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The Anisian/Ladinian boundary in the Balaton Highland, Hungary

for the field workshop of the International Triassic Subcommission in the Southern Alps, Italy and in the Balaton Highland, Hungary

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The Middle Triassic events of the Transdanubian Central Range in the frame of the Alpine evolution

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The first, Pelsonian facies differentiation in the Transdanubian Central Range coincided with a global sea level rise but the effects of the local extensional tectonism were decisive. Late Illyrian event (drowning of all carbonate platforms) can be due to sudden tectonic subsidence and to the simultaneous effect of volcanic ash falls. The Late Illyrian–Early Ladinian rhyolitic–trachytic tuffs are widespread whereas the Late Ladinian, intermediate-mafic volcanoclastics seem to be restricted to the eastern part of the TCR.

Further evidences, such as distribution of diagnostic facies and paleobiogeography of brachiopods and ammonites strongly suggest that in the Middle Triassic the Transdanubian Central Range belonged to the southern shelf of the Meliata ocean in close vicinity of the Southern Alps.

Key words: Middle Triassic evolution, facies differentation, volcanism, synsedimentary block tectonics, eustatic sea level change, paleogeographic setting

Introduction

In the Transdanubian Central Range, Middle Triassic formations are known on the surface in the Balaton Highland, in the southern rim of the Veszprém Plateau, in the Eastern Bakony Mts, and in the Buda Hills. Between the Bakony and Buda region as well as in the Northern Bakony Mts only some boreholes provide point-like and sometimes uncertain information, thus, except the southwestern area, relatively small amount of data is available on the Middle Triassic of the Central Range due to the poorly exposed state. This is why a so detailed and reliable paleogeographic evaluation cannot be carried out in our region as in the Alpine territories (De Zanche and Farabegoli 1988, etc.) though a series of events analogous to the Alpine ones have been recognized in the Central Range during the researches of the last years.

In our earlier work we called the attention to the similarities of the Middle Triassic in the Balaton Highland and in the Southern Alps, and to the analogies with the Jurassic in the Bakony Mts (Budai and Vörös 1992).

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We stated that in the Pelsonian lateral facies differentiation can be observed in the Balaton Highland and this was attributed to the disintegration of the former carbonate platform along synsedimentary faults.

The diagnostic sediments and structures that indicate the synsedimentary block tectonics (after Bechstädt et al. 1978) were also listed. The tilting movement of blocks was tried to be proved by the lateral change of thickness of basin sediments.

This time (partly supplementing, partly refining the former statements) we try to answer the following questions regarding the Middle Triassic mobile events of the Central Range:

1) Are there any facies changes in the Alps and in the Balaton Highland that were caused not by synsedimentary tectonics, but rather by eustatic sea level change?

2) Can we, in the Central Range, follow the tendencies recognized in the Triassic magmatism of the Southern Alps, i.e.:

– the Late Illyrian–Early Ladinian acid-to – intermediate alkaline volcanism became more and more basic with increasing calc-alkaline character;

– the magmatites of different chemistry appear in separate belts;

– after the initial (Late Anisian to Early Ladinian) pyroclastic eruptions the volcanism culminated in the Late Ladinian when explosions were interrupted by effusions producing huge amounts of lava material.

3) May the South Alpine Middle Triassic paleogeographic features (synsedimentary faults, lateral facies boundaries, etc.) or the boundaries between magmatic belts serve as a basis for the paleogeographic "fitting" between the Central Range and the Southern Alps?

4) How far the Central Range can be qualified as a transition between the Southern and Northern Alps, considering the Middle Triassic paleogeographic aspects?

Middle Triassic events

1. Facies differentiation

Pelsonian

In the Balaton Highland sharp lateral facies differentiation can be experienced on the broad shelf showing uniform development from the Late Permian to Middle Anisian. The carbonate ramp (Megyehegy Dolomite) was disintegrated in the Pelsonian along lines of about NW–SE direction. This present day direction - taking into correction the rotation angle due to the displacement of the Central Range (see Kázmér and Kovács 1985, Fig. 11) – fairly well corresponds to the NNE–SSW strike of synsedimentary faults demonstrated in the Southern Alps (see De Zanche and Farabegoli 1988, Fig. 7). In the subsided areas basin sediments were deposited (Felsőörs Formation) while in the areas

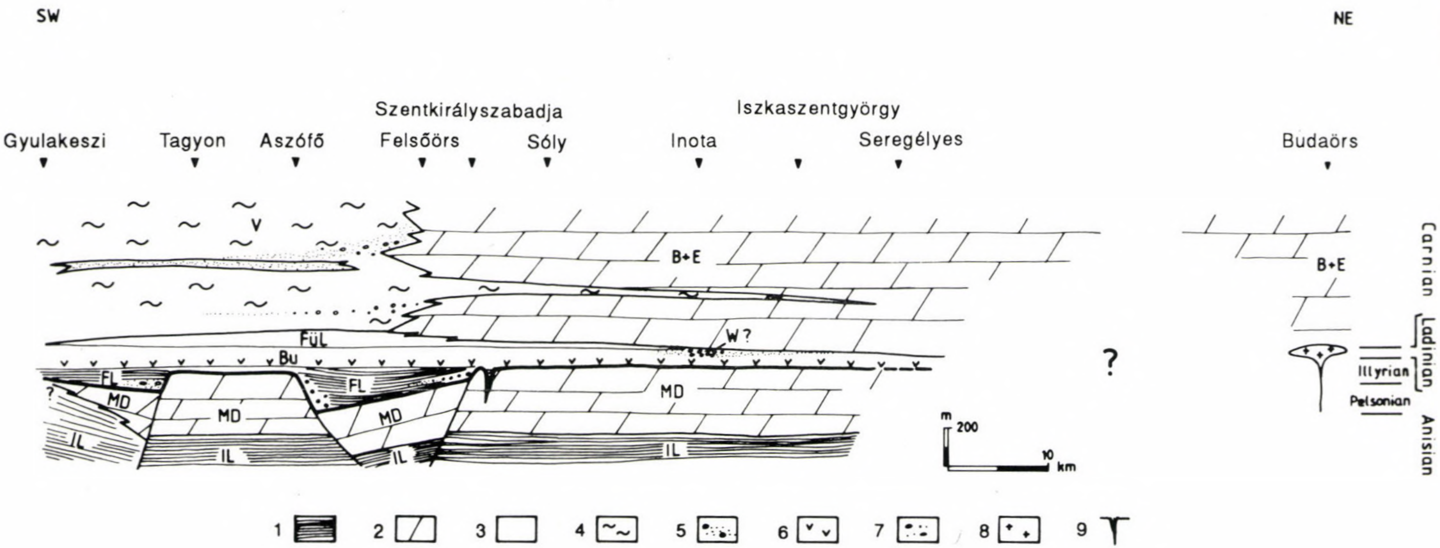

Fig. 1

Lateral distribution and relationship of Middle Triassic formations in the southeastern margin of the Central Range. - 1. restricted (periodically anoxic) internal shelf basin, 2. carbonate platform, 3. open shelf basin, 4. intrashelf basin with terrigeneous clastics; 5. allodapic detritus; 6. pyroclastics; 7. volcanoclastics; 8. lava; 9. neptunian dyke; Anisian: IL – Iszkahegy Limestone; MD – Megyehegy Dolomite + Tagyon Limestone; FL – Felsőörs Limestone; Ladinian: BU – Buchenstein Formation; W? – Wengen Formation; Carnian: FüL – Füred Limestone; V – Veszprém Formation; B + E – Budaörs + Ederics Formations

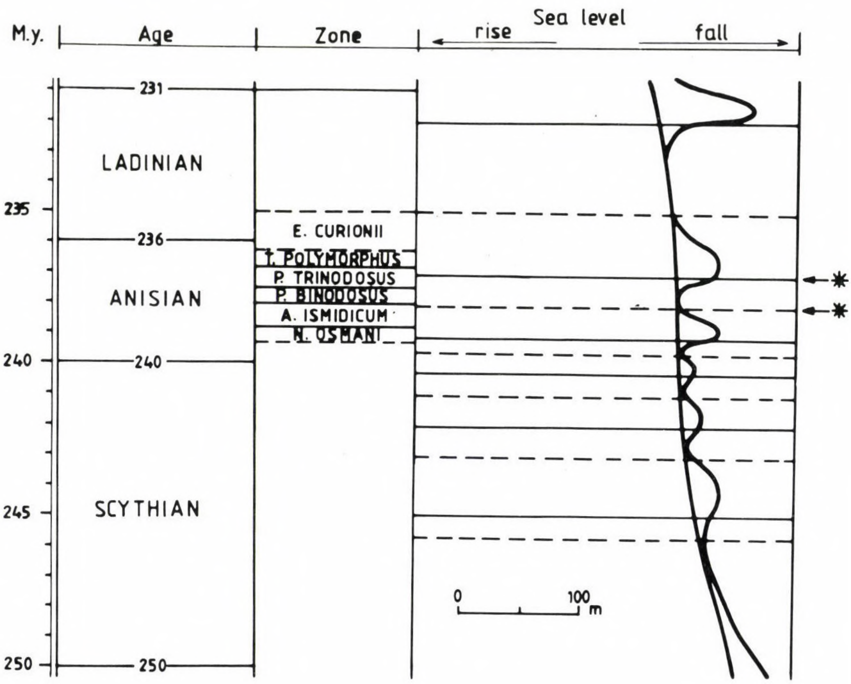

Fig. 2
Lower and Middle Triassic part of the global sea level curve, simplified after Haq et al. 1988. Note: stars on the right margin of the figure denote the phases of synsedimentary block-tectonic movements in the Balaton Highland

remaining in uplifted position the formation of shallow marine carbonates continued (Megyehegy Dolomite, Tagyon Limestone).

Facies differentiation that proceeded during the Pelsonian was produced by synsedimentary block tectonics (Fig. 1) that is proved in addition to sudden and large-scale facies changes by characteristic sediments (e.g. allodapic clasts intercalating several times in the Felsőörs Limestone, lumachelles consisting of redeposited brachiopod shells, etc.) and sedimentary structures (slumping, graded bedding) as well as by tensional fissures (Budai and Vörös 1992). The differentiating effect of the local tectonism could be enhanced by eustatic sea level rise demonstrated in the Pelsonian (see Haq et al. 1988; De Zanche and Farabegoli 1988, Fig. 4).

Data concerning the movements of different blocks, i.e. stratigraphically investigated sections, are of irregular distribution even in the Balaton Highland, and are completely lacking north of this region. Southwest of the central platform of the Balaton Highland (between Tagyon and Aszófő) no well-studied Anisian sections are available. In the old literature the only biostratigraphical reference was made by Böckh (1873, pp. 73–75), that allows to conclude to Pelsonian basin sediments at Köveskál. This is not the situation in the areas lying northeast of the central platform, where biostratigraphically fully investigated sections of the Felsőörs Limestone thinning out from Aszófő to Szentkirályszabadja are available. The lower part of these sequences belong by all means or with great probability to the Pelsonian, independently of the total thickness of the sedimentary sequence at the given point of the basin. The conclusion can be drawn that the area between Aszófő and Szentkirályszabadja subsided synchronously along normal faults and due to a gentle tilting movement, considerably thicker sequence was deposited in its southwestern than its northeastern part (Fig. 1).

Illyrian

Subsequently to the Pelsonian tectonic event the next facies change can be detected in the Illyrian in the area of the central platform of the Balaton Highland. Basin sedimentation started in the Late Illyrian at the end of Trinodosus Zone (e.g. Szentantalfa); beginning of Polymorphus Zone (e.g. Vászoly). In the former case the ammonite-bearing limestone overlies the Tagyon Limestone with sharp and sudden facies change (Budai and Vörös 1992, Fig. 5), whereas in Vászoly it overlies the Megyehegy Dolomite with gradual transition (Vörös and Pálfy 1989). This phenomenon could be interpreted as a result of gradual sea level rise, but this period is characterized by eustatic sea level fall according to Haq et al. (1988; see Fig. 2). Thus, in the Balaton Highland the "complete drowning" of the Late Anisian platform was brought about by tectonic subsidence that was much more intensive than the sea level fall of opposite effect. This is an independent evidence of synsedimentary block tectonics.

It is highly probable that the interruption of shallow marine carbonate sedimentation was partly due to the volcanism started in the Late Illyrian. The effect of this can be experienced also in the areas of which platform carbonates were formed prior and subsequently to the Ladinian (see Fig. 1). In the Balaton Highland no "surviving platforms" are known where carbonate production was not interrupted during volcanism like in Lombardy (Esino Limestone) or in the Dolomites (Schlern Dolomite, Marmolada Limestone). The existence of such platforms cannot be excluded in areas where Middle Triassic sequences are unknown so far (e.g. Northern Bakony, Vértes and Pilis Mts).

2. *Volcanism*

Late Anisian–Early Ladinian

In the region of Late Anisian basins on the Balaton Highland the first indications of Middle Triassic volcanism occur in the middle of the Late Illyrian. In the Forrás Hill section at Felsőörs the lowermost independent tuff layer occurs in the Felsőörs Limestone, beneath the base of the Polymorphus Zone (Budai and Vörös 1992, Fig. 4, called "Avisianum Zone" in that paper). In the area of Late Anisian platforms this tuff horizon is missing (e.g. at Szentantalfa).

Both in the central and in the northeastern platform of the Balaton Highland the first tuff layers were deposited during the Latest Anisian (Polymorphus Zone). Pyroclastics constituting the lower part of the Buchenstein Formation (Lower Pietra Verde) were deposited in much greater thickness in the areas of former basins than in platforms. Consequently, the subsidence of the basement disintegrated in the Pelsonian was differential still in the Latest Anisian and Earliest Ladinian. These differences of the basement morphology were eliminated due to the eustatic sea level rise that can be assigned to the base of the Curionii Zone (see Fig. 2). As a result, the deposition of pelagic limestones started both on platforms (Vászoly Limestone) and in basin areas (Nemesvámos Limestone). In terms of sequence stratigraphy the phosphoritic "hard ground" of the Vászoly Limestone indicates also a sea level rise (oral communication of P. Gianolla, Univ. Padova).

In the Balaton Highland the Upper Anisian to Lower Ladinian volcanics can be detected everywhere and are relatively thick while they are of subordinate significance in the Upper Ladinian. Perhaps, this is why only the former ones were studied from petrographic-geochemical aspects (Szabó and Ravasz 1970; Ravasz 1973). Observations based on the results of these investigations fairly well correlate with those known from the literature of the Southern Alps. The Upper Anisian to Lower Ladinian volcanics are of acid-intermediate chemistry, their composition changes from potash-trachytic to rhyolitic, with the gradual increasing of calk-alkaline character.

The close relationship of volcanic series of the two regions is emphasized by the fact that accessory volcanic quartz mentioned by Szabó and Ravasz (1970, p. 40) from the uppermost part of "Lower Ladinian tuffs" has been considered by Cros and Houel (1983, p. 416) as regionally traceable guide horizon of lithostratigraphic value in the Upper Pietra Verde sequence.

Similarity does not exist in the thickness of the sequences since the Lower Ladinian tuffs of the Balaton Highland are several ten meters thick while those of the Livinallongo Formation in the Dolomites of more complex lithology are 180–200 m thick.

It is our opinion that the great areal distribution of Lower Ladinian pyroclastics may be related to the acid chemistry of the magma, and to the intense explosion activity.

Late Ladinian

The magma became more and more basic during the Late Ladinian and reached the surface as lava flows in the Southern Alps (Rossi et al. 1980). In addition to pyroclastics, pumices and ignimbrites, different lava rocks occurred in the Southern Alps (Cros and Szabó 1984, Fig. 4) and at several localities of the Central Range, e.g. at Inota (Budai et al. 1985) and in the Buda Hills (Horváth and Tari 1987). In the region of the Eastern Bakony the sequence consisting of volcanoclastics (tuff-sandstone, gravel) is known as underlying the Lower Carnian limestone (Raincsák 1980). Based on the petrographic investigation of gravels and on the heavy mineral composition (predominance of opaque iron-titanium minerals), intermediate-to-basic volcanism can be presumed (Budai et al. 1985). Kubovics (1985, p. 150), having studied several boreholes (e.g. Iszkaszentgyörgy-1) concluded that neutral volcanism turned into gabbroidal composition in the Triassic. Parallel with the decrease of silica content of the magma, its viscosity decreased and the violence of explosions also diminished. This may be the explanation of the fact that in areas lying farther from the volcanic centers (e.g. Balaton Highland) the Upper Ladinian volcanics occur in subordinate quantities.

The trend of geochemical change of Ladinian magmatism in the Southern Alps has been in focus of interest and its interpretation has much been debated. According to Castellarin et al. (1980, Fig. 2 and 1988, pp. 79–80) not only temporal but also spatial displacement occurs in the features of magmatism. The old (Late Anisian–Early Ladinian) acid volcanics are widespread in the southern and eastern region of the Southern Alps, northwards these are gradually replaced by the more basic (and younger) Middle Ladinian andesites, then by Upper Ladinian (Early Carnian ?) basalts. These belts are asymmetric with northern polarity, the plate tectonic interpretation of which will not be discussed here (several geologists consider this as evidence of Middle Triassic subduction, see Castellarin et al. 1980, 1988). The question arises, however, whether a similar trend can be recognized within the Central Range from the Balaton Highland to the Buda Hills?

The occurrences of magmatites in the Central Range fall into a narrow belt of about NE–SW strike and it is troublesome to distinguish zones within this. Nevertheless, scarce stratigraphic data suggest that the intermediate-mafic volcanics of the northern regions are younger than the Late Anisian–Early Ladinian acid pyroclastics of the Balaton Highland. In addition to the petrological analogies with the Southern Alps, the lithostratigraphic correlation also suggests that the volcanoclastics with plant remnants of Inota (see Fig. 1) can be correlated with some part of the South-Alpine Upper Ladinian Wengen Group (Budai 1992).

Paleogeographic setting of the Central Range in the Middle Triassic

In order to fit the Transdanubian Central Range and its Middle Triassic, respectively, in a broader paleogeographic picture, some important geological phenomena of this surroundings (Alpine–Carpathian–Dinaric region) have to be selected that may serve as frame of reference. In the region of the Alps–Carpathians–Dinarides two fundamental phenomena can be considered to be decisive.

1. The existence and opening of the Meliata oceanic belt. The remnants of this oceanic belt connected to the "Paleotethys" s.l. in the east but closed westwards (Kovács 1982; Haas et al. 1990) have recently been found also in

the Eastern Alps (Mandl and Ondrejicková 1991, Kozur 1991). Its paleogeographic significance and effect is remarkable from two aspects:

– shelfs lying around it show the characteristic facies polarity of passive continental margins: areas closer to the ocean (outer shelfs) were disintegrated already in the Middle Anisian, were subsided and deep water marls or pelagic limestones were deposited here; in the regions farther from the ocean the platform-type sedimentation continued. Thus, a N–S tendency of deepening appears within the East-Alpine (Austro-Alpine), West Carpathian and Tisza Unite (northern margin of Meliata ocean), whereas the tendency of deepening is of S to N in the Southern Alps and in the Inner-Dinaric regions (southern margin of Meliata ocean).

– The distribution of Late Anisian ammonite and brachiopod faunas shows similar zonality (Vörös 1992; Pálfy 1992), i.e. "Dinaric" faunal elements occur in greater number in the strip directly surrounding the oceanic belt, while in other ares (inner shelf) the predominance of ubiquist "South Alpine" faunal assemblages is characteristic.

2. The Ladinian volcanism that is characteristic only of the regions south of the Meliata ocean and shows decreasing intensity westwards: it is very strong in the Dinaric areas, and still intense in the Dolomites but shows only traces in Lombardy.

It was emphasized that in the Transdanubian Central Range the Middle Triassic formations are known only in a narrow strip (practically only in the Balaton Highland) and from some boreholes and this seriously restricts the possibilities of the paleogeographic reconstruction or recognition of tendencies.

When studying the changes along the strike (i.e. of NE–SW direction) of the Central Range, three tendencies have to be mentioned:

1. In the northeastern part carbonate platforms predominate, southwestwards there are more basinal formations.

2. The Anisian (Megyehegy) platform was disintegrated along lines nearly perpendicular to this NE–SW range.

3. The intensity of Ladinian volcanism (at least the quantity of Upper Ladinian volcanoclastics) decreases southwestwards.

To trace the spatial changes perpendicular to the strike of the Central Range (SE–NW direction) the boreholes drilled at the margin of the Little Hungarian Plain may serve a basis. From among these, the only borehole that provides useful information to solve the problem, is the borehole Bakonyszűcs Bsz. 3. Here a nodular ammonite-bearing limestone appears in the Illyrian, whereas in the Balaton Highland similar facies occur only in the Ladinian. Ammonites found in cores can be assigned to species of Dinaric character, i.e. differing from those of the Balaton Highland (Vörös 1992).

In accordance with the phenomena mentioned above, in the Middle Triassic paleogeographic picture the Transdanubian Central Range Unit can be placed to the southern margin of the Meliata oceanic belt, i.e. in the neighbourhood of the Southern Alps. This is supported by the unambiguous Ladinian volcanism, by the facies polarity from south to north (i.e. to become more pelagic) and by the presence of Dinaric faunal elements appearing in the north.

As regards the more exact arrangement it seems probable that the Transdanubian Central Range unit stretched directly north of the Southern Alps, in a position corresponding partly to Lombardy, partly to the Dolomites. This imagination is supported by the coincidence of fault lines disintegrating the platforms and by the fact that products of Ladinian volcanism (mainly the "Wengen" volcanoclastics) occur in decreasing quantities westwards.

The explanation of the considerable facies and faunal similarity between the Transdanubian Central Range and the Austroalpine units lies in the fact that the two regions were situated close to each other, on the two opposite shelfs of the narrow Meliata ocean. The symmetric position allowed the formation of similar facies, the oceanic belt, closed in the west, did not block the migration of benthic faunal elements, either.

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Redefinition of the Reitzi Zone at its type region (Balaton area, Hungary) as the basal zone of the Ladinian

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The basal part of the Buchenstein Formation corresponding to the Reitzi Zone as introduced by Böckh (1873) has been studied in details in several sections in the type region in the Balaton area. The Reitzi Zone as the basal Ladinian (Norische Stufe at that time) was defined by Mojsisovics (1882) on the basis of an assemblage of characteristic ammonoid species. The new bed-by-bed collections in several sections in the Balaton area have revealed that these characteristic taxa appear in definite faunal horizons. The application of Mojsisovics' original definition means that in the type region the Reitzi Zone is built up by the felsoeoersensis, liepoldti, reitzi and costosus horizons. Since the Reitzi Zone is regarded as the basal zone of the Ladinian, the base of this stage should be drawn at the lowermost, felsoeoersensis horizon.

Key words: Ammonoidea, biostratigraphy, Middle Triassic, Balaton area, Hungary

Introduction

The Reitzi Zone was introduced by Böckh (1873) in the Balaton area for yellowish, siliceous limestones with "*Ceratites*" *reitzi*, "*C.*" *zalaensis* and "*C.*" *boeckhi*, above grey "Reifling" limestones and below red, nodular "Pötschen" limestones.

The understanding of the Reitzi Zone was greatly improved by Mojsisovics (1882) who proved the same faunal level in the Buchenstein beds of the Southern Alps and included a series of newly described ammonoid species into the characteristic assemblage. He placed the "Trachyceras reitzi" zone between the "Ceratites trinodosus" and the "Trachyceras archelaeus" zones to the base of his "Norische Stufe" which was later renamed by Bittner (1892) as Ladinian.

Without going into the details of the subsequent, century-long story of controversies, it can be stated that the content, range and status of Mojsisovics' Reitzi Zone changed several times. Its upper part was duly separated as the Curionii Zone; the Avisianum Zone was used instead of or below it; it was substituted by the "Parakellnerites" and/or the "Nevadites" Zones and consequently, sometimes it was pushed down to the Anisian. All these have been done by Alpine workers but mainly in the frame of academic polemies.

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This uncertainty and confusion might have led Tozer (1967) and Silberling and Tozer (1968) to ignore priority and choose the appearance of the true trachyceratids for drawing the base of the Ladinian. This was a quite reasonable procedure at the time of the North American "emancipation" (Tozer 1984), because this datum is excellent for world-wide correlation. Nevertheless, it is an evolutionary event and does not correspond to the base of the Ladinian stage *per definitionem*. Therefore, for the sake of a "reconciliation" we have to return to the priority and to the type region. Certainly not to the polemies but to the field: to collect new material and to refine biostratigraphy.

Important steps in this direction were done by Rieber (1973) and Brack and Rieber (1986) in the Southern Alps, and a comparative study of important South Alpine and Hungarian sections also appeared (Kovács et al. 1990).

Since the Reitzi Zone was originally defined in the Balaton area, a modern redefinition has to be carried out also here. The critical interval (tuffitic limestones of the lower Buchenstein Formation) is very poorly exposed, therefore the collections were made in excavations, artificial trenches established by the Hungarian Geological Survey in the last decade. Now, in the Balaton area we have five important and a few auxiliary sections straddling the Reitzi Zone, collected bed-by-bed for ammonoids and other fossils (Fig. 1). The results from one of these sections (Vászoly) were rather hastily published (Vörös and Pálfy 1989); in the light of new data and improved knowledge they are revised herein.

The main purpose of this paper is to present the biostratigraphical results obtained on the basis of newly collected ammonoid faunas, and to give a redefinition of the Reitzi Zone in the Balaton area. The taxonomical study is in course, therefore the open nomenclature is applied for some names of genera or species. In spite of the nomenclatural and taxonomical uncertainties, these taxa have well-defined importance and practical use in biostratigraphy.

The sections and their ammonoid fauna

The type locality of the Reitzi Zone and at the same time the most classical Triassic section of the Balaton area is at Felsőörs (known also as Forrás-hegy or Malom-völgy). This was first studied by J. Böckh in 1869 and shortly after (1870) by L. Roth, but their descriptions appeared in a reverse order (Roth 1871, Böckh 1873). Later on, partly equivalent beds with rich ammonoid fauna were found at Hajmáskér (Diener 1900; Arthaber 1903).

In the 1950's and subsequently, I. Szabó made serious efforts in collecting ammonites from the Middle Triassic of the Balaton area; he found important localities at Vászoly and had several shafts and pits dug by workers (see Kovács et al. 1990).

In the last decade, in the course of the detailed geological mapping of the Balaton Highland by the Hungarian Geological Survey, several new localities exposing the "Reitzi Zone" (lower Buchenstein beds) have been found and a

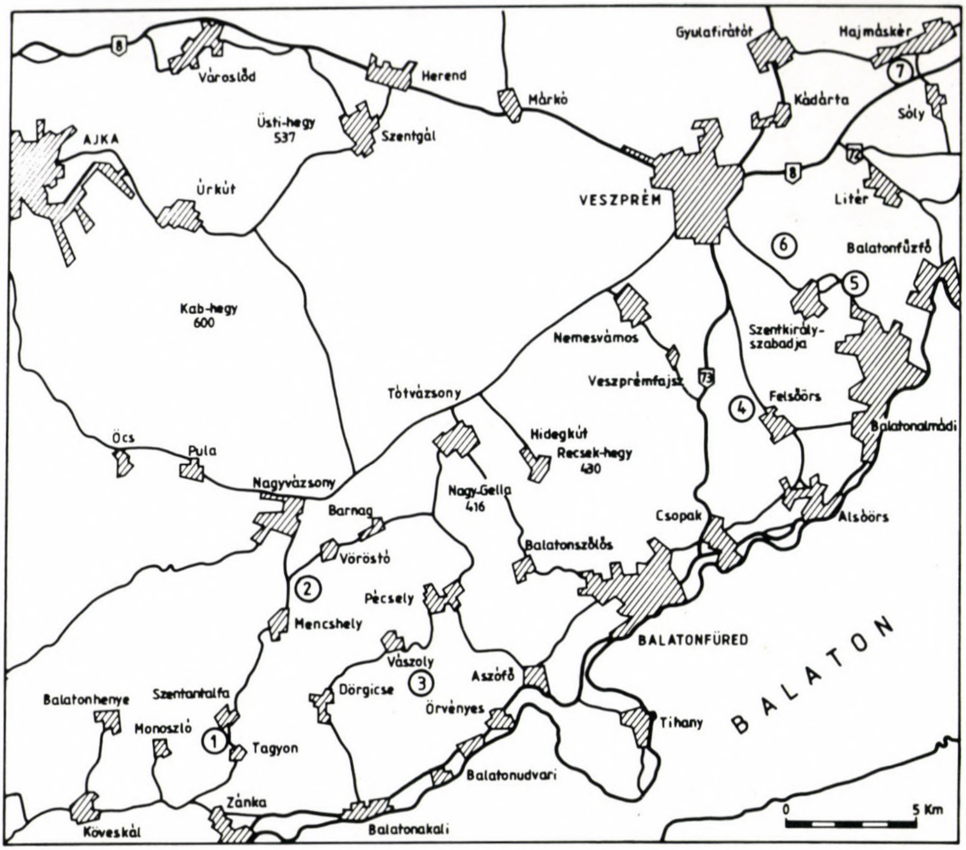

Fig. 1
Location map. 1. Szentantalfa; 2. Mencshely; 3. Vászoly P-11/a and P-2, 4. Felsőörs; 5. Vörösberény; 6. Szentkirályszabadja; 7. Hajmáskér

lot of loose ammonites from this level have been gathered by T. Budai, G. Csillag and L. Koloszár (HGS, Budapest). In the most promising places artificial exposures, trenches have been dug and detailed bed-by-bed collections have been made. In order to have a comprehensive picture, we have made new, detailed collections at the Vászoly and Felsőörs sections as well. The eight studied sections will be shortly presented below.

Szentantalfa

A small trench was dug near the road between the villages Tagyon and Szentantalfa. Here the uppermost part of the Anisian carbonate platform limestone (Tagyon Limestone) is exposed and its sharply truncated top surface is capped by a thin (1.5 m) sequence of brownish-yellow to purplish-red crinoidal and tuffitic, ammonitic limestone beds. These layers are followed by

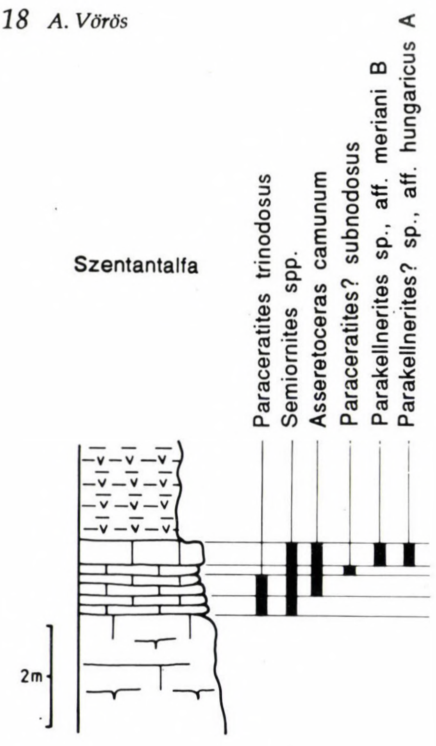

Fig. 2
Stratigraphic column of the Szentantalfa section with the ranges of the diagnostic ammonoid taxa

tuffitic clays. (For a description of the section see Budai and Vörös 1992.) The fossils were collected bed-by-bed by T. Budai, V. Hermann and L. Dosztály in 1989. The most frequent elements of the rich ammonoid fauna (more than 800 specimens, 15 taxa) belong to *Semiornites*, *Paraceratites* and *Parakellnerites*. The vertical ranges of the biostratigraphically most diagnostic taxa are shown in Fig. 2.

Mencshely

This section is exposed in a long (25 m) trench on the flattened northern slope of the Cser-tető Hill, between the villages Mencshely and Vöröstó, in a newly grown pinewood. The lowermost beds belong to the Anisian Felsőörs Formation (grey limestone beds with clay intercalations). This is followed by ash-grey tuffites of the Buchenstein Formation in about 4 m thickness, containing a few, thin (8–10 cm) yellow and grey cherty limestone intercalations. The higher part of the tuffitic sequence becomes pinkish and passes into reddish-bown clay with limestone lumps. These crinoidal, tuffitic limestone lumps yielded a very rich ammonoid fauna. With gradually decreasing amount of clay, massive crinoidal limestone beds appear. The uppermost member of the exposed sequence is light-coloured micritic limestone. The detailed collection was made in 1990 by A. Vörös, P. Vincze, L. Dosztály, I. Szente and I. Főzy. The ammonoid fauna is extremely rich (1200 specimens, 27 taxa); the ranges of the most diagnostic taxa are shown in Fig. 3.

Vászoly

On the Öreg-hegy (Hill) between the villages Vászoly and Pécsely one can find several collapsed shafts and trenches made by I. Szabó in the 1950's for exploratory purposes. Two of these have been restored and collected in details.

The section "P-11/a" is exposed in a deep trench where the lowermost bed is thought to belong to the top of the Anisian Megyehegy Dolomite. It is followed by yellow tuffites alternating with limestone and massive dolomite layers in 2 m thickness. Above this, the yellowish tuffites become dominant

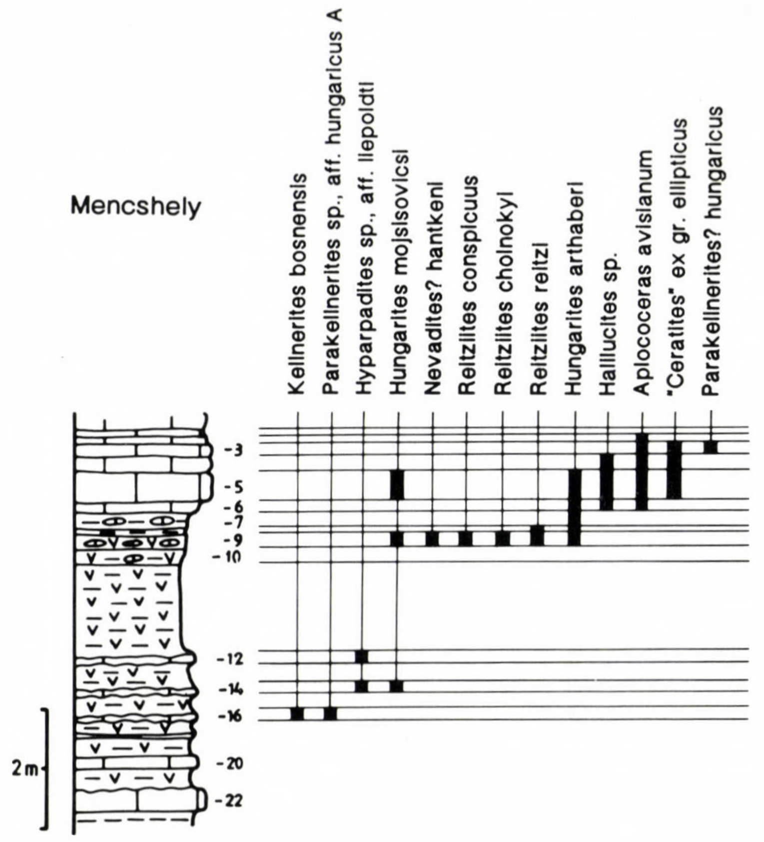

Fig. 3
Stratigraphic column of the Mencshely section with the ranges of the diagnostic ammonoid taxa

and contain only sporadic calcareous lumps. The higher part of this 3 m thick sequence consists of tuffitic clay containing big blocks of yellow crinoidal limestone with plenty of ammonites. The exposed sequence is terminated with massive beds of light-coloured micritic limestone (Vászoly Limestone). (More detailed descriptions of the section are published in Vörös and Pálfy 1989 and Kovács et al. 1990.) The fossils (mainly ammonoids and brachiopods) were collected in 1988 by A. Vörös, J. Pálfy, A. Galácz, M. Kázmér, I. Szente, P. Vincze, L. Bujtor and A. Dulai. Vertical ranges of the most diagnostic taxa of the rich ammonoid fauna (more than 800 specimens, 22 taxa) are shown in Fig 4. (N. B.: the range-chart published by Vörös and Pálfy 1989 is revised and slightly modified here.)

The section "P-2" is a small pit penetrating into the upper part of the reddish clayey tuffitic sequence. Ammonites were found in calcareous lumps and lenses in the clays. With a sharp lithological change the tuffites are overlain by

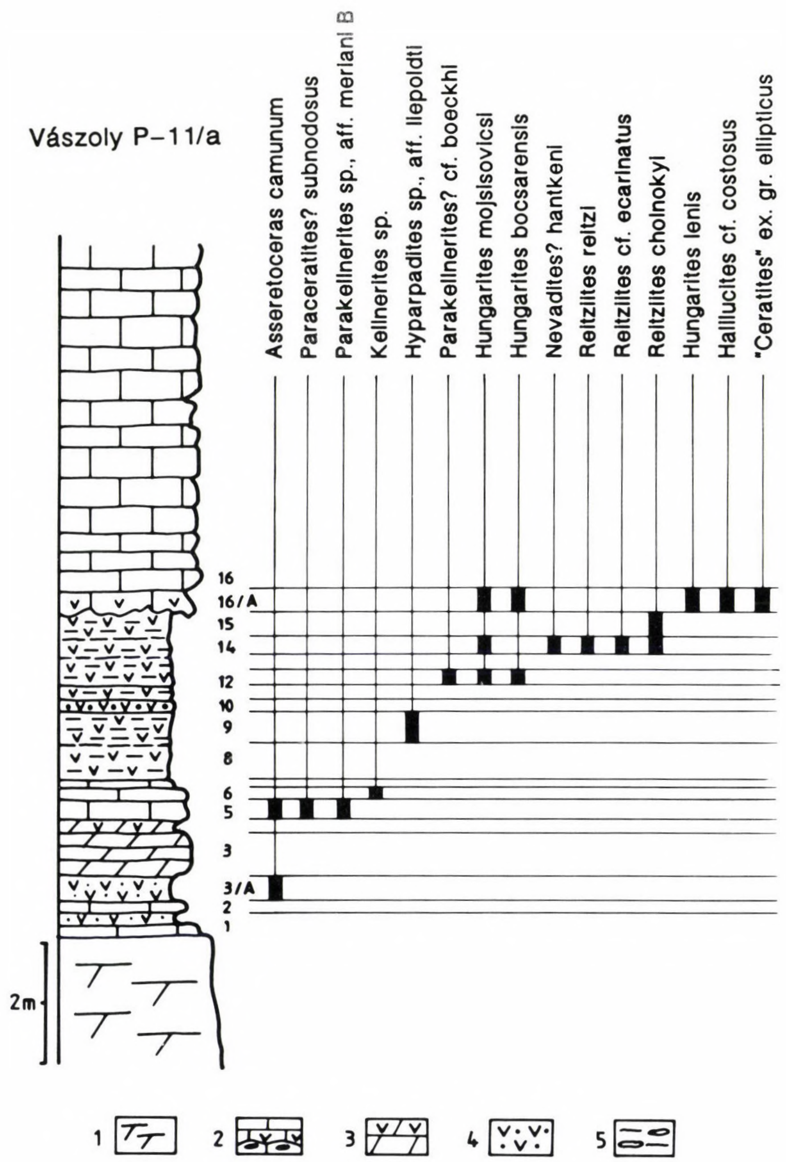

Fig. 4
Stratigraphic column of the Vászoly P-11/a section with the ranges of the diagnostic ammonoid taxa

thick-bedded white micritic limestones (Vászoly Limestone). The poor ammonoid fauna (70 specimens, 8 taxa) was collected by A. Vörös, L. Dosztály, P. Vincze and I. Szente in 1990. The ranges of the few diagnostic taxa are shown in Fig. 5.

Felsőörs

The sequence shown in Fig. 6 is exposed artificially in two parallel trenches on the southwestern slope of the Forrás-hegy (Hill). The lowermost part belongs to the Felsőörs Formation and consists of grey, well-bedded limestone alternating with more or less altered, yellowish tuffitic clays (in about 4 m thickness). Above the uppermost massive limestone layer the tuffites of the Buchenstein Formation become predominant. In this 18 m thick, variegated (dominantly ash-grey, sometimes light-green or violet-grey) tuffitic complex only thin and nodular ochre-yellow cherty limestone interbeds can be found. The uppermost member of the exposed sequence is a light-coloured, nodular, cherty limestone which still contain some tuffitic clay seams. (Earlier descriptions of the section can be found in Szabó et al. 1980 and in Kovács et al. 1990). The new, bed-by-bed collection was made in 1989 by I. Szabó, A. Vörös, P. Vincze, L. Dosztály, I. Szente and L. Bujtor. The ranges of the diagnostic taxa of the rich ammonoid fauna (nearly 700 specimens, 22 taxa) are shown in Fig 6. This compilation is based on the new collections in 1989. The data of the older collections, presented in Kovács et al. (1990) were partly revised and some of them (e.g. the occurrences of the representatives of the "reitzi-group" in beds 99 and 100) were ruled out.

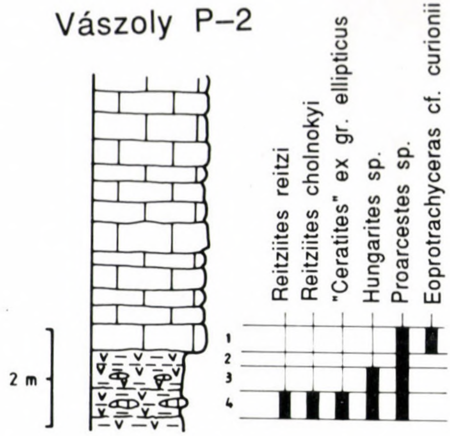

Fig. 5
Stratigraphic column of the Vászoly P-2 section with the ranges of the diagnostic ammonoid taxa

Vörösberény

This section is a road-cut between Vörösberény (Balatonalmádi) and Szentkirályszabadja. Here the uppermost part of the Felsőörs Formation (limestone beds with clay seams) and the lower part of the tuffitic Buchenstein Formation is exposed in about 10 m thickness. Bed-by-bed bulk samples have been collected by the workers of the Hungarian Geological Survey in 1987, the detailed elaboration of the ammonites was done by A. Vörös and V. Hermann. The rich ammonoid fauna (more than 1000 specimens, 16 taxa) was concentrated in the Felsőörs Limestone, the Buchenstein beds were almost barren. The ranges of the diagnostic taxa are shown in Fig. 7.

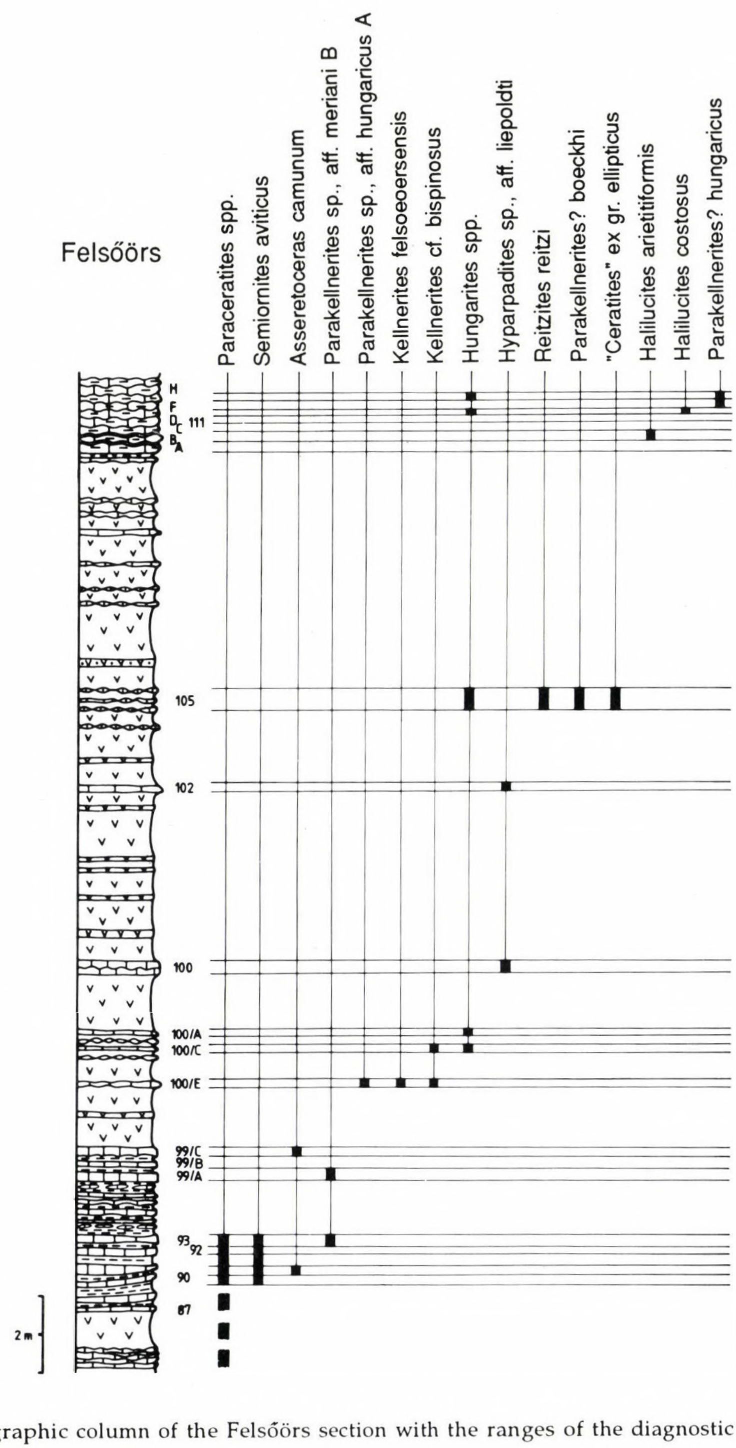

Fig. 6
Stratigraphic column of the Felsőörs section with the ranges of the diagnostic ammonoid taxa

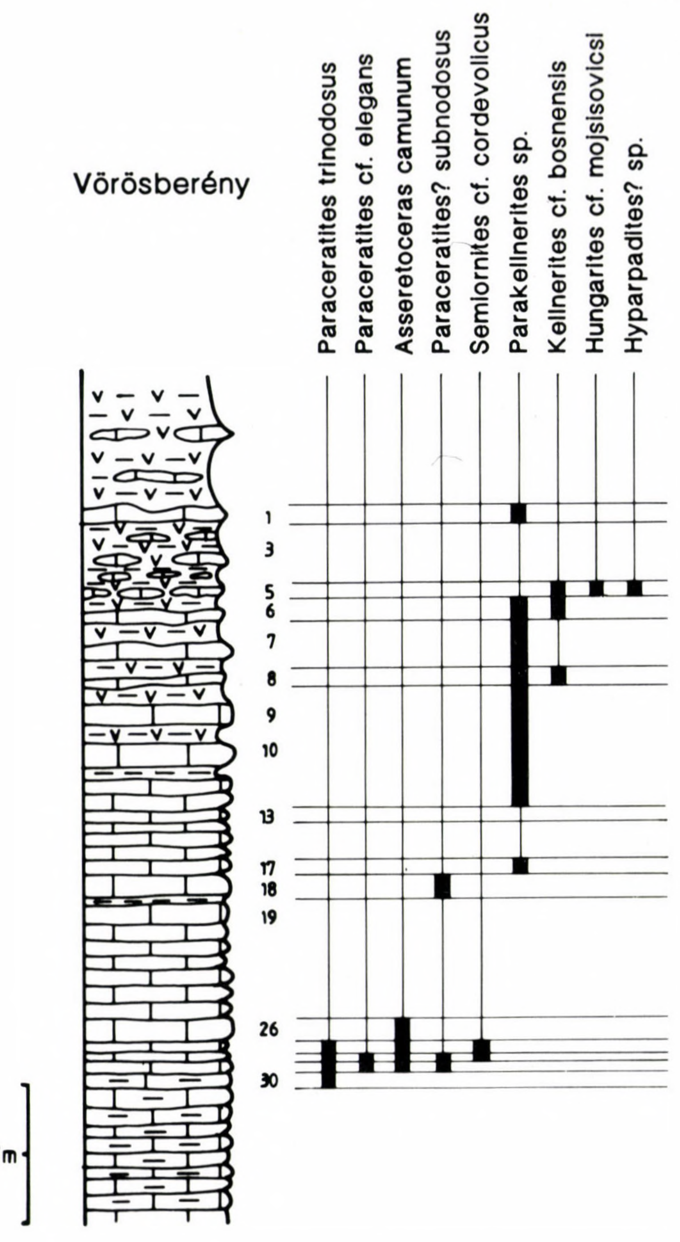

Fig. 7
Stratigraphic column of the Vörösberény section with the ranges of the diagnostic ammonoid taxa

Hajmáskér

The borehole Hmt-3 penetrated variegated limestones in the interval 112.7–113.8 m, pinkish-yellow, crinoidal, tuffitic limestone between 113.8 and 115.8 m and then red and green tuff. The author collected ammonites from the crinoidal, tuffitic limestones in 1980. Despite the small diameter of the cores, 40 specimens and 5 taxa were found. The ranges of the diagnostic taxa are shown in Fig. 8.

Szentkirályszabadja

The fossiliferous beds are exposed in a pit used formerly for military purposes. The pit was extended to dip-direction by a narrow trench. The lower part of the sequence consists of thick dolomite beds alternating with yellow clays. Higher up the dolomite becomes thin-bedded and crumbly and contains volcanoclastic admixture; then alternates with limestone, but the crumbly and tuffitic character remains constant throughout the sequence. The uppermost beds are again massive dolomites but since they contain a few ammonite "ghosts" they must have been pelagic limestones dolomitized secondarily. The rich ammonoid fauna (750 specimens, 18 taxa) was collected in 1991 by A. Vörös, L. Dosztály and P. Vincze. The ranges of the most diagnostic taxa are shown in Fig. 9.

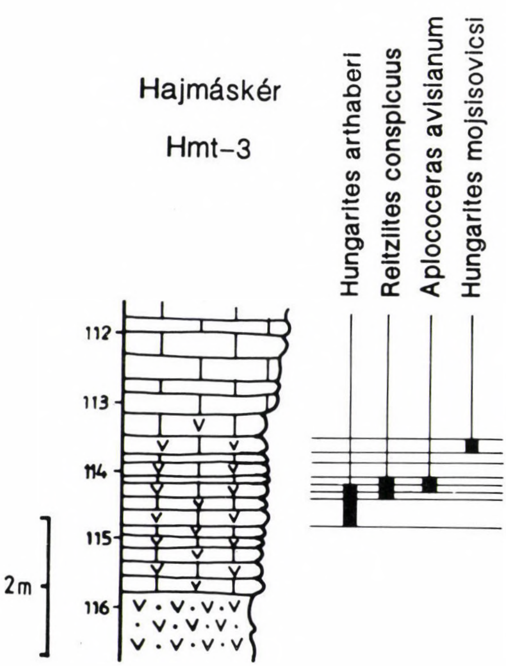

Fig. 8
Stratigraphic column of the Hajmáskér Hmt-3 section with the ranges of the diagnostic ammonoid taxa

Faunal horizons

The stratigraphical sequences described above can be correlated on the basis of diagnostic ammonoid taxa. This correlation is shown in Fig. 10. In the studied area the following seven faunal horizons can be recognized from top to bottom:

Curionii horizon. Until now it is proved only in the Vászoly sections, in the lower part of the white pelagic limestone sequence (Vászoly Limestone). In its poor fauna *Eoprotrachyceras curionii* (Plate VI: 5) is the single diagnostic taxon. *Proarcestes* are relatively frequent but this may partly due to the increasingly pelagic environment.

Costosus horizon. It seems to be restricted to the lowermost beds of the crinoidal or cherty limestone complex directly overlying the tuffitic sequences in the Mencshely, Vászoly and Felsőörs sections. It can be characterized by the occurrence of the following species:

Halilucites costosus (Plate VI: 3)
Halilucites arietitiformis (Plate VI: 4)
Parakellnerites ? hungaricus. (Plate VI: 1, 2)
Proarcestes sp. (Plate V: 1)

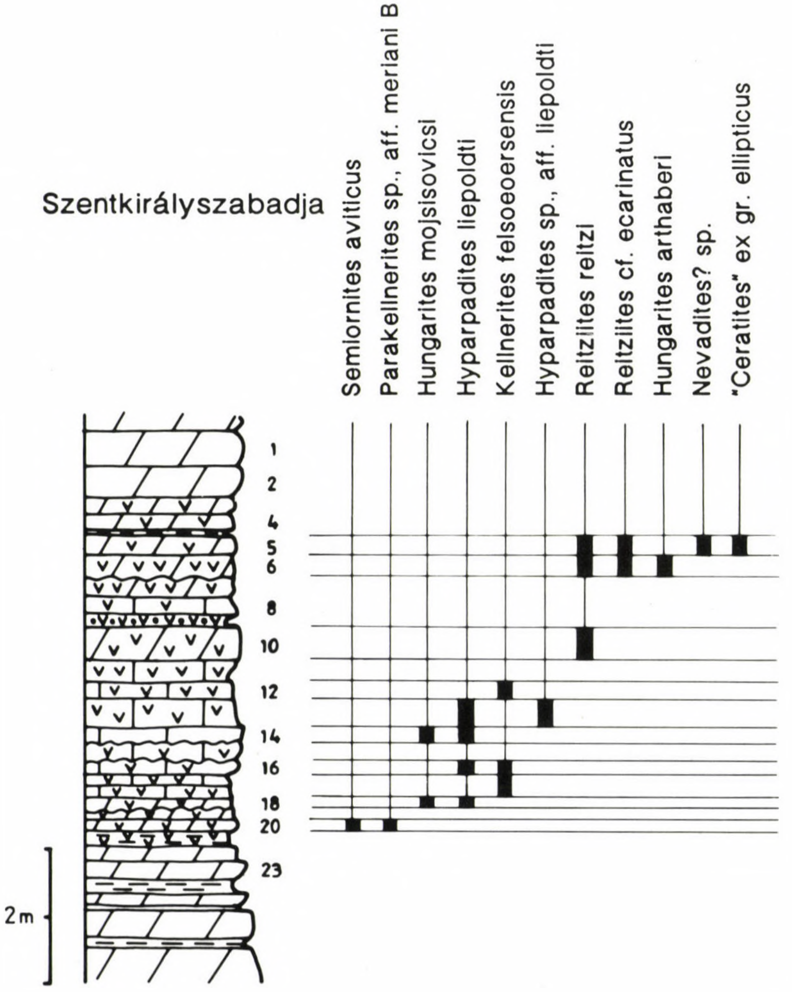

Fig. 9
Stratigraphic column of the Szentkirályszabadja section with the ranges of the diagnostic ammonoid taxa

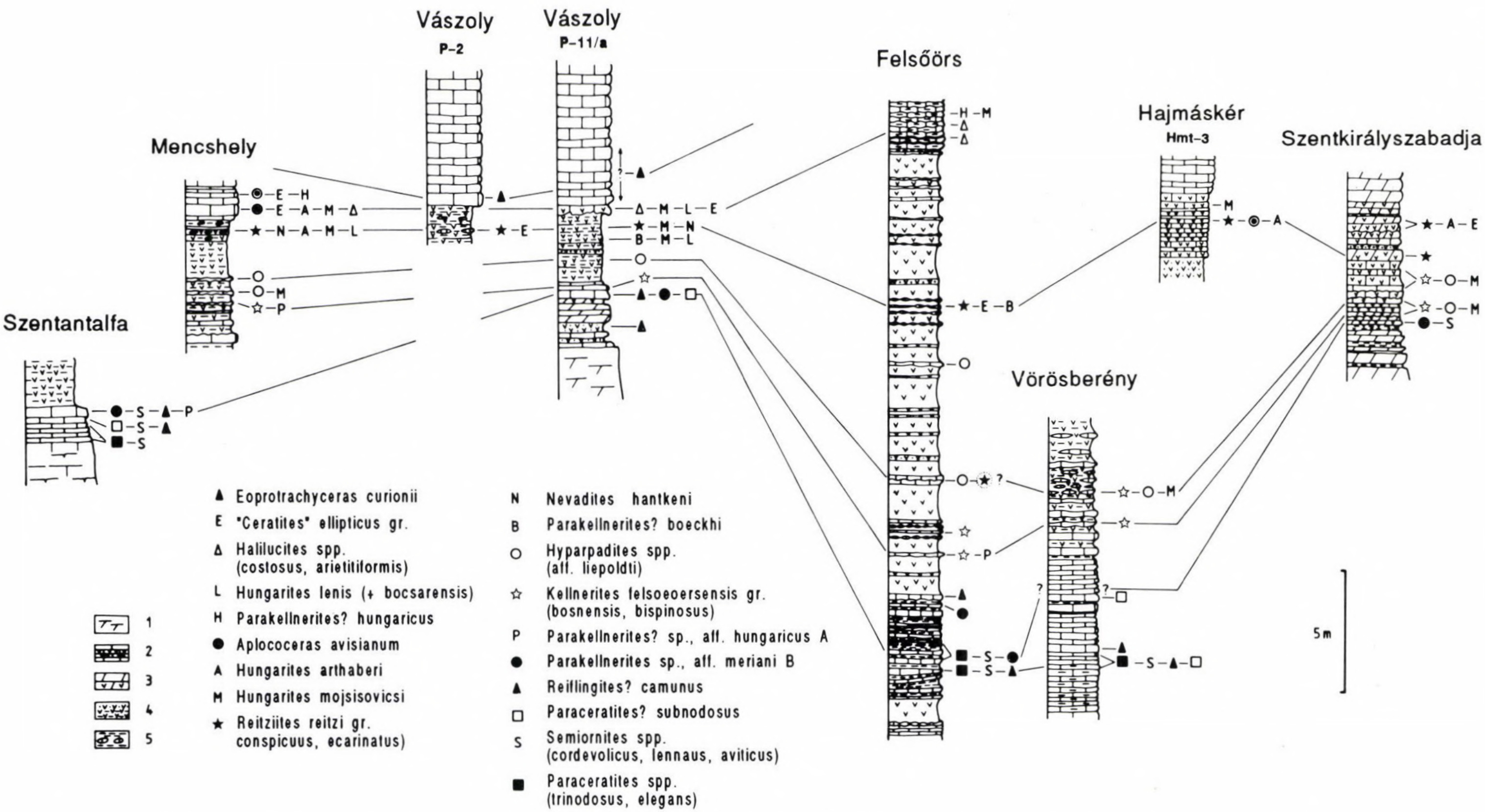

Fig. 10
Biostratigraphic correlation between studied sections by appearances of diagnostic ammonoid taxa and the faunal horizons. 1. massive platform limestones, dolomites; 2. limestone, tuffitic, cherty, nodular; 3. dolomite, tuffitic; 4. tuffitic sand, tuffitic clay; 5. clay, calcareous lumps/nodules

Further important elements of the assemblage are "*Ceratites*" ex. gr. *ellipticus*, *Hungarites lenis*, *H. bocsarensis*, *H. arthaberi* and *Aplococeras avisianum*, though these occur in the underlying retzi horizon as well.

Reitzi horizon. The new bed-bed collections revealed that the appearance of the members of the "*reitzi*-group" (*Reitziites reitzi, ecarinatus, conspicuus, cholnokyi*) forms a well-defined level in the upper third of the dominantly tuffitic complex of the basal Buchenstein Formation. Besides *Reitziites*, the occurrence of *Nevadites ? hantkeni* is also characteristic. The diagnostic assemblage is the following:

Reitziites reitzi (Plate III: 2, 3, 4)
Reitziites ecarinatus (Plate III: 7)
Reitziites conspicuus
Reitziites cholnokyi (Plate III: 5, 6)
Nevadites ? hantkeni (Plate III: 8, Plate IV: 1, 2)
Hungarites lenis (Plate IV: 4)
Hungarites bocsarensis (Plate V: 4)
Hungarites arthaberi (Plate IV: 3)
Hungarites mojsisovicsi (Plate V: 3)
Aplococeras avisianum (Plate IV: 5, 6)
"Ceratites" ex gr. ellipticus (Plate V: 2)

Furthermore, curious *Kellnerites* with *Reitziites*-like ornamentation but still with ventral keel (e.g. *Kellnerites angustaecarinatus*, Plate II: 4) occur. Remarkably, *Aplococeras avisianum* was found in the reitzi horizon or higher.

Liepoldti horizon. It can be followed rather well in the middle or lower half of the tuffitic complex. *Hyparpadites liepoldti* (Plate II: 5) occurs also in the subjacent horizon; the really characteristic ammonites form a group of uncleared taxonomy, closely related to *H. liepoldti* (e.g. Plate III: 1). They have a general *Parakellnerites*-like shape but bear four rows of nodes; in some cases with very faint, dense ribbing.

Felsoeoersensis horizon. It is a very well-recognizable level in the lowermost limestone intercalations in the tuffitic complex. The horizon is characterized by the following taxa:

Kellnerites felsoeoersensis (Plate II: 1, 2)
Kellnerites bosnensis (Plate II: 3)
Kellnerites bispinosus
Parakellnerites sp., aff. *hungaricus* A. (Plate I: 6, 7)

Meriani B horizon. It comprises the uppermost part of the Felsőörs Limestone Formation including the beds where the limestone contains tuffitic intercalations. Its fauna is characterized by *Parakellnerites* species (closely allied to *P. meriani* and/or *frauenfelderi*):

Parakellnerites sp., aff. *merianii* B (Plate I: 8, 9)
Paraceratites? subnodosus (Plate I: 5)
Semiornites aviticus (Plate I: 3, 4)

The latter two forms range up from the underlying horizon.

Camunum horizon. It was found in the deeper levels of the "transitional" beds (Felsőörs Limestone alternating with tuffites) or, in the "condensed" sequences (Szentantalfa, Vászoly) in the beds directly overlying the platform carbonates. Characteristic elements are:

Asseretoceras camunum (Plate I: 1, 2)
Paraceratites elegans
Semiornites aviticus (Plate I: 3, 4)
Semiornites lennanus
Semiornites cordevolicus
Paraceratites? subnodosus (Plate I: 5)

Paraceratites trinodosus ranges up also. The lower boundary of this horizon is poorly known because the detailed collections did not reach deeper levels.

