Sandstone Fm.
Rágyincsvölgy Q.sandst. Fm.
Mályinka (?) Formation
Lázbérc Formation
Dedevár Limestone Formation
Abod Limestone Formation
Uppony Limestone Fm.
Szendrő Phyllite Fm.
Rakaca Marble Fm.
Hiatus (sediments preserved only in fissure fillings)
Verebeshegy Limestone Mb.
Abod Limestone Fm.
Rakacaszend Marble Fm.
Bükkhegy Marble Fm.
Abod Limestone Fm.
Szendrőlád Limestone Fm.
Irota Fm.
IV.
III.
II.
I.
1 a
2
3
4
5
6 a
7
8
9
10

to Lower Bashkirian (*Idiognathoides sinuatus* zone). This age can be extrapolated to the interfingering "foamy" marble and banded marble as well. The Lower Visean Kopaszhegy Limestone Member may represent the deeper part of the Rakaca Marble Formation. Thickness: altogether not more than 200 m.

Szendrő Phyllite Formation: Flysch-type, dark grey to black phyllite and grey metasandstone sequence, lying over the Rakaca Marble Formation, or interfingering with it. According to the results of the mapping by Péró (unpublished as yet) it can be divided into a lower part of proximal character and into a higher part of distal character. In the lower, coarse detritic part (Meszes Member) limestone olistostromes and sandstone turbidites are frequent. The clasts of the olistostromes derive mostly from the Verebeshegy Member (as proven by the same conodonts) and in a lesser amount from the Rakaca Marble. In addition, according to conodont evidence, Middle Devonian and Tournaisian–Lower Visean limestones of basinal facies are also admixed. Single limestone olistothrymmata also occur. The higher part of the sequence is constituted by an almost monotonous phyllite sequence, with distal turbidites. The so-called "median slate" overlies the Rakacaszend Marble (see above). Age: the lower part is contemporaneous with the Rakaca Marble; the upper part can be Upper Bashkirian, or even Lower Moscovian in age, without more precise determination due to lack of useable fossils. Thickness: appr. 500–600 m (Péró, pers. comm.).

2.4. *Uppony Mountains*

The Paleozoic of the Uppony Mts. (=Uppony Unit) is enclosed within the Darnó fault zone, in a transpressional regime of Early Miocene age, which forms a kink-like bend here. It consists of two subunits: the southerly Tapolcsány Subunit made up of predominantly pelitic, psammitic and siliceous rocks, and the northerly Lázbérc Subunit consisting of predominantly of carbonate rocks and showing a NNW-vergent structure. The contact between the two subunits is also of strike-slip character, formed by an element of the fault zone. Mesozoic rocks are unknown except for Upper Cretaceous conglomerates forming the posttectonic cover of the Tapolcsány Subunit. The Paleozoic formations were affected by Alpidic metamorphism, the intensity of

← Fig. 5

Stratigraphic charts of the Paleozoic of Szendrő and Uppony units (modified after Kovács 1992). 1. metasandstone, 1a. pebbly sandstone, microconglomerate; 2. slate, phyllite; 3. marly slate, calcareous phyllite; 4. olistostrome horizons; 5. siliceous slate, lydite; 6. limestone of basinal facies; 6a. crinoidal limestone; 7. corals; 8. metatuffitic limestone ("cippolino"); 9. platform carbonate; 10. metavolcanics. Thick vertical lines between the lithological columns and formation names indicate biostratigraphically proven ages. *Sedimentary cycles*: I. Late Ordovician–Early Devonian volcano (?)-sedimentary cycle; II. Middle Devonian–early Late Devonian (Eifelian–Frasnian) carbonate platform cycle; III. Late Devonian–Early Carboniferous (Frasnian–Early Visean) pelagic carbonate cycle; IV. latest Early Carboniferous–Middle Carboniferous (late Visean–?early Moscovian) flysch cycle which was near to the boundary between the anchizone and epizone; Variscan metamorphism, however, cannot be proven (Árkai 1983 and Árkai et al. 1995). The present review is based mostly on Kovács (1989b, 1992) as well as on Fülöp (1994).

Tapolcsány Subunit

The subunit is made up of deep-water, euxinic formations similar to the Bischofalm Facies of the Carnic Alps, but determinable fossils are known only from olistoliths of olistostrome horizons.

Rágyincsvölgy Quartz Sandstone Formation: Light grey, massive quartz sandstone (protoquartzite), subordinately silty slate, with postkinematic chloritoid (Árkai et al. 1981; Ivancsics and Kisházi 1983). Age: based on the East and South Alpine analogies (Carnic Alps: Bischofalm Quartzite; Graywacke Zone: Polster Quartzite; and perhaps also the Hundsberg Quartzite in the Graz Paleozoic) it is considered to represent the first phase of the Variscan sedimentary cycle (Upper Ordovician). Thickness: 50–100 m.

Csernelyvölgy Sandstone Formation: Grey, matrix-rich, massive metasandstone, often with biotites and phyllite clasts. According to its texture and mineral composition it is considered as feldspathic graywacke (Ivancsics and Kisházi 1983). Age: based on similar arguments as in the case of the Rágyincsvölgy Quartz Sandstone, it is also assigned to the Upper Ordovician. Thickness: around 100 m.

Tapolcsány Formation: Alternation of dark grey, carbonate-free slate, black and light grey siliceous slate and black, radiolarian lydite, containing graphitic slate intercalations, in one horizon with basic, mostly of schalstein-type volcanics, in part associated with limestone olistoliths. It was deposited below the CCD and is characterized by a high pyrite content, indicating euxinic conditions. Iron–manganese mineralization is also characteristic for the formation. Age: based on the analogy to the Bischofalm Facies it is also assigned to the Silurian–Lower Carboniferous (Tournaisian) in general, without a more precise determination due to lack of diagnostic fossils. Thickness: min. 300–400 m.

Strázsahegy Member: Basic metavolcanic sequence, consisting mostly of schalstein, subordinately of basalt lava. Formerly it was considered as a distinct formation (Kovács and Péró 1983; Kovács 1989b); however, mapping has revealed that it forms one zone and occurs within the Tapolcsány Formation, as confirmed by some data from artificial outcrops and boreholes. Within the member an olistostrome horizon occurs with altered metabasalt (partly amygdaloidal) lava and subordinately tuff matrix and limestone olistoliths. The olistoliths belong to two major types: 1. Silurian (Wenlockian–Ludlovian) purplish red or greenish, micritic, pelagic limestone, rarely with orthoceratids and 2. lowermost Devonian (Lochkovian) light grey crinoidal limestone, rarely with coral fragments. As single olistothrymmata (outside the Strázsa Hill type locality) white, coarse crystalline marble of carbonate platform facies also occurs. Iron metasomatism is frequent throughout the sequence. Age: younger than the olistoliths: higher Lower Devonian or Middle Devonian. Thickness: min. 50 m, max. 100 m.

Éleskő Formation: Olistostrome(s) with grey calcareous slate, siltstone or marl matrix and with bluish grey pelagic, tentaculitic limestone olistoliths. The latter yielded Emsian–Early Famennian conodonts. Age: based on sedimentological criteria, this olistostrome is assigned to the late Lower Carboniferous or Middle Carboniferous flysch stage. Thickness: several 10 m.

Lázbérc Subunit[1]

Uppony Limestone Formation: Light grey, sometimes bluish grey, thick-bedded crystalline limestone of carbonate platform facies. Age: Middle Devonian(?)–Frasnian. Thickness: min. 200.

Abod Limestone Formation: Typically bluish-grey to purplish-grey or white and brownish-yellow, metatuffitic limestone with a characteristic sericite–chlorite network ("cippolino"). Bluish-grey, thick-bedded, non-tuffitic (but sometimes cherty) and brown, flasered types are also present. It contains light green metatuff and tuffitic calcschist intercalations and in the western part of the mountains it interfingers with basic metavolcanics (mostly schalstein, subordinately lava). Intercalations of lydite and black slate may also occur. Metasomatic ankeritization–dolomitization is frequent. Age: Frasnian–Famennian, the grey cherty limestone (only one outcrop) is Upper Givetian (*Polygnathus varcus* zone). For the age of the schalstein and lava so far only Frasnian data exist (cf. also Kovács and Vető-Ákos 1983), but thin green chloritic intercalations of volcanic origin can be found up to the *Bispathodus costatus* zone. Thickness: min. 100–200 m.

Dedevár Limestone Formation: Typically grey (often with a purplish or pinkish shade), thick-bedded, within the beds platy–flasered limestone, without metatuffitic traces. Bluish-grey, thick-bedded, non-flasered type is also present. In the Lower Visean a brownish grey, carbonatic lydite horizon occurs in 1–2 m thickness. Age: Tournaisian–Lower Visean. Thickness: a few meters to max. 10–20 m.

Lázbérc Formation: Alternation of bluish-grey, bedded to thin-banked limestone, sometimes platy limestone and calcschist with grey to dark grey, slightly calcareous slate, without resedimentation features. In one place thin metasandstone intercalations, resp. slaty clayey marl also occur. Age: the limestones contain conodonts from the Upper Visean to the Lower Bashkirian (from the *Paragnathodus nodosus* zone to the *Idiognathoides sinuatus* zone), whereas the shales may be partly of younger age as well. Thickness: min. 200–300 m.

1 To avoid duplication of names (e.g. Uppony Unit and Uppony Subunit, respectively, as was used in Kovács 1989b and based thereupon also in Ebner et al. 1991 and Fülöp 1994) the term "Lázbérc Subunit" instead of the former "Uppony Subunit" was proposed in Kovács 1992

Mályinka(?) Formation (=Derennek Member of Lázbérc Formation in Fülöp 1994): Post-tectonic, marine molasse sequence occurring in one zone within the folded–imbricated structure of the Lázbérc Formation. It consists mainly of light grey, coarse-grained (sometimes pebbly), calcareous sandstone, bluish-grey sandy limestone and subordinately of grey sandstone, microconglomerate and bluish grey crinoidal limestone. Well-rounded white quartz and black lydite pebbles of max. 1–2 cm diameter, occurring in the pebbly sandstone and microconglomerate, indicate uplift and erosion somewhere in the hinterland. Age: probably Upper Carboniferous and an equivalent of part of the Auernig-type Mályinka Formation of the adjacent Bükk Mts. Thickness: several 10 m.

3. *Cycle stratigraphy*

3.1. *Carnic Alps–South Karawanken Mts.*

Within the biostratigraphically best-known Paleozoic sequence the of the Carnic Alps, the following second-order transgression–regression cycles can be distinguished within the Variscan megacycle:

3.1.1. *Late Ordovician (Caradocian–Ashgillian) siliciclastic cycle*: It begins with transgression in the early Caradocian and ends with a regression at the Ordovician/Silurian boundary. This cycle can be recognized in parts of the Eastern Alps (Upper Austroalpine) and Southern Alps. Its characteristic formations are shallow marine siliciclastic sediments (quartzites, graywackes) with a coeval porphyroid volcanism. In the central Carnic Alps, however, only traces of the volcanism can be found, whereas beside the quartzites–graywackes fossiliferous sandy shales are also widespread and limestones occur in the uppermost part of the cycle.

3.1.2. *Silurian–earliest Devonian (Lochkovian) pelagic carbonate cycle*: Its lower boundary is marked by a new transgression in the lower part of the Silurian, and the upper one by a regression at the Lochkovian/Pragian boundary and by the appearance of carbonate platforms. Characteristic sediments are pelagic limestones with nautiloids, conodonts, and other fossils. Because of differential subsidence the sequence can be (nearly) complete (Plöcken Facies) or a considerable part of the Silurian can be missing (Wolayer Facies).

3.1.3. *Late Early Devonian–Early Late Devonian (Pragian–Frasnian) carbonate platform cycle*: Its lower boundary is marked by the beginning of carbonate platform building, and the upper one by the gradual drowning of the platforms, at latest by the end of the Frasnian (this latter was a worldwide phenomenon due to significant eustatic sea-level rise; Krebs 1974). Coevally with the platform building, in basins pelagic carbonate sedimentation continued. Likewise in the deep basins the euxinic sedimentation of the Bischofalm Facies continued from the beginning of the Silurian to the beginning of the flysch stage. Obviously, in both types of basinal sequences the third and lower order cycles connected to sea-level changes cannot be recognized.

3.1.4. Late Devonian–earliest Carboniferous (Frasnian–Tournaisian) pelagic carbonate cycle: The lower boundary is marked by the gradual drowning of the carbonate platform due to eustatic sea-level rise in course of the Frasnian, and the upper one by the beginning of flysch sedimentation (in terms of sequence stratigraphy) at the Tournaisian/Visean boundary. Due to a regression at the Devonian/Carboniferous boundary interval and the new transgression, the sequence of the cycle is interrupted by various hiatuses: in the two extreme cases the sequence is either wholly complete or is totally missing and the flysch lies directly upon the Frasnian platform limestones.

3.1.5. Visean–Early Moscovian/Early Westfalian flysch cycle: Siliciclastic turbiditic sedimentation began at the Tournaisian/Visean boundary (in the overlier of the preflysch sediments ending with the *Scaliognathus anchoralis* zone) and ended with the tectogenesis taking part in the middle part of the Moscovian (or Westfalian) stage (Carnic phase; Vai 1975).

3.2. Northern Graywacke Zone

The cycles recognizable in the Noric Nappe of the Northern Graywacke Zone largely correspond to those described above (cf. Schönlaub 1979). Significant differences can be seen only in the sequences of the first and fourth–fifth cycles. The quartzite–graywacke sequence of the Upper Ordovician siliciclastic cycle is much thicker here (even exceeding 1000 m) and the volcanism (first basic, then intermediate [porphyroid]) was considerably more intense. Further basic volcanism of less areal distribution can be recognized in the Silurian. Sediments of the fourth cycle have been preserved (with the exception of those of the Frasnian Stage) only as components of a limestone breccia of a few meter thickness. Finally, the flysch cycle itself is represented here by the monotonous black shale sequence of the Eisenerz Formation of non-flysch type.

3.3. Graz Paleozoic

The second-order cycles recognizable in the Graz Paleozoic correspond only partly to those of the Carnic Alps. The cycles can be characterized as follows:

3.3.1. Upper Ordovician(?)–Lower Devonian siliciclastic–volcanic cycle: Black phyllite–basic volcanic sequence, subordinately sandstone (Hundsberg Quartzite), and with Lower Silurian–Lower Devonian limestone intercalations. In the Rannach Facies it only extends as far as the Lochkovian, and in the Schöckel and Hochschlag Facies to the end of the Emsian (here the phyllite sequence replaces the "Dolomite–Sandstone Series").

3.3.2. Late Lower Devonian–early Upper Devonian (Pragian–Frasnian) carbonate platform cycle: The formations representing the cycle here differ from those of the Carnic Alps in the considerable terrigenous (siliciclastic) influx: it is especially significant in the Pragian–Emsian "Dolomite–Sandstone Formation", but can be observed in the Middle Devonian coral-bearing limestones as well. The siliciclastic influx indicates the proximity of a crystalline source area (Fenninger and Holzer 1978). Pure platform carbonates occur only near the end of the cycle (Givetian–Frasnian Hochlantsch Limestone, Schöckel Limestone, Kanzel Limestone).

3.3.3. *Upper Devonian–Lower Carboniferous (Frasnian–Namurian A) pelagic carbonate cycle*: The sedimentation of this cycle here is largely identical with that of the Carnic Alps, apart from the larger stratigraphic extent. However, the pelagic Sanzenkogel Limestone here certainly reaches as far as the Namurian A (*Gnathodus bilineatus bollandensis* zone). Above it the sequence is interrupted by hiatuses and rocks belonging unambiguously to the Dult Formation occur only from the Namurian B (*Idiognathoides noduliferus* zone; Ebner 1976, 1977a, b).

3.3.4. *"Flysch cycle"* (e.g. its equivalent): The sequence of this cycle here is not of flysch-type: it is made up of a normal alternation of limestone and shale, with only minor evidence of resedimentation (Dult Formation). Another difference is that it does not begin at the boundary of Tournaisian/Visean, but only at the base of the Namurian B (or according to the East European subdivision from the base of the Bashkirian). However, the event may probably be accounted for within the hiatus occurring at the end of the *Gnathodus bilineatus bollandensis* Zone in the Namurian A; cf. Ebner 1976).

4. *Correlation of the Variscan sequences of the two regions*

In contrast to the Carnic Alps (the best-known Paleozoic of the Alps, cf. Kreutzer 1992), microscopic sedimentological investigations in the Szendrő Unit are not possible because of metamorphism, and even in the Uppony Unit they are possible only in exceptional cases (mostly in limestone olistoliths of the olistostromes, protected by the matrix). Also the siliciclastic rocks and the metamorphosed platform carbonates cannot be directly dated biostratigraphically due to the lack of fossils, which were obliterated by metamorphism. Nevertheless, based on their macroscopically observable lithology and sedimentary facies, as well as on the general characteristics of the Variscan sedimentary cycle in the Alps, the following Austroalpine (e.g. Upper Austroalpine) and South Alpine, respectively North Hungarian formations can be correlated with each other (bearing the above-mentioned uncertainties in mind):

4.1. Carnic Alps–South Karawanken Mts. ↔ Szendrő–Uppony Paleozoic

Upper Ordovician siliciclastic formations (1st cycle): Bischofalm Quartzite ↔ Rágyincsvölgy Quartz Sandstone, Csernelyvölgy Sandstone (tentatively, since biostratigraphic evidence is lacking in the latter).

Silurian–lowermost Carboniferous deep water, euxinic formations (2–4. cycle): Bischofalm Facies (Lower and Upper Bischofalm Shales, Zollner Formation) ↔ Tapolcsány Formation (tentatively, since biostratigraphic evidence is lacking in the latter).

Silurian pelagic limestones (Wenlockian and Ludlovian of the 2nd cycle): Kok Limestone, Alticola Limestone (reddish varieties) ↔ pelagic limestone olistoliths of the Strázsahegy Formation.

Late Lower Devonian – early Upper Devonian carbonate platforms (3rd cycle): carbonate platforms of the Carnic Alps–South Karawanken Alps ↔ Uppony Limestone, Rakacaszend Marble, Bükkhegy Marble (the latter can be proven only in the Frasnian).

Upper Devonian–Lower Carboniferous pelagic limestones (4th cycle): Pal Limestone + Kronhof Limestone ↔ Abod Limestone + Dedevár Limestone, as well as the pelagic fissure fillings of the Rakacaszend Marble. In the Abod Limestone, however, a difference is the coeval basic volcanism and tuffaceous–tuffitic admixture.

Flysch Stage (5th cycle): Hochwipfel Flysch ↔ Szendrő Phyllite, Éleskő Formation. The olistostromal–turbiditic character is common, but carbonate platforms interfingering with the flysch (as the Rakaca Marble with the Szendrő Phyllite) are missing in the Carnic Alps–South Karawanken Mts. However, their connection is indicated by the recently discovered exotic limestone clasts of carbonate platform facies in the Hochwipfel Flysch: see Flügel and Schönlaub (1990). Flysch sedimentation, however began in the Szendrő Paleozoic no earlier than in the late Visean (*Paragnathodus nodosus* zone). The Kirchbach Limestone dividing the Hochwipfel Flysch into two parts is analogous with the lower part of the Verebeshegy Limestone (*Paragnathodus nodosus* zone).

Posttectonic molasse (6th cycle): The Mályinka(?) Formation of the Lázbérc Subunit of the Uppony Mts., containing quartz and lydite pebbles, seems to be comparable with the Auernig Group of the Carnic Alps.

4.2. *Graz Paleozoic ↔ Szendrő–Uppony Paleozoic*

Pre-Middle Devonian siliciclastic–volcanic complexes (first, 2nd and partly the 3rd cycle in the Carnic Alps): Hundsberg Quartzite ↔ Rágyincsvölgy Quartz Sandstone (tentatively, as biostratigraphic evidence is lacking in both cases), Arzberg Formation Irota Formation.

Middle Devonian–Frasnian carbonate–siliciclastic, coral-bearing formations (third cycle in the Carnic Alps): Barrandei Limestone, Tyrnaueralm Fm., Zachenspitz Fm. ↔ Szendrőlád Limestone Fm.

Middle Devonian–Frasnian carbonate platforms (3rd cycle in the Carnic Alps): Hochlantsch Limestone, Schöckel Limestone, Kanzel Limestone ↔ Rakacaszend Marble, Uppony Limestone (the Bükkhegy Marble of similar facies differs in its lithology; furthermore, the latter can be proven only in the Frasnian).

Middle–Upper Devonian basic volcanics: metavolcanics of Tyrnaueralm Fm. ↔ metavolcanics of the Frasnian part of the Abod Limestone Fm.

Upper Devonian–Lower Carboniferous pelagic basinal facies (4th cycle in the Carnic Alps): Steinberg Limestone (lacking volcanics), Sanzenkogel Limestone ↔ Abod Limestone (containing volcanics), Dedevár Limestone (the latter only reaches as far as the Lower Visean).

Formations of the flysch stage (5th cycle in the Carnic Alps): here the heteropic facies of flysch can be well correlated to each other: Dult Formation ↔ Lázbérc Formation (c.f. Ebner et al. 1991) (see, however, the differences in age assignment in the footnote on p. 18).

4.3. Northern Graywacke Zone ↔ Szendrő–Uppony Paleozoic

The Rágyincsvölgy Quartz Sandstone and the Csernelyvölgy Sandstone of the Tapolcsány Subunit can be compared with the thick Upper Ordovician quartzite and graywacke formations of the Noric Nappe, based on their lithology (but lacking biostratigraphic evidence). However, the associated volcanics are unknown in the former.

5. *Summary*

As can be seen from the above correlation of the formations of each sedimentary cycle, the Szendrő Paleozoic in the pre-flysch stage shows close relationships to the Graz Paleozoic and with certain of its tectofacial units. The absence of Lower Devonian shelf carbonates and the analogies between the calc-phyllite-bearing Irota and Arzberg Formations are analogous with the Schöckel Facies. The equivalents of Eifelian–Frasnian, coral-bearing carbonate–clastic sequence of the Szendrőlád Limestone occur in all tectofacial units of the Graz Paleozoic, with the exception of the pelagic Laufnitzdorf Group. Corresponding formations of the higher Middle Devonian(?)–Frasnian platform carbonates and of the pelagic fissure fillings of Rakacaszend Marble, containing mixed Upper Devonian–Lower Carboniferous conodont faunas, are well known both in the Graz Paleozoic and in the Carnic Alps–South Karawanken Mts. The metamorphosed platform carbonates (bluish-white, banded marbles) of the Rakacaszend Marble and the Schöckel Limestone formations are lithologically identical in the two regions (a further analogy between the Szendrő and the Schöckel Units is their similarly low-grade Alpidic metamorphic overprint). The tuffitic admixture and metatuff intercalations of the Upper Devonian pelagic Abod Limestone of "cipollino" structure, however, show a difference, though Frasnian basic volcanics are known in the Zachenspitz Formation (which shows relations to the Szendrőlád Limestone) of the Hochlantsch Facies and in places "cipollino" structure can be recognized in the Upper Devonian "Flaser-Netzkalk" of the Carnic Alps as well.

As opposed to the pre-flysch stage, in the flysch stage the Szendrő Paleozoic (and also the Bükk; see Ebner et al. 1991) is unambiguously related to the Carnic Alps–South Karawanken Mts, with its olistostromal–turbiditic, flysch-type sedimentation. However, even in this stage some characteristics of the Graz Paleozoic can be recognized in the Szendrő Paleozoic, which relate to the beginning of the flysch-type sedimentation and the end of carbonate sedimentation: 1. whereas in the Carnic Alps flysch sedimentation began at the Tournaisian/Visean boundary (immediately following the *Scaliognathus anchoralis* zone), in the Rakaca Subunit of the Szendrő Unit it did not begin before the Late Visean (lower *Paragnathodus nodosus* zone) and the bulk of the Szendrő Phyllite was deposited after the Early Bashkirian (as the limestone olistostromes occurring at its lower part contain clasts of the *Idiognathoides sinuatus* zone); 2. carbonate sedimentation in part of the depositional area of the Rakaca Subunit, similarly to the Rannach facies (Dult Formation) of the Graz Paleozoic, lasted until the Early Bashkirian (*Idiognathoides sinuatus* zone), as opposed to the Kirchbach Limestone (lower-upper *Paragnathodus nodosus* zone) of the Carnic Alps. This latter fact has some regional paleogeographical consequence: carbonate sedimentation up to the *Idiognathoides sinuatus* zone in context with the Variscan flysch sedimentation was recorded in the Dinarides as well: from the Medvednica Mts. NW of Zagreb (Durdanovic 1973) and from the Javorje Mts belonging to the East Bosnian–Durmitor Unit (in sense of Dimitrijevic 1982) (Djordjijevski-Kalambokis et al. 1992). From the Drina–Ivanjica Unit Serpukhovian (*Gnathodus bilineatus bollandensis* zone) conodont biostratigraphic data are reported from limestones connected to the flysch (Djordjijevski-Kalambokis et al. 1990; Sudar pers. comm.). In the Jadar Paleozoic turbiditic–olistostromal sedimentation appeared only in the Early Moscovian (Pesic 1982). These imply an eastward or southeastward (according to present coordinates) time-shift in the beginning of the Variscan flysch sedimentation in comparison with the Carnic Alps.

Formations of the Tapolcsány Subunit of Uppony Mts, according to their lithology and sedimentary facies, correspond to the Bischofalm Facies of the Carnic Alps. The higher Lower Devonian or Middle Devonian volcanic–olistostromal sequence of the Strázsahegy Member represents, however, a difference: these are unknown in the Carnic Alps (nevertheless coeval basic volcanics are common in several subunits of the Graz Paleozoic). The Silurian pelagic limestone olistoliths of the olistostrome have, however, their equivalents in the reddish varieties of the pelagic carbonate formations of the Carnic Alps–South Karawanken Alps (Kok Limestone, Alticola Limestone). Equivalents of the Lochkovian crinoidal limestones can be found within the Rauchkofel Limestone (Kreutzer 1992b) (there is some microfacial evidence for this: cf. Kreutzer, op. cit., pl. 11, fig. 1 and Kovács 1989a, fig. 6b) The olistostrome of the Éleskő Formation may correspond to the flysch stage; its Emsian–Early Famennian pelagic limestone olistoliths containing styliolinids have their corresponding facies in the Carnic Alps. However, here also a time-shift in the onset of flysch-type sedimentation cannot be excluded (see above, in case of the Szendrő Phyllite); in this case the continuation of the deposition of Tapolcsány Fm. in the Visean could be possible).

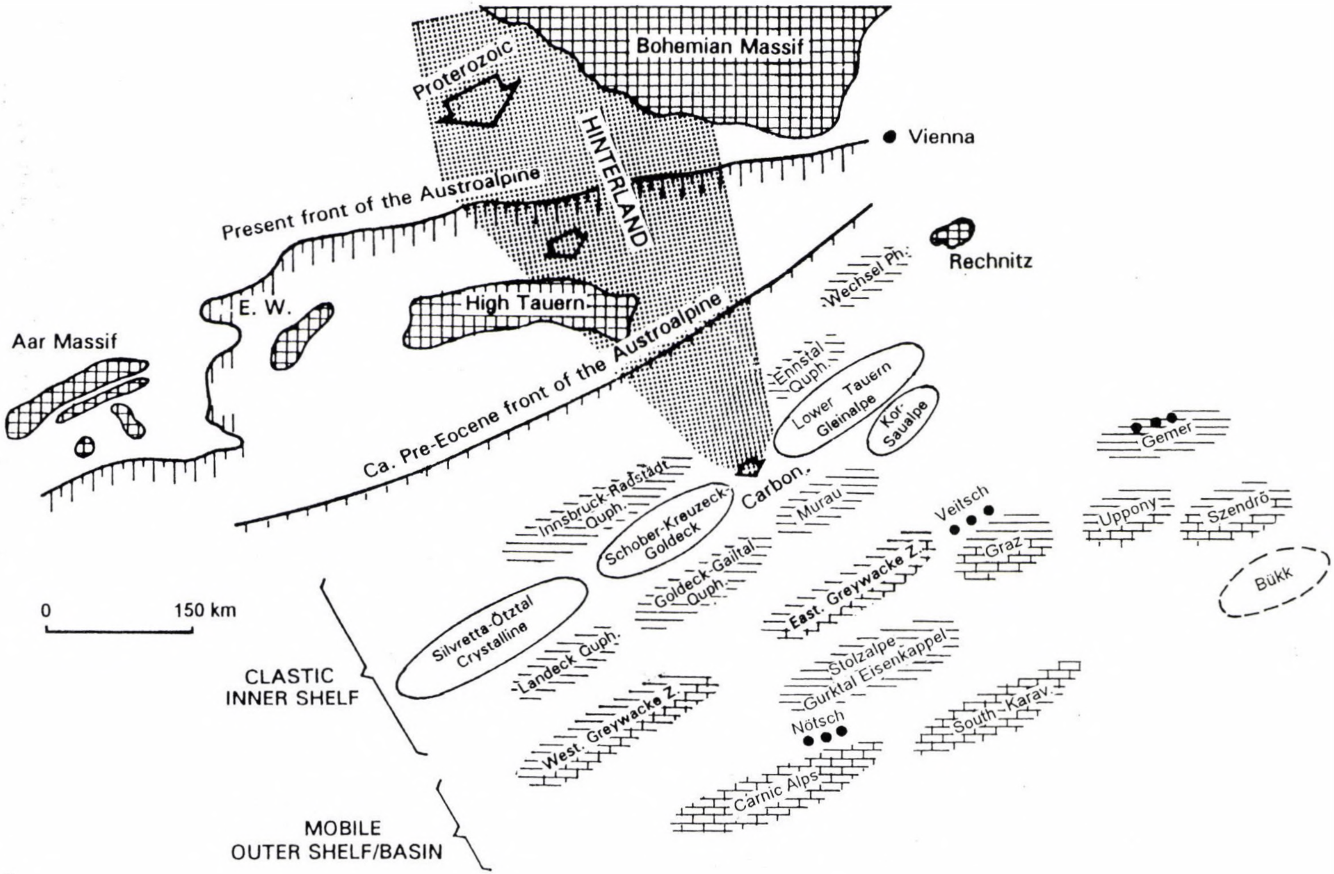


Fig. 6a
Paleogeographic cartoon of the Austroalpine and South Alpine Variscan sequences (Schönlaub 1979, modified after Neubauer 1988, von Raumer and Neubauer 1993, 1994; and Ratschbacher and Fritsch 1993) and position of the Szendrő–Uppony (and Gemer) Paleozoic relative to it

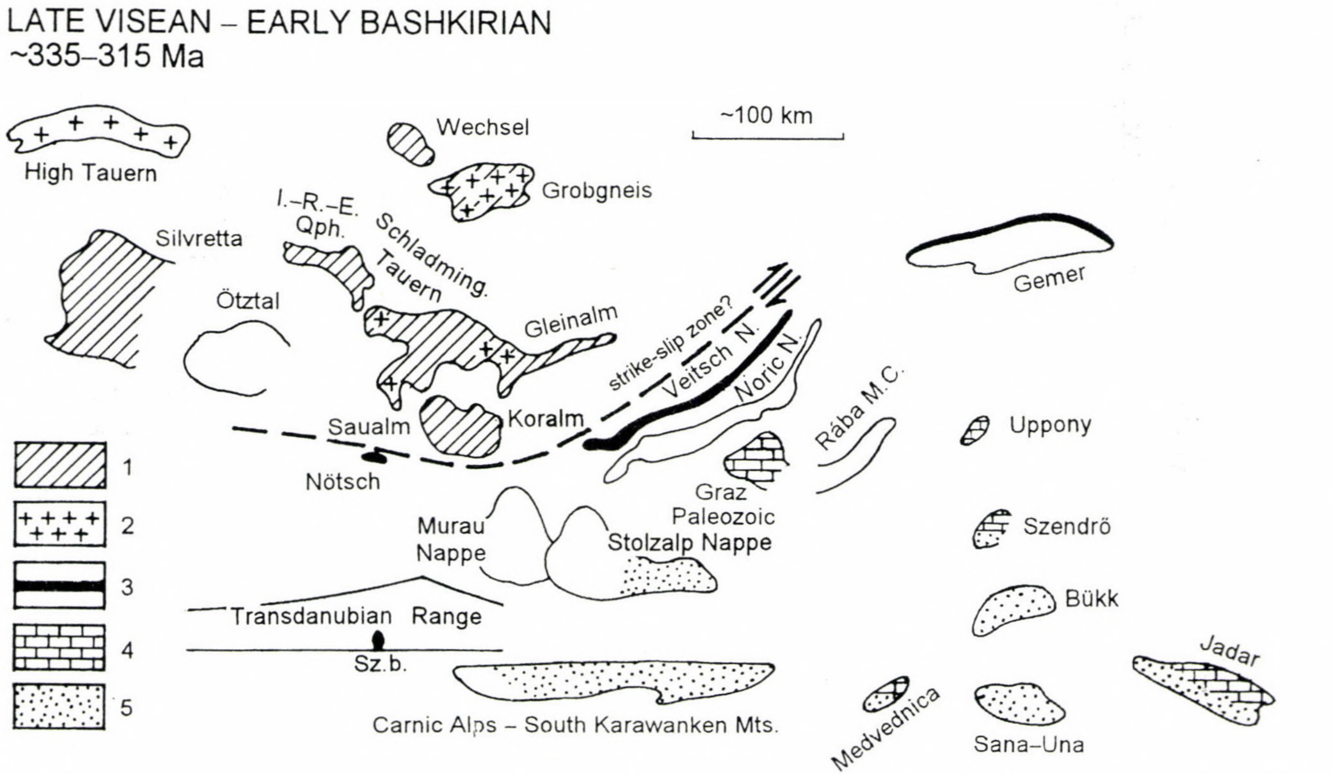


Fig. 6b
Approximate Pre-Mesozoic setting of Pre-Alpine basement units of the Eastern Alps (after von Raumer and Neubauer 1993) and tentative fitting of North Pannonian, resp. eastern South Alpine–Northern Dinaridic units to it (relative to facies), for the Variscan flysch period. Position of Dinaridic units after Filipović et al. (1993), as well as personal communications by I. Filipović, D. Jovanović and J. Pamić. Fitting of the Transdanubian Range Unit is mainly inferred from its Late Permian–Jurassic relationships. Sz.b. = Szabadbattyán Carboniferous (Nötsch-type). 1. metamorphism; 2. granitoids; 3. marine molasse of the Nötsch–Veitsch–Ochtina zone; 4. non-flysch-type intrashelf basinal facies; 5. flysch facies

The analogies of the Middle Devonian(?)–Frasnian carbonate platform (Uppony Limestone) and Upper Devonian pelagic limestone (Abod Limestone) of Lázbérc Subunit of the Uppony Mts can be found both in the Carnic Alps–South Karawanken Mts and in the Graz Paleozoic. A difference is represented, however, by the basic volcanism (schalstein, lava) and tuffitic admixture of the Abod Limestone. This volcanism is missing in the Carnic Alps (perhaps only the "cipollino" structure can be found in the "Flaser-Netzkalk"), whereas in the Graz Paleozoic it is present only in the Frasnian Zachenspitz Formation, which is, in turn, related to the Szendrőlád Limestone (see the Szendrő Paleozoic). When considering also the Lower Carboniferous of the subunit (with the Dedevár Limestone, bearing a lydite horizon, the equivalent of the Sanzenkogel Formation) the relationship to the continuous development of the Rannach Facies of the Graz Paleozoic is more conspicuous. Finally, in the flysch-stage (as shown by the analogous development of the non-flysch type Lázbérc and Dult Formations[1]; Ebner et al. 1991) the close connection of the subunit to the Graz Paleozoic (especially to the Rannach Facies) is unambiguous.

Paleogeographic implications

As could be seen from the above comparison, formations of the Szendrő–Uppony Paleozoic and Carnic Alps–South Karawanken Alps and of the Paleozoic can mostly be fairly well correlated and show, in general, close relationship. The North Hungarian units can be well-fitted (see Fig. 6) into the paleogeographic scheme reconstructed for the Eastern and Southern Alps (Schönlaub 1979, fig. 25; Scharbert and Schönlaub 1980, fig. 3). All these predominantly carbonate sequences, which basically differ from the much thicker volcano-sedimentary Variscan of the entire Carpathian arc, were originally located much closer to each other on the southern (Apulian) carbonate shelf of the Alpine–Dinaric arm of Prototethys (sensu Tollmann and Kristan-Tollmann 1985). From this original proximal position they were dismembered only during the Cretaceous nappe movements and (for the North

1 In spite of the identical facies, in the opinion of the authors there are still slight differences in consideration of the importance of certain sedimentary events. A significant event can be recognized in the uppermost Visean lower *Paragnathodus nodosus* zone in the Szendrő–Uppony Paleozoic by the appearance of the typical variety (dark bluish-grey limestone) of Verebeshegy Limestone and of Lázbérc Limestone, and similarly, in the Carnic Alps by the appearance of Kirchbach Limestone. The dark limestone appearing in the *Paragnathodus nodosus* zone ("Dult Limestone, Ist type" according to the classification of Ebner 1976) can also be pointed out in a few sections of the Rannach Facies of Graz Paleozic. However, it rests without a hiatus upon the underlying Sanzenkogel Limestone; in turn, the "Dult Limestone IInd type" (according to the classification of Ebner 1976) follows above it with a considerable hiatus and begins only in the Namurian B (with the conodonts *Idiognathoides noduliferous* and *Id. sinuatus*). For this reason, Ebner (from 1978 onwards) considers only the IInd type of limestone as part of the Dult Formation and still assigns the Ist type (known only at a few localities) to the Sanzenkogel Formation. Beside the similar lithology another difference is that in the Verebeshegy Limestone and Lázbérc Limestone sediments of the intermediate zones can also be proven by conodonts.

Hungarian units) during the Paleogene strike-slip movements (at the tectonic escape of the Pelso Megaunit or "composite terrane"; Kázmér and Kovács 1985; Fülöp et al. 1987; Csontos et al. 1992). The stronger siliciclastic influx in the Szendrő–Uppony Paleozoic, which manifests itself all over the pre-flysch stage and can be recognized even in the carbonates of the flysch stage (Verebeshegy Limestone, Lázbérc Limestone), also indicates a more pronounced northern connection than shown by the Graz Paleozoic. This stronger siliciclastic influx explains why formations of the Szendrő–Uppony Paleozoic are usually poorer in conodonts than the corresponding ones in the Alps. In addition, another more northerly character is the more intense basic, tholeiitic volcanism of the Uppony Paleozoic (Vető-Ákos, in Kovács and Vető-Ákos 1983) in the Devonian.

The overlap of South Alpine and Upper Austroalpine facial characteristics in the Szendrő–Uppony Paleozoic (pre-flysch stage: relationship predominantly with the Graz Paleozoic, whereas in the flysch stage South Alpine-type development in the Szendrő Paleozoic and also in the Bükk) is also an evidence that the Periadriatic Lineament was not a paleogeographic boundary "separating two worlds" in the Paleozoic (and also not in the Mesozoic). According to the terrane classification of Frisch and Neubauer (1989), the Paleozoic of South and Austroalpine domains belongs to one unit, to the Noric Composite Terrane. The sharp boundary character of the lineament is only the consequence of sizeable Paleogene strike-slip movements in connection with the eastward escape of the Pelso Megaunit (c.f. Kázmér and Kovács 1985; for the most recent review see Neubauer et al. 1995). The relative position of the units discussed in this paper that can be reconstructed for the Carboniferous flysch stage is shown in Fig. 11 of Ebner et al. (1991); (for a more detailed discussion of these problems the reader is referred to this previous work by the present authors).

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Basic geologic data from the Croatian part of the Zagorje–Mid-Transdanubian Zone

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The most characteristic tectonostratigraphic units of the Zagorje–Mid-Transdanubian Zone are represented in outcrops of the Medvednica Mts. of northwestern Croatia. Following its present-day structural stacking order from the bottom to the top, four main tectonostratigraphic units can be distinguished: 1. Tectonized ophiolite mélange; 2. Paleozoic–Triassic magmatic–sedimentary complex overprinted by Early Cretaceous metamorphism; 3. Late Cretaceous–Paleocene flysch; and 4. Triassic clastic and platform carbonate formations. Based on data available the main characteristics of each tectonostratigraphic unit are reviewed.

In the Zagorje–Mid-Transdanubian Zone in Croatia units showing Alpine and Dinaridic characteristics, respectively occur. Borehole data indicate that the previously supposed location of southern part of the Zagreb–Zemplen Fault Zone is to be expected further eastward. We propose that the area adjoining the Zagreb–Zemplen Fault Zone and the southern part of the Mid-Transdanubian Zone be taken as a northwestern boundary of the Internal Dinarides.

Key words: Zagorje–Mid-Transdanubian Zone, Medvednica Mts, mixed Alpine–Dinaridic lithologies, Zagreb–Zemplen Fault Zone, northwestern boundary of the Internal Dinarides

Introduction

The Mid-Transdanubian Zone (MTZ) was first defined by Wein (1969) using the name "Igal–Bükk eugeosyncline". In some current geotectonic sketch-maps of Hungary this zone, which can be traced along strike for more than 100 km in a NE–SW direction, wedges southeastward of Mt. Bakony (Haas et al. 1990). In other geotectonic schemes of Hungary it is a continuous zone about 400 km long which is named as Igal and Bükkium unit in its southeastern and northeasternmost parts, respectively (Árkai et al. 1991). In all these tectonic considerations, the MTZ stretches southwestwards up to the Drava River and obviously continues towards northwestern Croatia, mainly Hrvatsko Zagorje, and therefore we propose to name it Zagorje–Mid-Transdanubian Zone (ZMTZ). According to current opinions this area, which is commonly included in the northwestern Dinarides, belongs to the northwesternmost parts of the "Bosniden" (Kober 1952), i.e., the internal Rodopides, Sava folds (Petkovic 1958),

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Zagreb zone (Dimitrijević 1982) or represents a transitional zone between the Slavonian Mts. and the Supradinaricum (Herak 1986).

In northwestern Croatia, the ZMTZ can be traced for about 120 km along strike. The zone stretches in a NE–SW direction which is nearly perpendicular to the NW–SE strike of the Dinaridic structures (Fig. 1). With its southwesternmost boundary the ZMTZ is thrusted onto the northeastern margin of the External Dinarides, whereas west-southeastwards the zone continues in the system of the Sava and Julian Savinja nappes (Mioč 1982).

According to published data (Bérczi-Makk et al. 1993; Bérczi-Makk and Kochansky-Devidé 1981; Haas et al. 1988; and others) the ZMTZ in Hungary consists of Lower, Middle and Upper Permian calcareous silts, dolomites and limestones, Upper Permian silts, dolomites and limestones, Scythian dolomitic marls, Middle Triassic Buchenstein beds and algal dolomites, Norian dolomites, Mesozoic ophiolite mélange and Cretaceous (?) red claystones, silts and conglomerates. All relevant paleontological and petrographic data were obtained only from core samples taken in boreholes.

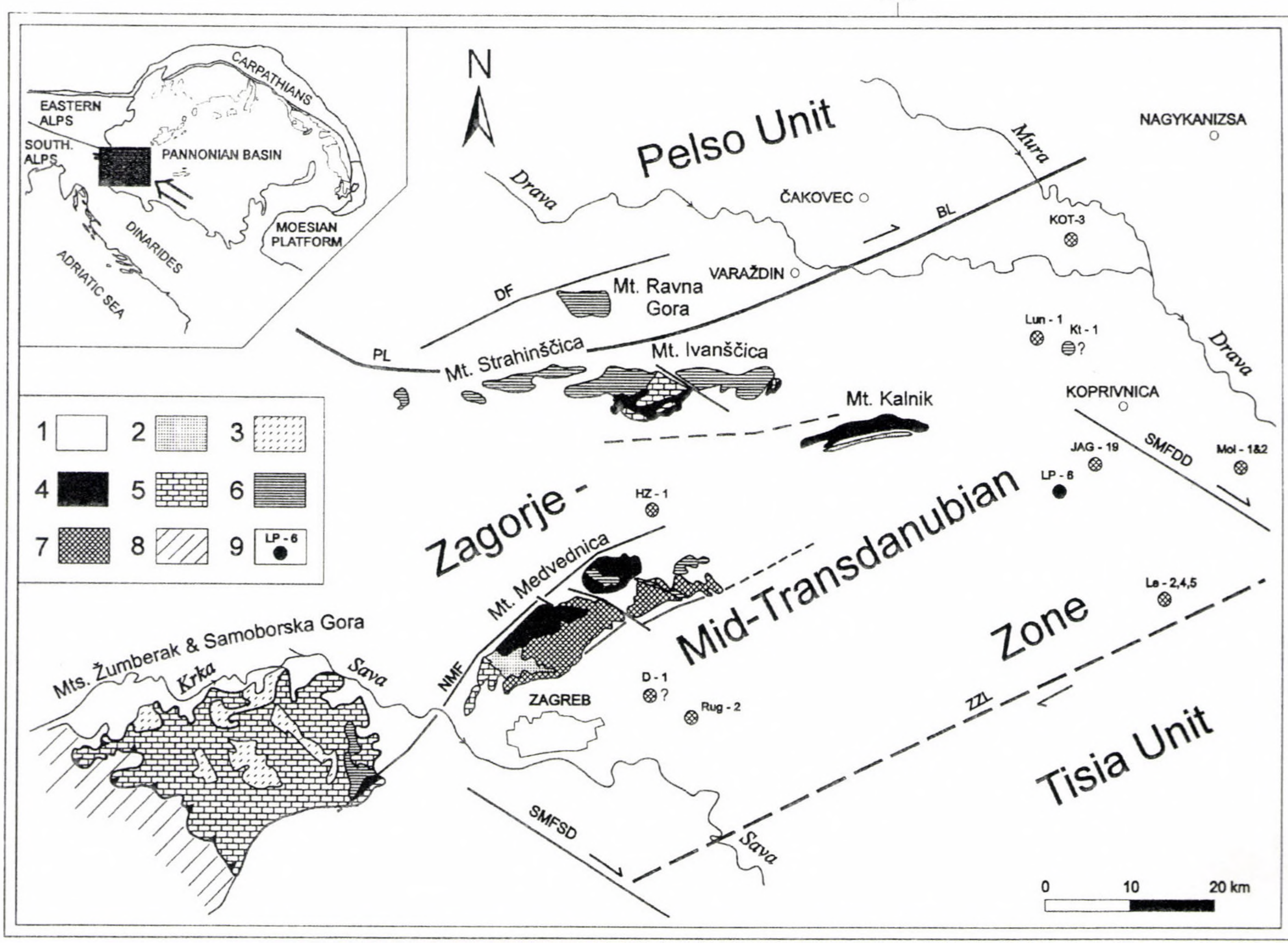
Comparatively high mountains occur in the Croatian part of the ZMTZ (Ivanščica, Kalnik, Strahinčica, Ravna Gora, Medvednica, Žumberak and Samoborska Gora) in which all characteristic lithostratigraphic units of the ZMTZ crop out. The aim of this paper is to characterize major tectonostratigraphic units of the MTZ in Croatia which are best preserved in the Medvednica Mts. Furthermore, available borehole data are also taken into consideration.

Tectonostratigraphic units of the Medvednica Mts.

According to the current geotectonic interpretations the Medvednica Mts, which can be traced along a NE strike for about 40 km, belong to the northwestern parts of the Dinarides (e.g. Herak 1986). The mountain is composed of numerous Paleozoic, Mesozoic–Paleogene and Neogene formations (Šikić et al. 1977; Basch 1981; Šikić 1995) which can be grouped into four main tectonostratigraphic units. The present-day structural succession from the bottom to the top is as follows (Fig. 2): 1) Tectonized ophiolite mélange; 2) Paleozoic–Triassic magmatic–sedimentary complex overprinted by Early Cretaceous metamorphism; 3) Late Cretaceous–Paleocene flysch; and 4) Triassic sequences mainly of carbonate platform facies.

Fig. 1 →

Geological sketch map of the Croatian part of the Zagorje–Mid-Transdanubian Zone. 1. Neogene and Quaternary of the Pannonian Basin; 2. Late Cretaceous–Paleocene flysch; 3. Hauterivian to Cenomanian pelagic limestones and calcareous turbidites; 4. Ophiolite mélange; 5. Late Triassic platform carbonates; 6. Late Paleozoic and Triassic clastics and carbonates interlayered with volcanics and tuffs; 7. Paleozoic–Triassic metamorphic complex; 8. External Dinarides; 9. Corresponding Pre-Neogene basement rocks penetrated by deep oil-wells; DF – Donacki Fault; PL – Periadriatic Lineament; BL – Balaton Line; NMF – Northern marginal fault of Medvednica Mts; SMFSD – Southern marginal fault of the Sava Depression; SMFDD – Southern marginal fault of the Drava Depression; ZZL – Zagreb–Zemplén line


16°30'
46°20'
N
EASTERN ALPS
CARPATHIANS
SOUTH. ALPS
PANNONIAN BASIN
DINARIDES
ADRIATIC SEA
MOESIAN PLATFORM
Pelso Unit
Drava
Mura
NAGYKANIZSA
ČAKOVEC
BL
KOT-3
VARAŽDIN
DF
Mt. Ravna Gora
PL
Mt. Strahinščica
Mt. Ivanščica
Lun - 1
Kt - 1
KOPRIVNICA
Mt. Kalnik
SMFDD
JAG - 19
Mol - 1&2
LP - 6
1
2
3
4
5
6
7
8
9
HZ - 1
Zagorje -
Mid-Transdanubian
Zone
Mt. Medvednica
Le - 2,4,5
Mts. Žumberak & Samoborska Gora
Krka
Sava
NMF
D - 1
ZAGREB
Rug - 2
ZZL
Tisia Unit
SMFSD
0
10
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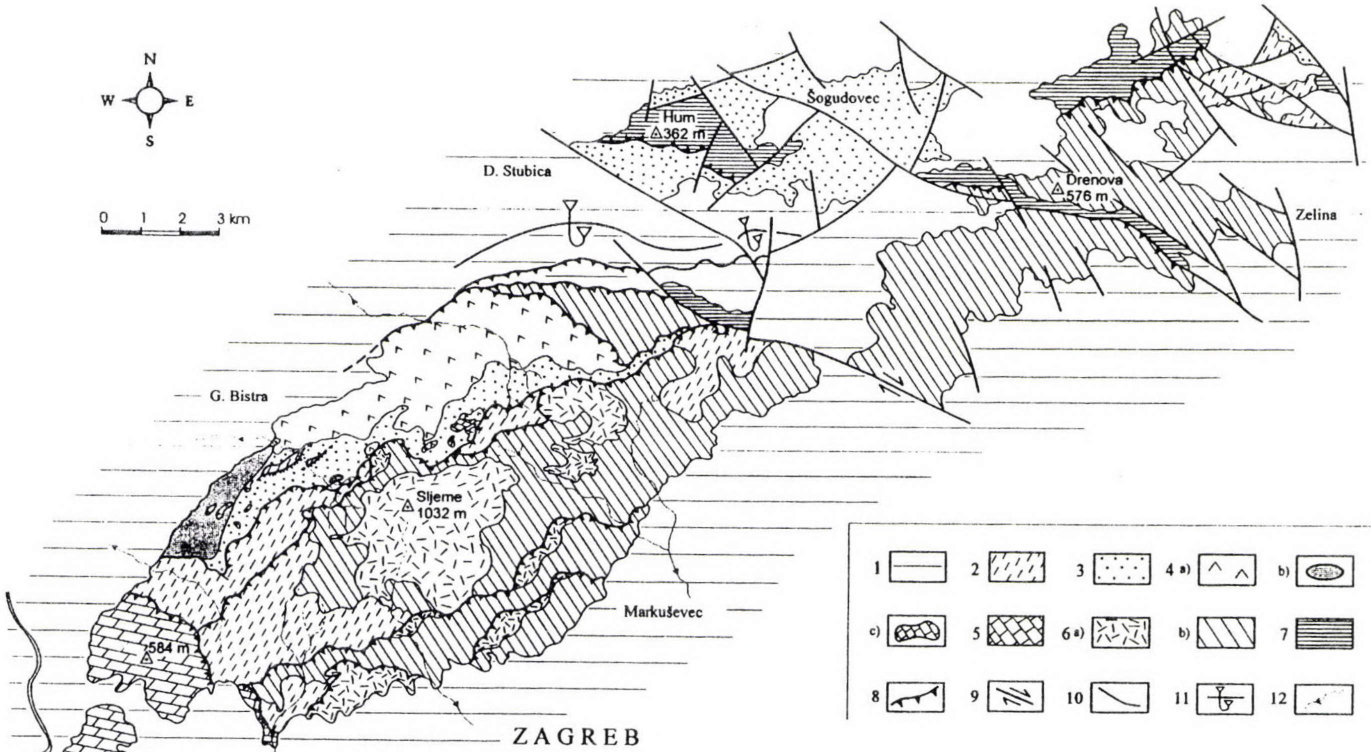

Podsused

Fig. 2
Simplified map of the main tectonostratigraphic units of Mt. Medvednica (modified after Šikić et al. 1977; Basch 1981; Šikić 1995). 1. Neogene and Quaternary sedimentary rocks; 2. Late Cretaceous–Paleocene flysch; 3. Tectonized ophiolite mélange; 4. Larger blocks in ophiolite mélange: a) mafics; b) radiolarites, c) limestones; 5. Triassic mainly platform carbonates of the Sava nappe; 6. Paleozoic–Triassic metamorphic complex: a) orthometamorphic rocks, b) parametamorphic rocks; 7. Mesozoic, mostly Triassic clastics and carbonates; 8. Major reverse or thrust faults; 9. Strike-slip fault; 10. fault; 11. axial trace of overturned syncline; 12. main creek

The relationship of these units with the basement is ambiguous. In the northeastern part of Medvednica Mts, in the area of Hum and Drenova (Fig. 2), the Mesozoic unit, i.e. Triassic clastics and carbonates with condensed Jurassic–Cretaceous sequence, crop out in a way that, according to interpretation of Basch (1981), might have been originally the basement of the ophiolite mélange. A similar structural relationship is known between the ophiolite-bearing unit and underlying unit composed of Late Paleozoic and Triassic clastics and carbonates with Anisian volcanic and tuff intercalations in the Ivanščica Mts (e.g. Šimunić et al. 1982; and others), (Fig. 1). Thus, we can expect that the same Late Paleozoic–Triassic unit occurs in the basement of the ophiolite mélange also to the south and east (Fig. 1). This unit is extensively described by Šimunić and Šimunić (1992), analyzing the Mesozoic sequences of Hrvatsko Zagorje, and will not be considered in this paper. The Neogene formations of the Pannonian Basin are omitted as well.

Tectonized ophiolite mélange

The tectonized ophiolite mélange represents the lowermost exposed structural unit in the present-day structural succession. This is a chaotic unit characterized by a pervasively sheared shaly-silty matrix containing fragments of predominate native graywackes and basalts, diabases, gabbros, serpentinites, radiolarites, shales and exotic limestones. Relict interlayering of shales and graywackes with basalts is preserved only in some places (Tomljenović and Pamić 1997). These formations are poor in characteristic fossils; rare findings of angiosperms indicate the Early Cretaceous sedimentary age of some parts of the unit.

Fossiliferous rock fragments found within the tectonized ophiolite mélange provided data for the age of the reworked formations. The limestone fragments are Triassic to Late Cretaceous in age. In the Kalnik Mts Paleocene limestone fragments were also encountered in carbonate megabreccias associated with tectonized ophiolite mélange (Šimunić and Šimunić 1982).

The tectonized ophiolite mélange shows Dinaridic affinities in general but with some peculiarities. In contrast to the tectonized mélange in the Medvednica Mts, in the Dinaride Ophiolite Zone an olistostrome mélange occurs in which the youngest limestone fragments are Tithonian. Moreover, the Dinaride ophiolite mélange includes large ultramafic massifs (500–1000 km^2), whereas the tectonized ophiolite mélange of Medvednica Mts includes only small fragments of ultramafics, at most a few metres in diameter (Pamić 1982).

These and some other differences stem from variations in their primary geotectonic setting. The tectonized ophiolite mélange of the Medvednica Mts is related to the Vardar Zone *sensu lato* whose main Late Cretaceous–Paleogene sedimentary, metamorphic and magmatic units formed along a subduction zone and related magmatic arc, i.e., the active continental margin of the Dinaridic Tethys (Pamić 1993).

Paleozoic–Triassic magmatic-sedimentary complex affected by Early Cretaceous low-grade metamorphism

Rocks of this metamorphic complex which make up the southeastern parts of Medvednica Mts, including its main ridge, are thrusted onto the tectonized ophiolite mélange. According to Belak et al. (1995) the complex is composed mainly of metasediments: metagraywackes, quartz-muscovite schists, slates, phyllites, marbles, recrystallized limestones and dolomites, and varieties of greenschists containing quartz, feldspars, muscovite, chlorite ± stilpnomelane and chloritoid. Metaconglomerates and quartzites are subordinate. The metasediments are interlayered with orthogreenschists with common remnants of pristine metadiabases and metagabbros. Based on geochemical data the orthogreenschists, which make up the main ridge of the Medvednica Mts, display a distinct ocean-tholeiitic affinity (Pamić 1985/86).

It is obvious that the protolith of the metamorphic complex represented a magmatic–sedimentary formation composed of clastic and carbonate sedimentary rocks interlayered with basalts and tuffs and intruded by diabases and ophitic gabbros. This pristine magmatic–sedimentary complex might correspond to Ladinian–Late Triassic rifting-related formations which are known in some parts of the Dinarides (Pamić 1984).

The age of the protolith as a whole has not been reliably defined so far (Belak et al. 1995). Available paleontological data indicate Silurian to Ladinian–Carnian age. Based on conodont determinations (Đurdanović 1973) the greater part of the complex is Devonian to Late Triassic. On the basis of the presently available paleontological documentation, however, the Paleozoic and Triassic parts of the complex cannot be separated (mapped).

The magmatic–sedimentary protolith was affected by Early Cretaceous very low to low-grade synkinematic metamorphism. It is constrained by six K-Ar ages of 122–110 Ma. The measurements were made mainly on muscovite concentrates separated from ortho and paragreenschists (Belak et al. 1995).

The Medvednica metamorphic complex can be correlated with metapelites, metapsammites and metacarbonates penetrated in boreholes in the area of Barcs (Árkai et al. 1991; Lelkes-Felváry et al. 1996). The same rocks were penetrated in about 10 boreholes located close to the Zagreb–Zemplen Fault Zone in Croatia (Fig. 1 and Table I).

Late Cretaceous-Paleocene flysch

Rocks of this unit crop out along the main ridge of the Medvednica Mts unconformably overlying both the Paleozoic–Triassic metamorphic complex and the tectonized ophiolite mélange.

The sequence begins with poorly sorted, chaotic and massive red conglomerates interlayered with sandstones and siltstones. The composition of the conglomerates directly reflects the petrography of nearby sediment source rocks. These sediments are interpreted as semi-arid alluvial fan deposits (Crnjaković 1979; Pavelić et al. 1995).

The succession grades upward into well-sorted conglomerates and coarse-grained sandstones of Santonian and Campanian age. Upsection more pelagic and fine-grained sediments, i.e., thin-bedded siltstones and laminated shales occur grading into hemipelagic *Scaglia-type* biomicrites. At the northeastern edge of the mountain ranges similar hemipelagic biomicrites cover rudist bioherms of Campanian age (Korolija et al. 1995).

Since the sequence was first described by Gorjanović-Kramberger (1908) it has been interpreted as a transgressive and basinal facies. Based on comprehensive studies the same idea was supported by numerous workers (e.g. Nedela-Devidé 1957; Crnjaković 1979; Marinčić et al. 1995; Pavelić et al. 1995; and others).

Triassic sequences, mainly of carbonate platform facies

These rocks crop out in the southwestern part of the Medvednica Mts as the highest structural unit overthrusting the ophiolite mélange, Paleozoic-Triassic metamorphic complex and the Late Cretaceous–Paleocene flysch sequence (Šikić et al. 1977).

The Triassic sequence includes Scythian reddish to gray, thin-bedded silty shales, interbedded with calcareous lithic arenite. They grade into sandy oosparite (Šikić et al. 1979). These are conformably overlain by Anisian recrystallized dolomites and dolomitic limestones, interlayered in some places with shales, cherts and pyroclastics. Large masses of dolomites also occur. They are probably Ladinian in age although reliable paleontological data are missing. The Norian–Rhaetian is represented by fossiliferous, shallow marine dolomites with desiccation cracks and erosional surfaces (Šikić et al. 1979; Fuček et al. 1995).

Discussion

The present structures in the Medvednica Mts and the surrounding ranges near the southwestern margin of the Pannonian Basin resulted mainly from young Pliocene–Quaternary tectonics.

Data presented for the Medvednica Mts clearly show the mixed Alpine–Dinaridic character of its main tectonostratigraphic units. Correlation with borehole data from the surrounding area shows that a significant proportion of rocks of different tectonostratigraphic units is different compared to those cropping out in the Medvednica Mts. Based on borehole data presented in Table I it is obvious that the most frequently penetrated rocks are those which belong to the Paleozoic–Triassic low-grade metamorphic complex accompanied by subordinate ophiolites. The presented data also suggest that the role and position of the Zagreb–Zemplen Fault Zone should be reconsidered:

Table I
Basic data from boreholes in the Croatian part of the Zagorje–Mid-Transdanubian Zone

Borehole	Depth (m)	Overlying rocks	Basement rocks	Comments
Dubrava 1 (D-1)	2156	2137 m Fossiliferous Lower Badenian silty and sandy marls	2137–2156 m Quartz-sericite schists	? Equivalent of the Mt. Medvednica metamorphic complex
Hrvatsko Zagorje-1 (HZ-1)	5181	3650 m Eggenburgian–Early Badenian: conglomerates, breccias, marls, sandstones, sands, volcanics	3650–3885 m Breccias, conglomerates, graywackes, shales	Barren
			3885–4180 m Metabasalts, partly cataclasized with an interlayer of tectonized black shales	
			4180–4548 m Black shales, breccias and dolomites	Barren
			4548–4553 m Stripped marbles and greenschists of tuffaceous origin	Equivalent of the Adolfovac formation (The Medvednica complex affected by Early Cretaceous metamorphism)
			4553–4730 m Black shales	
			4730–4830 m Breccias, cataclastic with fragments of slates, phyllites, and dolomites	Barren
			4830–5181 m Black phyllites, in some parts moderately to strongly calcitized	The complete penetrated profile belongs probably to the Mt. Medvednica metamorphic complex affected by Early Cretaceous metamorphism
Jagnjedovac-19 (JAG-19)	?	?	1539–1540 m Muscovite-quartz schists and phyllites	? Equivalent of the Mt. Medvednica metamorphic complex
Kotoriba-3 (KOT-3)	4000	3230 m Early and Middle Miocene: shales, siltstones, sandstones and fossiliferous biocalcarenites and biorudites	3230–3540 m Barren recrystallized dolomitic limestones	The complete drilled basement profile (460 m) belongs probably to the Mt. Medvednica metamorphic complex affected by Early Cretaceous metamorphism
			3540–3590 m Altered, partly brecciated tuffs and tuffites	
			3590–3970 m Barren low-grade metapelites with glassy quartzite interlayers at 3750–3755 m	
			3970–4000 m Chlorite-sericite-quartz schists	
Kutnjak-1 (Kt-1)	2430	2019,7 m Lower to Middle Miocene shallow-marine subgraywackes with globigerinas (Moslavačka Gora Formation)	2019,7–2284 m Micaceous quartz-sandstones with silica cement, recrystallized and silicified	? Scythian formation
			2284–2315 m Brecciated dolomites	Barren, presumably Middle Triassic
			2315–2430 m Low-grade metamorphosed limestones with a few interlayers of low-grade marly limestones	? Inverse (overturned) sequence

Borehole	Depth (m)	Overlying rocks	Basement rocks	Comments
Lepavina-6 (LP-6)	678	635 m Lower to Middle Miocene fossiliferous marls and sandstones (Moslavačka Gora Formation)	635–649 m Magnesite-serpentinite breccias 649–678 Serpentinites	Ophiolites
Leticani-2 (Le-2)	2355	2182 m Egerian marls	2182–2355 m Quartz-chlorite-muscovite schists	Equivalent of the Mt. Medvednica metamorphic complex
Leticani-4 (Le-4)	2930	2625 m Egerian barren marls and micaceous sandstones with coal seams	2625–2780 m Quartz breccias and conglomerates, cataclasized 2780–2848 m Altered greenish volcanic rocks 2848–2930 m Quartz-chlorite-sericite schists	Equivalent of the Mt. Medvednica metamorphic complex
Leticani-5 (Le-5)	2765	2400 m Lower to Middle Miocene marls interlayered with sandstones and coal	2400–2439 m Cataclasized quartz sandstones 2439–2675 m Quartzite breccias 2675–2708 m Quartz sandstones 2708–2734 m Altered volcanic rocks 2734–2765 m Quartz-sericite schists	Scythian ? Equivalent of the Mt. Medvednica metamorphic complex
Lunkovec-1 (Lun-1)	?	?	1997,5–2037 m Graphitic slates with metabasalts 2037–2040 m Schistose metagraywackes 2040–2201 m Graphite-calcite slates	Equivalent of the Mt. Medvednica metamorphic complex
Molve-1 (Mol-1)	3280,4	3206 m Badenian fossiliferous marls and sandstones	3206–3280,4 m Quartz-sericite schists	Equivalent of the Mt. Medvednica metamorphic complex
Molve-2 (Mol-2)	3560	3124 m Alternation of marls and sandstones (Iva Sandstone Formation)	3124–3172 m Breccias with quartz-mica fragments 3172–3560 m Quartz-mica schists	Equivalent of the Mt. Medvednica metamorphic complex
Rugvica-2 (Rug-2)	2989	2985 m Lower to Middle Miocene fossiliferous sandstones and marls	2985–2989 m Slate-phyllites and metagraywackes	Equivalent of the Mt. Medvednica metamorphic complex

it turned out that it is located further southeastward from its previously assumed position (Fig. 1).