The ranges of the biostratigraphically diagnostic ammonoid species plotted against the faunal horizons are shown in Fig. 11. These well-defined faunal horizons are the elementary units of our biostratigraphical subdivision, and from these we may construct biozones.

Redefinition of the Reitzi Zone

For defining the Reitzi Zone we have to turn back again to Mojsisovics (1882) who listed 16 species from the "yellow, cherty limestones of the Bakony Forest" to characterize the zone. From among these species, 11 belong to the biostratigraphically diagnostic Ceratitidae and, with one exception, these were found again in the new, detailed collections. The 11 species are listed below, quoted from Mojsisovics (1882) and in the other column the corresponding names, used in the present study, are given.

Mojsisovics (1882)	present study
Ceratites hungaricus	*Parakellnerites ? hungaricus*
Ceratites aff. *hungarico*	*Kellnerites* sp.
Ceratites Felsö-Örsensis	*Kellnerites felsoeoersensis*
Ceratites Boeckhi	*Parakellnerites ? boeckhi*
Ceratites Hantkeni	*Nevadites ? hantkeni*
Ceratites Zezianus	(not found)
Arpadites Liepoldti	*Hyparpadites liepoldti*
Trachyceras Reitzi	*Reitziites reitzi*
Longobardites Zsigmondyi	*Longobardites zsigmondyi*
Hungarites Mojsisovicsi	*Hungarites mojsisovicsi*
Hungarites costosus	*Halilucites costosus*

If we compare this list with the ranges presented in Fig. 11, a very clear picture emerges: the Reitzi Zone is defined as embracing the felsoeoersensis, liepoldti, reitzi and costosus horizons. If this definition needs any further discussion, we may state the following: The curionii horizon is clearly out of question. The costosus horizon is represented in the list by *Halilucites costosus* and *Parakellnerites? hungaricus*; the reitzi horizon by *Reitziites reitzii*, *Nevadites? hantkeni* and *Parakellnerites? boeckhi*; the liepoldti horizon by *Hyparpadites liepoldti* and the

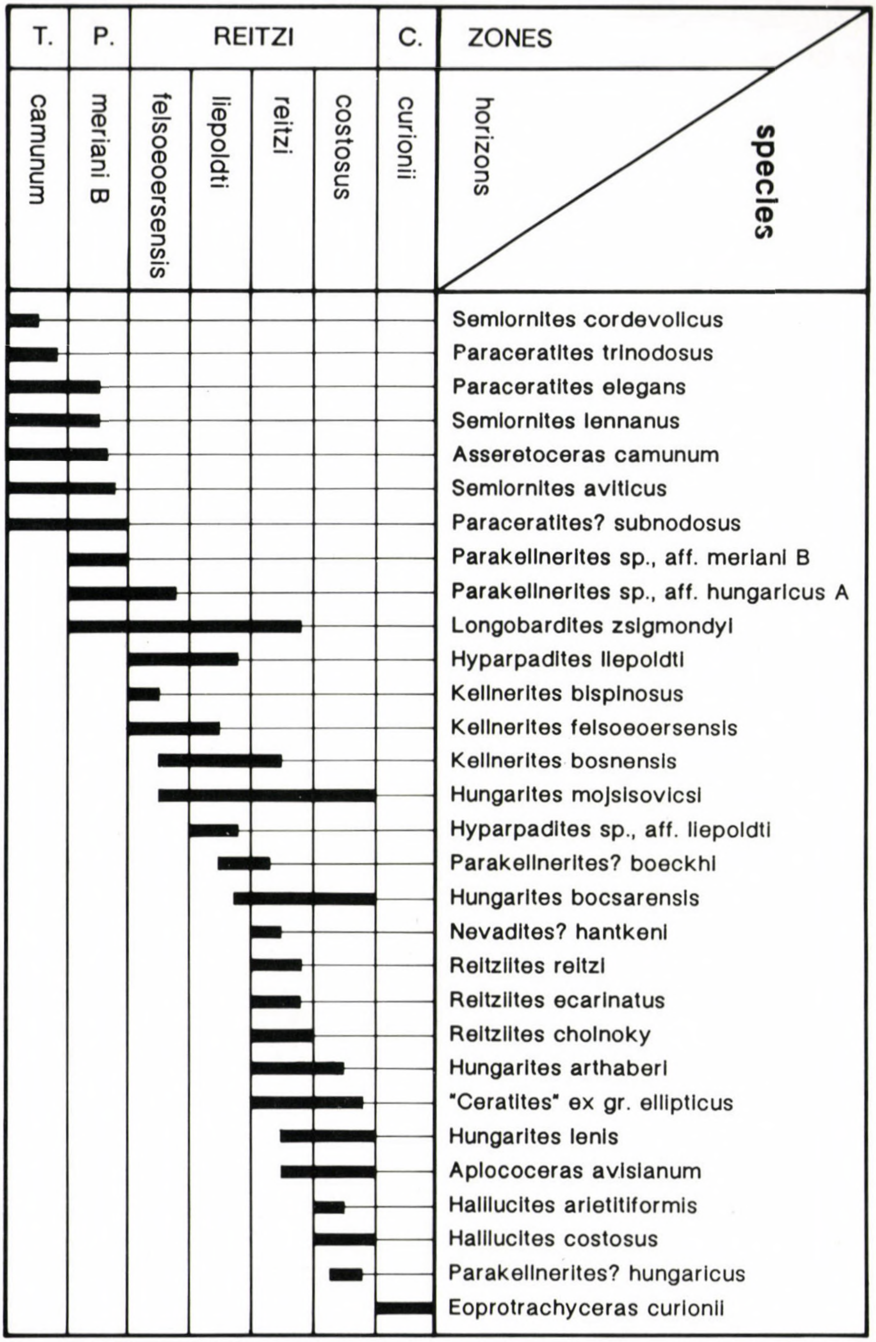

Fig. 11
The ranges of the diagnostic ammonoid taxa plotted against the faunal horizons

felsoeoersensis horizon by *Kellnerites felsoeoersensis* and *Kellnerites* sp. The only species passing up from the meriani B horizon is *Longobardites zsigmondyi* but this is (together with *Hungarites mojsisovicsi*) a long-ranging form of low biostratigraphical value. The application of Mojsisovics' original definition means that the base of the Reitzi Zone has to be drawn at the lowermost, felsoeoersensis horizon.

The meriani B and the camunum horizons can be ranged into the Trinodosus Zone, or, if someone wishes to distinguish a Polymorphus Zone on the top of the Anisian, it seems to be represented solely by the meriani B horizon in the Balaton area.

In conclusion, the faunal content, the subdivision and the base of the Reitzi Zone is well defined in the type region. Its status as the basal zone of the Ladinian stage can be maintained.

Acknowledgements

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Plate I

All figures in natural size. Photographs by Mrs. L. Pellérdy.

1. *Asseretoceras camunum*. Felsőörs, bed 90, camunum horizon; 2. *Asseretoceras camunum*. Szentantalfa, bed 5, meriani B horizon, 3. *Semiornites aviticus*. Felsőörs, bed 93, meriani B horizon; 4. *Semiornites aviticus*. Szentkirályszabadja, bed 20, meriani B horizon; 5. *Paraceratites? subnodosus*. Szentantalfa, bed 2/A, camunum horizon; 6. *Parakellnerites?* sp., aff. *hungaricus* A. Felsőörs, bed 100/E, felsoeoersensis horizon; 7. *Parakellnerites*? sp., aff. *hungaricus* A. Mencshely, bed -16, felsoeoersensis horizon; 8. *Parakellnerites* sp., aff. *meriani* B. Felsőörs, bed 99/A, meriani B horizon; 9. *Parakellnerites* sp., aff. *meriani* B. Vászoly P-11/a, bed 5, meriani B horizon

Plate II

All figures in natural size. Photographs by Mrs. L. Pellérdy.

1. *Kellnerites felsoeoersensis*. Felsőörs, bed 100/E, felsoeoersensis horizon; 2. *Kellnerites felsoeoersensis*. Szentkirályszabadja, bed 12, liepoldti horizon. 2a. lateral view, 2b. ventral view; 3. *Kellnerites bosnensis*. Mencshely, bed -16, felsoeoersensis horizon; 4. *Kellnerites angustaecarinatus*. Mencshely, bed -9, reitzi horizon; 5. *Hyparpadites liepoldti*. Szentkirályszabadja, bed 16, felsoeoersensis horizon

Plate III

All figures in natural size. Photographs by Mrs. L. Pellérdy.

1. *Hyparpadites* sp., aff. *liepoldti*. Szentkirályszabadja, bed 13, liepoldti horizon; 2. *Reitziites reitzi*. Felsőörs, bed 105, reitzi horizon; 3. *Reitziites reitzi*. Mencshely, bed -9, reitzi horizon; 4. *Reitziites reitzi*. Mencshely, bed -9, reitzi horizon; 5. *Reitziites cholnokyi*. Mencshely, bed -9, reitzi horizon; 6. *Reitziites cholnokyi*. Mencshely, bed -9, reitzi horizon; 7. *Reitziites ecarinatus*, Mencshely, bed -9, reitzi horizon; 8. *Nevadites? hantkeni*. Vászoly, bed 14, reitzi horizon. 8a. lateral view, 8b. ventral view

Plate IV

All figures (except 1) in natural size. Photographs by Mrs. L. Pellérdy.

1. *Nevadites? hantkeni*. Mencshely, bed -9, reitzi horizon. 1a. lateral view, 1b. ventral view, 1c. dorsal view of the last whorl after removing the inner whorls. (X 1.5); 2. *Nevadites? hantkeni* (same last whorl as in the previous figure); 3. *Hungarites arthaberi*. Mencshely, bed -9, reitzi horizon; 4. *Hungarites lenis*. Mencshely, bed -9, reitzi horizon; 5. *Aplococeras avisianum*. Mencshely, bed -2, costosus horizon; 6. *Aplococeras avisianum*. Mencshely, bed -6, costosus horizon

Plate V

All figures in natural size. Photographs by Mrs. L Pellérdy.

1. *Proarcestes* sp. Mencshely, bed -5, costosus horizon; 2. "*Ceratites*" ex gr. *ellipticus*. Vászoly P-11/a, bed 16/A, costosus horizon; 3. *Hungarites mojsisovicsi*. Vászoly P-11/a, bed 16/A, costosus horizon; 4. *Hungarites bocsarensis*. Vászoly P-11/a, bed 16/A, costosus horizon; 5. *Halilucites* cf. *costosus*. Vászoly P-11/a, bed 16/A, costosus horizon

Plate VI

All figures in natural size. Photographs by Mrs. L. Pellérdy.

1. *Parakellnerites? hungaricus*. Felsőörs, bed 111/E, costosus horizon; 2. *Parakellnerites? hungaricus*. Felsőörs, bed 111/F, costosus horizon; 3. *Halilucites costosus*. Felsőörs, bed 111/E, costosus horizon; 4. *Halilucites arietitiformis*. Felsőörs, bed 111/B, costosus horizon; 5. *Eoprotrachyceras* cf. *curionii*. Vászoly, loose, from the Vászoly Limestone, curionii horizon

Plate I

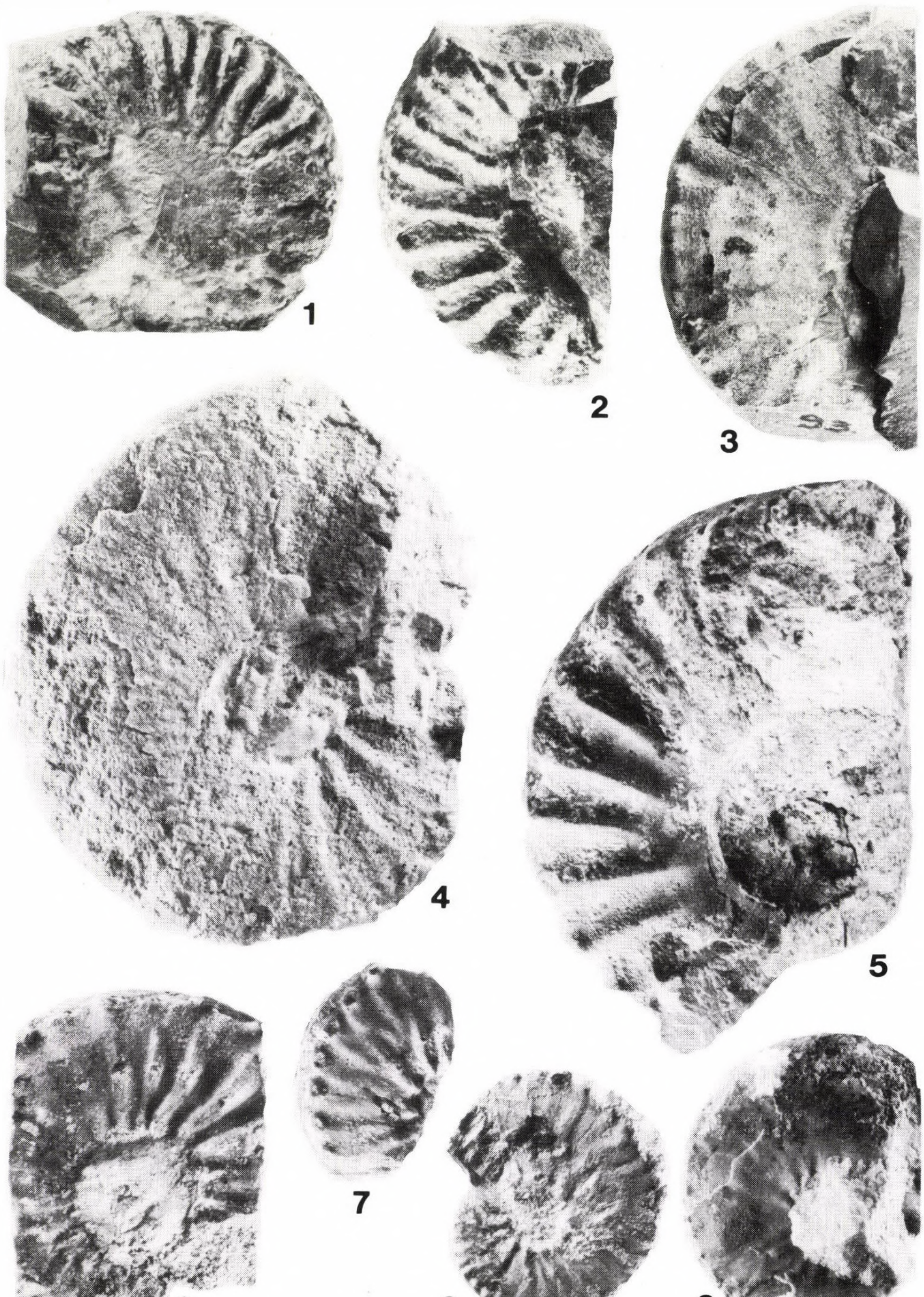

Plate II

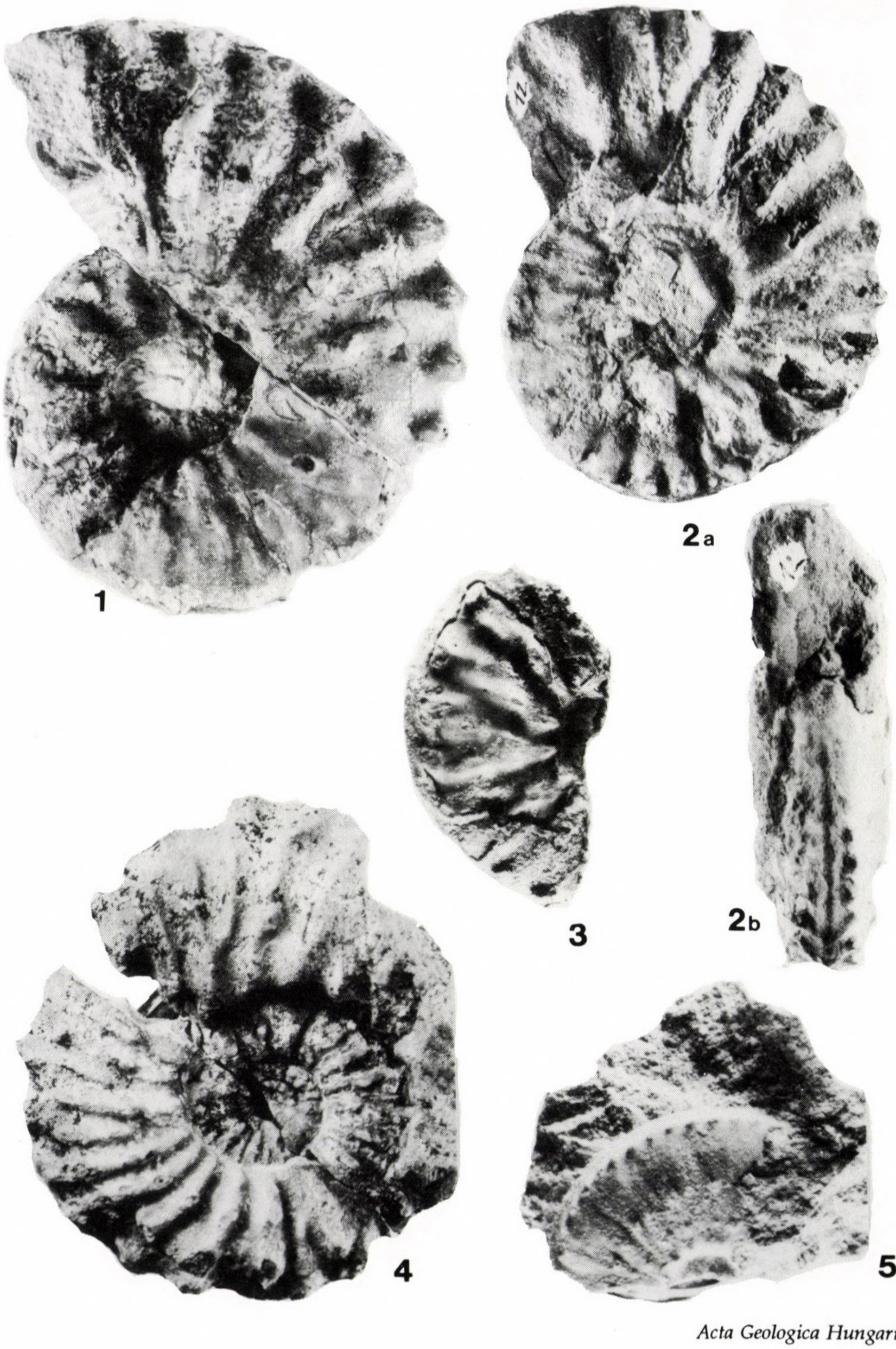

Plate III

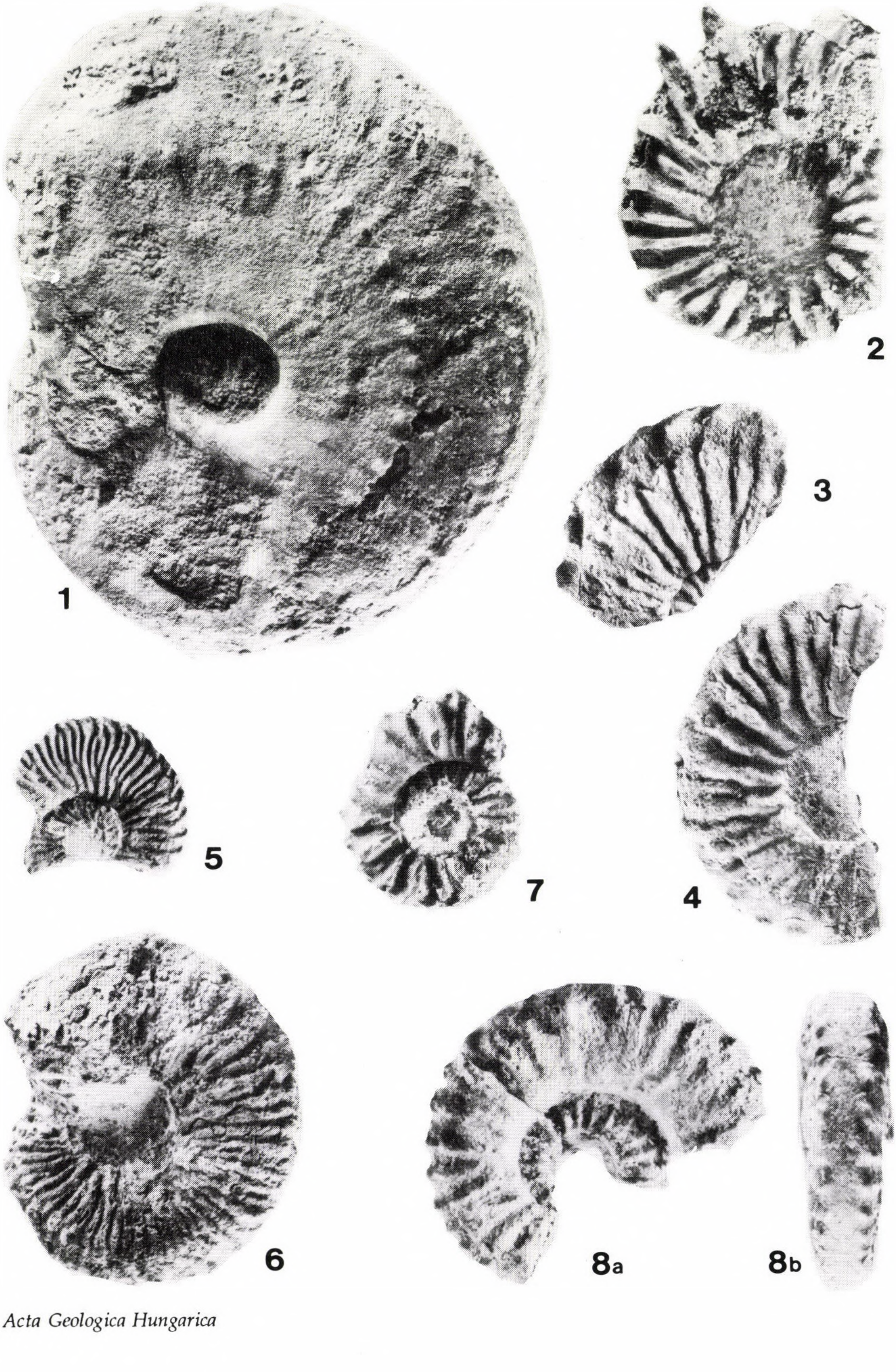

Plate IV

Plate V

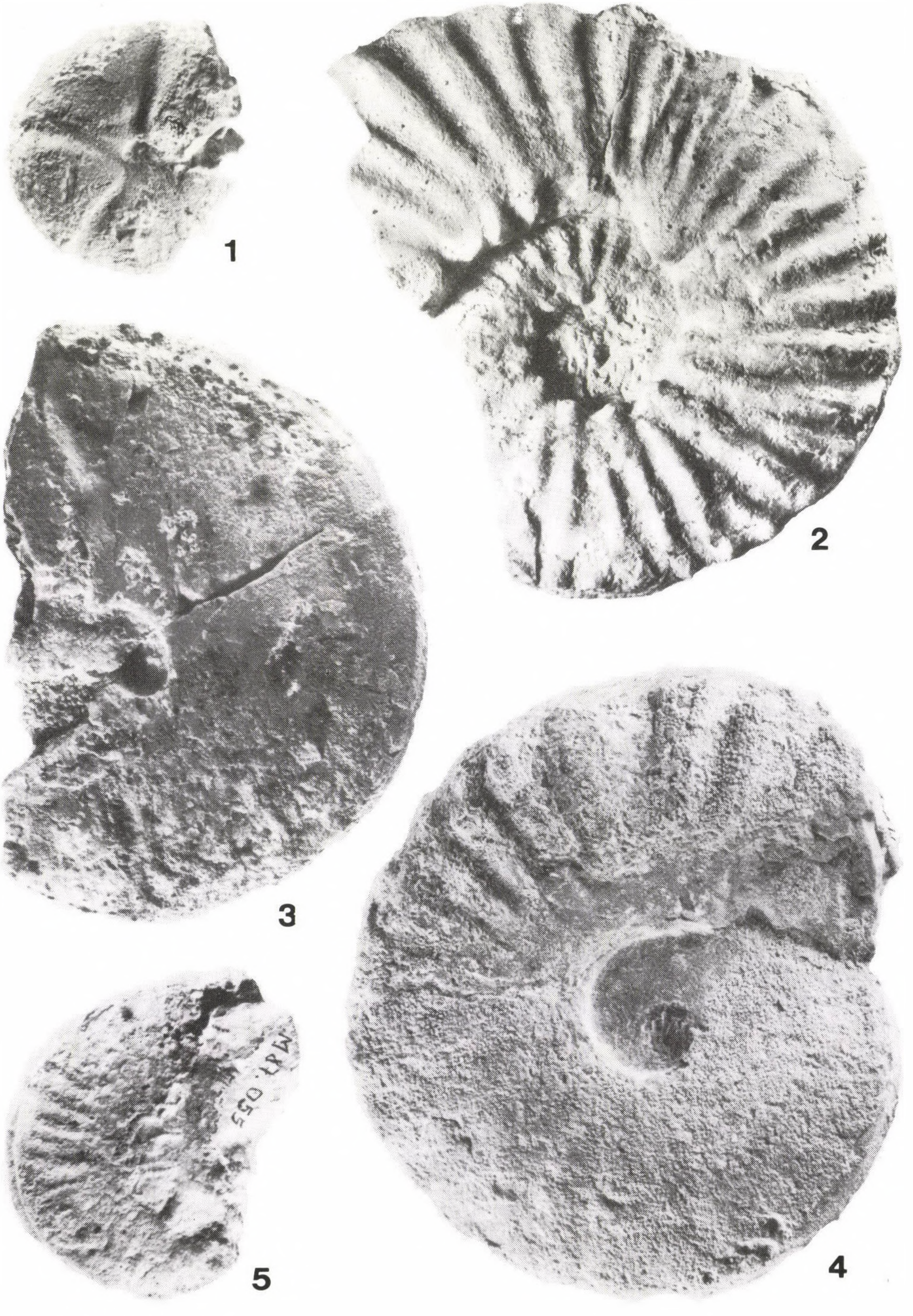

Plate VI

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Conodont biostratigraphy of the Anisian/ Ladinian boundary interval in the Balaton Highland, Hungary and its significance in the definition of the boundary (Preliminary report)

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Bed-by-bed investigation of the Felsőörs, Mencshely and Vászoly sections in the Balaton Highland (Hungary) allowed to recognize three main events of conodont evolution in the Anisian/Ladinian boundary interval:

1. Appearence of *Gondolella constricta postcornuta* near to the base of the Reitzi Zone s.l. (slightly below the *Kellnerites felsoeoersensis* horizon); 2. Appearence of *Gondolella trammeri* at the base(?) of the *Halilucites costosus* horizon of the Reitzi Zone s.l.; 3. Appearence of *Gondolella transita and G.? praehungarica* near to the base of the Curionii Zone. The *Xenoprotrachyceras reitzi* horizon is not favourable for conodont investigations in the studied sections. The first occurrence of *"Metapolygnathus" hungaricus* lies much higher above the base of the Curionii Zone; accordingly, in the Epidauros section of Greece, where they coincide (Krystyn 1983), there should be a considerable gap in the lower part of the Curionii Zone.

At the end the conodont events are correlated with the ammonoid chronology (based on the study by Vörös, in the present volume) and the three (plus one) alternatives of defining the Anisian/Ladinian boundary are discussed. The first alternative at the base of the Reitzi Zone s.l. (though the change in conodont evolution at this horizon is the least sharp, with a fairly gradual transition) is supported not only by the priority, but also by the appearence of main Ladinian elements in radiolarians (Dosztály, in the present volume) and in palynomorphs and foraminifers (Góczán and Oravecz-Scheffer, in the present volume).

Key words: Triassic, biostratigraphy, conodonts, Balaton Highland (Hungary)

Introduction

Modern biostratigraphic re-evaluation of the Anisian/Ladinian boundary problem in the Balaton Highland (including conodont-biostratigraphy) began in 1978 by a team of Hungarian stratigraphers, for the meeting of the Triassic Subcommission of IUGS and of the IGCP Project No. 4 ("Triassic of Tethys Realm"), held in Hungary in that year. For this occasion the classical Felsőörs section (first excavated by Telegdi Róth more than a century before, who published his results in 1872) had been re-excavated and was shown during the meeting. These first results are contained in the paper by Szabó et al. 1980.

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During the eighties, within the scope of the National Type Section Programme, several sections have been excavated to expose the Anisian/ Ladinian boundary and investigated by a team of biostratigraphers for different fossil groups (the results of which are contained in the present volume). The conodont studies within this project have been performed by the present author. Sections of three localities to be shown during the 1993 meeting have been investigated bed-by-bed: the Öreg-hegy between Vászoly and Pécsely (the locality is usually called simply as "Vászoly" in the literature), the Cser-tető at Mencshely and the Forrás-hegy at Felsőörs (Fig. 1). A short taxonomic discussion of the stratigraphically important conodonts is given in a separate contribution (Kovács, in press) A brief conodont- biostratigraphic evaluation of the Vászoly and Felsőörs sections were given by Kovács et al. 1990.

In the present work we correlate the conodont biostratigraphy with the ammonoid zonation of Vörös (in the present volume) and use his stratigraphic columns. However, in a few cases there is a slight shift in the numbering of the beds as compared with the original one by Szabó et al. made in 1978; such cases are indicated in the text.

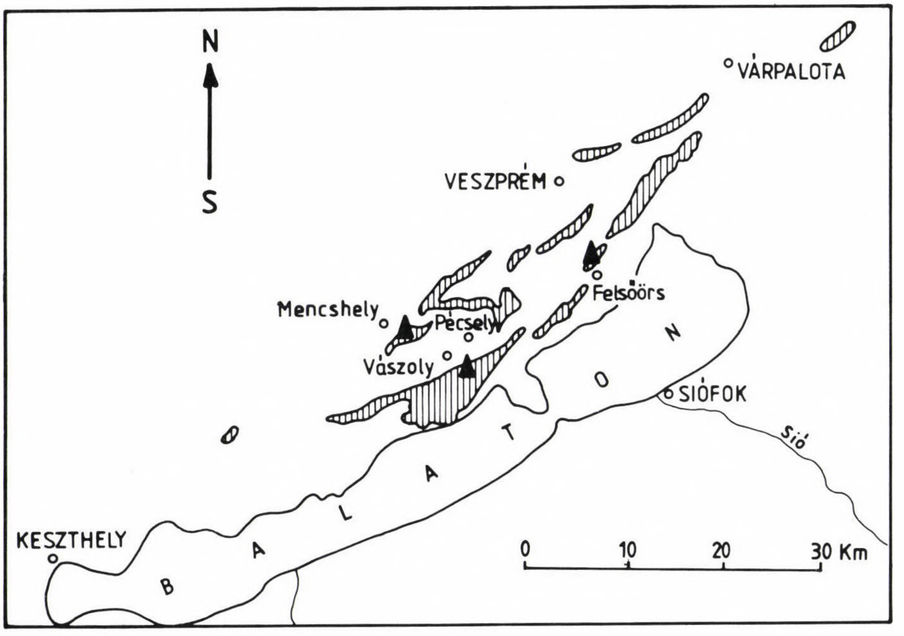

Fig. 1
Location of the investigated sections in the Balaton Highland. Vertically hatched areas indicate surface occurrences of Middle Triassic rocks.

Investigated sections

Felsőörs

The upper section on the western slope of Forrás-hegy at Felsőörs exposes the upper member of the Felsőörs Limestone ("Trinodosus Zone" auct.; Lóczy 1916) and the lower and middle members of the Buchenstein Formation ("Reitzi Zone" auct. and the lower part of "Tridentinus Zone" auct., in sense of Lóczy 1916). Bed-by-bed sampling of the Felsőörs Limestone and the lower member of the Buchenstein Formation (up to bed No. 111) was carried out in 1978, while that of the middle member in 1986 (samples from 0/86 on) (Figs 2, 3).

The upper member of the Felsőörs Limestone is built up by alternation

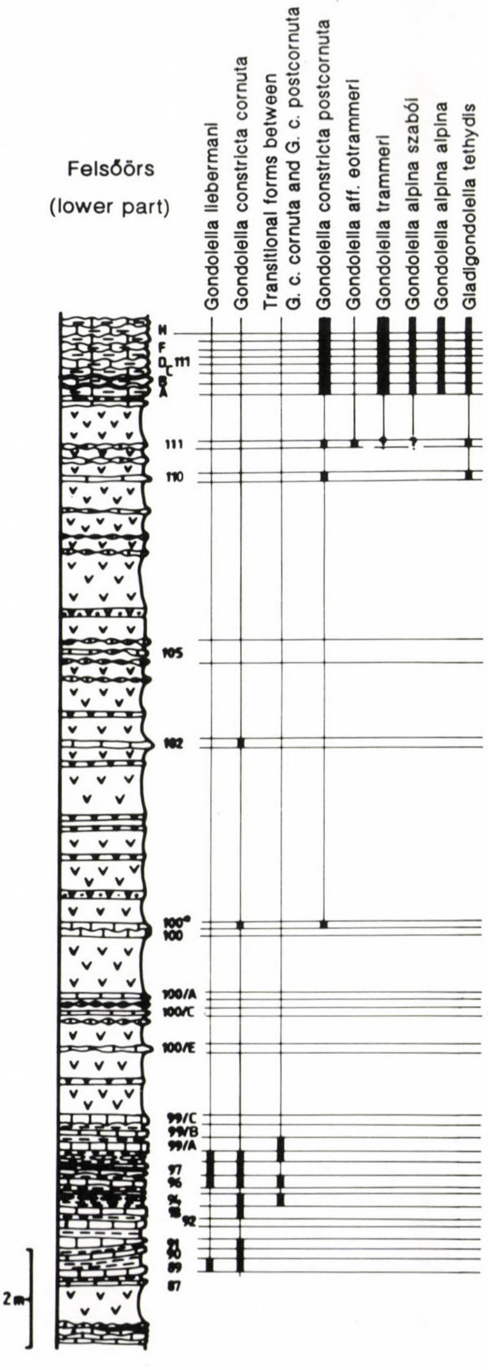

Fig. 2
Distribution of conodonts in the Trinodosus and Reitzi Zones of the Felsőörs section. (Lithostratigraphic column after Vörös, in the present volume.). Note: Due to the discontinuity of limestone nodule horizons in the lower tuffaceous member of the Buchenstein Formation, the number of the exposed limestone interbeddings have slightly been changed during the repeated excavations and sampling since the first excavation in 1978 (simplified lithostratigraphic column see in Szabó et al. 1980, Fig. 1). For this reason the Hungarian team agreed to use uniformly the lithostratigraphic columns reconstructed by A. Vörös during the latest ammonoid collecting. The double bed No. 100 was erraneously numbered by the present author in Kovács et al. 1990, Fig. 9b as "No. 101". From its upper part a specimen of *Gondolella constricta postcornuta* was figured in Szabó et al., 1980, pl. 59, Fig. 8 as "*Gondolella* cf. *longa*", as deriving from bed No. "100/A". Respecting Vörös's latest numbering, this bed is re-numbered in the present figure as bed No. "100^{x}".

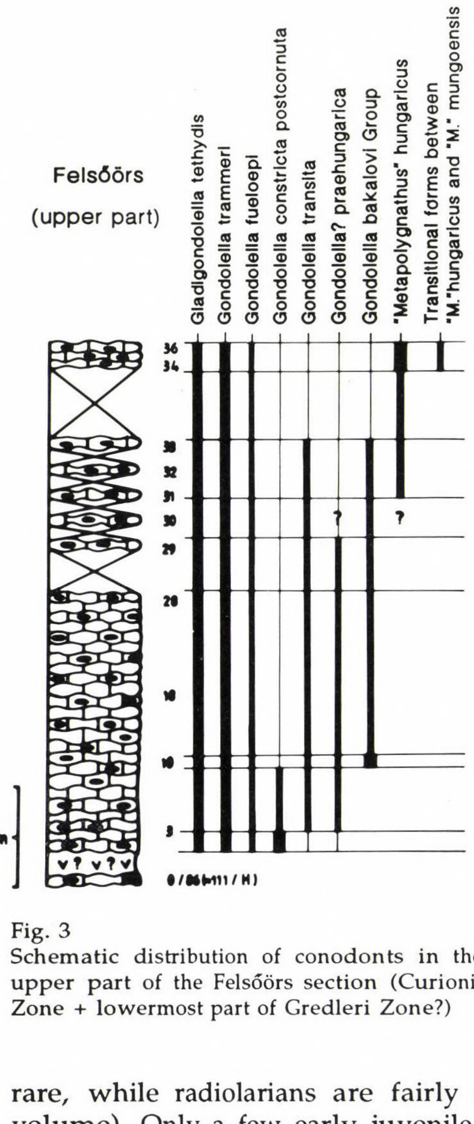

Fig. 3
Schematic distribution of conodonts in the upper part of the Felsőörs section (Curionii Zone + lowermost part of Gredleri Zone?)

of grey, in weathered stage brownish grey, tabular limestone beds and yellowish brown clay beds (beds Nos 85–99). Ostracods and especially juvenile pelecypod shells are very abundant in these limestone beds (see also Kozur 1970).

On the contrary, conodonts occur only in a rather low number. The bed No. 87 is very rich in radiolarians (see Dosztály, in the present volume). Conodonts are mainly represented by juvenile (= *"G. constricta"* auct.), medium and adult (= *"G. cornuta"* auct.), sometimes hyperadult forms of *Gondolella constricta cornuta*. Besides these, representatives of *G. liebermani* were also found sometimes. From bed No. 94 to bed No. 99 transitional forms between *G. constricta cornuta* and *G. constricta postcornuta* occur, which are becoming more slender and elongated. However, it should be noted, that such shaped form of *G. constricta cornuta* (even with an anteriorward shifted pit) was found already in bed No. 89.

In the yellowish brown, sometimes greenish, siliceous limestone intercalations of the here anomalously thick lower, tuffaceous member (called by a few authors as "reitzi tuff"; e.g. Krystyn 1983) of the Buchenstein Formation conodonts are extremely rare, while radiolarians are fairly abundant (see Dosztály, in the present volume). Only a few early juvenile forms of *G. constricta* and three typical specimens of *G. constricta postcornuta* (in beds No. 100* and 110), representing more advanced ontogenetic stages, have been found. In terms of ammonoids, this member includes the *Kellnerites felsoeoersensis, Hyparpadites liepoldti* and *Xenoprotrachyceras reitzi* faunal horizons (see Vörös, in the present volume).

The tuffaceous member s.s. (exposed in the main, deeper trench) is followed by a transitional "tuff with limestone nodules" (= pyroclastic debris flow with limestone clasts) horizon, exposed in the 3 m long small trench. Its lower boundary at the upper end of the main trench is formed by the bed No. 111

(the last coherent limestone bed upward in the tuffaceous member) and samples from its nodule horizons are numbered as No. 112–116 (beds No. 111/A–H according to Vörös, in the present volume; according to him they contain ammonoids of the *Halilucites costosus* horizon). In the bed No. 111 single specimens of *G.* aff. *eotrammeri* and *G. trammeri* were found whereas in the nodule horizons representatives of the latter become quite frequent. This small trench is terminated with an about 25 cm thick light yellowish grey, hard, strongly cherty-nodular limestone bed (with brownish grey chert nodules), containing the same conodonts (numbered No. 0/86). *Gladigondolella tethydis* occur from bed No. 111, in a low number. Further forms are represented by *G. constricta postcornuta* and rarely by *G. alpina szabói* and *G. alpina alpina*.

After a 3.5 m long covered interval (probably with tuffs) thick (15–25 cm) beds of yellowish grey, nodular limestone follow, with red chert nodules (beds No. 1/86 to 18/86), then light red or red-spotted, nodular limestone beds, also with red chert nodules (beds No. 19/86 to 28/86). With this the continouos profile comes to an end. Above the end of the horizontally exposed section scattered outcrops of single beds of light red, nodular, cherty limestone can be seen (beds No. 29/86 to 33/86). Higher up, on the northern slope of the hill and out of the wood, a small cliff is exposed containing three beds of light red, nodular limestone beds with red chert nodules (beds No. 34/86 to 36/86). Though ammonoids have not been found here, in the light the Vászoly sections (see below and in Vörös, in the present volume) this part corresponds to the *Eoprotrachyceras curionii* zone, while the isolated uppermost outcrop (beds No. 34–36/86) might already correspond to the basal part of the *Protrachyceras gredleri* zone (in sense of Krystyn 1983).

This middle, calcareous member of the Buchenstein Formation was formerly assigned to the "Tridentinus Zone" auct. (Lóczy 1916 and even in Szabó et al. 1980) and distinguished as "Nemesvámos Limestone Formation" (Balogh 1981). However, the *tridentinus*–fauna was found much higher in the stratigraphic column, on the opposite slope of Forrás-hegy (Szabó pers. comm.) and only the light red limestone part of the sections (beginning with bed No. 19/86) can be assigned to the "Nemesvámos Limestone" s.s. Even, according to the results of the latest mapping, this unit should be considered only as a member of the Buchenstein Formation in the Balaton Highland (Budai and Dosztály 1990).

Gondolella trammeri is predominant and occurs in a great number throughout the section of this middle member, in all beds. Besides this, *G. fueloepi* (moderately frequent already from the bed No. 1/86 on), *G.? praehungarica** (subordinate, but present in many beds; mostly juvenile forms) and rare representatives of *G. transita* are characteristic for this interval, both from bed No. 3/86 on. *G. constricta postcornuta* and *G. constricta balkanica* (?) range up into this member, but from bed No 10/86 remarkably elongated gondolelloids

* Representing transitional evolutionary stage between genus Gondolella and "Metapolygnathus"

related to the *G. bakalovi*-group are common. *Gladigondolella tethydis* is also common throughout the section.

An important evolutionary event can be recognized from bed No. 31/86 on, by the appearence of true metapolygnathoids (*"M" hungaricus).* In the uppermost exposed beds (Nos 34–36/86) already transitional forms between *"M" hungaricus* and *"M" mungoensis* do occur, indicating the Fassanian/ Longobardian boundary interval or the basal part of the latter.

Mencshely

The section has been excavated on Cser-tető Hill, and exposes the uppermost bed of the Felsőörs Limestone and the lower tuffaceous member of the Buchenstein Formation, overlain by purplish red crinoidal limestone.

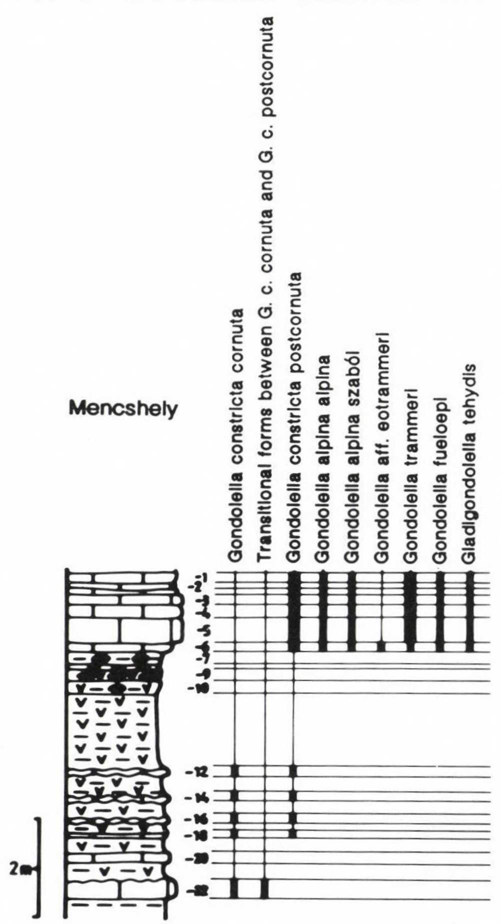

Fig. 4
Distribution of conodonts in the Mencshely section. (Lithostratigraphic column after Vörös, in the present volume.)

The topmost bed of Felsőörs Limestone (No. 22) contains rare specimens of the same conodont association as the topmost bed (No. 99) of the same formation in the Felsőörs section, with transitional forms between *Gondolella constricta cornuta* and *G. constricta postcornuta.*

The brownish yellow or grey coloured, siliceous limestone intercalations (beds No. –18 to –12) of the lower, tuffaceous member of the Buchenstein Formations (comprising the *Kellnerites felsoeoersensis* and *Hyparpadites liepoldti* ammonoid horizons; see Vörös, in the present volume) contain subspecies of *G. constricta* in a low number, among them *G. constricta postcornuta.*

The few ammonoid fragments dissolved from the tuff beds of the *Xenoprotrachyceras reitzi* horizon (see Vörös, in the present volume) yielded only a few Gondolella fragments and ramiform elements.

The overlying purplish red crinoidal limestone beds (No. -6 to -1) represent the *Halilucites costosus* ammonoid horizon (Vörös, in the present volume). In the first bed (No. -6) *G.* aff. *eotrammeri* and *G. trammeri* jointly occurs, but higher up the former is missing, whereas the latter is quite frequent. Typical forms of *G. constricta postcornuta* are also frequent, but representatives of *G. constricta cornuta* are already missing. This facies was favourable for the *szabói*-lineage: *G. alpina szabői* (morphologicallly closely resembling *G. bulgarica*, that became extinct at the end of the Pelsonian) and *G. alpina alpina* are common in it. On the *excelsa*-lineage *G. fueloepi* occurs already in this horizon. So do *Gladigondolella tethydis*, indicating that full pelagic conditions had been established by this time.

After a fault, the same lithology is repeated, with ammonoids of the same age (beds No +1 to +5). This part of the section has not been investigated for conodonts.

Vászoly, Öreg-hegy

Ditch P 11a

The limestone intercalations of the lower tuffaceous member of the Buchenstein Formation are fairly rich in conodonts (especially the beds No. 1, 5 and 6). This part of the section (up to bed No. 17) was sampled for conodonts by I. Szabó (the results see in Kovács et al. 1990, Fig. 5). Additional sampling was made later by the present author in beds No. 1–6.

Gondolella constricta cornuta is predominating up to bed No. 6 (containing two internal beds, both sampled for conodonts: samples No. 6 and No. 6*), and is present up to bed No. 12, but higher up it disappers.

Transitional forms between *G. constricta cornuta* and *G. constricta postcornuta* occur rarely already in bed No. 1 (yellowish, phosphatic limestone layer lying immediately on top the of the Megyehegy Dolomite of platform facies) and more frequently in beds No. 3 and 4.

Primitive representatives of *G. constricta postcornuta* characterize the beds No 5, 6 and 6; however, they are still subordinate in number as compared with *G. constricta cornuta*. There is a slight shift here in the faunal change concerning conodonts and ammonoids. The latters change with one bed higher only: the bed No. 5 contains still *Parakellnerites* aff. *meriani B,* whereas bed No. 6 already *Kellnerites* sp., indicating the *Kellnerites felsoeoersensis* horizon (see Vörös, in the present volume).

Typical forms of *G. constricta postcornuta* occur from bed No. 8 on. In the limestone nodule horizons of the tuffaceous beds (No. 8, 9 and 12) they predominate in the conodont association and show already undoubtedly the same morphology (the cusp is fused with the platform end even in juvenile forms without any stronger denticle before it, and the pit is remarkably anteriorward shifted), which characterize the representatives in the *Eoprotrachyceras curionii* ammonoid zone. (That is, the *constricta*-lineage shows already a "Ladinian" evolutionary stage from bed. No. 8. on.) Also, in these beds *G. constricta cornuta* is already subordinate in number against *G. constricta*

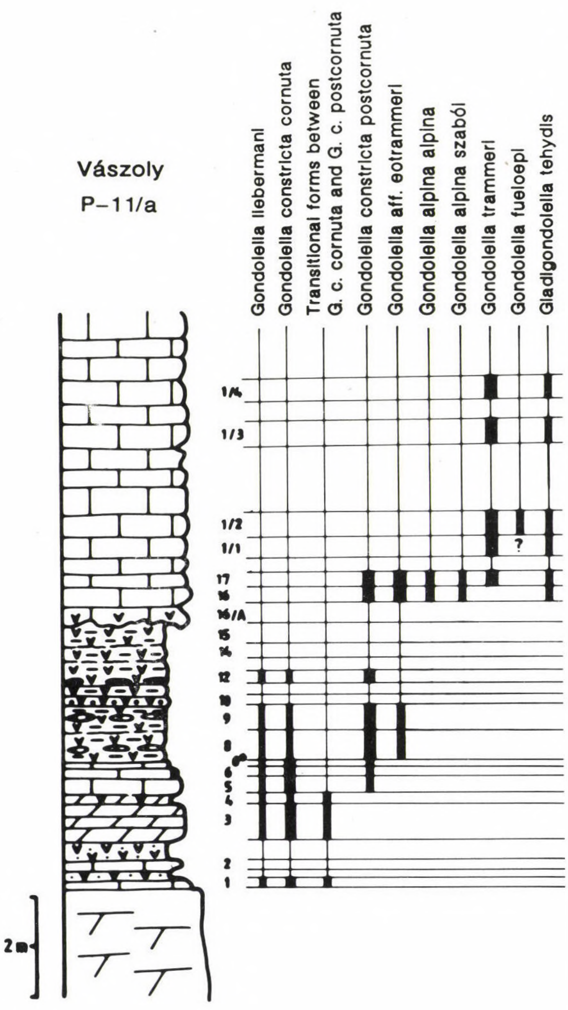

Fig. 5
Distribution of conodonts in the Vászoly P 11a section. (Lithostratigraphic column after Vörös, in the present volume.)

postcornuta. Ammonoids from bed No. 9 indicate the *Hyparpadites liepoldti* horizon (Vörös, in the present volume).

G. liebermani is common in all these beds (from No. 1 to 12), whereas *G. excelsa* is represented only by a few specimens.

The tuff beds (No 13 to 15) have not been sampled for conodonts (see Kovács et al. 1990). Subsequently Vörös, A. (see in the present volume, and in Vörös and Pálfy 1989) has found specimens of *Xenoprothrachyceras reitzi* in bed No. 14.

Also, at the time of the sampling for conodonts, the redish, ammonoid-bearing limestone beds No. 16a (reconstructed from isolated blocks in the wall of the trench; see Vörös and Pálfy 1989, Fig. 2) and No. 16 were not yet distinguished (see Kovács et al. 1990, Fig. 5). These yielded a rich ammonoid fauna belonging to the *Halilucites costosus* horizon (Vörös, in the present volume) and the same conodonts, as the purplish red crinoidal limestone beds of the Mencshely section (beds No. -6 to -1; see above), with the exception of *G. trammeri*. For this reason, the Anisian/Ladinian boundary was proposed here between beds. No. 16 and 17 by Kovács et al. 1990, at the base of its first appearence in bed No. 17.

Higher up in the section, in the white, thick bedded micritic limestone (Vászoly Limestone Member) *G. trammeri* predominates, in associaton with *G. fueloepi*, *G. constricta balkanica* and *Gladigondolella tethydis*. This part of the section, according to a nearby found specimen of *Eoprotrachyceras curionii* (Vörös, in the present volume), should already belong to the Curionii Zone.

Ditch P2

Two samples from the lower tuffaceous member of the Buchenstein Formation (No. 1/78 and 2/78 in Kovács et al. 1990, Fig. 6) yielded a poor conodont fauna, including *Gondolella constricta cornuta, G. constricta postcornuta, G. excelsa* and *G. mesotriassica* (s.s.). Interestingly, *Gladigondolella tethydis* already appears in the purplish limestone bed of sample No. 1/78.

This is the only section, in which a few determinable conodonts have been found in the ammonoid nodules of the *Xenoprotrachyceras reitzi* horizon (the tuff bed No. 4 according to the numbering of Vörös, in the present volume). These include *G. constricta postcornuta, G. constricta cornuta* juv., *G. alpina alpina, G. excelsa* and *G.* aff. *eotrammeri* (1–2 specimens from each).

The *Halilucites costosus* ammonoid horizon is missing in this section (a hyatus?).

On the other hand, the first bed (No. II/3 in Kovács et al. 1990, Fig. 6. and No. 1 in Vörös, in the present volume) of the overlying white, thick-bedded micritic limestone (Vászoly Limestone Member) yielded a specimen of *Eoprotrachyceras curionii* (Vörös, in the present volume). Unfortunately, the sample dissolved from this bed was very poor in conodonts.

Higher up in the section *G. trammeri* predominates in the conodont association, but Gladigondolella is also common, indicating a full pelagic connection. *G.? praehungarica*, characteristic of the Felsőrs section, is not present in the Vászoly Limestone. On the other hand, rare specimens of *G. transita* have been found and in bed II/6 forms belonging to the *G. bakalovi*-group are frequent. (For distribution of conodonts see Kovács et al. 1990, Fig. 6.)

3. *Conclusions and open problems*

Joint evaluation of the sections investigated bed-by-bed for conodonts allows to recognize three main evolutionary events from the upper part of the Trinodosus Zone to the lower part of the Curionii Zone.

The results of the similarly bed-by-bed carried out ammonoid investigations (Vörös, in the present volume) permit an exact correlation with the ammonoid horizons. We prefer to discuss the events of conodont evolution in this orthobiostratigraphic scheme, rather than in terms of any of the previously proposed conodont zonations. Several "standard" zonations have already been proposed by different authors, but almost none of the zones proposed is based on sections investigated in a "standard" manner. (The only exception is the Epidauros section, which is, unfortunately, rather condensed at this chronostratigraphic interval; see Krystyn 1983.)

At the end we attempt to correlate the conodont events with the radiolarian, palynomorph and foraminifer ones Fig. 6, the latters after (Dosztály, resp. Góczán and Oravecz- Scheffer, in the present volume), too.

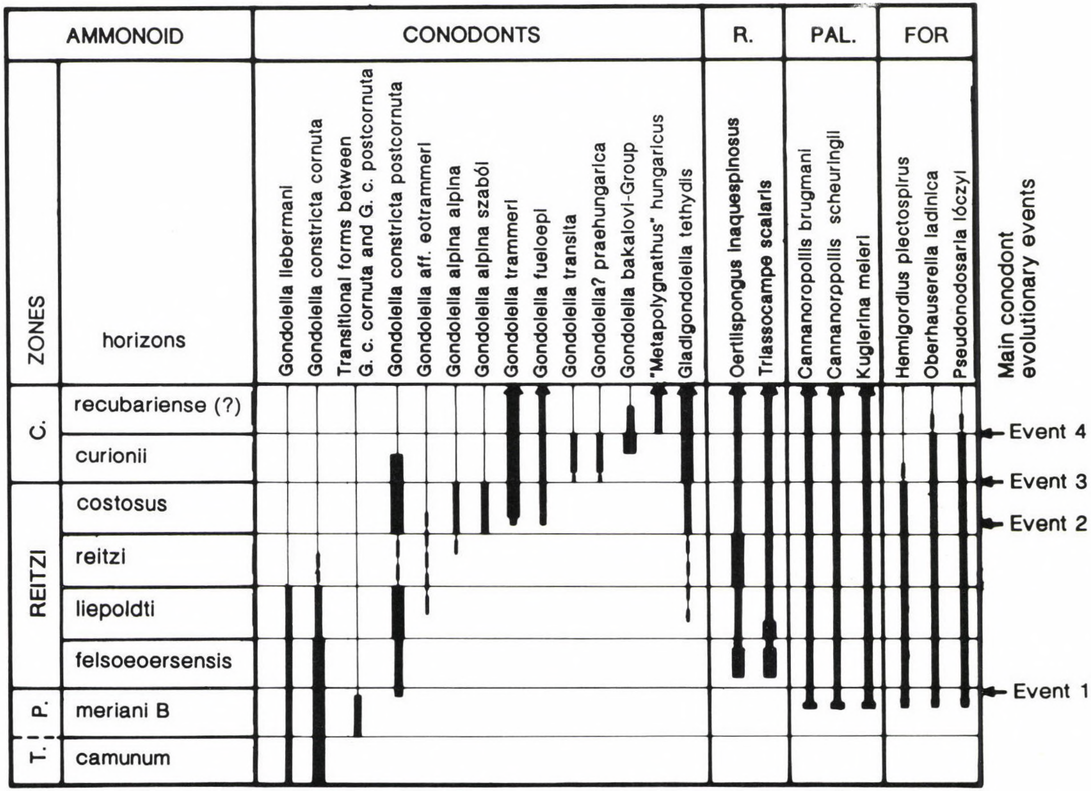

Fig. 6
Ranges of the stratigraphically most important conodonts in the Anisian/Ladinian boundary interval of the Balaton Highland, according to the ammonoid chronology by Vörös (in the present volume). Position of the main conodont evolutionary events and their relation to the main radiolarian and palynomorph evolutionary events (according to Dosztály, resp. Góczán, in the present volume; only concerning Event 1) are also shown. T –trinodosus; P –polymorphus; C –curionii; R – radiolarians; PAL – palynomorps; FOR – foraminifers

1st event: Appearence of *Gondolella constricta postcornuta*, evolved from *G. constricta cornuta*. The event is preceded by the appearence of transitional forms between the two subspecies, in beds No. 94–99 in the Felsőörs section and in beds No. 1–4 in the Vászoly P 11a section. This transitional interval (with the exception of bed No. 5 in the Vászoly P 11a section) coincides with the *Parakellnerites* aff. *meriani* B ammonoid horizon.

Primitive representatives of *G. constricta postcornuta* occur first in the uppermost part of the *P.* aff. *meriani* B horizon (in bed No. 5 in the Vászoly P 11a section) and characterize the *Kellnerites felsoeoersensis* horizon. In this interval representatives of *G. constricta cornuta* still predominate.

In the *Hyparpadites liepoldti* horizon (beds No. 9 and 12 in the Vászoly P 11a section, as well as the ammonoid-free bed No. 8) typical, highly evolved forms represent *G. constricta postcornuta*, which show the same morphological characteristics, as those in the *Halilucites costosus* horizon and in the *Eoprotrachyceras curionii* zone. Thus, the *constricta*-lineage from here on shows already an undoubtedly "Ladinian" character, with rather uniformly long, slender forms in pre-adult ontogenetic stages and with considerably anteriorward shifted pit* having prominently protruding magins. This morphological trend lead to the appearence of *G. transita* and later the *G. bakalovi*-group in the Curionii Zone. In this horizon representatives of *G. constricta cornuta* are already subordinate in number against those of *G. constricta postcornuta*.

The *Xenoprotrachyceras reitzi* horizon cannot be satisfactorily evaluated in the Balaton Highland from the point of view of conodont biostratigraphy, because of its poorness in conodonts (being mostly represented only by ammonoid nodules in tuff beds). Hopefully, the study of Italian conodont workers on limestone sequences of the lower Buchenstein Formation in the Southern Alps will be more successful.

2nd event: Appearence of *Gondolella trammeri*, developed from *G.* aff. *eotrammeri*. In our area, according to the Mencshely and Felsőörs sections, this event coincides with the base (?) of the *Halilucites costosus* horizon (but see the problem of the reitzi-horizon discussed just above and also of the Vászoly sections). In the lowermost bed the two species occur together, whereas higher up only *G. trammeri*. Characteristic forms of this horizon are *G. alpina alpina* and *G. alpina szabói*. The preceding evolutionary history of this lineage, leading to *G. trammeri*, is still poorly known, being bound mainly to slope facies (crinoidal limestones; Kovács and Papšová, in press).

On the *excelsa*-lineage *G. fueloepi* appears first also in this horizon (but only in the Mencshely section). Regular occurrence of Gladigondolella indicates the

* It should be noted, however, that these morphological features occur sometimes already in the Trinodosus Zone (see above in the description of the Felsőörs section).

establishment of full pelagic connection of the basin (though its representatives become more frequent from the basal Curionii Zone on).

3^{rd} *event*: Occurrence of *Gondolella? praehungarica* (developed from G. *trammeri*) and *G. transita* (developed from *G. constricta postcornuta)* from bed No. 3/86 in the Felsőörs section. This event has an important phylogenetic significance, being *G.? praehungarica* the evolutionary link between gondolelloids and the Middle Triassic metapolygnathoids. However, representatives of the two species occur only rarely in the Felsőörs section (*G. transita* is even rarer) and *G.? praehungarica* is missing in the Vászoly sections. This is probably because *G.? praehungarica* may have lived in deeper water environments: the cherty limestone of the Felsőörs section was deposited in a basin existing already during the Pelsonian–Illyrian, while the white, thick bedded micritic limestone of the Vászoly section (Vászoly Limestone) over the Megyehegy Dolomite platform existing until the late Illyrian (see Budai and Vörös, in the present volume).

This event, recognized at the base of the Nemesvámos (s.l.), resp. of the Vászoly Limestone, approximately coincides with the base of the *Eoprotrachyceras curionii* zone (see Vörös, in the present volume). *G. trammeri, G. fueloepi* and *Gladigondolella tethydis* are already common after this event (especially the first species). Higher up, within the range of the *G.? praehungarica,* a less remarkable event can be recognized by the mass occurrence of the *G. bakalovi*-group (bed No. 10/86 in the Felsőörs section and bed No 6 in the P2 section at Vászoly).

4th event: A further event can be recognized higher up by the occurrence of true metapolygnathoids (*"M" hungaricus,* from bed No. 31/86 on in the Felsőörs section). This event in the classical Epidauros section of Greece seems to coincide with the base of the Curionii Zone (Krystyn 1983). According to the conodont distribution in the Felsőörs section, however, in the condensed Epidauros section cut by numerous hardgrounds a considerable part of the Curionii Zone should be missing. This hyatus comprises the interval corresponding at least to the beds from No. 0/86 to 30/86 in the Felsőörs section.

The local conodont zonation, that can be recognized in the Balaton Highland based on these events and their correlation with ammonoids, palynomorphs and radiolarians (after Vörös, Góczán and Dosztály, all in the present volume) is shown on Fig. 7.

These zones partly deviate from the so far proposed "standard" zonations; however, these are based on bed-by-bed investigations. On the other hand, these have been recognized until now only in the Balaton Highland and for this reason the present author strictly considers them as local ones.