It has been a generally accepted idea that the Dinarides continue without any break into the Southern Alps. This is, however, only partly true because only the External Dinarides extend without any break into the Southern Alps, whereas in the area west of the Zagreb–Zemplen Fault Zone the internal units of the Dinarides are mainly concealed beneath the allochthonous Paleozoic–Triassic formations of the Sava and Julian–Savinja nappes (Mioč 1982). In this area there are no outcrops of ophiolites. Ophiolites of the southeastern part of the ZMTZ represent the northwesternmost outcrops of the Dinaridic ophiolites and thus the area adjoining this zone and the Zagreb–Zemplen Fault Zone can be taken as the northwestern boundary of the Internal Dinarides.

This proposal fits with the geotectonic division of the basement of the Pannonian Basin (Fülöp and Dank 1987; and others) according to which the western Pelso megaunit is characterized by Alpine lithologies, whereas the western Tisia megaunit bears elements of the Carpathians. In such an interpretation the Zagorje–Mid-Transdanubian Zone has a distinct transitional Alpine–Dinaridic character.

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Geologic and paleomorphological observations of the Neogene and the Pleistocene of the Parnass zone (Greece). Application to the exploration for and exploitation of bauxites

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Geologic field observations and exploratory drilling in the third (Middle Cretaceous) bauxite horizon NW of Amfissa have made it possible to reconstitute a pre-Pliocene paleosurface which cuts the hanging wall of the bauxite (a 500 m deep paleocanyon). This erosional paleosurface is covered by fluvial, proximal conglomerates assigned to the Plio-Pleistocene. This pre-Pliocene paleosurface, when compared to the paleocanyons of the western Mediterranean, can be related to the Messinian regression, the sedimentary filling to the Plio-Pleistocene sea level rise when the paleocoast was situated a little south of the northern shoreline of the present-day Gulf of Corinth. Later the Plio-Pleistocene uplifts resulted in partial erosion of the conglomerates. The development of the present-day topography and of secondary karstification phenomena may have considerably affected the bauxite deposits. The bearing of these events on the exploratory drilling and on the underground or open-cast mining is examined, taking into account the observations made during the past ten years.

Key words: Bauxites, paleomorphology, paleokarst, tectonics, Messinian, Greece

Introduction

Geologic studies, exploration by drilling, and mining of the bauxites of the third horizon (Middle Cretaceous) in the Parnass zone permitted new sedimentological and geomorphological observations concerning the lithology and the Neogene–Pleistocene evolution of this region. The Parnass Zone consists of a mighty (about 2,000-m thick) carbonate sequence of Triassic to latest Cretaceous age, constituting the bulk of the mountain. Above this sequence continental sediments were deposited and morphologic surfaces and karstic dissolution phenomena developed during the Neogene and the Pleistocene. These have produced relative uplifts due mainly to neotectonism, and less obviously to the overdeepening caused by the Messinian regression. These young events are characterized by specific facies and structures which must be taken into consideration for the appropriate location of exploratory boreholes and of underground or open-cast mining operations.

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Field observations and borehole evidence

The Pre-Pliocene paleosurface and the conglomerates at Pera Lakko – Papa Lakka (NW of Amfissa) – Fig. 1

Geologic setting

The study area is situated 5 km NNW of Amfissa (1:50,000 geologic map, sheet Amfissa), between 800 and 1.200 m a.s.l., i.e. about 500 m higher than the Rodia gully and the Viniani depression located 1 km farther east. It is made up of a more than 500 m thick series dipping 30–40 degrees W, consisting of Lower Cretaceous limestones (footwall of the B3, the third bauxite horizon) and of Upper Cretaceous ones (hanging wall), overlain by red-colored flysch of Paleocene age. In the west an inverse fault at Lianokladi has produced an eastward overthrusting of the Lower Cretaceous limestones (Fig. 1). On these rocks a morphological paleosurface has developed, covered by a thick mass of conglomerates dated as Pliocene (Célet 1962).

The Pre-Pliocene paleosurface and its conglomerate cover

These can be observed both on the Lower and Upper Cretaceous limestones and on the Paleocene red flysch as well. The Pliocene conglomerates are in a horizontal position; the 30 to 40 degree angular unconformity is particularly clearly visible in the south of the area (Fig. 1b–c). Four main facies could be distinguished:

P1: Over a thickness difficult to determine but estimated at several meters at least, the Cretaceous limestones are finely recrystallised. The result is a network of millimetric fissures which give the rock the resemblance of pseudomylonite. However, microfossils (e.g. Orbitolinae) are still visible in the intact zones. The displacement of the constituents of this breccia is nil or very small (a few cm); only the structure of the limestone (mudstone–wackestone) has been modified. Accordingly, it is an "alteration facies" due to the recrystallisation of the limestones, beginning with the pre-Pliocene erosion surface.

P2: The preceding facies develops further by collapse and dissolution, producing subangular fragments of 1 to 20 cm in diameter, which move vertically and laterally, depending upon the topography. The result is a poorly cemented, loose and heterogeneous breccia, the constituents of which abound on the surface. The dislocations are only of the order of a few meters or a few tens of meters. Accordingly there is a considerable proportion of identical constituents, realistically reflecting the composition of their original substratum. This breccia covers its own substratum which appears irregularly, with or without the P1 facies. Should there have been outcropping bauxite at the inception of this P2 breccia then it is also integrated in the system (Fig. 1a), forming an accumulation of bauxite blocks in the breccia, among fragments of the footwall and hanging

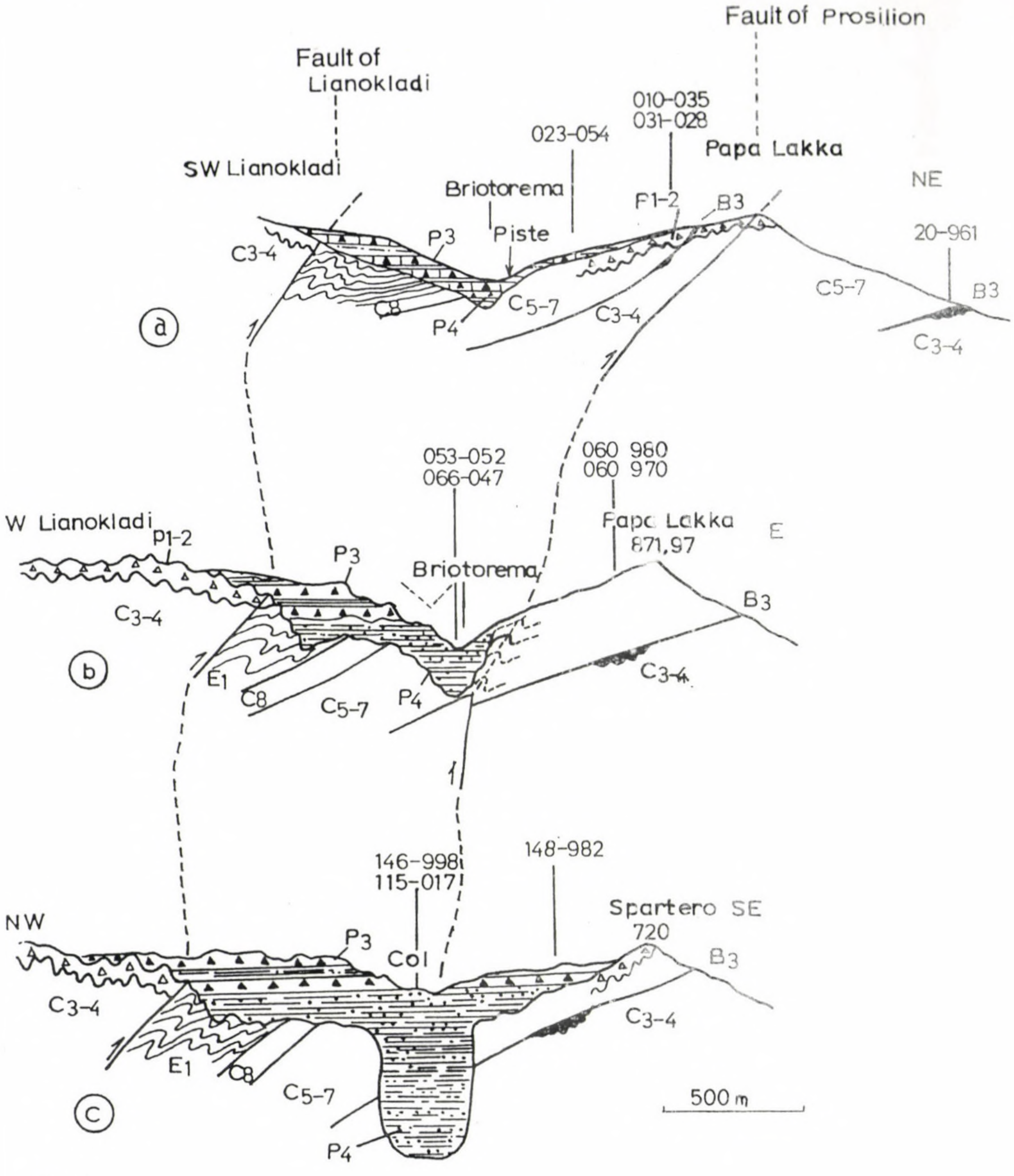

Fig. 1
Sector Pera Lakko – Papa Lakka, series of oriented cross-sections displaying the structure of the substratum, the pre-Pliocene paleosurface and the Pliocene facies. P1 to P4 – Pliocene (description in text); E1 – Paleocene flysch; C8 – Maastrichtian limestone; C5–7 Hanging wall: Upper Cretaceous limestone; B3 – Bauxite and contact with bauxite; C3–4 Footwall: Lower Cretaceous limestone; O10-035: borehole number

wall limestones (Fig. 2). This peculiar indicator surface is of interest despite its unusual heterogeneity, because it indicates the presence of a primary contact of the breccia with the underlying bauxite.

P3: Breccia deposited near the supplier topography, with a considerable lateral displacement of the rock fragments (several hundreds of meters). The fragments are subangular or slightly rounded, 10–15 cm in diameter. The mixing of fragments of different origin (from the footwall, from the hanging wall and some from the dispersed bauxite) occurred on ancient slopes, as did the considerable cementation by carbonates. This makes the rock massive and compact-looking (more or less karsted cliffs) and tends to obliterate the breccia structure.

P4: "Puddings" of rounded or slightly rounded, 5 to 50 cm large rock fragments of different origin. The matrix contains smaller (less than 1 cm) fragments of the same nature and varying amounts of reddish clayey cement. The pebbles are of Parnassian (Jurassic and Cretaceous limestones, less often bauxite) and of Beotian facies (red rudist limestones, pieces of greenstone). The latter originated from the erosion of the Beotian nappe which had covered the entire Parnass zone, and of which some klippes have survived, such as that of Jérolékas (5 km NE). The reddish clays occurring in the matrix or forming mainly pelitic interbeds were produced by the erosion of the Paleocene flysch. The P4 facies, associated higher in the series to P3, corresponds to torrential sediments and proximal river alluvia.

The Pliocene P3–4 filled up an ancient valley, which strikes NNW–SSE, is 6 km long and up to 1.500 m wide (1:50,000 Geologic Map of Amfissa). The pre-Pliocene paleosurface has deeply eroded the Cretaceous series. This is clearly visible on the cross-sections plotted on the basis of field observations and borehole data in the south of the Pera Lakko – Papa Lakka sector (Fig. 1). These allow the reconstruction of the ancient valley. The paleocanyon merits particular attention because it accounts for the abnormal thickness of P 3–4: more than 500 m in borehole No. 146–998 (not completed) and 541 m plus 9 m footwall in borehole No 115–017. Outside the paleocanyon borehole 148–982 encountered only 74 m of Pliocene. This 1500 m long and 300 m wide structure can be followed toward the SSE. The infilling of the paleocanyon is nearly cut by the present topographic surface, and only the paleocanyon has been preserved. Farther toward Amfissa even the Pliocene seems to have been eroded. A gully situated in its prolongation may be partly inherited from the pre-Pliocene paleosurface.

Development of the present-day morphology and related karst phenomena

The Pliocene conglomerates of the Amfissa region constitute a slightly dipping paleosurface about 20 km long, from 1200 m a.s.l. NW of Pera Lakko – Papa Lakka (Prosilion sector) to the Corinthian gulf (Itea region) in the south (Célet 1962; Sébrier 1977). Remnants of this paleosurface can be seen to the W in the landscape in form of rock caps (200 to 800 m) dissected by the Rodia gully north of Amfissa and by the Amfissa valley between Amfissa and Itea.

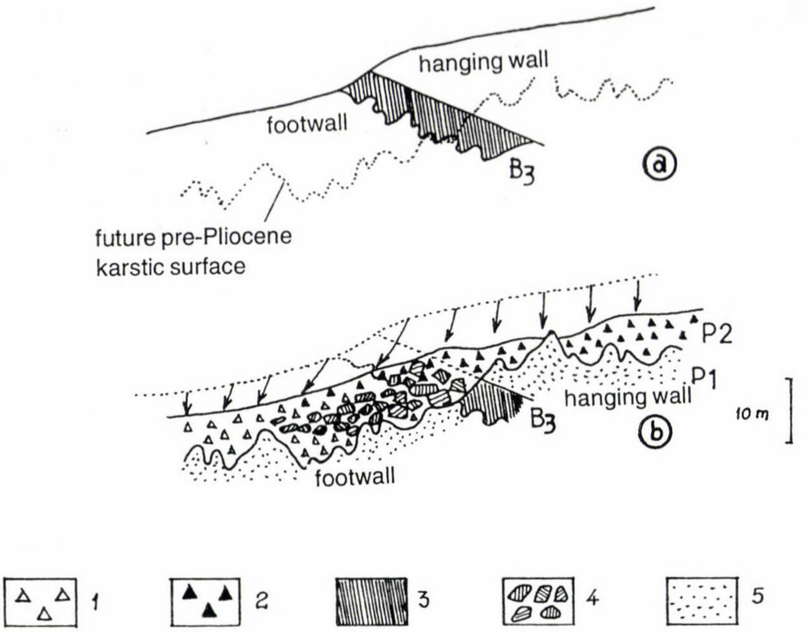

Fig. 2
Reconstruction of the development of the pre-Pliocene paleosurface and the related facies P1–2. a) Original, pre-Pliocene situation of the outcrop; b) Development of the karstic pre-Pliocene paleosurface and the P1–2 facies by collapse dissolution and colluvium formation; 1. P2 breccia with fragments of the footwall; 2. P2 breccia with fragments of the hanging wall; 3. Bauxite horizon; 4. P2 breccia with bauxite fragments; 5. Recrystallised footwall and hanging wall (facies P1)

Accordingly the Pliocene accumulation paleosurface was deeply dissected by the recent morphology. This erosion is accompanied by karstic dissolution phenomena identified during exploration and exploitation of bauxite.

The huge collapse pit of Vartos A1

Between the sector of Pera Lakko – Papa Lakka in the west and the Amfissa–Lamia road in the east (1:50,000 Geologic Map of Amfissa) there is a strongly faulted anticline formed by a 500 to 600 m thick series: Lower Cretaceous, 3rd bauxite horizon–Upper Cretaceous. It is deeply cut by the Rodia gully and its Paleocene flysch cover has been removed. The Vartos–Rodia sector, the left bank of the gully, is now in exploitation (open cast and underground mines). The abandoned Vartos A 1 open pit (3 km east of Pera Lakko – Papa Lakka) and its immediate surroundings offer rather favorable conditions for observation. West of the open pit there is an oval-shaped depression (dimensions: 200 m NS, 50 m EW) in the rudist limestones of the hanging wall. It contains huge blocks of Paleocene flysch separated from each other by numerous gliding planes. These blocks contain subangular fragments up to 50 cm diameter of rudist limestones as well as sandstones with radiolarite and greenschist pebbles. They are characteristic of the Beotian facies. These facies occur in the karstic cavities at the bauxite/hanging wall contact in the Vartos B deposit and also at Kokkini Lakka, east of Desfina (Combes 1978a, b). They are internal sediments supplied by the erosion of the Beotian nappe and can be correlated with the surficial Pliocene conglomerates. The direct contact with the hanging wall all along the contours of the depression displays numerous vertical slickensides, in particular W of the open pit. These suggest a curved normal fault.

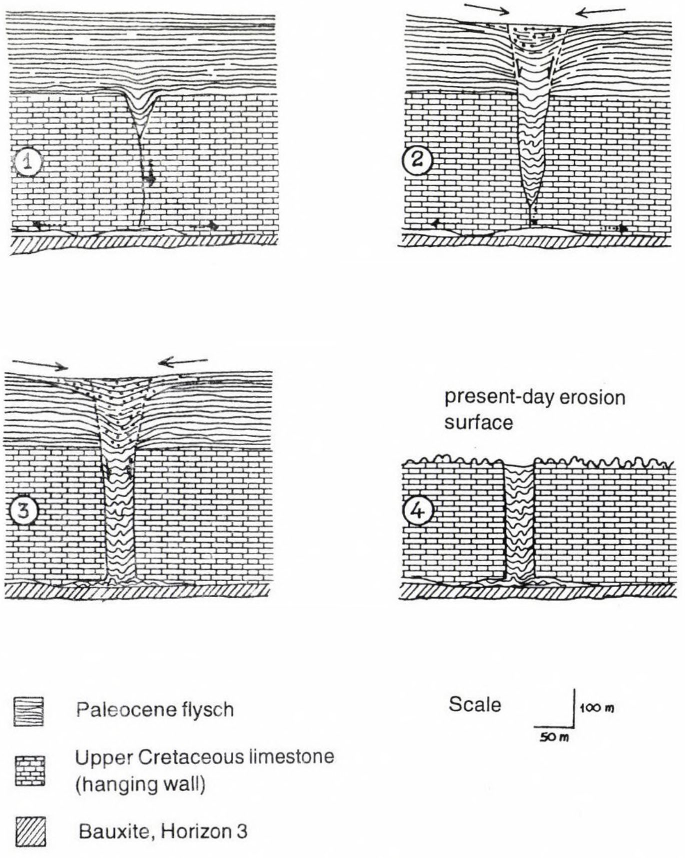

Fig. 3
Reconstruction of a huge Vartos A-type collapse pit produced by dissolution beneath the flysch cover. 1. Beginning of dissolution by water circulation (dotted arrows) along a fracture and on top of the bauxite (impervious level), draw-off of flysch blocks; 2. Continuation of the process, appearance of a collapse pit (curved faults), the flysch draw-off arrives to the surface creating a depression filled with redeposited flysch in the affected area; 3. The pit is completely filled, the system stops functioning; 4. Erosion of the flysch and of part of the Upper Cretaceous limestone of the hanging wall. Reconstruction after the model by Bonte (1963) and Combes (1973)

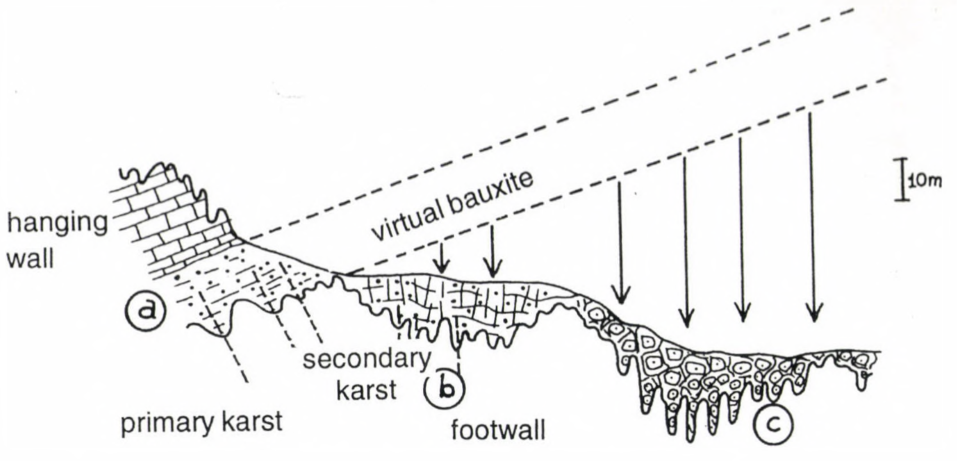

Fig. 4
Development of secondary karst in the Toumba Lakka sector. (a) Original bauxite with primary karst; (b) Moderately drawn off bauxite by collapse dissolution in the secondary karst; (c) Breccia of bauxite fragments mixed up with terra rossa, formed by important draw-off in a well-developed secondary karst (high pinnacles of vertical axis)

The characteristics of this structure indicate a collapse pit provoked by the progress of karstic dissolution beneath the Paleocene flysch. The development of a cavity resulted in successive breaking in of blocks from the overlying flysch, as well as of those from the hanging wall and of intrakarstic sediments. Figure 3 shows a hypothetical reconstruction of this secondary renewal of karstification.

Phenomena of secondary karstification in the Toumba Lakka-Sector (Northern Parnass)

These phenomena are widely developed in the Parnass zone (Combes 1978a, b), particularly in areas of higher elevation and higher annual precipitation, covered by snow during part of the year. The Toumba Lakka sector is situated at 1700–1800 m a.s.l., west of the ski resort of Parnass near the Arachova–Amfiklia road. The different stages of stripping due to the development of a secondary karst can be observed there (Fig. 4). Three categories of deposits can be distinguished.

1. The original bauxite (a) of Horizon 3, as a homogeneous bed (up to 20 m thick) between the hanging wall and the footwall. The primary karst is represented by anastomosed pockets (type 2, Combes 1978c), the axis of which is oblique in the case of a dipping series.
2. Superimposed bauxite in its own footwall, (b) due to the development of secondary karst beneath the bauxitic cover. The moderate dissolution and draw-off of the footwall resulted in "en masse" penetration of the bauxite, with

fissures due to tilting and beginning fragmentation. In the footwall the renewed dissolution determines the neoformation of a system of vertical pinnacles.

3. Breccia (c) composed of angular bauxite fragments up to one cubic meter, embedded in terra rossa. In this case the displacement is considerable, tens to hundreds of meters below the original level of the bauxite. During its vertical (and if the slope permits, also lateral) displacement the bauxite undergoes fragmentation, becomes heterogeneous and eventually gets mixed up with residual clays resulting from carbonate dissolution. The pinnacles of the secondary karst of the footwall may be very well developed (10 to 20 m high).

When secondary karstification occurs in a horizontal series the bauxite is superimposed into the footwall, but there is no angle difference between the axes of the primary and the secondary cavities. In such cases the secondary karst can be recognized by the high frequency of pinnacles in the footwall which have developed preferentially in zones affected by late fracturation, where the draw-off is more important. Given the fact that the fracturation affected the hanging wall as well, dissolution and/or collapse dolines (sinkholes) may occur (Fig. 5).

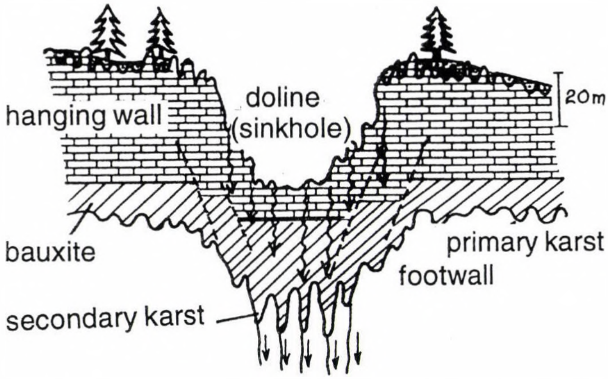

Fig. 5
Toumba Lakka sector, development of a doline in the hanging wall of a secondary karst pocket. Arrows indicate the direction of water circulation

Interpretation and impact on exploration and exploitation

The pre-Pliocene paleosurface cannot be precisely dated. It cuts the Lianokladi and Prosilion overthrusts (Fig. 1) of Late Eocene age and is covered by the conglomerates considered to be Pliocene. Accordingly, the stratigraphic gap ranges from the Late Eocene to the Pliocene. Nevertheless the morphological resemblance of this paleosurface (paleocanyon discovered by drilling) to the one which in the Western Mediterranean is bound to the Messinian regression is impressive. In fact it is known (Hsü et al. 1973; Clauzon

1973, 1982; Saint-Martin and Rouchy 1990) that as a consequence of an important eustatic sea level drop in the Late Miocene (Haq et al. 1987) communication with the Atlantic Ocean partially ceased to exist. The warm and arid climate of those times (Suc and Bessais 1990) produced the complete drying out of the Mediterranean Sea, the level of which might have been, according to some authors, 1.500 to 2.000 m deeper than that of the Atlantic. The result was the appearance of the Messinian evaporites and the overdeepening of riverbeds, the canyons of which are especially well studied in the Western Mediterranean. In the Eastern Mediterranean observations are scarce but similar overdeepening has been reported from the upper course (Chumakov 1973) and the delta (Barber 1980) of the Nile.

We can also assume that analogously to the Western Mediterranean the filling in of the paleosurface may have been subsequent to the rise of the eustatic sea level in the Zanclean (Haq et al. 1987). At that time the Pliocene seashore was located slightly landward of the present-day Gulf of Corinth. The filling of the pre-Pliocene paleosurface would correspond to the Pliocene series of fluvial conglomerates, the top of which constitutes an accumulation surface recognized from Prosilion to Itea and is cut by the recent erosion surface. The latter is a consequence of the Plio-Pleistocene uplifts, accompanied by quite a number of faults, the combined throw of which has been estimated to be about 2,500 m (Sébrier 1977). We can link the phenomena of secondary karstification which affect the bauxite horizons to this phase.

As far as practical application is concerned the identification of different Pliocene facies and the correlative paleosurface allowed us to develop an exploratory drilling program in the Pera Lakko – Papa Lakka sector. For example, the P1 and P2 breccias which developed in situ or suffered only minor lateral displacements can be added to the hanging wall surfaces, thus increasing the area liable to be drilled. On the other hand the pre-Pliocene paleosurface has removed part of the Upper Cretaceous–Paleocene hanging wall, so that the contact with the bauxite can be reached by drilling through the Pliocene infilling. A more accurate reconnaissance is being conducted to delimit the paleocanyon and to check the hydrogeology of this structure. As far as the phenomena of large-scale draw-off are concerned, such as the collapse pit at Vartos A1, during the field work dolines of considerable size were discovered which may be indicative of deep-seated connection, at the level of the underground mining. In preparing the plan of mining operations it is necessary to delimit the protective pillars with particular care in order to avoid such pits. At any rate the Vartos C and Rodia mines which are in operation seem to indicate that the collapse phenomena are only in rare cases as fully developed as at Vartos A1. They are confined most commonly to punctual collapses of the hanging wall ("fontis") which have already been described (Combes 1978a, b). These do not propagate to the surface and therefore are not predictable. The appearance of secondary karst such as at Toumba Lakka, very common in the Parnass zone, is a well-understood and handled phenomenon by now. If the draw-off is moderate (b) the grade of the bauxite is not affected: it can be mined just as the primary ore is. If the draw-off is important (b–c), admixture with the terra rossa results in increased silica content of the bauxite which makes its direct use impossible. A simple mechanical processing system adapted to each particular case is necessary to separate the bauxite fragments, the original grade of which has not been modified.

Conclusion

The events which occurred in Neogene and Pleistocene times in the Parnass zone have deeply affected the geologic and morphological setting of the Middle Cretaceous bauxites (3rd horizon). The important pre-Pliocene paleosurface discovered NW of Amfissa, can be correlated, as a working hypothesis based on analogy with the Western Mediterranean, with the Messinian regression. The thick proximal fluvial series dated as Pliocene which overlies this paleosurface may be related to the return of the sea in the Zanclean, controlling its deposition down to the marine base of erosion situated in the Corinthian gulf, slightly behind the present-day southern coast. The discovery of the complementary processes of erosion and filling up which have affected the hanging wall of the bauxite, as well as the recognition of the resulting facies and morphology, are indispensable for the reasonable exploration and exploitation of the bauxites NW of Amfissa. The Pliocene erosion and the development of the present-day orography, both resulting from the Plio-Pleistocene uplifts, brought about secondary karstic dissolution and draw-off phenomena. These must be understood before mining is begun since they are liable to influence the deposits rather considerably.

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This study was undertaken within the framework of IGCP Project 287 "Tethyan Bauxites" and presented to the 7th International Congress of ICSOBA (1992) at Balatonalmádi – Tapolca (Hungary). This paper therefore represents the knowledge to this date.

Résumé

Les études géologiques et la prospection par sondages au NW d' Amfissa ont permis de mettre en évidence une paléosurface antépliocène recoupant profondément (paléocanyon de 500 m de profondeur) le toit de la bauxite du Crétacé moyen (3ème horizon) et comblée par une série conglomératique fluviatile proximale, rapportée au Pliocène. Par comparaison avec la Méditerranée occidentale, cette érosion est corrélée avec la régression messinienne, le comblement avec la remontée du niveau marin au début du Pliocène dont le rivage se situait un peu au S du littoral septentrional du golfe de Corinthe. Postérieurement, l' érosion partielle du Pliocène est due aux soulèvements plio-pleistocènes responsables de l' orographie actuelle et de phénomènes de karstification secondaire suceptibles de modifier profondément les gisements de bauxite. L'incidence de ces événements sur la prospection par sondages et l'exploitation souterraine ou à ciel ouvert est abordée à l'aide d'observations effectuées depuis une dizaine d'années.

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Miocene acidic explosive volcanism in the Bükk Foreland, Hungary: Identifying eruptive sequences and searching for source locations

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Neogene volcanism began in the Hungarian part of the Carpatho-Pannonian area with explosive eruptions of large-volume acidic magmas during the Miocene. It was followed by andesitic volcanism which is partially coeval with the acidic one. Three main acidic tuff horizons are known, referred to as the Lower, Middle and Upper Rhyolitic Tuff, Eggenburgian/Ottnangian, Karpatian/Lower Badenian and Lower Sarmatian in age, respectively. Their onshore-deposited counterparts, found in the basin-rim regions such as the Bükk Foreland, are difficult to identify and correlate. Complex investigation, including lithostratigraphy, petrography, petrochemistry, K-Ar geochronology and paleomagnetic studies, were recently undertaken in the Bükk Foreland area. The three tuff complexes have been identified, distinguished and characterized. The Lower Tuff Complex (21.0–18.5 Ma, 80° to 90° counterclockwise rotation) is well exposed in the northern part of the area and consists of biotite-rich rhyolitic Plinian pumice fall deposits, welded and non-welded ignimbrites, phreatomagmatic sequences and reworked tuffs. The Middle Tuff Complex (17.5–16.0 Ma, ca. 30° counterclockwise rotation) is widespread in the southern part of the area, and is characterized by a two-component chemistry (rhyolitic and andesitic), presence of mixed scoria-pumice pyroclastics, and an abundance of pyroxene crystalloclasts. It consists of a sequence of welded and non-welded ignimbrites, phreatomagmatic deposits and reworked tuffs. Various degrees of mixing of rhyolitic and andesitic magma result in extremely heterogeneous compositions ranging from typically rhyolitic to andesitic. The youngest Upper Tuff Complex is present in few outcrops at the western and eastern peripheries of the Bükk Foreland, as thick non-welded rhyolitic ignimbrites. Compositionally it is hardly distinguishable from the Lower Tuff Complex, but K-Ar ages are much younger (14.5–13.5 Ma) and no rotation is present in its paleomagnetic record.

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Volcanic source areas of the pyroclastic material should be somewhere south of the area, buried beneath younger sediments. Thickness and maximum pumice and lithic clast sizes of the pumice fall deposits of the Lower Tuff Complex suggest a northeast-oriented dispersal axis; imbrication in non-welded ignimbrites suggest transport in the same direction, hence source location for the oldest tuffs is possibly somewhere south of Eger. An elliptic aeromagnetic anomaly zone, present in that region, is consistent with such a hypothesis. On the other hand, magnetic anisotropy directions measured in welded ignimbrites of the Middle Tuff Complex converge in a zone NE of Mezőkövesd (south of Harsány), where a wide elliptic positive gravity anomaly is located and thick andesite lava was described in boreholes. Maximum size of obsidian fiamme in the welded ignimbrite, measured at Tibolddaróc, is consistent with the location of the inferred source caldera in the area.

Key words: Bükk Foreland, ignimbrites, volcanology, lithostratigraphy, correlation of Miocene tuff horizons

1. *Introduction*

Acidic explosive volcanism marks the start of Neogene magmatic activity in the Carpatho-Pannonian area and occurs repeatedly until the Uppermost Sarmatian. Its products are widespread in the Pannonian Basin, cropping out in the Börzsöny, Cserhát, Mátra, Bükk, Tokaj and Mecsek mountains in Hungary. Together with those known in the Transylvanian basin and even outside the Carpathians in Romania, they occupy an area of ca. 700 x 400 km, mostly covered by younger sediments. Three tuff horizons were recognized in boreholes penetrating into the Pannonian Basin, commonly referred to as the "Lower Rhyolite Tuff" (Paul and Göbl 1866), "Middle Rhyolite Tuff" (Noszky 1933), and "Upper Rhyolite Tuff" (Schréter 1934). Their ages are Eggenburgian–Ottnangian, Karpatian–Lower Badenian and Sarmatian, respectively (Hámor et al. 1980). These tuff horizons are thought to have been emplaced during three main episodes of volcanic activity. They are separated by horizons of interlayering fossiliferous sediments in the deeper parts of the basin, but are hardly distinguishable at, or near the basin margins (where datable sedimentary interbeds are lacking).