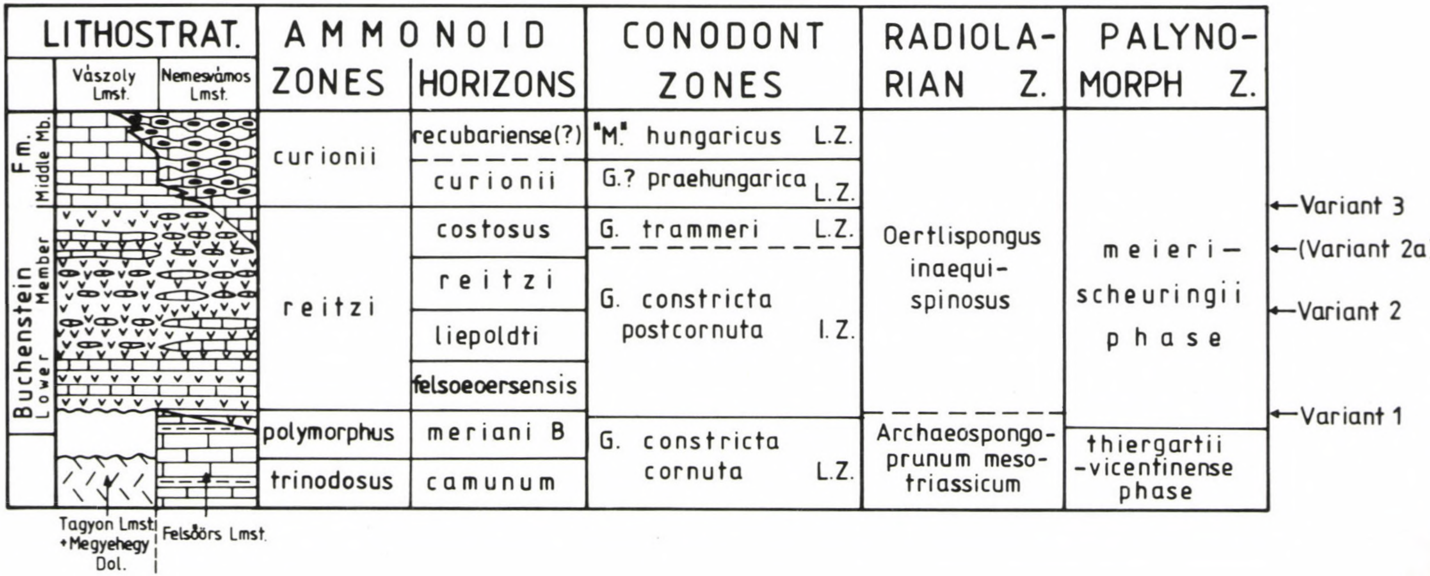

Fig. 7
Local conodont zones recognized in the Anisian/Ladinian boundary interval of the Balaton Highland and their relation to the ammonoid chronology, radiolarian and palynomorph zones (using the zonations proposed by Vörös, Dosztály and Góczán; all in the present). L.Z – lineage zone; I.Z – interval zone

Where to define the Anisian/Ladinian boundary based on conodonts?

A.) At Event 1., with the occurrence of *G. constricta postcornuta*. This is the least sharp boundary from among the three possibilities, with a fairly gradual morphological transition between *G. constricta cornuta* and *G. constricta postcornuta*. Even, in the eupelagic sequences of Northeastern Hungary this event cannot be recognized, being *G. constricta postcornuta* missing (see below). However, there is a fairly good coincidence with the ammonoid boundary (with Variant 1, at the base of the Reitzi Zone s.l.; see Vörös, in the present volume); one bed difference in the Vászoly P 11a section and no recognizable difference in the Felsőörs section. Also, this event is very near (see Fig. 6) to the first appearence of typical Ladinian radiolarians (in bed No. -16 in the Mencshely section; see Dosztály, in the present volume, and our Fig. 7) and of typical Ladinian palynomorphs (slightly below the top of the Felsőörs Limestone; see Góczán, in the present volume), as well as to the change of foraminifers (in bed No. 98 in the Felsőörs section; see Oravecz-Scheffer, in the present volume). Though the latters are usually facies-indicators and have less stratigraphic value, this change of foraminifer fauna may have a stratigraphic importance, because it happened within the same facies and same formation.

This nearly perfect coincidence of phylogenetic events in five fossil groups strongly supports the definition of the Anisian/Ladinian boundary at the base of the Reitzi Zone (used sensu lato as an Oppel Zone; see Vörös, in the present volume), which unambiguously has the priority among all ammonoid zonal names in the boundary interval, introduced already by Böckh 1873. This would be in accordance with the traditional view of Hungarian geologists about the position of the Anisian/Ladinian boundary, between the "Reifling Limestone" auct. (= Felsőörs Limestone) and the Buchenstein Formation of the Balaton Highland (see Lóczy 1916; Balogh 1981). Definition of the Anisian/Ladinian boundary in this way would also be in accordance with the original definition of the Ladinian Stage by Bittner 1892, who defined the lithological content of its lower part by the Buchenstein Formation, though in the Southern Alps*. That is, in this case the "golden spike" should be inserted at the lower boundary of the stratotype of the Reitzi Zone, in between the top of the Felsőörs Limestone and the base of the "reitzi-tuff" (namely between bed No. 99/C and the thick tuff horizon below bed No. 100/E). Essentially this view has been advocated also by Kozur since 1972 and 1973.

* It should be noted, however, that he lived and worked in a historical time, when the Balaton Highland and part of the Southern Alps belonged to the same country. The areal separation of the two regions was probably not very decisive in the thinking of the geologists of the Austro-Hungarian Monarchy, therefore they often applied previously introduced Alpine names for the present territory of Hungary, too; among others the "Recoaro Limestone", the "Buchenstein beds" and, incorrectly, the "Reifling Limestone" for the Balaton Highland (cf. Böckh 1873 and Lóczy 1916).

The disadvantage of defining the Anisian/Ladinian boundary at *Event 1* (from conodont biostratigraphic point of view) is, that *G. constricta postcornuta* seems to be absent in eupelagic facies (in which gladigondolelloids were present already in the Anisian or even in the Late Scythian, from the beginning of basinal sedimentation), thus making this boundary unrecognizable. At least it was not found in Northeastern Hungary (Kovács, unpubl.) and forms belonging to it were not reported from the "ammonitico rosso" type limestones of Epidauros in the Hellenides (Krystyn 1983) and of the area of Han Bulog in the Dinarides (Sudar 1982; Sudar and Budurov 1983). This means, that this form was probably restricted to intrashelf basins.

B.) At Event 2. As opposed to *G. constricta postcornuta* of the *constricta*-lineage, *G. trammeri* is a characteristic and common form of Tethyan eupelagic sequences. As ammonoid accumulations are usually rare, but conodonts are common in pelagic facies, the first occurrence of *G. trammeri* (evolved from *G.* aff. *eotrammeri)* offers a clearly and easily recognizable boundary in these facies. As *G. trammeri* is frequent throughout the Fassanian and part of the Longobardian, the definition of the Anisian/Ladinian boundary by the appearence of this characteristic Ladinian element (see Variant 2a on Fig 7) would be of great practical stratigraphic usefulness (similarly to the *G. polygnathiformis*-event for the beginning of the Carnian or the *G. navicula*-event for the beginning of the Norian).

Definition of the lower boundary of the Ladinian at this event was suggested first by Krystyn 1983 (who proposed also first the dividing of the Reitzi Zone into two zones*) and was subsequently advocated by Kovács et al. 1990, too.

However, within the Reitzi Zone (in sense of Mojsisovics 1882 and Vörös, in the present volume), the presence of *G. trammeri* can be proven only in the *Halilucites costosus* horizon, whereas the *Xenoprotrachyceras reitzi* horizon cannot be satisfactorily evaluated for conodonts (see above). In the knowledge of the new results of Vörös, but respecting also the previously published opinions (Krystyn 1983, Vörös and Pálfy 1989 and Kovács et al. 1990), the present author in Vörös et al. 1991, advocated to use the Reitzi Zone in a narrow sense and to draw the Anisian/Ladinian boundary at the base of the *X. reitzi* horizon (see Variant 2 on Fig. 7); in this case the known first occurrence of *G. trammeri* would have been only slightly above the base of the Ladinian (cf. the Vászoly P 11a and the Mencshely sections).

This gap in the conodont-biostratigraphy of the Reitzi Zone may probably be filled up in the more calcareous South Alpine sections by Italian colleagues.

* In 1979, during the meeting of the Triassic Conodont Working Group in Budapest he proposed dividing the Reitzi Zone into a lower Kellnerites Zone and a higher Nevadites Zone. His proposal was, however, used by Kozur (1980, p. 96, and in his revised zonal scheme; furthermore, in Kovács and Kozur 1980, p. 48–49) without correct reference, therefore in 1983 he suggested Parakellnerites for the lower zone. In the light of the new results by Vörös (in the present volume) Krystyn's original proposal can be verified, being Parakellnerites a rather long-ranging genus

C) At Event 3. The appearence of *G. transita* and especially of *G.? praehungarica* is a phylogenetically important event, but is not easy to recognize even within the Balaton Highland because of the rarity of both forms. In Northeastern Hungary they have even not been found. Much more significant is the contemporaneous predominance of *G. trammeri*, thus this event does seem to have a great usefulness in practice.

Definition of the Anisian/Ladinian boundary at this event would practically correspond to the base of the *Eoprotrachyceras curionii* zone (see Variant 3 on Fig. 7) and thus would meet the proposal of Tozer 1967; Dagys et al. 1979; and Brack and Rieber 1986.

D.) The first occurrence of *"Metapolygnathus" hungaricus* (Event 4), which coincides with the base of the Curionii Zone in the *Epidauros* section (Krystyn 1983) has no meaning on the definition of the stage boundary, because a considerable part of the Curionii Zone is missing there (see above).

Open problems

From among the three (or four) alternatives of the position of the Anisian/Ladinian boundary (see Fig. 7) the first one is the most supported at present by complex biostratigraphic studies. The priority is unambiguously for this alternative (see Böckh 1873; Mojsisovics 1879, 1882; Bittner 1892). Concerning the priority and the one-century long debate on the boundary of the two stages, however, two facts should be noted:

– Though Mojsisovics in 1895 confused the stratigraphic order of the Avisianum and Curionii Zones, but included both of them into his "Norische Stufe" and none of them into the Anisian. Thus, apart from his erraneous referring of the stage, what was named by Bittner 1892 as "Ladinian", to the Norian, he did not deviate from Bittner's original definition and included both ammonoid zones contained by the Buchenstein beds (see also Brack and Rieber 1986 and Vörös, in the present volume) into the same stage.

– This mistake by Mojsisovics did not influence the mapping practice and the Buchenstein Formation, as a whole, in both areas was usually mapped as Ladinian.

Before making a decision about the boundary and to choose any of the three (or four) alternatives (that is, before the voting by the Triassic Subcommission), it is unavoidably necessary to study the most important South Alpine sections bed-by-bed with the same methods (radiolarians, palynomorphs and more detailed conodont studies) and to correlate the events recognized. Probably the similarly ammonoid-rich, but more calcareous South Alpine sections offer a better possibility to answer those problems, which could not be investigated in the Balaton Highland (e.g. what kind of events took place in the evolution of the radiolarians and palynomorphs at the possible second and third alternatives, and what is the conodont content of the reitzi horizon).

In the author's opinion, in a full agreement with Visscher 1991, when defining chronostratigraphic boundaries, besides the principle of priority the applicability for long-distance (global) correlation should also be equally considered. These boundaries (because stratigraphers and mapping geologists are in the need to use them) should be recognizable worldwide: in continental deposits (by palynomorphs), in epicontinental marine deposits, in shelf carbonates (by dasycladaceans in the lagoonal facies of carbonate platforms) and in pelagic deposits (by ammonoids, conodonts, radiolarians and pelagic bivalves). (The mentioned fossil groups are understood for the Triassic.) This is only possible, if the stage boundaries are defined by *complex biostratigraphic methods*. In our case the Balaton Highland sections offered a good possibility to study the first alternative (which has also the priority) by ammonoids, conodonts, radiolarians, palynomorphs and foraminifers (see the contributions by Vörös, Dosztály, Góczán and Oravecz-Scheffer, in the present volume). There was no opportunity to study the dasycladacean boundary (due to lack of Ladinian platform carbonates in the studied area), but the position of the *Diplopora annulatissima* zone and the lower boundary of the *Diplopora annulata* zone (e.g. when it appears already without *D. annulatissima*; for the zonation see Pia 1930, p. 97; Ott 1972; Bystricky 1986; and Piros, in Kovács et al. 1989) should also be considered at the definition of the Anisian/Ladinian boundary, because the amount of platform carbonates is at least with a magnitude of order higher in the Alpine Triassic, than that of the basinal ones.

Whatever of the three (or four) alternatives (e.g. at the base of the Reitzi Zone s.l. or at base of the reitzi-horizon or at the base of the Curionii Zone; or at the first occurrence of *G. trammeri*) will be finally accepted for the Anisian/Ladinian boundary and wherever will it be defined (either in the Southern Alps or in the Balaton Highland; it is equal, because originally they belonged to the same paleogeographic domain on the southern shelf of the Vardar–Meliata–Euhallstatt ocean; see Kázmér and Kovács 1985), it should meet the above-mentioned requirements.

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The Anisian/Ladinian and Ladinian/Carnian boundaries in the Balaton Highland based on Radiolarians

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In the Balaton Highland Radiolarians were found from the Illyrian to the Julian substages. Volcanism during Ladinian provided favourable conditions to Radiolarians, and as a result these fossils are often found in great amounts in these sequences. The rich fauna allows the minute biostratigraphic classification (in some cases this is possible only by Radiolarians). Complex biostratigraphic investigations (on Ammonites, Conodonta, Radiolaria) carried out in the Balaton Highland allowed to elaborate a Radiolaria-zonation. The Radiolaria-zonation could be successfully correlated with the Ammonites and Conodonta zonations. In case of this zonation the stage boundaries represent zone boundaries, as well.

Key words: Radiolaria, Triassic, biostratigraphic zonation, Balaton Highland

Introduction

Investigation of radiolarians have had remarkable traditions in the Balaton Highland. It is more than hundred years that M. Hantken called the attention to the radiolarian content of some Triassic rocks (1884). Hantken sent the thin sections with Radiolarians to D. Rüst, the most famous specialist of that time. Rüst (1891) published ten new species from the limestones of Malomvölgy at Felsőörs. In the region R. Hojnos carried out studies on Radiolarians at the beginning of the century (1916, 1921). In addition to the classical Felsőörs section he described Radiolarians from the environs of Balatonalmádi and Vörösberény.

The investigations mentioned were made in thin sections; the first up-to-date analyses were carried out by H. Kozur. Together with H. Mostler he published several new taxa from Felsőörs and Köveskál (Kozur et Mostler 1981, 1983; Kozur 1984, 1988). The material studied in the course of mapping of the Balaton Highland in a scale of 1:10000 served as a basis to the recent investigations. By the study of Radiolarians stratigraphic problems could be solved and new taxa were described (Budai and Dosztály 1990; Dosztály 1991).

In the Balaton Highland Radiolarians are known from the Illyrian to the Julian substages. The spread of Radiolarians was favoured by the volcanic activity started in the Uppermost Anisian. Due to tuff dusting up to the Late Ladinian the dissolved silica content of seawater increased and this proved to

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be favourable to the population of Radiolarians. It is characteristic of Radiolarians that in a short interval they occur in great number while in other periods they may be completely absent. Tuffaceous formations show often subsequent resilicification. These formations are unsuitable to recover Radiolarians from them. The geographic position of outcrops discussed in the next chapter is shown in Fig. 1.

Radiolaria zonation

The investigation of the rich radiolarian fauna of the region completed with results from other areas and the literature data allowed to elaborate a Radiolaria-zonation. Zonation starts from the Upper Anisian and covers the interval to the Cordevolian substage (Fig. 2).

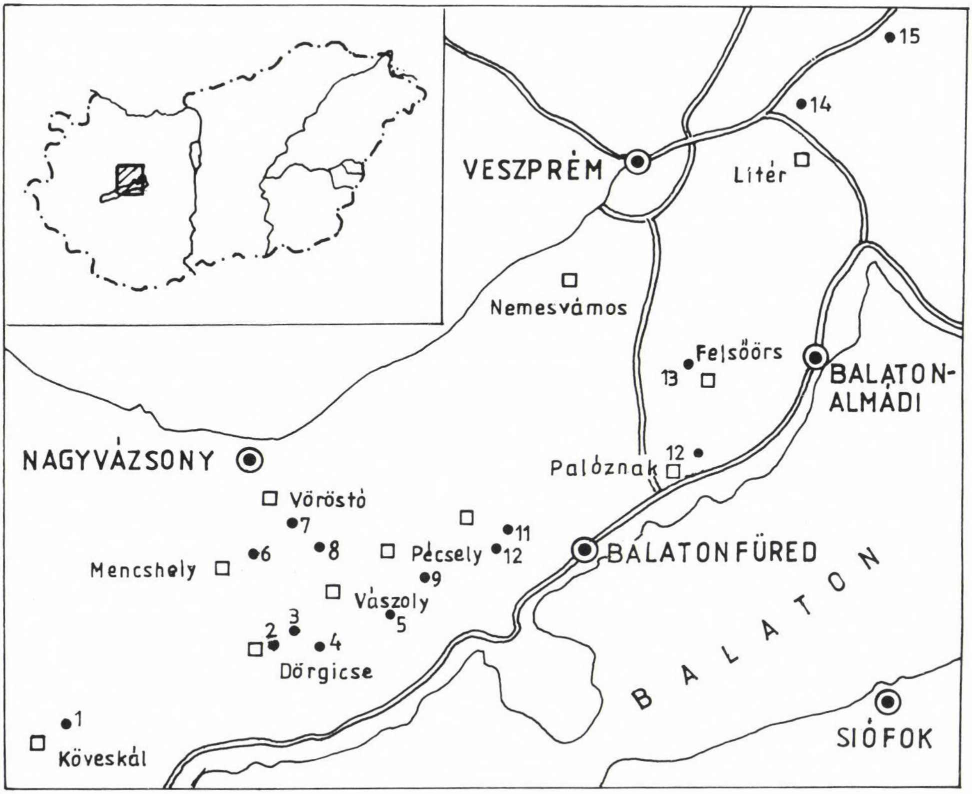


Fig. 1

Localities with radiolarians in the Balaton Highland. 1. Köveskál; 2. Alsódörgicse; 3. Borehole Dörgicse-1; 4. Dörgicse, Öregerdő; 5. Pécsely, Öreg Hill; 6. Mencshely, Csertető; 7. Borehole Vöröstó-7; 8. Borehole Mencshely-1; 9. Pécsely, Meggy Hill; 10. Balatonfüred Száka Hill; 11. Borehole Balatonfüred-; 12. Borehole Palóznak-1; 13. Felsőörs, Malomvölgy; 14. Litér, grit-quarry; 15. Sóly, road cut on the main road No. 8 (33 km)

CARNIAN	Paleosaturnalis triassicus
LADINIAN	Muelleritortis cochleata
	Oertlispongus inaequispinosus
ANISIAN	Archaeospongoprunum mesotriassicum

Fig. 2
Triassic Radiolarian zonation in Balaton Highland

Former authors who prepared Triassic Radiolaria-zonation did not define the Anisian/Ladinian boundary by radiolarians. The oldest Radiolaria-zones suggested so far did not reach downward the Lowermost Ladinian (Yao et al. 1980; Yao 1982; Tikhomirova 1986; Wakita 1988; Sano et al. 1992). The only exception is the zonation of Kozur (in Haas 1984), who divided the Upper Anisian and Lower Ladinian period into three Radiolaria-zones. Nevertheless, the zones have not been defined so far, moreover some of the index taxa are still unpublished. Thus, this zonation could not be used during our studies. In case of the Ladinian/ Carnian boundary the Radiolaria-zones proposed previously did not coincide with the stage boundaries (Yao et al. 1980; Tikhomirova 1986; Wakita 1988).

In the works published so far zonations were usually based on the genus Triassocampe. As to the recent studies, however, it is more expedient to use the representatives of the subfamily Oertlisponginae and that of the family Parasaturnalidae when establishing the zonations. Taxa belonging to these groups show faster evolution rate and more widespread, respectively.

In case of the boundaries correlation was made with ammonites and conodonts: in case of the Anisian/ Ladinian boundary ammonites, in that of the Ladinian/Carnian boundary conodonts served as basis of correlation. In the following each Radiolaria-zone will be shortly discussed.

Archaeospongoprunum mesotriassicum zone

The lower boundary of the zone is unknown, according to recent knowledge it cannot be defined, it covers the upper part of the Anisian, the interval of the *Paraceratites trinodosus* zone. The upper boundary of the zone should be drawn at the appearance of the first *Oertlispongus inaequispinosus*.

Most characteristic taxa of the zone are: *Archaeospongoprunum mesotriassicum* (Plate I: 3, 4)

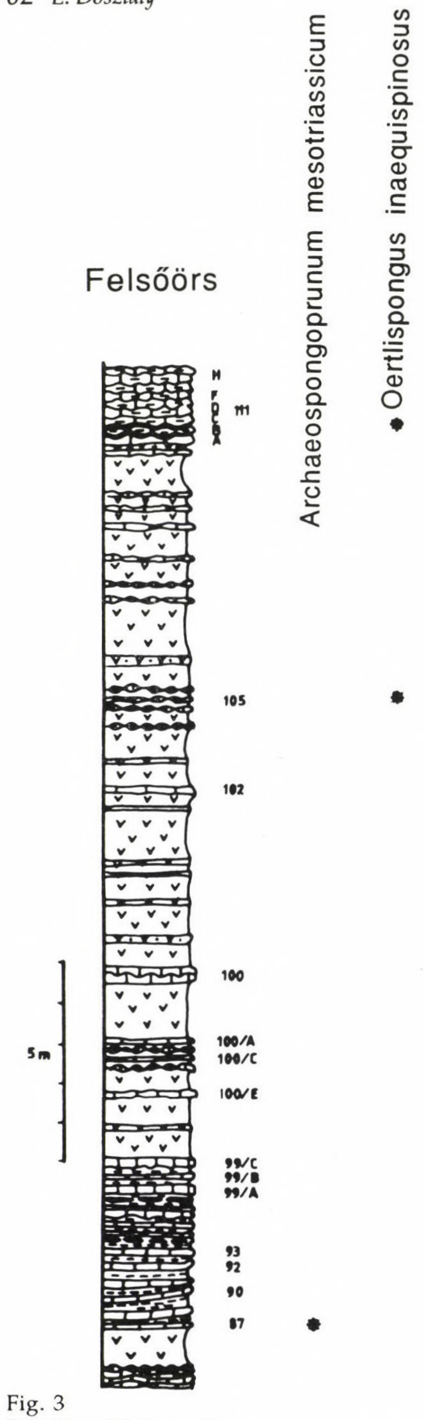


Fig. 3
Felsőörs, Malom-völgy

Foremanella macrocephala (Plate I: 2), *Planispinocyrtis baloghi* (Plate I: 1). The first representatives of the Oertlisponginae (*Paroertlispongus multispinosus*) can already be found in this zone.

Reference section of the zone is found in the Malomvölgy at Felsőörs (Fig. 3).

Occurrence: Felsőörs, Malomvölgy, bed No. 87, Borehole Balatonfüred-1 215.6 m.

Oertlispongus inaequispinosus zone

The lower zone boundary can be defined by the appearence of the first *Oertlispongus inaequispinosus*. In case of the Csertető section (Mencshely) this coincides with the lower boundary of the *Xenoprotrachyceras reitzi* ammonite zone (Vörös et al. 1991). The upper zone boundary is marked by the appearance of the species *Muelleritortis cochleata*. Due to the lack of ammonites this boundary cannot be defined at present. According to our recent knowledge it lies in the lower part of the Longobardian substage.

Most characteristic taxa are: *Oertlispongus inaequispinosus* (Plate II: 3–8), *Falcispongus calcaneum*, *Eptingium manfredi* (Plate I: 7) *Yeharia annulata*.

Reference section: Mencshely, Csertető (Fig. 4).

Occurrence: Mencshely, Csertető, bed No. 16, Alsódörgicse, Borehole Paloz- nak-1, between 55 and 63 m, Pécsely, Öregerdő, Felsőörs, Malomvölgy.

Muelleritortis cochleata zone

The lower boundary of the zone is marked by the appearance of the first *Muelleritortis cochleata*. The upper zone boundary can be defined by the appearance of the species *Paleosaturnalis triassicus*. In the Balaton Highland this boundary coincides with the appearance of the conodont species *Gondolella polygnatiformis* and *"Metapolygnathus" diebeli*.

Most characteristic taxa are: *Muelleritortis cochleata* (Plate III: 4, 5), *Hungarosaturnalis longobardicus* (Plate III: 3), *H. multispinosus* (Plate III: 1, 2), *Spongoserrula rarauna* (Plate I: 6).

Within the zone, the genus Ruesticyrtium also appears in addition to the genus Triassocampe. The earliest appearance of the representatives of the family Veghicyclidae can be fitted to this zone (Dosztály 1991).

Reference section: Borehole Palóznak-1

Occurrence: Borehole Dörgicse-1, between 33 and 34 m, Dörgicse, Öregerdő, Litér, grit-quarry, Borehole Mencshely-1, between 450 and 456 m, Borehole Palóznak-1 between 26 and 34 m, Pécsely, Öreg Hill, Sóly, road cut on the main road No. 8 (33 km), Borehole Vöröstó-7 between 40 and 42 m.

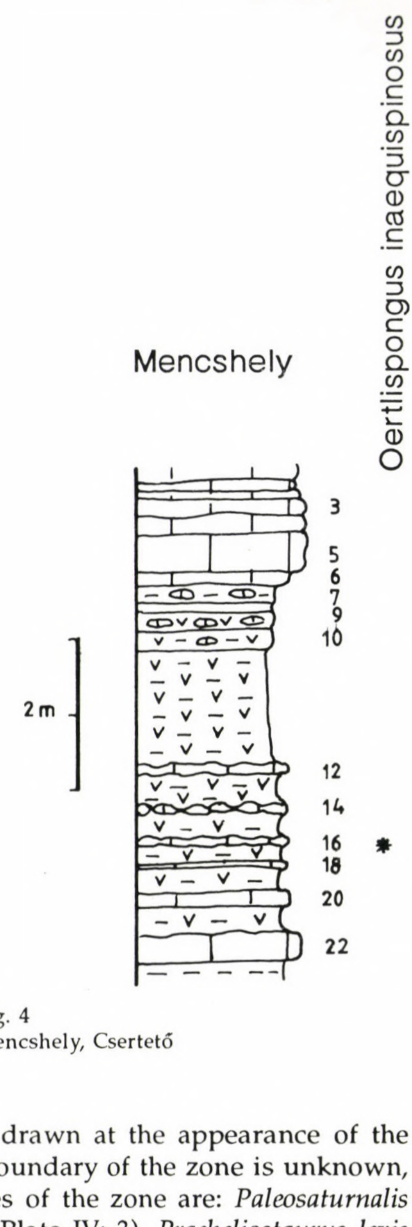


Fig. 4
Mencshely, Csertető

Paleosaturnalis triassicus zone

The lower boundary of the zone can be drawn at the appearance of the species *Paleosaturnalis triassicus*. The upper boundary of the zone is unknown, it cannot defined. Most characteristic species of the zone are: *Paleosaturnalis triassicus* (Plate IV: 1), *Paleosaturnalis zapfei* (Plate IV: 3), *Praeheliostaurus levis* (Plate IV: 2), *Spongosaturnaloides multidentatus* (Plate IV: 4), *Capnuchosphaera triassica*. At the Ladinian/Carnian boundary remarkable changes can be recorded within the radiolarian fauna. Significant groups such as the subfamily Oertlisponginae, the genera Triassocampe, Hungarosaturnalis and Muelleritortis become extinct or exist only in a very reduced number of individuals.

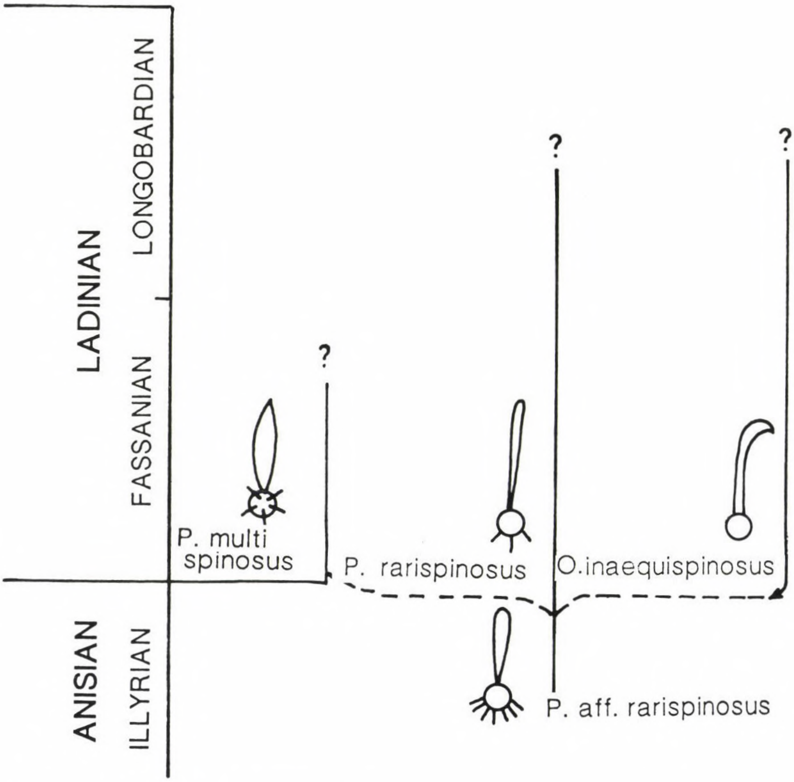


Fig. 5
The evolution of the Paroertlispongus genus

At same time, characteristic genera such as Paleosaturnalis, Praeheliostaurus, Praeorbiculiformella, Rhopalodyctium and Spongosaturnaloides appear. The species belonging to the familiy Veghicyclidae are also frequent. In the Száka Hill of Balatonfüred from this zone a *Trachyceras* cf. *aon* (Münster) was found, this verifies the Cordevolian age.

Reference section: Balatonfüred, Száka Hill.

Occurrence: Balatonfüred, Száka Hill, beds No 1–43, Borehole Mencshely-1 between 412 and 435 m, Pécsely, Meggy Hill, beds No. 1/3-1/21, II/4-II/36, III/1-III/7.

In the following some radiolarian taxa playing important role in the elaboration of the Radiolaria-zonation will be dealt with in detailes.

Subfamily Oertlisponginae

Genus Paroertlispongus

From the representatives of the subfamily only this genus appear in the Anisian (Felsőörs). Specimens found at Felsőörs cannot be identified as known species. These are most similar to the species *Paroertlispongus rarispinosus* (Plate

I: 5). As far as I know these specimens of Felsőörs are the oldest representatives of the subfamily and are probably the forerunners of the Oertlisponginae taxa. The specimens *Paroertlispongus rarispinosus* found in Csertető, Mencshely already agree with the holotype (these specimens were found in the lowermost strata of the Reitzi Zone; bed No. 16, see Vörös, in the present volume) (Plate II: 1). The evolution of the genus is seen in Fig. 5.

Genus Oertlispongus

The main diagnostic character of the genus, on the basis of which it can be distinguished from the Paroertlispongus genus is the curved main spine. This genus developed from the genus Paroertlispongus. In the fauna deriving from Csertető, Mencshely the evolution can be followed. The fauna derives from the lowermost layer of the Reitzi Zone (bed No. 16). In the sample the typical *Oertlispongus inaequispinosus* and the transitional forms, respectively, can be found (Plate II: 3–6). The initial stage is marked by the form that is similar to the *Paroertlispongus rarispinosus* (Plate II: 1) but differs from it by the lack of by-spines. The main spine of this is straight. Between the forms of straight spine and the *O. inaequispinusus* of square-curved spine all transitional stages can be found. The boundary of the curvature of the main spine can be detected. The extreme high rate of evolution is marked by the fact that the initial stage, the new species and the transitional forms can be found within one layer. In the Felsőörs section, in the higher horizons of the Reitzi Zone (bed No. 105) specimens similar to the holotype are found. The evolution of the genus is seen in Fig. 6.

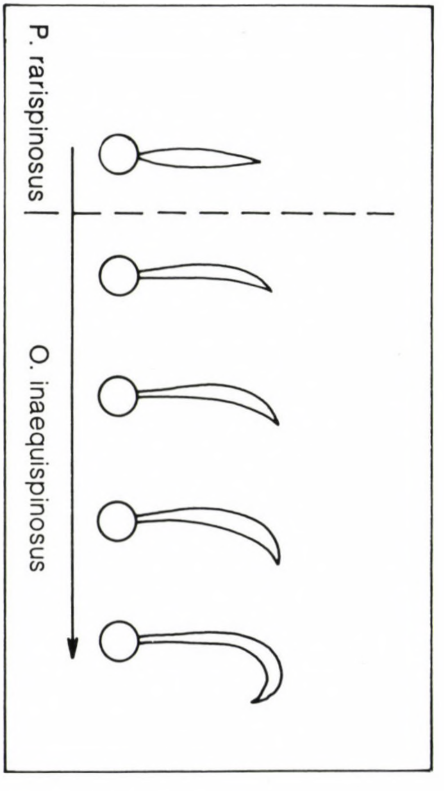


Fig. 6
The evolution of the Oertlispongus genus

Genus Falcispongus

Representatives of this genus occur from the Fassanian to the Cordevolian. As it is known so far it evolved from the genus Oertlispongus. Within the genus the species *Falcispongus calcaneum* is the most important, the range of wich is restricted to the Fassanian. Relationships within the genus are seen in Fig. 5.

Conclusions

Investigations carried out so far proved that the Radiolarian-zonation can be fairly well used and allows the distinction at substage level at least. In case of the Anisian/Ladinian boundary the lithological boundary (appearance of tuffaceous strata) favourably coincides with the radiolarian zone boundary proved by ammonites (Vörös et al. 1991). In case of the Ladinian/ Carnian boundary this coincidence does not exist since ammonites are nearly completely lacking in this interval in the Balaton Highland. Joint radiolaria and conodont investigations proved that the boundary of the two fossil groups coincide (Kovács et al. 1991). All these facts evidence that in lack of other fossils the radiolarians are highly suitable to solve stratigraphic problems.

Acknowledgements

The author is indebted to Hungarian Scientific Research Fund (OTKA) No. 2671 for the support to prepare this article.

Plate I

1. *Planispinocyrtis baloghi* Kozur et Mostler, Felsőörs, Malomvölgy No. 87, 360x; 2. *Foremanella macrocephala* Dumitrica, Felsőörs, Malomvölgy No. 87, 400x; 3. *Archaeo- spongoprunum mesotriassicum* Kozur et Mostler, Felsőörs, Malomvölgy No. 87, 160x; 4. *Archaeospongoprunum mesotriassicum* Kozur et Mostler, Felsőörs, Malomvölgy No. 87, 150x; 5. *Paroertlispongus* aff. *rarispinosus* Kozur et Mostler, Felsőörs, Malomvölgy No. 87, 120x; 6. *Spongoserrula rarauna* Dumitrica, Borehole Palóznak-1 29.2 m, 150x; 7. *Eptingium manfredi* Dumitrica, Mencshely, Csertető No. 16 , 150x

Plate II

1. *Paroertlispongus rarispinosus* Kozur et Mostler, Mencshely, Csertető No. 16, 120x; 2. *Paroertlispongus multispinosus* Kozur et Mostler, Mencshely, Csertető No. 16, 72x; 3. *Oertlispongus inaequispinosus* Dumitrica et Kozur et Mostler, Mencshely, Csertető No. 16, 100x; 4. *Oertlispongus inaequispinosus* Dumitrica et Kozur et Mostler, Mencshely, Csertető No. 16 150x; 5, 6. *Oertlispongus inaequispinosus* Dumitrica et Kozur et Mostler, Mencshely, Csertető No. 16, 100x; 7. *Oertlispongus inaequispinosus* Dumitrica et Kozur et Mostler, Felsőörs, Malomvölgy No. 105, 240x; 8. *Oertlispongus inaequispinosus* Dumitrica et Kozur et Mostler, Borehole Palóznak-1 57.9 m, 200x

Plate III

1. *Hungarosaturnalis multispinosus* Kozur et Mostler, Borehole Palóznak-1 29.2 m, 200x; 2. *Hungarosaturnalis multispinosus* Kozur et Mostler, Borehole Palóznak-1 28.6 m, 220x; 3. *Hungarosaturnalis longobardicus* Kozur et Mostler, Borehole Palóznak-1 29.2 m, 200x; 4. *Muelleritortis cochleata hungarica* Dosztály, Borehole Palóznak-1 29.2 m, 100x; 5. *Muelleritortis cochleata* (Nakaseko et Nishimura), Borehole Palóznak-1 28.6 m, 150x

Plate IV

1. *Paleosaturnalis triassicus* (Kozur et Mostler), Balatonfüred, Száka Hill No. 23, 200x; 2. *Praeheliostaurus levis* Kozur et Mostler, Balatonfüred, Száka Hill No. 23, 180x; 3. *Paleosaturnalis zapfei* (Kozur et Mostler), Balatonfüred, Száka Hill No. 23, 200x; 4. *Spongosaturnaloides multidentatus* Kozur et Mostler, Balatonfüred, Száka Hill No. 23, 200x

Plate I

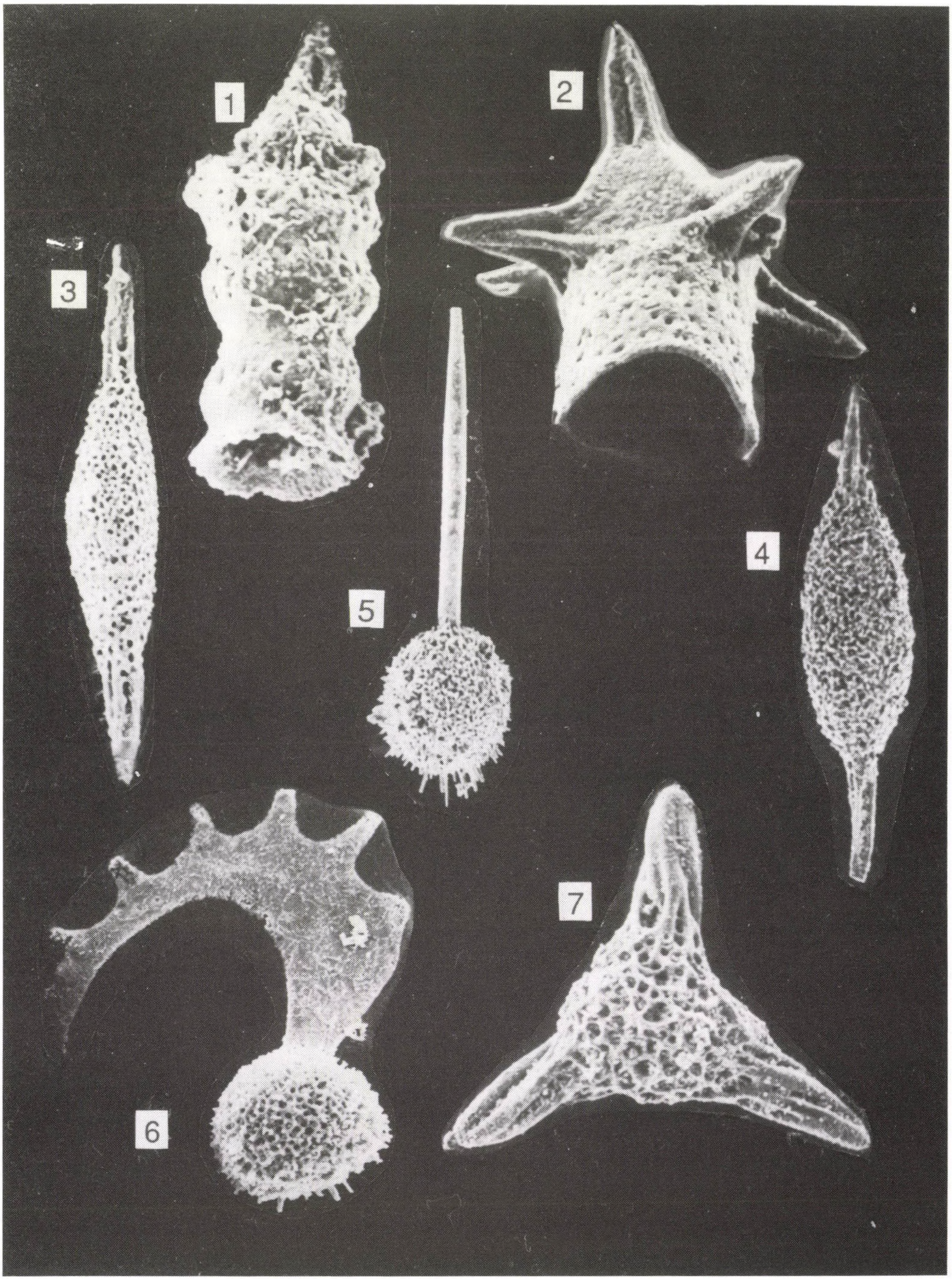


Plate II

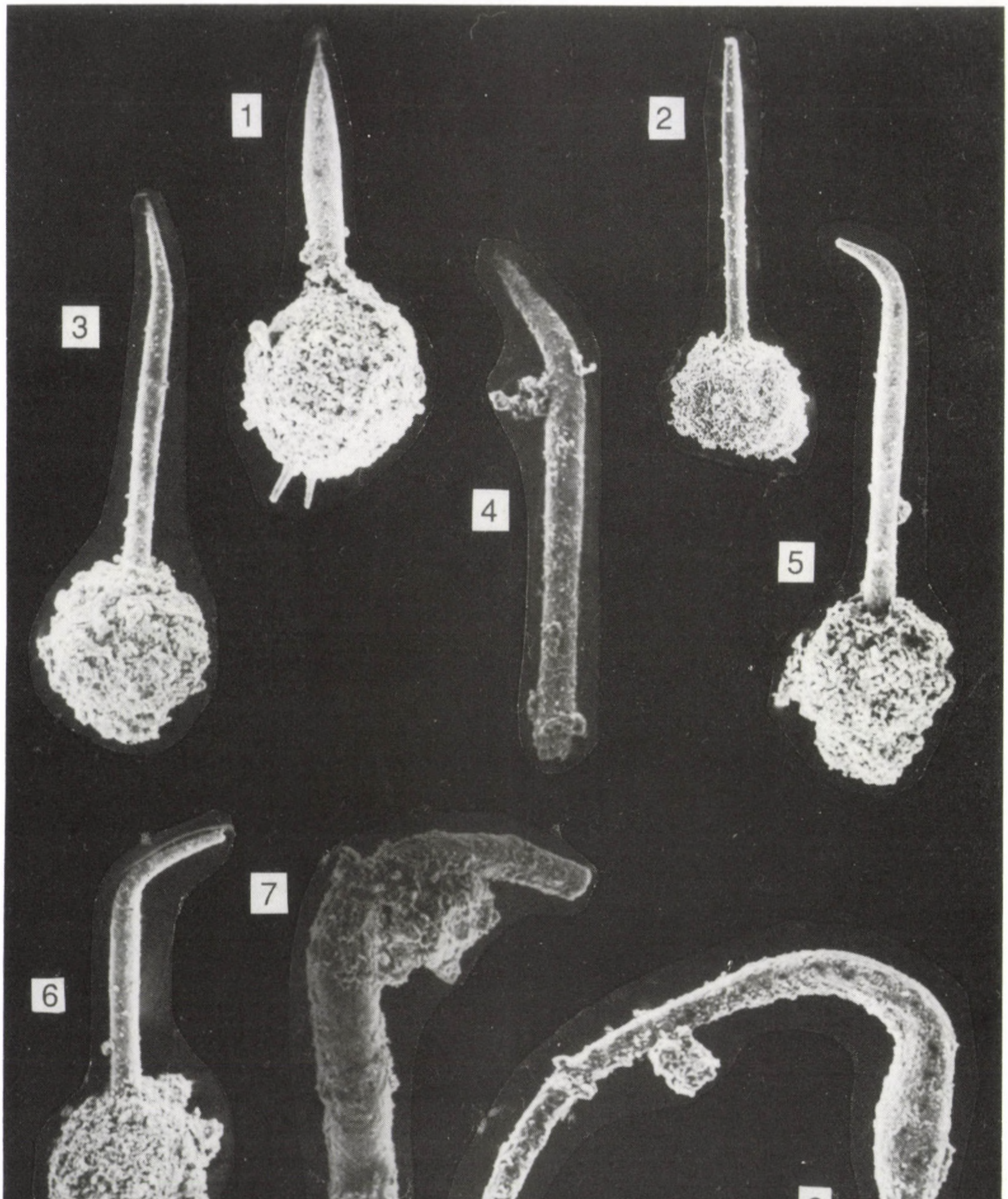


Plate III

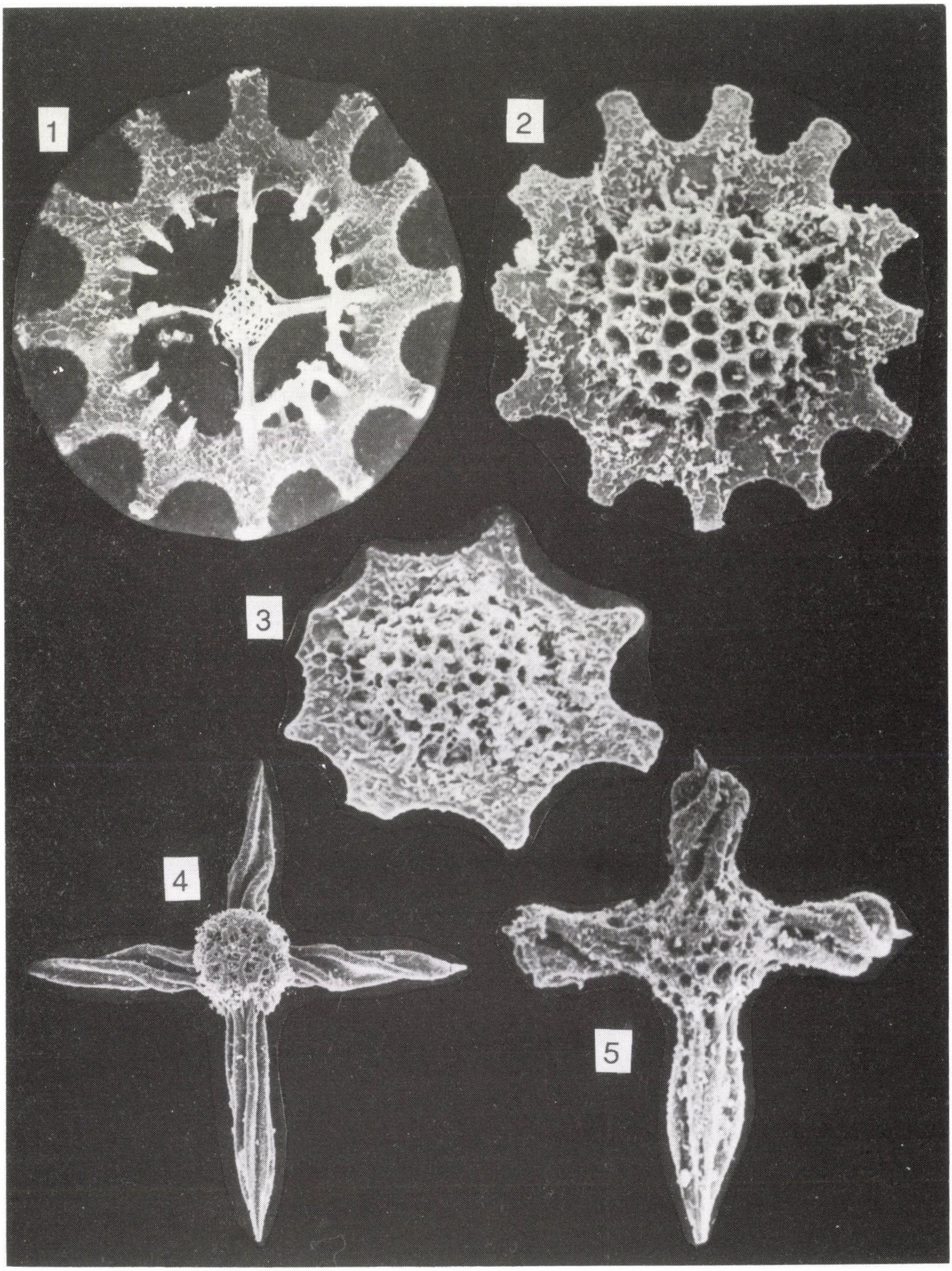
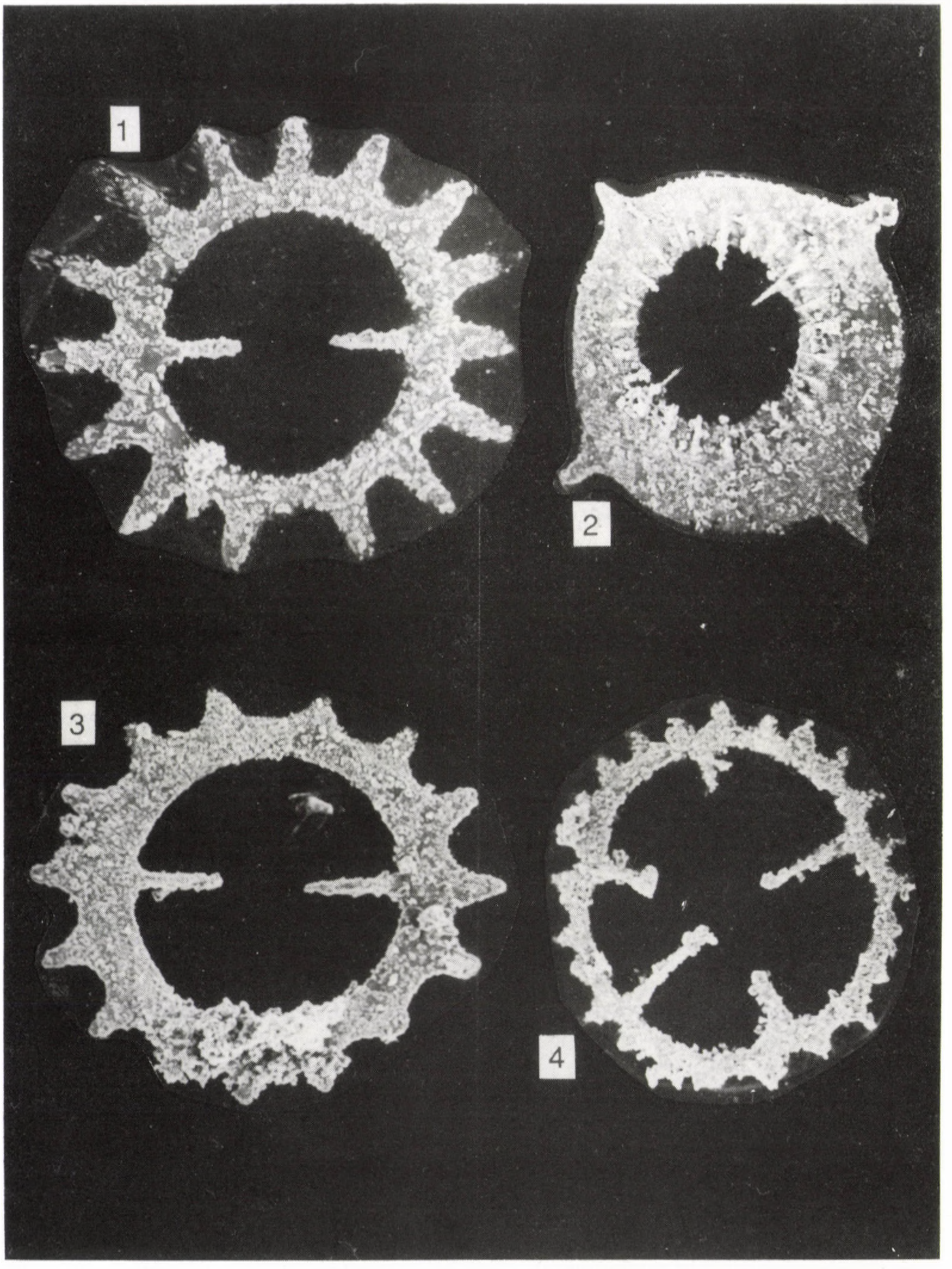


Plate IV



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The Anisian/Ladinian boundary in the Transdanubian Central Range based on palynomorphs and foraminifers

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Based on the palynostratigraphic and foraminifer-studies of Middle Triassic sections of the Balaton Highland and of the Bakony Mountains a proposal is made to draw the parachronostratigraphic boundary between the Anisian and Ladinian. The proposal is based on the changes recognized in the sporomorph and foraminifer associations in the boundary beds of the Felsőörs Limestone Formation and Tagyon Limestone Formation, and of the Buchenstein Formation. These can beseen in the predominance relations and in the joint appearance of new genera and species. The change is considered as event of stage grade. The suggested boundary almost coincides with the lower boundary of the *Xenoprotrachyceras reitzi* biozone s.l. of classical sense.

Key words: Parachronostratigraphy, Anisian, Ladinian, Illyrian, Fassanian, Foraminifera, sporomorpha, Transdanubian Central Range, Hungary

Introduction

In the course of Triassic biostratigraphic investigations carried out in the past decade we had the opportunity to study sequences in which the Anisian/ Ladinian parachronostratigraphic boundary could be marked either on palynostratigraphic or on foraminifer-stratigraphic bases, occasionally by both methods.

Sections in question are found in the area of the Transdanubian Central Range, in the Balaton Highland and in the Bakony Mountains (Fig. 1). The majority of them is found in boreholes, some of them belong to surface outcrops (Fig. 2).

Borehole sections:

Szentantalfa (Szaf-1), Balatonfüred (Bf-1), Bakonykúti (But-2), Várpalota (Vpt-3), Bakonyszűcs (Bsz-3).

Among the surface exposures those at Felsőörs (Forrás Hill, Malomvölgy ditch) and the road cut at Vörösberény (Megyehegy) are worthy of mention.

The stratigraphic columns and chronostratigraphic classification of the studied sections are discussed below.

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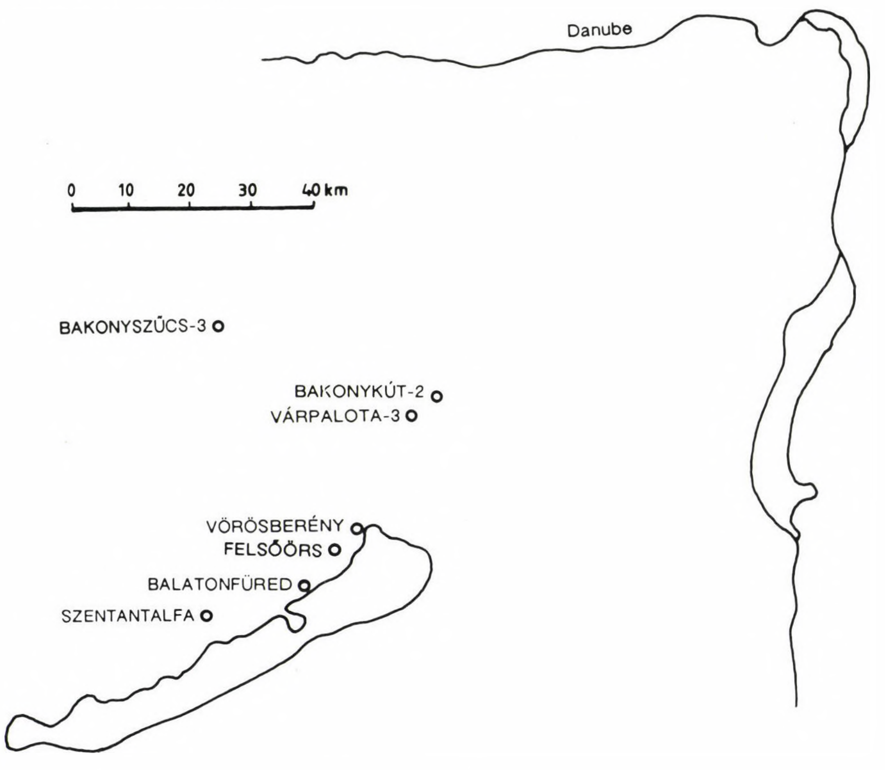


Fig. 1
Localities of studied profiles and boreholes

Investigated sections

Borehole Szentantalfa, Szaf-1 (after I. Szabó)

0.0– 34.5 m: Nemesvámos Limestone Member: red limestone of basin facies that was assigned partly to the Longobardian.

34.5– 69-5 m: Buchenstein Formation: tuffaceous, tuffitic limestone beds, characteristic of the Fassanian Substage.

69.5–197.0 m: Tagyon Limestone Formation: white biogenic limestone of carbonate platform facies. Age: Pelsonian–Illyrian.

197.0–204.5 m: Megyehegy Dolomite Formation. Age: Pelsonian. Drilling was stopped in this formation.

Borehole Balatonfüred, Bf-1 (after I. Szabó)

4.5– 66.0 m: Veszprém Marl Formation. Age: Cordevolian.

66.0–122.0 m: Füred Limestone Formation. Age: Cordevolian and Longobardian (?).

122.0–136.0 m: Nemesvámos Limestone Member: red, nodular limestone beds. Age: Fassanian–Longobardian.

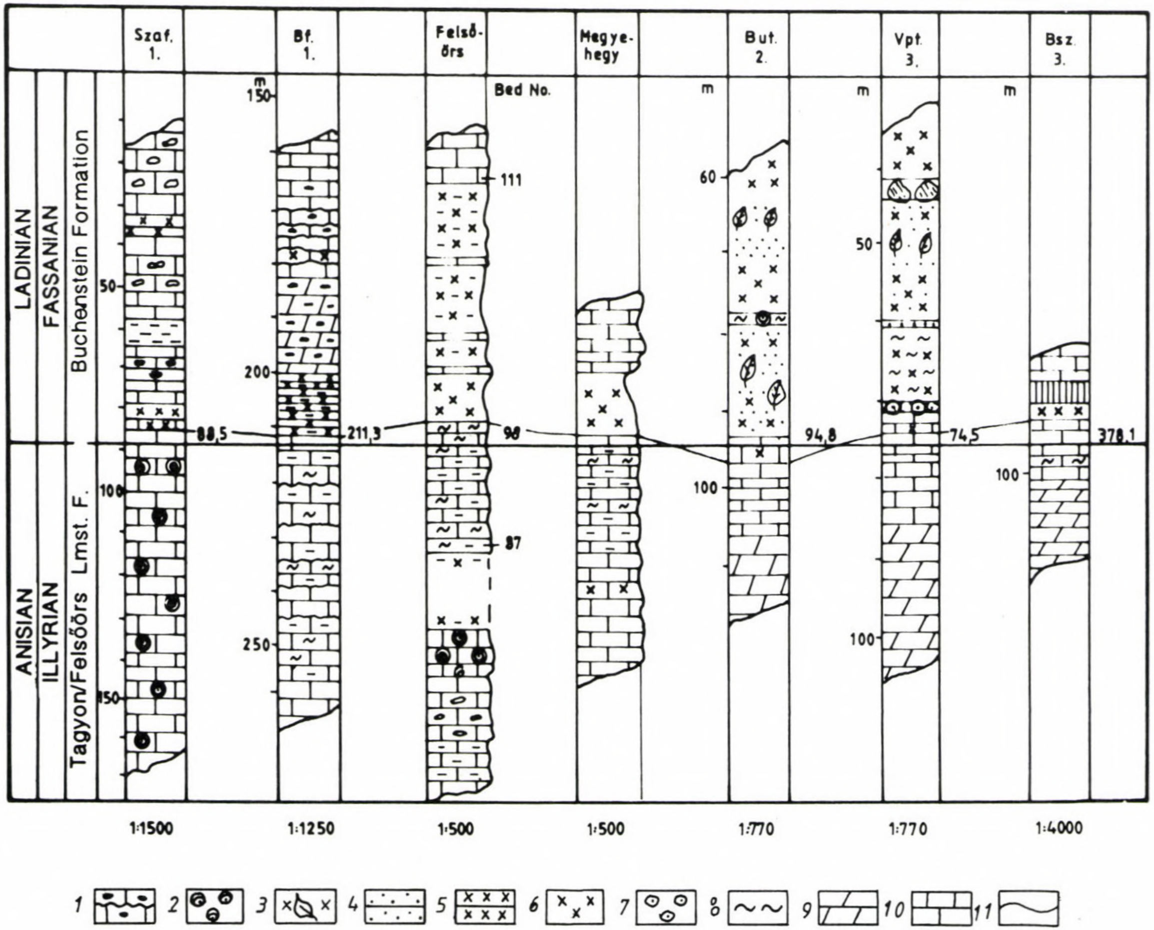


Fig. 2

Anisian–Ladinian boundary and its relation to the lithostratigraphic boundary in the studied profiles. 1. bituminous cherty limestone; 2. dasycladacean limestone; 3. tuff with plant remnants; 4. sandstone; 5. tuffitic beds; 6. tuff; 7. crinoidal limestone; 8. marl; 9. dolomite; 10. limestone; 11. boundary of formations. For abbreviations of borehole names see the chapter "Investigated sections".

136.0–211.1 m: Buchenstein Formation: tuffaceous, tuffitic, deep water carbonate formations. Age: Fassanian.
211.1–274.5 m: Felsőörs Limestone Formation: the major part is bituminous limestone, the upper 20 m consists of marly limestone. Age: Pelsonian–Illyrian.
274.5–299.0 m: Megyehegy Dolomite Formation: dolomitic marl and dolomite beds. Age: Pelsonian.

Borehole Bakonykúti, But-2 (after Gy. Raincsák)

In the roughly 124 m thick sequence of the borehole the Upper Anisian formations and the Lower Ladinian volcano-sedimentary beds of the Buchenstein Formation are tectonically repeated. The stratigraphic column is as follows:

0.0– 4.6 m: yellowish-brown limestone with marl spots. Age: Anisian.
4.6– 16.2 m: Gravelly tuffite. Age: Tertiary.
16.2– 36.1 m: Buchenstein Formation: bentonitized tuffaceous sandstone and crystal tuff, that develop continuously from the underlying Anisian carbonates.
36.1– 54.6 m: grey limestone with echinoids in the lower and stilolites in the upper section. Age: Anisian.
54.6– 73.0 m: Buchenstein Formation: dark-grey crystal tuff with plant remnants, in the lower section with a thin crinoidal-brachiopodal marl layer.
73.0– 93.0 m: Buchenstein Formation: black tuffaceous sandstone with plant remnants and crystal tuff layers, the basis of which consists of thin crystalline limestone. Ages: Lower Fassanian based on the rich sporomorph content.
93.0–102.3 m: bentonitized crumbling tuff and grey limestone with yellow spots, crinoidal limestone; these strata alternate. Age: Upper Anisian.
102.3–128.0 m: Megyehegy Dolomite Formation, Age: Pelsonian.

Borehole Várpalota, Vpt-3 (after Gy. Raincsák)

The sequence of Upper Anisian and Lower Ladinian formations significant from the aspect of the Anisian/Ladinian boundary can be considered as continuous. Faults observed in the upper 20 m of the borehole, in the upper section of the Buchenstein Formation as well as in the upper section of the Megyehegy Dolomite Formation do not affect the problem of boundary. The stratigraphic column of the 168 m long borehole is as follows:

4.6– 12.0 m: purplish-brown crinoidal, gastropodal limestone. Age: Fassanian (based on the foraminifers).
12.0– 14.3 m: Daonella-bearing marl. Age: Fassanian.
14.3– 20.0 m: gravelly tuff. Age: Tertiary. Its lower and upper boundary is formed by fault planes.
20.0– 63.0 m: Buchenstein Formation: alternation of bentonitized tuff and marl as well as of bentonitized tuffaceous sandstone and crystal tuff layers.
63.0– 70.0 m: greyish-brown (in the upper part) crinoidal limestone and bentonite showing green, grey and yellow color. Age: Fassanian.
70.0– 78.4 m: the lower and upper layers are crinoidal limestone, in the middle section bentonite, bentonitized tuff and dark-grey dolomitic-ankeritic limestone are found. Age: Upper Illyrian–Lower Fassanian.
78.4–101.9 m: Megyehegy Dolomite. Age: Upper Anisian.
101.9 m: fault zone
101.9–168.0 m: dolomite. Age: Carnian?

Borehole Bakonyszűcs Bsz-3 (the Middle and Upper Triassic section after G. Császár, the Lower Triassic after I. Szabó)

211.0–232.5 m: Veszprém Marl Formation. Age: Cordevolian.
232.5–247.0 m: Nemesvámos Limestone Member. Age: Ladinian.
247.0–247.3 m: fault zone.
247.3–360.6 m: Buchenstein Formation consisting of siliceous shale, tuffite and limestone beds. Age: Lower Ladinian.
360.6–396.0 m: Felsőörs Limestone Formation. Age: Illyrian.
396.0–442.0 m: Megyehegy Dolomite Formation. Age: Pelsonian.
442.0–552.5 m: transitional formation of the Megyehegy Dolomite and Iszkahegy Limestone Formations.
552.5–600.0 m: Iszkahegy Limestone Formation. Age: Lower Anisian.
600.0–653.7 m: transitional formations of the Iszkahegy and Aszófő Formations. Age: Lower Anisian.
653.7–758.0 m: Aszófő Dolomite Formation. Age: except the lowermost 3.2 m Lower Anisian, this 3.2 m is Scythian: Olenekian.
758.0–886.0 m: Csopak Marl Formation. Age: Olenekian.
886.0–930.0 m: Hidegkut Formation. Age: the upper dolomitic section is Olenekian, the main section with sandstone and sandy siltstone is Induan.

Felsőörs, Forrás Hill; stratigraphic column of the Malomvölgy ditch (after I. Szabó et al.)

The Megyehegy Dolomite Formation is the oldest rock in the sequence. Age: Lower Anisian, perhaps Pelsonian.

Felsőörs Limestone Formation. Key section of this formation. It consists of three fairly well-distinguishable sections. The lower part is built up by grey, bedded limestone with chert nodules and with thin marl intercalations. Age: Lower Pelsonian. The middle part consists of crinoidal, brachiopodal limestone. Age: Upper Pelsonian.

In the upper section clayey limestone beds with tuffaceous intercalations are characteristic, with rich macro- and microfauna. Age: Illyrian.

Buchenstein Formation: above the Felsőörs Limestone the tuffaceous sequence with limestone intercalations of the Buchenstein Formation is exposed by a test pit, with micro- and macrofauna. Age: Lower Ladinian, Fassanian substage. These tuffaceous beds are overlain also by the Nemesvámos Limestone Member that can completely be assigned here to the Fassanian (based on the Conodonts investigations).

Vörösberény, Megyehegy, road cut exposure (after I. Szabó and A. Vörös)

The section of well-exposed state begins here also with the Megyehegy Dolomite Formation out of the Middle Triassic formations. The upper part of the formation consists of greyish-brown, microcrystalline bituminous dolomite beds. Age: Pelsonian.

It is overlain by the Felsőörs Limestone Formation consisting of brownish-yellow marly beds with intercalated siliceous limestone lenses and banks. This is followed by a thin-banked cherty limestone in the upper section of which the crinoids and brachiopods are frequent. The upper section of the Formation is built up by a sequence consisting of the alternation of well-stratified siliceous limestone and marl. The transition towards the overlying Buchenstein Formation is continuous. Marl is replaced by tuffitic marl and calcareous tuffite. Among them siliceous limestone is intercalated. The quantity of volcanics decreases gradually towards the overlying beds. At the upper end of the exposure nodular cherty limestone follows that represents the Nemesvámos Limestone Member.

Biostratigraphic-parachronostratigraphic evaluation

Since in this paper solely the drawing of the Anisian/Ladinian boundary is in question, in Fig. 2 only the Upper Anisian and Lower Ladinian sections of those discussed above are demonstrated.

The justification of drawing the boundary is based not only on the theoretical legality of parachronostratigraphy but also on the impelling fact that in the borehole sections of the Middle Triassic the ammonite species of ortho-stratigraphic value are much less frequent than the disperse sporomorphs and foraminifers. Further, we recognized that in the Middle Triassic sporomorph and foraminifer assemblages the appearances of new elements coincide, the appearance of which can be evaluated as stage-rank change within the Triassic. In addition to these professional arguments the subjective inspiration should also be mentioned, i.e. in different issues of Albertiana (1, 2, 5, 10) we systematically follow the discussions and suggestions (thanks to the editorial board!) of ammonite specialists in this topic – without agreement so far.

We believe that our suggestion not only increases the number of solution possibilities but do hope that it will be successful to call the attention of experts dealing with this problem to the possibilities of evaluating these two groups of fossils.

Taking this aspect into consideration the statements below can be made on the studied sections:

– to mark the Anisian/Ladinian boundary by foraminifer stratigraphy the Malomvölgy exposure (Felsőörs, Forrás Hill) is of decisive value since here the ammonite and foraminifer fauna occur together in the Trinodosus Zone.

– the biostratigraphic value of the benthic foraminifer assemblages of basin facies can be evidenced to the sections where foraminifers occur without ammonites but with sporomorph assemblages together.

– different sporomorph assemblages have their own biostratigraphic value based on the phylogenetic trend of their own group, that can be verified orthostratigraphically through the foraminifers, and *vice versa*:

– the foraminifer assemblages of other taxon composition of the formations differing in facies from basin sediments (e.g. different lagoon types) may get biostratigraphic rank through the sporomorph associations of defined value.

In case of the studied sections believed to be suitable to mark the boundary these principles are manifested as follows:

In the Malomvölgy exposure the biostratigraphic value of the ammonites of the Upper Anisian Trinodosus Zone authenticates the forams of basin facies by their common occurrence: *Eoophthalmidium tricki, Hemigordius amylovolutus, H. chialingchiangensis, Tolypammina milanis, Planiinvoluta mesotriassica, Diplotremmina astrofimbriata*.

At the same time, this basinal foraminifer assemblage prove (by their common taxa: *Hemigordius chialingchiangensis, Glomospirella triphonensis*) the simultanity of the foraminifers and sporomorphs of the open marls, penetrated of 380.0–384.5 m of the borehole Bakonyszűcs, Bsz-3. These associations are the following: forams: *Glomospirella shenghi, Gl. triphonensis, Hemigordius chialingchiangensis, H. amylovolutus, Diplotremina astrofimbriata:* sporomorphs: *Crassosphaera triassica, Cyclotriletes margaritatus, C. oligogranifer, Verrucosisporites remyanus, V. thuringiacus, Densosporites fissus, Uvaesporites fueredensis, Concentricisporites nevesi, Dyupetalum vicentinense, Stellapollenites thiergartii, Triadispora crassa, Tr. suspecta, Tr. obscura, Angustisulcites div. sp., Striatoabieites aytugii, Strotersporites tozerii.*

The rich sporomorph assemblage of the same taxon composition verifies back this in the upper 20 m thick marly limestone section of the Felsőörs Formation in the Balatonfüred, Bf-1 borehole.

Here in addition to the above listed taxa the following are worth mentioning: *Punctatisporites fungosus, P. triassicus, Convolutispora jugosa, Triadispora stabilis*, and among the first appearing: *Sellaspora rugoverrucata, Aratrisporites fimbriatus*. In lack of micro- and macrofauna these were defined to be of Upper Illyrian age on the basis of observing the phylogenetic trend of Anisian sporomorph assemblages.

In the borehole Várpalota, Vpt-3 the uppermost beds of Illyrian are pollen-free. Nevertheless, in the depth interval 75.8–89.8 m Upper Anisian foraminifers are found such as *Glomospirella triphonensis* and *Diplotremina astrofimbriata*. The Illyrian age of the host rock is verified, however, by the fact that in the unbroken sequence upward by 1.3 m, i.e. at 74.5 m a typical Lower Fassanian sporomorph assemblage occurs: *Kuglerina meieri, Cannanoropollis scheuringii, Cannanoropollis brugmani, Triadispora* div. sp. Thus, in this borehole the two micro-groups are able to mark only together the Anisian/Ladinian boundary since Ladinian foraminifers are found only much more upwards, in 56.8 m: *Paratriassina jiangyouensis, Diplotremina altoconica* and *Turriglomina mesotriassica*.

Similar considerations are applied in case of Upper Illyrian and Lower Fassanian in the borehole Bakonykúti, But-2. The upper Illyrian is characterized here only by foraminifers, by Illyrian species occurring in 98.8 and 96.6 m and calibrated with ammonites of the Trinodosus Zone in the Malomvölgy section at Felsőörs: *Ophthalmidium tricki, Hemigordius chialingchiangensis*, accompanied by *O. amylovolutus* and *O. uebeyliense*. Though in the assemblage of reduced number of individuals of 92.2 m the *"Pilaminella" gemerica* and *Gauryinella elegantissima* known only from the Ladinian occur, the first beds of the Fassanian are indicated between 94.0 and 94.8 m by the sporomorph assemblage: *Kuglerina meieri, Cannanoropollis scheuringii, Cannanoropollis brugmani* and this verifies the status of Upper Illyrian foraminifers, as well.

As the predominating role of foraminifers is unambiguous in case of carbonate sequences such as the Malomvölgy section, the Megyehegy road cut or the Tagyon Limestone Formation consisting merely of biogenic limestone (Szentantalfa, Szaf-1 borehole), sporomorphs may become of only stratigraphic value in lithofacies such as the Anisian sequence of the borehole Balatonfüred, Bf-1. In this borehole we had the possibility to follow the evolutionary changes of the Anisian vegetation by one meter average samples, and to evaluate the Anisian/Ladinian boundary formations on the basis of rich sporomorph assemblages having carried out the sampling by 10 cm density. It was also this borehole where the palynostratigraphic characteristics could be formulated that mean the chronostratigraphic value of the Upper Illyrian and Lower Fassanian assemblages (Fig. 3).

Palynostratigraphy

On these basis, discussed above, those sporomorph assemblages can be defined as Illyrian which are rich in pteridophyte spores (*Punctatisporites punctatus, P. fungosus, Cyclotriletes hians, C. margaritatus, C. oligogranifer, Verrucosisporites thuringiacus, V. remyanus, V. slevecensis, Verrucosisporites div. sp., Concentricisporites nevesi, Convolutispora jugosa, Sellaspora rugoverrucata, Reticulatisporites bunteri, Retitriletes* div. sp., *Densosporites fissus, Uvaesporites fueredensis, U. gadensis, Dyupetalum vicentinense*) and in which among the bisaccate forms (Chordasporites, Illinites, Angustisulcites, Triadispora, Striatoabieites, Strotersporites) the representatives of the Triadispora (*Tr. crassa, Tr. falcata, Tr. obscura, Tr. suspecta, Tr. stabilis*) predominate. Furthermore, *Concentricisporites nevesi, Strotersporites tozerii, Stellapollenites thiergartii* and *Dyupetalum vicentinense* systematically occur in them, but *Kuglerina meieri, Cannanoropollis scheuringi* and *Ca. brugmani* or the extremely varied forms of these species are still lacking.

An assemblage of this kind is found in the borehole Balatonfüred, Bf-1 in the samples of Felsőörs Limestone Formation taken from 211.4 and 240.0 m, as well as in the borehole Bakonykúti, But-2 in 39.0 m. A sporomorph assemblage of the same composition occurs in the uppermost section of the Felsőörs Limestone Formation penetrated by the borehole Bakonyszűcs, Bsz-3 in 378.1 m, together with Illyrian conodonts that are present also in the Felsőörs section of the Felsőörs Limestone (*Gondolella constricta cornuta, Gondolella liebermani*).

Those assemblages with Triadispora predominance should be qualified as Fassanian, in which among the species of this genus those described by Scheuring (1970, 1978): *Tr. obscura, Tr. barbata, Tr. bella, Tr. Boelchii, Tr. sulcata,*

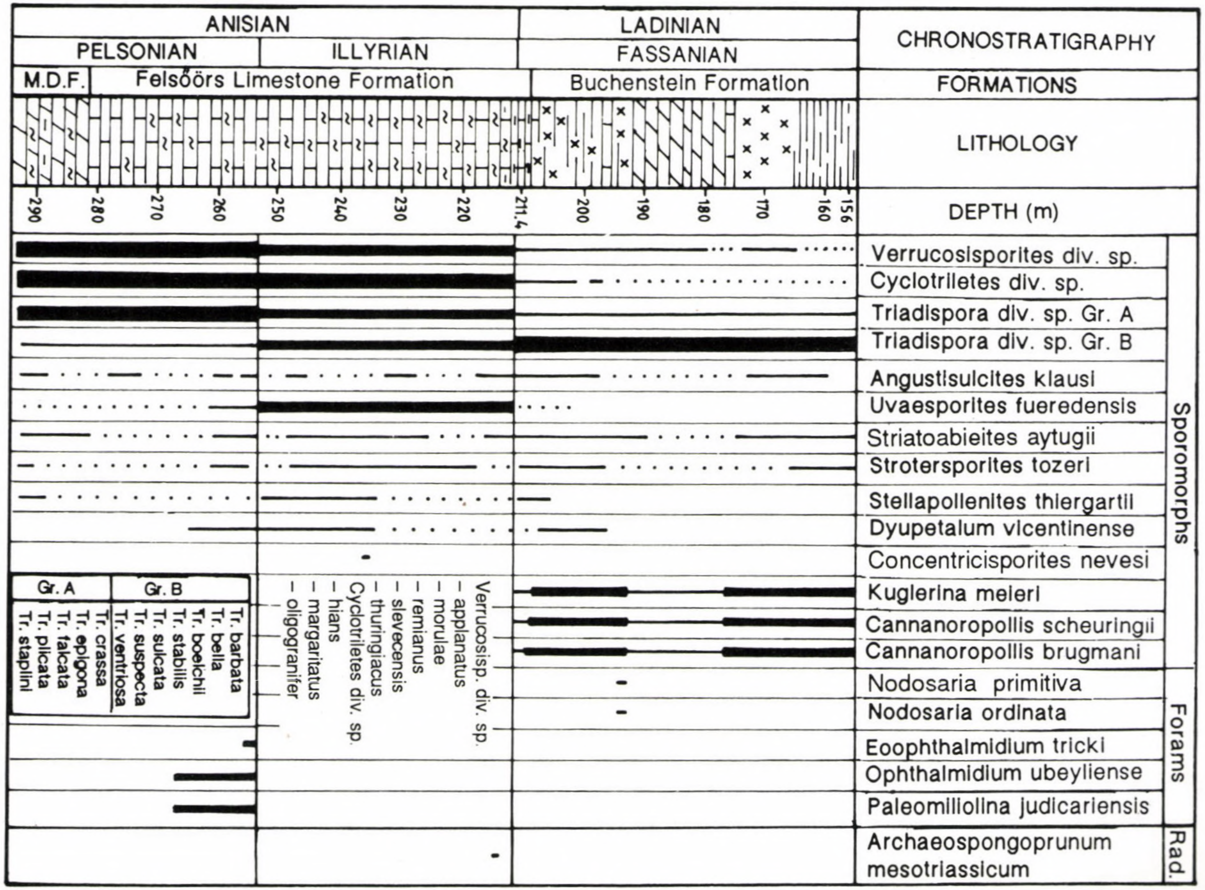


Fig. 3
Characteristic sporomorph, foraminifer and radiolarian taxa of borehole Balatonfüred, Bf-1

Tr. ventriosa are more frequent than those of Klaus (1964): *Tr. crassa, Tr. epigona, Tr. falcata, Tr. plicata, Tr. staplini,* and in which the appearing Cannanoropollis genus alone or together with the also appearing *Kuglerina meieri* is of subdominating value. Furthermore, Dyupetalum and Concentricisporites, as well as Mädler's species (1964a) of Stellapollenites and Cyclotriletes are rare or are completely absent.

Sporomorph assemblages of this kind occur in 211.4 m of the borehole Balatonfüred, Bf-1, in the sample of 94.0–94.8 m of the borehole Bakonykúti, But-2, further in 74.5 m of the borehole Várpalota, Vpt-3, in the altered tuffitic marl intercalation at 309.5 m in the Felsőörs Limestone Formation explored by the borehole Várpalota-316, and in 371.3–377.8 m in the borehole Bakonyszűcs, Bsz-3.

Foraminifer-stratigraphy

The change marked by predominance exchange in the sporomorph assemblages and by the appearance of two genera of wide range is evaluated as stage-rank event. This coincides with the appearance of new elements in the foraminifer assemblages of basin facies, lagoonal environments and transitional facies that become more frequent in the younger formations of the Ladinian.

On this basis among the assemblages of *platform-lagoonal environments* the associations will be qualified as Upper Anisian in which the taxa below are of association-forming role:

Meandrospira dinarica Kochansky Dev. et Pant.
Endothyranella wirzi (Koehn Zaninetti)
Ammobaculites radstadtensis Kristan
Earlandinita oberhauseri Salaj
Duostomina magna Trifonova
Trochammina almtalensis Koehn Zan.

Among the studied sections the assemblage of this composition could be identified in the Tagyon Limestone Formation explored by the borehole Szentantalfa Szaf-1.

The Upper Anisian assemlages of the *deeper basin facies* are characterized by the taxa

Eoophtalmidium tricki (Langer)
Hemigordius amylovolutus (He)
Ophthalmidium uebeyliense Dager
Hemigordius chialingchiangensis (Ho)
Planiinvoluta mesotriassica Baud et Zan.

but in addition to these the species more frequent in shallower environment may also occur, e.g. *Glomospirella triphonensis* Baud et al., *Trochammina almtalensis* Koehn. Zan., *Earlandia amplimuralis* Pantic.

Assemblages of this composition are found in the upper section of the Felsőörs Limestone Formation explored at Megyehegy and Malomvölgy, *"Transitional" types* (between lagoonal and basinal environments). These are, brachiopodal Upper Anisian marls and marly limestones, characterized by assemblage:

Glomospirella shengi He
Glomospirella triphonensis Baud et al.
Tolypammina milanis Urosevic
Ammodiscus multivolutus Reitlinger
Trochammina almtalensis Koehn Zan.

These associations are found in the upper section of the Felsőörs Formation in the borehole Bakonyszűcs Bsz-3. (Fig. 4), Bakonykúti But-2, Várpalota Vpt-3, as well as in the lower part of this formation exposed in Megyehegy and Malomvölgy.

The foraminifer assemblages of the Fassanian substage of the Ladinian are much more homogeneous than those of the Illyrian substage. In the marine sedimentation contemporaneous with the commencement of volcanic activity the fine facies differences are "smoothed" and in the more homogeneous biotope similar ecosystems developed also in the benthic world. Accordingly, the taxa characteristic of the Upper Anisian lagoon formations (Glomospirella species) disappear but the small-sized members of basin facies (Ophthalmidium and Hemigordius) survive though in reduced number of individuals, moreover, in the Early Ladinian a new species, *Hemigordius plectospirus* (Oravecz Sch.) appears in the Hungarian sections. The *Pseudonodosaria loczyi* Oravecz Sch. is also a new element, that together with the members of the family Nodosariidae is a predominating form of the Ladinian:

Pachyphloides klebelsbergi (Oberhauser)
Nodosaria raibliana Gümbel
Pseudonodosaria lata (Tappan)
Pseudonodosaria obconica (Reuss)
Austrocolomia ploechingeri (Oberh.)

The species below appear also in the Fassanian:

Turriglomina mesotriassica (Koehn Zan.)
Paratriassina jiangyouensis He
Oberhauserella ladinica (Fuchs)
Oberhauserella mesotriassica (Oberhauser)
"Pilaminella" gemerica Salaj
Triadodiscus eomesozoicus (Oberh.)

Major part of the appearing taxa becomes frequent in the Longobardian substage of the Ladinian.

4. Conclusions

Based on the Upper Illyrian and Lower Fassanian sporomorph and foraminifer assemblages listed above, the parachronostratigraphic Anisian/ Ladinian boundary is suggested to be marked in the studied sections in the depth intervals below (see Fig. 2):

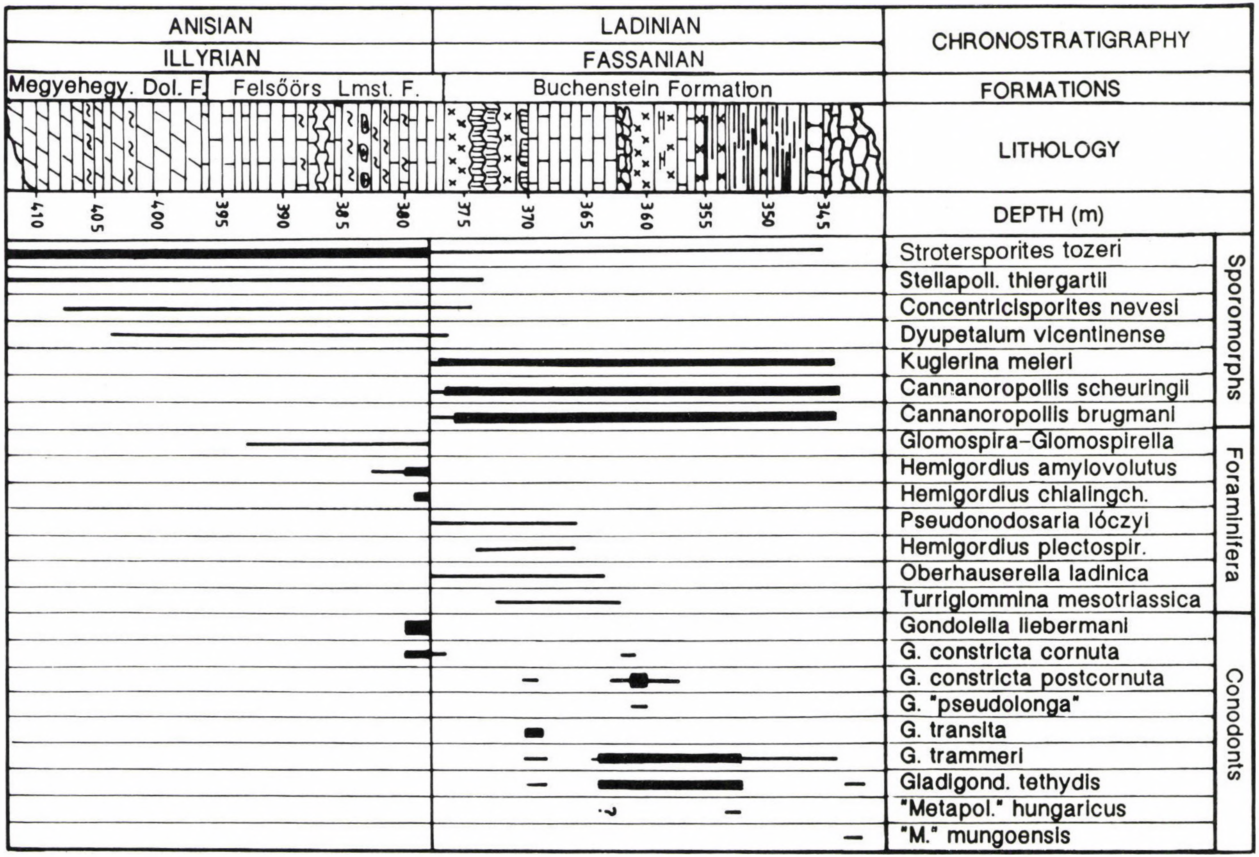


Fig. 4
Characteristic sporomorph, foraminifer and conodont taxa of borehole Bakonyszűcs, Bsz-3

borehole Szentantalfa, Szaf-1: 88.5 m
borehole Balatonfüred, Bf-1: 211.4 m
borehole Bakonykúti, But-2: 94.8 m
borehole Várpalota, Vpt-3: 74.5 m
borehole Bakonyszűcs, Bsz-3: 378.3 m
Malomvölgy section (Felsőörs, Forrás Hill): bed No. 98
Road cut (Vörösberény, Megyehegy): 55.0 m

This suggested boundary does not perfectly coincide with the Anisian/ Ladinian boundary proposed recently by A. Vörös (in the present volume) to the sections of the Balaton Highland on the basis of ammonites, that at the base of the Buchenstein Formation is defined by the appearance of *Kellnerites felsoeoersensis*. It was found slightly below the top of the underlying formation (Felsőörs or Tagyon Limestone) in all the investigated sections (see Fig. 2). Altough the change in the basinal foraminifer associations could be partly ecologically controlled, we would emphasize that it took place still below the formation boundary and within the same formation, underlineing their great significance for local correlation (e.g. of surface and borehole sections) in the Balaton Highland. It should also be stressed, that in our area no significant change in the sporpmorph and foraminifer associations can be recognized higher up in the Fassanian, until the base of the Longobardian (see the section of the borehole Bakonyszűcs 3 on Fig. 4, where it is correlated with conodonts). As sporomorphs are useful tools for long-distance correlation, this fact should also be considered at the definition of the Anisian/Ladinian boundary, widely discussed among ammonoid workers.

5. *Taxonomic remarks*

In order to give validity to *Cannanoropollis scheuringii* nov. sp., Brugman 1986, and *Strotersporites tozeri* nov. sp., Brugman 1986, we repeat here their original diagnosis by Brugman 1986a, because both of them are nomen nudum till now and both species are present in our material. Furthermore, we give the emended diagnosis of *Cannanoropollis scheuringii* Brugman 1986a, the diagnosis of *Cannanoropollis brugmani* sp. nov. and the diagnosis of *Uvaesporites fueredensis* sp. nov.

Original diagnosis by Brugman 1986a, pp. 73–74:

"*Cannanoropollis scheuringi*" Brugman, nov. spec. (Plate VII, 1A-D; 2A, B)

Holotype: Parasaccites sp. a in Scheuring, 1978 (Plate XXVI, 385, 386)

Derivatio nominis: Dr. B. Scheuring, Basel (Switzerland)

Diagnosis: Dilate/trilate monosaccoid pollen grains with a (sub)circular outline in polar view. The nexine constitutes a central body, characterized by a (sub)circular outline in plan view. A mark, generally reduced trilete, occasionally dilate or fully trilete, is present in the nexine; rays extend approx. to 1/3 to 1/2 of the central body radius. The nexine is enveloped by the sexine,

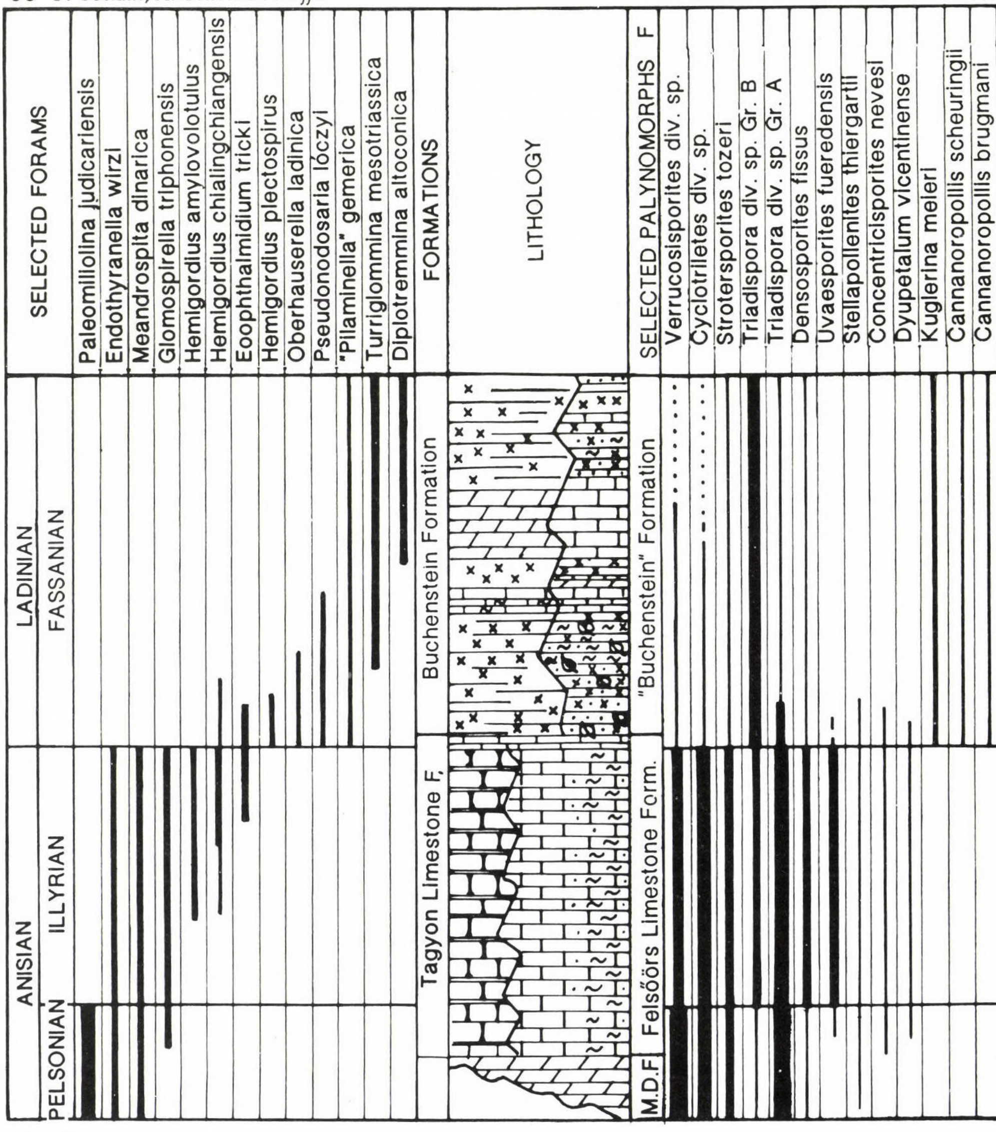


Fig. 5
Generalized profile and ranges of selected foraminifers and palynomorphs in the Anisian/Ladinian boundary interval of the Balaton Highland

to which it is fused. The proximal sexine is thin, showing an ornamentation of fine rugulae, grana, and/or an imperfect reticulum. Equatorial sexine differentiated as a monosaccoid expansion, showing a complex infrastructure built up by a three-dimensional network of sexinous elements, appearing as a reticulate to rugulate pattern in plan view. Distal sexine differentiated as a thin, finely rugulate to granulate region. The shape of the proximal and distal areas of thinned sexine is (sub)circular. The transition from the proximal and distal thinned areas to the monosaccoid expansion is relatively abrupt. The area, in which the monosaccoid expansion overlaps the central body is relatively large, maximum width of the expansion is upto 1/4 of the central body radius.

Known size range: 90-120 mm

Remarks and comparation: Cannanoropollis scheuringii has been erected to accomodate the forms already described by Scheuring (1978) as Parasaccites sp.a. It may also be remarked that forms designated as Enzonalasporites sp. A in Van Der Eem (1983) are also identical with the forms here placed within Cannanoropollis scheuringii.

Cannanaropollis scheuringii is here considered to differ from other species assignable to Cannanoropollis in possessing an essentially reduced trilete mark and a proximal sexine with a fine rugulate, granulate and/or imperfect reticulate pattern. The dilate and fully trilete forms are regarded to represent variations.

In a number of specimens, the presence of irregularly distributed verrucae-like elements has been observed (see Plate VII, 2A, B). It is here considered that these elements are the result of bacterial activity."

Original diagnosis by Brugman, 1986a, pp. 74–75:

Strotersporites tozeri Brugman, nov. spec. (Plate X, 1A, B; Plate XI, 1A, B)

Holotype: Strotersporites tozeri nov. spec., slide REC-1-75, Plate X, 1A ,B.

Derivatio nominis: Dr. E. T. Tozer, Geological Survey of Canada, Ottawa

Diagnosis: Predominantly bisaccoid pollen grains, occasionally with a U-shaped saccoid expansion (in plan view). The nexine constitutes a central body characterized by a (sub)circular to ellipsoidal outline in plan view. A monolete, dilate or reduced trilete mark, extending more than 1/2 of the central body radius, is present in the nexine and the sexine. Proximally the sexine is differentiated into ten or more taenia. The taeniae may encircle the mark, whereas the outermost taeniae usually encircle the mark completely. The taeniae (finely to coarsely infrastructured) may be complete, incomplete and/or branched, and are occasionally built up by rows of closely spaced verrucae. Laterally the sexine is differentiated as two saccoid expansion, or occasionally as a U-shaped saccoid expansion. The two saccoid expansions may show a relatively large variation in size and shape; they are built up by a complex infrastructure, composed of a three-dimensional network of sexinous element, appearing as an infrapunctate and/or infrarugulate pattern in plan view.

Observed dimensions: 40–100 mm.

Remarks and comparisons: Strotersporites tozeri is characterized by its large monolete, dilete or reduced trilete mark and ten or more taeniae encircling (completely or in part) the mark in the nexine. The latter characteristic can not be observed in the Permian species of Strotersporites (Wilson 1962; Klaus 1963). The species was figured as Strotersporites n. sp. by Visscher and Brugman (1981, Plate I, 4).

Cannanoropollis scheuringii Brugman 1986 emend
Plate: XXV, Fig. 1.

Emended diagnosis: monosaccoid, large, trilete pollen grains, with a triangular, circular or subcircular outer and inner contour in plan view.

The trilate mark often reduced, occasionally fully trilete. Rays extend cca to 1/3 to 1/2 of the central body radius. The nexine is enveloped by the sexine, to which it is fused. The central body is surrounded by a monosaccoid apparatus consisting of bundles of the radial elongated sexinous elements. The contact areas of the monosaccoid apparatus asymmetrical, which is narrower on the marked side of the body then on the other side. The proximal sexine is thin, showing an ornamentation of fine rugulae, grana, and/or an imperfect reticulum. The distal sexine differentiated as a thin, finely rugulate to granulate region. The shape of the proximal and distal areas of thinned sexine is subcircular. The area, in which the monosaccoid apparatus overlaps the central body is relatively large, maximum width of the expansion is up to 1/5 of the central body radius.

Remarks: in some volcano-sedimentary beds of our Lower Ladinian profiles Cannanoropollis specimens of good preservation are predominant. Among them, the large, ellipsoidal monolete forms are predominant with 80 to 90%. These specimens differ from the holotype of Cannanoropollis scheuringii Brugman 1986 in monolete mark and the structure and sculpture too.

Therefore we propose the following:

– to take out the Cannanoropollis specimens with monolete mark from the taxon of *Cannanoropollis scheuringii* Brugman 1986 and to introduce for them a new name: *Cannanoropollis brugmani* nov. sp.

– to emend the diagnosis of *Cannanoropollis scheuringii* Brugman 1986 for the Cannanoropollis specimens with fully trilete and reduced trilete mark.

Cannanoropollis brugmani nov. sp.

Plate XXV, Fig. 2, Plate XXVI, Figs 1–4.

Derivatio nominis: Dr. W. A. Brugman, Utrecht, (Netherland).

Holotype: specimen in slide No. 64981/1; 9.1–108.0; Plate XXVI, Figs 1–4.

Locus typicus: Várpalota, borehole Vpt-3.

Stratum typicum: 66.2 m; gray bentonitic marl, lower part of Fassanian.

Diagnosis: large, monosaccoid, monolete pollen grains, generally with ellipsoidal, occasionally (sub)circular outer and inner contours in plane view.

The pollen grain is constituted by a discusslike central body and a ring-like monosaccoid apparatus. The later surrounds the central body, building asymmetrical contact areas. Generally on the marked side (proximal side) the contact area is reduced, coinciding with the border of the central body. So, on the proximal side the whole surface is uncovered growing visible the infrarugulate or infrapunctate sculpture of the central body. Occasionally on the proximal side there is a narrow contact area, too, but it is never as wide as on the distal side.

The monosaccoid apparatus (consisting of bundles of the radially elongated sexinous elements) overlaps the distal side of central body, covering a relatively large part of it. So here the contact area is relatively wide, its max size is up to 1/5 of the diameter of the body. Generally at the inner boundary of the contact area the radially elongated sexinous elements are ending in an irregular network, forming uncovered field of the central part of central body showing its fine infrarugulate or infrapunctate structure. Besides this, on the distal side a special ornamentation is frequent, mostly among the circular specimens. This ornamentation consist of fine, rounded, flat verrucae, so-called maculae, 1–2 µm in diameter. Their number and situation is irregular along the contact area and on the central part of central body.

Size range: total length 85–120 μm.

Holotypus: total length 96 μm; total width 74 μm width of monosaccoid apparatus 9-11 μm; length of sulcus 37 μm; thickness of central body 2 μm; max width of contact area 18 μm.

Differential diagnosis: Cannanoropollis brugmani nov. sp. differs from the other known species of *Cannanoropollis genera* in its monolete mark, in the asymmetrical contact area, and in the structure of the uncovered field of the central body.

Remarks: concerning the irregularly situated verrucae–maculae sculptur elements, we suppose the following possibilities:

– they are original elements presenting separateble species characters,

– result of partial chemical solutions of the sexine, because the trails of bacterial activity are star-like.

Uvaesporites fueredensis sp. nov.

Plate X, Figs 7, 8–9, Plate XIV, Fig. 7., Plate XV, Fig. 1.

Holotypus: specimen in slide N.o 55950; 13.5–115.4; Plate X, Figs 8–9.

Derivatio nominis: Balatonfüred, Balaton Hill (Hungary).

Locus typicus: Balatonfüred, borehole Bf-1.

Stratum typicum: 261.0 m

Diagnosis: medium sized triangular, trilet, verrucate microspores. In E-plan the outer and inner contours are convex triangulars with rounded corners. The thickness of the wall is 2–3 μm. Exine is ornamented by verrucae-like elements. Wideness of verrucae is 2–8 μm, height of verrucae 1–6 μm. The rough elements are present on the whole distal side, and on the periphery of the proximal side. They often fuse along their basis, forming a wavy-line cingulum-like ornamentation in E-plan.

On the proximal side in the interradial area the exine may be smooth or ornamented by verrucae of 1–3 μm. Situation of verrucae is irregular. The Y-mark is well visible, mainly it is straight. Over the mark is running to the corners there is a wavy bundle-like thickening of exine.

Size range: 35–42 μm (measured on 10 specimens).

The holotype is 40 μm.

Differencial diagnosis: *Uvaesporites fueredensis* sp. nov. is similar to *Uvaesporites argenteaeformis* (Bolkhovitina 1953) Schulz 1967 and to *Uvaesporites reissingeri* (Reinhardt 1961) Lund 1977 by its smaller size by the irregular sporadic situation of verrucae and by the sculpture of the interradial field of proximal side. *Uvaesporites fueredensis* sp. nov. differs from *Uvaesporites reissingeri* (Reinhardt 1961) Lund 1977 by the rougher sculptur elements and by its situation on both-sides.

Based on the sculpture of exine, *Uvaesporites fueredensis* sp. nov. is also similar to *Uvaesporites gadensis* Praehouser-Enzenberg 1970, which is common in our material, too. They are clearly separable, because *Uvaesporites fueredensis* sp.

nov. is larger, and its exine is ornamented by much larger verrucae, than those of *Uvaesporites gadensis* Praehouser-Enzenberg 1970.

The occurrence of *Uvaesporites fueredensis* sp. nov. within the Triassic sequences of Balaton Highland shows the following trend: epakme: Pelsonian, akme: Upper Illyrian, parakme: Ladinian.

Remarks: Concerning to Helmar, A.'s conception (1981, p. 23.) about the *Uvaesporites argenteaeformis* and *Uvaesporites reissingeri,* I can not accept it, based on specimens of these taxa occurred in our material.

Plate I

Upper Pelsonian. 1. *Calamospora tener* (Leschik 1956) De Jersey 1962, Balatonfüred, Bf-1 borehole, 293.5 m; 2. *Todisporites minor* Couper 1958, Balatonfüred, Bf-1 borehole, 290.0 m; 3. *Verrucosisporites applanatus* Mädler 1964a, Balatonfüred, Bf-1 borehole, 290.0 m; 4. *Dictyophyllidites surangei* Bharadwaj et Singh 1963, Balatonfüred, Bf-1 borehole, 293.50 m; 5. *Punctatisporites fungosus* Balme 1963, Balatonfüred, Bf-1 borehole, 290.0. All magnifications 1000x

Plate II

Upper Pelsonian. 1. *Deltoidospora toralis* (Leschik 1956) Lund 1977, Balatonfüred, Bf-1 borehole, 293.5 m; 2. *Foveolatisporites* cf. *potoniei* Mädler 1964b, Balatonfüred, Bf-1 borehole, 290.0 m; 3–4. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 293.5 m; 5. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 290.0 m; 6. *Guttatisporites ambiguus* Tiwari et Vijaya 1980, Balatonfüred, Bf-1 borehole, 293.5 m; 7. *Crassosphaera triassica* Orlowska-Zwolinska 1979, Balatonfüred, Bf-1 borehole, 273.0 m. All magnifications 1000x

Plate III

Upper Pelsonian. 1–2. *Verrucosisporites thuringiacus* Mädler 1964a, Balatonfüred, Bf-1 borehole, 290.0 m. All magnifications 1000x

Plate IV

Upper Pelsonian. 1. *Platysaccus triassicus* Mädler 1964a, Balatonfüred, Bf-1 borehole, 293.5 m; 2. *Triadispora falcata* Klaus 1964, Balatonfüred, Bf-1 borehole, 293.5 m; 3–5. *Jugasporites conmilvinus* Klaus 1964, Balatonfüred, Bf-1 borehole, 293.5 m; 6. *Triadispora epigona* Klaus 1964, Balatonfüred, Bf-1 borehole, 293.5 m. All magnifications 1000x

Plate V

Upper Pelsonian. 1–2. *Stellapollenites thiergartii* (Mädler 1964a) Clement-Westerhof et al. 1970, Balatonfüred, Bf-1 borehole, 293.5 m; 3, 7. *Triadispora obscura* Scheuring 1970, Balatonfüred, Bf-1 borehole, 293.5 m; 4, 6, 8. *Triadispora crassa* Klaus 1964, Balatonfüred, Bf-1 borehole, 293.5 m; 5. *Triadispora suspecta* Scheuring 1970, Balatonfüred, Bf-1 borehole, 293.5 m. All magnifications 1000x

Plate VI

Upper Pelsonian. Balatonfüred, Bf-1 borehole, 293.5 m. 1. *Striatoabieites aytugii* Visscher 1966; 2. *Striatoabieites balmei* Klaus 1964; 3. *Striatoabieites* sp.; 4. *Triadispora crassa* Klaus 1964; All magnifications 1000x

Plate VII

Upper Pelsonian. Balatonfüred, Bf-1 borehole, 293.5 m. 1–6. *Strotersporites tozeri* Brugman 1986. All magnifications 1000x

Plate VIII

Illyrian. 1. *Crassosphaera triassica* Orlowska-Zwolinska 1979, Balatonfüred, Bf-1 borehole, 216.8–217.2 m; 2. *Crassosphaera triadica* Orlowska-Zwolinska 1979, Balatonfüred, Bf-1 borehole, 242.7–242.9 m; 3. *Spheripollenites balmei* Jansonius 1962, Balatonfüred, Bf-1 236.0 m; 4. *Todisporites minor* Couper 1958, Balatonfüred, Bf-1 borehole, 227.0–227.3 m; 5–6. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 254.2–254.4 m; 7. *Punctatisporites* sp. Balatonfüred, Bf-1 borehole, 223.4–224.0 m; 8. *Punctatisporites fungosus* Balme 1963, Balatonfüred, Bf-1 borehole, 229.4–230.0 m; 9. *Leptolepidites ipsviciensis* De Jersey 1962, Balatonfüred, Bf-1 borehole, 227.2–227.3 m; 10–11. *Foveosporites* sp. nov. ind. Balatonfüred, Bf-1 borehole, 244.7–244.9 m; 12. *Densosporites fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 237.2–237.8 m. All magnifications 1000x

Plate IX

Illyrian. 1. *Punctatisporites triassicus* Schulz 1964, Balatonfüred, Bf-1 borehole, 225.6–225.8 m; 2. *Reticulatisporites bunteri* Mädler *1964a*, Balatonfüred, Bf-1 borehole, 223.2–223.5 m; 3. *Cyclotriletes margaritatus* Mädler 1964a, Balatonfüred, Bf-1 borehole, 223.2–223.5 m; 4. *Cyclotriletes oligogranifer* Mädler 1964a, Balatonfüred, Bf-1 borehole, 252.6–252.8 m; 5. *Triadispora stabilis* Scheuring 1970, Balatonfüred, Bf-1 borehole, 232.0–232.2 m; 6. *Densosporites fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 269.8–270.1 m (Pelsonian). All magnifications 1000x

Plate X

Pelsonian–Illyrian. 1. *Concentricisporites nevesi* Antonescu 1970, Balatonfüred, Bf-1 borehole, 236.0 m (Illyrian); 2–3. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 267.9–268.1 m (Pelsonian); 4. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 244.7–244.9 m (Pelsonian); 5–6. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 256.0–256.2 m (Pelsonian); 7. *Uvaesporites fueredensis* nov. sp., Balatonfüred, Bf-1 borehole, 265.9 m; 8–9. *Uvaesporites fueredensis* nov. sp. (holotype), Balatonfüred, Bf-1 borehole, 261.0 m (Pelsonian). All magnifications 1000x

Plate XI

Illyrian. 1. *Verrucosisporites remyanus* Mädler 1964, Balatonfüred, Bf-1 borehole, 229.4–230.4 m; 2–3. *Sellaspora rugoverrucata* Van Der Eem 1983, Balatonfüred, Bf-1 borehole, 232.0–232.2 m; 4. *Aratrisporites fimbriatus* (Klaus 1960) Playford et Dettmann 1965, Balatonfüred, Bf-1 borehole, 223.2–223.5 m; 5. *Triadispora stabilis* Scheuring 1978, Balatonfüred, Bf-1 borehole, 211.9–212.1 m; 6. *Leptolepidites major* Couper 1958, Balatonfüred, Bf-1 borehole, 252.6–252.8 m. All magnifications 1000x

Plate XII

Illyrian. 1–3. *Sellaspora rugoverrucata* Van Der Eem 1983, Balatonfüred, Bf-1 borehole, 242.7–242.9 m; 4–5. *Densosporites fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 242.7–242.9 m; 6. *Ginkgocycadophytus levis* (Leschik 1956) nov. comb., Balatonfüred, Bf-1 borehole, 231.0–231.4 m. All magnifications 1000x

Plate XIII

Illyrian. 1. *Verrucosisporites slewecensis* (Mädler 1964) Orlowska-Zwolinska 1983, Balatonfüred, Bf-1 borehole, 253.4–253.8 m; 2. *Verrucosisporites slewecensis* (Mädler 1964a) Orlowska-Zwolinska 1983, Balatonfüred, Bf-1 borehole, 222.0–222.4 m; 3. *Sellaspora rugoverrucata* Van Der Eem 1983, Balatonfüred, Bf-1 borehole, 242.7–242.9 m; 4. *Leptolepidites* sp. nov. ind., Balatonfüred, Bf-1 borehole, 242.7–242.9 m; 5. *Convolutispora jogosa* Smith et Butterworth 1967, Balatonfüred, Bf-1 borehole, 254.2–254.4 m (Pelsonian). All magnifications 1000x

Plate XIV

Illyrian. 1. *Costatisulcites ovatus* Scheuring 1978, Balatonfüred, Bf-1 borehole, 222.0–222.4 m; 2. *Dyupetalum vicentinense* Brugman 1983, Bakonykuti, But-2 borehole, 39.0 m; 3. *Stellapollenites thiergartii* (Mädler 1954a) Clement-Westerhof et al. 1970, Bakonykuti, But-2 borehole, 39.0 m; 4–5. *Densosporites fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 248.8–249.0 m; 6. *Densosporites* cf. *fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 242.7–242.9 m; 7. *Uvaesporites fueredensis* nov. sp., Balatonfüred, Bf-1 borehole, 253.4–253.8 m; 8. *Uvaesporites gadensis* Praehauser-Enzenberg 1970, Balatonfüred, Bf-1 borehole, 258.0–258.2 m (Pelsonian); 9–10. *Densosporites* cf. *fissus* (Reinhardt 1964) Schulz 1967 , Balatonfüred, Bf-1 borehole, 242.7–242.9 m. All magnifications 1000x

Plate XV

Illyrian. 1. *Uvaesporites fueredensis* sp. nov., Balatonfüred, Bf-1 borehole, 232.0–232.2 m; 2. *Alisporites* cf. *ovatus* (Balme et Hennelly 1955) Jansonius 1962, Bakonykuti, But-2 borehole, 52.5 m; 3. *Triadispora* cf. *obscura* Scheuring 1970, Balatonfüred, Bf-1 borehole, 252.6–252.8 m; 4. *Cyclotriletes hians* Mädler 1964, Balatonfüred, Bf-1 borehole, 254.2–254.4 m; 5. *Triadispora suspecta* Scheuring 1970, Balatonfüred, Bf-1 borehole, 227.0–227.3 m; 6. *Striatoabieites* cf. *balmei* Klaus 1964, Bakonykuti, But-2 borehole, 53.0–54.0 m; 7. *Ginkgocycadophytus adjectus* De Jersey 1962, Balatonfüred, Bf-1 borehole, 252.6–252.8 m. All magnifications 1000x

Plate XVI

Illyrian. 1–2. *Meandrospira dinarica* Kochansky Dev. et Pantic, Szentantalfa-1, borehole, 105.5 m 64x; 3. *Endothyranella wirzi* (Koehn Zaninetti), Szentaltalfa-1, borehole, 113.0 m 83x; 4. *Ammobaculites radstadtensis* Kristan, Szentantalfa-1, borehole, 175.3 m 83x; 5. *Glomospira regularis* Lipina, Szentantalfa-1, borehole, 146.3 m 78x; 6. *Trochammina almtalensis* Koehn Zaninetti, Szentantalfa-1, borehole, 88.2 78x; 7. *Duostomina* cf. *magna* Trifonova, Szentantalfa-1, borehole, 175.3 m 83x; 8. *Tetrataxis* sp., Szentantalfa-1, borehole, 146.3 m 78x; 9–10. *Diplotremina persublima* (Kristan Tollmann), 9. Szentantalfa-1, borehole, 112.4 m 83x, 10. Szentantalfa-1, borehole, 146.3 m 78x; 11. *Diplotremina astrofimbriata* Kristan-Tollmann, Szentantalfa-1, borehole, 194.5 m 83x

Plate XVII

Illyrian. 1–4. *Eoophthalmidium tricki* (Langer), 1. Balatonfüred-1, borehole, 104.6–105.5 m 130x, 2. Bakonykuti-2, borehole, 92.2 m 160x, 3. Felsőörs Malomvölgy, bed No 99, 190 x, 4. Vörösberény Megyehegy 66 m 200x; 5–6. *Hemigordius chialingchiangensis* (He), Bakonyszűcs-3, borehole, 376.1–378.3 m 200x; 7–8. *Hemigordius amylovolutus* (He), Bakonyszűcs-3, borehole, 378.1–378.3 m 200x; 9. *Ophthalmidium uebeyliense* Dager, Balatonfüred-1, borehole, 265.1 m 130x; 10-11. *Glomospirella shengi* He, Bakonyszűcs-3, borehole, 378.1–378.3 m 10 = 200x, 11 = 130x; 12. *Glomospira densa* (Pantic), Aszófő-2, borehole, 221.0 m 120x; 13–15. *Glomospirella triphonensis* Baud et al., 13. Szentantalfa-1, borehole, 104.4 m 160x, 14, 15. Bakonykuti-2, borehole, 87.0 m 140x

Plate XVIII

Fassanian. 1. *Veryhachium irregulare* Jekhowsky 1961, Várpalota, Vpt-3 borehole, 72.7 m; 2, 4. *Baltisphaeridium* sp. "A", 2. Balatonfüred, Bf-1 borehole, 163.4–164.4 m, 4. Balatonfüred, Bf-1 borehole, 171.0–172.0 m; 3. *Baltisphaeridium* sp. "B", Balatonfüred, Bf-1 borehole, 166.2–166.3 m; 5. *Spheripollenites balmei* Jansonius 1962, Balatonfüred, Bf-1 borehole, 173.0–174.0 m; 6. *Spheripollenites elphinstonei* Jansonius 1962, Balatonfüred, Bf-1 borehole, 171.0–172.0 m; 7. *Convolutispora microrugulata* Schulz 1962, Balatonfüred, Bf-1 borehole, 168.0–169.0 m; 8. *Leiotriletes mesozoicus* (Thiergart 1949) Schulz 1967, Balatonfüred, Bf-1 borehole, 164.4–165.4 m; 9. *Convolutispora microrugulata* Schulz 1967, Várpalota, V-316, borehole, 309.5 m; 10. *Retitriletes gracilis* (Nilsson 1958) Döring et al. 1963, Balatonfüred, Bf-1 borehole, 169.0–170.0 m; 11. *Trilites tuberculiformis* Cookson 1947, Balatonfüred, Bf-1 borehole, 158.2–159.2 m; 12. *Veryhachium reductum* (Deunf 1958) Jekhowsky 1961, Balatonfüred, Bf-1 borehole, 211.3–211.4 m; All magnifications 1000 X

Plate XIX

Fassanian. 1. *Costatisulcatus ovatus* Scheuring 1978, Balatonfüred, Bf-1 borehole, 161.3–162.3 m; 2. *Verrucosisporites morulae* Klaus 1960, Balatonfüred, Bf-1 borehole, 163.4–164.s4 m; 3. *Cyclotriletes oligogranifer* Mädler 1964a, Balatonfüred, Bf-1 borehole, 166.3–167.3 m; 4–5. *Uvaesporites reissingeri* (Reinhardt 1961) Lund 1977, Balatonfüred, Bf-1 borehole, 175.0–176.0 m; 6. *Densoisporites fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 161.3–162.3 m; 7. *Uvaesporites argenteaeformis* (Bolkh. 1953) Schulz 1967, Balatonfüred, Bf-1 borehole, 164.2–165.4 m; 8. *Uvaesporites* sp., Balatonfüred, Bf-1 borehole, 163.4–164.4 m; 9. *Neoraistrickia* sp., Balatonfüred, Bf-1 borehole, 170.0–171.0 m; 10. *Uvaesporites* cf. *reissingeri* (Reinhardt 1961) Lund 1977, Balatonfüred, Bf-1 borehole, 170.0–171.0 m. All magnifications 1000x

Plate XX

Fassanian. Balatonfüred, Bf-1 borehole, 211.3–211.4 m. 1. *Trilites tuberculiformis* Cookson 1947; 2. *Trilites tuberculiformis* Cookson 1947; 3. *Dictyophyllidites surangei* Bharadwaj et Singh 1963; 4. *Guttatisporites ambigus* Tiwari et Vijaya 1980; 5. *Microreticulatisporites opacus* (Leschik 1956) Klaus 1960; 6. *Foveolatisporites* cf. *crassus* Orlowska-Zwolinska 1983. All magnifications 1000x

Plate XXI

Fassanian. 1. *Aratrisporites fimbriatus* (Klaus 1960) Playford et Dettmann 1965, Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 2. *Aratrisporites* sp., Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 3. *Illinites* sp., Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 4–5. *Stellapollenites thiergartii* (Mädler 1964a) Clement-Westerhof et al. 1974, Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 6. *Triadispora suspecta* Scheuring 1970, Balatonfüred, Bf-1 borehole, 211.3–211.4 m. All magnifications 1000x

Plate XXII

Fassanian. 1–2. *Perotriletes* cf. *granulatus* Couper 1953, Bakonykuti, But-2 borehole, 73.5 m; 3. *Paraconcavisporites* sp., Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 4–5. *Paraconcavisporites lunzensis* Klaus 1960, Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 6. *Densosporites fissus* (Reinhardt 1964) Schulz 1967, Balatonfüred, Bf-1 borehole, 211.3–211.4 m; 7–8. *Vitreisporites pallidus* (Reissinger 1950) Nilsson 1958, Balatonfüred, Bf-1 borehole, 211.3–211.4 m; All magnifications 1000x

Plate XXIII

Fassanian. 1. *Dyupetalum vicentinense* Brugman 1983, Bakonykuti, But-2 borehole, 84.0–84.4 m; 2. *Araucariacites* cf. *australis* Cookson 1947, Bakonykut, But-2 borehole, 73.5 m; All magnifications 1000x

Plate XXIV

Fassanian. 1–12. *Kuglerina meieri* Scheuring 1978, 1–3, 5, 8.Várpalota, V-316, borehole, 309.5 m, 4. Bakonykuti, But-2 borehole, 68.0–69.0 m, 6. Várpalota, Vpt-3 borehole, 44.3–45.0 m, 7. Bakonykuti, But-2 borehole, 61.0–63.0 m, 9. Bakonykuti, But-2 borehole, 74.0–75.0 m, 10. Balatonfüred, Bf-1 borehole, 211.3–211.4 m, 11. Balatonfüred, Bf-1 borehole, 211.3–211.4 m, 12. Várpalota, Vpt-3 borehole, 49.0 m. All magnifications 1000x

Plate XXV

Fassanian. 1. *Cannanoropollis scheuringii* Brugman 1986, Bakonykuti, But-2 borehole, 68.0–69.0 m; 2. *Cannanoropollis brugmanii* sp. nov., Várpalota, Vpt-3 borehole, 58.4 m. All magnifications 1000x

Plate XXVI

Fassanian. 1–4. *Cannanoropollis brugmanii* sp. nov. (holotypus), Várpalota. Vpt-3 borehole, 66.2 m, 1. proximal surface with uncovered central body and monolete mark, 2 distal surface with contact area and with the central free field, 3. section of proximal surface with ingrarugulate sculpture, 4. section of distal surface with the end of longitudinal elongated sexinous elements. All magnifications 1000x

Plate XXVII

Fassanian. 1–3. *Cannanoropollis brugmanii* sp. nov., 1, 2. Várpalota, Vpt-3 borehole, 66.2 m, 3. Balatonfüred, Bf-1 borehole, 211.3–211.40 m; 4: *Triadispora ventriosa* Scheuring 1978, Bakonykuti, But-2 borehole, 61.0-63.0 m. All magnifications 1000x

Plate XXVIII

Fassanian. 1–3. *Cannanoropollis scheuringii* Brugman 1986, 1. Bakonykuti, But-2 borehole, 91.0–92.0 m, 2. Bakonykuti, But-2 borehole, 60.0–61.0 m, 3. Bakonykuti, But-2 borehole, 94.0–94.8 m; 4. *Triadispora ventriosa* Scheuring 1978, Bakonykuti, But-2 borehole, 89.0–90.0 m; 5. *Triadispora* cf. *stabilis* Scheuring 1970, Bakonykuti, But-2 borehole, 89.0–90.0 m. All magnifications 1000x

Plate XXIX

Fassanian. 1–4. *Cannanoropollis scheuringii* Brugman 1986, Várpalota, Vpt-3 borehole, 66.2 m, 1.Proximal surface with end of longitudinal elongated sexinous elements, 2. Distal surface with network of end of the contact area, 3. Infrarugulate - punctate sculpture of central body of proximal surface, A part of the distal surface with the elongated longitudinal sexinous elements and 4. the network. All magnifications 1000x

Plate XXX

Fassanian. 1. *Cannanoropollis scheuringii* Brugman 1986, Bakonykuti, But-2 borehole, 55.0–56.0 m; 2–4. *Stellapollenites thiergartii* (Mädler 1964) Clement-Westerhof et al. 1974, 2. Bakonykuti, But-2 borehole, 67.0–68.0 m, 3. Bakonykuti, But-2 borehole, 54.7–55.0 m, 4: Bakonykuti, But-2 borehole, 56.0–57.0 m; 5. *Triadispora ventriosa* Scheuring 1978, Bakonykuti, But-2 borehole, 89.0–90.0 m. All magnifications 1000x