Acidic pyroclastics are well exposed in several areas, mostly located along the northern margin of the Pannonian Basin. The Bükk Foreland (Bükkalja in Hungarian) area offers a large number of excellent outcrops allowing a detailed investigation of the Miocene acidic volcanic products.

Two tuff horizons were identified and described (Schréter 1934, Balogh 1964) in an elongated area (ca. 40 km long and 8 km across), located at the southern feet of the Bükk Mts, referred to as the Lower Ignimbrite and Upper Ignimbrite, respectively. Their correlation with the tuff horizons outside the Bükk area is difficult. We will argue, however, that they represent the onshore-deposited counterparts of the Lower and Middle Rhyolite Tuffs of the Pannonian Basin. According to Balogh (1964), the Upper Rhyolite Tuff is also present at the eastern side. A more detailed account of previous research on the Miocene rhyolitic tuffs of the Bükk Foreland and the Pannonian Basin is given by Póka et al. (1998, this volume).

Following the recognition of the ignimbritic nature of the welded part of these tuffs (Pantó G. 1962, 1963) no modern volcanological studies have been undertaken until recently, when the presence of other genetic types of primary pyroclastic deposits (i.e., non-welded ignimbrites, phreatomagmatic deposits), as well as reworked deposits, were pointed out in addition to the welded ignimbrites (Capaccioni et al. 1995). Some strongly welded ignimbrites were mapped as lavas in previous cartographic representations (e.g. Balogh 1964). No attempts, other than speculations, have as yet been made in order to identify eruptive centers of the acidic pyroclastics in the Bükk Foreland, and generally in the Pannonian Basin (Varga 1981).

From 1988 on, systematic paleomagnetic investigations were carried out in this area (Márton 1990) that confirmed the presence of at least two volcanic horizons corresponding to the Lower and Upper Ignimbrites of Balogh (1964).

The main purpose of this paper, focused on the Bükk Foreland area, is to establish the eruptive sequences, styles and mechanisms, as well as to localize possible eruptive centers of the acidic tuff sequences. In addition, we attempted to identify and correlate the onshore-deposited counterparts of the three Rhyolite Tuff horizons of the Pannonian Basin. To achieve these goals we have undertaken a complex investigation in a multidisciplinary approach, including geologic studies (lithostratigraphy, volcanology, petrography and petrochemistry, paleomagnetic investigations, and K-Ar geochronology) as well as the interpretation of available geophysical data. Part of these data are reported in two companion papers in this same volume (Márton and Pécskay 1998; Póka et al. 1998, this volume).

2. *Identification and correlation of tuff horizons*

2.1. Terminology

We will use a terminology similar, but not identical, to that adopted for the Pannonian Basin by Császár et al. (1983) and Hámor (1985), i.e., Lower Tuff Complex (hereafter LTC) = Gyulakeszi Rhyolite Tuff Formation, Middle Tuff Complex (hereafter MTC) = Tar Dacite Tuff Formation and Upper Tuff Complex (hereafter UTC) = Galgavölgy Rhyolite Tuff Formation. "Tuff Complex" seems to us a more appropriate term because neither "Tuff" is a single horizon, but a composite lithostratigraphic unit consisting of various lithologic entities and lithofacies; the ignimbrites are only parts of the volcanoclastic sequences. Moreover, the "Lower Tuff" and the "Middle Tuff" resulted from more than one eruptive event with discernible time gaps between them (Márton and Pécskay 1998, this volume).

2.2 Occurrence

The spatial distribution of the three Tuff Complexes in the Bükk Foreland area is depicted in Fig. 1. It results from a revision of the geologic map of Balogh (1964). Mappable lithologies inside each tuff complex are shown as well.

The Lower Tuff Complex crops out along the northwestern part of the Bükk Foreland, being covered by the Middle Tuff Complex mostly along the southeastern margin of the area. The MTC was not found west of Ostoros (Fig. 1). The UTC is present only patchily at the western (Demjén and north of Egerszalók) and eastern (north of Harsány) parts of the region (Fig. 1). All three complexes were encountered southeast of the Bükk Foreland area in boreholes, being covered by younger sediments in the subsided part of the Pannonian Basin (Fig. 2).

2.3 Lithostratigraphy

The main lithostratigraphic features of the volcanoclastics are presented in the synthetic lithologic column in Fig. 3.

The LTC is composed of two identifiable parts which resulted from two eruptive pulses. The lower part – we name it informally LLTC – is interbedded in red conglomerates with clayey matrix, of probably Eggenburgian age, at Cserépfalu (Karácsonytisztás) or, more frequently, overlies the paleosurface of the Oligocene sediments (the lower horizon at Eger–Tihamér, Eger–Wind brick factory, lower part of Kács–Church hill, Kisgyőr–Kiskút). It consists of biotite-rich non-welded massive ignimbrites (Fig. 4a). These rocks are associated with silica sinters in outcrops around Eger (Egerszalók, Eger–Wind brick factory, Eger–Bányakert). The age of this horizon is supposedly Eggenburgian according to its lithostratigraphic relationships, as well as to a K-Ar dating (ca. 19.7 Ma), (Márton and Pécskay 1998, this volume).

The upper part of the LTC – we informally name it ULTC – unconformably overlies LLTC or older sediments and displays various lithologies. The complete sequence of the ULTC includes reworked tuffs at the base (Egerszalók, lower part in the Eger–Tihamér quarry), then a sequence of alternating Plinian pumice fall and phreatomagmatic fall deposits (Kisgyőr–Kerekhegy, Kisgyőr–Kékmező, middle part in the Eger–Tihamér quarry), non-welded ignimbrites (middle part in the Eger–Tihamér quarry, Ostoros SE, lower part at Tard), and welded ignimbrites (Demjén–Pünkösdhegy, Sály–Latorút, Kisgyőr, upper part at Kács–Church hill, Bükkzsérc). K-Ar ages obtained on rock samples of the ULTC (Márton and Pécskay 1998, this volume) of ca. 18.7 Ma places it roughly at the Eggenburgian–Ottnangian boundary.

The boundary between the LTC and MTC is rarely visible in outcrops; hence mapping it is quite difficult. However, warned by differences in the paleomagnetic features (degree of rotation) of ignimbrites, careful examination of a hillside sequence of discontinuous outcrops at Sály–Latorút led to the recognition of a fluvial gravel horizon intercalated between the two tuff

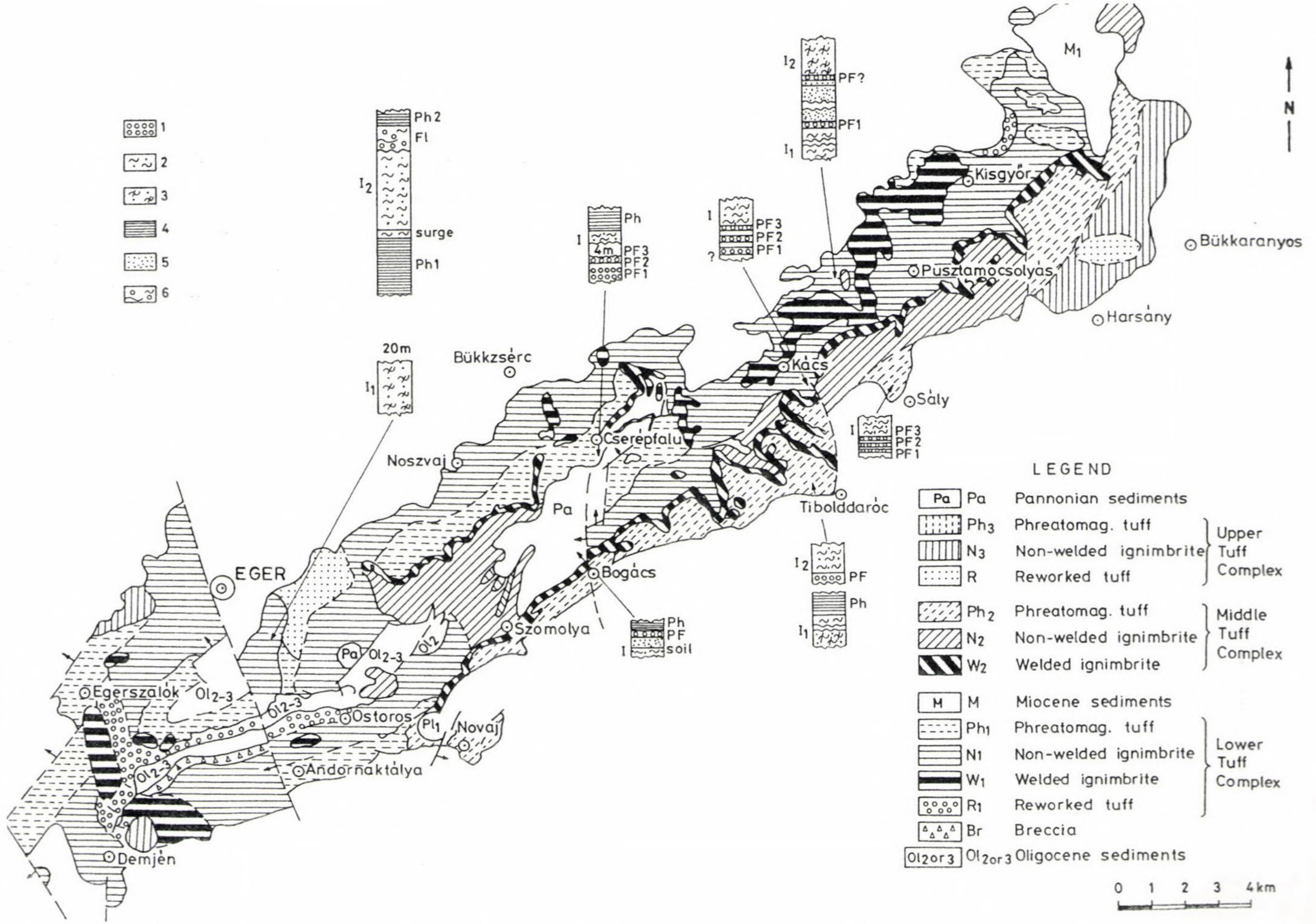


Fig. 1
Geological map showing the spatial distribution of Miocene acidic tuff complexes in the Bükk Foreland (modified after Balogh 1964). Legend of the lithologic columns: 1. fallout pumice lapilli (PF); 2. non-welded ignimbrite; 3. welded ignimbrite; 4. phreatomagmatic sequence (Ph) 5. reworked volcaniclastics; 6. epiclastic rocks (Fl)

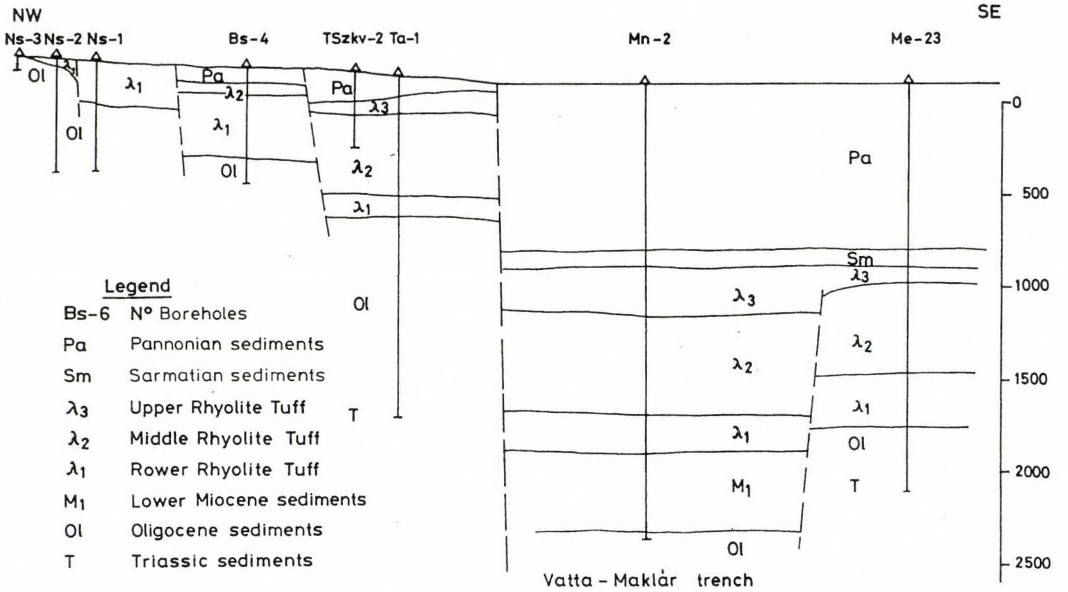


Fig. 2
Cross section showing the distribution of tuff complexes in the Pannonian Basin south of the Bükk Foreland area, according to borehole data. Borehole abreviations: Ns – Noszvaj; Bs – Bogács; TSzkv – Tard–Szekrény völgy; Ta – Tard; Mn – Mezőnyárád; Me – Mezőkeresztes

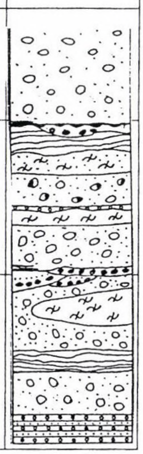
Tuff complex	Lithology	Mineral assemblage	Vitroclast chemistry SiO_2 (%)	K_2O (%)	K-Ar age (Ma)	Paleomagnetic rotation (degrees)	Lithologic column
UPPER TUFF COMPLEX	Reworked tephra Phreatomagmatic deposits Non-welded ignimbrites	Plagioclase Biotite Quartz Amphibole	75.99-76.60	4.69-5.12	13.5-14.5	0	
MIDDLE TUFF COMPLEX	Epiclastic rocks Reworked tephra Phreatomagmatic deposits Welded ignimbrites Non-welded ignimbrites	Plagioclase Pyroxene Biotite Amphibole	70.21-75.18	4.31-5.13	16.0-17.5	30	
LOWER TUFF COMPLEX	Reworked tephra Phreatomagmatic deposits Welded Ignimbrites Non-welded ignimbrites Fallout pumice lapilli	Plagioclase Quartz Biotite Amphibole	73.22-75.68	3.60-5.20	18.5-21.0	80-90	

Fig. 3
Summary of lithologic, petrographic, paleomagnetic and K-Ar data for the three tuff complexes studied in the Bükk Foreland area (petrography and chemical data from Póka et al. 1998, this volume, K-Ar and paleomagnetic data from Márton and Pécskay 1998, this volume)

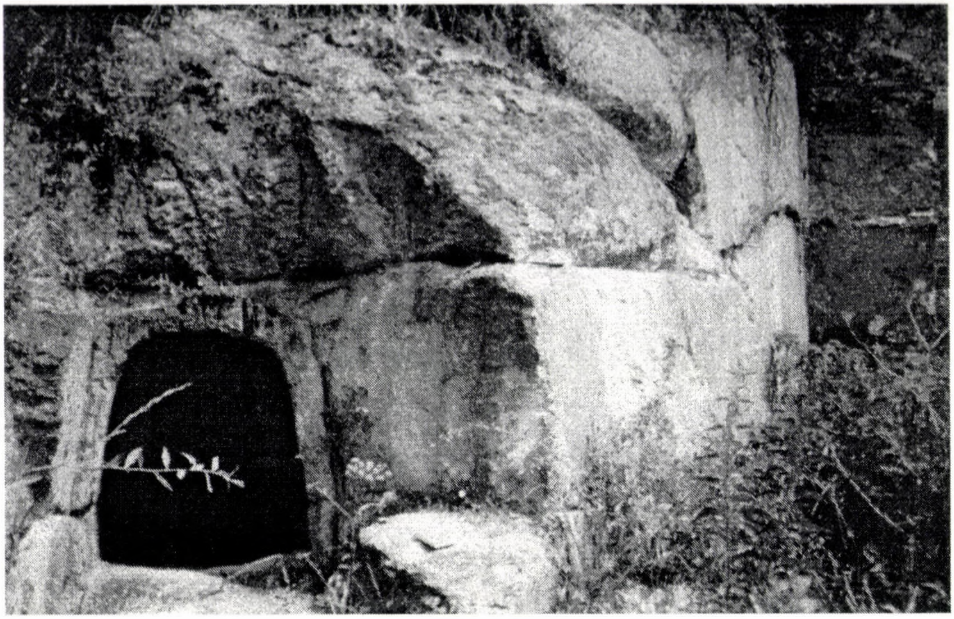


Fig. 4a
Outcrop of the slightly welded to non-welded massive pumice-rich ignimbrite in the Lower Tuff Complex; south of Kisgyőr

complexes. Fluvial pebbles with the same lithostratigraphic position were also found between Kisgyőr and Tibolddaróc. It seems that this continental sediment is discontinuously present in the area, but it clearly marks an important break in volcanic activity and concurrent strong erosion of the top of the LTC. Furthermore, a sequence containing coal beds is present at the same stratigraphic level at Kisgyőr (Leányvár).

The MTC is composed, in turn, of two identifiable sequences, both of them characterized by the presence of more basic (dacitic to andesitic) components along with the rhyolitic material. The lower one – we informally name it LMTC – consists of welded red ignimbrites (Fig. 4b) (Szomolya–Novaj, Cserépfalu, Sály–Latorvár, Kisgyőr) followed by phreatomagmatic deposits (Bogács, Ábrahámhegy-, Szomolya N, Cserépfalu N). K-Ar ages obtained from these rocks cluster around ca. 17.0 Ma (Márton and Pécskay 1998, this volume). Following an erosional unconformity, the upper sequence – informally designed as UMCT – begins with obsidian fiamme-rich gray to pinkish welded ignimbrites (Bogács–Galagonyás, Cserépváralja-Mangótető, Tibolddaróc, Pusztamocsolyás–Meredek-hegy) grading into reddish to dark gray non-welded mixed pumice-scoria ignimbrites (Szomolya, Borsodgeszt–Kerekhegy, Cserépváralja–Kaptárkövek), with accretionary lapilli-bearing phreatomagmatic deposits (Fig. 4c) at the top (Bogács W, Tibolddaróc–Church Hill, Sály S,

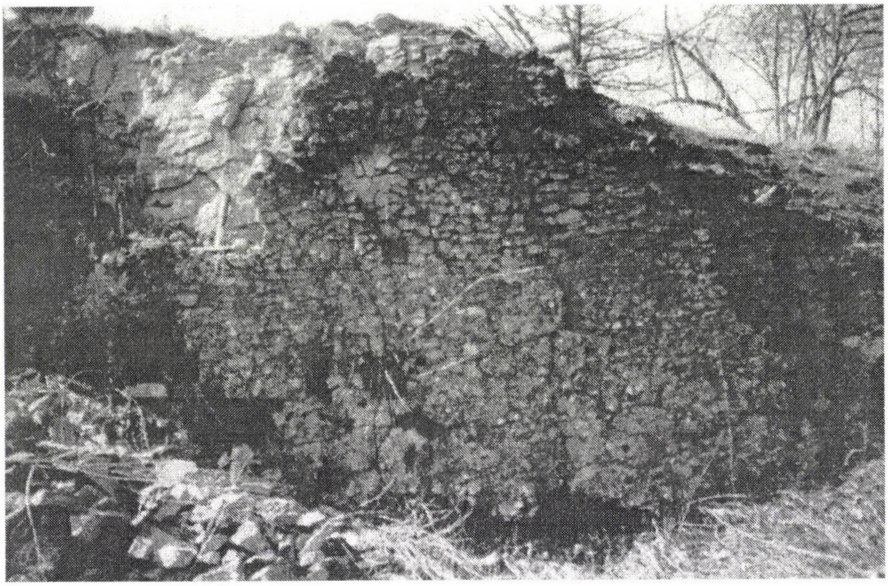


Fig. 4b
Welded ignimbrite with fine columnar jointing normal to a vertical cooling surface; Middle Tuff Complex; Ispánberek (between Szomolya and Novaj)

Fig. 4c
Phreatomagmatic sequence (Ph) overlying a pink, slightly welded mixed pumice-scoria ignimbrite (I) in the Middle Tuff Complex; West of Bogács

Fig. 4d
Non-welded ignimbrite in the Upper Tuff Complex; Harsány

Kisgyőr–Kőbányatető). In the eastern part of the area (between Pusztamocsolyás, Bükkaranyos and north of Harsány), the UMCT is represented by more or less welded dark gray to brown andesitic ignimbrites. The K-Ar ages of these rocks fit well with the average age of the "Middle Rhyolite Tuff" of the Pannonian Basin (16.4 Ma; Hámor et al. 1980; Márton and Pécskay 1998, this volume).

The UTC occurs as patches overlying the LTC in the west or MTC in the east (Fig. 1). It is composed of biotite-rich rhyolitic rocks, especially non-welded ignimbrites (Fig. 4d) (Demjén–Nagyeresztvény quarry, Harsány–cellars, Bükkaranyos–Kaptárkő). Other lithologies, such as reworked tuffs and phreatomagmatic tuffs are less developed and limited to the eastern occurrence area (Harsány–Vargyastető). A K-Ar determination obtained for the western occurrence area (13.84 Ma) suggest an Upper Badenian age for these rocks (Márton and Pécskay 1998, this volume).

2.4 Petrographic features

On the old maps (e.g. Balogh 1964) rhyolites, dacites and andesites are designated as petrographic types of the Bükk Foreland explosive volcanics. Some of them are mapped as lava flows, especially those of andesitic and dacitic composition. Some strongly-welded dark-colored ignimbrites of more basic composition are indeed lava-like in places, but their careful examination invariably points out their pyroclastic origin (presence of pumice and scoria clasts, glass shards and broken crystals). We have not found any rock of lava flow origin in the study area.

Petrography and petrochemistry, besides K-Ar dating and paleomagnetism, are able to discriminate between at least the LTC and MTC, as well as between different entities within the complexes. The LTC and MTC display important differences in their juvenile mineral compositions. LTC rocks are rich in both biotite and quartz, and amphibole appears as an accessory phase among phenocrysts in pumice and crystalloclasts in the matrix of various types of volcanoclastics. Andesitic components only occur as accidental lithoclasts in the ignimbrites (Eger–Wind brick factory). Most lithoclasts are high-silica rhyolites (Póka et al. 1998, this volume).

The MTC is compositionally different in the sense that more basic than rhyolite juvenile components are always present, even in rocks (welded ignimbrites) which macroscopically appear identical to LTC rocks. The andesitic-dacitic components consist of gray to black juvenile scoria, massive dark vitroclasts and pyroxene crystalloclasts. Especially in the lower part of the UMTC the presence of pyroxene crystalloclasts are diagnostic. Mixed pumice-scoria ignimbrites are common. Banded pumices have also been encountered in outcrops. The abundance of non-rhyolitic material is variable, leading to different chemical compositions, from rhyolitic to basic andesitic. In most outcrops the rocks are more and more basic upwards in the lithostratigraphic column. The ignimbrites are entirely of andesitic composition in a large area in the eastern third of the region. In these rocks one still can recognize, however, small amounts of rhyolitic material such as pumice and acidic glass shards, accidental biotite and quartz crystalloclasts. This petrographic-petrochemical variability within a single eruptive unit led the former researchers (e.g. Balogh 1964) to map these rocks as separate volcanological units.

The UTC is petrographically similar to the LTC by its biotite-rich rhyolitic composition. Quartz is, however, less abundant as phenocrysts in pumice and as crystalloclasts in the ignimbrite matrix. Subtle differences can be pointed out in the petrochemical features of the two occurrences of the UTC, as well (Póka et al. 1998, this volume).

Welding is present in both the LTC and MTC ignimbrites, but it is absent in the UTC rocks of the Bükk Foreland. Therefore, this feature cannot be used as a discrimination criterion between the LTC and MTC, as Capaccioni et al. (1995) attempted previously. Welding characterizes the middle part of thick ignimbrite units. Welded ignimbrites are typically present at the upper part of ignimbrites in the Bükk Foreland because the non-welded top of these pyroclastic rocks have been erosionally removed and supplied material for the reworked tuff deposits which frequently overlie the ignimbrites.

2.5 Petrochemistry

Chemical features of the studied pyroclastic rocks are just as valuable discrimination criteria as the petrographic one. They are presented in a companion paper in this volume (Póka et al. 1998).

3. Volcanology

3.1 Eruptive history

3.1.1 Genetic types of volcanoclastic deposits

The Miocene acidic volcanic rocks in the Bükk Foreland area display a rather wide range of genetic types of primary and secondary volcanoclastics. They include welded and non-welded ignimbrites, fallout pumice lapilli and tuff layers, and phreatomagmatic deposits, debris flow deposits and epiclastics.

Ignimbrites are the most characteristic eruptive products in the area. They are ubiquitous in the three tuff complexes. Thick (meters to tens of meters), massive non-welded ignimbrites are most frequent. They contain white to yellow pumice clasts ranging from less than 1 cm to 40–50 cm in diameter and subordinate smaller lithic clasts embedded in a largely developed vitro-crystalloclastic matrix. The presence of darker scoria clasts is obvious within mixed pumice-scoria ignimbrites. Welded ignimbrites characteristically contain flattened pumice and obsidian fiamme which can attain 45 cm length at Tibolddaróc (MTC). These ignimbrites, whether welded or non-welded, are typical pyroclastic flow deposits.

Pumice fall deposits have been identified in a couple of outcrops of the LTC. They consist of relatively well-sorted angular pumice clumps of sub-centimetric to centimetric sizes in decimetric thick beds separated by unsorted finer-grained layers of similar thickness. As many as three pumice layers have been encountered in these outcrops.

Fallout tuffs are rare because of their high potential to be removed unless quickly buried. Within the LTC one layer of coarse-grained fallout tuff was found at Novaj (cellars) in which laminae of white rhyolite pumice and dark rhyolite obsidian fragments alternate. An andesitic tuff layer has been found in a MTC sequence in a hillside at Sály–Latorvár.

Phreatomagmatic tuffs characteristically occur at several localities. Their airfall origin is questionable but at least the unsorted interbeds in the pumice fall sequences of the ULTC can be interpreted in this way. The accretionary lapilli bearing tuffs in the MTC sequences are common. They are poorly sorted, decimetric in thickness, but are associated with debris flow deposits, and hence their fallout origin is hypothetical. An alternative interpretation is that they are of surge origin, or they are secondary due to reworking of primary fallout phreatomagmatic tephra containing accretionary lapilli by mass-flow processes.

Reworked tuffs are present in all three tuff complexes. They form thick (meters) massive, unsorted, chaotic beds, in places rich in lithic clasts. Soil and vegetal fragments can be seen in them in places. These features strongly suggest their origin in debris flow processes through which part of the primary tephra was reworked shortly after deposition.

Volcanic epiclastic deposits are also present in several outcrops. In the Eger–Tihamér quarry toward the top of the LTC a thick sequence of cross-bedded coarse gravel, pebble and sand crops out displaying beautiful intraformational channels.

3.1.2. *Eruptive style and sequences*

The prevalence of ignimbrites in all three tuff complexes, their widespread occurrence, the presence of welding, the rhyolitic composition of most of the juvenile material, and the lack of effusive products all strongly suggest that the volcanoclastic deposits in the Bükk Foreland area resulted from repeated highly explosive eruptive events spanned over a ca. 7.5 Ma time period (from 21 to 13.5 Ma ago; Márton and Pécskay 1998, this volume) and clustered in three major pulses. The volume of erupted tephra is extremely difficult to assess because a significant part of the dispersal area is now buried under younger sediments, but the observable part of the deposits suggests that large-volume ash-flow eruptions were responsible for the emplacement of the studied rock sequences. Their source volcanoes are most probably of caldera type. Each ignimbrite unit is the result of a major individual eruption, but repeated eruptions from the same magma chamber, hence from the same volcano or volcano cluster, can be inferred from the lithostratigraphy of the deposits as presented above.

The LTC was emplaced during two major explosive eruptions, possibly from the same shallow rhyolitic magma chamber. Even though Plinian fall deposits were preserved for merely one of them it is likely that both eruptions started with a Plinian phase, later turning to the ash-flow phase. The Plinian phase of the ULTC was punctuated by phreatomagmatic episodes due to repeating access of external water to the magma conduit. The climactic ash-flow phase was purely magmatic but phreatomagmatic activity resumed as the eruptive energy vanished. This scenario seems to be valid for the other eruptions of the LTC and MTC as well. It means the access of water to the conduit was not accidental, providing a hint for emplacement of the source vents in near-water or shallow water environments.

The eruptions giving rise to the MTC were peculiar in that they bear the evidence of mixing of acidic and more basic magma in the magma chamber. The presence of a whole range of petrochemistry from rhyolite to basaltic andesite strongly suggest that the eruption originated from a zoned magma chamber. The succession of acidic to more basic compositions upwards in the sequence indicates that the more differentiated top of the chamber containing rhyolitic magma erupted first and, as the eruption progressed, deeper levels of less differentiated magma were successively tapped. The final product of the UMTC is almost purely andesitic. Co-existence of acidic (rhyolitic) and more basic (andesitic) magmas in a magma chamber is documented in many parts of the world. The presence of at least two ignimbrites in the sequence of the MTC suggests that more than one eruption from the same-zoned magma chamber may have occurred. The andesitic fallout tuff at Sály–Latorvár, in the area of occurrence of the most basic (andesitic) ignimbrite, suggests that this was a separate phase, if not a separate eruption during emplacement of the UMTC. Phreatomagmatic activity at the end of eruption pulses is a result of ready access of water to the conduit as the magmatic phase vanished, suggesting the availability of abundant water at the eruption sites.

The style and episodicity of volcanism for the UTC is poorly constrained in the Bükk Foreland area due to the scarcity of the outcrops. It is likely, however, that UTC deposits in the western and the eastern parts of the area resulted from different eruptions, or from different source volcanoes.

The lithologic, petrographic, petrochemical, paleomagnetic features and K-Ar ages of the three tuff complexes are summarized in Fig. 3.

3.2. Volcanic source location

The source areas of old, pre-Quaternary ash-flow sheets are commonly destroyed by erosion or hidden beneath younger sediments and thus extremely difficult to locate. This task is even more difficult in the Bükk Foreland where only a small part of the area covered by the acidic explosive products is available to direct observation. No convincing evidence concerning the possible presence of eruptive centers has been found in the area where the tuffs crop out throughout the Bükk Foreland. Previous speculations on source location (Varga 1981) presumed possible eruptive centers as being located inside the Pannonian Basin, somewhere south of the study area.

The presence of moderate to intense welding in the ignimbrites belonging to both the LTC and MTC, as well as some grain-size parameters of the same ignimbrites (e.g. maximum size of dense obsidian clasts up to 50 cm, or similar maximum pumice sizes) strongly suggest that the source of these rocks should not be very far away in the basin, south of the Bükk Foreland. According to transport distances of similar grain-sized ash-flow tephra from their source, as known in modern ignimbrite fields, the expected distance of the eruptive centers from the southeastern margin of the outcrop area of the tuffs should be in the range of kilometers to a few tens of kilometers.

Taking into account the acidic composition of the pyroclastics as well as the eruptive style recorded in these deposits, one may expect large calderas, with possible resurgent intrusive or effusive activity in their interiors, as centers of volcanic activity to which the large-volume ignimbrites are related.

Starting from the above presumptions, the methodology we adopted to locate the eruptive centers involved two steps. First, we attempted to identify indicators of transport directions recorded in the outcropping pyroclastic rocks; statistically, they should converge in areas at or near the source areas of the tephra. Second, we searched for evidence and arguments in the available geophysical and borehole record, that may support the possible presence of eruptive centers in such areas of tephra transport direction convergence.

3.2.1. *Lithological indicators of source location*

Thickness and grain-size of fallout tephra

Pumice lapilli fall deposits crop out at several localities at the base of ignimbrites belonging to the LTC. In four locations three Plinian pumice lapilli fall layers were measured concerning their thicknesses and maximum pumice sizes. The same parameters were measured at one additional outcrop (Novaj) where one coarse-grained fallout tuff layer is exposed (Fig. 5a–5b). Even though the measured values are too few to construct isopach or isopleth contour lines, a tephra dispersal axis can roughly be estimated from these summary data. Perhaps not surprisingly the dispersal axes obtained by both thickness and grain-size data coincide, suggesting a northeast dispersal of the fallout tephra during the Plinian phase of the ULTC event.

Imbrication and bomb sag asymmetry

Imbrication of elongated pumice and/or lithic clasts frequently occur near the base of non-welded ignimbrites. They record the local transport direction of tephra immediately prior to deposition. Such features have been encountered in three outcrops belonging to the UMTC (Bogács, Cserépváralja and Kisgyőr).

Asymmetrical bomb sags left behind by oblique impact of coarse dense clasts in soft wet ash during phreatomagmatic eruptions can be used as indicators of direction of aerial transport of tephra from the source to the deposition site. We have found one asymmetrical bomb sag in a phreatomagmatic sequence in an outcrop at Cserépváralja belonging to the UMTC, and measured the azimuth of its axis.

Figure 6a synthesizes all these data indicating transport directions of tephra during the Plinian phase (imbrication) and phreatomagmatic phase (bomb sag asymmetry) of the UMTC eruption. One can observe that the backtraces of these directions roughly converge towards an area located northeast of Mezőkövesd in the Pannonian Basin not far from the Bükk Foreland area (Fig. 6a). The backtrace of one imbrication indicates a slightly different transport direction, as compared with the other three source direction indicators, but still suggests the source location in the same SE quadrant of the sketch map. Local transport direction can be influenced by local topography by channeling of the ash flow.

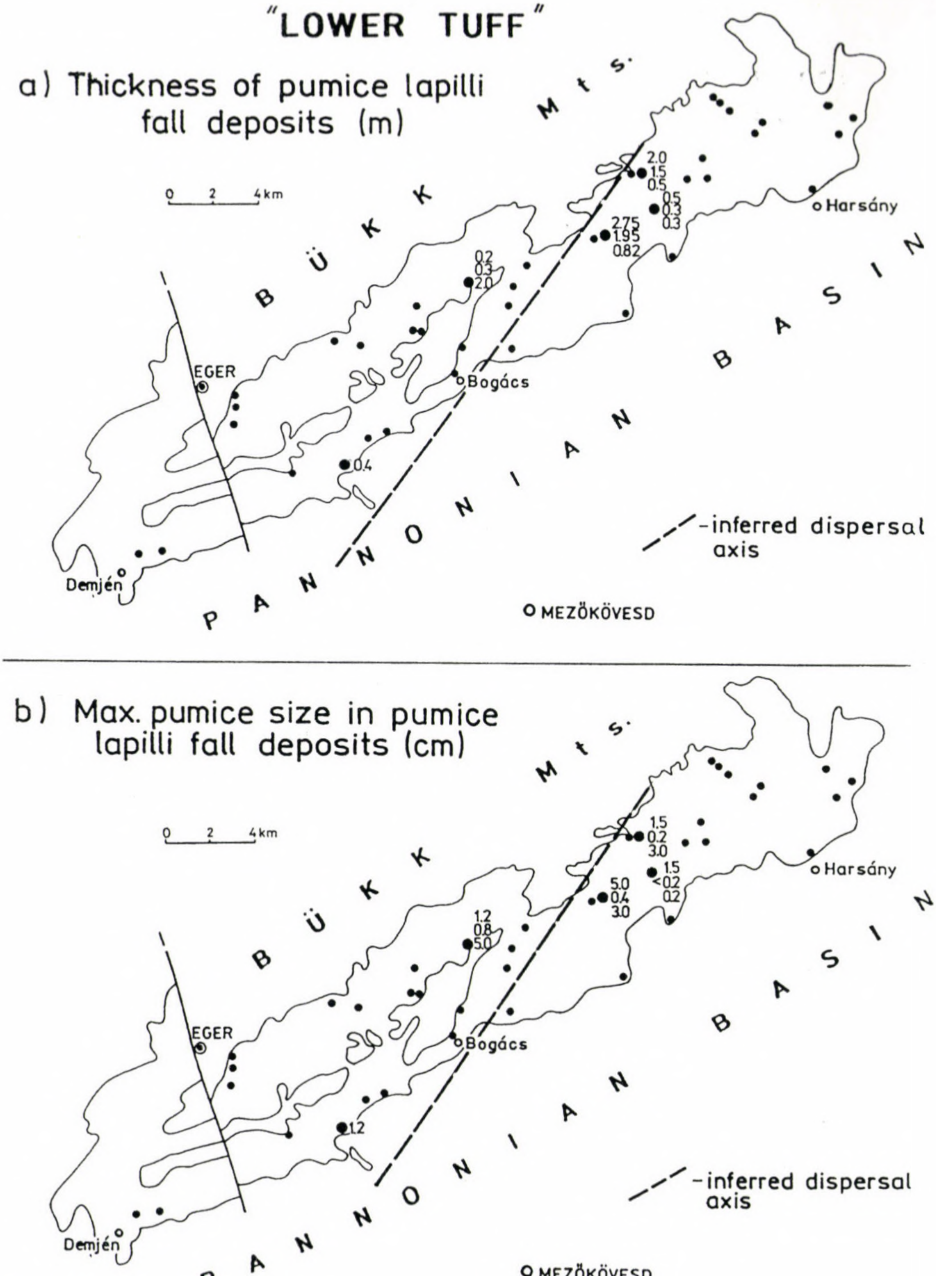


Fig. 5
Sketch showing transport direction of fallout tephra as inferred from lithologic features of the Lower Tuff Complex

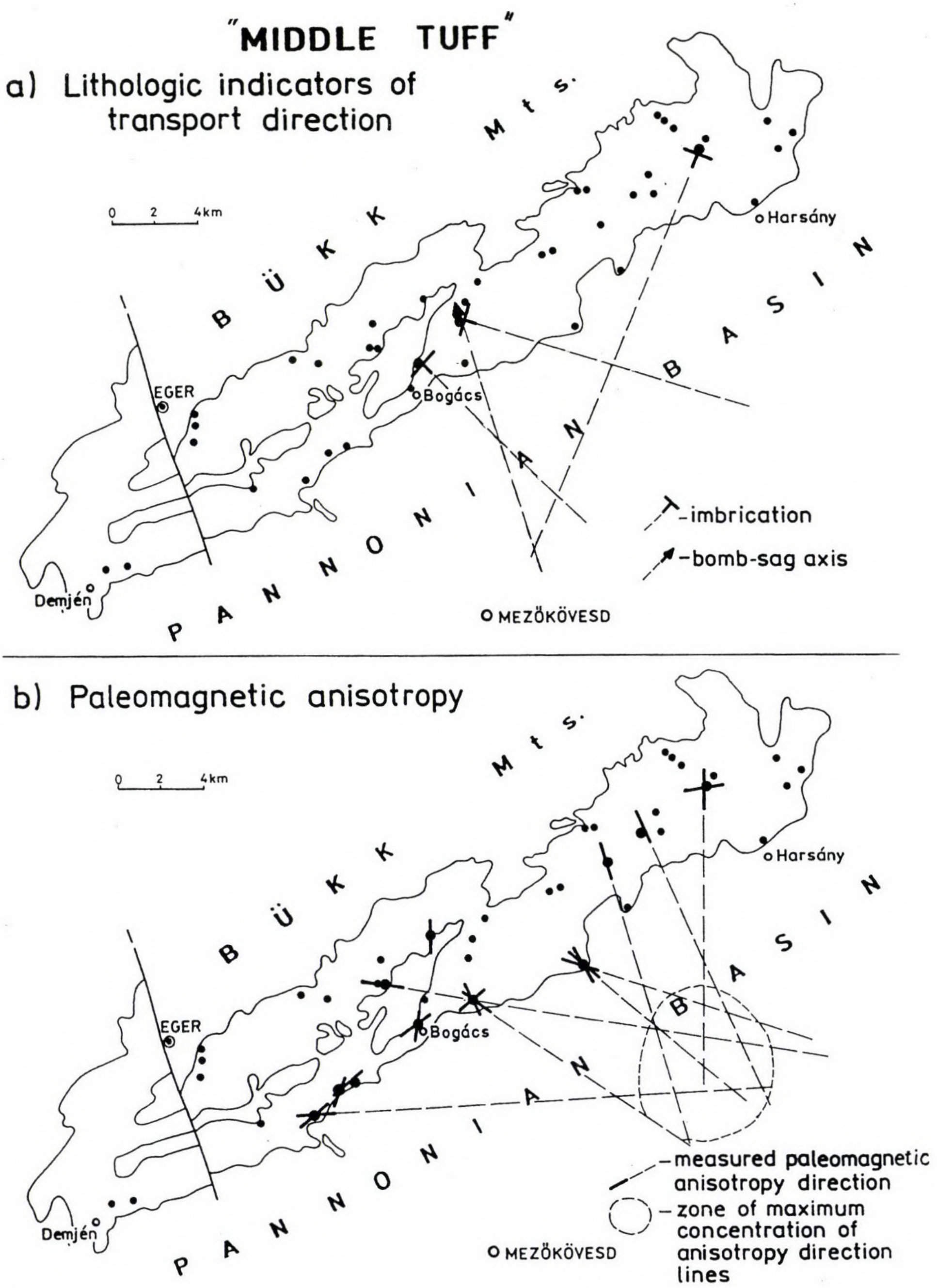


Fig. 6
Sketch showing transport directions of ash-flow tephra and phreatomagmatic deposits as inferred from lithologic (a) and paleomagnetic (b) features of the Middle Tuff Complex

3.2.2. Paleomagnetic indicators of source location
Paleomagnetic anisotropy

Paleomagnetic anisotropy can be successfully used to infer transport direction of ash flow tephra hot enough to stay above the Curie point of their magnetic minerals. Magnetic susceptibility anisotropy was studied at all paleomagnetic sites (Márton and Pécskay 1998, this volume). Without exception the magnetic fabric is dominantly foliated reflecting the key role of compaction in the formation of the texture. The degree of foliation is variable, between 1 and 6 percent. Magnetic lineation is subordinate (up to 2 percent, often between 0.5–0.8 percent). At site level the grouping of the maxima and the intermediate directions within the plane of foliation is often observed, but it is only in the welded varieties of the MTC where the maxima are clearly flow-directed (Bordás 1992). The direction of the maxima in such cases may be used as a proxy for transport direction and eventually locate the center of eruption. The results are plotted in the sketch map (Fig. 6b). Two groups of directions are observed. One group is constantly SW–NE oriented, roughly along the main structural trend of the Bükk Foreland. The other group displays a remarkable convergence of their backtraces in a zone located northeast of Mezőkövesd in the Pannonian Basin (Fig. 6b), which could represent a possible source area for the UMTC pyroclastics.

3.2.3. Borehole data
Thickness of tuff horizons

Borehole data reviewed in the southeastern part of the Bükk Foreland area (Fig. 2) indicate an obvious thickening of the volcanoclastics of the LTC southward in an area located south of Eger in the Pannonian Basin (Fig. 7). Thickening is observed, from 250 to 750 m along a length of merely several km. It is well known that ignimbrites, and more generally volcanic products, display an overall thickening in a sourceward direction.

Lavas

Rocks interpretable as lava flows in origin have not been encountered by us in the Bükk Foreland area. However, andesite lavas were credibly described (unpublished data from the Archives of the Geological Survey of Hungary) in a borehole (Mn-1) northeast of Mezőkövesd in the Pannonian Basin. Their thickness (74 m) strongly suggests a near-source location of that borehole.

3.2.4. Geophysical data

Searching for possible source areas of the Bükk Foreland volcanics in the nearby parts of the basin, we reviewed available geophysical anomaly data in the Archives of the Geological Survey of Hungary. Thick rhyolitic ignimbrites ponded in caldera depressions and/or possible shallow subcaldera intrusions

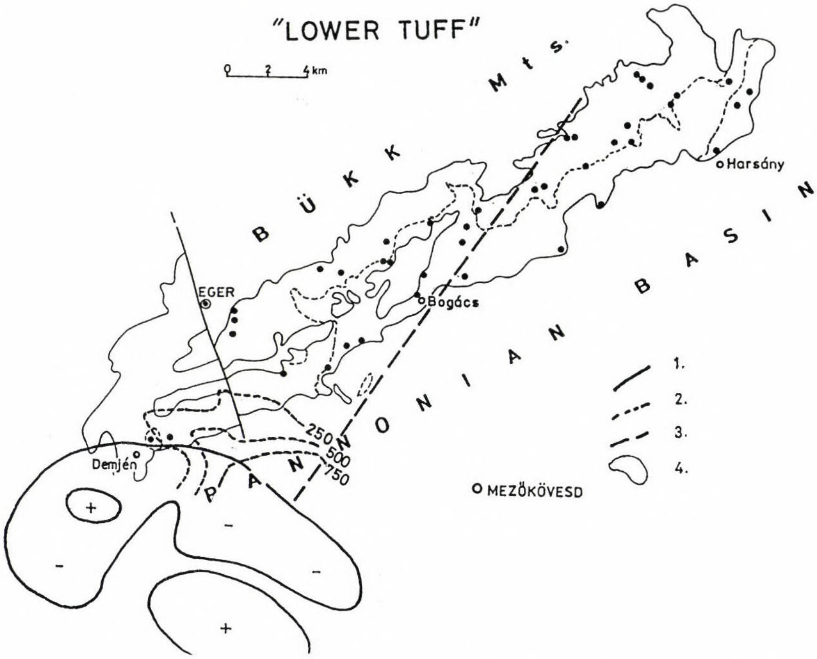


Fig. 7
Sketch showing possible source location for the Lower Tuff Complex. 1. positive (+) and negative (–) geomagnetic anomalies; 2. thickness of the Lower Tuff Complex in drillings (m); 3. inferred dispersal axis of fallout tephra; 4. inferred source location (burried caldera ?)

or resurgent domes of rhyolitic composition are often expressed through negative gravity anomalies as well as strong magnetic anomalies. On the other hand more basic lavas or intrusions in intracaldera/subcaldera settings may be reflected in positive gravity anomalies coupled with magnetic anomalies.

Magnetic anomalies

A positive ellipse-shaped anomaly (ca. 8 x 4 km in size) of the vertical component of the magnetic field (Z) surrounded in the north by a much larger crescent-shaped negative anomaly is present south of Demjén in the southeastern extension of the study area (Fig. 7) (Polhammer and Szilárd 1965; Gulyás et al. 1994). The intensity of the negative anomaly is –7.5 nT, while the surrounding positive anomaly is +25 nT in strength. The anomaly pair may reflect the magnetic effect of a large, ellipse-shaped rock body containing abundant magnetic minerals.

Strong aeromagnetic anomalies, detected within the Bükk Foreland area, closely follow the outcropping welded MTC ignimbrites, especially their more

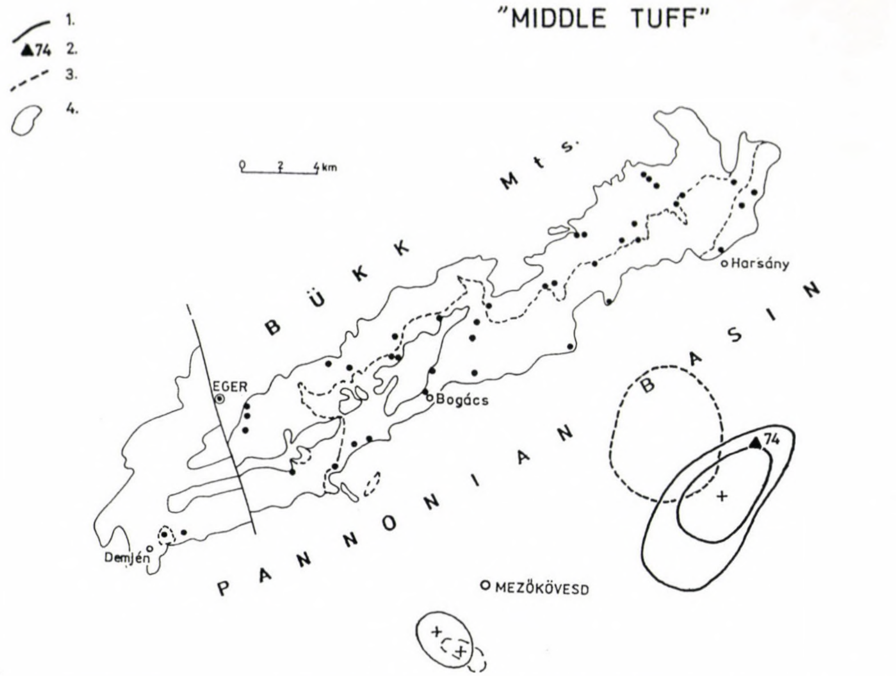


Fig. 8
Sketch showing possible source location for the Middle Tuff Complex. 1. positive (+) gravity anomaly; 2. andesite lava occurrence in drilling (thickness in meters); 3. concentration zone of backtraces of magnetic anisotropy lines (Fig. 6b); 4. inferred source location (burried caldera ?)

magnetite-rich andesitic varieties, hence they can be considered as the magnetic expression of the presence of these rocks at the surface (Fig. 8).

Gravity anomalies

Two significant gravity anomalies are recorded immediately south of the Bükk Foreland area. These positive anomalies are located west and east of Mezőkövesd in the Pannonian Basin (Haáz and Komáromy 1965). The eastern one has a NE–SW-oriented ellipse shape ca. 10 x 4 km in size (Fig. 8). It may reflect a dense rock body of similar shape buried beneath the younger basin-filling sediments.

3.2.5. Summary of source location indicators and discussion

The data presented above concerning possible source location indicators reveal two zones of interest in which most of these features converge (Figs 7 and 8). Both of them are located in the Pannonian Basin south of the Bükk Foreland area and relatively close to the outcrop area of the Miocene volcanoclastic rocks: (1) south of Demjén in the western part, and (2) northeast of Mezokövesd in the eastern part. They may represent the buried source calderas (or caldera clusters) from which the LTC and the MTC originated.

The western area can eventually be considered as the source for the LTC. This hypothesis is supported by the following arguments: (1) the presence of a positive (Z) aero-magnetic anomaly whose shape and size is fully consistent with that of a normal rhyolite ignimbrite-generating caldera; (2) the inferred dispersal axis of the LTC fallout tephra layers points quite precisely to the magnetometric anomaly; (3) the thickness of the LTC rocks, as revealed by borehole data, increases towards the aeromagnetic anomaly area; (4) the spatial distribution of the welded LTC ignimbrites in the western outcrop area suggests a roughly N–NW-oriented flow-unit lobe (Fig. 1) the extension of which in the Pannonian Basin coincides with the aeromagnetic anomaly; the spatial distribution of the other welded ignimbrite occurrences (Fig. 1) is compatible with the presence of a second, much longer, channeled ash-flow lobe, oriented NE, originating from the same source area. The lack of a measurable gravity anomaly coupled to the aeromagnetic anomaly can be explained by the possible lack of a significant density contrast between the caldera fill and host rocks.

The source area of the MTC, a possible second caldera or caldera cluster, may be located NE of Mezőkövesd in the basin. Such a hypothesis is consistent with (1) the presence of an elliptic positive gravity anomaly of comparable size with that of an ignimbrite-generating caldera, (2) the convergence in roughly the same area of the backtraces of local transport directions inferred from paleomagnetic anisotropy, (3) the convergence toward a nearby area (in the same quadrant of the sketch map) of the transport directions inferred from lithologic indicators, (4) the presence of thick andesitic lava flows in the area of the gravity anomaly, consistent with a possible post-caldera fill of andesite lava flows or domes, and (5) the maximum sizes (up to 45 cm in length) of the obsidian fiamme measured at Tibolddaróc, the closest outcrop locality to the inferred source. The lack of expected aeromagnetic anomalies in the area is, however, problematic, especially when compared with the other inferred caldera, located south of Demjén. It may be explained by the pervasive alteration of the magnetic support minerals in subaquatic environments.

The source area of the UTC cannot be discussed here because of the limited occurrence of the complex and the very few outcrops available in the study area. Because of the lack of connectivity between the western and eastern occurrence areas, as well as possible age differences (see the companion paper of Márton and Pécskay 1998, this volume), it may be assumed that despite their rough petrographic and chemical similarities they may have originated from different source areas, located outside the study region.

4. *Conclusions*

Complex geologic, geochemical, geophysical and geochronological investigations were carried out in the Bükk Foreland area. Our investigations basically confirm the occurrence and general distribution of welded and non-welded ignimbrites in the area, belonging to two main pyroclastic sequences (the Lower Ignimbrite and the Upper Ignimbrite) as mapped by Balogh (1964). However, the rocks mapped as dacitic lava flows actually represent strongly welded ignimbrites of the Middle Tuff Complex, the dacitic composition resulting from mixing of a rhyolitic component and an andesitic one. Based on the newly obtained data we are now able to distinguish between the different tuff sequences, including the Upper Tuff Complex, and correlate them across the area, according to their ages and paleomagnetic characteristics, in addition to their petrographic and petrochemical features. This complex approach may serve for further correlating of the Miocene tuff sequences of the Pannonian Basin over broader areas, irrespective of their emplacement (marine or terrestrial environments). Each tuff complex can be characterized by a unique set of features, including lithology, petrography, petrochemistry, paleomagnetic parameters and K-Ar age. The three tuff complexes have a composite internal lithology, reflecting a succession of eruptive and reworking episodes of tephra.

The Lower Tuff Complex (LTC), Eggenburgian-Ottnangian in age, displaying a ca. 90° counterclockwise paleomagnetic rotation, consists of two biotite and quartz-rich high-silica rhyolite ignimbrites, three Plinian pumice lapilli beds separated by phreatomagmatic tuffs at the base of the second ignimbrite unit, and a sequence of phreatomagmatic tuffs, debris flow deposits, and epiclastics. The lower unit of the LTC is non-welded while the upper one displays both welded and non-welded facies.

The Middle Tuff Complex (MTC) is Badenian in age, rotated ca. 30° counterclockwise, and characterized by the presence of two petrographic components (rhyolitic and dacitic-andesitic) the mixing of which gave rise to a whole range of lithologies from rhyolitic to basaltic andesitic with the ubiquitous presence of pyroxenes in each rock type. Mixed pumice-scoria ignimbrites containing banded pumice are common. At least two ignimbrite units, the upper one partially welded, resulted from eruptions tapping a zoned magma chamber. Phreatomagmatic deposits with accretionary lapilli and associated debris flow deposits are characteristically present as well. The boundary between the LTC and MTC is locally marked by a gravel horizon or by a coal sequence.

The Upper Tuff Complex (UTC) occurs as small patches at the western and eastern peripheries of the area. It consists of biotite-rich non-welded rhyolitic ignimbrites and some phreatomagmatic pyroclastics and reworked tephra. These deposits are Sarmatian in age and do not show any paleomagnetic rotation.

The volcanological interpretation of the data indicates that the tuff complexes resulted from voluminous ash-flow eruptions preceded, at least in the case of the LTC, by a Plinian phase, with an intervening phreatomagmatic phase toward the end of the activity as eruptive energy vanished, followed by immediate reworking of the resulting loose tephra. This common scenario for at least five individual eruptions suggests a similar paleoenvironment in which the eruptions occurred from shallow magma chambers. In at least one case (MTC) the ignimbrite eruption tapped a zoned magma chamber containing rhyolitic magma in the top and dacitic to andesitic magma at deeper levels.

The type and volumes of eruptions suggest that the source volcanoes may have been of large, acidic caldera type. Using various indicators of source location, lithologic (thickness and maximum pumice size of Plinian pumice fall deposits, imbrication in non-welded ignimbrites, bomb sag asymmetry in phreatomagmatic fall deposits) and paleomagnetic anisotropy as well as borehole and geophysical data (gravity and aeromagnetic anomalies), two possible source areas have been inferred: (1) south of Demjén in the western part of the study area, from which the LTC pyroclastics may have been erupted, and (2) northeast of Mezokövesd in the eastern part of the region, from which the MTC rocks may have originated. Both can be calderas or caldera clusters buried beneath younger sediments.

Acknowledgements

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Petrology and geochemistry of the Miocene acidic explosive volcanism of the Bükk Foreland; Pannonian Basin, Hungary

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The Miocene acidic explosive sequences in the Bükk Foreland (Northern Pannonian Basin) were studied by optical microscopic, microprobe, petrochemical, trace element and REE analyses, as an integrated part of a K/Ar age and paleomagnetic, stratigraphic–volcanological, and petrological–geochemical team effort (Márton and Pécskay; Szakács et al.; this volume).

Results suggest that the three acidic explosive complexes of the Bükk Foreland are akin to the Lower, Middle, and Upper Tuff Horizons developed elsewhere in the Pannonian Basin. The individual tuff complexes can be distinguished from one another on the basis of their petrological–geochemical characters.

The investigated tuff complexes derive from the melting of an upper crust of granitic or acidic metasediment composition.

According to the petrological and geochemical data the Middle (Dacite) Tuff Complex was generated by the mixing of an acidic magma and an intermediary (andesitic) magma.

The Upper Tuff Complex is found only in two small patches. These two occurrences originated from different volcanic centers outside of the studied area.

Key words: Bükk Foreland, acidic explosive volcanism, Miocene Tuff Horizons in the Pannonian Basin, magma mixing, petrology and geochemistry of ignimbrites

Introduction

It is characteristic of the Miocene calc-alkaline volcanism of the Pannonian Basin that acidic explosive volcanic products of several hundred metre thickness accumulated. Tuff horizons covered by young sediments were preserved in

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especially thick levels in the molasse basins adjoining the Pannonian Basin and in its ultra-deep sub-basins. The relatively continous horizons had been already recognized at the beginning of our century and were used as guide horizons in the stratigraphic classification (Noszky 1933).

With the intensification of borehole exploration, the extension of tuff horizons within the basin has became more clear. Having studied the borehole material and in the course of the reambulatory analysis of the Tokaj Hills, the magma-genetic and tectonic significance of the acidic explosive sequences and numerous types of acid tuffs and ignimbrites were described and classification was performed as well (Pantó G. 1962, 1963, 1965.)

In 1963 G. Pantó had already recognized that some of the tuff horizons are at least bipartite, i.e., in case of complete formations these consist of a loose unwelded and a strongly welded tuff horizon (ignimbrite).

K/Ar age determination of the tuff horizons of the Pannonian Basin started in the seventies and in the course of these investigations both the chronostratigraphy and the tectonic relations of the tuff horizons were defined as follows (Hámor et al. 1980):

I. Lower Rhyolite Tuff Horizon (Eggenburgian/Ottnangian)

II. Middle (Rhyo)dacite Tuff Horizon (Karpatian/Lower Badenian)

III. Upper Rhyolite Tuff Horizon (Lower Sarmatian).

Varga (1962) called the attention that in the Mátra Mts and in its environs the Middle Tuff Horizon is not rhyolitic, but of rhyodacitic-dacitic composition.

Having collected the analytical data of sporadically described tuff occurrences in the mountains and basins, Póka made a first attempt (1988) to compile a petrochemical evaluation on the acidic tuff sequences of the Carpathian Basin.

Using borehole data and geologic maps Pantó G. (1965) and Ravasz (1987) prepared the first general rhyolite tuff horizon distribution map of the Pannonian Basin.

The Bükk Foreland is the most significant occurrence area on the surface of the acidic explosive volcanic rocks. Here the Lower and Middle Tuff Horizons are found often in one exposure and as to our latest results the Upper Tuff horizon can be found as well. The formation, belived also formerly as Miocene volcanic sequence, was documented by Balogh (1964) in his monograph on the Bükk Mts in the geological map at 1:100 000 scale. He described these formations as tuffs and ignimbrites and made the distinction between the "Lower and Upper ignimbrite Horizons". A report in manuscript and sampling geologic map are found recently in the data bank of the Hungarian Geological Survey, prepared by Varga and Szenthe (1976) and Varga (1981).

The chemical composition of some biotite of the ignimbrites and tuffs was measured by microprobe (Dobosi 1980).

Since the stratigraphic classification of the terrestrial volcanic sequence (see Szakács et al. this volume) was impossible, paleomagnetic measurements were made to make clear the uncertain age relations of the Bükk Foreland sequence (Márton E. 1990). A counter-clockwise rotation of 80–90^{o} and 60^{o} in the Lower

Ignimbrites, and 30° in the Middle Ignimbrite Horizons, was defined. These data allowed to perform exact stratigraphic–volcanological research.

Capaccioni et al. (1995) presented a preliminary volcanological report (the paper was concerned only field observations). They stated that the Lower Ignimbrite is nonwelded tuff, while the Upper one is welded ignimbrite.

In the frame of a Hungarian National Scientific Research Foundation competition (1995) the stratigraphic–volcanological (Szakács et al. 1997 and this volume), K/Ar age and paleomagnetic measurements (Márton E. and Pécskay 1997 and this volume) and petrological–geochemical investigation of the rocks (Póka et al. 1997) commenced in teamwork.

The petrological-geochemical investigations aimed at

– the determination of the petrological–geochemical character of the three tuff sequences of the Bükk Foreland; this may provide data not only for these rocks but also for the chronology of tuff horizons of uncertain stratigraphic classification within the Pannonian Basin as well;

– the contribution to evaluate of some contradicting K/Ar and paleomagnetic measurement data;

– the comparison of the Bükk Foreland ignimbritic explosive sequence with other similar volcanics;

– the contribution to solve the petrogenetic and geodynamic problems of the Miocene calc-alkali magmatism of the Carpathian Basin.

Methods

Rock samples were qualified under optical microscope (see the Appendix).

The chemical compositions of biotites occurring in all the three tuff complexes and of the pyroxenes occurring only in the middle horizon were determined by microprobe analyses in the Laboratory for Geochemical Research of the Hungarian Academy of Sciences (Table 3). The microprobe analysis of groundmass glass and pumice clasts was carried out in Birbeck College, London.

In the Laboratory for Geochemical Research of the Hungarian Academy of Sciences measurements were carried out on the JXA-733 type JEOL microprobe, quantitative analyses were made by WDS (analytical conditions: 15 keV, 40 nA electron current, 5 x 4 s counting time; standards: to Si, Al and K orthoclase, to Fe, Mg and Ca (synthetic) glass, to Ti oxide, to Mn spessartite and to Na albite; to avoid the matrix effect the ZAF-correction was used.

Major elements (Table 1) were determined after lithium-metaborate leaching as follows: silica by gravimetry, FeO by titrimetry with $KMnO_4$, the other elements by spectrophotometry (Perkin Elmer 5000).

Trace elements and the REE were measured by nondestructive NAA in the Nuclear Research Institute of the Central Institute for Physics (analytical conditions: irradiation: 3 minutes and 24 hours in the pneumatic dispatch and traditional vertical irradiation channels).

Measurement: 36% 1.75 keV Canberra HPGe detector + 2x8K Dual LFC gamma spectrometer with ACCUSPEC/B 16K analysator; Evaluation: Hypermet-PC V 50 gamma spectroscopic and KAYZERO/SOLCOI V 4a program package for calculation of elemental concentrations.

Results and discussion

The samples studied in laboratory are listed in the Appendix. The localities of these samples are shown in Fig. 1. In the tables and figures the samples are of the same numbers and the same numbers represent the samples in the papers of Szakács et al. (this volume) and of Márton E. and Pécskay (this volume).

The stratigraphy and volcanology of the studied area described in the publication of Szakács et al. (this volume).

Lower Tuff Complex

Petrography

In this complex pumice tuffs and lithoclastic, often pumice-bearing, more or less welded ignimbrites, are characteristic (Photo 1, 2); in the upper horizons often phreatomagmatic and redeposited tuffs are found. In the lowermost horizon of the complex, in the western and southern parts of the area, in addition to dacite lithoclasts remarkable quantities of hypersthene andesite lithoclasts occur, in the northernmost areas with Cretaceous alkali-diabase lithoclasts from the Bükk Mountains. In the upper horizons the quantity of rhyolite lithoclasts exceeds that of the dacite lithoclasts. In the lowermost horizon of the Eger–Wind brickyard factory among the dacite-andesite lithoclasts specially coarse porphyric amphibole-dacite was described (with microcline!).This lithoclast type could not be identified elsewhere.

Rock-forming minerals

In the whole Lower Tuff Complex biotite is the main mafic component (it is often opacitized or chloritized). In the welded varieties green amphibole is rarely found in addition to biotite.

In this horizon the FeO content of biotite is 23–25% in addition to the MgO content of 8–9% (this relates to medium-differentiated calc-alkali acidic dacites). Dobosi (1980) studied the biotite compositions of four samples of rhyolite tuffs from the Kisgyőr and Cserépfalu environs and measured also 25–29% total FeO and 6–9% MgO. He mentioned that this Fe/Mg ratio relates to the magma melt of relatively low temperature but of high water fugacity.