Plate XXXI

Fassanian. 1–2. *Triadispora sulcata* Scheuring 1978, Bakonykuti, But-2 borehole, 59.0–60.0 m; 3–4. *Lunatisporites acutus* var. *nudus* Scheuring 1970, Bakonykuti, But-2 borehole, 67.0–68.0 m; 5–6. *Stellapollenites thiergartii* (Mädler 1964) Clement-Westerhof et al. 1974, 5. Bakonykuti, But-2 borehole, 84.4–85.4 m, 6. Bakonykuti, But-2 borehole, 89.0–90.0 m; 7. *Striatoabieites balmei* Klaus 1964, Bakonykuti, But-2 borehole, 92.0–93.0 m. All magnifications 1000x

Plate XXXII

Fassanian. 1–3. *Strotersporites tozeri* Brugman 1986, Bakonykuti, But-2 borehole, 70.0–71.0 m; 4,–5. *Vitreisporites pallidus* (Reissinger 1950) Nilsson 1958, Várpalota, Vp-316 borehole, 309.5 m. All magnifications 1000x

Plate XXXIII

Fassanian. 1. *Cuneatisporites radialis* Leschik 1956, Várpalota, Vp-316 borehole, 309.5 m; 2. *Angustisulcites klausii* Freudenthal 1964, Várpalota, Vp-316 borehole, 309.5 m; 3. *Triadispora stabilis* Scheuring 1970, Várpalota, Vp-316 borehole, 309.0 m; 4. *Triadispora boelchii* (Scheuring 1970) 1978, Várpalota, Vp-316 borehole, 309.5 m. All magnifications 1000x

Plate XXXIV

Fassanian. 1–2. *Lunatisporites noviaulensis mollis* Scheuring 1978, Várpalota, Vp-316 borehole, 309.5 m; 3. *Lunatisporites noviaulensis mollis* Scheuring 1978, Balatonfüred, Bf-1 borehole, 169.0–170.0 m; 4. *Triadispora ventriosa* Scheuring 1978, Várpalota, Vp-316 borehole, 309.5 m; 5, 7. *Triadispora barbata* Scheuring 1978, 5. Balatonfüred, Bf-1 borehole, 158.9–159.2 m, 7. Balatonfüred, Bf-1 borehole, 166.3–167.3 m; 6. *Triadispora plicata* Klaus 1964, Várpalota, Vp-316 borehole, 309.0 m; 8–9. *Triadispora* sp. Forma A, Várpalota, Vpt-3 borehole, 32.0 m. All magnifications 1000x

Plate XXXV

Fassanian. 1–2. *Lunatisporites noviaulensis* mollis Scheuring 1978, Várpalota, Vp-316 borehole, 309.5 m; 3. *Triadispora* cf. *obscura* Scheuring 1970, Balatonfüred, Bf-1 166.3–167.3 m; 4. *Lunatisporites acutus* (Leschik 1956) Scheuring 1970, Balatonfüred, Bf-1 borehole, 211.1–211.2 m; 5–7. *Triadispora bella* Scheuring 1978, 5. Balatonfüred, Bf-1 borehole, 209.7–209.8 m, 6. Balatonfüred, Bf-1 borehole, 162.2–163.4 m, 7. Balatonfüred, Bf-1 borehole, 163.4–164.4 m. All magnifications 1000x

Plate XXXVI

Fassanian. 1–2. *Triadispora bella* Scheuring 1978, Várpalota, Vp-316 borehole, 309.5 m; 3. *Triadispora sulcata* Scheuring 1978, Balatonfüred, Bf-1 borehole, 163.4–164.4 m; 4. *Triadispora falcata* Klaus 1964 Várpalota, Vpt-3 borehole, 43.3–44.3 m; 5. *Triadispora suspecta* Scheuring 1970, Balatonfüred, Bf-1 borehole, 164.4–165.4 m; 6. *Triadispora suspecta* Scheuring 1970, Balatonfüred, Bf-1 borehole, 171.0–172.0 m; 7. "*Triadispora* cf. *obscura* Scheuring 1970" in Antonescu 1976, Pl. IV., Fig. 11.,Balatonfüred, Bf-1 borehole, 173.0–174.0 m. All magnifications 1000x

Plate XXXVII

Fassanian. 1. *Triadispora* cf. *suspecta* Scheuring 1970, Bakonykúti, But-2 borehole, 60.0–61.0 m; 2. *Striatoabieites balmei* Klaus 1964, Bakonykúti, But-2 borehole, 76.0–77.0 m; 3. *Striatoabieites balmei* Klaus 1964, Bakonykúti, But-2 borehole, 91.0–92.0 m; 4. *Triadispora obscura* Scheuring 1970, Bakonykúti, But-2 borehole, 101.6 m. All magnifications 1000x

Plate XXXVIII

Fassanian. 1–3. *Pseudonodosaria loczyi* Oravecz Sch., 1. Bakonyszűcs-3, borehole, 378,1–378.3 m 200x, 2, 3. Felsőörs Malomvölgy bed No 100, 130x; 4–8. *Hemigordius plectospirus* (Oravecz Sch.), 4–5. Felsőörs Malomvölgy bed No. 98, 240x, 6. Balatonfüred-1, borehole, 112.2–112.8 m, 260x, 7. Vörösberény Megyehegy, borehole, 28.6 m, 260 m, 8. Felsőörs-2, borehole, 28.6 m, 300x; 9 *Pseudonodosaria ploechingeri* (Oberh.), Bakonyszűcs-3, borehole, 378.1–378.3 m, 130x; 10. *Pachyphloides klebelsbergi* (Oberh.), Bakonyszűcs-3, borehole, 358.7 m, 200x; 11. *Nodosaria raibliana* Gümbel, Bakonyszűcs-3, borehole, 370.8 m, 130x; 12. *Oberhauserella ladinica* (Oberh.), Bakonyszűcs-3, borehole, 372.0 m, 130x; 13. *Oberhauserella mesotriassica* (Oberh.), Bakonyszűcs-3, borehole, 378.1–378.3 m, 130x

Plate XXXIX

Fassanian. 1. *Earlandinita* cf. soussi Salaj, Bakonykúti-2, borehole, 51.7 m 65x; 2. *Earlandinita oberhauseri* Salaj, Várpalota-3, borehole, 12.4 m, 80x; 3. *Reophax* cf. *asper* Cushman et Waters, Bakonykúti-2. borehole, 51,7 m, 80x; 4–5. *"Pilaminella" gemerica* Salaj, 4. Balatonfüred-1, borehole, 120.6 m, 120x, 5. Bakonykúti-2, borehole, 92.2 m, 240x; 6–8. *Turriglomina mesotriassica* (Koehn Zaninetti), 6. Balatonfüred-1, borehole, 111.4–112.8 m, 360x, 7. Várpalota-3, borehole, 10.4 m, 160x, 8. Vörösberény Megyehegy 22 m, 190x; 9. *Paleolituonella* aff. *reclinata* He, Bakonykúti-2, borehole, 70.0 m, 65x

Plate XL

Fassanian. 1. *Oberhauserella multiloculata* (Fuchs), Bakonykúti-2, borehole, 72.4 m, 160x; 2. *Oberhauserella mesotriassica* (Oberhauser), Bakonykúti-2, borehole, 72.4 m, 180x; 3–4. *Oberhauserella* sp., Bakonykúti-2, borehole, 72.4 m, 180x. 1–4 SEM

Plate XLI

Fassanian. 1–4. *Diplotremina altoconica* Kristan-Tollmann, Bakonykúti-2, borehole, 72.4 m, 1, 4. lateral view, 150x, 2. apical view, 160x, 3. bottom view with area of aperture, 100x. 1–3 SEM

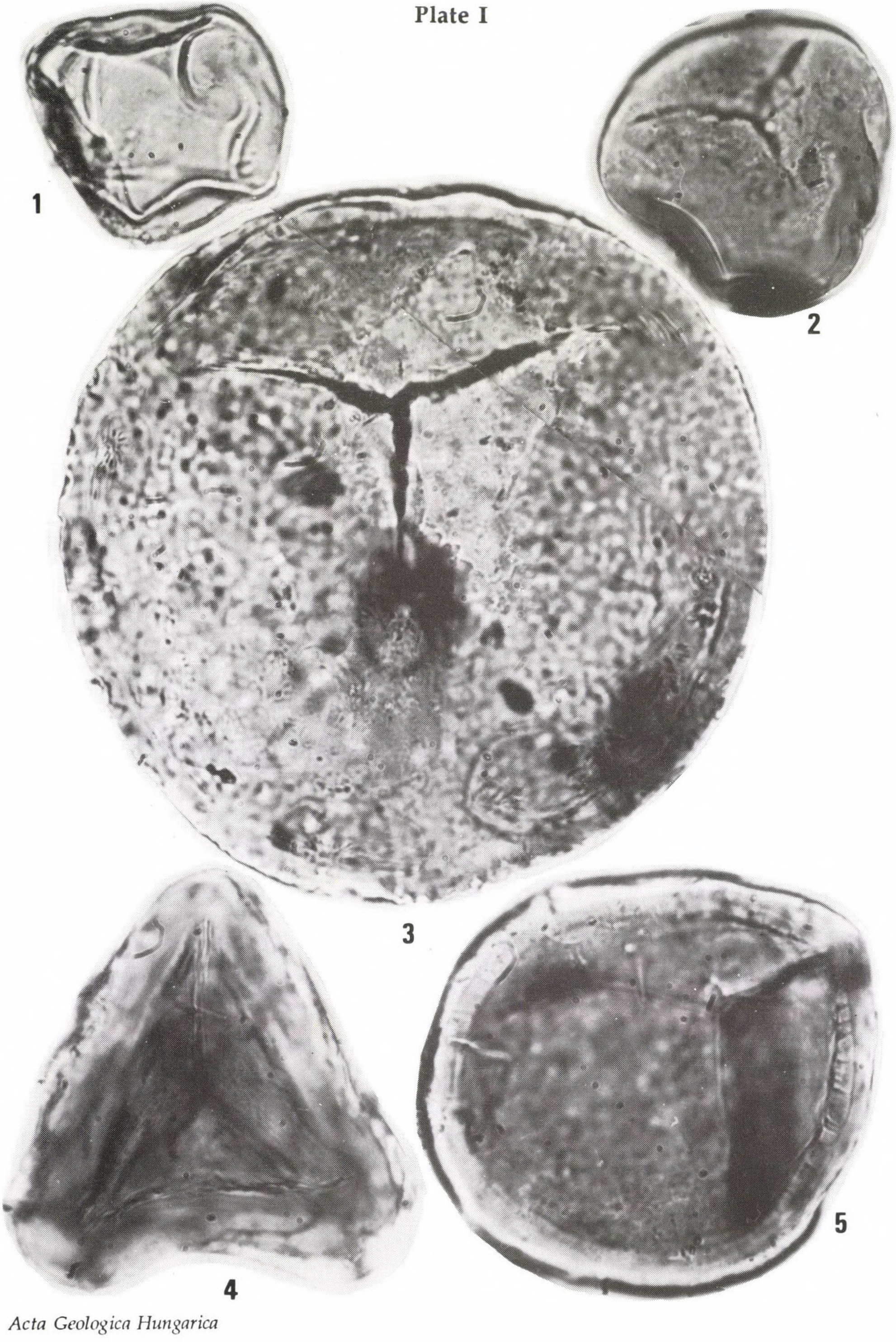
Plate XLII

Fassanian. 1–4. *Diplotremina stenocamerata* He, Bakonykuti-2, borehole, 72.4 m, 1–3. lateral view with the area of aperture, 4. bottom view with branching aperture. 1–4 SEM

Plate XLIII

Fassanian. 1–4. *Lamelliconus cordevolicus* (Oberhauser), Bakonykúti-2, borehole, 36,5 m, 1–3. lateral view, 160x, 2. lateral view with simple, small rounded aperture, 160x, 3. specimen corroded with acetic acid, 180x, 4. bottom view 200x. 1–4 SEM

Plate I



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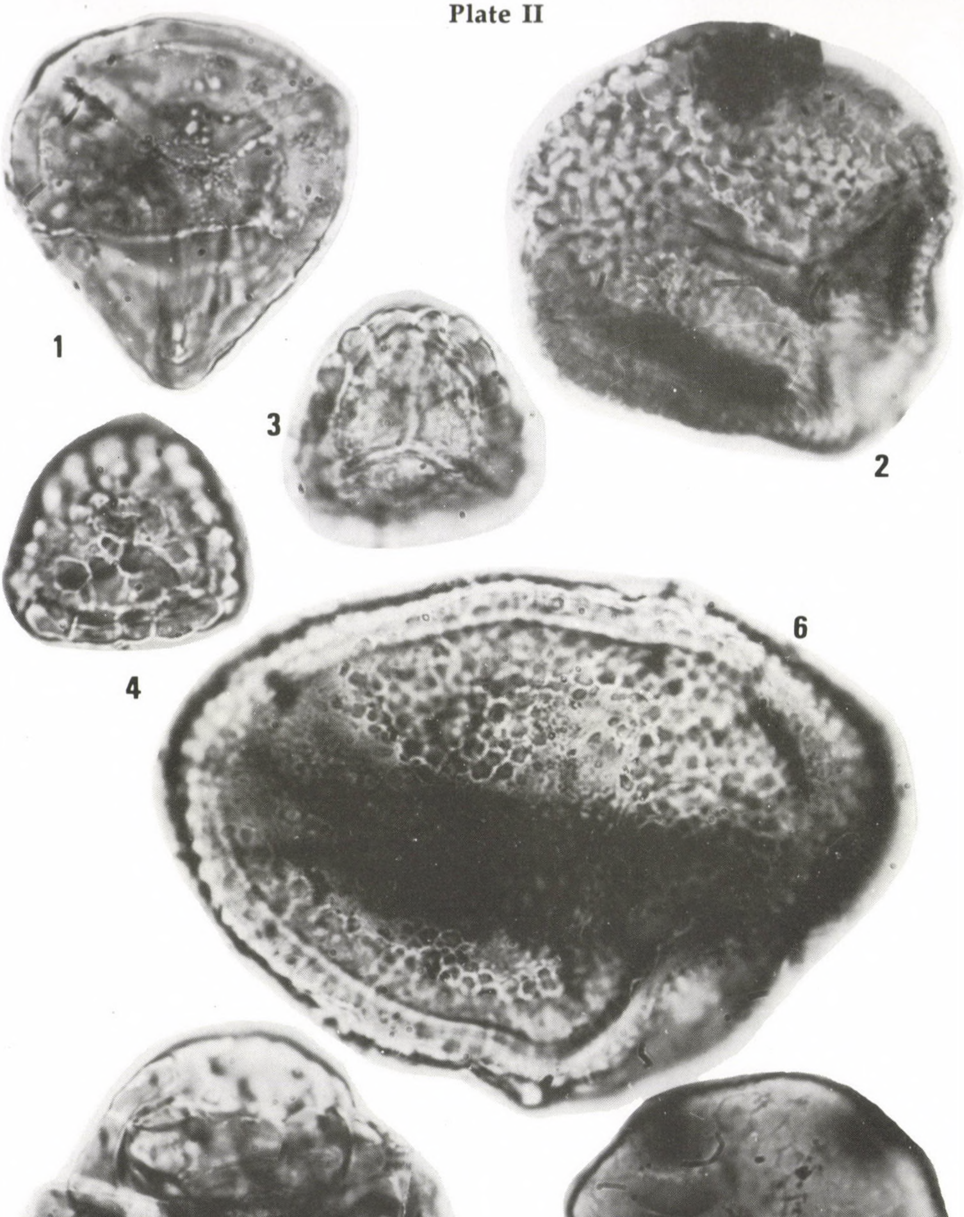

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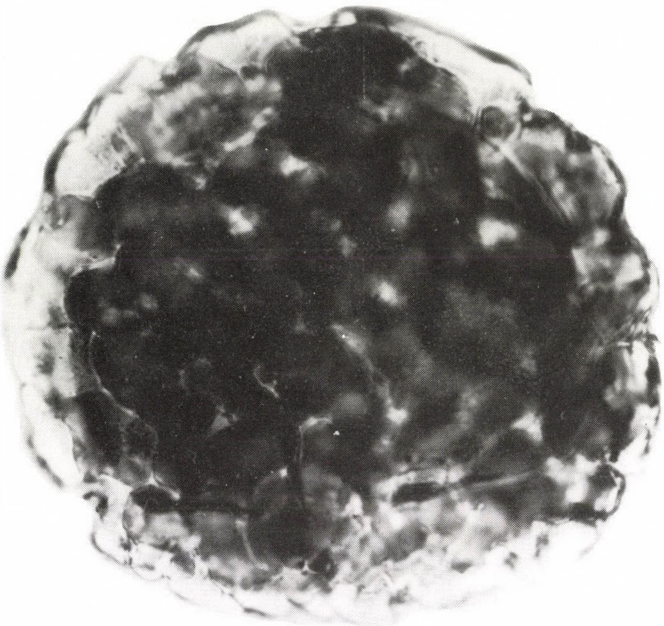

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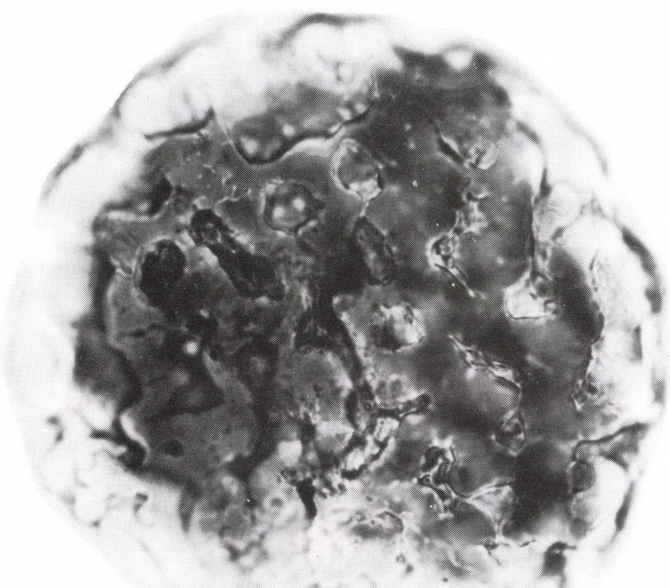

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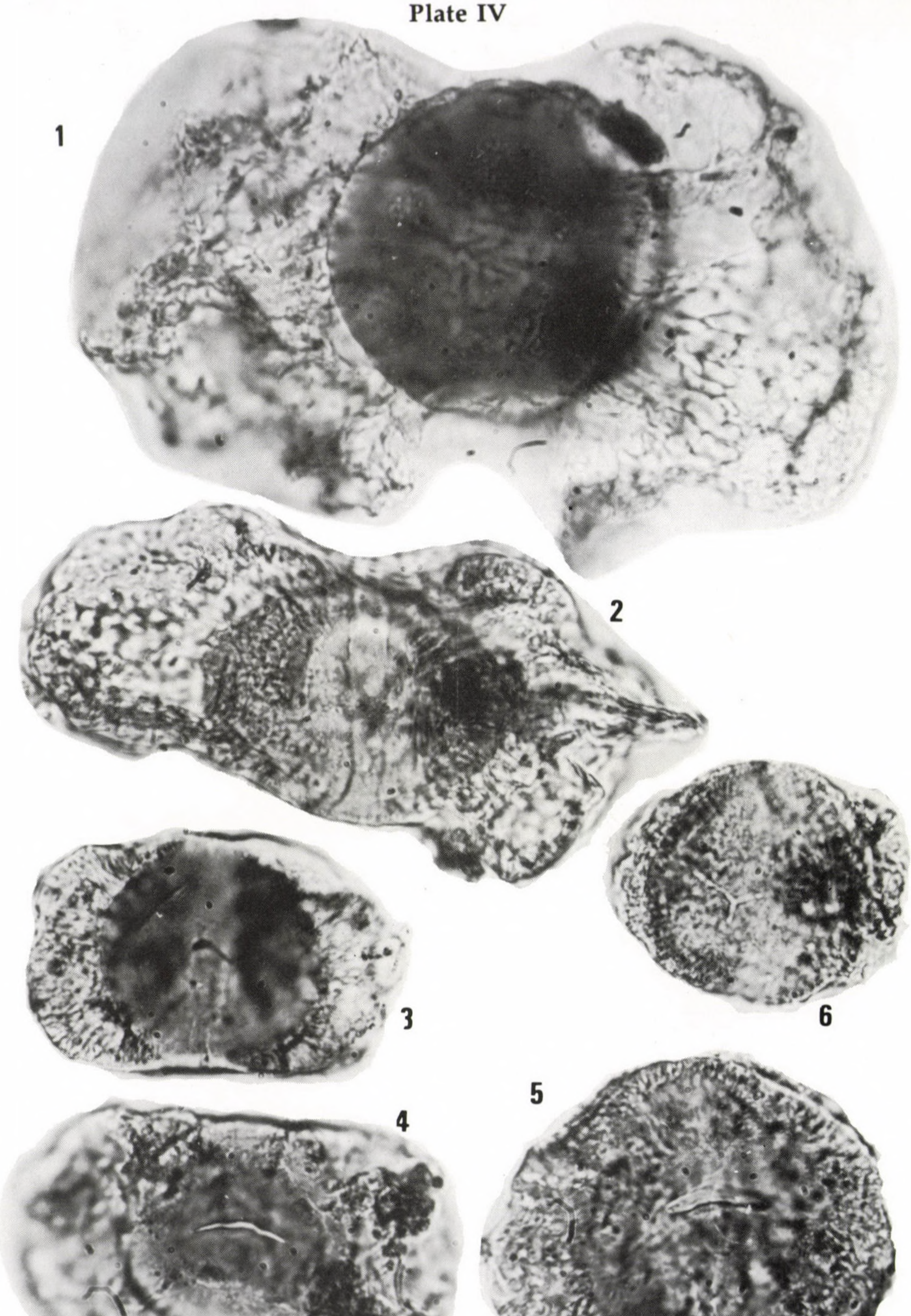

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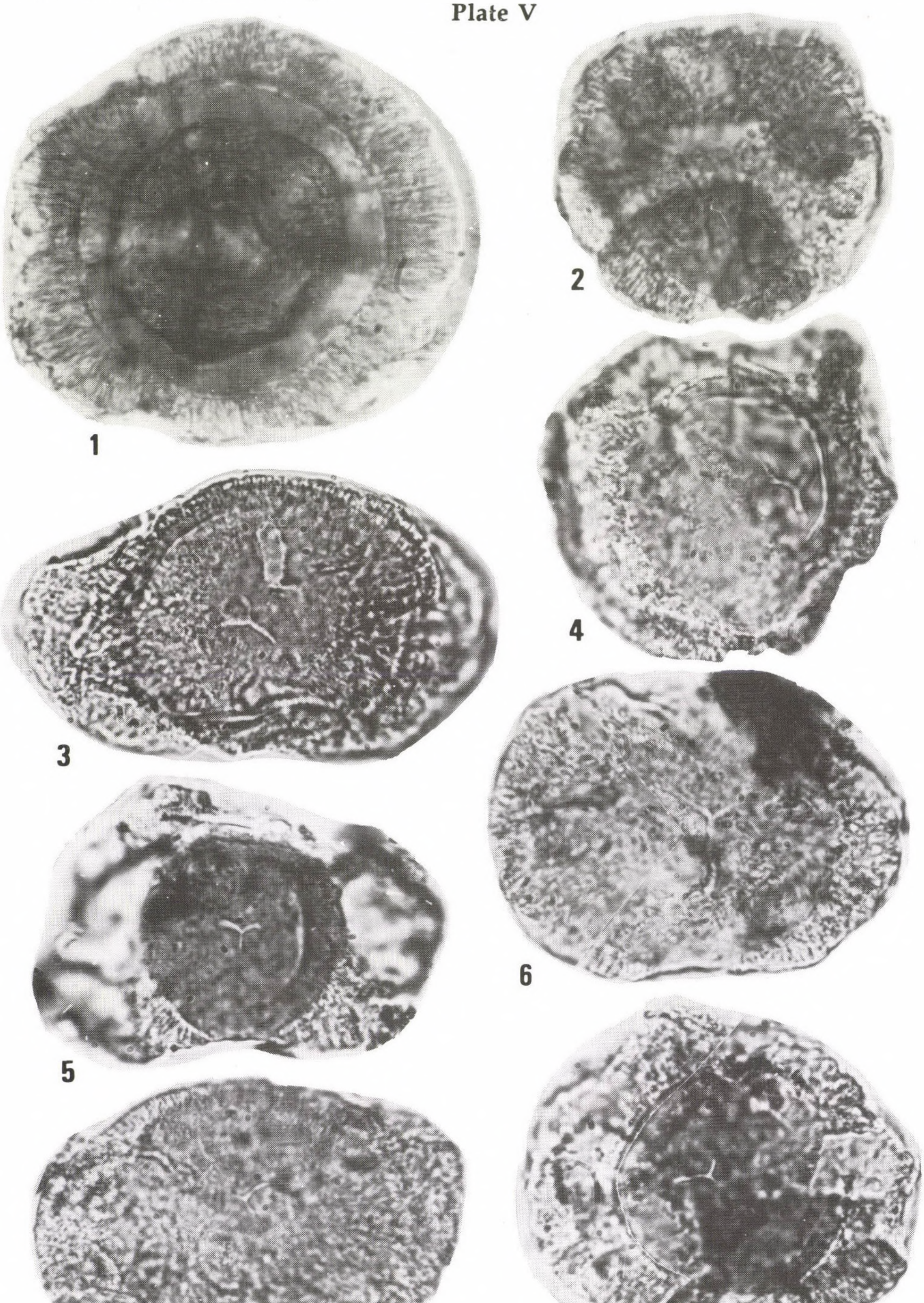

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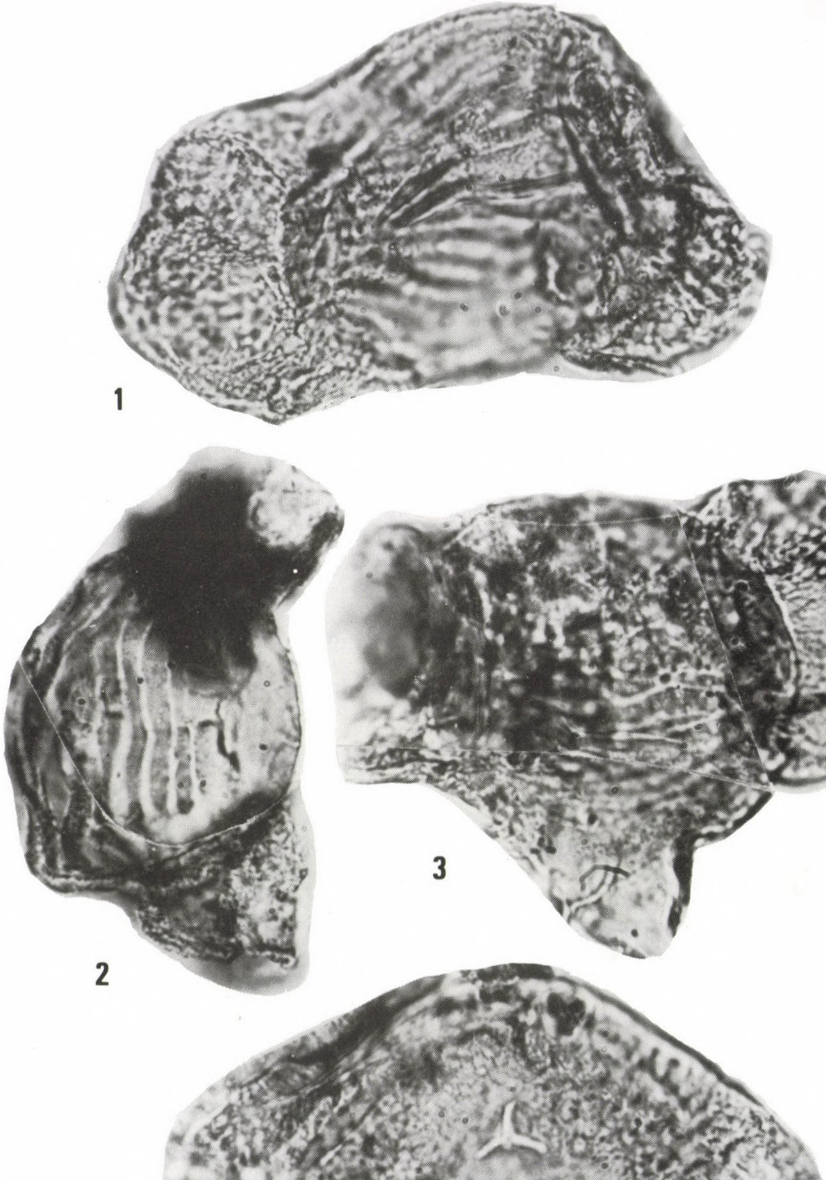

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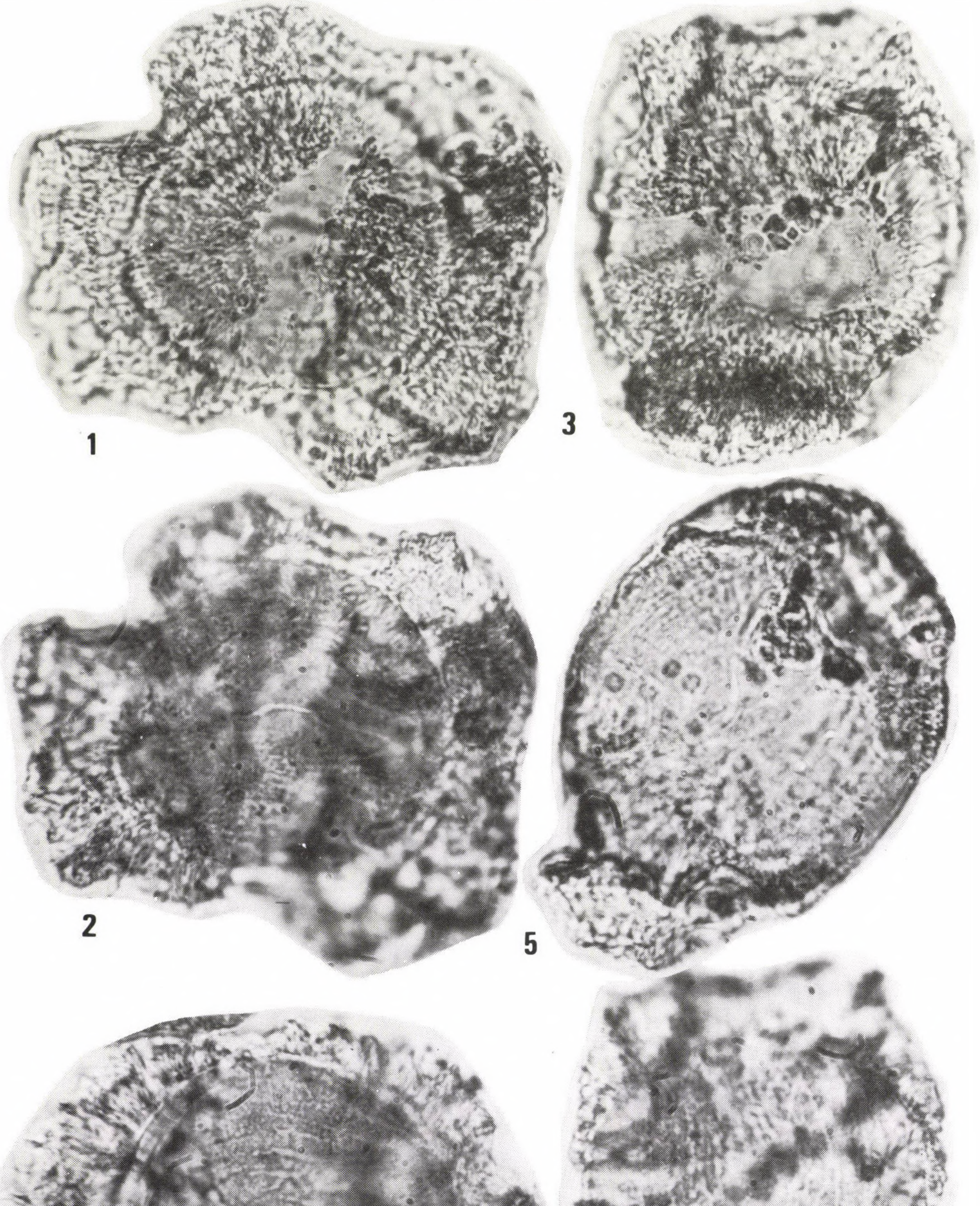

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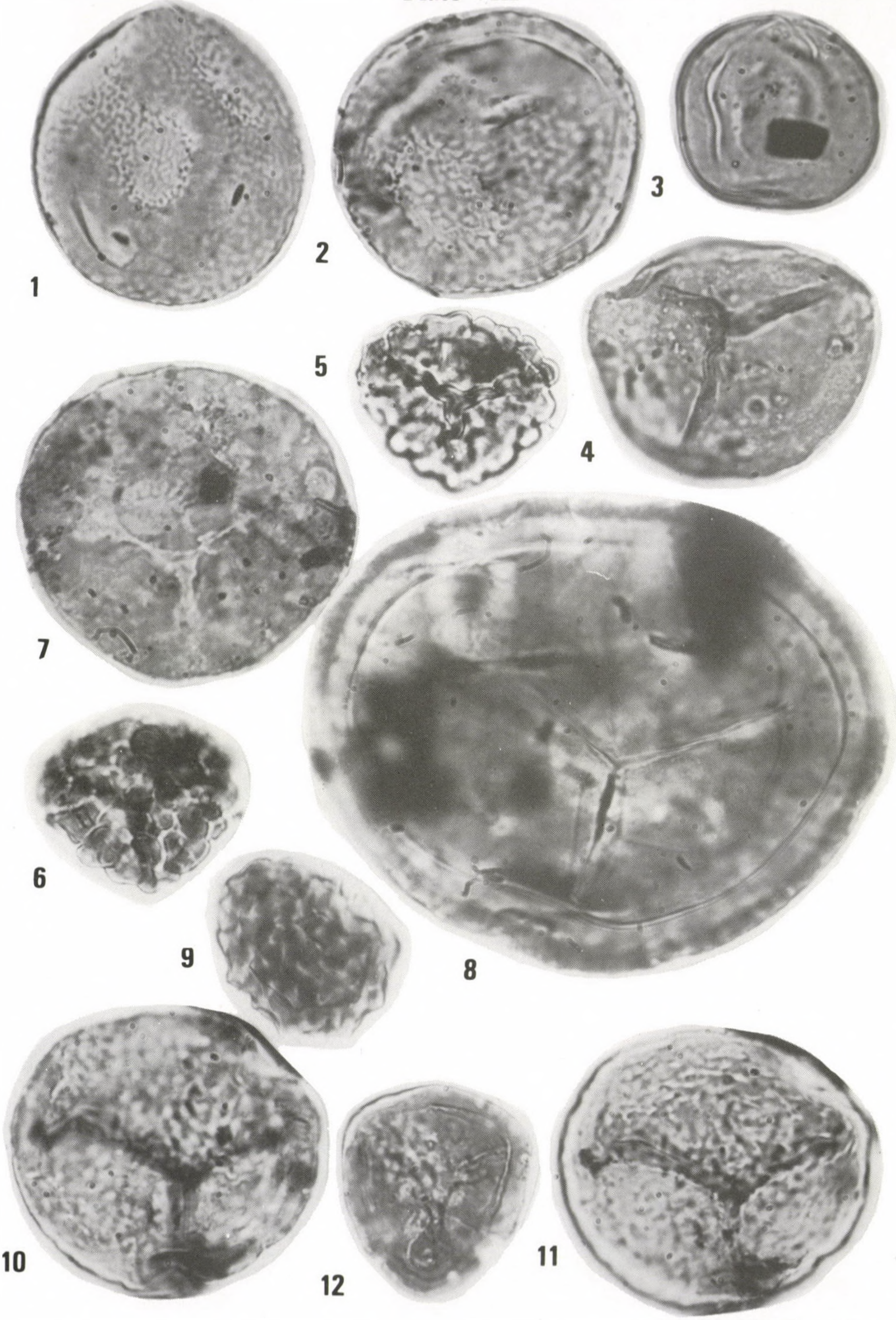

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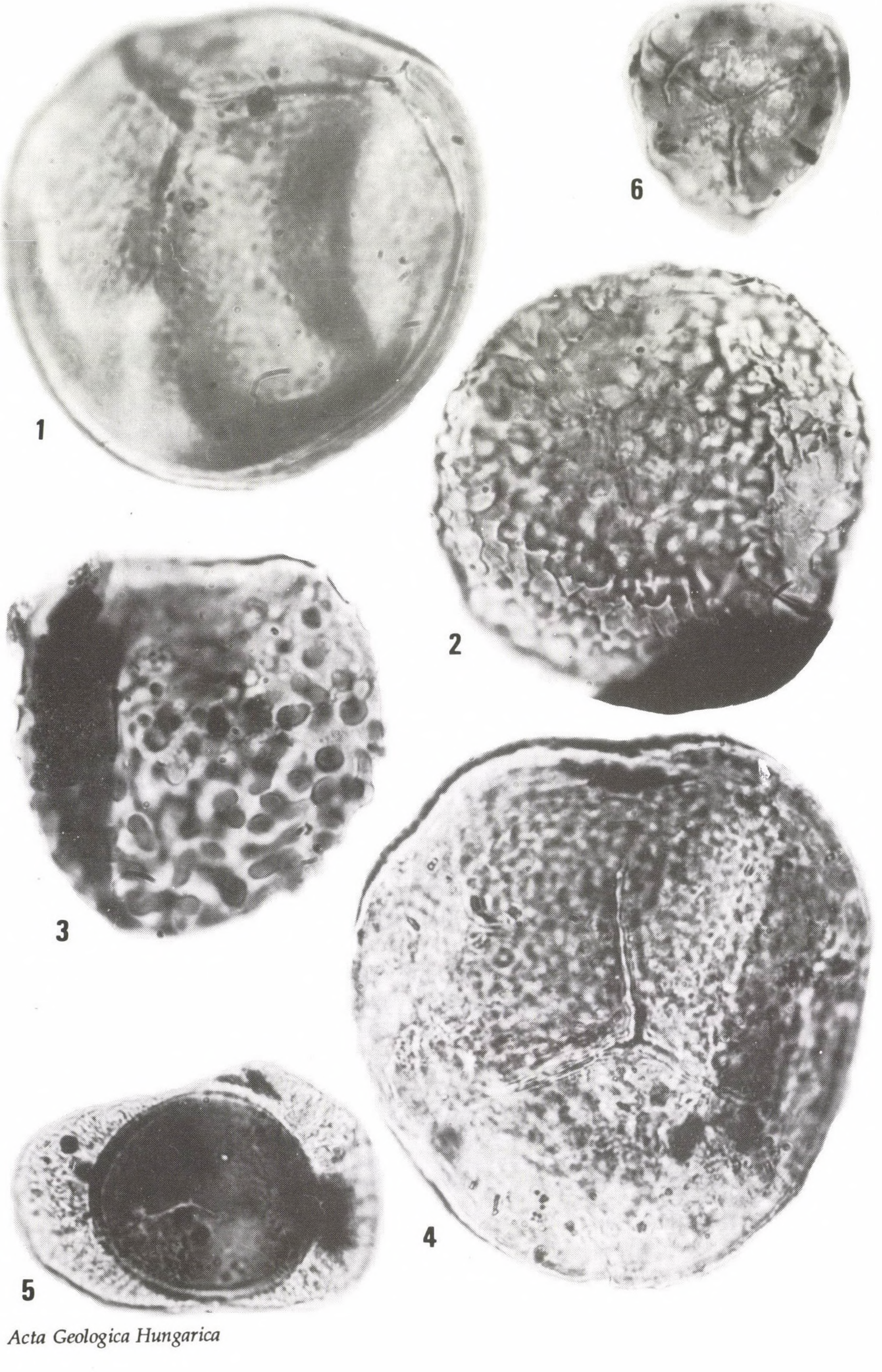

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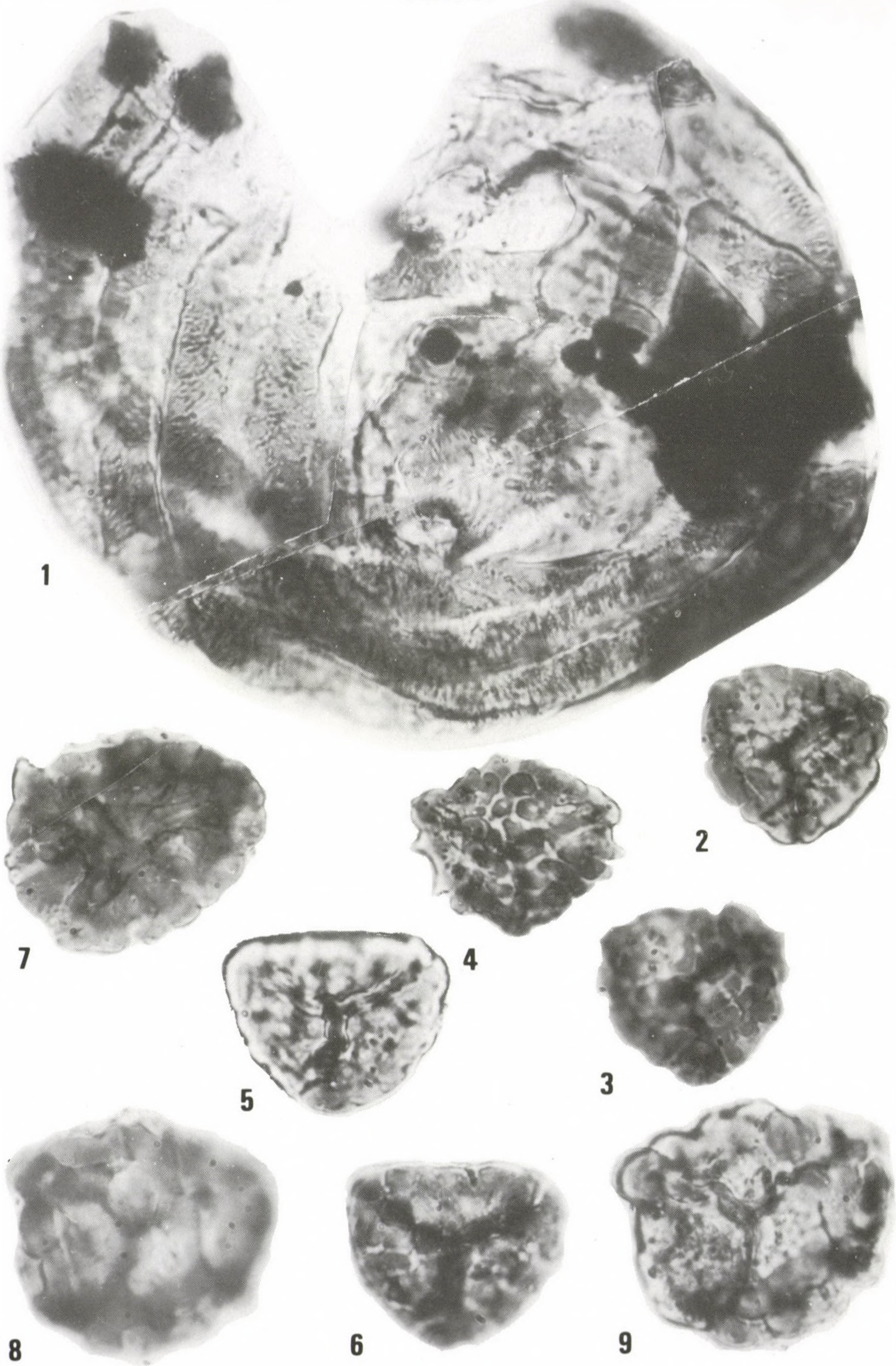

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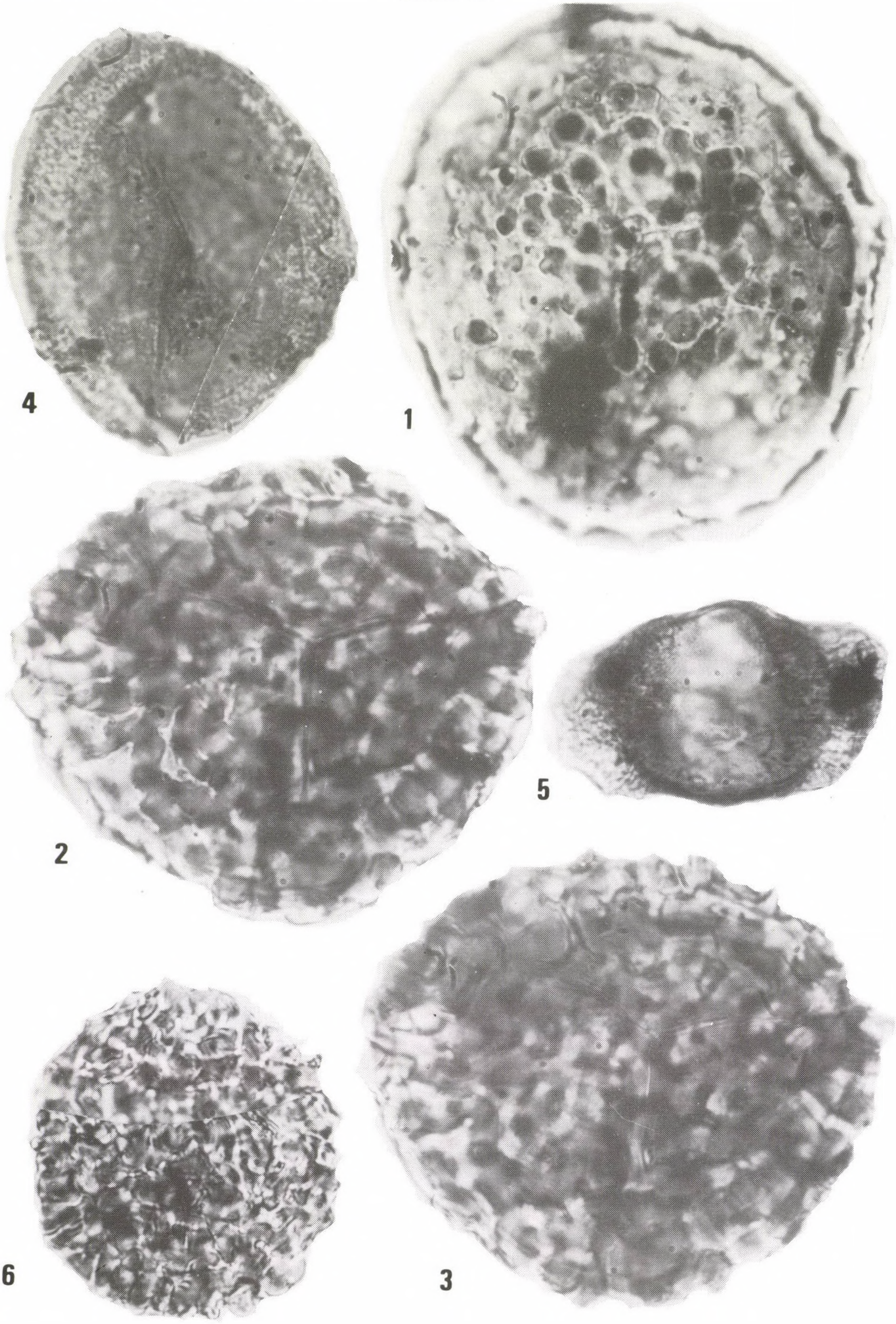

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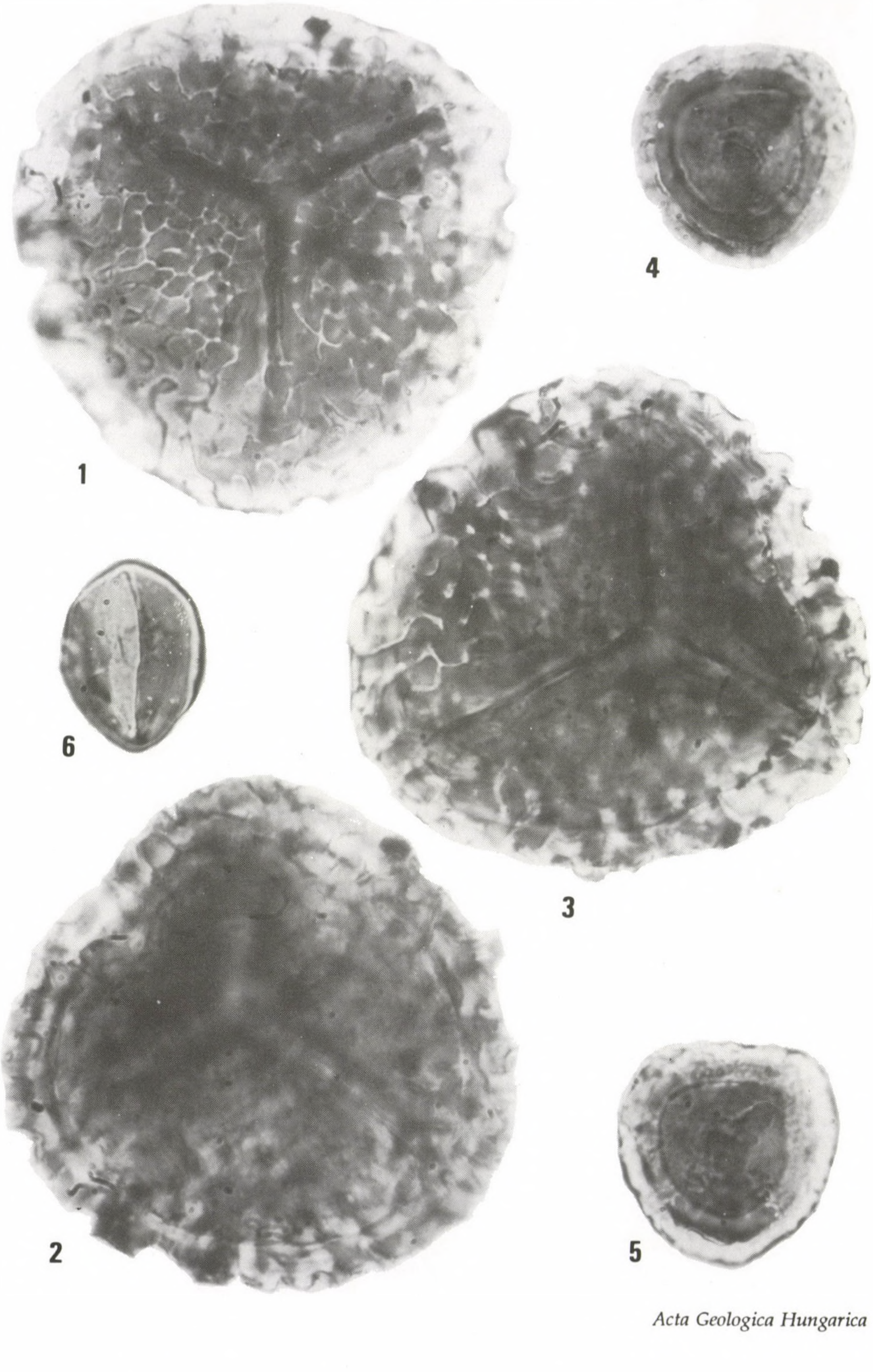

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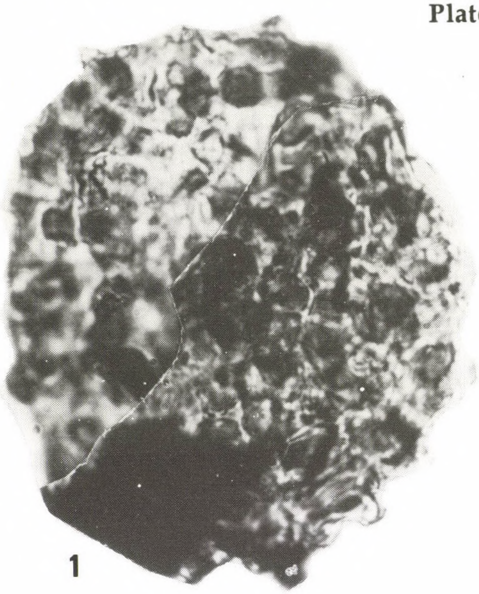

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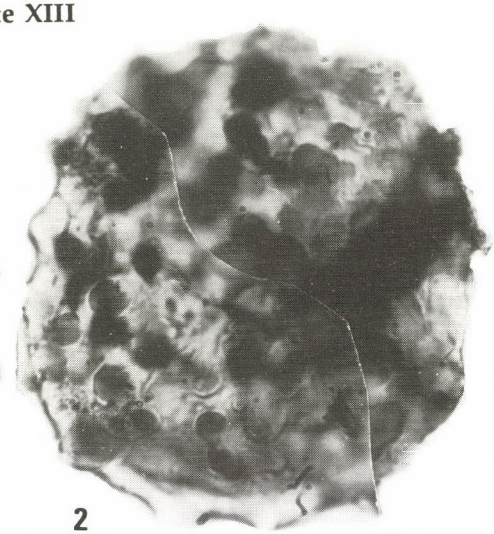

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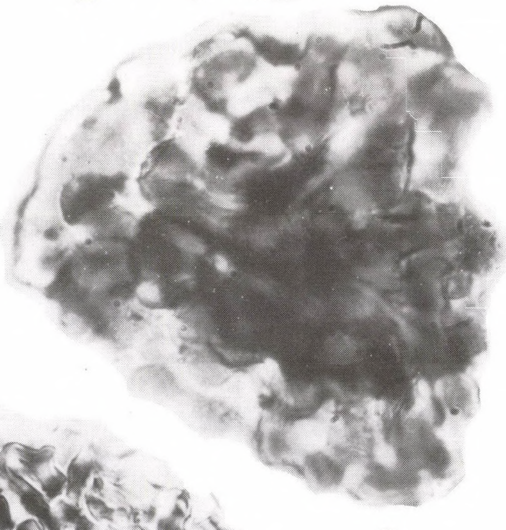

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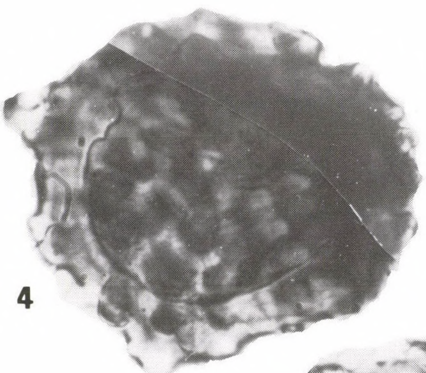

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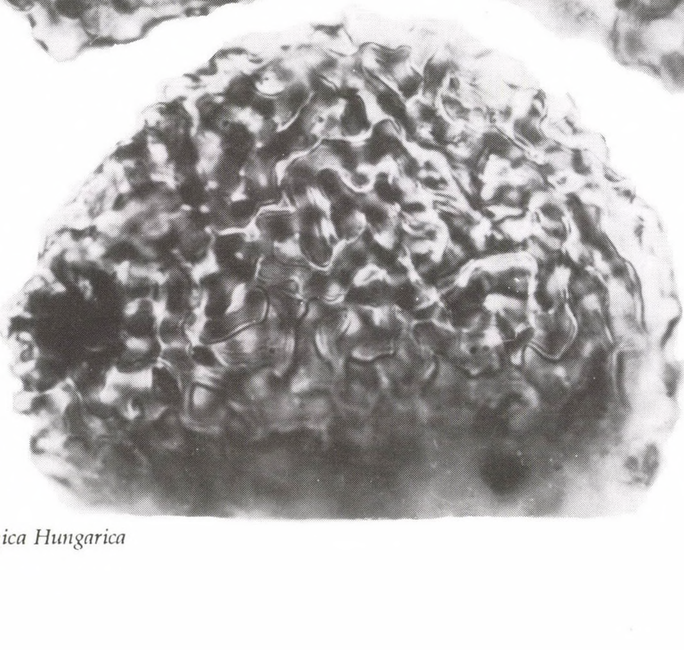

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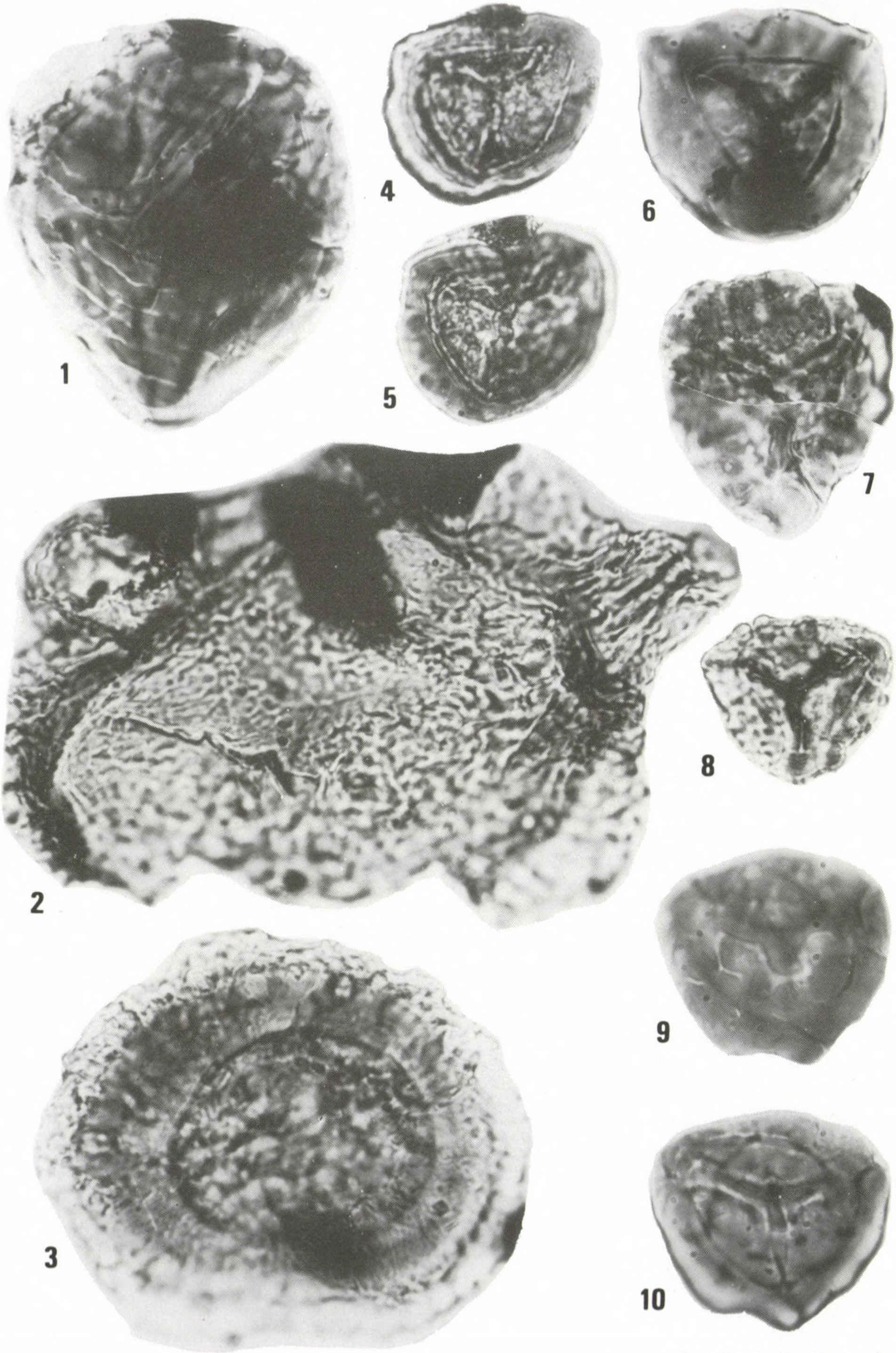

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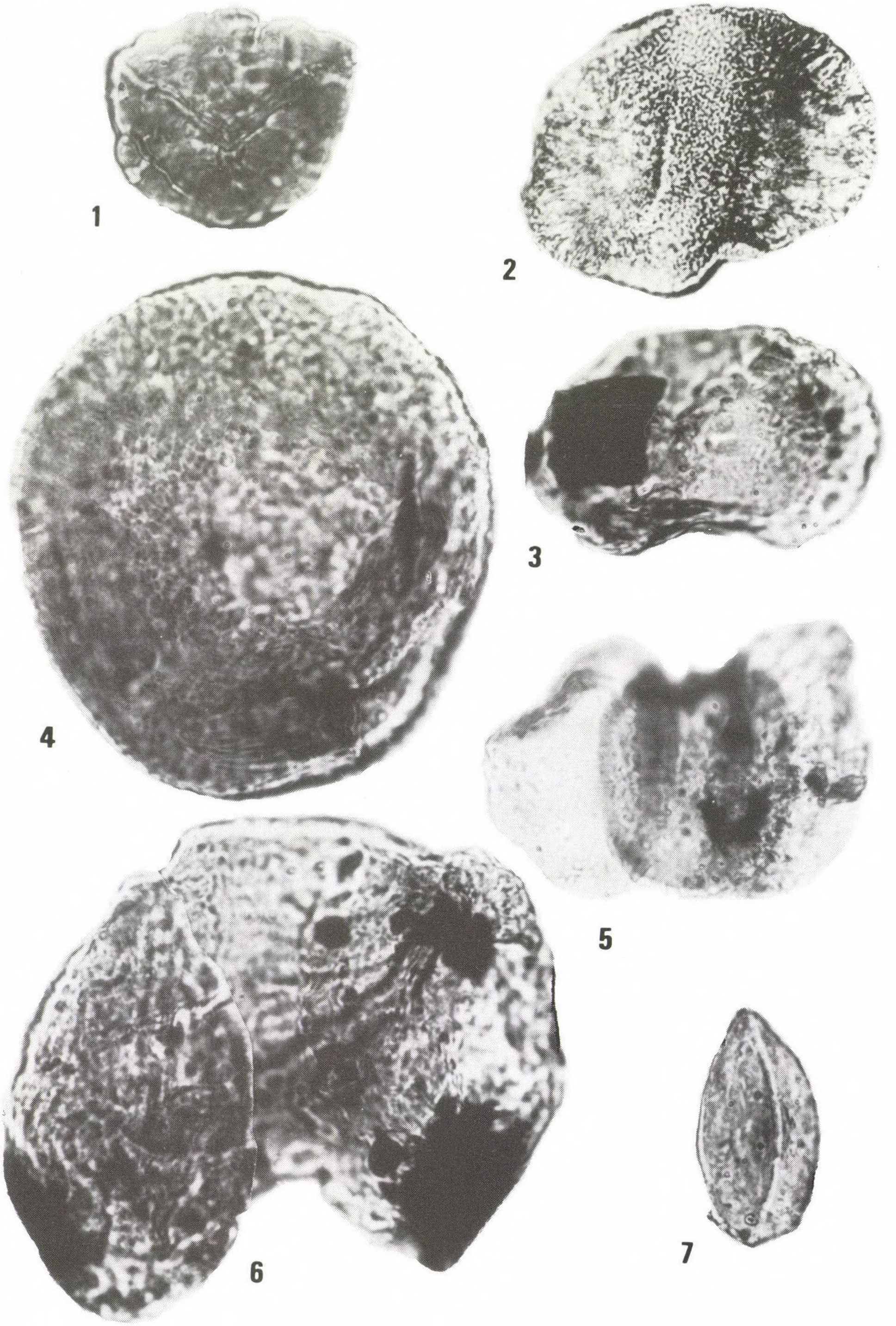