In the Lower Complex the plagioclases are predominantly of oligoclase, subordinately of oligoclase-andesine composition (determinated by optics) and are often zoned. Magnetite, hematite, zircon are the accessories; in biotites zircon and apatite inclusions are frequent. Chloritization of biotites and

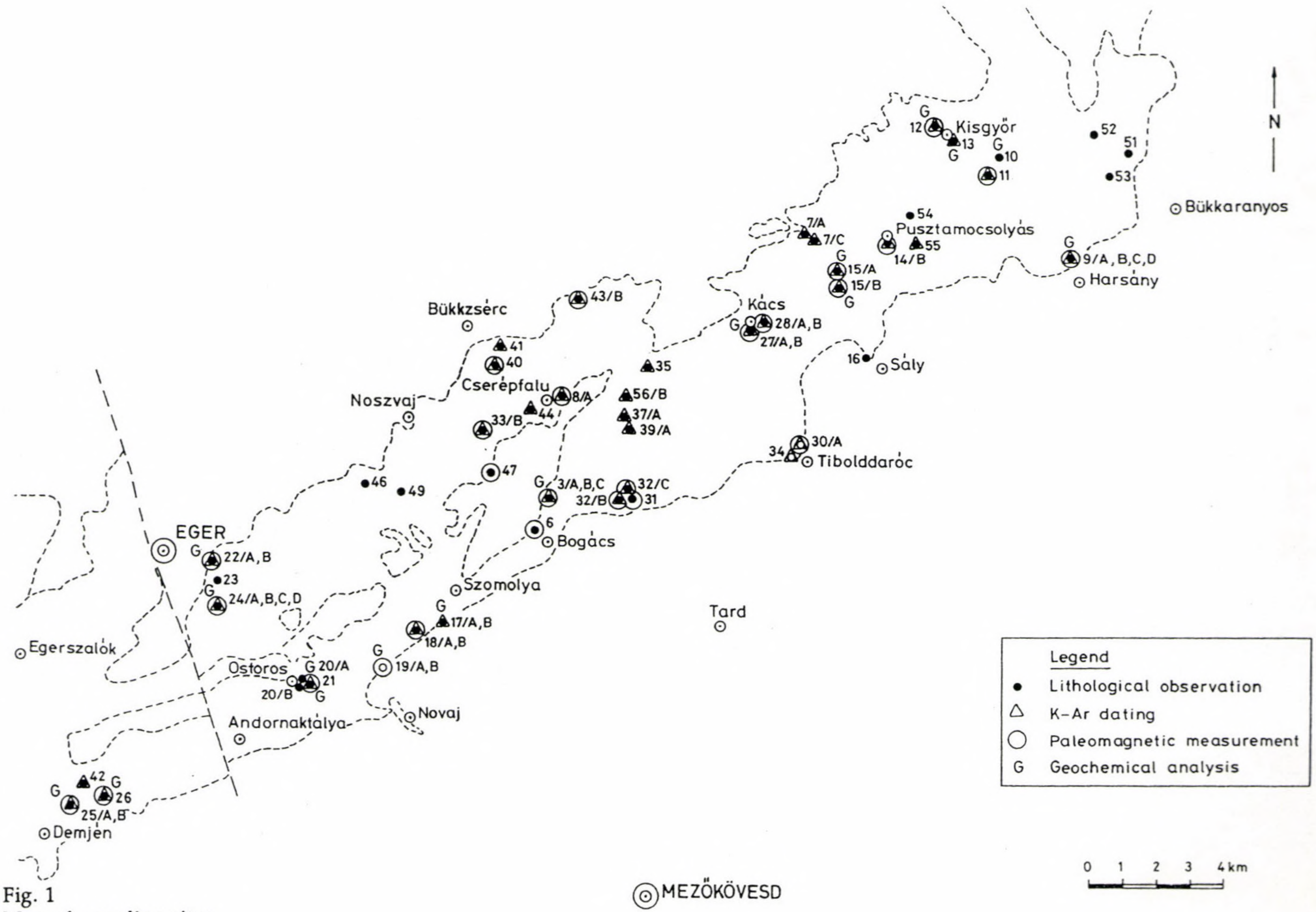

Fig. 1
Map of sampling sites

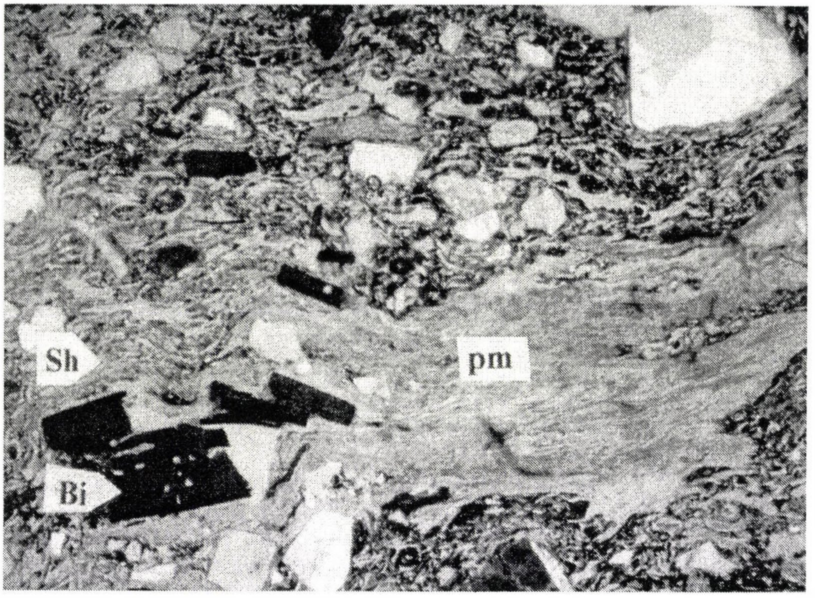

Photo 1.
Lower Tuff. Moderately welded pumice-rich ignimbrite with plastically deformed pumice fiamme (pm) and oriented glass shards (Sh) (Bi = biotite). Demjén, Pünkösdhegy. Q > Bi > Am (II N)

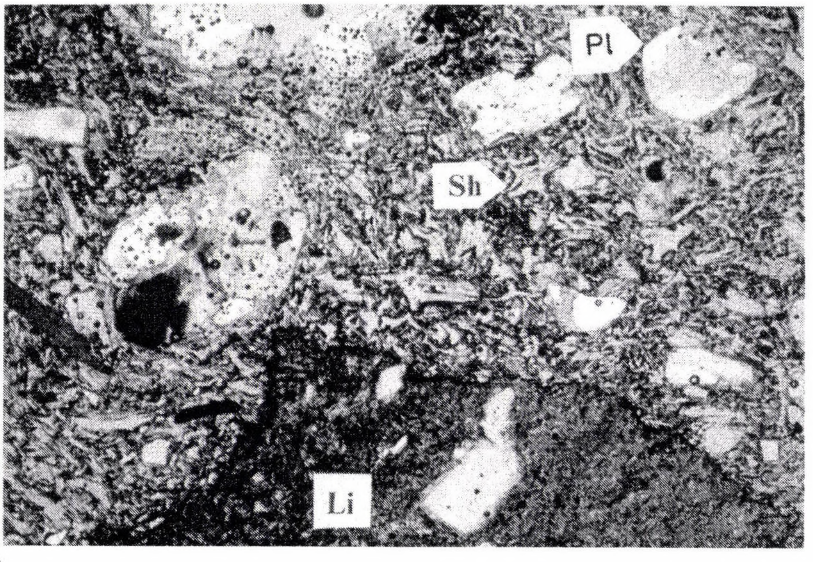

Photo 2.
Lower Tuff. Slightly welded ignimbrite with mostly randomly oriented glass shards (Sh), with phyllite litoclasts (Li) (Pl = zoned plagioclase). Kács, Templomdomb (II N)

kaolinitic decomposition of feldspars are frequent phenomena. Zeolitization of the groundmass and of the pumice is remarkable in some layers.

Pumice and vitroclasts

The composition determinated by microprobe of the pumice and groundmass glassy clasts is potassium-rich rhyolitic; in the groundmass the Fe-content may be as high as 1% (Table 6, Fig. 10).

Petrochemistry

The chemistry of the rocks of the Lower Tuff Complex is high-potassic rhyolitic (calculated for water-free values); nevertheless, some of the lithoclasts display normal and high-potassic dacitic and andesitic compositions (Table 1, Figs 2 and 3). The differentiation index is 88–90, that of the lithoclasts is 73–85; the color index is 1.3–3.3, that of the lithoclasts 6.1–10.6.

REE and incompatible trace elements

The distributions of REE and incompatible trace elements (Tables 2 and 3) were normalized to primitive mantle (Sun and McDonough 1989), to the granulite xenoliths of Hungarian Pliocene basalts (presumed lower crust), (Kempton et al. 1997), as well as to granitoids (Variscan granitoid, type II, Velence Hills) (Buda 1985).

In relation with the primitive mantle and lower crust (granulite) the REE distribution shows gradual transformation. In the Lower Tuff Complex the LREE are enriched while the HREE are reduced to smaller extent (except for the rhyolite lithoclast sample 22/B that displays an extreme positive HREE anomaly). Both of the normalized values indicated a considerable and well-defined Eu-anomaly (Figs 4 and 5).

As regards the HREE sample 22/B shows an extreme positive anomaly to the granitoids as well. As a whole the granitoids show similarity with the Lower Tuff Complex (Figs 6 and 8).

Some incompatible elements, i.e. U, La, Ce, Nd, Zr and Yb show characteristic positive anomaly in the primitive mantle normalized distributions while Ta, Sr, P and Ti are of negative anomalies (Fig. 7).

In case of the values normalized to the lower crust (granulite) Rb, K, Zr and Yb display extreme positive anomalies while depletion is shown in the Ba, Th, Ce, P and Ti quantities (Fig. 8).

The incompatible trace element values normalized to granitoids show close relations: in addition to the small positive anomalies of Ba and Eu only P displays a remarkable negative anomaly (Fig. 9).

Table 1/A
Major element concentrations (wt%)

	Lower Tuff Complex													
Sample No.	22A	22B	24A	24B	24C	24D	26	21	17A	17B	27A	28	12	13
SiO_2	74.39	82.51	76.68	66.80	63.19	69.42	73.54	73.60	62.33	74.93	72.77	74.63	72.68	73.69
TiO_2	0.21	0.24	0.11	0.65	0.82	0.53	0.22	0.16	0.70	0.18	0.20	0.15	0.19	0.18
Al_2O_3	14.70	10.51	12.94	16.23	17.76	15.65	14.38	14.36	17.58	13.62	14.52	14.65	15.08	14.15
Fe_2O_3	1.17	0.59	0.86	2.21	6.29	3.30	1.10	1,46	4.61	0.97	1.89	1.36	2.01	1.65
FeO	0.78	0.32	0.39	1.87	2.54	0.40	0.78	1.00	1.65	0.95	0.71	0.86	0.72	0.19
MnO	0.03	0.02	0.04	0.05	0.13	0.03	0.03	0.04	0.05	0.04	0.05	0.02	0.03	0.03
MgO	0.74	0.32	0.27	1.42	2.05	0.70	0.39	0.35	1.84	0.34	0.48	0.28	0.40	0.56
CaO	1.60	0.71	1.33	4.44	4.66	3.46	2.42	1.98	6.05	1.68	2.14	1.39	1.91	2.33
K_2O	4.11	3.40	5.44	3.44	1.54	3.40	4.29	4.97	2.68	5.35	4.86	5.13	4.70	4.69
Na_2O	2.20	1.35	1.89	2.76	0.94	2.98	2.80	2.25	2.40	1.87	2.34	1.51	2.18	2.46
P_2O_5	0.01	0.03	0.04	0.14	0.09	0.12	0.03	0.07	0.16	0.05	0.03	0.03	0.06	0.07
Σ	99.94	100.00	99.99	100.01	100.01	99.99	99.98	100.24	100.05	99.98	99.99	100.01	99.96	100.00
LOI	7.10	2.32	4.62	1.43	12.55	2.50	2.91	4.54	3.74	4.83	4.73	7.00	4.96	5.82
	CIPW norms													
Q	79.81	90.50	83.21	71.27	70.55	72.23	78.19	76.09	74.24	80.88	81.13	81.52	78.92	78.92
Or	4.98	3.28	6.78	6.91	1.72	5.04	6.19	7.72	2.90	7.02	3.12	6.12	6.42	6.62
Ab	4.17	2.00	3.48	5.65	1.51	6.73	6.05	5.24	4.04	3.69	4.24	2.72	4.49	5.29
An	3.36	1.18	8.85	11.82	8.95	8.55	5.78	5.10	10.89	3.69	4.24	2.72	5.01	5.43
C	4.17	1.09	1.90		6.58	1.12	0.94	2.47		2.09	4.24	4.76	3.08	1.05
Di														
Hy	2.08	0.73	0.83		5.39	2.38	1.34	1.31	4.77	0.97	1.33	0.79	0.12	1.85
Wo				0.30					0.31					
Mt	0.81		0.47	2.15	2.69		0.94	1.31	1.45	0.73	0.78	0.90	1.02	
Il	0.34	0.36	0.11	1.22	1.07	0.84	0.40	0.29	0.93	0.24	0.33	0.22	0.25	0.26
He		0.20		0.61	1.51	2.94	0.13	0.43	1.55	0.61	0.55	0.22	0.64	1.31
Ap														
Ru						0.14								
D.I.	88.96	95.78	93.47	83.86	73.78	84.00	90.43	89.05	81.21	91.59	88.49	90.36	89.83	90.87
C.I.	3.23	1.09	1.41	3.78	10.66	6.12	2.81	3.34	8.60	2.56	2.99	2.13	1.39	2.13

Table 1/B

Middle Tuff Complex																	
Sample No.	20A	19A	19B	18A	18B	2	3A	3B	3C	4	7C	8A	15A	15B	14A	14B	10
SiO_2	62.38	62.77	68.77	66.6	66.28	69.55	66.51	63.92	65.78	65.57	70.13	75.69	65.24	71.22	73.37	73.57	60.22
TiO_2	0.65	0.91	0.5	0.54	0.2	0.36	0.48	0.66	0.73	0.89	0.19	0.37	1.22	0.24	0.22	0.22	1.44
Al_2O_3	21.6	17.14	19.77	19.39	19.49	17.21	19.29	18.75	17.53	18.31	18.88	15.14	16.94	16.2	15.41	15.12	20.26
Fe_2O_3	3.23	6.82	2.1	2.95	2.13	2.45	2.32	1.6	3.2	2.71	2.04	1.63	3.76	0.75	1.58	1.77	1.87
FeO	2.09	1.46	1.2	1.45	1.63	0.89	1.21	2.07	1.93	2.98	0.65	1.03	1.8	1.37	0.38	0.43	5.27
MnO	0.05	0.07	0.03	0.04	0.05	0.03	0.04	0.04	0.05	0.07	0.02	0.04	0.05	0.03	0.02	0.03	0.13
MgO	1.58	1.06	0.71	0.8	0.94	0.79	1.01	1.	1.61	2.34	0.36	0.28	1.89	0.52	0.55	0.6	4.32
CaO	4.39	4.17	1.66	2.55	2.86	2.67	3.49	6.71	4.08	2.32	2.14	1.39	4.28	2.69	2.41	2.45	2.94
K_2O	2.24	2.59	3.69	3.16	3.47	3.88	3.41	2.55	2.91	2.62	3.52	3.38	2.03	4.00	3.58	4.34	1.76
Na_2O	1.53	2.71	1.84	2.24	2.35	2.1	1.97	2.38	1.89	2.01	1.98	0.97	2.49	2.9	2.35	2.02	1.32
P_2O_5	0.26	0.24	0.03	0.19	0.2	0.13	0.18	0.3	0.14	0.15	0.03	0.04	0.24	0.09	0.12	0.08	0.23
Σ	100.	99.94	100.3	99.9	99.6	100.06	99.91	99.98	99.85	99.97	99.94	99.96	99.94	100.01	99.99	100.63	99.76
LOI	9.73	6.53	8.33	6.46	6.01	7.29	6.81	4.47	6.19	6.57	8.38	8.91	6.19	3.16	4.7	6.17	6.38
CIPW norms																	
Q	66.38	64.82	65.21	71.08	68.98	72.96	70.06	64.15	68.21	67.32	82.67	83.17	64.89	71.81	79.4	79.48	69.41
Or	2.75	3.82	3.78	4.49	4.86	5.06	4.08	4.24	3.89	3.26	4.57	3.36	2.85	6.38	4.52	4.09	2.03
Ab	2.99	5.95	6.17	4.37	4.99	4.19	4.07	5.97	3.76	3.73	3.56	1.49	5.18	6.97	4.520	4.09	2.03
An	9.33	10.49	10.37	5.46	6.57	5.92	7.89	8.71	9.17	4.78	4.54	2.33	9.58	7.12	5.00	5.41	5.35
C	10.28	3.54	3.22	9.11	8.67	5.67	7.51		4.77	9.81	0.86	3.63	6.08	3.11	3.92	1.96	11.66
Di					0.66			0.15									
Hy	4.66	3.68	3.64	2.30	3.02	2.46	3.18	3.93	5.02	6.76	1.10	0.65	6.08	2.67	1.55	1.84	4.71
Wo																	
Mt	2.39	1.27	1.26	1.57	0.66	0.61	4.40	1.25	2.51	1.98	0.86	0.84	1.29	0.74	0.23	0.36	1.28
Il	0.95	1.56	1.54	0.85	0.13	0.61	0.75	1.25	1.13	1.28	0.24	0.46	1.94	0.45	0.36	0.37	1.53
He	0.23	4.82	4.76	0.72		1.23	0.50	0.31	1.50	1.03	0.74	0.05	1.81		0.36	0.98	
Ap					0.13	0.13							0.25		0.12		0.21
Ru																	
D.I.	72.42	74.54	75.16	79.94	78.63	82.21	78.71	74.36	75.86	74.31	90.80	88.02	72.86	85.16	88.44	89.22	72.37
C.I.	8.23	11.33	11.20	5.44	4.51	4.91	8.84	6.74	10.16	11.05	2,94	2.00	11.37	3.26	4.60	3.55	7.92

Table1/C

Upper Tuff Complex					
Sample No.	25A	9A	9B	9C	9D
SiO_2	73.18	76.31	76.75	76.47	75.28
TiO_2	0.23	0.07	0.16	0.10	0.25
Al_2O_3	14.32	13.14	13.27	13.28	14.00
Fe_2O_3	1.90	1.18	0.41	0.98	1.26
FeO	0.39	0.48	0.34	0.23	0.26
MnO	0.05	0.03	0.02	0.03	0.02
MgO	0.59	0.14	0.15	0.15	0.16
CaO	2.09	0.91	1.17	1.21	2.09
K_2O	5.05	5.11	4.12	4.49	3.68
Na_2O	2.13	2.63	3.54	3.01	2.89
P_2O_5	0.09	0.02	0.03	0.02	0.06
Σ	100.02	100.02	99.96	99.97	99.95
LOI	5.35	4.23	0.91	3.61	4.00
CIPW norms					
Q	77.90	83.31	84.68	82.17	81.48
Or	6.97	6.43	5.80	6.11	4.79
Ab	4.39	5.00	7.50	6.11	5.72
An	4.65	1.90	2.77	2.80	4.51
C	2.06	2.02	1.05	1.53	1.71
Di					
Hy	1.94	0.36	0.93	0.51	0.49
Wo					
Mt	0.26	0.72	0.40	0.25	0.12
Il	0.13	0.12		0.13	0.37
He	1.29	0.12	0.26	0.51	0.85
Ap	0.13				
Ru					
D.I.	89.26	94.74	94.98	94.33	91.95
C.I.	3.62	1.32	1.19	1.40	1.83

Middle Tuff Complex

Petrography

In harmony with the bipartite age distribution, the complex is petrographically also not uniform, the composition changing from andesitic to rhyolitic and shows a dacitic composition on the average.

This transitional character occurs also regionally (when moving from the west eastwards the composition becomes more basic) but the extreme mixtures within one rock type are also frequent (bounded pumice, microscopic-scale mixtures of andesitic-dacitic and andesitic-rhyolitic materials). In addition to the dacitic-rhyodacitic most frequent composition, a tuff complex of andesitic composition also occurs in the southeastern part of the area (Sály–Latorvár).

The main rock types are as follows: pumice and ash flow deposits; mixed pumice-scoria and ash flow deposits; ignimbrites with bounded and/or plastically deformed vitroclasts. In the initial phase reworked tuffs, in the closing phase thinner-thicker phreatomagmatic sequences are present.

Rock-forming minerals

The mafic components are represented by hyperstene, subordinately augite, green and brown amphiboles and biotite, as juvenile minerals (Photo 3). Hypersthene is FeO-rich (21.4–26.5%) depending on the rather dacitic than andesitic chemistry.

Here biotite is often opacitized and chloritized too; in the ignimbrites it is usually fresh. Based on microprobe analyses (Table 4) its total FeO-content varies between 27.5 and 29.0%; the MgO-content is 6.1–6.5%. Dobosi (1980) measured similar values in two samples from the complex.

The FeO-content of biotites is higher than of those of the calc-alkali dacites indicating the low temperature and high water saturation of the magma producing the tuff complex.

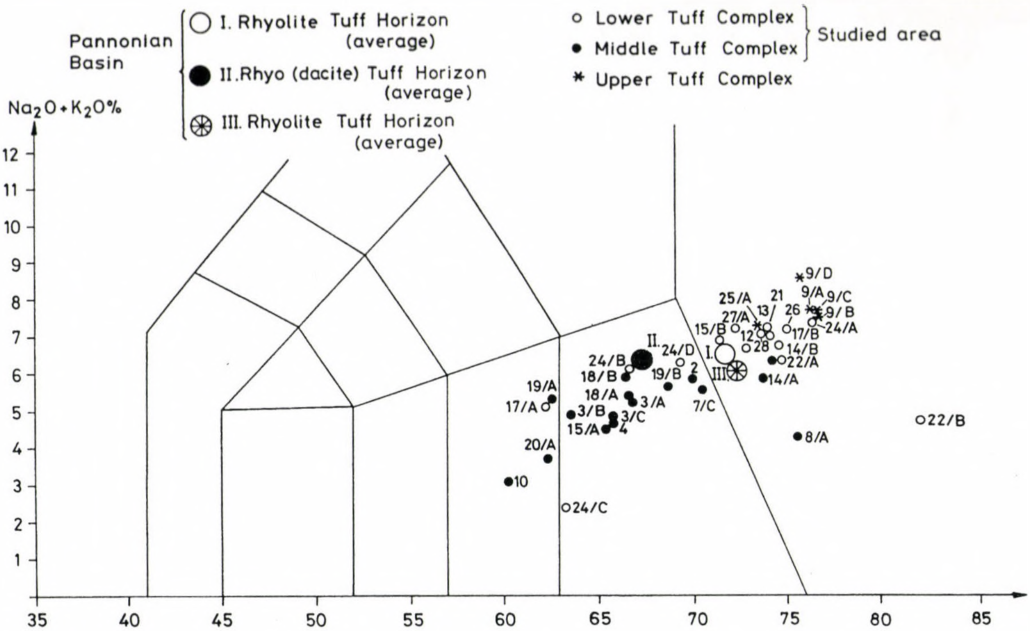

Fig. 2
TAS diagram (after Le Maitre et al. 1989; with values calculated for volatile-free state)

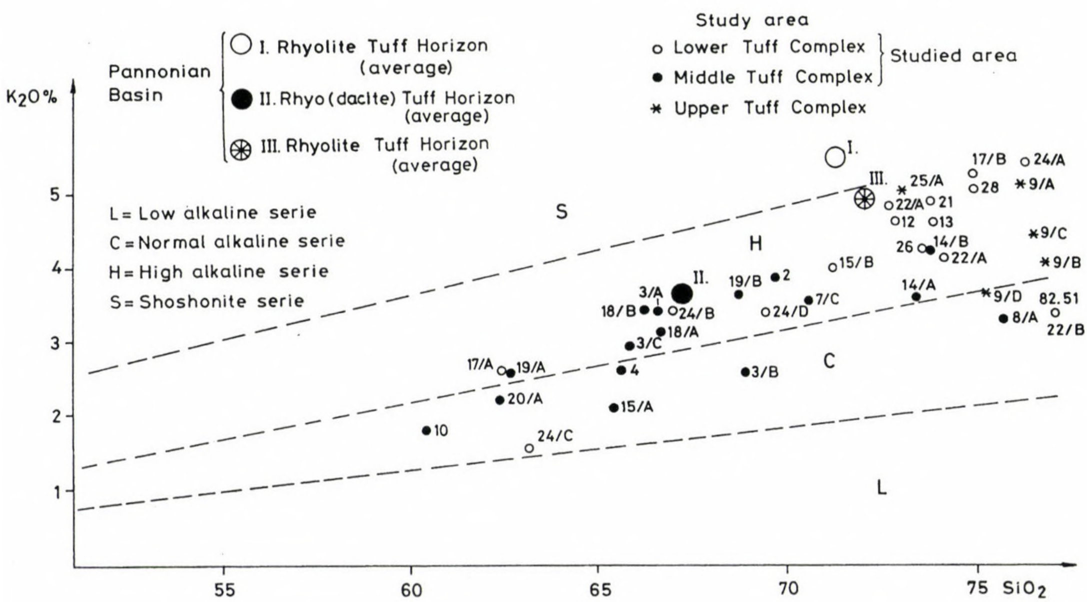

Fig. 3
K_2O–SiO_2 diagram (Peccerillo-Taylor 1976; with values calculated for volatile-free state)

Table 2
Trace element concentrations (ppm)

	Lower Tuff Complex (I. Phase)							Middle Tuff Complex (II. Phase)			Upper Tuff Complex (III. Phase)	
Sample No.	22B	24B	26	15B	17A	27A	12	20A	3A	15A	25A	9B
Sr	94.70	289.00	209.00	245.00	278.00	182.00	164.00	174.00	215.00	227.00	144.00	110.00
Ba	608.00	1140.	875.00	849.00	1220.	750.00	843.00	412.00	708.00	568.00	677.00	940.00
Rb	121.00	122.00	154.00	143.00	92.40	159.00	164.00	68.30	99.60	62.10	167.00	145.00
Th	14.20	17.14	21.20	17.70	12.50	19.30	19.62	9.64	12.85	10.42	22.18	17.00
U	4.40	4.30	4.96	4.70	2.90	5.10	5.60	1.94	3.10	2.80	6.60	5.40
Zr	180.00	220.00	130.00	150.00	240.00	153.00	182.00	217.00	350.00	280.00	184.00	250.00
Hf	4.58	5.24	3.57	4.30	5.00	4.10	4.50	5.14	7.40	6.60	4.75	7.00
Ta	1.00	0.96	0.96	0.95	0.68	1.00	1.10	0.57	0.78	0.84	0.89	1.37
Sc	4.30	14.40	4.28	4.64	23.00	4.30	4.67	15.30	14.80	25.50	4.92	3.49
Cr	10.70	33.00	4.86	253.00	332.00	90.20	97.80	80.00	75.00	51.90	51.00	177.00
Co	4.00	7.93	2.62	3.36	12.40	2.90	20.83	9.00	6.20	12.40	3.20	1.04
Cu*	4.70	8.90	32.40	51.40	188.00	72.40	76.70	56.70	70.50	15.30	82.70	
Zn	13.40	49.20	35.90	47.30	59.40	41.50	46.00	91.70	62.70	80.80	40.00	205.00
Mo	8.91	7.58	10.00	24.50	15.10	16.00	14.82	4.70	8.40		17.63	24.42
V	20.20	89.90	24.80	23.80	125.00			100.80	77.00	181.00	33.00	9.97
Ga	10.40	38.00	24.00	27.00	32.74		20.00	24.11	29.00			
As	8.86	3.12	4.12	4.04	3.87	4.31	3.72	3.49	3.57	1.09	12.41	3.53
Br				1.28			1.45		0.95			
Sb	0.94	0.27	0.22	0.20	0.73	0.24	0.21	0.33	0.22	0.12	1.00	0.17
Te	4.98	4.00	4.90	5.00		501.00	5.40			2.50	5.70	5.40
Cs	3.94	2.56	5.90	5.33	2.10	59.70	6.24	2.25	3.00	1.02	98.70	4.30
W	1.50		2.24	113.00	70.50	82.00	86.60	40.00	27.00	1.75	16.53	35.00
Ir	4.20	14.60	13.83	9.81	11.39	11.68	9.53		10.87	10.32		11.00

Table 3
REE concentrations (ppm)

	Lower Tuff Complex (I. Phase)							Middle Tuff Complex (II. Phase)			Upper Tuff Complex (III. Phase)	
Sample No.	22B	24B	26	15B	17A	27A	12	20A	3A	15A	25A	9B
La	34.00	43.40	46.30	32.60	33.10	30.40	29.40	26.20	36.40	36.80	31.00	42.00
Ce	74.00	89.00	86.00	64.50	68.00	63.00	57.70	60.00	78.00	72.20	52.00	80.00
Pr												
Nd	33.80	37.60	30.20	24.00	31.00	24.00	23.00	29.70	37.00	38.90	16.20	36.00
Sm	6.50	6.10	4.40	4.00	5.10	3.70	4.10	5.10	7.00	6.10	2.80	6.00
Eu	1.10	1.40	0.98	1.05	1.28	0.23	0.82	1.10	1.49	1.62	0.52	0.85
Gd	7.70	6.00	3.85	3.70	5.50	3.00	3.80	6.00	6.20	6.00	2.20	7.50
Tb	1.12	0.80	0.49	0.50	0.68	0.49	0.55	0.68	0.87	0.73	0.35	0.85
Dy	8.80	5.05	3.20	3.46	4.23	3.52	3.76	4.50		5.42	2.74	8.30
Ho	2.36	1.40	1.15	0.96	0.78	0.97	1.20	1.10	1.40	1.50	0.49	2.90
Er												
Tm	1.00	0.30	0.37	0.33	0.41	0.25	0.29	0.39	0.49	0.49	0.28	1.00
Ub	6.13	2.67	1.95	2.00	2.08	2.07	2.30	2.19	2.90	2.40	1.81	6.20
Lu	0.86	0.40	0.28	0.29	0.31	0.31	0.35	0.31	0.44	0.40	0.28	0.90
Σ	177.37	194.12	179.17	137.39	152.47	131.94	127.27	137.27	172.19	172.56	110.67	192.50
La/Yb	5.54	16.25	23.74	16.30	15.91	14.86	12.78	11.96	12.55	15.33	17.12	6.77

Augite is Ca-rich and is frequent in the hypersthene-augite andesite lithoclasts.

Green and brown amphiboles occur in the dacite lithoclasts.

Plagioclase is of labrador-andesine and oligoclase-andesine composition (determined by optics) and is often zoned.

In the acidic varieties sanidine also occurs.

Pumice and vitroclasts

The composition of the pumice and glassy vitroclasts of the Middle Tuff Complex shows also high alkali content and rhyolitic chemistry (Table 6). The FeO-content varies more than in the pumice and glassy vitroclasts of the Lower Tuff Complex; locally it may be as high as 1–2% (determinated by microprobe – Fig. 10).

Petrochemistry

The major element composition of the Middle Tuff Complex (Table 1) changes from medium- to high-K andesitic-dacitic to K-rich rhyolitic composition (Figs 2 and 3; values calculated to water-free conditions). Its DI value is 72–90 relating to remarkable differentiation. The CI-value is 3.2–11.3.

REE and incompatible elements

Analytical values of the REE and trace elements (Tables 2 and 3) were normalized as for the Lower Tuff Complex (Figs 4–9).

As compared to the Lower Tuff Complex the distribution pattern differs in the HREE when using the REE values normalized to primitive mantle, since the HREE quantities are greater than the former ones; a relatively slight Eu-anomaly occurs (Fig. 4). Values normalized to the granulite xenoliths (lower crust) show similar picture (Fig. 5).

The distribution normalized to granitoid is mostly free of anomaly. Nevertheless, instead of the Eu-anomaly a negative Sm and Tb anomaly occurs (unfortunately for the granitoids had no data on Gd). This verifies the granitoid origin though relates to certain magma mixing as well (Fig. 6).

The values of strongly incompatible elements normalized to primitive mantle and lower crust (granulite xenoliths) are similar shape as in case of the Lower Tuff Complex: from Rb to Sr display smaller positive anomalies from the Sr to Yb both positive and negative anomalies occur. The shape of the curve, however, is smoother than in case of the Lower Complex. Similar shape can be obtained for the lower crust (Figs 7 and 8).

In case of the values normalized to granitoids anomalies were measured more characteristic than in the rocks of the Lower Tuff Complex. Positive anomalies: Ba and Sm; negative anomalies: Rb,Th, U and P (Fig. 9). This can be related to the mixed character of the Middle Tuff Complex and may indicate certain magma mixing as well.

Table 4
Microprobe analyses

	Lower Tuff Complex								Middle Tuff Complex				Upper Tuff Complex			
	Biotite															
Sample No.	26		21		27		12		3/A		34/A		25/A		9/A	
	max	min	max	min	max	min	max	min	max	min	max	min	max	min	max	min
SiO_2	35.97	35.34	37.24	35.85	36.07	35.49	36.90	34.3	34.24	33.66	35.14	34.13	36.93	36.52	34.98	34.64
TiO_2	4.23	3.67	3.69	3.55	3.73	3.52	3.76	3.69	4.36	4.24	4.72	4.50	4.18	3.89	3.58	3.40
Al_2O_3	14.56	14.00	15.14	14.29	15.06	14.32	14.98	14.40	15.16	14.68	14.70	14.55	14.14	13.64	13.55	13.19
FeO^*	24.94	24.77	23.99	23.44	24.87	24.28	24.96	24.30	28.91	28.28	29.07	27.51	23.76	22.74	29.76	29.30
MnO	0.25	0.20	0.24	0.22	0.29	0.22	0.25	0.21	0.21	0.19	0.18	0.14	0.26	0.22	0.37	0.34
MgO	9.11	8.76	9.18	8.96	9.45	8.97	9.23	8.93	6.16	6.11	6.62	6.17	9.24	8.40	5.36	5.32
CaO	0.03	0.00	0.07	0.02	0.02	0.00	0.34	0.00	0.01	0.00	0.01	0.00	0.06	0.02	0.05	0.02
Na_2O	4.49	0.43	0.40	0.35	0.46	0.37	0.47	0.41	0.57	0.50	0.59	0.55	0.44	0.30	0.46	0.45
K_2O	9.11	8.83	8.82	8.28	9.19	8.87	9.06	8.69	8.74	8.44	8.06	7.94	8.92	8.35	8.81	8.14
	Ionic Num**															
Si	5.54	5.44	5.66	5.52	5.52	5.42	5.59	5.37	5.35	5.32	5.44	5.36	5.71	5.70	5.61	5.54
Ti	0.49	0.43	0.42	0.41	0.43	0.41	0.49	0.42	0.51	0.51	0.55	0.53	0.49	0.45	0.43	0.41
Al	2.64	2.54	2.71	2.56	2.71	2.60	2.71	2.57	2.78	2.75	2.69	2.68	2.57	2.49	2.55	2.48
Fe	3.21	3.16	3.02	3.00	3.19	3.11	3.21	3.08	3.77	3.76	3.82	3.56	3.07	2.97	3.98	3.95
Mn	0.03	0.03	0.03	0.03	0.04	0.03	0.03	0.03	0.03	0.03	0.02	0.02	0.03	0.03	0.05	0.05
Mg	2.08	2.01	2.11	2.07	2.15	2.06	2.11	2.05	1.45	1.42	1.53	1.44	2.13	1.95	1.28	1.27
Ca	0.01	0.00	0.01	0.01	0.00	0.00	0.06	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00
Na	0.15	0.13	0.12	0.10	0.14	0.11	0.14	0.12	0.17	0.15	0.18	0.17	0.13	0.10	0.14	0.14
K	1,79	1.73	1.74	1.61	9.19	1.74	1.78	1.68	1.73	1.73	1.61	1.57	1.76	1.66+O8	1.79	1.67

* Total iron calculated as FeO and Fe2+
** Number of cations on the basis of 22 oxygen

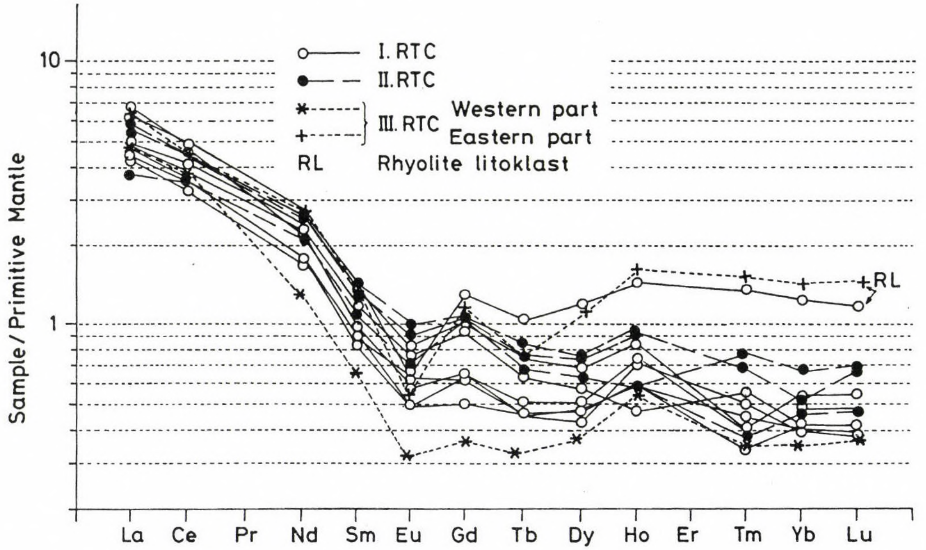

Fig. 4
REE-diagram normalized to primitive mantle (Sun and McDonough 1989);

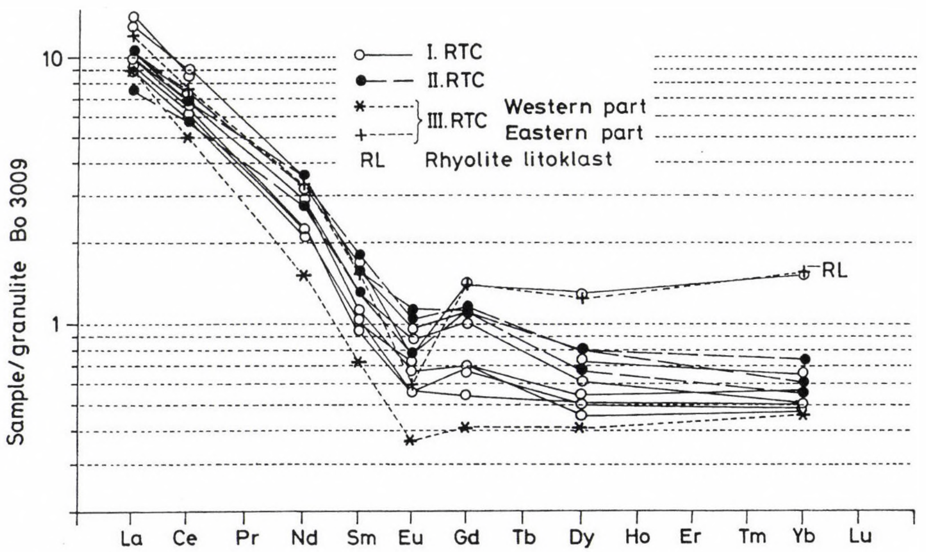

Fig. 5
REE-diagram normalized to granulite xenolith (Bo 3009) from Pliocene alkali basalt, Hungary (Kempton et al. 1997)

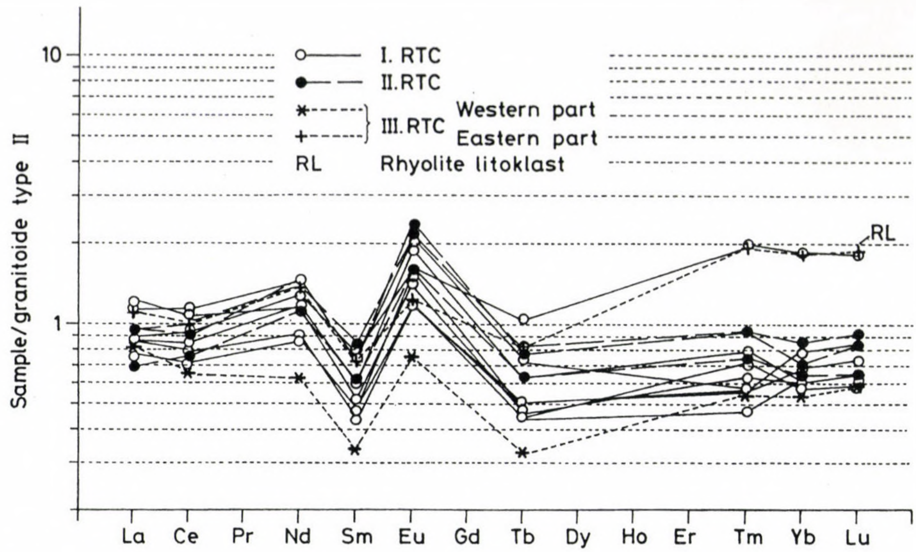

Fig. 6
REE-diagram normalized to granitoid (granite type-II, Velence Hills, Hungary), (Buda 1985)

Upper Tuff Complex

Petrography

The Upper Tuff Complex occurs in two smaller patches in the western part of the area (Demjén environs) and in the eastern part (Harsány environs) as well. The K/Ar age of the two occurences differs but both of them are of Sarmatian age (Márton E. and Pécskay, this volume).

In the Demjén exposure the ignimbrite (K/Ar age is 13.84 (±0.94 Ma)) is stratified, slightly fiammed, with pumice and lithoclasts. In addition to great amounts of biotite it contains less brown amphibole, as well. The Harsány occurrence is pumiceous crystal tuff with welded crystalloblastic groundmass composed of biotite, sanidine, and labradorite-andesine (Photo 4). It is characterized by felsitic rhyolite lithoclasts (K/Ar age is 15.66 (±0.60 Ma)) and white-to-grey pumice lapilli.

Rock-forming minerals

The biotite compositions of the rocks of the two occurrences considerably differ from each other (Table 4); in the western exposure the total FeO-content is 22–23% and the MgO is 8–9%, in the eastern exposure the total FeO-content is as high as 29% and the MgO content is 5%.

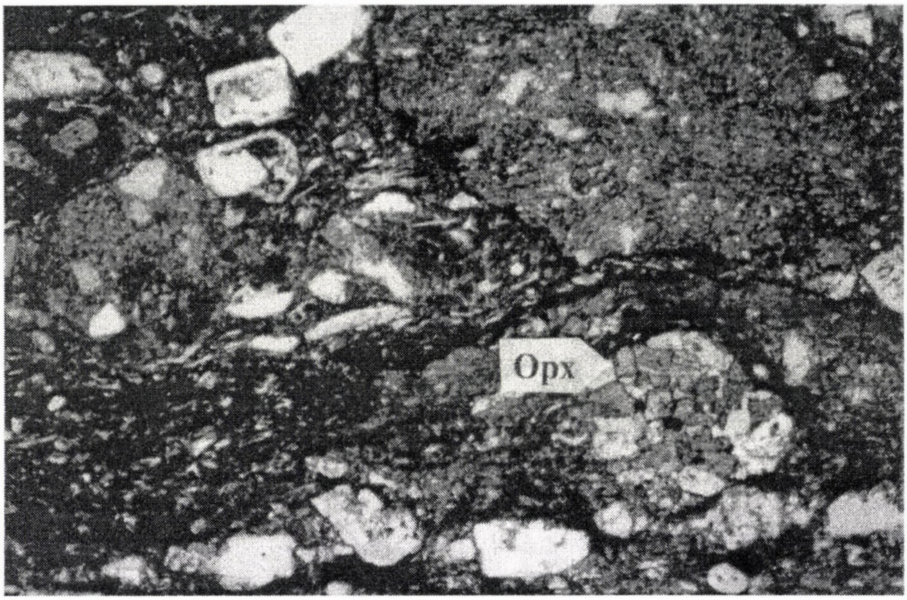

Photo 3.
Middle Tuff. Lithic-rich, slightly-welded pink ignimbrite. Pyroxene (Opx) is typically present. North of Bogács. Px > Bi > Q > Am (II N)

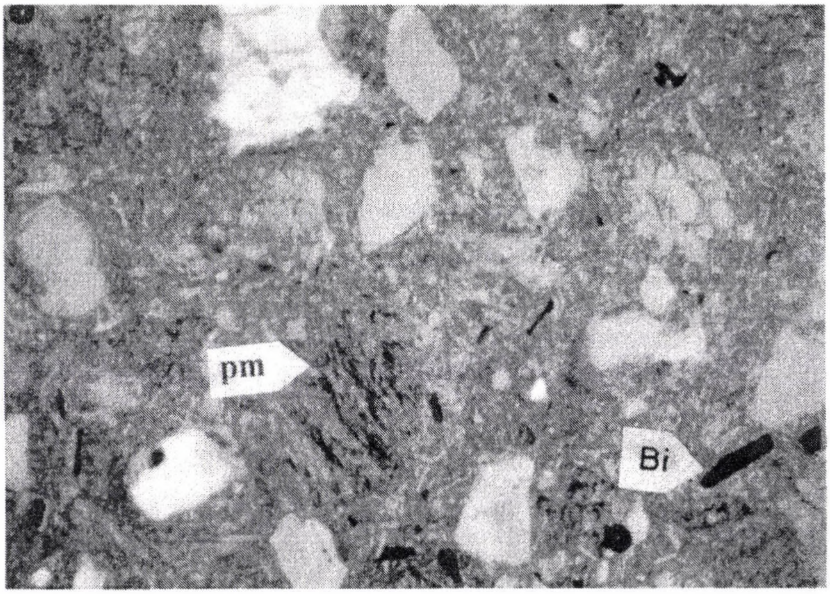

Photo 4.
Upper Tuff. Non-welded ignimbrite (pm = pumice, Bi = biotite). Harsány. Bi > Q > Am (II N)

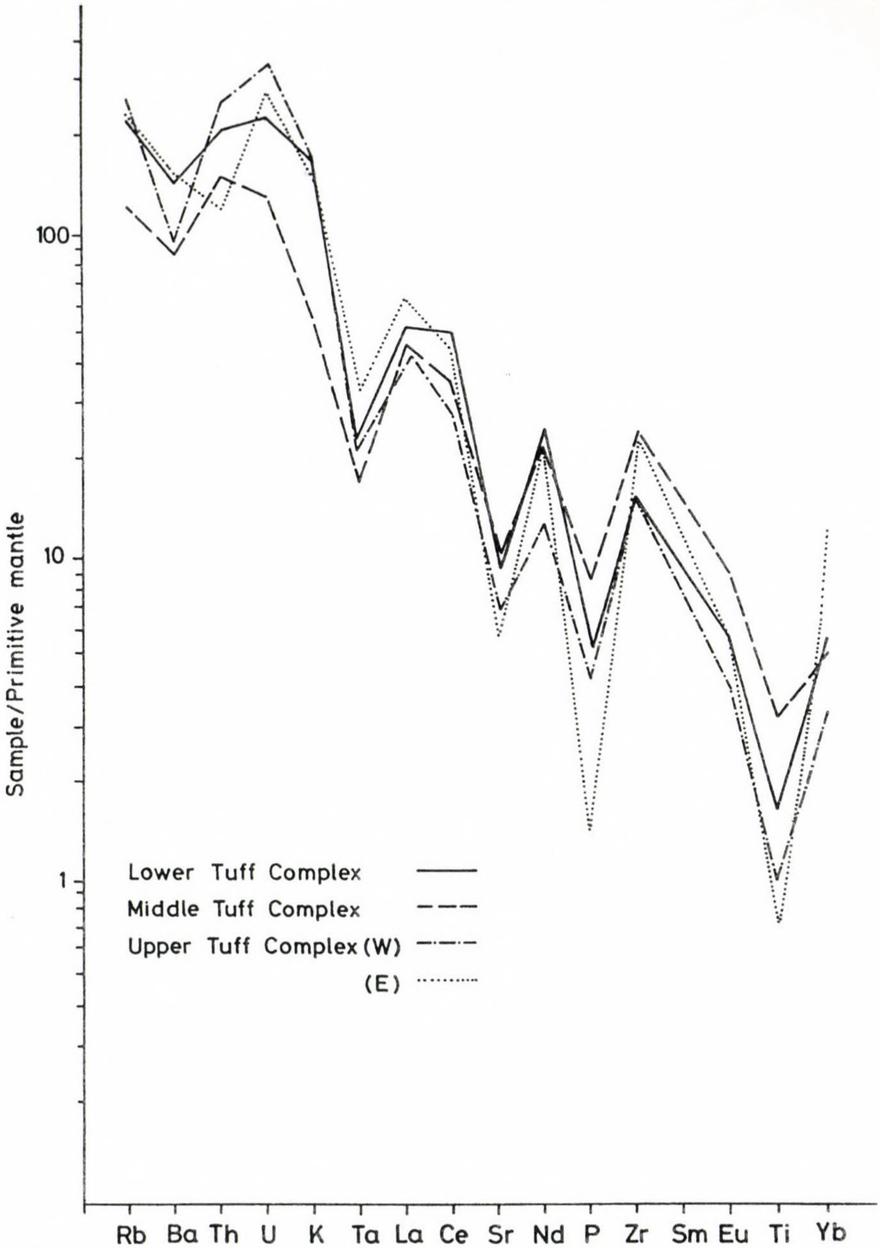

Fig. 7
Incompatible trace element diagram normalized to primitive mantle (Sun and McDonough 1989) with the average values of the tuff complexes

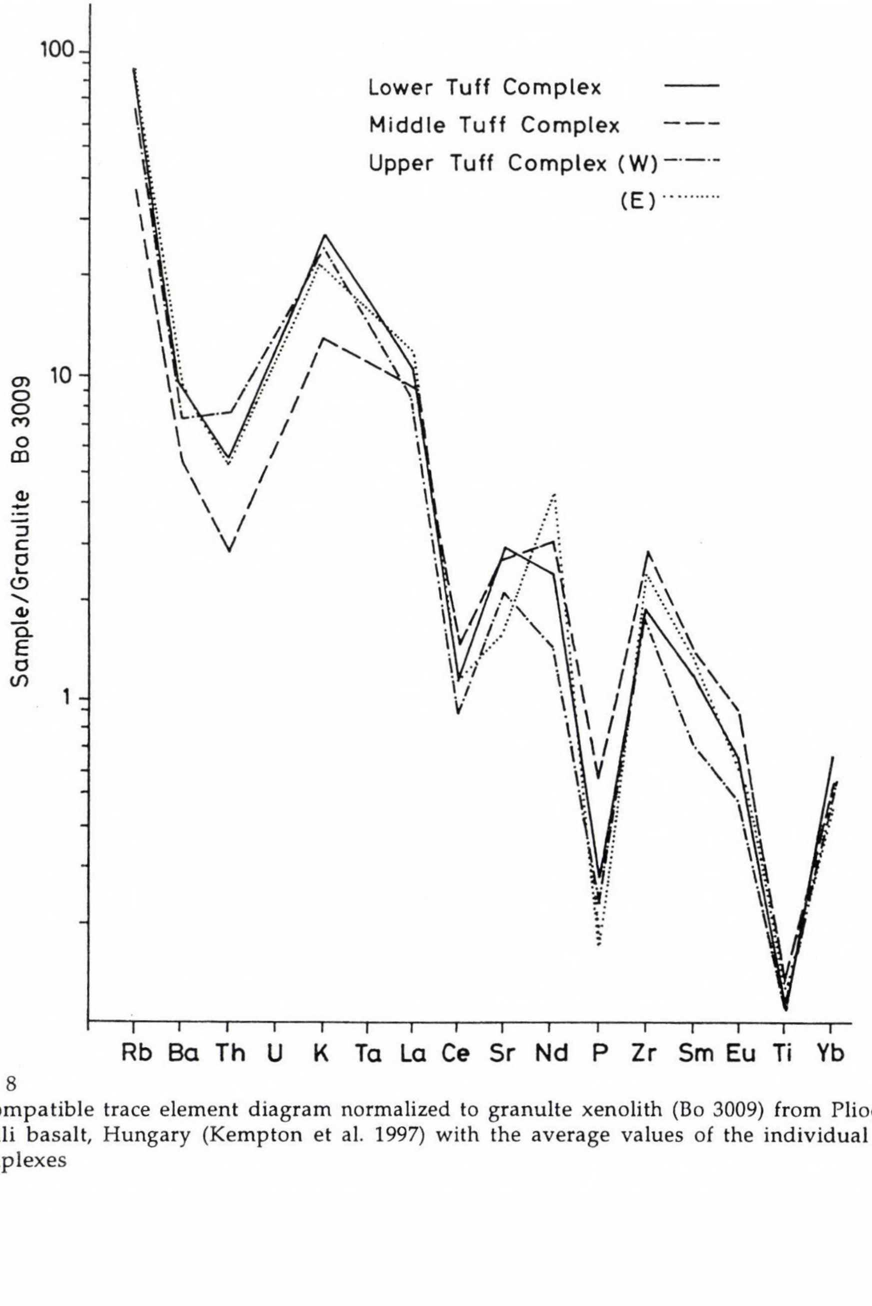

Fig. 8
Incompatible trace element diagram normalized to granulte xenolith (Bo 3009) from Pliocene alkali basalt, Hungary (Kempton et al. 1997) with the average values of the individual tuff complexes

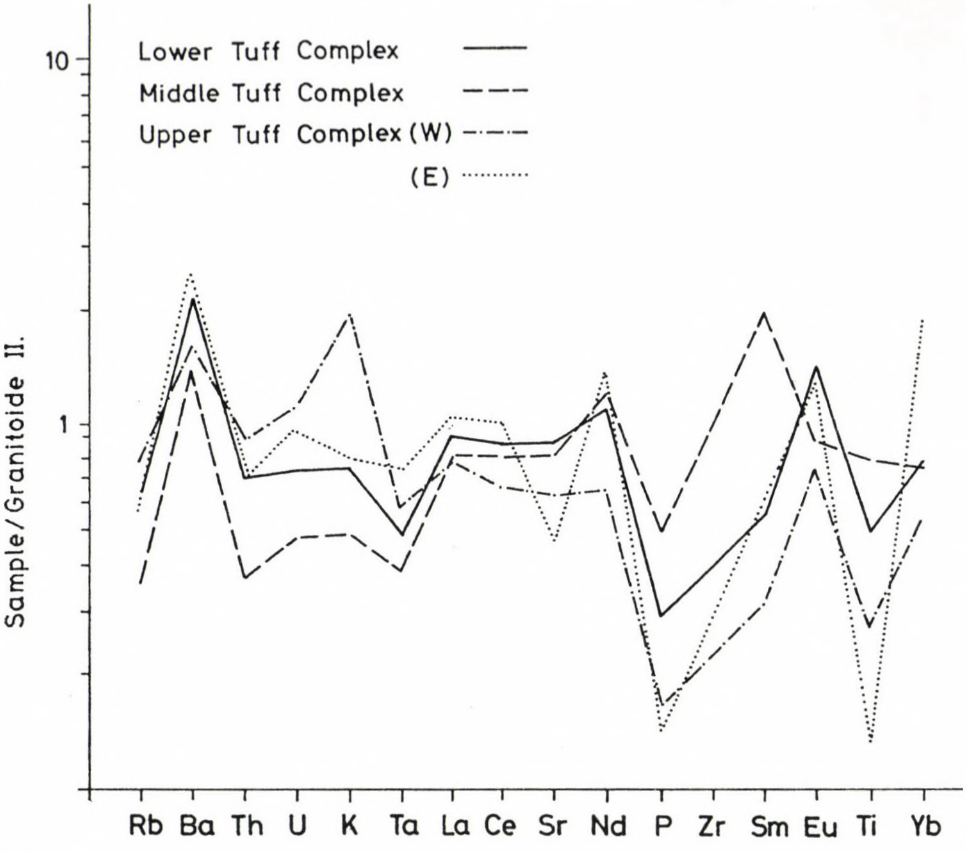

Fig. 9
Incompatible trace element diagram normalized to granitoid (granite type-II in the Velence Hills, Hungary), (Buda 1985) with the average values of the individual tuff complexes

Petrochemistry

The major element concentrations do not display greater variations though chemism of the western occurrence is somewhat more basic (Table 1). Both of the tuff complexes are rich in potassium and rhyolitic in composition (Figs 2 and 3). The DI-value and CI-value of the western occurrence are 89% and 3.62, respectively; those of the eastern formation are 92–95% and 1.2–1.8.

REE and incompatible trace elements

The REE patterns and incompatible trace element contents (Table 2 and 3) also support the different origins of the two tuff occurrences. In addition to the remarkable concentration differences the values normalized to primitive mantle show that there are considerable differences in the negative anomalies of the HREE

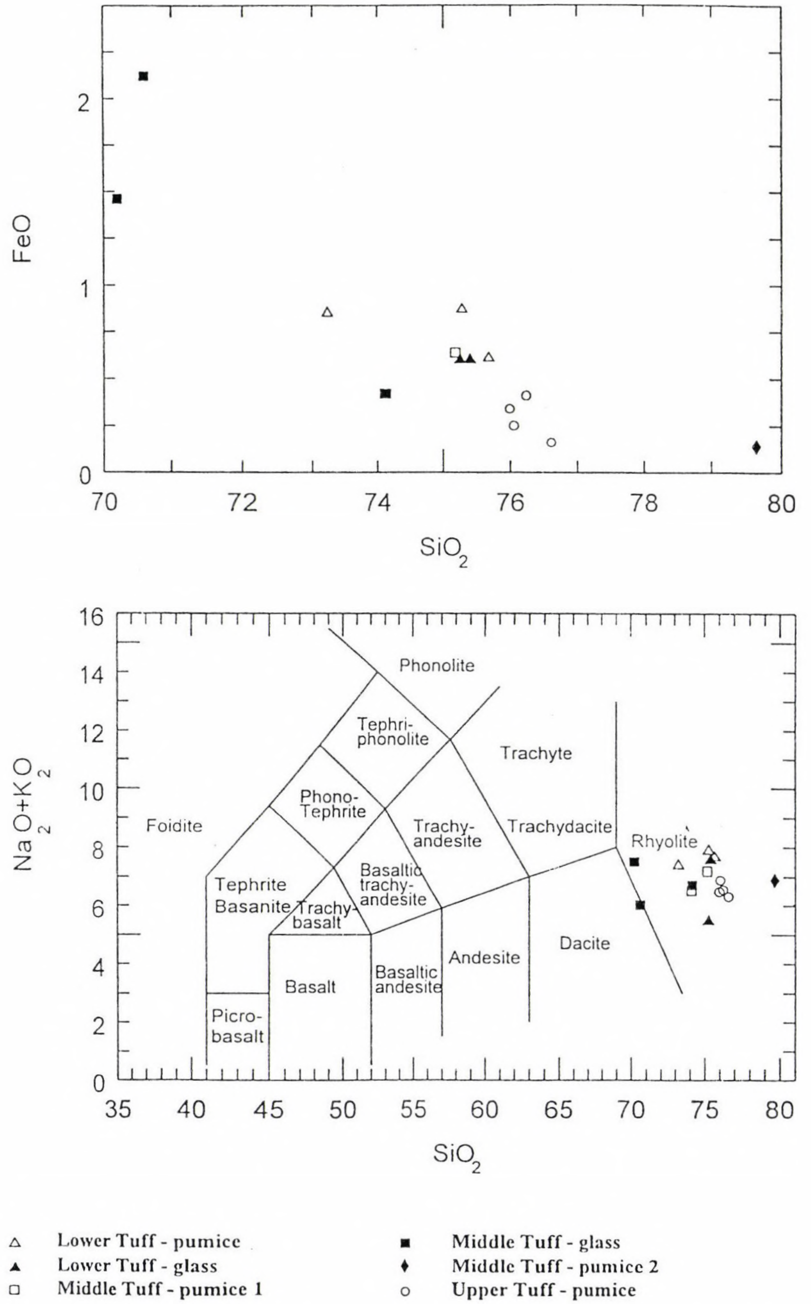

Fig. 10
$Na_2O + K_2O - SiO_2$ (Le Maitre et al. 1989) and $FeO - SiO_2$ diagram of pumices and glassy vitroclasts measured by microprobe

Table 5
Microprobe analyses

	Middle Tuff Complex				Lower Tuff Complex	
	Orthopyroxene		Clynopyroxene		Amphybol	
Sample No.	3/A		34/A		33/B	
	max	min	max	min	max	min
SiO_2	52.31	51.84	52.82	51.86	44.44	45.18
TiO_2	0.22	0.10	0.24	0.21	1.49	1.36
Al_2O_3	0.73	0.74	1.02	1.09	7.11	6.99
FeO*	25.19	26.65	22.83	25.29	15.33	15.63
MnO	0.61	0.66	0.51	0.46	0.24	0.31
MgO	20.53	19.85	23.04	21.29	12.36	12.34
CaO	1.31	1.19	1.36	1.06	9.14	9.02
Na_2O	0.01	0.02	0.01	0.01	1.30	1.28
K_2O	-0.01	-0.01	0.01	0.01	0.51	0.48
Ionic Num**						
Si	7.85	7.83	7.77	7.76	6.97	7.09
Ti	0.03	0.01	0.03	0.02	0.18	0.16
Al	0.13	0.13	0.18	0.19	1.32	1.28
Fe	3.16	3.37	2.81	3.17	2.01	2.04
Mn	0.08	0.08	0.06	0.06	0.03	0.04
Mg	4.60	4.47	5.05	4.75	2.89	2.86
Ca	0.21	0.19	0.21	0.17	1.54	1.51
Na	0.00	0.01	0.00	0.00	0.40	0.39
K	0.00	0.00	0.00	0.00	0.10	0.10

* Total iron calculated as FeO and Fe^{2+}

** Number of cations on the basis of 22 oxygen

and mainly of Eu. In the rocks from the western occurrence there is no anomaly, indicating slight differentiation. The REE pattern of the rocks from the eastern area is very similar to sample 22/B of the Lower Tuff Complex (rhyolite lithoclast) (Fig. 4).

A similar REE is characteristic of the values normalized to the lower crust (granulite xenolite) but with anomalies of smaller intensities (Fig. 5).

In the case of the values normalized to granitoids the anomalies are considerably reduced but in the shape of the Eu and HREE anomalies there is a difference between the two occurrences (Fig. 6).

Values of the incompatible elements (Table 3) normalized to primitive mantle are similar to those of the Lower and Middle Tuff Complexes but almost all the anomalies are of greater intensity (especially in case of P and Ti) (Fig. 7). In case of the values normalized to the lower crust (granulite xenoliths) similar but less intense anomalies are characteristic (Fig. 8).

In case of both occurrences remarkable differences can be observed in the values normalized to granitoids, both in the elements having anomalies and in the intensities of the anomalies, especially in case of Th, U, K, Sr, P, and Zr (Fig. 9).

Table 6
Microprobe analyses. Pumice and glassy fragments

	Lower Tuff Complex					Middle Tuff Complex						Upper Tuff Complex			
Sample No.	27/A					4					10	25/A			
	Pm/1	Pm/2	Pm/3	Gl/1	Gl/2	Pm/1	Pm/2	Gl/1	Gl/2	Gl/3	Gl/1	Pm/1	Pm/2	Pm/3	Pm/4
SiO_2	75.28	75.68	73.22	75.40	75.25	74.11	75.18	74.14	70.61	70.21	79.64	76.60	76.60	76.23	75.99
TiO_2						0.15		0.20	0.34	0.32		0.16			12.53
Al_2O_3	12.06	12.14	11.77	11.95	12.12	12.91	11.86	12.98	13.97	14.39	12.23	12.48	12.03	12.30	0.25
FeO	0.87	0.61	0.85	0.60	0.60	0.64	0.16	0.42	2.12	1.46	0.14	0.16	0.41	0.34	
MnO			0.03						0.07	0.07					
MgO		0.14	0.05						0.13	0.09					0.91
CaO	0.94	1.08	0.85	0.77	0.86	0.69	1.00	0.99	1.80	1.51	1.15	1.00	0.84	0.81	1.76
Na_2O	2.65	2.47	2.69	2.55	1.86	2.05	1.60	1.33	1,71	2.46	4.63	1.60	1.63	1.79	5.12
K_2O	5.20	5.19	4.69	5.01	3.60	5.13	4.71	5.38	4.31	5.05	2.25	4.71	4.92	4.69	
P_2O_5	0.40														
Σ	97.40	97.31	94.15	96.28	94.29	95.68	94.51	95.44	95.06	95.56	100.04	96.71	96.43	96.16	96.56

Table 7
Petrographical and geochemical features of Miocene acidic explosive volcanics in Bükk foreland

	Rock types	Juvenile mafic rock-forming minerals	Petrochemistry	Origin according to geochemical data
Upper Tuff Complex	Non-welded or slightly welded ignimbrites, crystal tuffs Lithoclasts: rhyolitic, dacitic	Biotite western, eastern p. FeO: 22–23%, 29% MgO: 8–9%, 5.0% (some hypersthene and amphibole)	High-K rhyolitic	Molten from acidic crust
Middle Tuff Complex	Fallout ash flow deposits, mixed pumice-scoria dep. ignimbrites (sl. and st. welded) Lithoclasts: andesitic, dacitic, rhyolitic	Hyperstene FeO: 21.4–26.5% Biotite FeO: 27.5–29.0% MgO: 6.6–6.5% Augite (Ca-rich) (Some green and brown amphibole)	Normal and high-K (rhyo)dacitic (From andesitic to rhyolitic)	Mixing of acidic (molten from upper crust) and intermediary (andesitic) magmas
Lower Tuff Complex	Fallout pumice, lapilli. sl. and st. welded ignimbrites, phreatomagmatic tuffs Lithoclasts: rhyiolitic, dacitic, andesitic alk. diabasic	Biotite FeO: 23–25% MgO: 8–8% (Some hyperstene, augite, green amphibole)	High-rhyolitic (Ultraacidic)	Molten from acidic crust

The two formations of the Upper Tuff Complex differ from the Lower and Middle Tuff Complexes both with respect to the REE and incompatible trace elements. This relates to the fact that the crustal part providing the magma melt could have considerably changed by the Sarmatian and that the explosion time of the two occurrences was different as well. The determination of the relations of the Upper Tuff Complex with other tuff horizons needs further investigation.

Conclusions

1. Up-to-date magmagenetic works and interpretations of the Miocene calc-alkali intermediary magmatism of the Pannonian basin are reported by Salters et al. (1988) and Downes et al. (1995), but the attempt in modern analyses of the Miocene acidic explosive volcanism of great stratigraphic and geodynamic significance first time was performed by Szakács et al., Márton, E. and Pécskay (this volume) and by the present team.

2. The investigation of the Bükk Foreland proved to be suitable since in this area the Lower and Middle Tuff Horizons and to a smaller extent the Upper Tuff Horizon are found on at the surface. It was proved that these formations are akin to the tuff horizons developed elsewhere in the Pannonian Basin.

3. In the Bükk Foreland the individual tuff complexes can be distinguished from one another on the basis of their characteristic rock types, mineral compositions, petrochemistry and geochemistry (Table 7).

4. Data from petrological–geochemical investigations suggest that the acidic tuff complexes derive from the upper crust of granitic or acidic metasediment composition. On a magma-tectonic basis this concept had been already deduced by G. Pantó (1962).

5. A new concept is that in the evolution of the Middle Tuff Complex magma mixing should have been significant (mixing of acidic magma melted from upper crust and intermedier-andesitic-magma). This concept is supported by the great intensity of the Carpathian–Badenian andesitic volcanism in the Pannonian Basin in the same geologic time.

6. The Sarmatian Upper Tuff Complex (being the latest occurrence of the intense acidic volcanism in the Pannonian Basin) can be derived also from the melting of an acidic upper crust. The two occurrences in the Bükk Foreland could have derived, however, from different volcanic centers.

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Appendix

Localities and brief description of the samples analyzed in this study (from W to E in the study area: see the location map, Fig 1). Sample number used here are the same as in Tables and Figures, and in the map of sampling sites.