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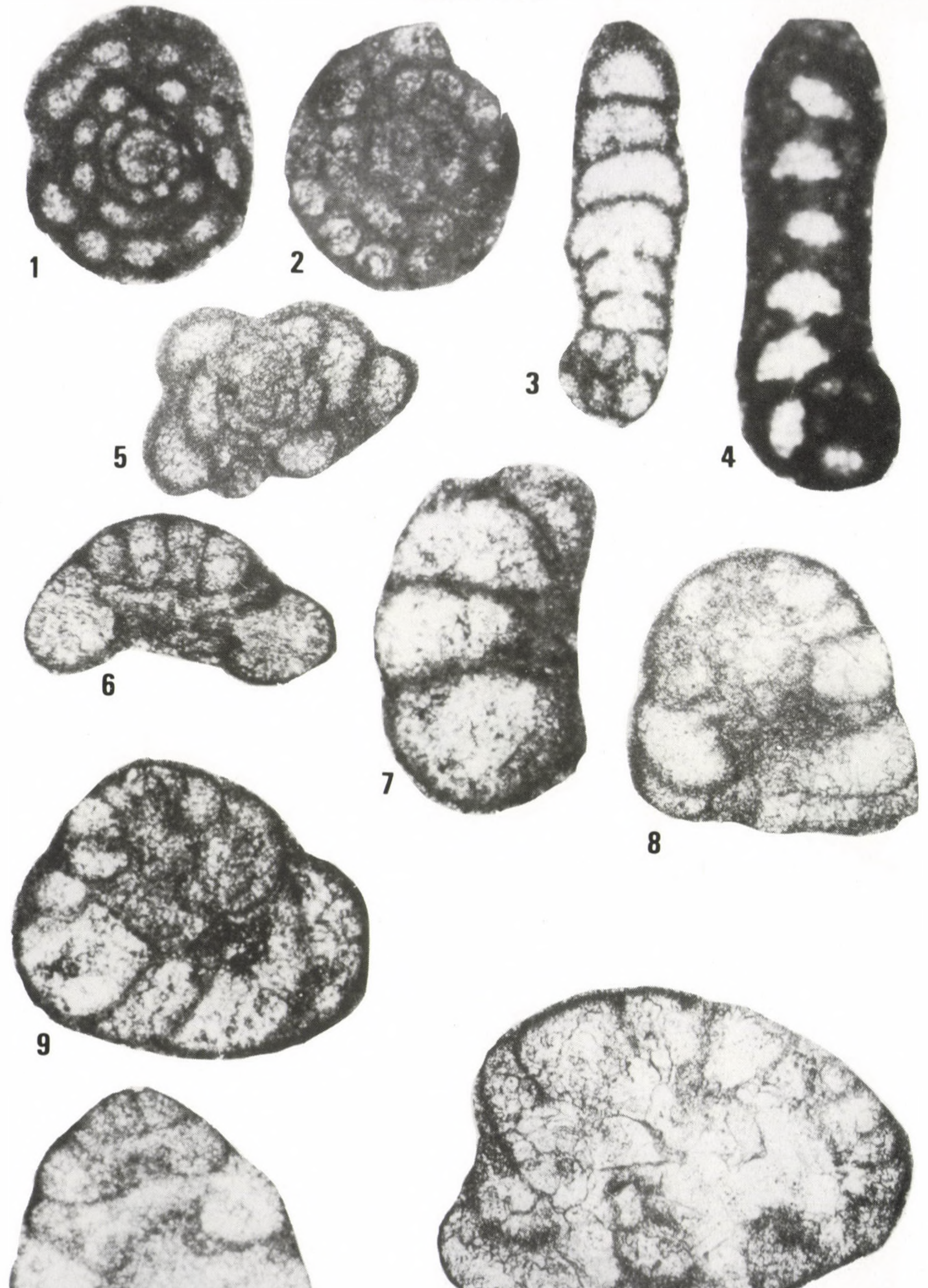

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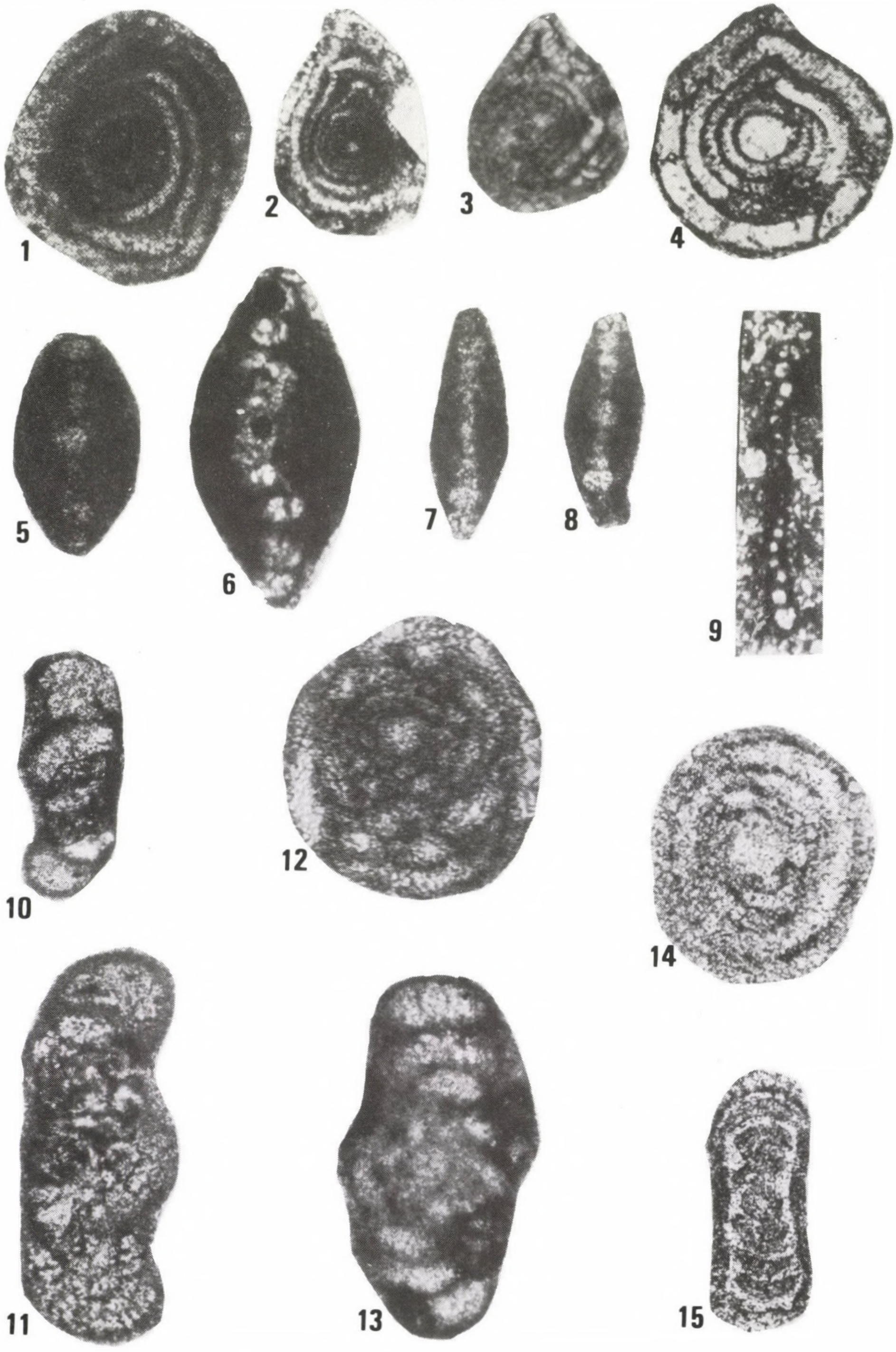

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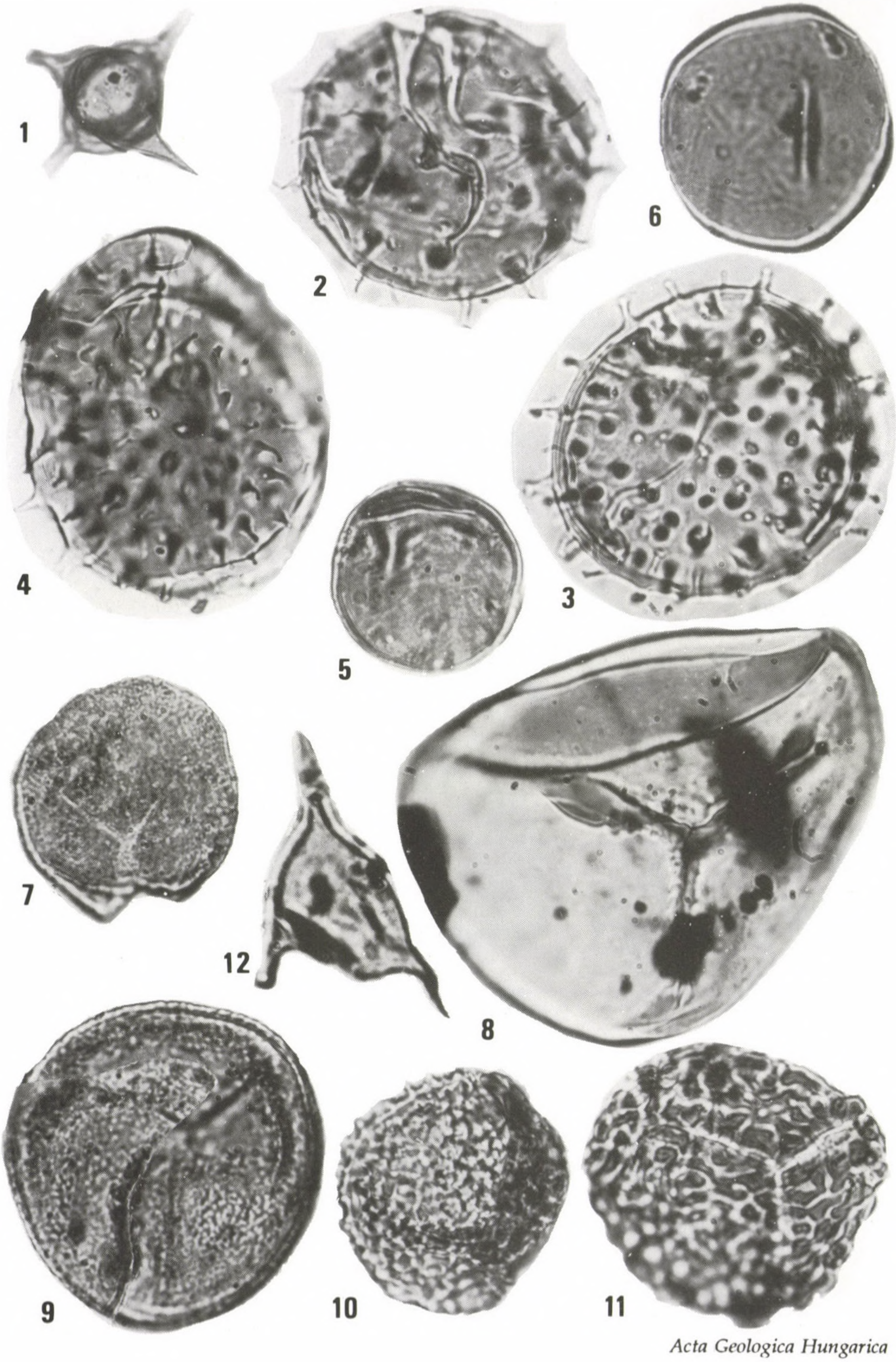

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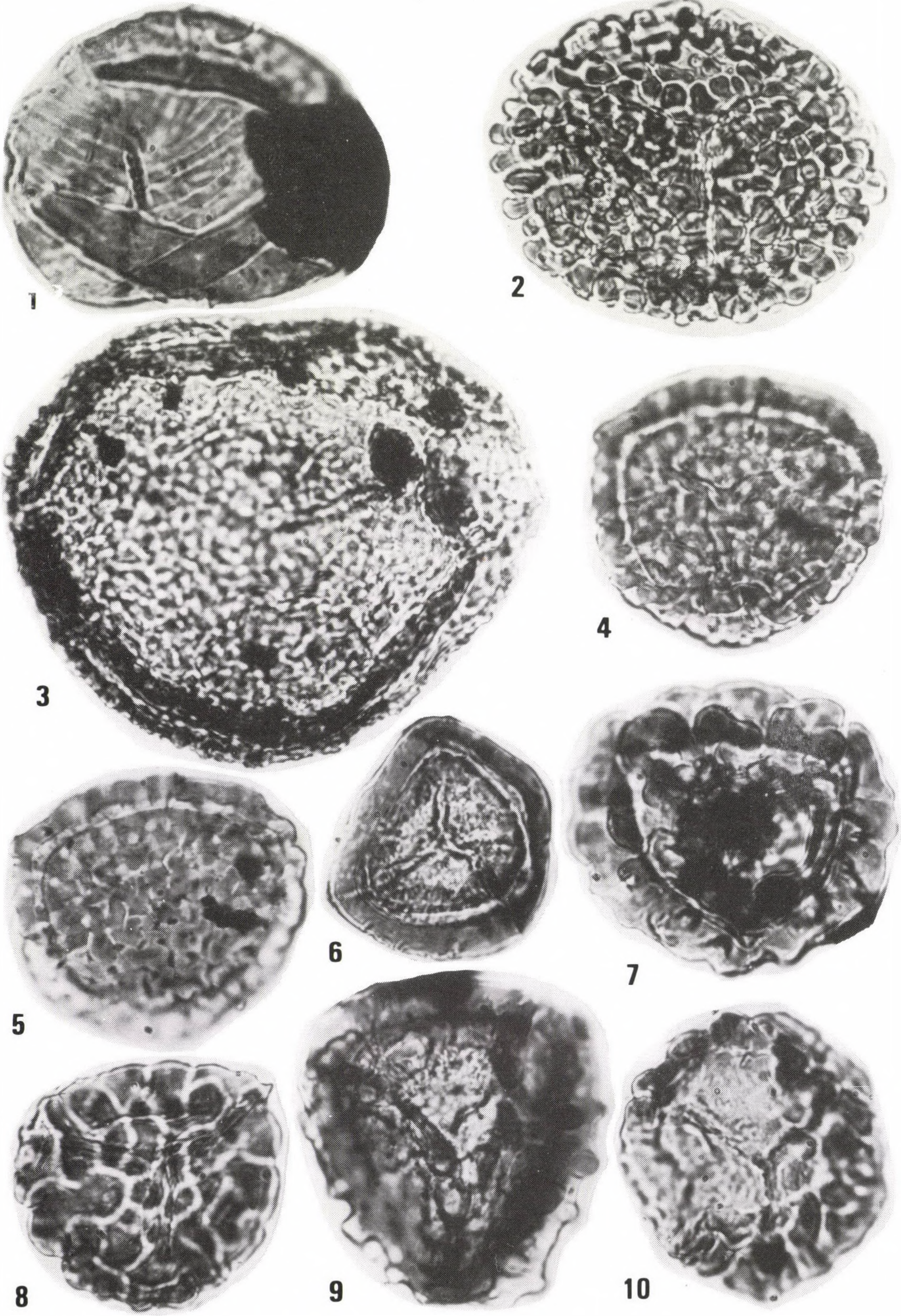

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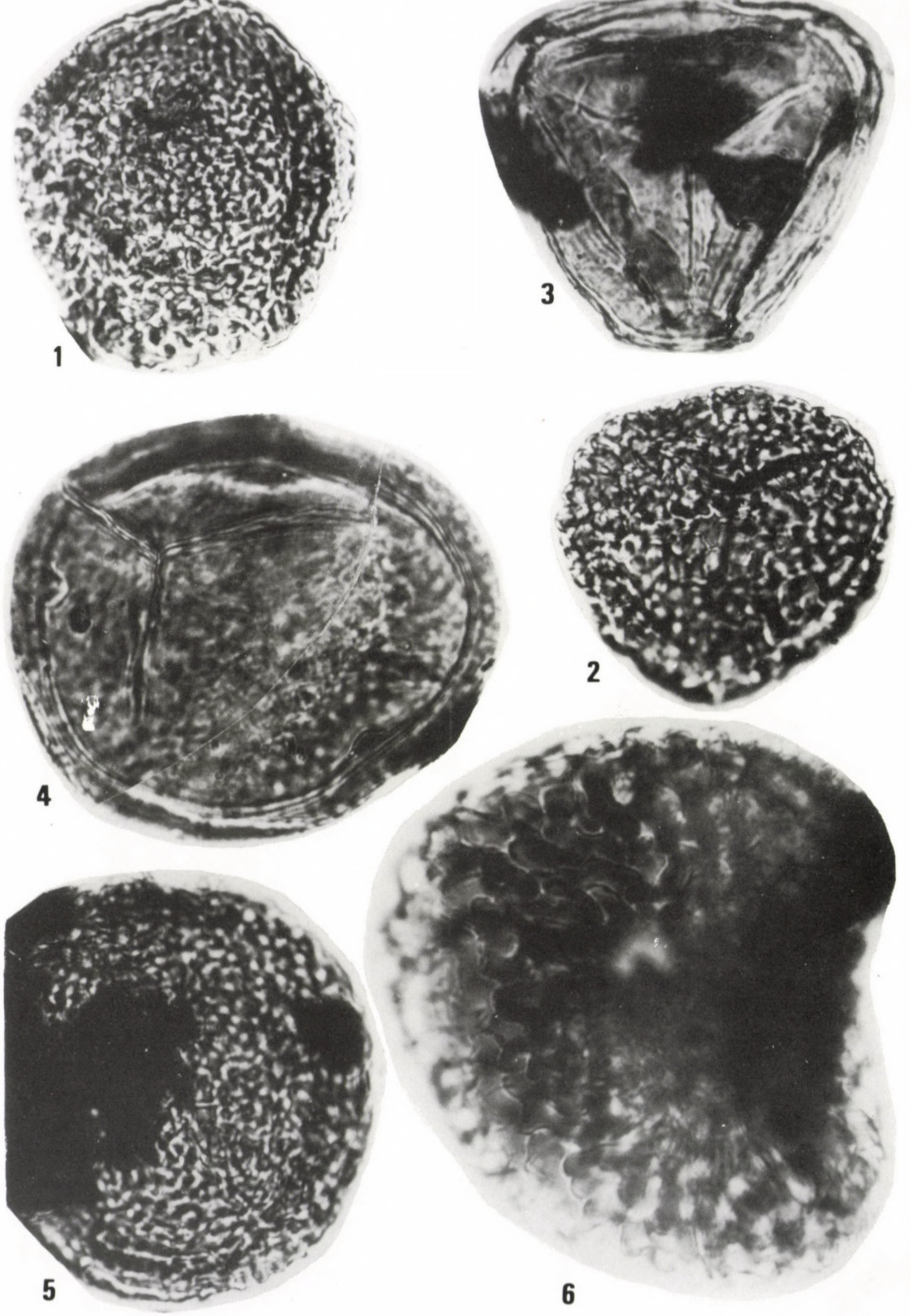

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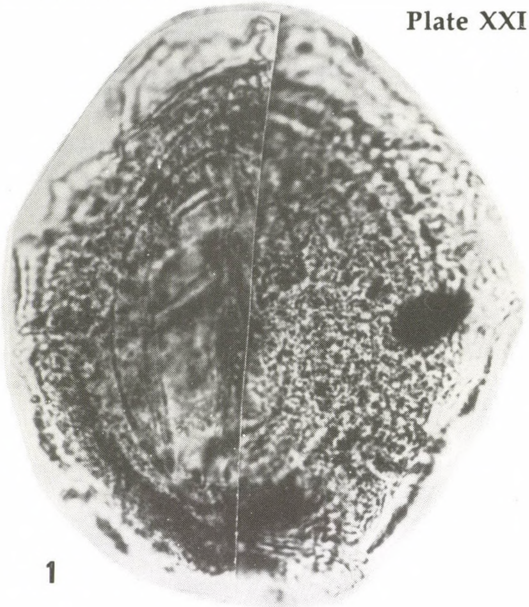

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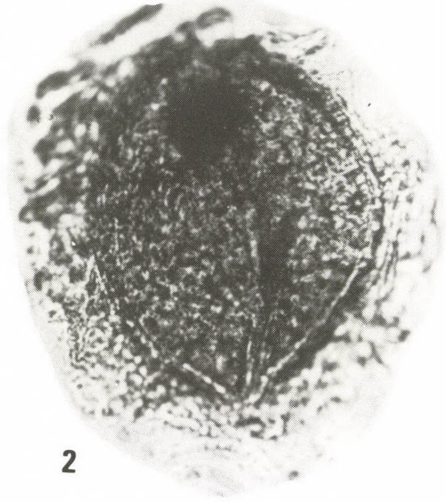

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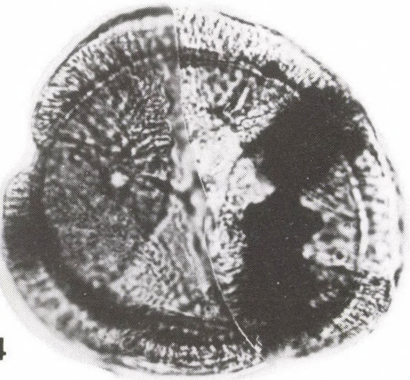

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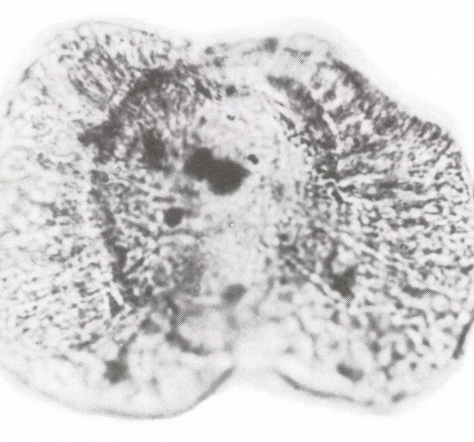

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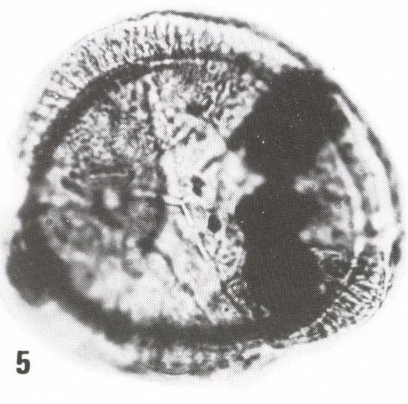

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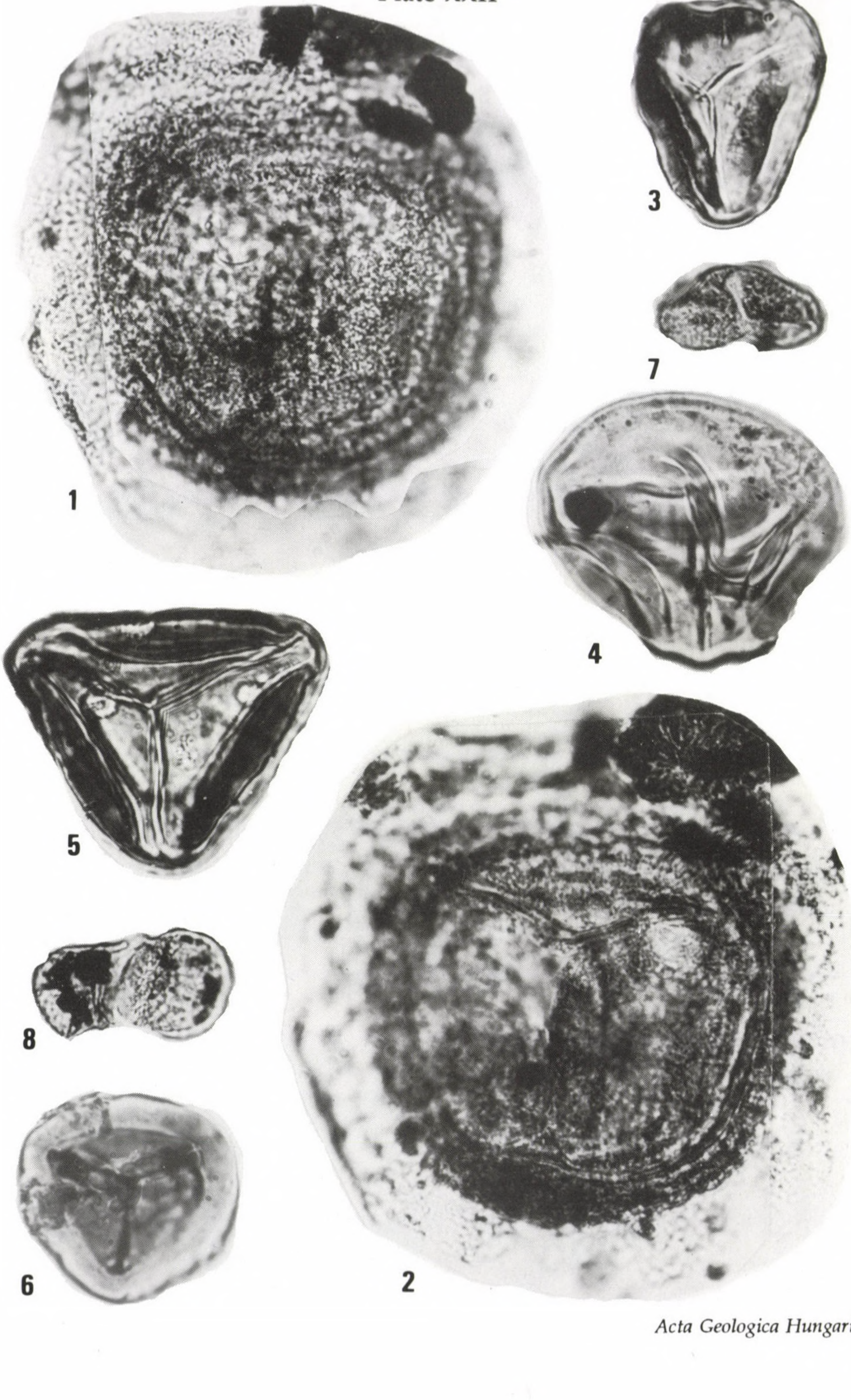

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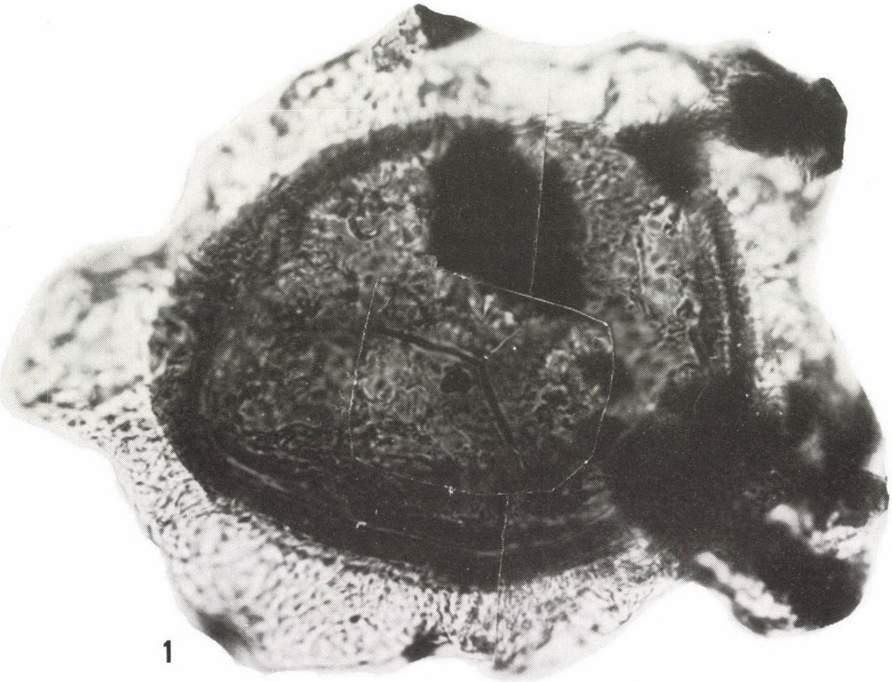

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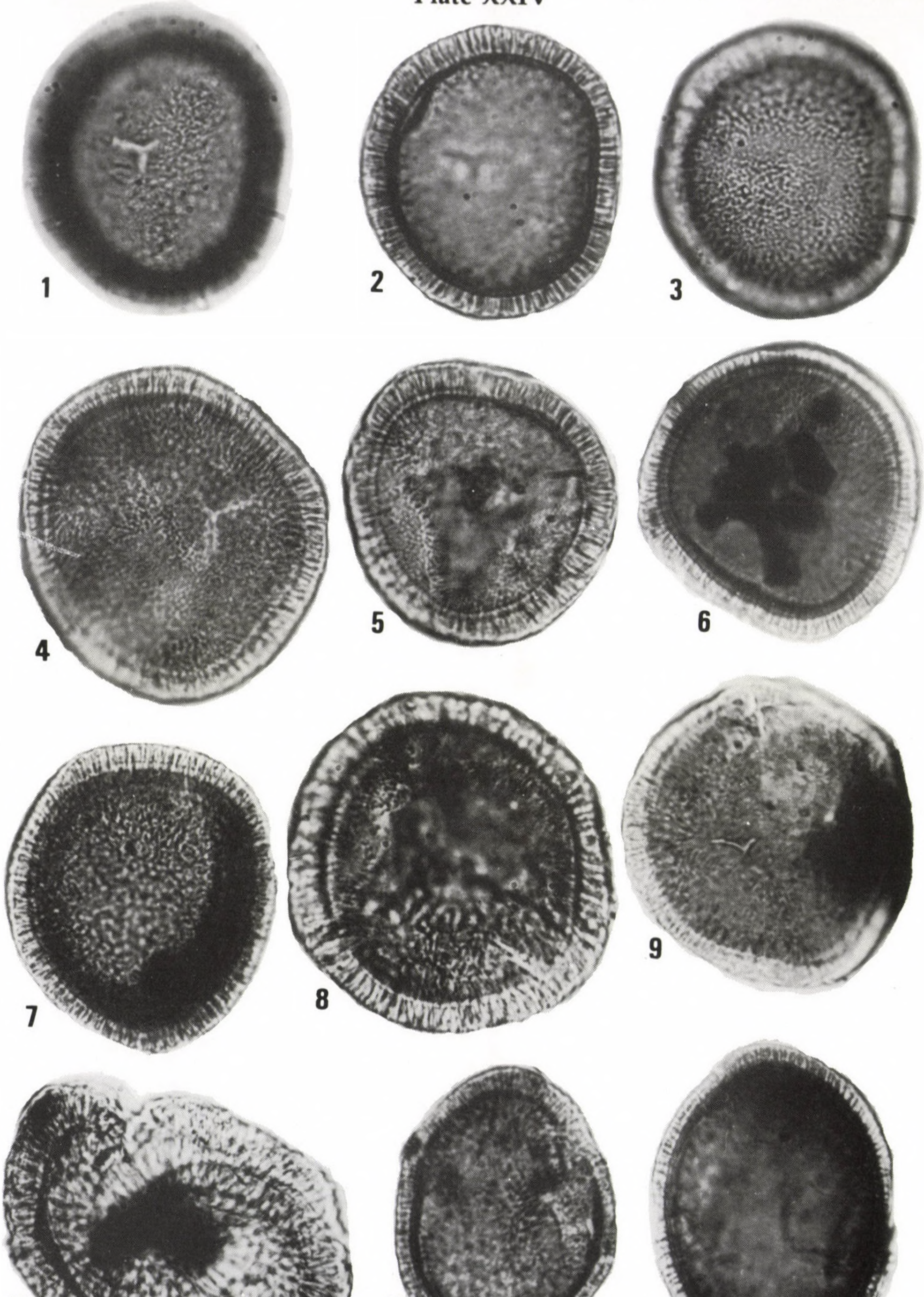

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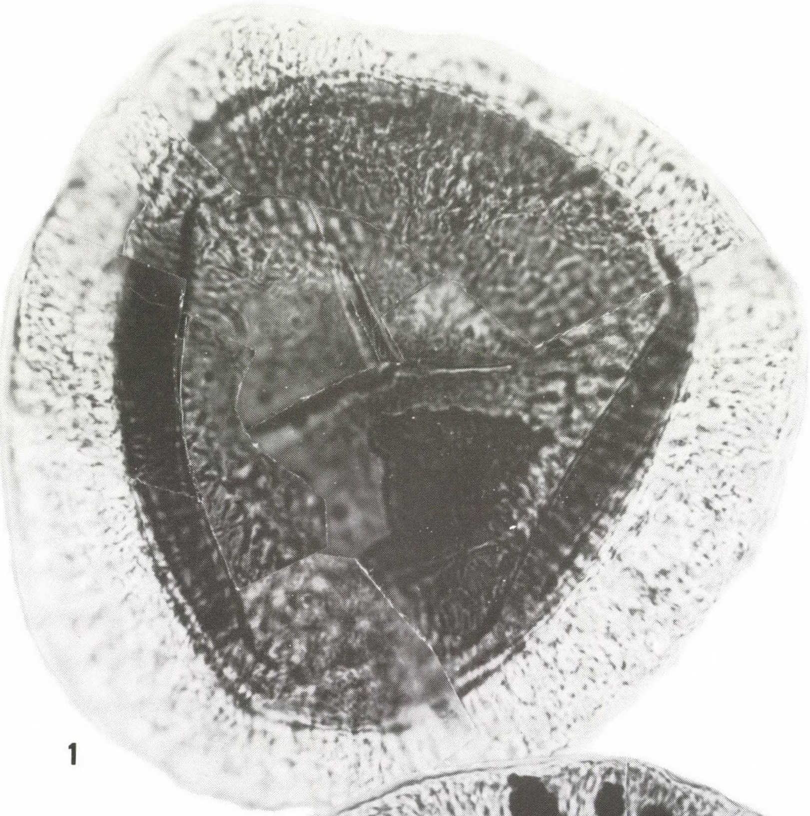

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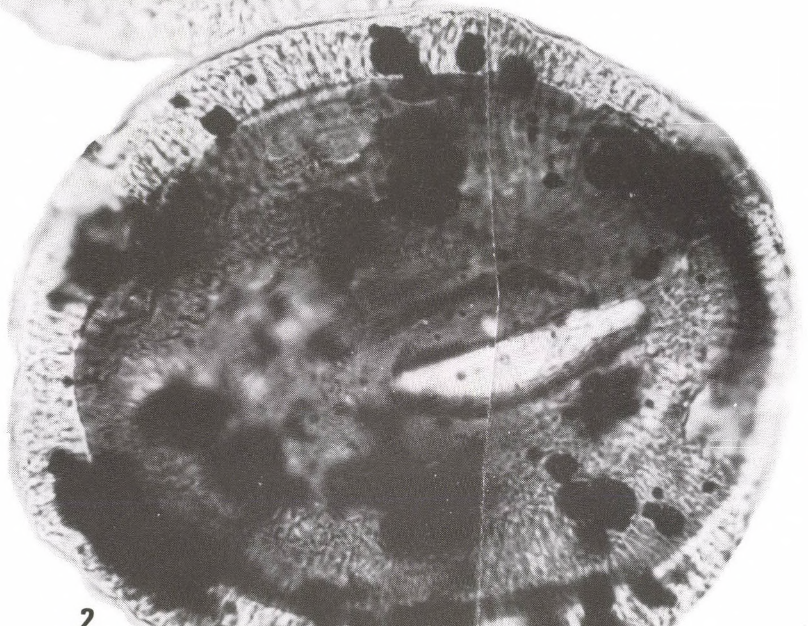

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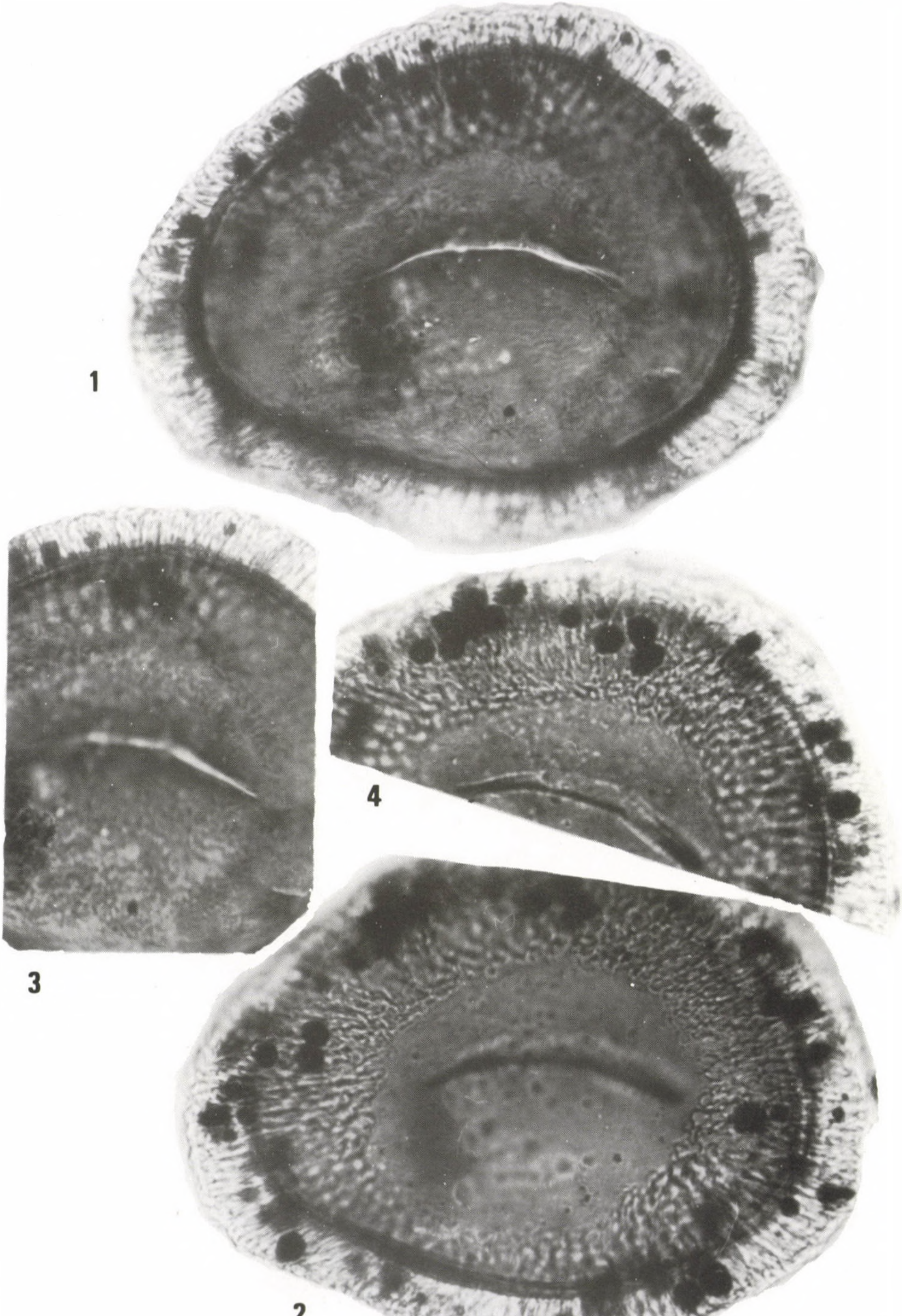

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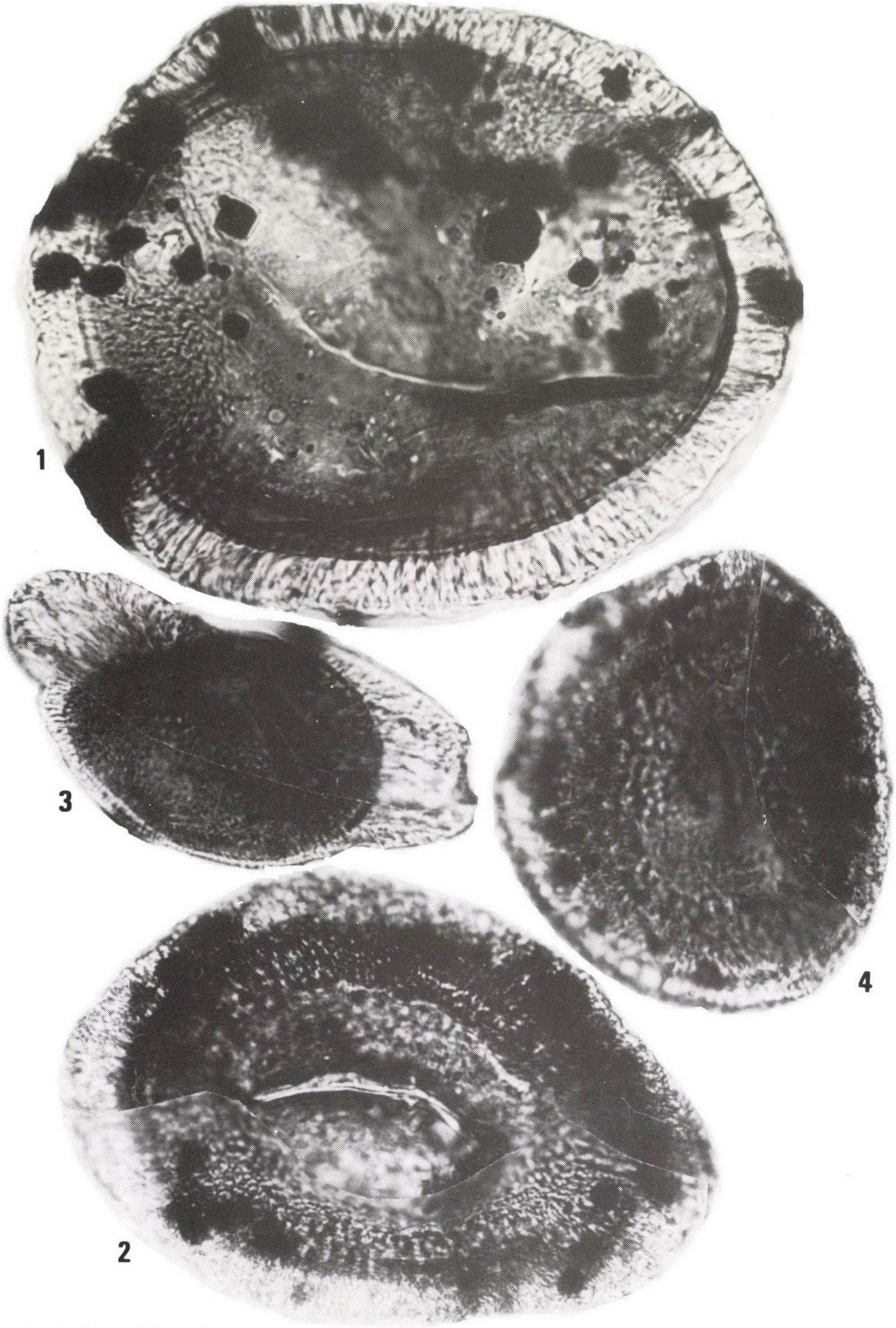

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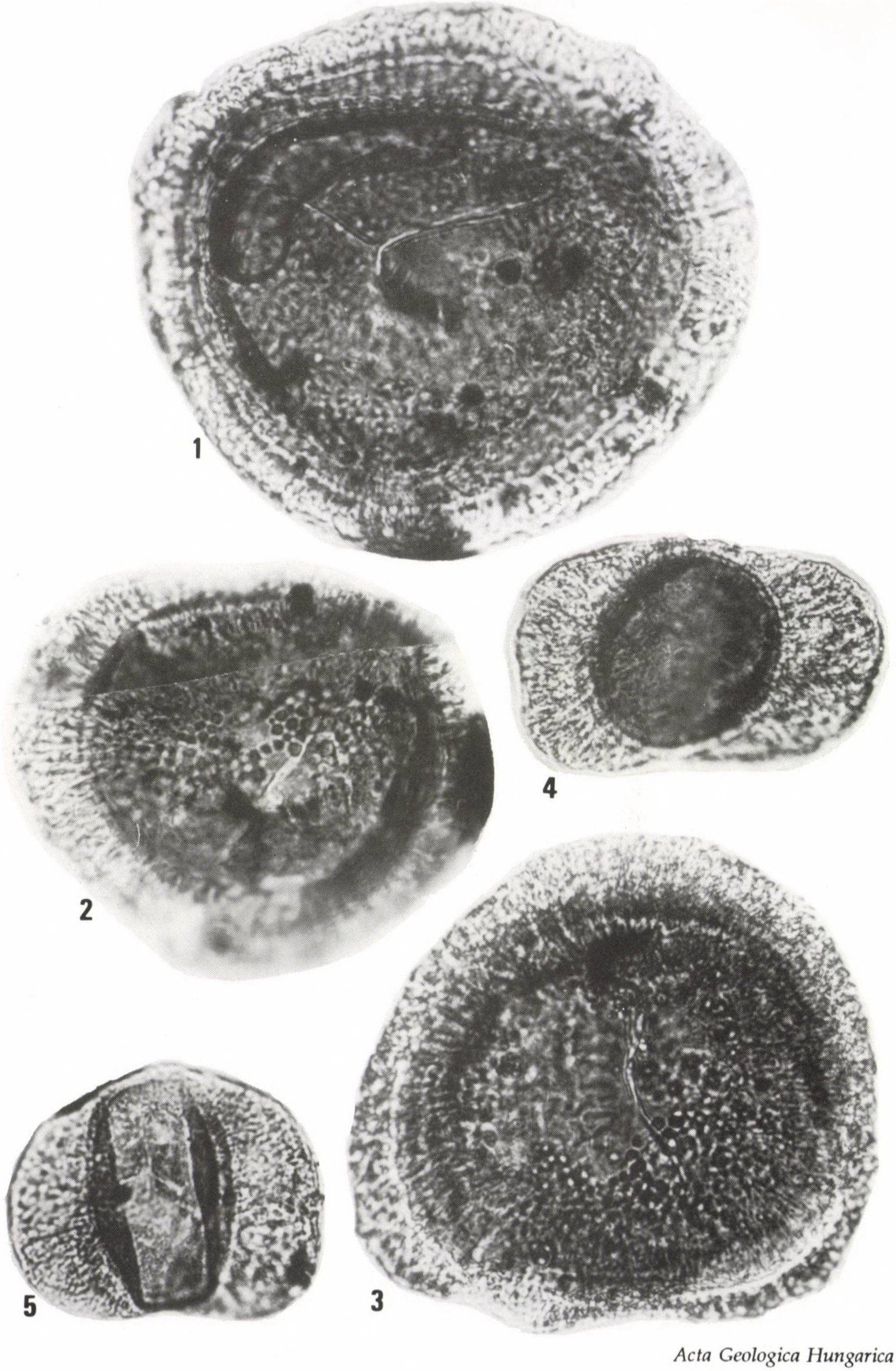

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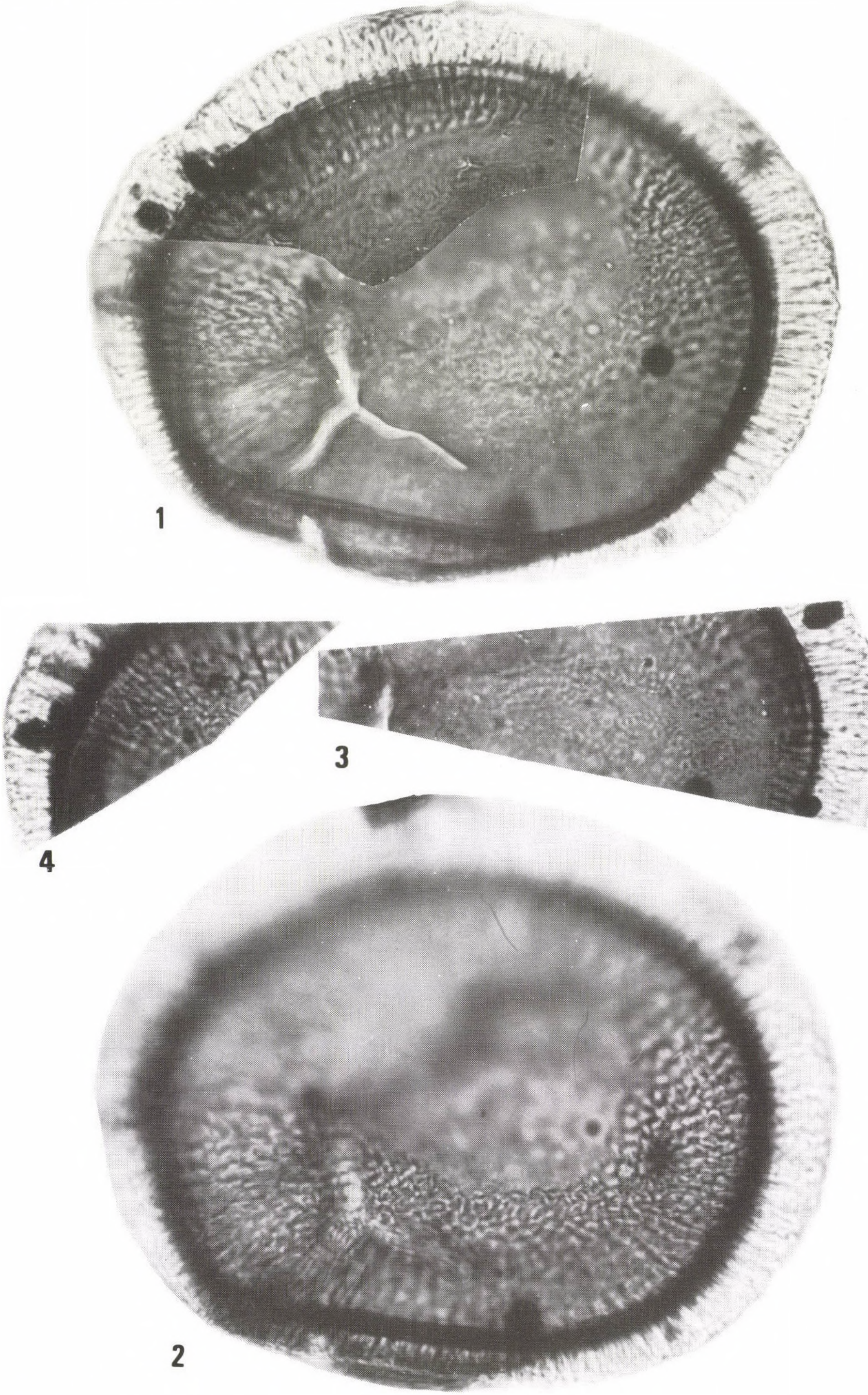

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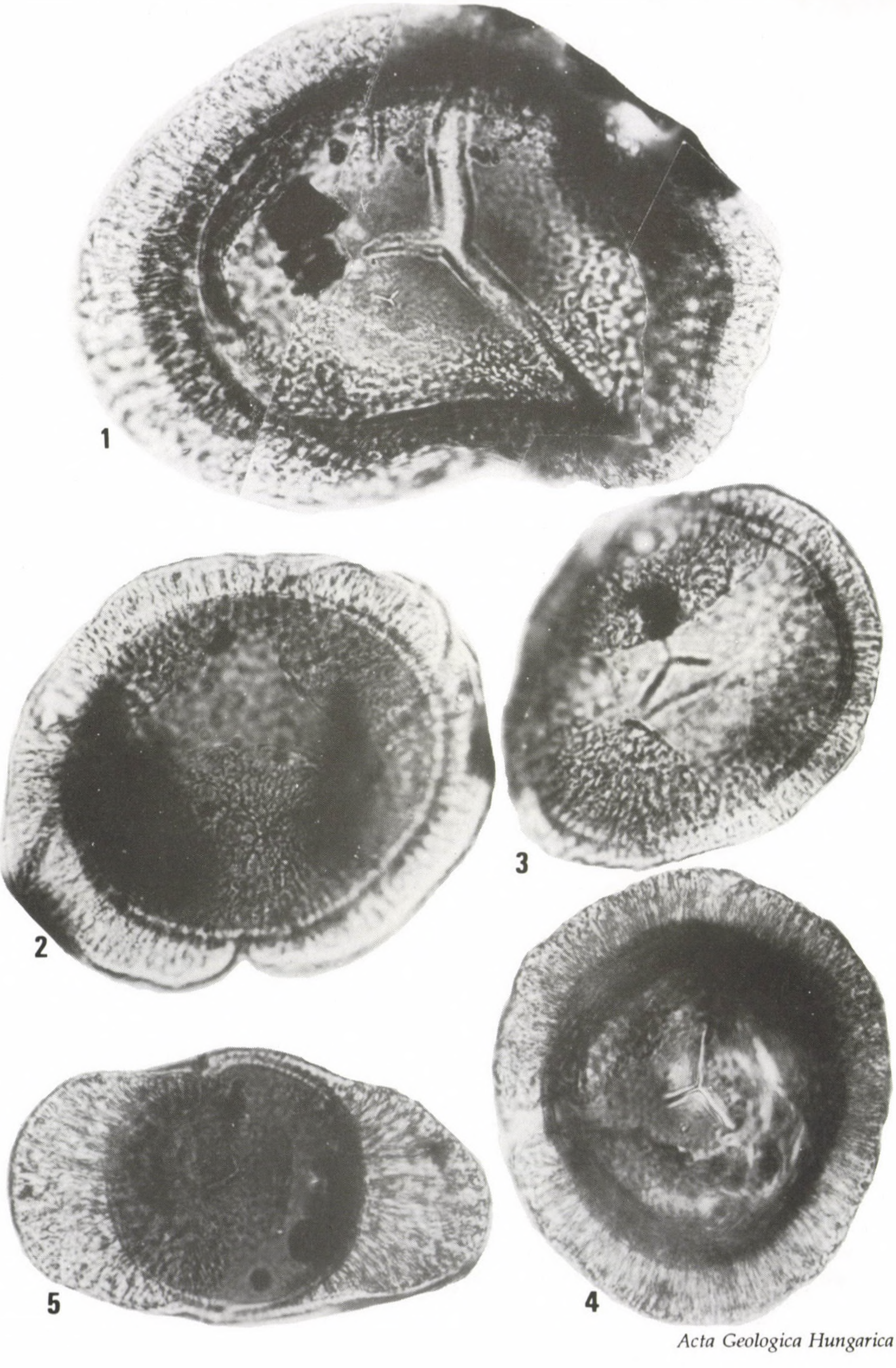

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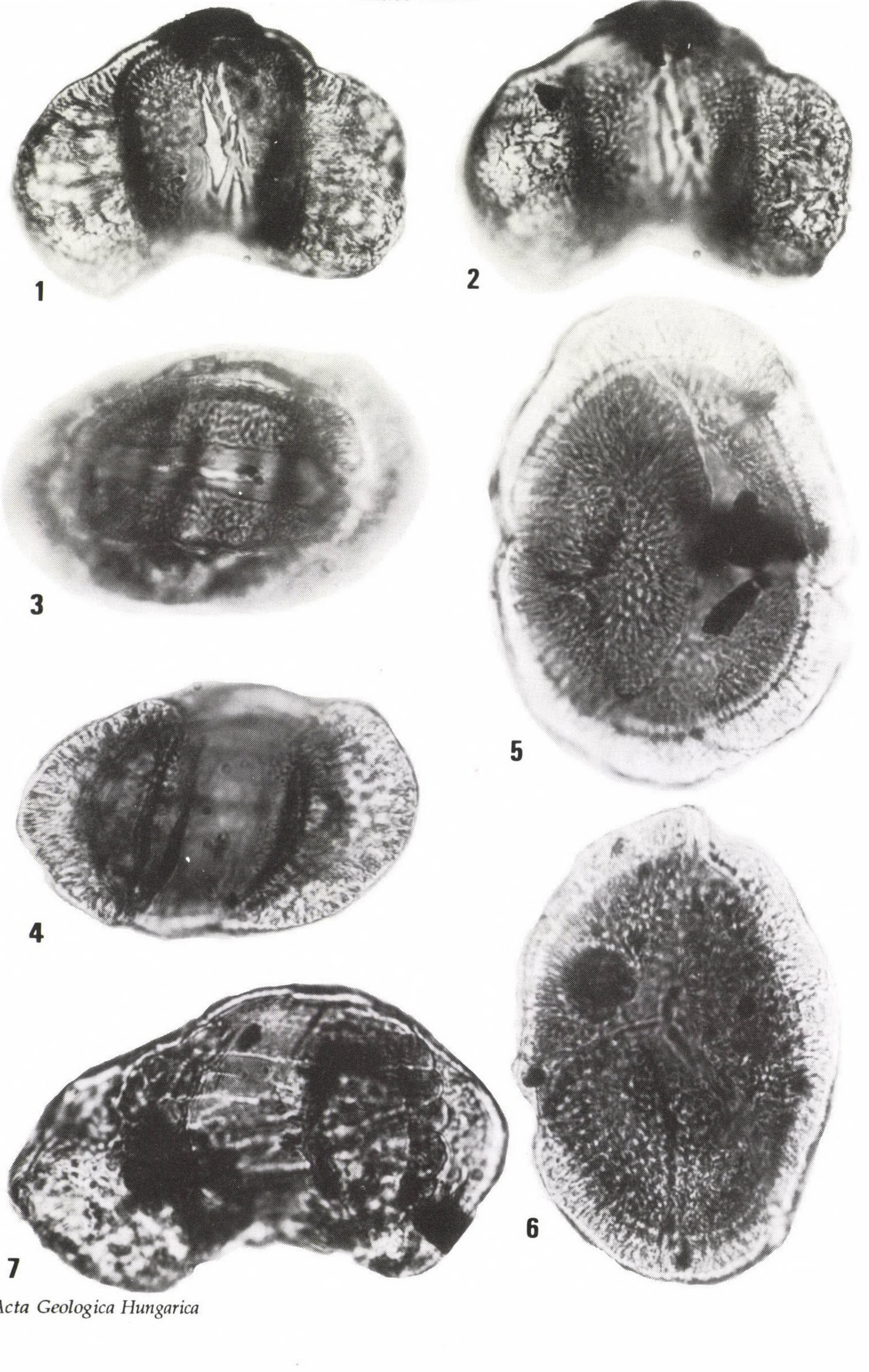

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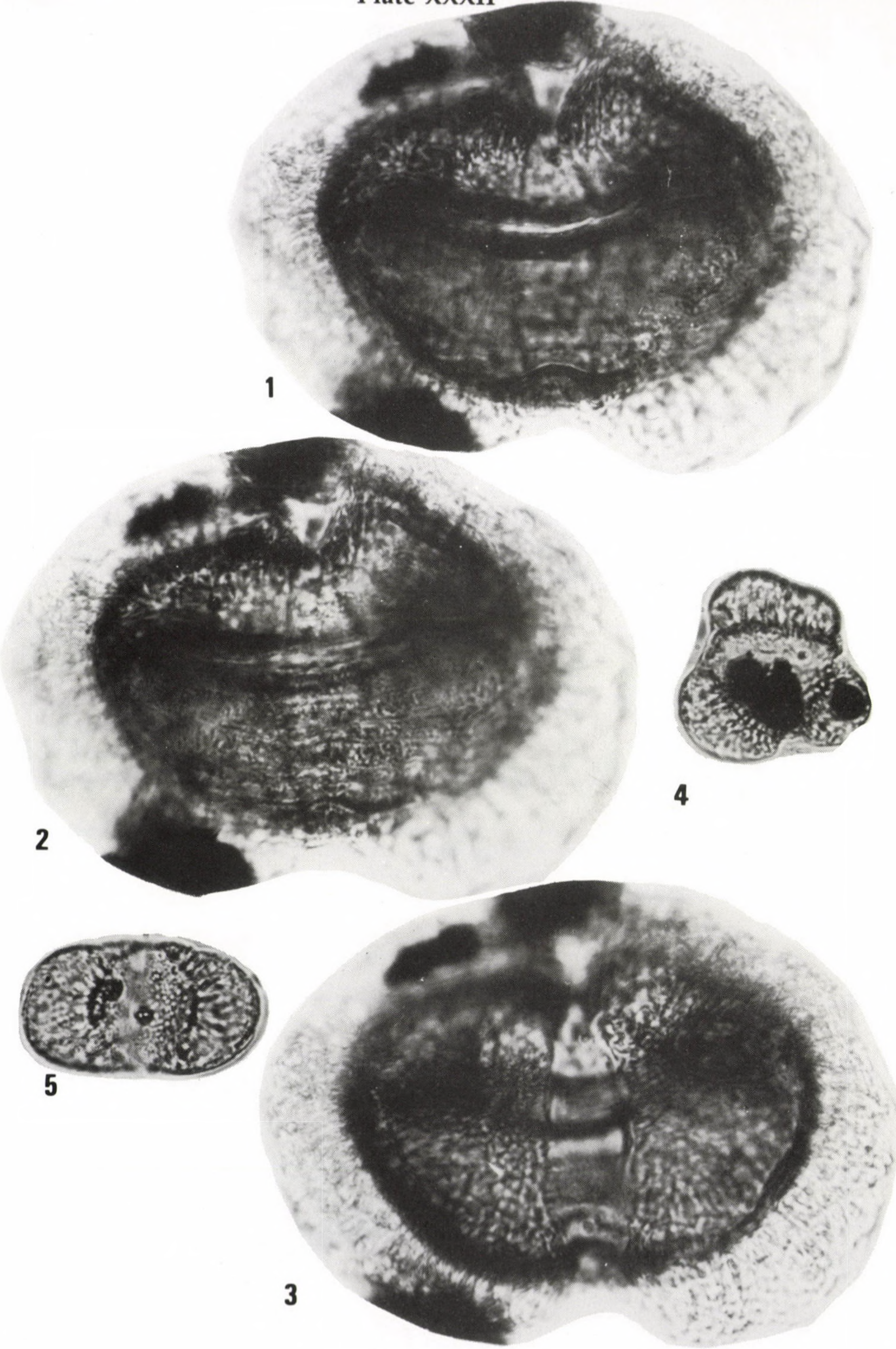