Abbreviations: Chem. – chemical analysis of the major elements; Tr.e. – trace element analysis; REE – rare earth element analysis; MP – microprobe analysis.

Sample	Analyses	Localities	Brief description
Lower Tuff Complex			
22/A	Chem.	Eger, Wind brickyard factory, quarry	Coarse-grained, lapilliferous lithoclastic rhyolite tuff
22/B	Chem., Tr.e., REE	Eger, Wind brickyard factory, quarry	Rhyolite lithoclast with flow texture from the rhyolite tuff
24/A	Chem.	Eger, Tihamér tuff quarry (upper level)	Slightly welded tuff with crystal clasts, pumice and rhyolite lapilli and groundmass fragments (oligoclase-andesine, sanidine, biotite)
24/B	Chem., Tr.e., REE	Eger, Tihamér tuff quarry (upper level)	Biotitic-hypersthenic dacite lapilli
24/C	Chem.	Eger, Tihamér tuff quarry (lower level)	Hypersthene andesite lithoclast
24/D	Chem.	Eger, Tihamér tuff quarry (lower level)	Amphibole dacite lithoclast (about 15%) green amphibole, quartz, oligoclase-andesine, K-feldspar of microcline structure in strongly altered spherolitic groundmass, the texture is coarse-porphyric

Sample	Analyses	Localities	Brief description
26	Chem.,Tr.e., REE, MP: biotite	Demjén, Pünkösd Hill	Welded tuff (oligoclase, sanidine, biotite, with plastically deformed vitroclasts)
15/B	Tr.e., REE	Sály, Lator út	Red, welded ignimbrite (oligoclase, hypersthene, biotite, hematite, with molten glass fragments)
21	Chem., MP: biotite	Ostoros, Cellars	Non-welded coarse pumice-bearing flood tuff (quartz, oligoclase-andesine, biotite, less amphibole, with perlit-rhyolite lithoclasts and zeolitized pumice lapilli)
17/A	Chem., Tr.e., REE	Szomolya, southern quarry	Porphyric, pilotaxitic, hypersthene-bearing augite andesite lithoclast of flow texture (basic labrador, hypersthene, augite, less amphibole)
17/B	Chem	Szomolya, southern quarry	Scoria
27/A	Chem., Tr.e., REE, MP: biotite, plagioclase, allanite, pumice		Slightly welded tuff (oligoclase, sanidine, biotite, vitroclasts, pumice)
28	Chem.	Kács, cellars at the village-end	Non-welded rhyolite tuff (oligoclase-andesine, sanidine, biotite, vitroclasts)
12	Chem., Tr.e., REE, MP: biotite,	Kisgyőr, western valley	Dark- and light-bounded welded tuff (quartz, oligoclase, sanidine, biotite, rhyolite lithoclasts with plastically deformed pumice)
13	Chem.	Kisgyőr-South, beside the road to Pusztamocsolyás	Pumice-bearing welded ignimbrite (with rhyolite and andesite lapilli)
33/B	MP: amphibole	Bükkzsérc, Baglyas	Pumice-bearing non-welded tuff (with dacite lithoclasts)
Middle Tuff Complex			
20	Chem., Tr.e., REE	Ostoros, southern cellars	Pumice-bearing biotitic flood tuff with pyroxene andesite and alkali diabase lithoclasts
19/A	Chem.	Novaj, cellars	Bounded, plastically deformed pumice-bearing ignimbrite (oligoclase, sanidine, biotite, hypersthene, pumice, glass fragments)
19/B	Chem.	Novaj, cellars	Phreatomagmatic sequence. In zeolitized groundmass strongly altered biotite, plagioclase
18/A	Chem.	Szomolya, Ispán vineyard	Welded ignimbrite (hypersthene, biotite, less green amphibole, quartz, oligoclase-andesine, plastically deformed pumice, glass fragments)

Sample	Analyses	Localities	Brief description
18/B	Chem.	Hill-crest between Szomolya and Novaj	Slightly welded ignimbrite of columnar cleavage (quartz, oligoclase-andesine, hypersthene, augite, less biotite)
2	Chem.	Bogács, Hintó-valley	Slightly welded flood-tuff (oligoclase-andesine, hypersthene, augite, green amphibole, pumice and rhyolite lapilli)
3/A	Chem., Tr.e., REE, MP: orthopyroxene, biotite, pumice	Bogács, Ábrahám Hill, upper horizon	Welded tuff (oligoclase, hypersthene, less augite, biotite, pumice and glassy fragments)
3/B	Chem.	Bogács, Ábrahám Hill (middle part)	Sluggy ignimbrite (scoria)
3/C	Chem.	Bogács, Ábrahám Hill (lower part)	Biotitic pumice
4	Chem.	Bogács, Szorospatak (upper level)	Red pumice-bearing welded tuff (with fumarola pipe breccia formation; quartz, andesine, hypersthene, amphibole, pumice and rhyolite lithoclasts)
6	MP: plagioclase, allanite, biotite	Bogács, western quarry, lower horizon.	Black-white bounded, slightly welded flood tuff (in the white bounds zeolitized pumice, in the dark bounds many mafic ingredients: hypersthene, augite, amphibole, biotite)
7/C	Chem.	Sály, Latorvár, Vár Hill	Welded ignimbrite (quartz, oligoclase, biotite)
8/C	Chem.	Cserépfalu, tuff quarry (upper level)	Brownish-red welded ignimbrite (oligoclase, quartz, biotite, hypersthene, allanite, green amphibole, magnetite)
15/A	Chem., Tr.e., REE	Sály, Lator út (upper level)	Red welded tuff (in the same exposure it overlies the Lower Tuff; quartz, oligoclase, hypersthene)
14/A	Chem.	Pusztamocsolyás, quarry beside the rode to Fehérgyarmat	Dacite lapilli (andesine-labrador, hypersthene, biotite)
10	Chem., MP: plagioclase, biotite, pumice	Kisgyőr–Kőbányatető	Pumice from crystalloblastic dacite and rhyolite from welded tuff
34/A	MP: clinopyroxene	West of Tibolddaróc	20–30 cm large pumices from lithoclastic pumice-bearing tuff (hypersthene, augite, biotite)
Upper Tuff Complex			
25/A	Chem., Tr.e., REE, MP: biotite, plagioclase, pumice	Demjén, Nagyeresztvény quarry (western margin of the study area)	Stratified, slightly fiammic, pumice-bearing vitroclastic rhyolite tuff (quartz, oligoclase-andesine, biotite, sanidine, less amphibole)

Sample	Analyses	Localities	Brief description
9/A	Chem., MP: biotite.	Harsány, tuff quarry	Pumice-bearing crystal tuff (slightly welded; quartz, labrador-andesine, sanidine and biotite in the crystalloblastic groundmass)
9/B	Chem., Tr.e., REE	Harsány, tuff quarry	Felsitic rhyiolite-lithoclast with lithophyses from the pumice-bearing crystal tuff
9/C	Chem.	Harsány, tuff quarry	White pumice from the crystal tuff
9/D	Chem.	Harsány, tuff quarry	Grey pumice from the crystal tuff (quartz, sanidine, biotite in the devitrified groundmass)

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Complex evaluation of paleomagnetic and K/Ar isotope data of the Miocene ignimbritic volcanics in the Bükk Foreland, Hungary

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The joint paleomagnetic and geochronological study of the Bükk Foreland is primarily aimed at providing a time framework for the ignimbritic volcanic evolution of the area, where relevant biostratigraphical data are lacking.

Rock samples for paleomagnetic measurements and K/Ar dating were mostly taken from the same sites. The total number of the paleomagnetic sites is 33, and of K/Ar sites 32.

K/Ar age determinations were mainly carried out on monomineralic fractions from ignimbrites and pumices; in the case of strongly welded ignimbrites and lithic clasts whole rock was analysed.

K/Ar dating indicates that the ignimbritic volcanism of the Bükk Foreland occurred between 21–13.5 Ma. Within this interval three main events can be distinguished: 21.0–18.5, 17.5–16.0 Ma and 14.5–13.5 Ma respectively. They are well separated by paleomagnetic declinations of the respective volcanic products (80°, 30°, 0°, respectively). The resolution is less in the K/Ar data because of the overlapping of the analytical ages.

In terms of the areal distribution the "Lower Ignimbrites" with 80° CCW rotation as well as the "Upper Ignimbrites" with 30° CCW rotation are widespread. Both measurement methods point to the occurrence of a youngest ignimbritic horizon in a restricted area at the western part of the Bükk Foreland.

The significance of the results from a geodynamic point of view is that the large Neogene CCW rotations of the Bükk Foreland are now precisely dated.

Key words: Bükk Foreland, ignimbrites, paleomagnetism, K/Ar dating, correlation of Miocene tuff horizons

Introduction

The volcanic evolution of the Neogene ignimbritic area of the Bükk Foreland was never previously established, though several studies (Schréter 1934; Pantó 1962, 1963; Balogh 1964; Varga 1976; Hámor et al. 1979) were carried out.

Nevertheless the map published by Schréter and updated by Balogh suggested that the volcanism was of lower-mid Miocene age and must have occurred in two phases; the older was called "Lower Ignimbrite" and the younger "Upper Ignimbrite".

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The separation of the two horizons was based on petrologic and chemical composition. The volcanics were deposited in terrestrial environment; therefore biostratigraphical constraints were lacking.

More recently K/Ar dating yielded ages of 19.5–17.0 Ma from some ignimbritic rocks from the Bükk Foreland (Hámor et al. 1979).

In 1988, systematic paleomagnetic studies began in the same area which showed that the two horizons, separated in the geological map, basically corresponded with two paleomagnetic groups; one of them shows about 80°, the other about 30° CCW declination deviation from the present north. The match between the ignimbritic horizons and the distinct paleodeclinations was so good that paleomagnetism seemed to be a very effective mapping method for the ignimbrite horizons of different ages. However, what has remained unknown was the exact age of the two horizons due to the scarcity of the earlier data. Thus, the radiometric dating of these ignimbritic rocks became a necessity.

In order to be able to correlate the paleomagnetic and radiometric data, the paleomagnetic sites were sampled for K/Ar age determination. In the meantime interest in the Bükk Foreland has increased, and both K/Ar and paleomagnetic samples were also collected at new sites. All the sampling sites are shown in Fig. 1.

The present study as a complex paleomagnetic and radiometric approach is aiming at providing a precise time framework for the volcanic evolution of the Bükk Foreland.

Sampling, measurements and results

Paleomagnetic samples were drilled and oriented with both magnetic and sun compass *in situ*, from welded and less welded ignimbrites. For K/Ar dating, samples were collected from massive ignimbrites as well as lithic clasts and pumice. Criteria for acceptance of samples for analysis were based on thin section examination. Special attention was paid to xenolith in the sense that contamination from xenolithic material was avoided.

Paleomagnetic measurements were carried out in the Paleomagnetic Laboratory of the Eötvös Loránd Geophysical Institute of Hungary. K-Ar dating was performed in the Institute of Nuclear Research of the Hungarian Academy of Sciences, Debrecen.

The results are summarized in Table I and Table II.

Discussion and conclusions

From the viewpoint of the present discussion the most important paleomagnetic properties are the excellent quality, also in a statistical sense, of the paleomagnetic directions (for documentation see Márton 1990; Márton and Márton 1996 and Table I) and the high Curie-points and unblocking temperatures (above 600 °C) of the magnetic minerals. The first permits us to distinguish clearly between

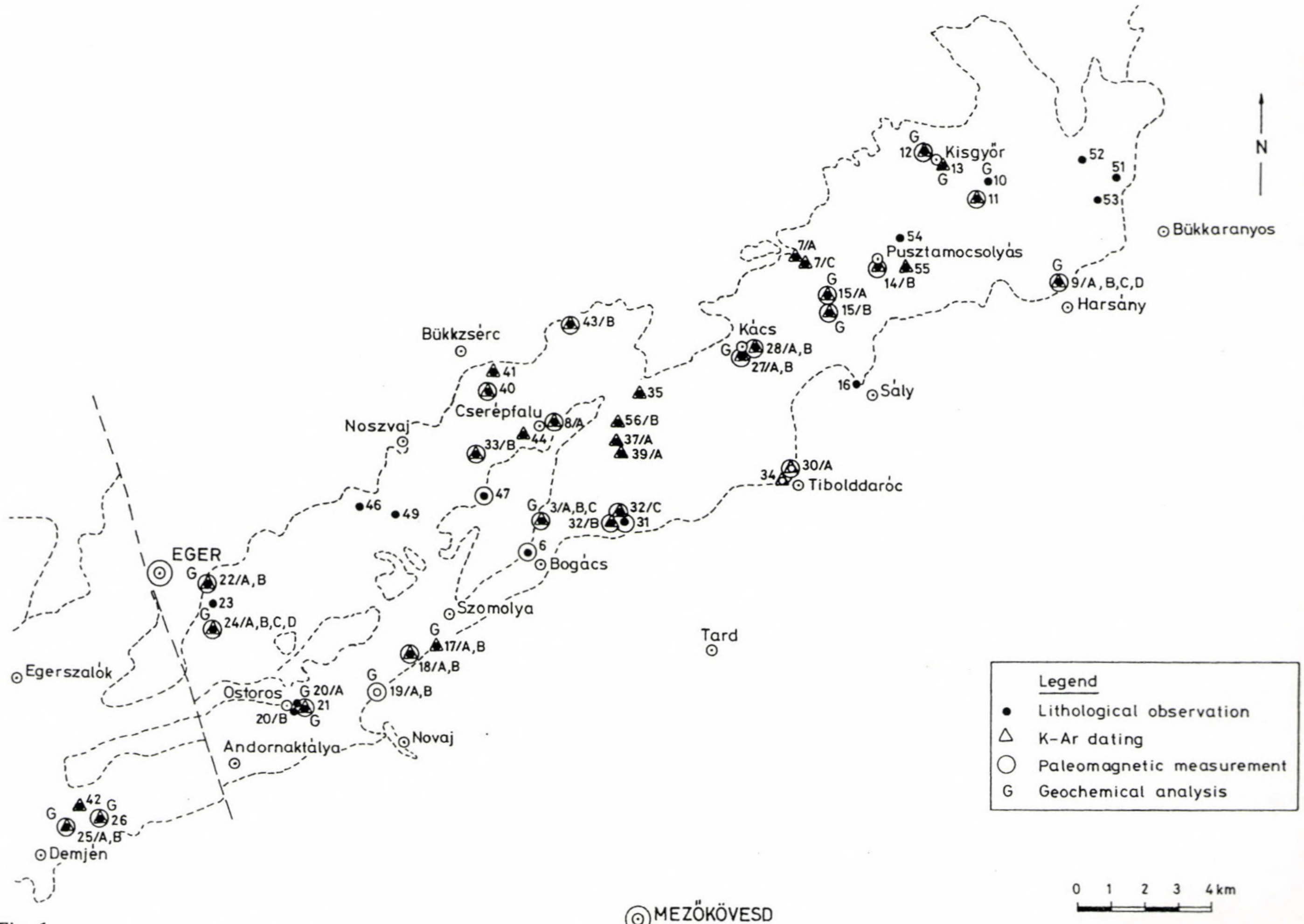

Fig. 1
Sketch map of the Bükk Foreland with the paleomagnetic and radiometric sampling sites

Table I.
Analytical data and simplified succession of the ignimbritic rocks from Bükk Foreland

Field No	Location	Rock type	Dated fraction	K (wt %)	$^{40}Ar_{rad}$ (%)	$^{40}Ar_{rad}$ (10^{-6}ccSTP/g)	K-Ar age (Ma)	D	I	K	α_{95}	N
BZ-26	Demjén, Pünkösd hill	welded	b. w.r.	6.39 3.34	38.1 61.8	4.668 2.230	18.69±0.89 17.09±0.68	116	-49	109	4.6	10
BZ-22A	Eger, Wind brick-works factory	non-welded	b. san.	4.93 6.51	28.0 89.0	3.882 5.004	20.2±1.7 19.7±0.9					
BZ-24A	Eger, Tihamér Quarry (upper)	pumice from non-welded	b.	5.52	17.1	4.262	19.75±1.66	115	-55	86	8.3	5
BZ-21	Ostoros SE vine cellars	welded	b.	6.49	23.2	4.536	17.88±1.17	119	-45	193	4.8	6
BZ-17B	Szomolya, S. Quarry	pumice from non-welded	b.	5.88	14.6	4.755	20.7±2.0					
BZ-40	Bükkzsérc, S. Oldal föld	non welded	b.	5.51	13.9	4.213	19.56±1.98	328	48	185	7.0	6
BZ-32B	Tard Quarry	non welded	b.	5.79	47.2 50.0	4.312 4.387	19.06±0.83 19.38±0.78					
BZ-43B	Cserépfalu, N Karácsony field	argillized vitro-clastic tuff	b.	5.63	14.1	3.915	17.80±1.79	105	-46	97	5.0	11
BZ-44	Cserépfalu, Quarry	pumice lapilli fall	b.	6.10	47.2	4.278	17.95±0.78					
BZ-39A	Cserépváralja "Fairy chimney"	pumice from non welded	b.	6.07	20.3	4.437	18.71±1.36					
BZ-35.	Cserépváralja Szurdokkővölgy	welded	b.	6.25	55.7	4.644	19.01±0.78					
BZ-37A	Cserépváralja Akácfa street 8.	pumice from non welded	b.	4.21	11.7	3.157	19.1±2.3					
BZ-56A	Cserépváralja, Törökréti brook	non welded	b.	4.63	17.8	3.689	20.4±1.6					
BZ-27A	Kács, Church hill	non-welded	b.	6.47	26.8	5.158	20.39±1.19					
BZ-27B	Kács, Church hill	welded	b.	6.67	46.0 37.0	4.294 4.368	16.5±0.7 16.8±0.8	111	-52	29	7.8	13
BZ-28A	Kács S. vine cellars	pumice from non-welded	b.	6.24	49.4	3.924	16.10±0.69					

Table I. (cont.)

Field No	Location	Rock type	Dated fraction	K (wt %)	$^{40}Ar_{rad}$ (%)	$^{40}Ar_{rad}$ (10^{-6}ccSTP/g)	K-Ar age (Ma)	D	I	K	α_{95}	N
BZ-15B	Sály, Lator street	welded grey	b.	6.50	45.4	4.672	18.39±0.81	101	-45	81	6.2	8
BZ-7A	Sály, N. Lator castle	non-welded	b.	6.35	17.1	5.065	20.40±1.72					
BZ-12	Kisgyőr, W	welded	b.I. b.II.	6.49 6.32	19.9 41.0 36.0	4.379 4.383 4.226	17.27±1.27 17.8±0.8 17.1±0.8	88	-50	139	2.5	25
TBZ-18	Kisgyőr, Kiskút	non welded	b.I. b.II.	4.85 5.61	18.7 13.0	3.746 4.163	19.75±1.52 19.0±2.0					
BZ-18.	Szomolya-Novaj Ispán vineyard	welded	w.r.	2.45	26.8	1.541	16.12±0.94	157	-35	672	1.8	12
BZ-3A	Bogács, N	welded	b+py	1.44	10.5	0.970	17.2±2.3	156	-26	676	1.9	10
BZ-41	Bükkzsérc, Nyomó hill	welded	b.	6.46	48.6	4.101	16.25±0.69					
BZ-8A	Cserépfalu, N Quarry	welded	b.	5.26	16.1	3.511	17.09±1.48	164	-26	396	3.4	6
BZ-32C	Tard, Quarry	welded	b. w.r.	5.52 2.93	35.4 34.5	4.019 1.854	18.40±0.91 16.20±0.81	148	-27	171	3.5	11
BZ-34	Tibolddaróc, At the church	pumice from non-welded	b.	6.25	13.4	3.934	16.12±1.71					
BZ-30A	Tibolddaróc, vineyard	pumice from red	b.	5.72	13.4	3.667	16.42±1.75	148	-31	64	3.9	22
BZ-15A	Sály, (upper) Lator street	welded, red	w.r.	1.93	38.3	1.279	16.97±0.80	162	-46	114	3.9	13
BZ-7C	Sály, Lator castle	welded	b.	5.31	15.8 19.9	3.491 3.470	16.83±1.51 16.73±1.23					
BZ-11	Kisgyőr Red quarry	red welded andesitic	w.r.	1.23	28.0	0.827	17.26±0.98	157	-40	133	2.6	24
BZ-55	Pusztamocsolyás Meredek hill	welded	w.r.	1.57	40.3	0.995	16.26±0.75	153	-39	110	4.0	13
BZ-25A	Demjén Nagyeresztvény	pumice from non-welded	b.	6.30	21.8	3.404	13.84±0.94	188	-50	92	3.3	21

Note: Atomic constants suggested by Steiger and Jäger (1977) were used for calculating age. All errors represent one standard deviation.
Details of the instruments and results of calibration have been described elsewhere (Balogh Kad. 1985).
Explanation: b. - biotite, san. - sanidine, w.r. - whole rock.
Key to paleomagnetic data: D = mean declination, I = mean inclination, k = precision parameter, a_{95} = circle of confidence, N = number of samples.

Table II.
K-Ar dating results for lithoclasts of ignimbrites from Bükk Foreland

Field No	Location	Rock type	Dated fraction	K (wt %)	$^{40}Ar_{rad}$ (%)	$^{40}Ar_{rad}$ (10^{-6}ccSTP/g)	K-Ar age (Ma)
BZ-25B	Demjén, Nagyeresztvény	biotite- hypersthene dacite	w.r.	2.71	23.6	1.518	14.35±0.92
BZ-42	Demjén	biotite-quartz rhyolite	w.r.	2.63	15.5	1.644	16.04±1.45
BZ-24B	Eger, Tihamér Quarry	biotite-hypersthene dacite	w.r.	2.69	76.6	1.832	17.39±0.67
BZ-17A	Szomolya, S. Quarry	hypersthene-augite andesite	w.r.	2.42	71.9	1.669	17.66±0.68
BZ-33B	Bükkzsérc, Baglyas	biotite-amphibole dacite	w.r.	4.13	67.1	2.819	17.47±0.69
BZ-56B	Cserépváralja, Törökréti brook	rhyolite	w.r.	2.33	27.8	1.563	17.17±0.98
BZ-28B	Kács, S. Vine cellars	biotite rhyolite	w.r.	3.26	64.3	2.142	16.82±0.67
BZ-13	Kisgyőr, S	rhyodacite	b.	5.90	16.2	3.796	16.48±1.46
BZ-14B	Pusztamocsolyás, Quarry	hypersthene-biotite dacite	w.r.	3.14	27.5	2.035	16.59±0.95
BZ-9B	Harsány, S Quarry	rhyolite	w.r.	2.95	80.6	1.804	15.66±0.60

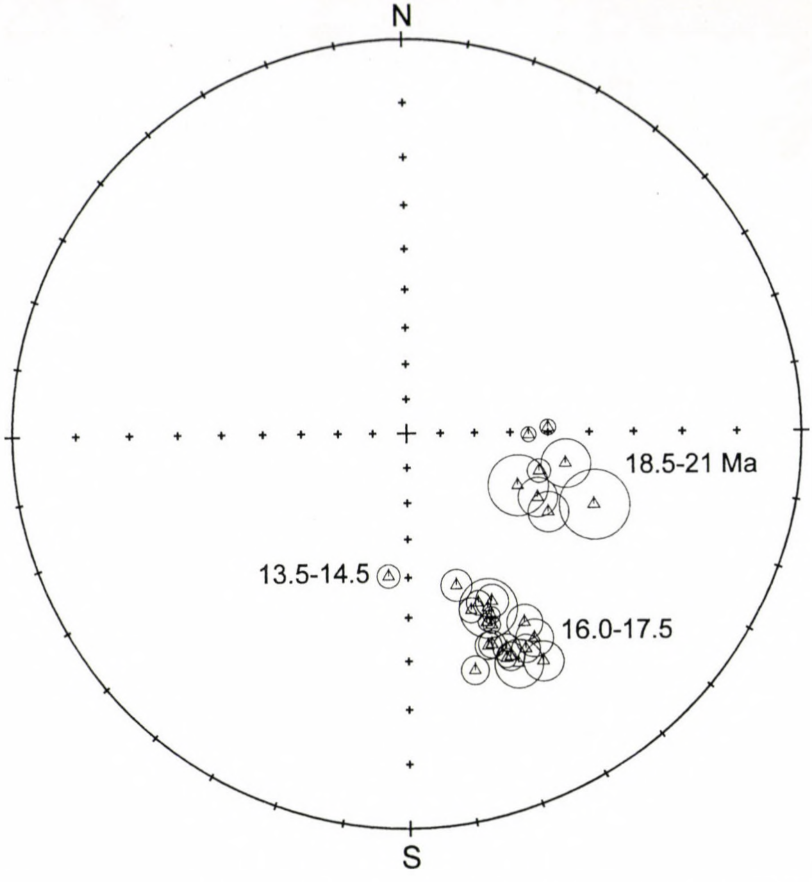

Fig. 2
Paleomagnetic site-mean directions with confidence circles for the ignimbrites of the Bükk Foreland. Stereographic projection. All inclinations are negative

paleomagnetic groups with different rotations (Fig. 2.); the second insures that possible moderate reheating of the area during subsequent volcanic events did not remagnetize the previously formed magnetic signal.

The paleomagnetic horizons of different rotation angles are also different in magnetic properties. The magnetic susceptibility (which is proportional to the magnetic mineral content) is one or two order of magnitude lower in the Lower Ignimbrite than in the "Upper Ignimbrite"; the intensity of the remanent magnetization is systematically lower in the former than in the latter.

As far as the K/Ar ages are concerned, we subdivide them into three groups (Fig. 3). The first group is the ages obtained from biotites separated from strongly welded ignimbrites, less welded ones and pumices; the second one consists of those obtained from whole rock samples of strongly welded ignimbrites; the third one is the ages of the lithic fragments (in tuffs).

Biotite ages range from 21-13.5 Ma with peaks at 19.5 Ma, 17.5-16.5 Ma and 14.5 Ma, respectively. Transitional ages between the oldest peak and the middle group of peaks (BZ-27B, BZ-28A, BZ-43B, BZ-12) are explained either by argon loss due to moderate reheating during the second eruption or may be regarded as the manifestation of two volcanic pulses during the second eruption event. The second explanation is in accordance with the geologic and petrographic results (Póka et al., 1998, this volume; Szakács et al., 1998 this volume). The youngest peak is sharp and distinct.

A striking feature of the histogram (Fig. 3) is that all but one of the lithoclasts and whole rock ages fall in the range of 18-16 Ma, though the lithoclasts (Table II) as well as the whole rock (Table I) ages represent both Lower and Middle Ignimbrites. This is explained by the better argon retentivity of the biotites than that of the other components of the rocks (see for example BZ-26 and BZ 32C in Table I, where the biotite ages are older than the whole rock ages).

In Fig. 3 the biotite ages obtained for the Bükk Foreland are compared with mean ages known for the Lower, Middle and Upper Rhyolite Tuffs of the entire Pannonian Basin. As it is seen from Fig. 3 the Lower Ignimbrite horizon of the Bükk Foreland

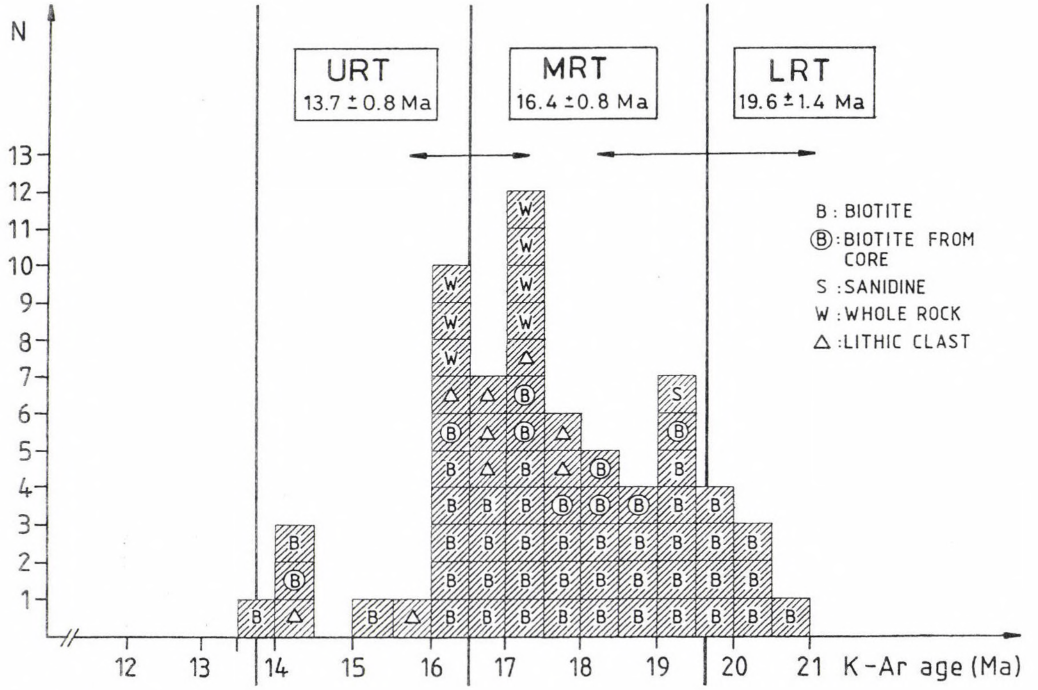

Fig. 3
Histogram summarizing the frequency distribution of ages for the Bükk Foreland, including those measured on subsurface material (cores). URT, MRT and LRT are for the Upper, Middle and Lower Tuffs of the Pannonian Basin outside the Bükk Foreland (data by Hámor et al. 1979)

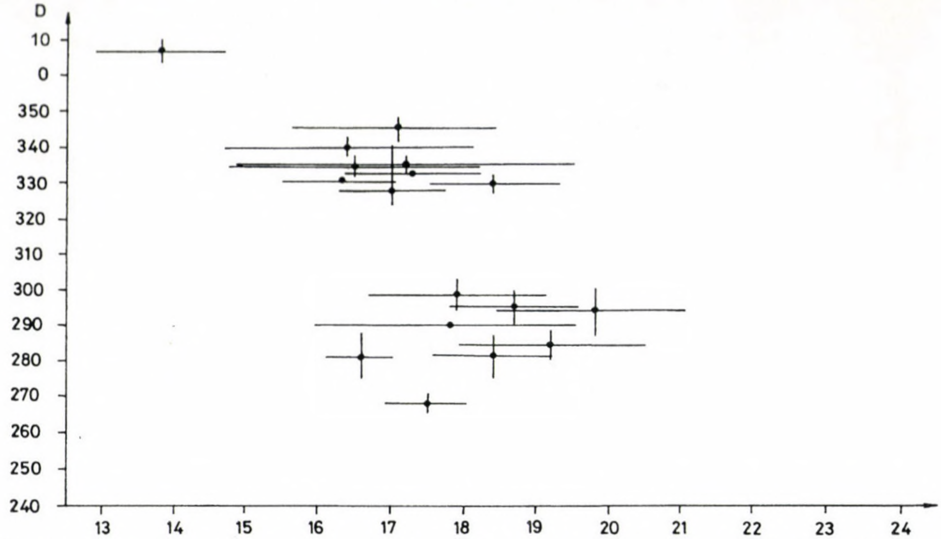

Fig. 4
Paleomagnetic declinations versus K/Ar ages from the Bükk Foreland for the sites where both methods yielded positive results

correlates very well with the Lower Rhyolite Tuffs elsewhere, while the Upper Ignimbrite horizon corresponds in age to the Middle Rhyolite Tuffs. We also found an ignimbrite in the western part of the study area (BZ–25A) which is distinctly younger than the Upper Ignimbrite of the Bükk Foreland and corresponds in age to the Upper Rhyolite Tuff horizon of the Pannonian Basin.

The complex evaluation of the paleomagnetic and radiometric data reveals that the distinct paleomagnetic horizons can be quite precisely dated as 21–18.5 Ma, 17.5–16 Ma and 14.5–13.5 Ma, respectively (Fig. 4). There are three sites only, where paleomagnetic and K/Ar data seem to be in conflict (BZ–32C, BZ–27B, BZ–12). Two of them are easily explained by the thermal influence of the upper ignimbrite volcanism on the K/Ar age, which must have left the paleomagnetic signal unchanged (because of the high Curie points!).

The only occurrence of unexplained incompatibility of the two data sets is BZ–32C, where K/Ar ages suggest "Lower" while paleomagnetism "Upper Ignimbrite".

The main conclusions of the present study are as follows:

1) The ignimbritic volcanism in the Bükk Foreland took place between 21.0–13.5 Ma.

2) Within this interval, three major events can be distinguished; they are sharply separated by their paleomagnetic declinations and less distinctly radiometrically, i.e., the two independent methods employed in combination increases the resolution in the evolution history.

3) The ignimbrite horizons of the Bükk Foreland perfectly match in age the Lower, Middle and Upper Tuffs of the Pannonian Basin elsewhere.

4) From a geodynamic point of view the significance of the successful dating of the ignimbritic horizons (Márton and Márton 1996; Márton et al. 1996) is that the rotations and final emplacement of the Bükk Foreland and related areas are now known with a high precision: the first rotation must have taken place between 18.5–17.5 Ma and the second one between 16.0–14.5 Ma.

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