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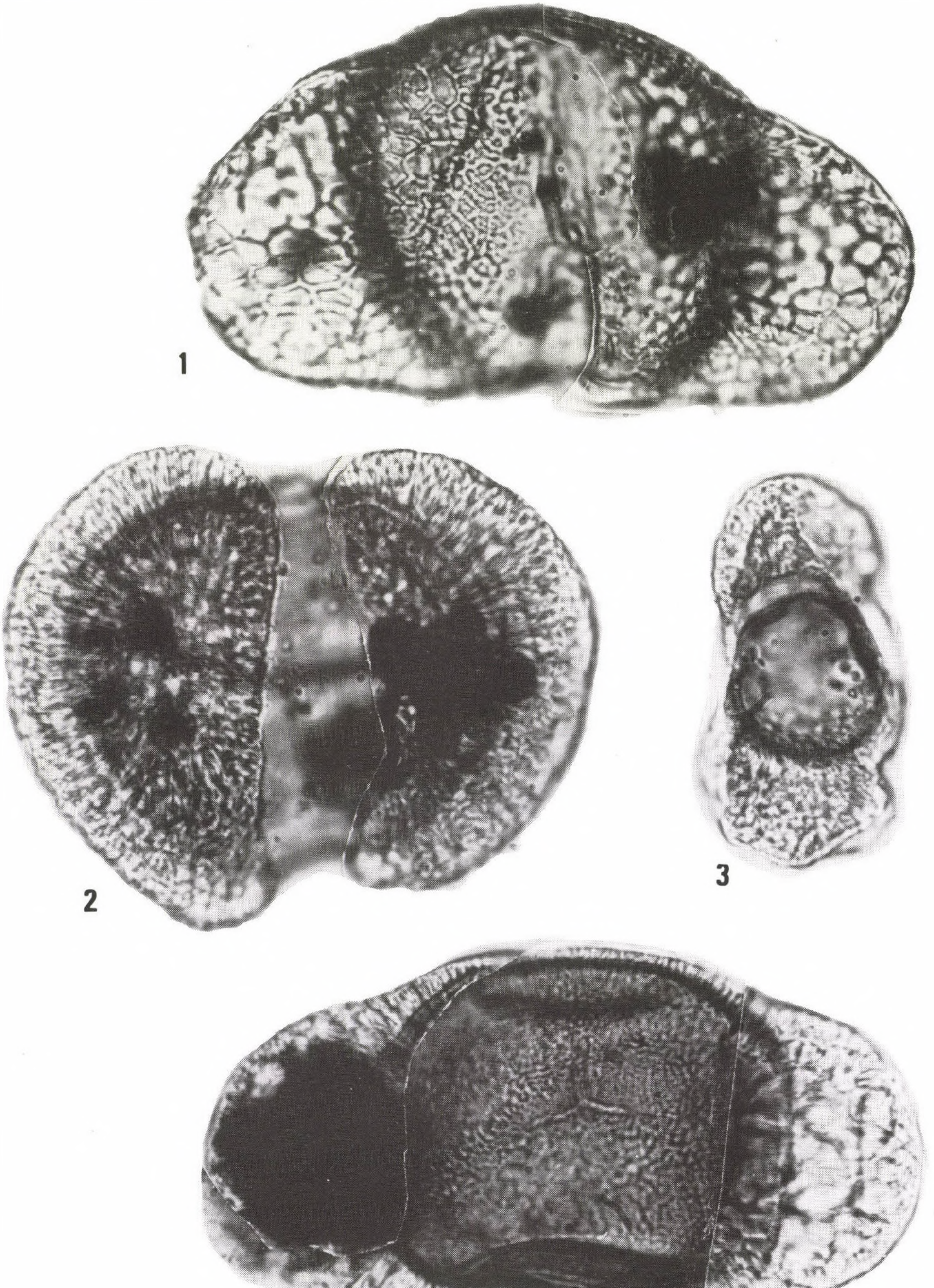

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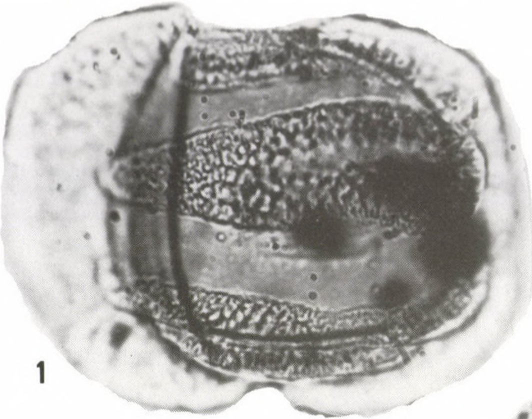

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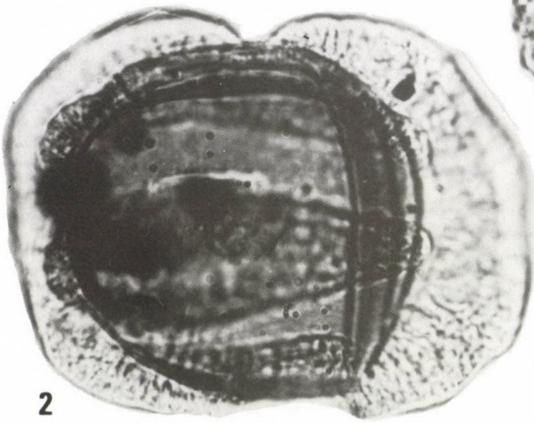

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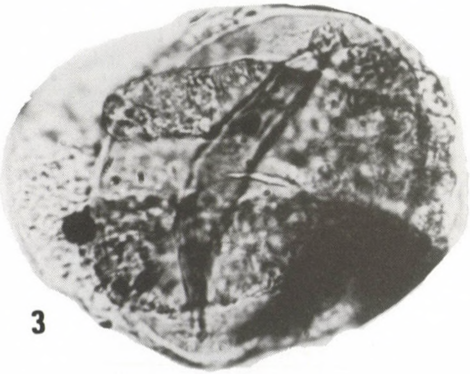

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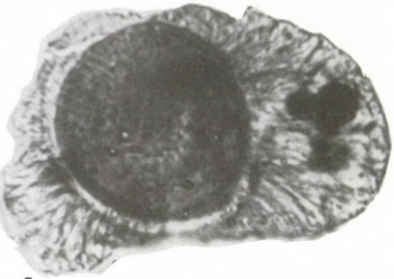

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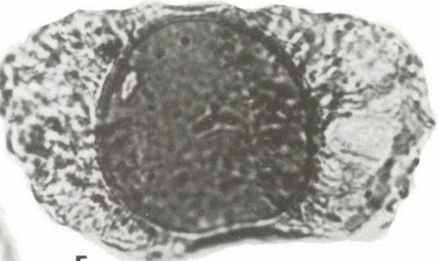

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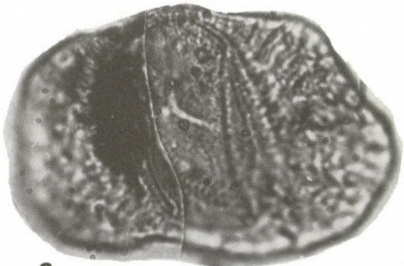

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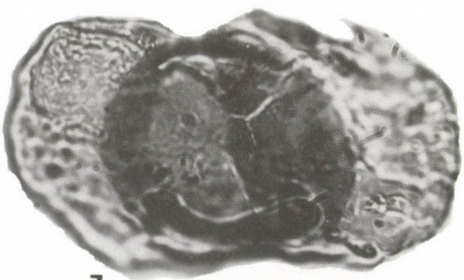

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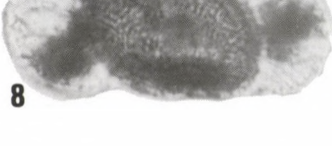

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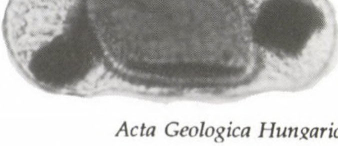

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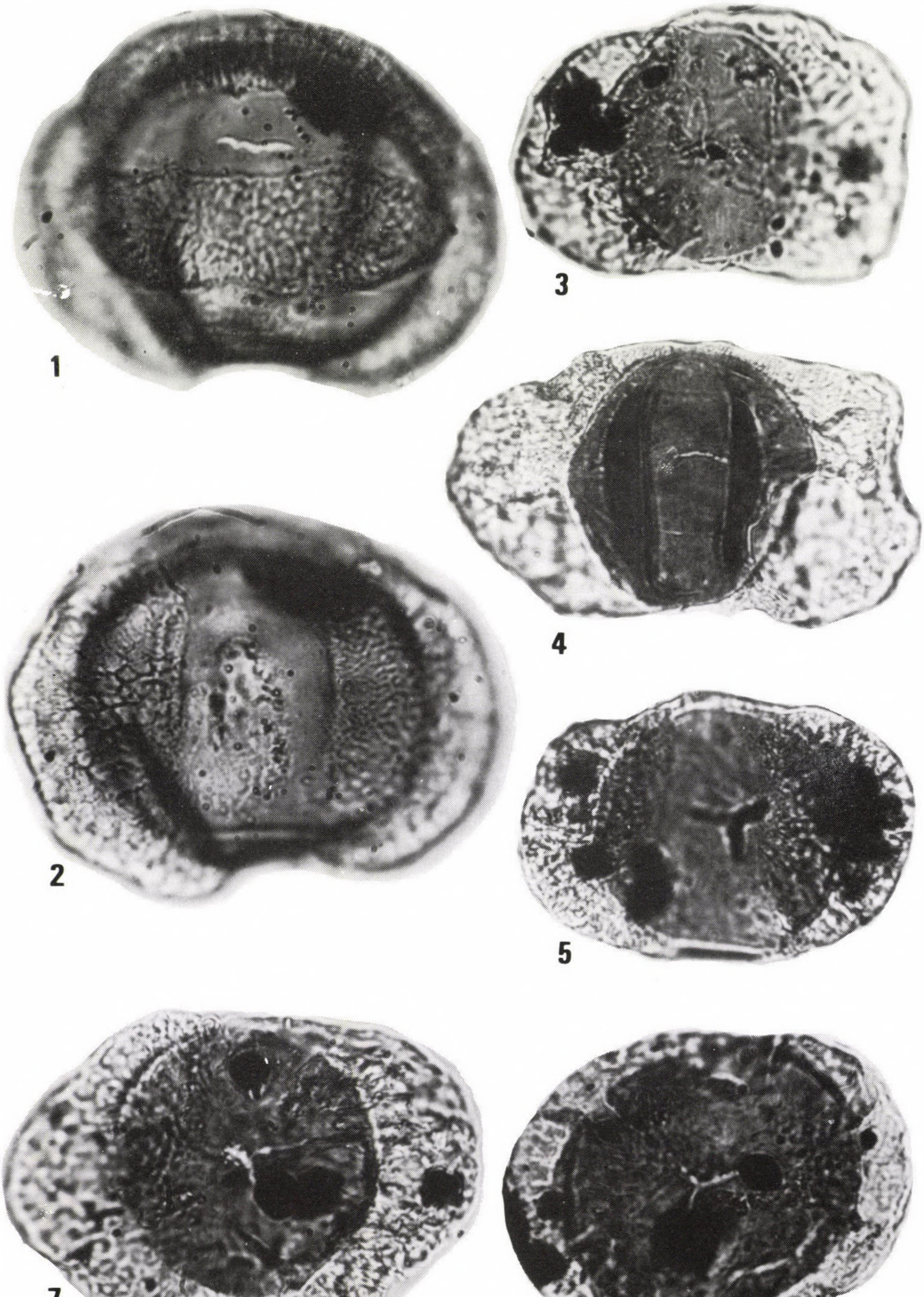

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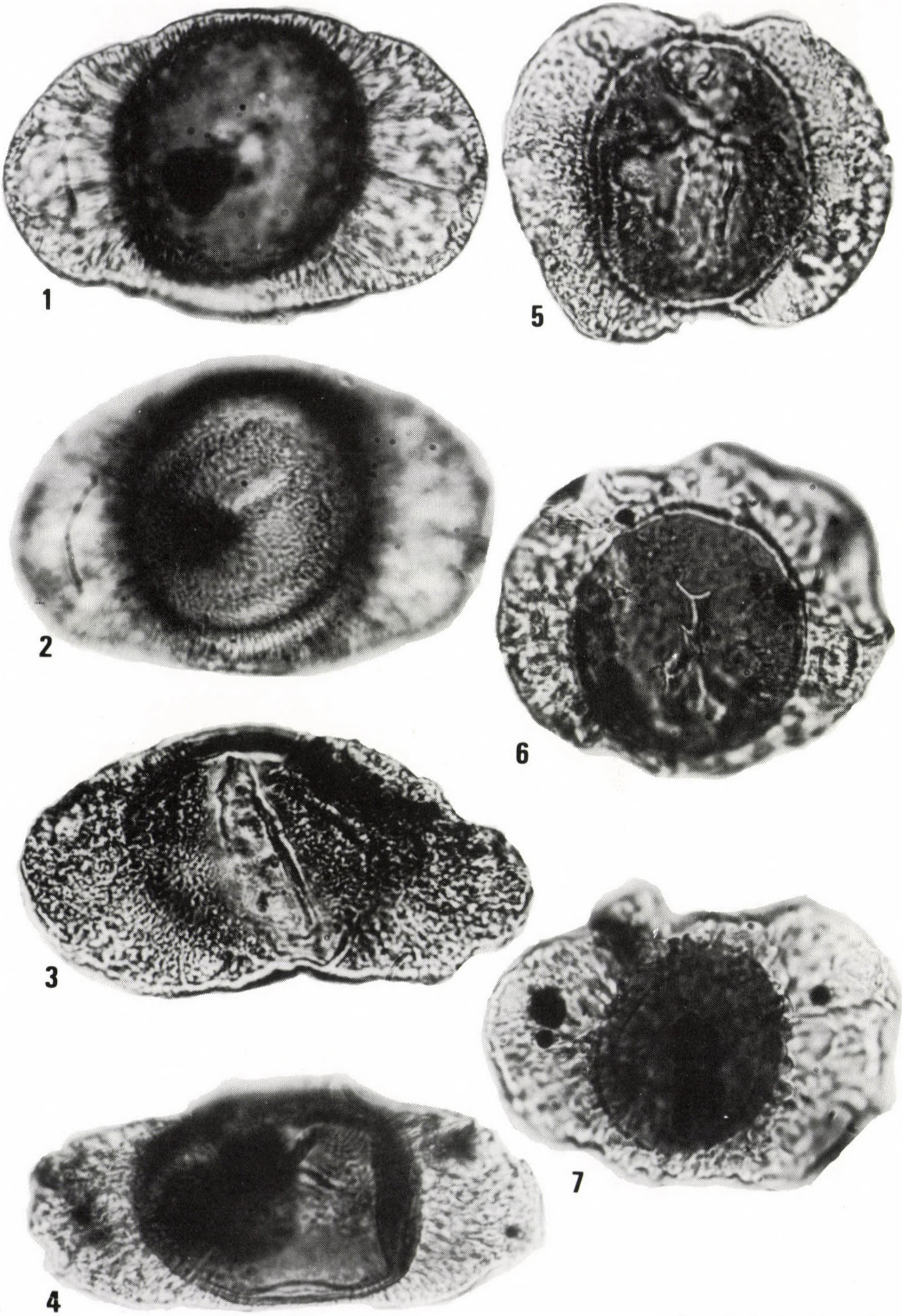

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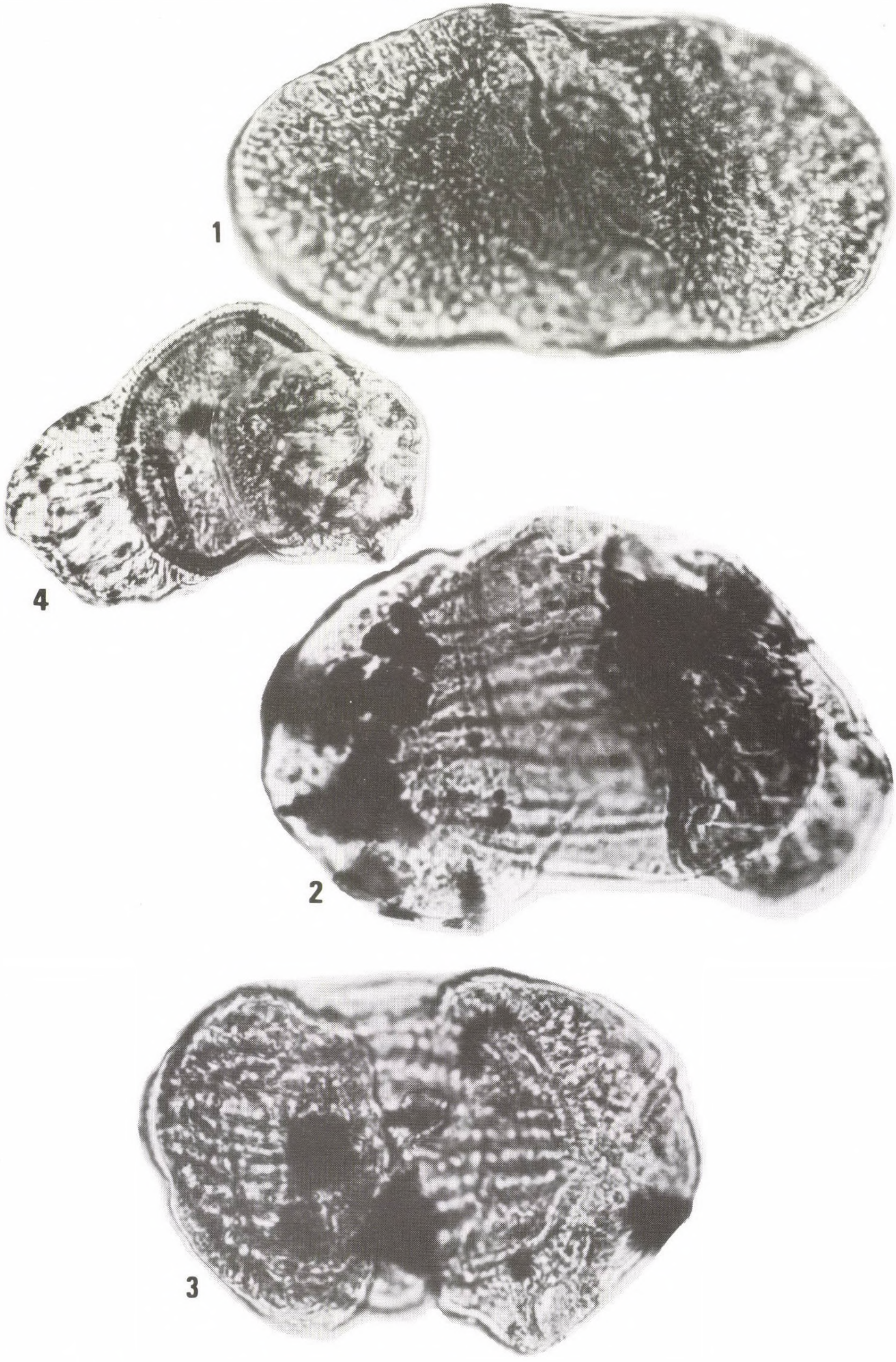

Plate XXXVIII

Plate XXXIX

Plate XL

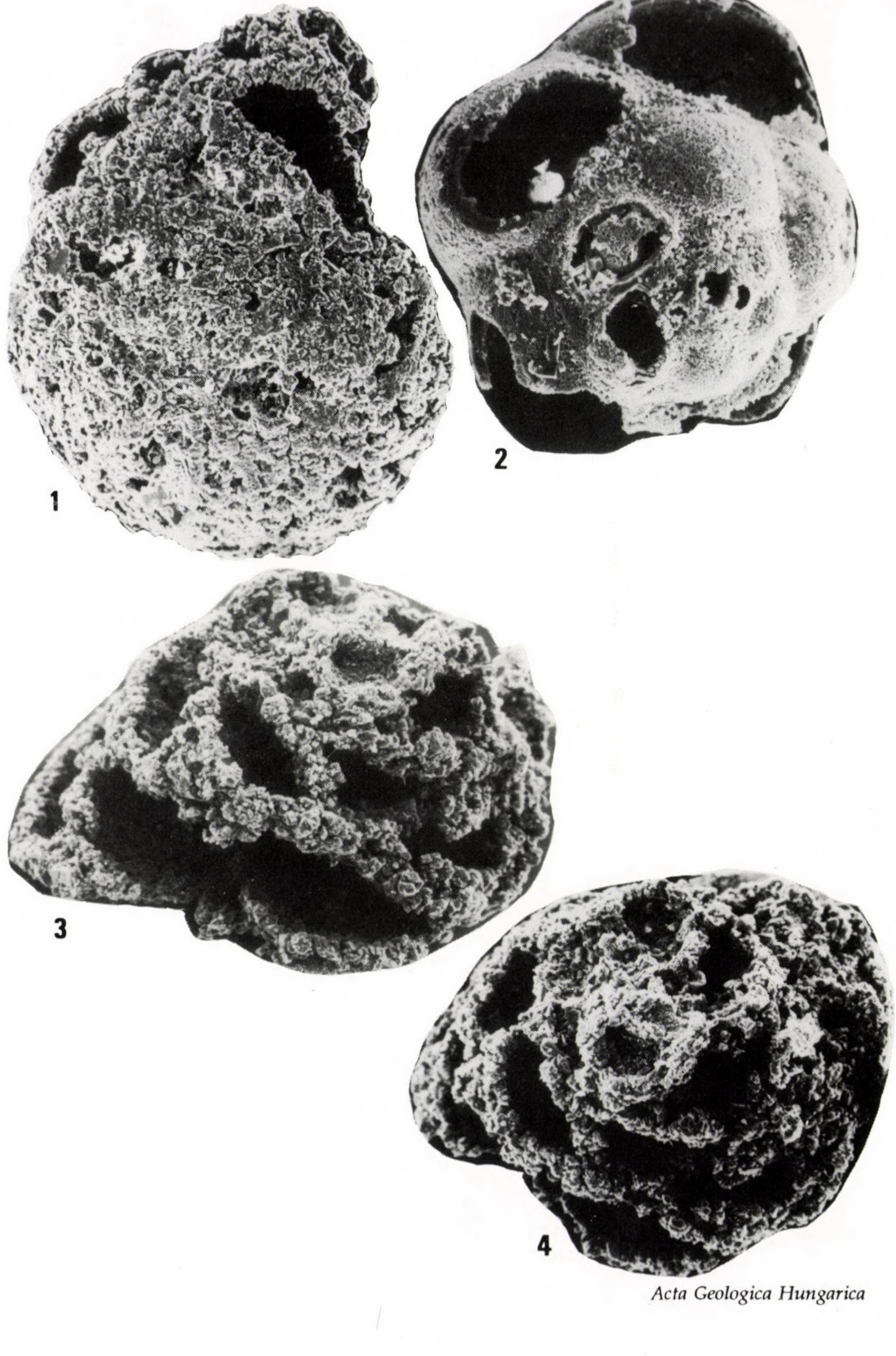

Plate XLI

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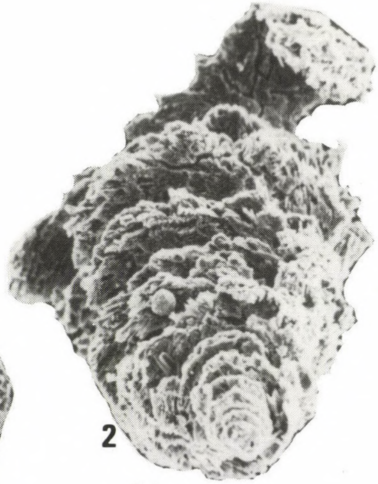

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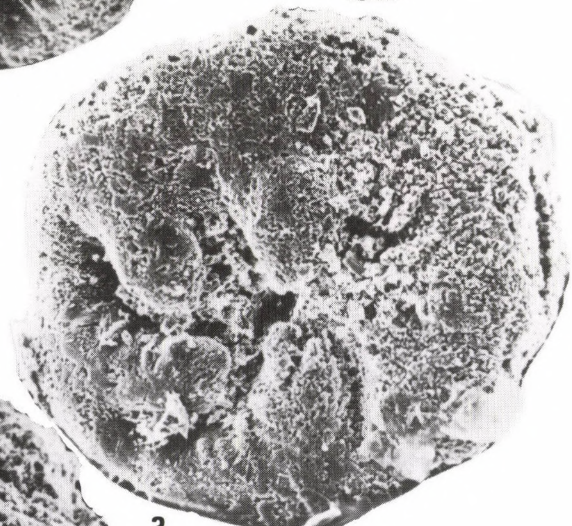

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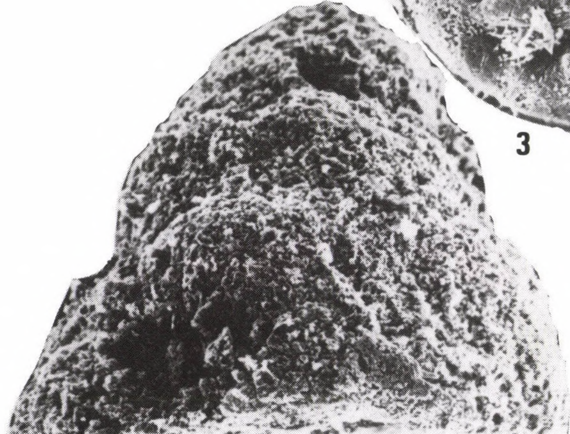

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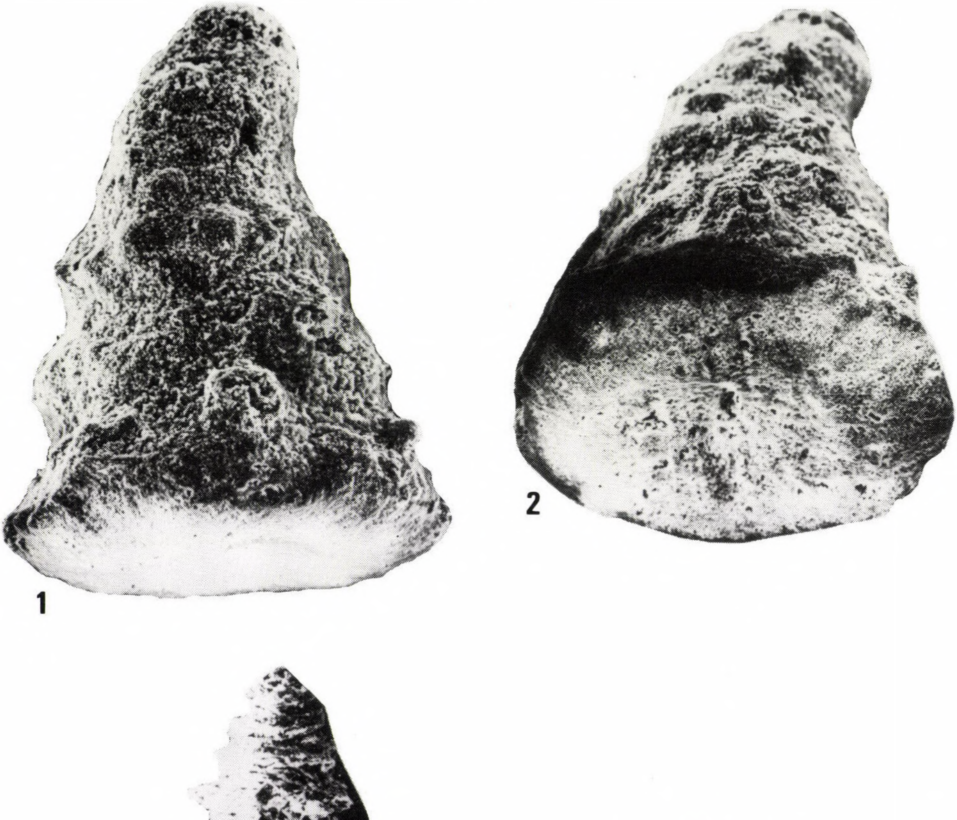
Plate XLII

Plate XLIII



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Evolution of Middle Triassic shallow marine carbonates in the Balaton Highland (Hungary)

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In the Transdanubian Central Range the shallow marine carbonates show a characteristic evolution in the Triassic. The first shallow subtidal carbonates of great areal extension appeared in the Middle Triassic. Well developed "true" platforms with elevated build-ups and steep slopes towards the basins were formed only in the Carnian, and they were separated from one another by intrashelf basins filled with clastic sediments. These were the precursors of the huge shallow marine platforms grown during the Norian-Rhaetian (Hauptdolomite, Dachstein Formation).

Key words: Carbonate sedimentation, palaeogeography, Middle Triassic, Balaton Highland

Introduction

The literature dealing with the Upper Triassic platform carbonates of the Transdanubian Central Range is quite rich both in detailed and in synthetizing works, but this cannot be said in relation to the Middle Triassic. Nevertheless, many sporadic data and statements can be collected in the Hungarian literature about the facies characteristics of the formations, conclusions are, however, usually laconic and contradictory concerning the environment of deposition.

As a result of the geological mapping on the Balaton Highland, finished recently and of the evaluation of several boreholes, some new observations were made on the Middle Triassic shallow marine carbonates, and an overall view can be outlined about their evolution. Fossil (mainly South Alpine) and recent analogies were useful for the interpretation of the palaeoenvironments.

Lithology, facies analysis

The simplified section of Middle Triassic formations of the Balaton Highland is shown in Fig. 1. The lithology, microfacies and sedimentary environments of each unit are summarized as follows.

Aszófő Dolomite Formation

The formation consists of the alternation of thin bedded, lamellar, yellowish-white or light grey, loosely cemented, calcite-spotted, so-called "cellular" dolomite and dolomarl. Its thickness varies between 200 and 250 m. The fenestral birdseye structure is characteristic of it, mud cracks are frequent, while

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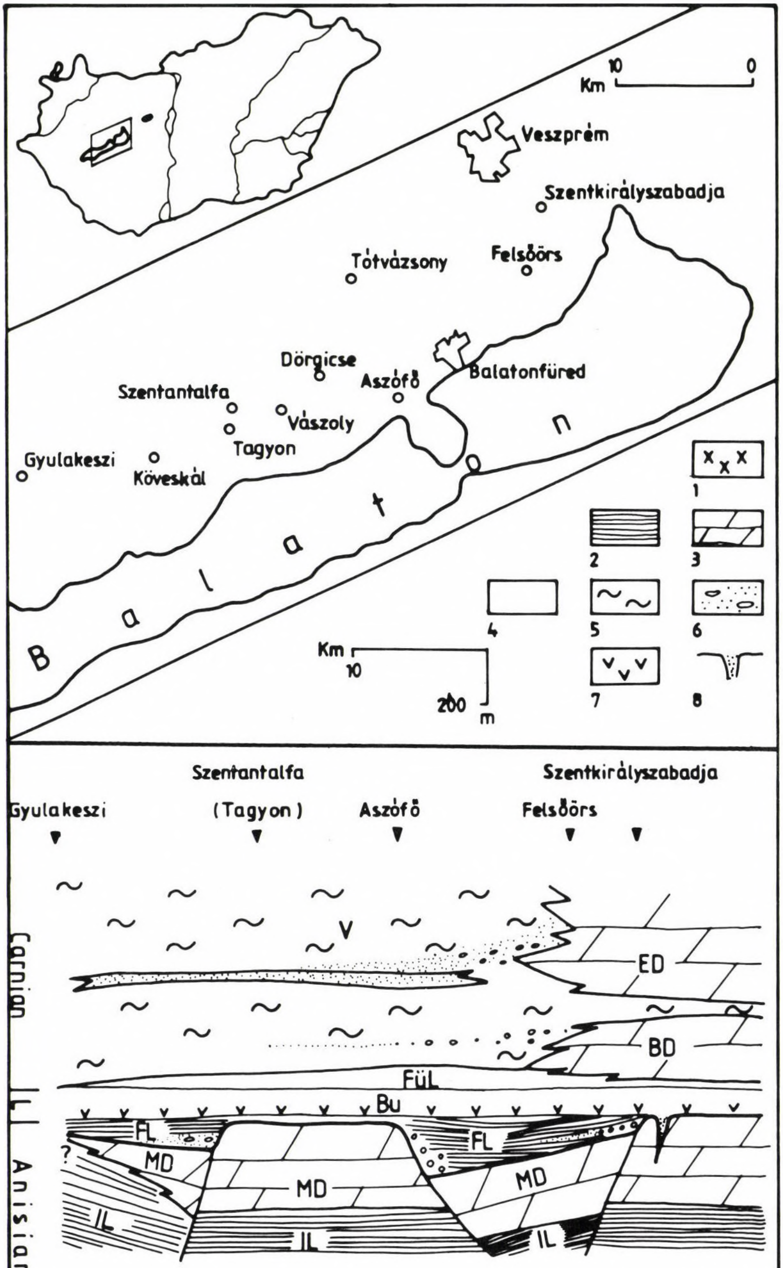
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Vászoly
Tagyon
Gyulakeszi
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Szentkirályszabadja
Gyulakeszi
(Tagyon)
Aszófő
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Ladinian
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the "chicken-wire structure" is subordinate. Its microfacies is characterized by the types below (Haas et al. 1988; Lelkes 1989):

- bioclastic (mainly ostracodal or foraminiferal) micrite (Plate I: 1);
- oosparite, oncosparite (Plate I: 2);
- pelmicrite, clotted micrite;
- dolomicrite (silty and porous, respectively, Plate I: 3);
- peloidal micrite with carbonate pseudomorphs after evaporite minerals (Plate I: 4).

From among the microfacies types above, mainly the dolomicrite and peloidal micrite can be observed forming alternation of micro-layers.

Interpretations known from literature agree that these textural and structural features of the Aszófő Dolomite point to a low energy hypersaline environment, under warm arid climate. This statement is supported by the rather rudimentary fauna of low diversity, characterized by high environmental tolerance (Oravecz-Scheffer 1987, p. 28; Lelkes 1989).

During evaporation the Mg/Ca ratio could considerably increase in sea-water due to gypsum and anhydrite precipitation and this led to the postdepositional direct (syndiagenetic) dolomitization of the lime-mud sediment (Haas et al. 1988, p. 48).

Nevertheless, opinions concerning the depositional environment are different. As to Oravecz-Scheffer (1987 p. 28) and Haas et al. (1988, p. 48), the Aszófő Dolomite was formed in a relatively deep, off-shore, quiet, restricted and hypersaline lagoon. On the contrary, Lelkes (in Császár et al. 1984, p. 11, 1989, pp. 26–27) and Haas et al. (1988, p. 83) stated that the place of deposition was a large, shallow, periodically desiccating near-shore lagoon and/or a tidal flat.

In our opinion both the recent (Haas 1991, pp. 32–46 and Balogh, K. ed. 1991, p. 443) and the fossil analogies (Farabegoli and Levanti 1982, De Zanche and Farabegoli 1988, etc.) suggest that the Aszófő Dolomite of peritidal facies was formed on an arid tidal flat and/or on a sabkha (Fig. 2). The environment of the sediment, deposited originally as lime mud, was shallow subtidal and intertidal where carbonate was essentially produced by blue-green algae. The supratidal sabkha prograded periodically towards the lagoon leading to the early dolomitization of the sediments deposited there.

The silt and marl intercalations occurring in the sequence relate to periodical input of terrigeneous detritus through the channel system. Lelkes (in Császár

←Fig. 1
Simplified section and lateral relationships of Middle Triassic formations in the strike of the Balaton Highland (after Budai and Vörös this volume, Fig. 1). 1. sabkha; 2. restricted (periodically anoxic) basin; 3. carbonate platform; 4. open shelf basin; 5. intrashelf basin with terrigeneous clastics; 6. allodapic clastics; 7. pyroclastics; 8. neptunian dyke; Anisian: AD – Aszófő Dolomite; IL – Iszkahegy Limestone; MD – Megyehegy Dolomite + Tagyon Limestone; FL – Felsőörs Limestone; Ladinian: BU – Buchenstein Fm.; Carnian: FÜL – Füred Limestone; V – Veszprém Fm.; BD – Budaörs Dolomite; ED – Ederics Dolomite

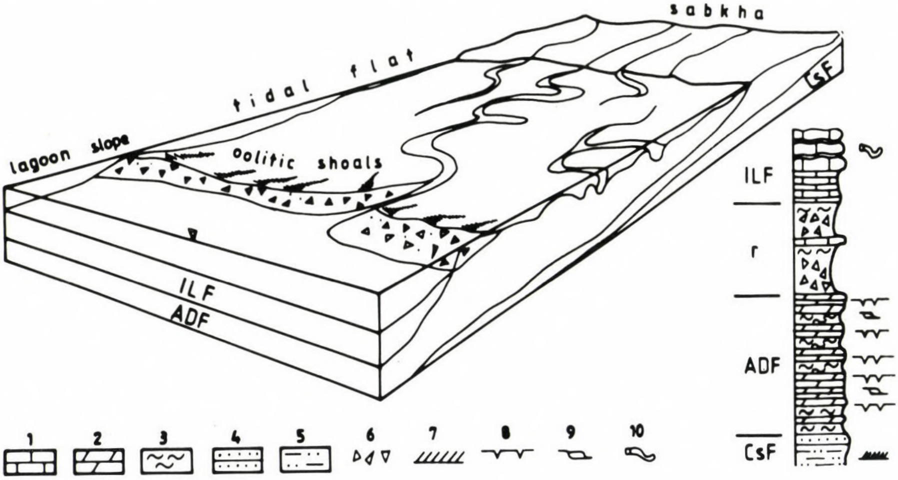


Fig. 2
Simplified strata column and palaeoenvironment of Lower Anisian shallow marine carbonates (modified Fig. 13/c of Haas 1991). 1. limestone; 2. dolomite; 3. marl; 4. sandstone; 5. siltstone; 6. slope breccia; 7. cross-bedding; 8. mud crack; 9. birdseye structure; 10. bioturbation; CSF – Csopak Fm. (Scythian); ADF – Aszófő Dolomite Fm.; r – rauwacke; ILF – Iszkahegy Limestone Fm.

et al. 1984) attributed this phenomenon to seasonal freshwater floods that is also characteristic of the arid climate. As to the evaluation of the ooidic–oncoidic microfacies (Haas et al. 1988; Lelkes 1989), the tidal flat was separated from the subtidal lagoon by calcareous sand shoals or bars (Fig. 2).

In our opinion the breccia intercalated with a thickness of a few tens of meters between the Aszófő Dolomite and the overlying Iszkahegy Limestone (see later) penetrated in several boreholes (e.g. Balatonfüred Bdt. 2) was deposited on the slope between the tidal flat and the subtidal lagoon (Fig. 2). This may be the special lithofacies (rauhwacke) of the Aszófő Dolomite that occurs on the surface at the boundary of the two formations as porous-alveolar, white or limonite coloured limestone or dolomite breccia with calcareous cement.

The Lombardian Carniola di Bovegno and the Lower Serla Formation of the Dolomites (Frassene Dolomite) are completely analogous to the Aszófő Dolomite both in lithology and stratigraphic position as well as in palaeogeographic aspects (Budai 1992). As regards the geologic evolution, the analogy seems to be closer to Lombardy where the intertidal sedimentation turns to subtidal (Angolo Limestone) in the Lower Anisian (Unland 1975; De Zanche and Farabegoli 1988, etc.). On the contrary, in the Dolomites erosion prevailed in major areas (Piz da Peres Conglomerate, Voltago Conglomerate, etc.), whereas the shallow marine carbonate formation continued elsewhere (Upper

Serla Formation, see Assereto et al. 1977; Farabegoli and Guasti 1980; De Zanche et al. 1992).

Iszkahegy Limestone Formation

The Aszófő Dolomite is overlain by dark grey, greyish brown, well stratified, in its lower part lamellar, upwards thick bedded, vermicular, bituminous limestone with marl intercalations. Its thickness is about 250–300 m. Its microfacies is usually micritic mudstone or wackestone, with ostracod and mollusc shells, foraminifer tests and subordinately with echinoderm skeletons (Plate II: 1, 2).

On the basis of the investigation of the Balatonfüred Bdt. 2 borehole the depositional environment of the Iszkahegy Limestone could be a shallow subtidal lagoon (Lelkes 1989). Previously it was a common view (e.g. in the opinion of mapping geologists) that the bituminous and lamellar character of the Iszkahegy Limestone can be attributed to the anoxic environment developed under humid climate. In the lamellar limestone, however, collected in the Malom-völgy at Felsőörs, a monospecific ostracod assemblage indicating unambiguously hypersaline environment was found (Monostori 1992). Thus, the preservation of lamellar structure, i.e. the lack of burrowing infauna and the high amount of organic matter was not necessarily resulted from the anoxia developed under humid climate but rather from the high salinity at the bottom that may bring about anoxic environment in the deep waters above the sea-floor (see Haas 1991, p. 43). The parallel stratified, lamellar variety of the Iszkahegy Limestone is characteristic of the lower part of the unit. During the deposition of the overlying thick bedded, vermicular limestone the sea-floor could be more ventillated than in the earlier phase.

According to most authors the Angolo Limestone, being completely analogous to the Iszkahegy Limestone, deposited in a low-energy, poorly ventillated subtidal lagoon (Unland 1975; Gaetani et al. 1979, etc.). As regards their palaeogeographic position, however, it is an essential difference that the formation of the Angolo Limestone continued also among the Pelsonian carbonate platforms (Camorelli, Dosso dei Morti) in Lombardy.

Megyehegy Dolomite Formation

The Megyehegy Formation consists of light grey, yellowish grey, locally bluish-purplish dolomite. It is usually bedded or thick bedded, medium to well stratified with flat bedding-planes. Its microfacies is represented by usually strongly recrystallized, rather monotonous dolosparite (Plate II: 3). The marly dolomite developing gradually from the underlying Iszkahegy Limestone is slightly bituminous, often biodetrital and is characterized by crinoidal skeletal fragments, dasycladaceans, e.g. in the surroundings of Tótvázsony (det. Piros, in Csillag 1991): *Physoporella pauciforata, Ph. pauciforata undulata, Oligoporella* sp. The Megyehegy Dolomite is otherwise poor in fossils, in addition to the sporadical foraminifers and dasycladaceans (Plate II: 4), crinoids and brachiopods occur in the upper part of the formation.

According to the previous facies analyses the Megyehegy Dolomite was formed in a carbonate platform lagoon (Lelkes, in Császár et al. 1984), less hypersaline as compared to the environment of the Aszófő Dolomite, but on the basis of the microfauna it could be of greater salinity than the normal value (Oravecz-Scheffer 1987).

In our opinion the evolution of the Megyehegy Dolomite can be divided into two phases. Its lower unit, that is widespread on the Balaton Highland, was formed in a shallow subtidal environment, similar to the Upper Serla Formation of the Dolomites (Gaetani et al. 1981). This ramp could join to the open shelf basin through a gentle slope as it can be concluded from the gradual transition (bituminous dolomarl) between the Megyehegy Dolomite and the overlying Felsőörs Limestone of basin facies. In the major part of the Balaton Highland the shelf evolution did not exceed the carbonate ramp state, i.e. during the Pelsonian no platform was developed characterized by cyclic sedimentation. The lack of intertidal sedimentary structures can be explained first of all by this and not only by the subsequent recrystallization.

In the northern strip of the Balaton Highland the thickness of the Megyehegy Dolomite is considerably smaller, locally it is less than 10 m (see Csillag 1991) than in the southern one, reaching occasionally 270 m. It cannot be excluded that in these localities the material of the thin dolomite body between the Iszkahegy and Felsőörs Limestone is allodapic, transported into the basin from a farther shallow ramp. This facies evolution is similar for the most part to that of the Lombardian areas (see Budai 1992), where the Angolo Limestone is directly overlain by the Prezzo Limestone of basin facies.

The second phase of the formation of the Megyehegy Dolomite can be detected in the areas of the Balaton Highland where the deposition of shallow marine carbonates continued up to the end of the Illyrian. This period is, however, characterized by the platform carbonate of the Tagyon Limestone, that is heteropic to the upper part of the Megyehegy Dolomite.

Tagyon Limestone Formation

The gradual Anisian evolution of the Balaton Highland was suddenly and drastically disturbed by a geological event. As a result of the Pelsonian synsedimentary tectonics, the shallow marine carbonate ramp was broken into blocks along normal faults and in the subsided areas relatively narrow, more or less restricted basins were generated (Felsőörs Limestone). The central part of the Balaton Highland remained in elevated position up to the Late Illyrian (Budai and Vörös 1992), where the formation of shallow marine carbonates continued. A platform developed from the Middle Anisian carbonate ramp that emerged above the surrounding basins (Figs 1, 4).

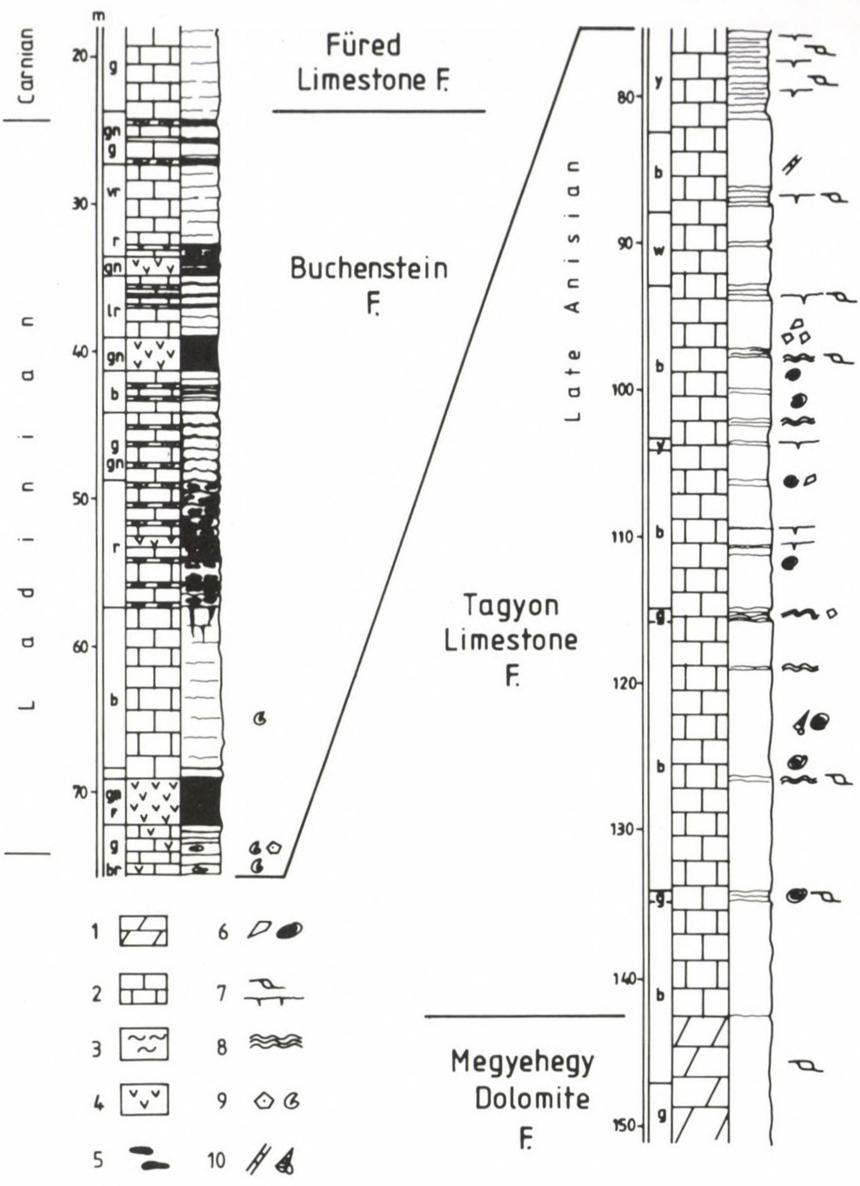


Fig. 3
Stratigraphic column of borehole Dörgicse Drt.1 (modified after Budai 1988, Fig. 8). 1. dolomite; 2. limestone; 3. marl; 4. tuff, tuffite; 5. chert; 6. breccia, oncoid; 7. fenestral structure, mud crack; 8. algal mat; 9. crinoids, ammonites; 10. Dasycladacea, gastropods; g – grey; gn – green; vr – violet-red; r – red; lr – light red; b – beige; br – brown; y – yellow; w – white

The Tagyon Formation is built up by the alternation of white, light grey or beige bedded limestone and of yellow, algal laminated limestone with birdseye structures penetrated by mud cracks. In the borehole Dörgicse Drt. 1 the bedded dasycladacean and the thin stratified calcite-spotted yellow limestone alternate rhythmically (Fig. 3), forming Lofer cycles of intertidal B and subtidal C members (Budai 1992). As to the recent studies, red subaerial formations of the A member can also be detected locally.

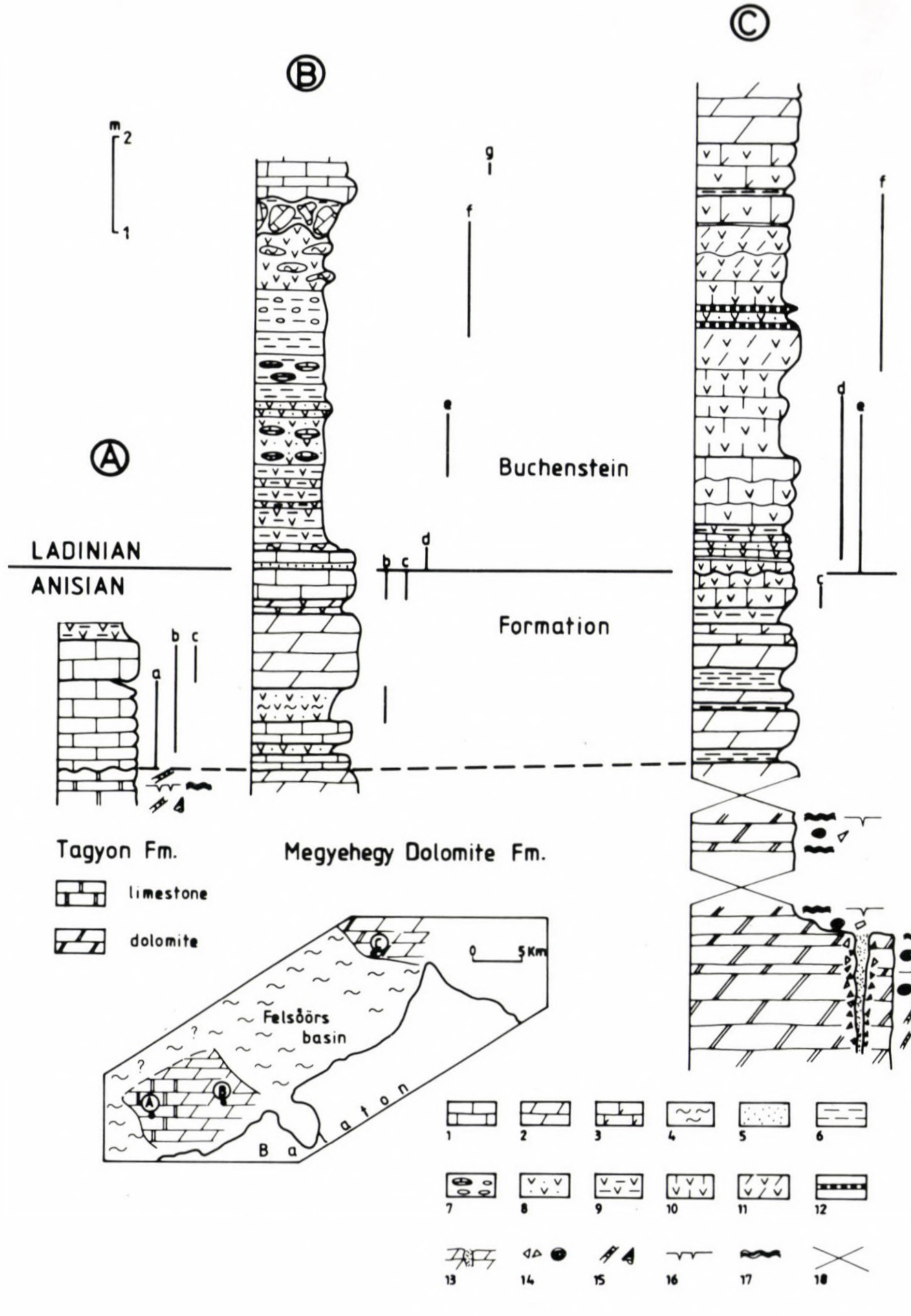
The dolomite observed in a quarry near the military airport at Szentkirályszabadja shows also the features of the Tagyon Formation. The dasycladacean beds containing large oncoids and intraclasts alternate rhythmically with thinner, algal laminated, in certain horizons reddish strata (Fig. 4, c). The most frequent dasycladaceans are: *Physoporella pauciforata, Physoporella pauciforata sulcata, Physoporella pauciforata undulata, Teutloporella peniculiformis.*

In the borehole Szentantalfa Szaf. 1 and Dörgicse Drt. 1 Oravecz-Scheffer (1980, in Budai et al. 1990) distinguished the main microfacies types as follows: oncosparite, stromatolite, biomicrite, biosparite and pelbiosparite. According to the new observations these sediments, deposited originally in tidal environment (Plate III: 1–4), have a subaerial, vadose, pedogenic overprint as evidenced by special rocks (pisoidal-peloidal caliche, occasionally brecciated vadose carbonate crusts, see Plate IV: 1–4).

According to the data of outcrops and boreholes (Szentantalfa Szaf. 1 and Dörgicse Drt. 1), the most frequent fossils of the biogenic limestone are the dasycladaceans (see Plates V and VI), such as *Physoporella pauciforata pauciforata, Ph. pauciforata undulata, Ph. pauciforata sulcata, Oligoporella pilosa, Teutloporella peniculiformis* (Piros 1992). Foraminifers (*Glomospira div.* sp., *Glomospirella triphonensis, Trochammina almtalensis, Earlandinita div.* sp., *Endothyranella wirzi, Meandrospira dinarica, Diplotremina astrofimbriata*), subordinately pelecypods and gastropods as well as ostracods and echinoderms also occur (Oravecz-Scheffer 1980, 1987). Based on the sporadic sponge and coral remains it can be also presumed, that small bioherms (patch reefs) surrounded the platform lagoon separating it from the open shelf basin (Budai and Vörös 1992).

Concerning the facies and palaeogeographic position, the Tagyon Limestone can be correlated with the Steinalm Limestone (Piros 1992) of North Hungary (Aggtelek), and with the Upper Anisian Contrin Formation of the Dolomites (Gaetani et al. 1981). Slope breccias and blocks collapsed from the platform into the basin described in the Dolomites, are unknown in the Balaton Highland.

Fig. 4 →
Boundary sections of Upper Anisian platform carbonates in the strike of the Balaton Highland., A: Szentantalfa, B: Vászoly, C: Szentkirályszabadja (after Budai and Vörös 1991, 1992; Vörös and Pálfy 1989 Fig. 4; Vörös 1991 Fig. 2). 1. limestone; 2. dolomite; 3. calcareous dolomite, dolomitic limestone; 4. marl; 5. sandstone, calcarenite; 6. clay; 7. limestone lens, concretion; 8. crystal tuff; 9. tuffaceous clay; 10. calcareous tuffite; 11. dolomitic tuffite; 12. chert; 13. neptunian dyke; 14. breccia, oncoid; 15. Dasycladacea, gastropods; 16. mud crack; 17. algal mat; 18. unexplored part of the section; Ammonite-ranges: a – *Paraceratites trinodosus*; b – *Reiflingites camunus*; c – *Parakellnerites* sp. aff. *meriani*; d – *Kellnerites felsoeoersensis*; e – *Hyparpadites liepoldti*; f – *Xenoprotrachyceras reitzi*; g – *Eoprotrachyceras curionii*



C
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g
f
e
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Buchenstein
LADINIAN
ANISIAN
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Tagyon Fm.
Megyehegy Dolomite Fm.
limestone
dolomite
0 5Km
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basin
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This is probably due to the fact that the transitional zones between the platforms and basins are very poorly exposed in the latter region.

The Upper Anisian platforms drowned everywhere in the Balaton Highland up to the Late Illyrian. Surviving platforms, such as in several localities in the Dolomites – where the Contrin Formation is replaced by the shallow marine Sciliar Dolomite (Assereto et al. 1977 Fig. 2; Farabegoli and Levanti 1982 Fig. 2) – are unknown so far in the study area.

Two Upper Anisian platforms can be outlined in the Balaton Highland. Biostratigraphically evaluated sections are available in the central platform at Szentantalfa and Vászoly and on the margin of the northeastern one at Szentkirályszabadja. Comparing these sections the following statements can be made (Fig. 4):

– according to the ammonite assemblages collected and evaluated by A. Vörös in the basal part of the overlying basin sediments (Buchenstein Formation), the shallow marine carbonate accumulation was interrupted in the Middle Illyrian (Trinodosus Zone, Trinodosus Subzone) in the Szentantalfa region, while at Vászoly only somewhat later (Trinodosus Zone, Camunus Subzone, Vörös 1990, Budai and Vörös this volume). On the contrary, in the margin of the "Veszprém platform" the first ammonites occurring in the sequence above the algal dolomite indicate the Late Illyrian ("Polymorphus Zone", Vörös 1991);

– comparing the sections above it becomes obvious that the vertical facies change is less sharp above dolomites than in the case of the Tagyon Limestone (see also Fig. 3);

– the proportion of intertidal B-member with algal mat and birdseye structure increases upwards in the studied outcrops (this tendency can be observed also in the borehole sequences, see also Fig. 3), thus the conclusion can be drawn that the sedimentary environment became shallower in time (shallowing upward);

– synsedimentary tectonics (neptunian dyke) is known only in the section at Szentkirályszabadja.

These facts suggest, that the appearance of basin sediments above the shallow marine carbonates, i.e. the drowning of platforms happened gradually later in time from southwest to the northeast, however this can be virtual as well. The fact is that we have only one section, where the oldest ammonite-bearing strata overlie directly the Tagyon Limestone (at Szentantalfa). In the other two sections no valuable fossils were found in the lowermost 1–2 m thick part of the Buchenstein Formation.

Based on these observations and on data from the Southern Alps, the drowning of carbonate platforms in the Balaton Highland can be interpreted as follows:

– The sudden interruption of shallow marine carbonate formation in the Late Anisian has been interpreted recently as a sequence boundary in the Southern Alps (Doglioni et al. 1990 Figs 9 and 17; De Zanche et al. 1992 Fig. 4).

In our opinion, this event can not be explained only by eustatic sea-level changes, since in wide areas of the Southern Alps the formation of platform carbonates continued (Sciliar Dolomite, Esino Limestone, etc.).

– At the same time synsedimentary extensional tectonics has to be taken into consideration in the Balaton Highland. This tectonic phase is evidenced by the neptunian dyke in the Upper Anisian algal dolomite at Szentkirályszabadja. The red, biomicritic, fissure filling material came from a still unspecified level of the overlying Buchenstein beds but the extensional movement can be dated as latest Anisian to earliest Ladinian (see Vörös 1991).

In the study area Middle Triassic platform carbonates younger than Late Anisian are unknown. The age of the dolomite overlying the Buchenstein or the Füred Limestone Formation (see Fig. 1) as a tongue of the Budaörs Platform prograding from northeast, is surely Carnian. Based on the previous geological mapping and on borehole sequences (Raincsák 1980; Budai and Vörös this volume), the interruption of shallow marine carbonate formation by the Ladinian volcanism can be traced along the whole strike of the Bakony Mountains s.l.

Conclusions

The following main phases of Middle Triassic shallow marine carbonate formation can be distinguished in the Balaton Highland:

Early Anisian

The accumulation of terrigeneous clastics in the Early Triassic basin resulted in a wide tidal flat. The Aszófő Dolomite was formed in this environment and dolomitized under sabkha conditions.

Middle Anisian

Due to the shallowing of the Early Anisian inner shelf basin (Iszkahegy Limestone), a Middle Anisian carbonate ramp, the environment of the Megyehegy Dolomite was developed.

Late Anisian

After the subsidence of the Middle Triassic basins and contemporaneously with the deposition of the Felsőörs Limestone, the formation of shallow marine carbonates continued in certain elevated regions of the Balaton Highland. The platform of Tagyon Limestone developed from the subtidal carbonate ramp of the Megyehegy Dolomite by the Late Anisian.

Ladinian

In the latest Anisian the deposition of shallow marine carbonates stopped in the Balaton Highland s. str. The next platform carbonates occur only in the Early Carnian as tongues of the Budaörs platform prograding from the northeast.

Acknowledgements

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Plate I

Photomicrographs illustrating microfacies types of the Aszófő Dolomite Formation. 1. Micrite mudstone with ostracods, Balatonfüred Bdt. 2. borehole, 195.3 m; 2. Oncosparite, grainstone, Balatonfüred Bdt. 2. borehole, 218.0 m; 3. Homogeneous micrite mudstone with a microlayer built-up of quartz and mica grains of silt-size, Balatonfüred Bdt. 2. borehole, 214.3 m; 4. Homogeneous microsparitized micrite mudstone with frequent lath-shaped carbonate pseudomorphs after gypsum, Balatonfüred Bdt. 2. borehole, 212.3 m

Plate II

Photomicrographs illustrating microfacies types of the Iszkahegy Limestone Formation (1-2) and the Megyehegy Dolomite Formation (3–4). 1. Molluscan-echinodermal wackestone with microsparitized micritic groundmass, Balatonfüred Bdt. 2. borehole, 77.1 m; 2. Foraminiferal-molluscan micrite, wackestone; the original micritic groundmass has recrystallized to very fine-fine grained sparite, Balatonfüred Bdt. 2. borehole, 119.0 m; 3. Dolosparite. The original depositional texture has completely been obliterated, Bakonyszűcs Bsz. 3. borehole, 486.8–488.5 m; 4. Dasycladacean biomicrite, packstone. The majority of dasycladacean skeletal elements has recrystallized, Bakonyszűcs Bsz. 3. borehole, 413.4–415.0 m

Plate III

Photomicrographs illustrating microfacies types of intertidal or shallow subtidal sediments of the Tagyon Limestone Formation. 1–3. Oncosparite, grainstone, with a few benthonic foraminifers and molluscan and dasycladacean skeletal fragments. Oncoids exhibit various size and contain frequent dolomite rhombs (photomicrograph 1). Recrystallized dasycladaceans in the middle and the right side of photomicrograph 3, Dörgicse Drt. 1. borehole, 1: 100.0 m, 2: 123.3 m, 3: 138.6 m; 4. Mainly homogeneous algal mat containing peloids, a few fossils (foraminifers and ostracods) and many desiccation (desiccations/hrinkage cracks/pores) pores. Some pores exhibit geopetal infillings, Dörgicse Drt. 1. borehole, 97.5 m

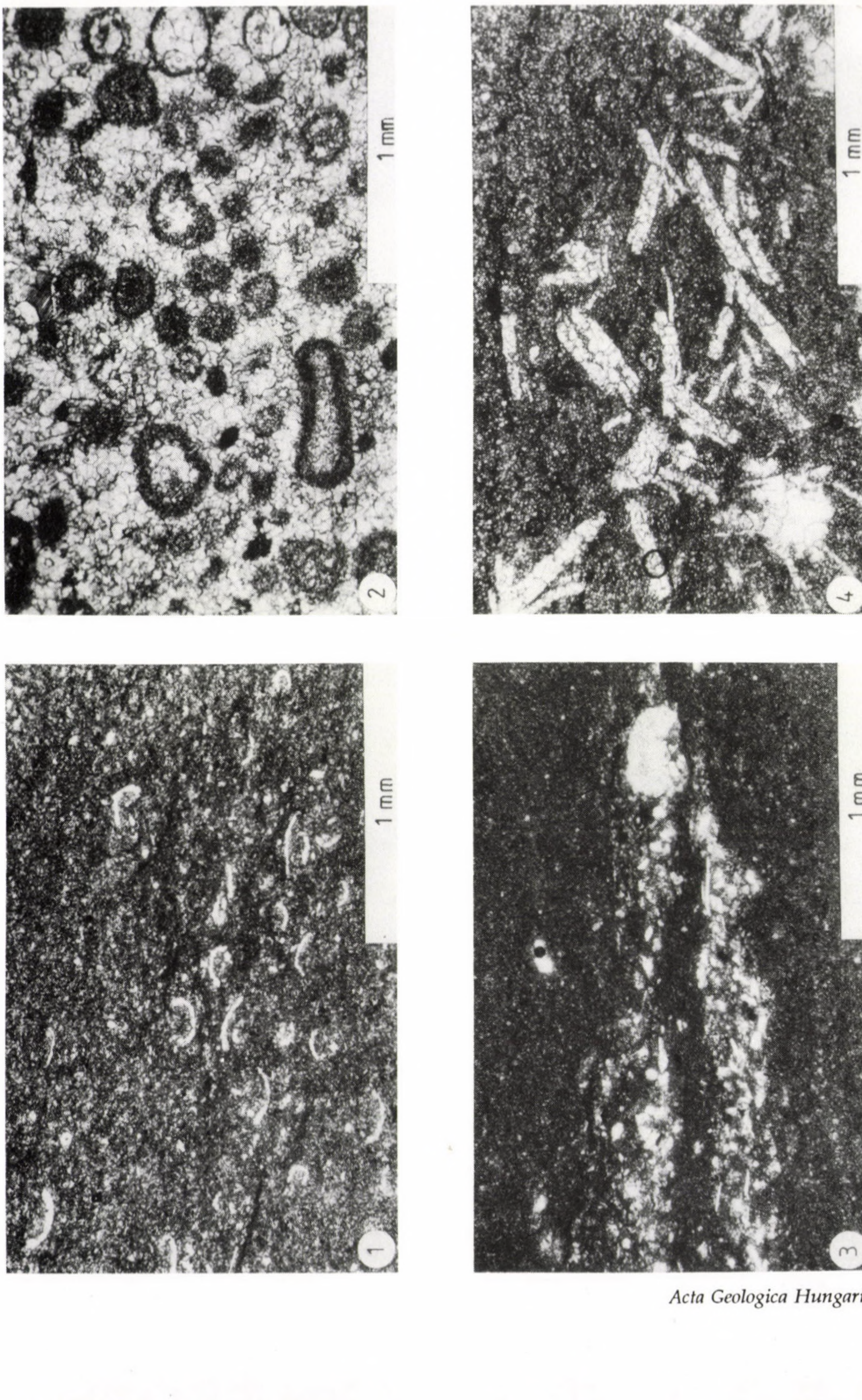
Plate IV

Photomicrographs illustrating vadose-pedogenic carbonates from the Tagyon Limestone Formation. 1–2. Pisoidal-peloidal caliches. Note the difference in grain-size and the special fitting of the grains! Centripetally oriented elongate carbonate crystals on the surface of the grains, whereas the remaining pore-space is filled with mosaic sparite (not visible in the photomicrographs), Dörgicse Drt. 1. borehole, 1: 124.4 m; 2: 125.0 m; 3–4. Vadose carbonate crusts (photomicrograph 3) and brecciated fragments of crusts of similar origin (photomicrograph 4) in pedogenic carbonates, Dörgicse Drt. 1. borehole, 3: 102.4 m; 4: 115.4 m

Plate V

1. *Physoporella pauciforata pauciforata* Bystricky 1964. Szentantalfa Szaf. 1. borehole, 146,.3 m 31x;
2. *Physoporella pauciforata pauciforata* Bystricky 1964. Szentantalfa Szaf. 1. borehole, 90.5 m 31x;
3. *Physoporella pauciforata undulata* Bystricky 1964. Szentantalfa Szaf. 1. borehole, 178.5 m 31x;
4. *Physoporella pauciforata undulata* Bystricky 1964. Dörgicse Drt. 1. borehole, 93.9 m 31x

Plate I



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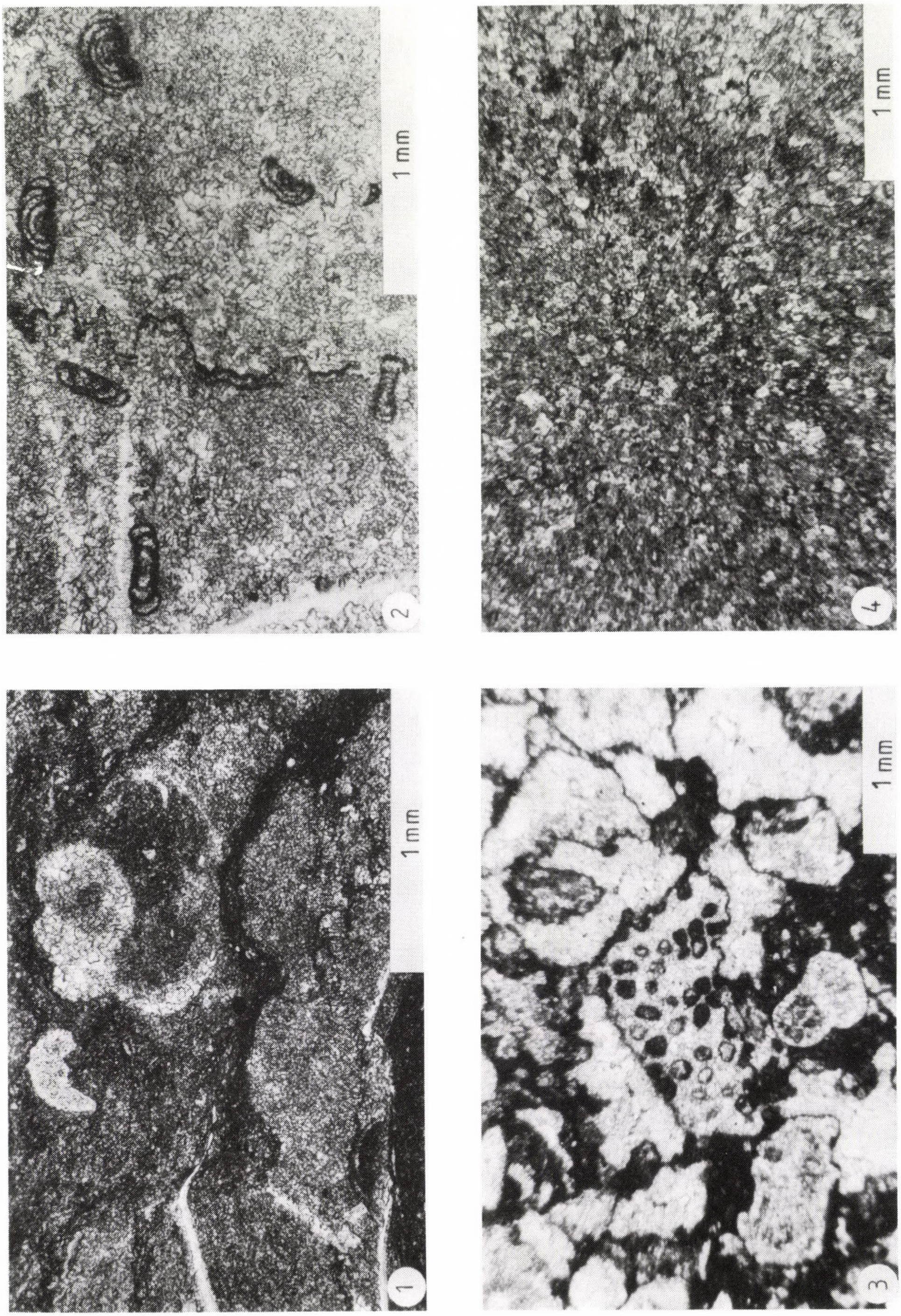
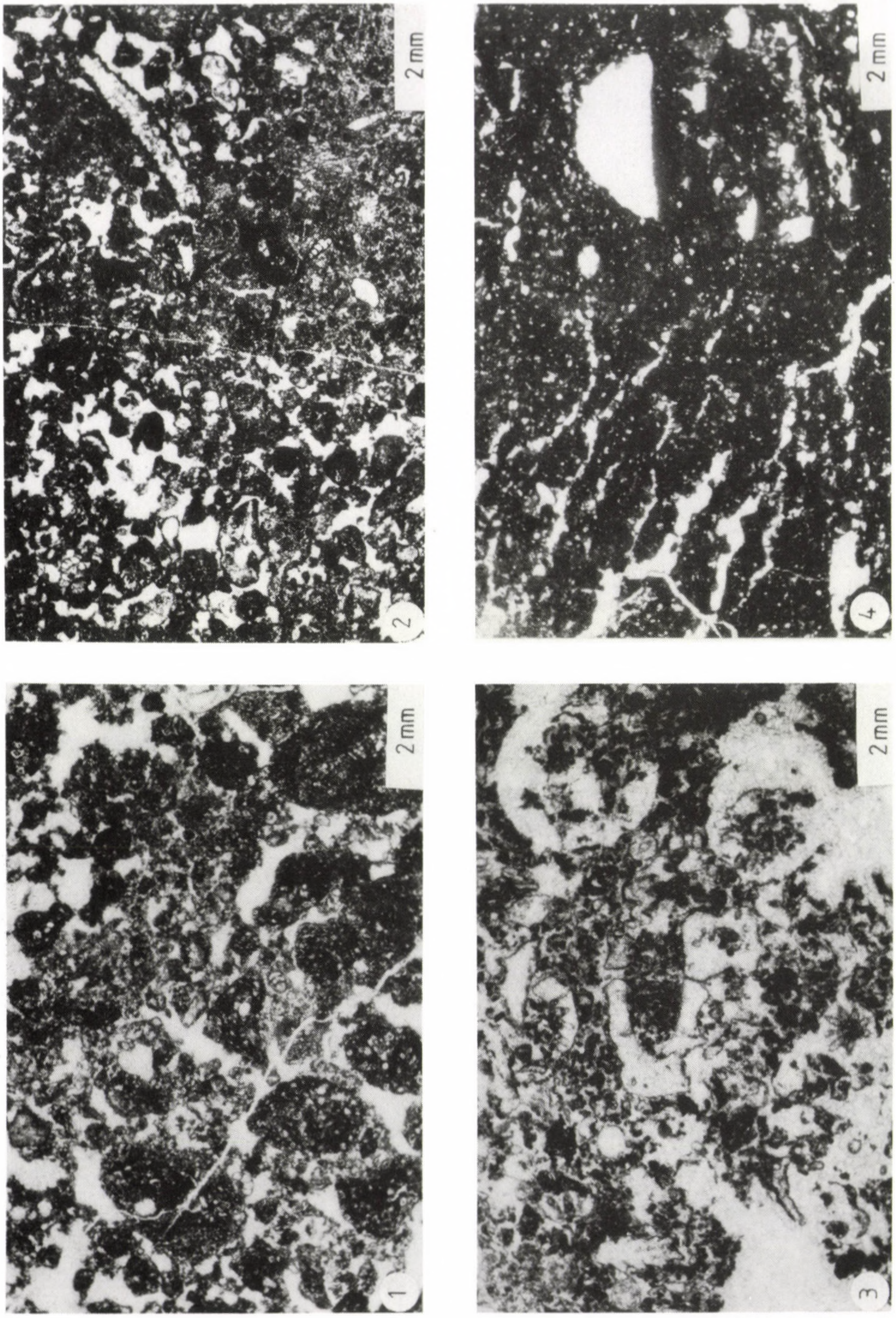


Plate III



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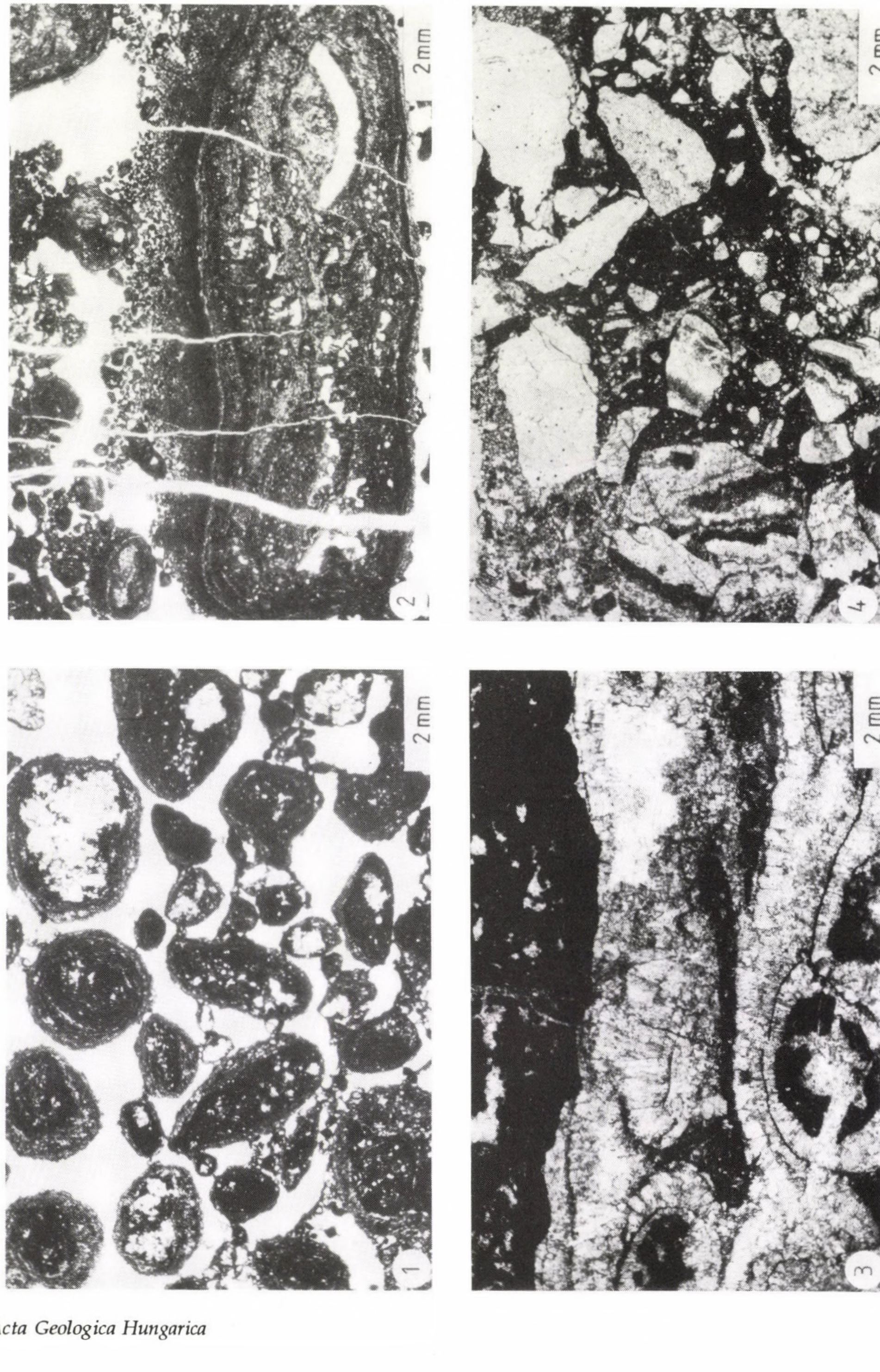

Plate V

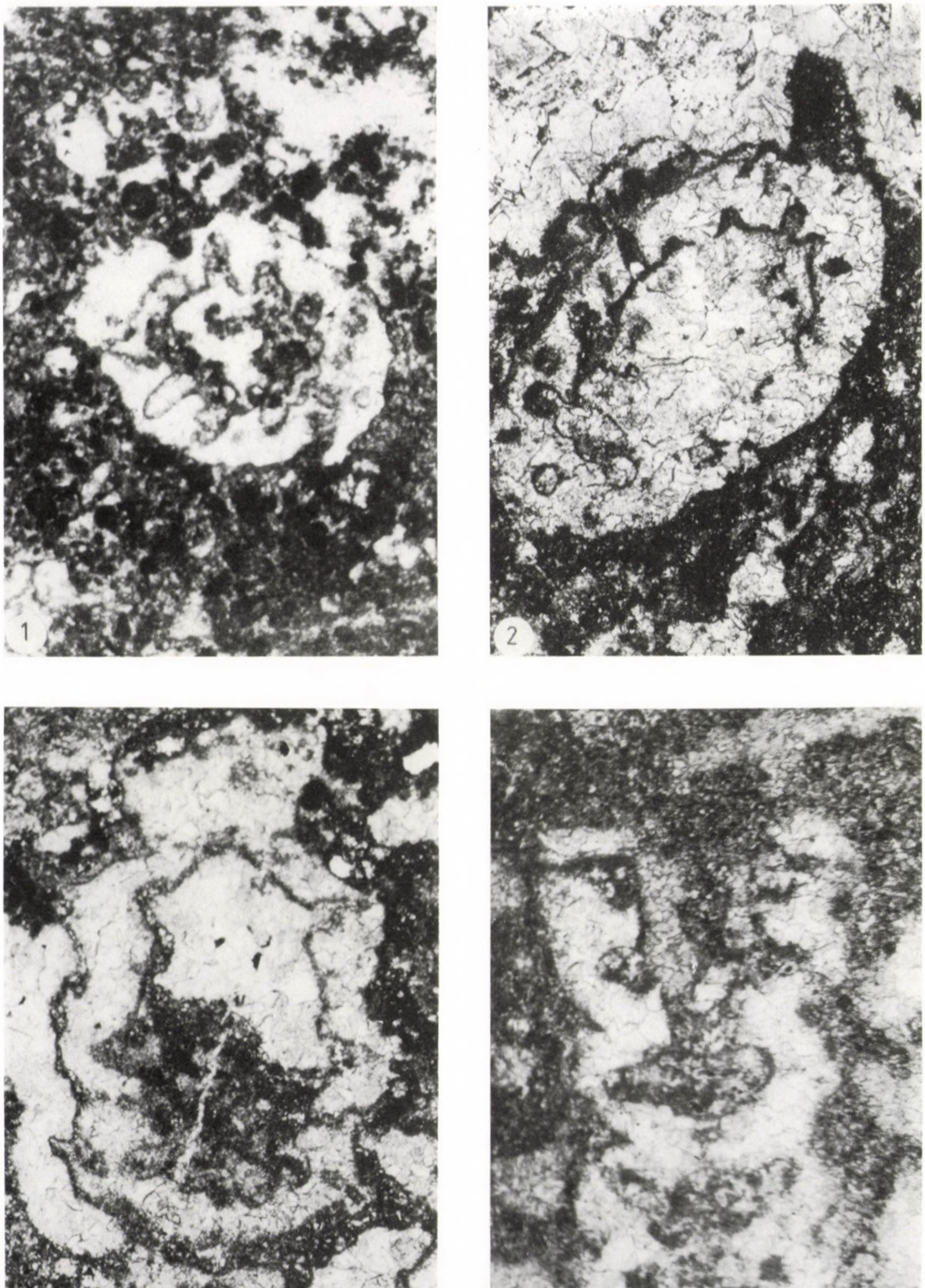

Plate VI

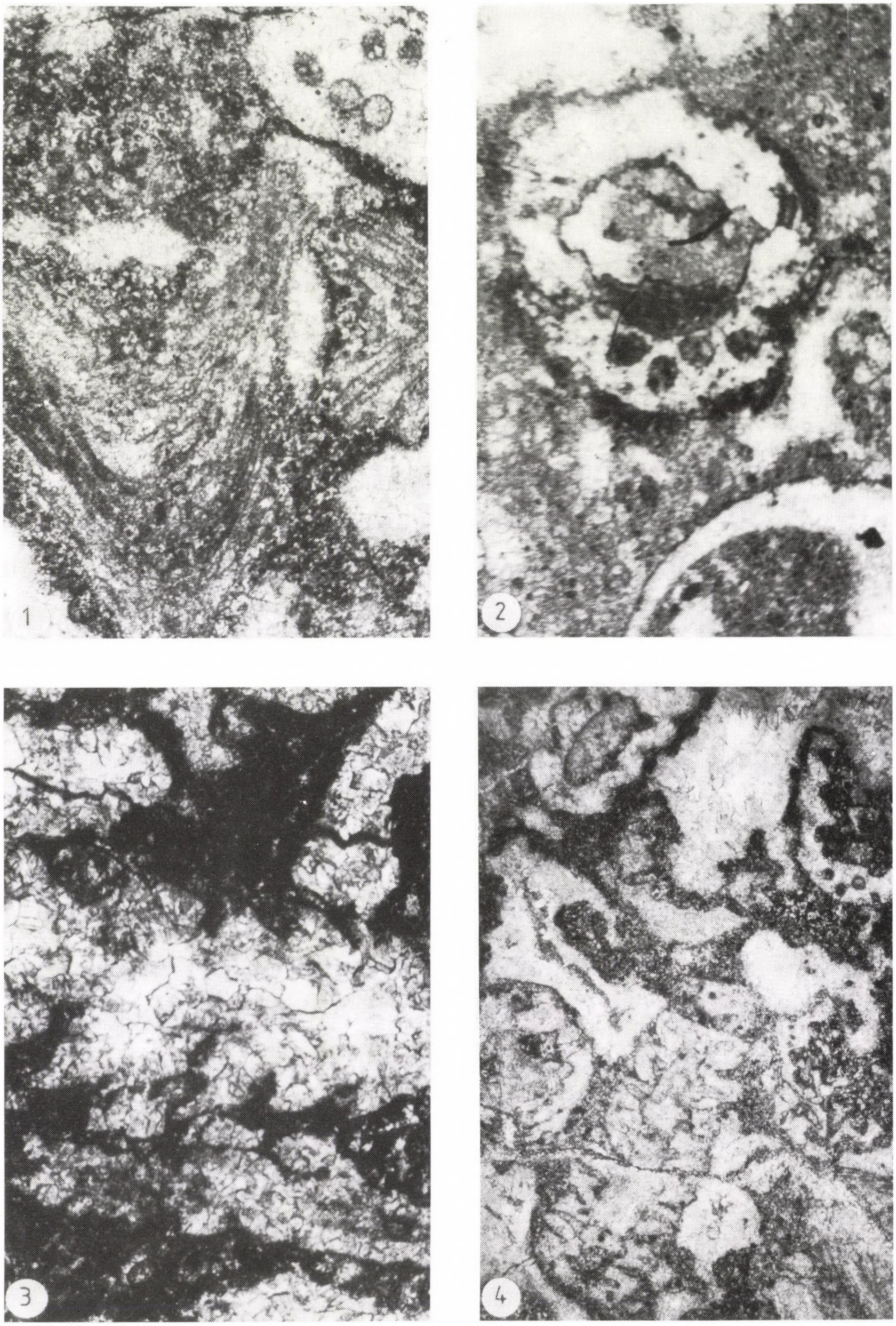

Plate VI

1. *Teutloporella peniculiformis* Ott 1963. Dörgicse Drt.1. borehole, 93.9 m 31x; 2.*Oligoporella pilosa* Pia 1935. Szentantalfa Szaf. 1. borehole, 90.5 m 31x; 3. *Physoporella pauciforata sulcata* Bystricky 1962. Szentantalfa Szaf. 1. borehole, 178.5 m 31x: 4. Limestone with Physoporella. Dörgicse Drt. 1. borehole, 93.9 m 12.5x

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Book reviews

Chatalov, G. 1990: Geology of the Strandzha Zone in Bulgaria. (In Bulgarian, with English summary).

Geol. Balcanica, ser. oper. sing., 4, 263 p., 44 figs., 41 pls., Publ. House Bulg. Acad. Sci., Sofia

This very important contribution to the geology of the western segment of "Anatolian Cimmerides" or "Rhodope–Pontide fragment" (in sense of Sengör 1984) contains Dr. Chatalov's more than 30 years work in the Strandzha Zone of SE Bulgaria.

Following a short historical review, the first main chapter deals with stratigraphy in general, including the available biostratigraphic data: Precambrian, Devonian, Upper Paleozoic, then in details the Triassic and Jurassic (from "Scythian" to Bathonian); finally, a brief review of Upper Cretaceous (Cenomanian to Senonian) rocks and of the Tertiary–Quarternary posttectonic cover is added.

The second main chapter deals in details with the lithology and petrography of Precambrian to Jurassic rocks, and to a certain extent also with that of the Upper Cretaceous ones. The Precambrian includes medium to high temperature metamorphosed sedimentary and magmatic rocks, which had been affected two successive regional metamorphic events (pre-migmatization, then migmatization), divided by a retrograde phase accompanied with by diaphtoresis. The Lower Paleozoic is attributed to the Devonian on the basis of faunistic proofs from marbles. It includes a "phyllite–diabase formation", with some marble intercalations, (Sokol Fm.) and a phyllite-marble formation. Metabasic rocks of the former are of mid-oceanic ridge basalts type. An acidic metavolcanic formation is also assigned to the Lower Paleozoic. The Upper Paleozoic (Yambol Group) includes continental coarse detrital rocks and acidic metavolcanics, of assumed Stephanian–Permian age.

The so-called South Bulgarian granitoids are assigned to the Late Permian.

The Triassic is divided into three tecto-facial types, from which only the first two lies over the known Precambrian–Paleozoic rocks. The Sub-Balkanide Triassic is related to the Balkanide-type and is considered to have deposited on the southern part of the "Moesian shelf". The amphibolite-facies metamorphosed Sakar Triassic begins with continental, then shallow marine Lower Triassic, overlain by a thick carbonate platform attributed to the Middle Triassic. Of special interest is, that inspite of its medium grade metamorphism, fossils were found in it: "Myophoria" (=Costatoria) costata, the ribs of which are expressed by hornblende crystals and in the same rock (quartzamphibolite) staurolite crystals formed between the fossils (shown on pl. 12 and pl. 25), furthermore strongly recrystallized conodonts in crinoidal marbles (shown on pl. 20). The Strandzha Triassic of highest tectonic position lacking Paleozoic basement, can be divided into two parts, separated by a metamorphic disc- ordance. Different formations of great thickness, consisting of phyllites, metagraywackes, meta- arcoses, associated metabasalts and meta- keratophyres of island-arc type and marbles, are assigned to the Lower Triassic (though being structurally above the younger Triassic carbonates). These partly turbiditic sediments accumulated in an active tectonic regime, probably within an island-arc system. Only the youngest(?) formation of this lower group. (Gramatikovo Formation, consisting of banded calcite phillites with some basic and acidic metavolcanics, associated with economically important sulphide are mineralization) is fixed biostratigraphically (conodonts, foraminifers) as Spathian–? Aegean.

The Middle–Upper Triassic (which is in the structurally lower position) consists of the heteropic Anisian to Lower Carnian Malko Tarnovo Limestone of carbonate platform facies and the basinal, conodont-bearing Kondolovo Limestone, followed by the flysch-type Upper Carnian–Norian Lipacka Formation, which indicates already the Early Cimmerian tectonism. (The summary of the most important Triassic conodont and foraminifer data are given in a separate paper by K. Budurov and E. Trifanova, published in English in the Revies os Bulgarian

Geological Society from 1991, Vol. 12/3, pp. 3–18, pls. 1–8.).

The Jurassic is known only in the overlier of the Subbalkanide Triassic of structurally lowest position. It is represented by Hettangian to Bathonian siliciclastic and calcareous sediments, with iron ore horizons, deposited in a shallow to moderately deep epicontinental sea.

The Upper Cretaceous shows two transgressive cycles (with swamp deposits and coal at their base), interrupted by a regression at the end of the Cenomanian. Further deepening lead to deposition of flysch-like sediments in the Turonian–Lower Senonian. Volcano-sedimentary (partly olistostromal) and volcanic rocks (tholeiitic and subalkaline) characterizes the Campanian.

The third main chapter is devoted to metamorphism, one of the author's main research fields. After a short review about the boundaries of zones from diagenesis to medium temperature metamorphism, features of local metamorphism are discussed. The main part of the chapter deals with regional dinamothermal metamorphism, which affected all pre-Cretaceous rocks, from very low grade to medium grade, including a table about illite crystallinity indices of Paleozoic, Triassic and Jurassic formations of the Strandzhadide and Balkanide–Prebalkanide units. The medium to high temperature metamorphosed Precambrian units are separately discussed.

The last chapter deals with the structure of the Strandzha Zone and its setting in the Alpine orogeny of Bulgaria. The structure of the Strandza Mts is the best shown in the Sveti Ilija Hights; according to this (shown on the block-model of Fig. 43), the first allochton (Zabernovo Nappe) was formed when the Strandzha-type Triassic thrust onto the Jurassic cover of the Subbalkanide Triassic in post-Bathonian times, and the second one (East Thracian Nappe) when they thrust together onto the Upper Cretaceous of the foreland in the Late Senonian. This chapter is illustrated with a geological sketch map the Strandzha Zone in scale 1:250.000, showing 48 formations (Fig. 40).

Finally, an English summary is given on eight pages, containing the most important informations.

The book is richly illustated with 43 text-figures (stratigraphic columns, geological sections, etc.) and with 41 photo plates. The latters show different types of metamorphic textures of Precambrian to Middle Jurassic siliciclastic and carbonate sedimentary, and magmatic rocks, respectively, from nonaltered textural types (in part of the Subbalkanide Triassic and Jurassic) to medium-grade metamorphosed ones (for example, in the Sakar Triassic). Of special interest are a number of plates showing types of textural alterations in metacarbonates, which are, unfortunately, usually already beyond the interest of sedimentologists, but are less attractive for metamorphic petrographers, too.

This work contains a lot of valuable informations for geologists interested in the extension of the Cimmeride system into Europe, and in the Paleotethys/Neotethys problem. A publication of more details or even of the whole book in English would certainly be of great interest for many researchers in Europe.

Sándor Kovács

Jurkovsek, B., Kolar-Jurkovsek, T. Fossili v Sloveniji (Fossils in Slovenia; in Slovenian).

Didakta, Ljubljana, 1992,. 71 p., 6 text-figs., 52 photos,

This nice, picturesque booklet, written in popular scientific language, shows examples of most of the important megafossil groups lven, a few large forams are included, too), which can be found in Slovenia, from Late Carboniferous to Miocene. A few one from the Quaternary are also added. The fossils, shown on colour photos of high quality, belong to bivalves, brachiopods, ammonoids, corals, sponges, echinoderms, fishes, plant remnants and other groups. They have been collected by the authors, Tea and Bogdan themselves and are stored in their private museum of "Fossils in Slovenia". Most of the photos and all figures were made by Bogdan.

A short explanatory text is written for those, who are interested in nature and in collecting fossils about fossilization, history of life and preparation of them. This is followed by a brief information on the most important plant and animal fossil groups.

This beautifully illustrated booklet would be of interest, however, not only for lovers of nature, but also for geologists and paleontologists to get an impression about the most important fossil groups occurring in the territory of Slovenia and are observable with naked eye.

Sándor Kovács

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GUIDELINES FOR AUTHORS

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Relations between Rock-Eval and cutting gas composition data

István Koncz, Emilia Horváth
Hungarian Oil and Gas Corporation (MOL Plc)

Rock-Eval and gas analyses from cuttings were performed in three boreholes, situated in the southeastern part of Hungary. In each borehole, source rocks located in the lower part of Lower Pannonian formation were indicated by hydrocarbon potential of cutting samples. The investigated source rock sections presently are in the oil window, their vitrinite reflectance values range from 0.6 to 0.9%. Oil production has occurred from the source rock sections and/or carrier beds bordering source rock section. The cutting gas composition data indicate the presence of source rock. The concentrations of the heavier gas components (propane, butane, C_{5+}) increase in the source rock section and the adjacent carrier beds due to the hydrocarbon migration. In the source rock section, the ratio of normal-butane to iso-butane seems to be related to the hydrogen index, to the quality of kerogen. There is no relation between the quality of kerogen and the concentration of heavier components in cutting gases.

Key words: source rock geochemistry, Rock-Eval, cutting gas

Results and discussion

Rock-Eval and gas analyses from cuttings were performed in three boreholes, situated in the southeastern part of Hungary. The wells have reached depths of 3400 to 5300 m. The pre-Neogene basement consists of Paleogene and Cretaceous sediments in well-named "A", and Cretaceous–Jurassic rocks in well named "C" (Fig. 1). In well-called "B", the drilling was finished in the Lower Pannonian (Late Miocene) sediments. The lithology, the source rock sections and the vitrinite reflectance values are shown in Fig. 1. The source rock sections are characterized by total organic carbon content of 0.5% and hydrocarbon potential of 1 mg/g at least.

The vitrinite reflectance of 0.6%, corresponding approximately to the start of the main pulse of oil generation, occur at a depth of 2200 m in the Upper Pannonian (Pliocene) sediments in case of well "A", 2400 and 2800 m in the Lower Pannonian rocks in case of wells "C" and "B", respectively. In well "C", the unconformity at the Neogene–Mesozoic boundary resulted in a "jump" in the maturation profile. The projection of the Neogene maturation profile at the base of the Neogene section provides a vitrinite reflectance of 0.68%. The projection of the Mesozoic maturation profile at the top of the Mesozoic section provides a vitrinite reflectance of 1.19%. These data indicate that the Upper

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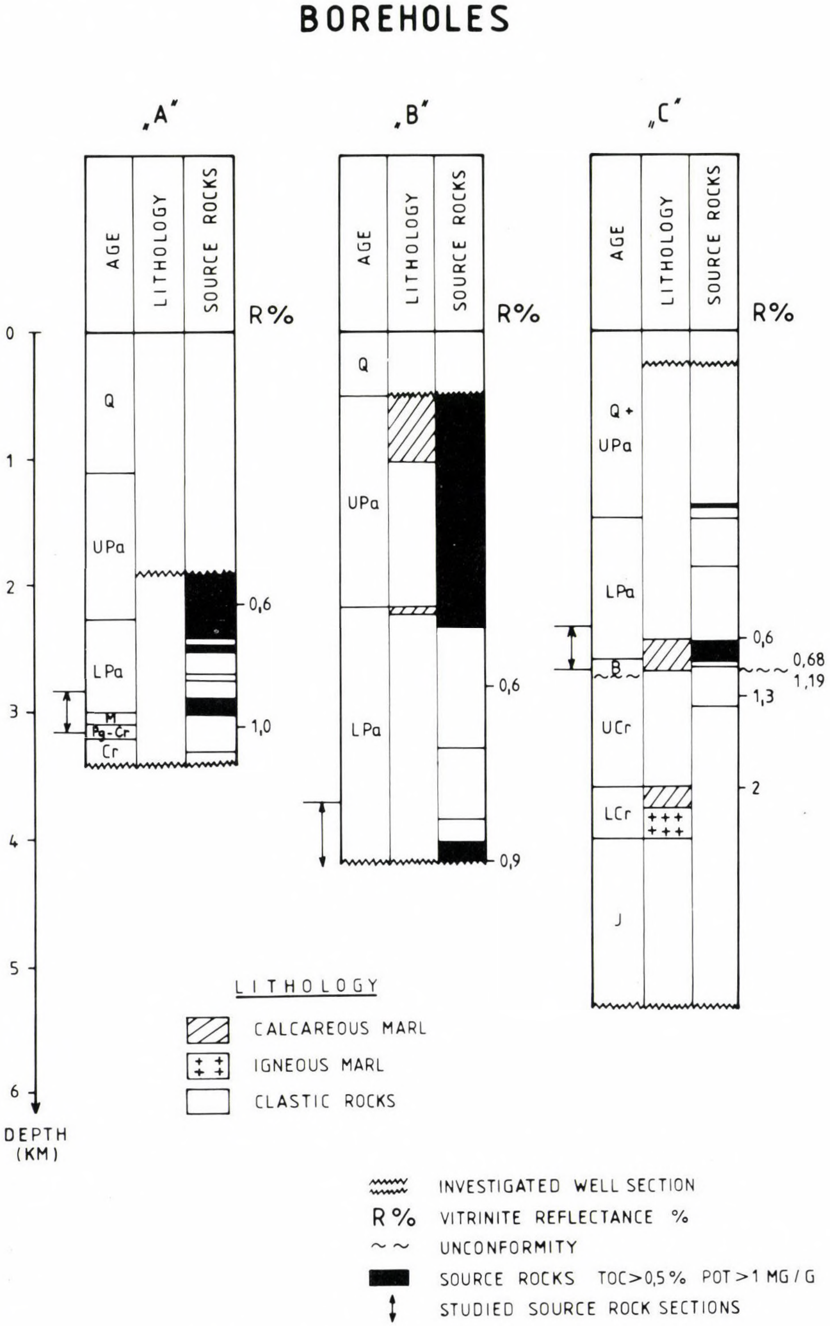

Fig. 1
Age, lithology and thermal maturity of the rocks, and source rock sections in the studied wells. Q – Quaternary; UPa – Upper Pannonian; LPa – Lower Pannonian; M – Middle Miocene; B – Badenian (Middle Miocene); Pg – Paleogene; Cr – Cretaceous; UCr – Upper Cretaceous; LCr – Lower Cretaceous; J – Jurassic

Cretaceous sediment of considerable thickness was eroded prior to Miocene subsidence. Based on total organic carbon content and hydrocarbon potential, source rock sections of significant thickness can be identified in the Upper and Lower Pannonian sequences. Only the Lower Pannonian source rock sections are situated in the oil window. The kerogen of the Upper Pannonian source rocks is thermally immature.

In each borehole, a source rock of Lower Pannonian age was selected in order to establish the relation between Rock-Eval parameters and cutting gas composition data. The selected source rock sections are shown in Fig. 1. The vitrinite reflectance values of the selected source rock sections range from 0.6 to 0.9%. Oil indication or production has occurred from the source rock sections and/or carrier beds bordering source rock section.

Rock-Eval parameters (hydrocarbon potential, hydrogen index) and cutting gas composition data (butanes and propane+butanes in hydrocarbons ranging from methane to butane, C_{5+} fraction in hydrocarbons ranging from methane to decane, ratio of normal-butane to iso-butane) of the selected source rock sections are shown in Figs 2–4. The cutting gas composition data indicate the presence of the source rock and the migration of hydrocarbons from the source rock. The concentrations of the heavier components (propane, butanes, C_{5+}) increase in the source rock sections and even in the adjacent carrier beds because of the primary migration of hydrocarbons. In the source rock sections, the ratio of normal-butane to iso-butane seems to be related to the hydrogen index, to the quality of kerogen (Fig. 5). There is no relation between the quality of kerogen and the concentration of heavier components of the cutting gases in the source rock sections (Figs 6–8).

The ratio of normal-butane to iso-butane has been used as a maturation indicator. The ratio is low in immature sediments, but it increases to a value of 1 to 2.5 corresponding to the onset of oil generation (Alexander et al. 1983). The results presented in this study indicate that the increasing of the ratio of normal-butane to iso-butane, departing from the general tendency is related to the quality of kerogen, the vitrinite reflectance of which ranges from 0.6 to 0.9%.

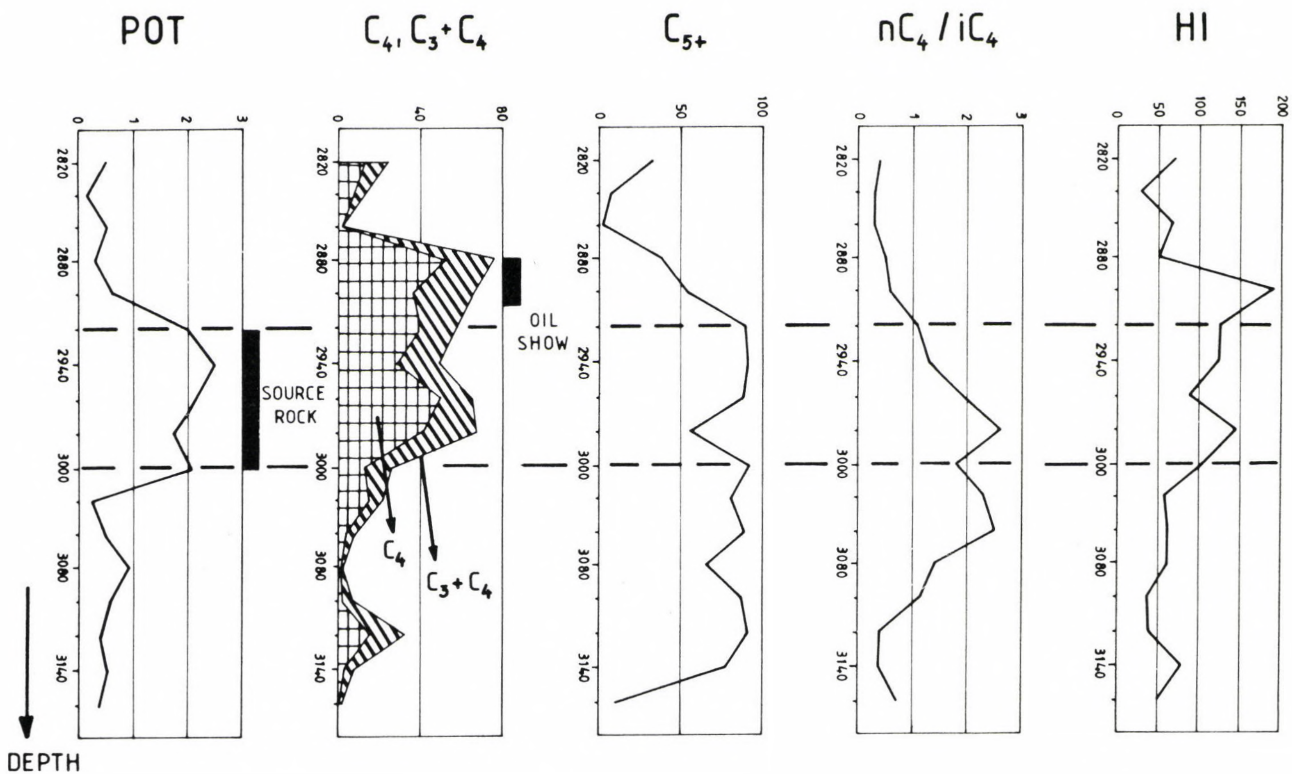

Fig. 2

Rock-Eval and cutting gas composition profiles of the source section in the well "A". POT – hydrocarbon potential; C4 and C_3+C_4 – concentration in C_1–C_4; C_{5+} – concentration in C_1–C_{10}; nC_4/iC_4 – ratio of normal-butane to iso-butane; HI – hydrogen index

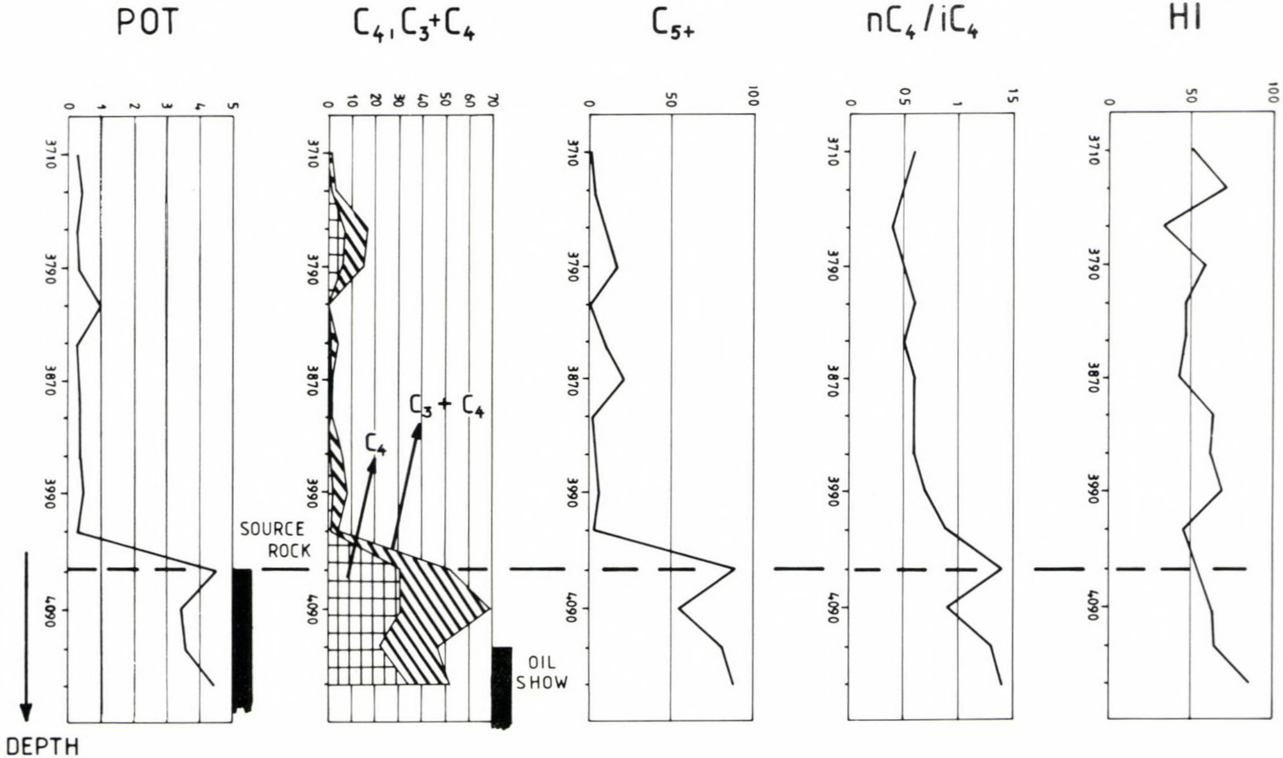

Fig. 3
Rock-Eval and cutting gas composition profiles of the source rock section in the well "B"

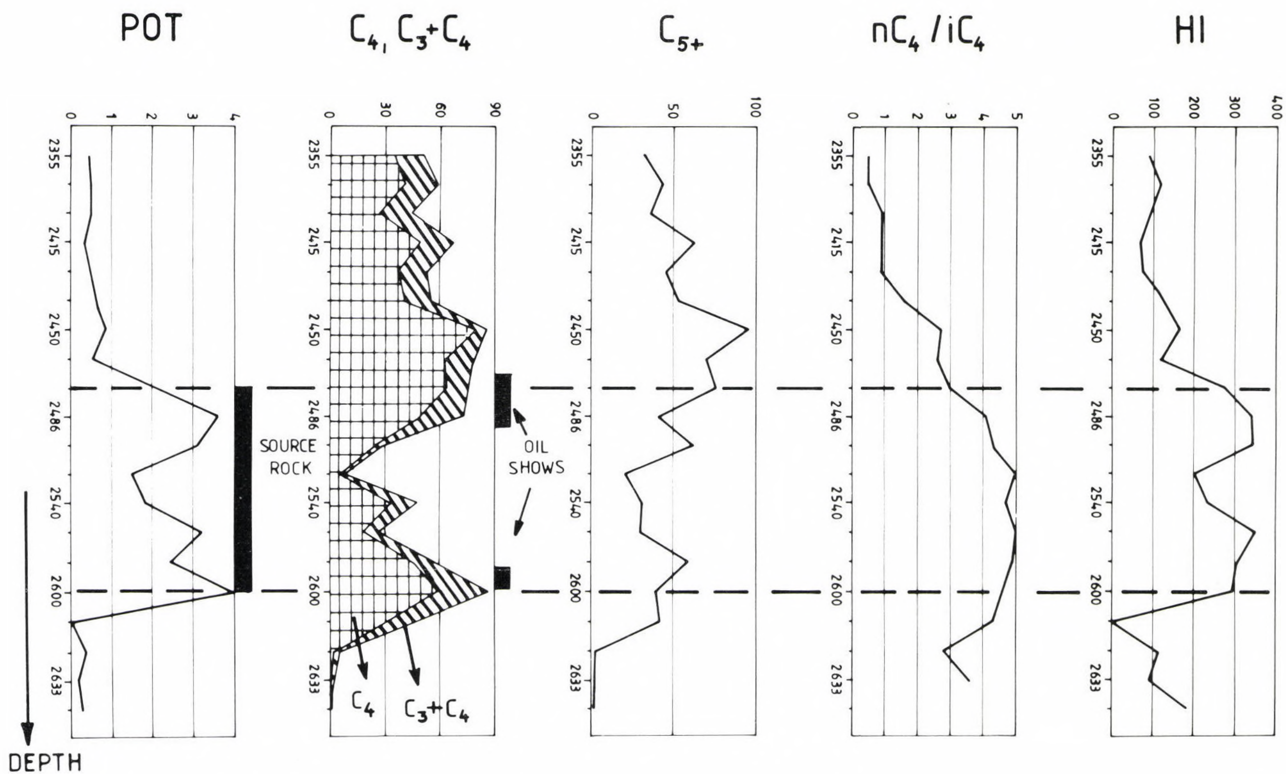

Fig. 4
Rock-Eval and cutting gas composition profiles of the source rock section in the well "C"

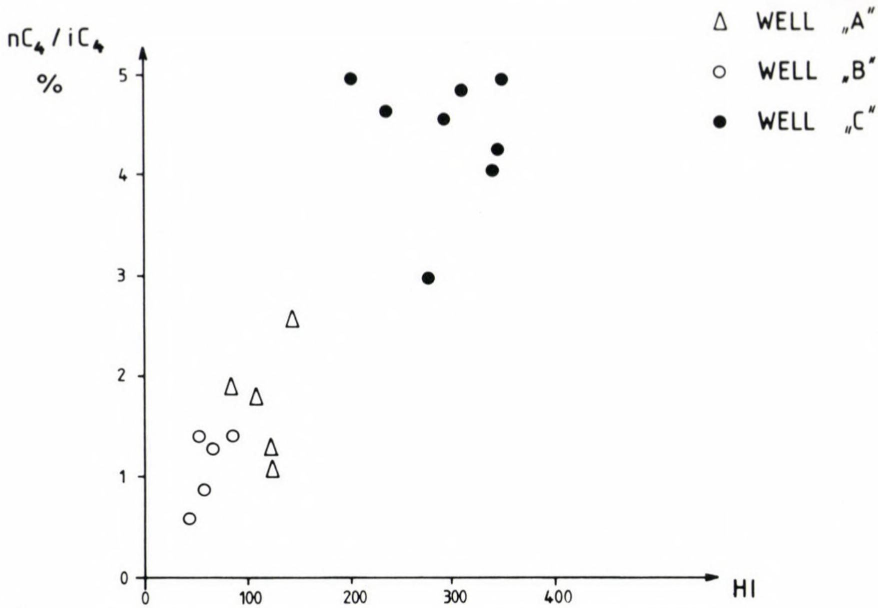

Fig. 5
Relation between ratio of normal-butane to iso-butane and hydrogen index

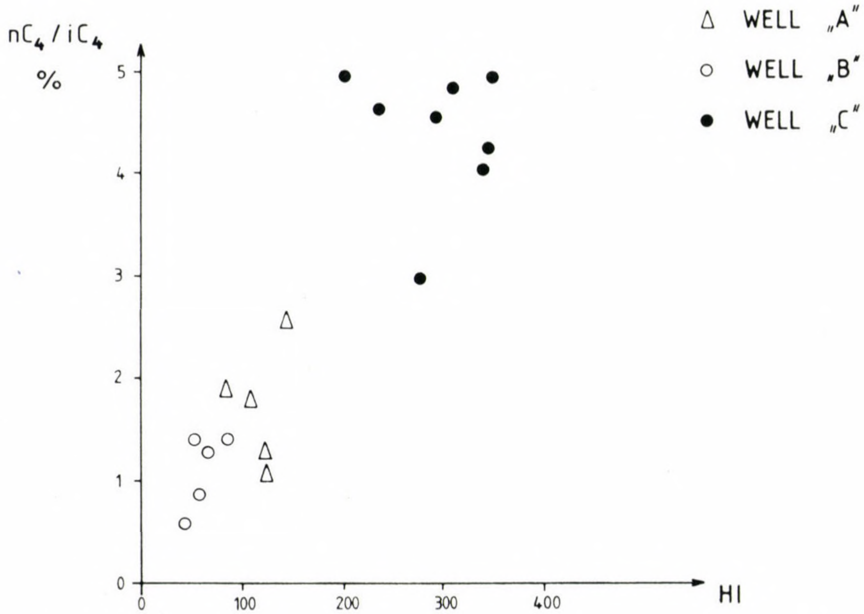

Fig. 6
Relation between butane concentration in hydrocarbons C_1 to C_4 and hydrogen index

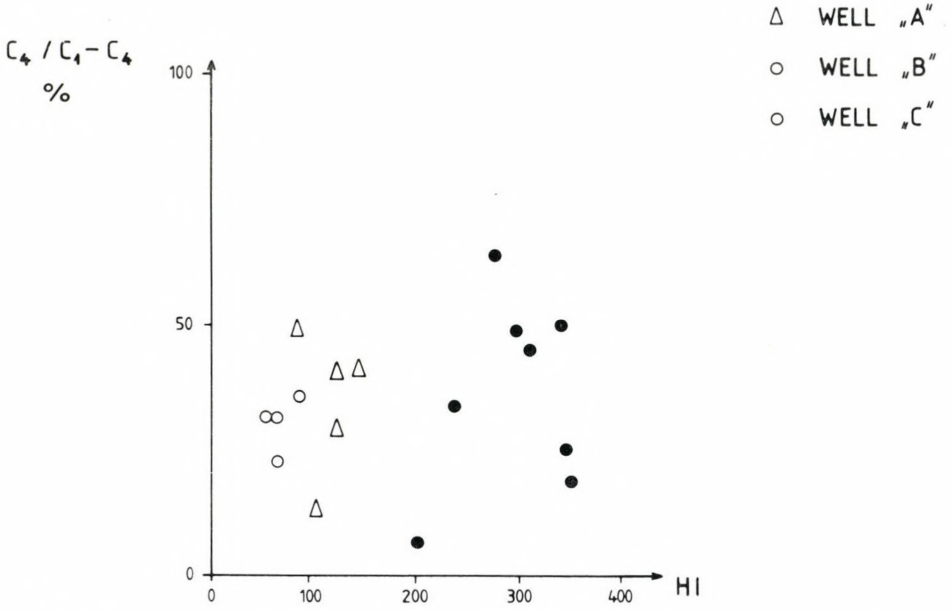

Fig. 7
Relation between C_3+C_4 concentration in hydrocarbons C_1 to C_4 and hydrogen index

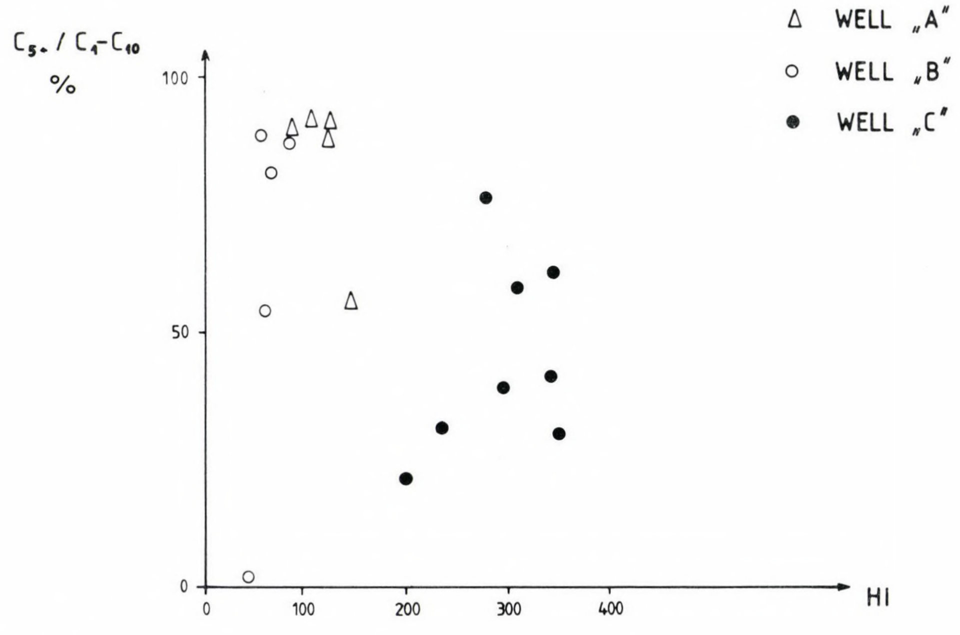

Fig. 8
Relation between C_{5+} concentration in hydrocarbons C_1 to C_{10} and hydrogen index

Acknowledgement

We are grateful to Hungarian Oil and Gas Corporation for allowing us to publish this study.

Reference

Alexander, R., R.I. Kagi, G.W. Woodhouse 1983: Variation in the ratio of isomeric butanes with sediment temperature in Western Australia. – In: Advances in Organic Geochemistry 1981 (Ed. M. Bjoroy et al.), pp. 76–90, John Wiley, Chichester.

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Rock-Eval measurements of two drilled wells in the Southeast Part of Hungary

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The delta plain facies of Pannonian deposits is characteristic of fair quality source rock, though the organic matter is immature. In the proximal part of the delta slope the average T_{max} values are 435–440 °C. Other parameters of the organic matter differ, depending on the different delta slope parts. The high TOC values are probably the results of the ablations of continental organic matter.

The organic geochemical parameters in the distal deposits are higher than in the proximal part, or they increase with depth. In similar samples which are from lower distal turbidite layers, the organic matter is mature, and the rock can be considered of good source quality. The turbidite deposits contain mixed type (II–III) organic matter.

In the open marine marl (calcerous marl), the organic matter is Type II, and it is able to generate oil. In the Ecs-1 well, on the top of the Miocene layer, the clay-marl contains Type–II organic matter, and the hydrocarbon potential is high. In the deeper parts which contain detrital limestone layers, the hydrocarbon potential decreases continuously. In the Sze-2 well, the geochemical parameters change according to the rocks. In the sandstone, mudstone and tuff layers the hydrocarbon potential is lower than in the Pannonian open marine section. On the other hand, the coal layers have very high organic matter content and hydrocarbon potential. The deepest part, which consists of conglomerate, has little organic matter.

Key words: Pannonian deltas, Miocene sediments, Rock-Eval, organic matter, maturity

Introduction

Wells Ecs-1 and Sze-2 were drilled in the Pannonian Basin, in the southeast part of Hungary (Fig. 1). The wells are located on the side of a northeast–southwest line of strike, a narrow Neogene graben, which lies below a depth of 3500 m (Fig. 2). The graben directly joins the Békés Basin, which is approximately 7000 m deep. We examined the quality and maturity of the organic matter (OM) and the hydrocarbon potential (HC–POT) of the rock using Rock-Eval III equipment. The ages of the examined rocks were Pannonian and Miocene. The examined part of well Ecs-1 was between 1500 and 2660 meters. The part of well Sze-2 that was examined was between 1600 and 3100 meters. We sampled cuttings at five meter intervals, and completed these measurements with the examination of some cores. In order to determine the facies of the stratigraphic units and their relationship, we not only performed the geological

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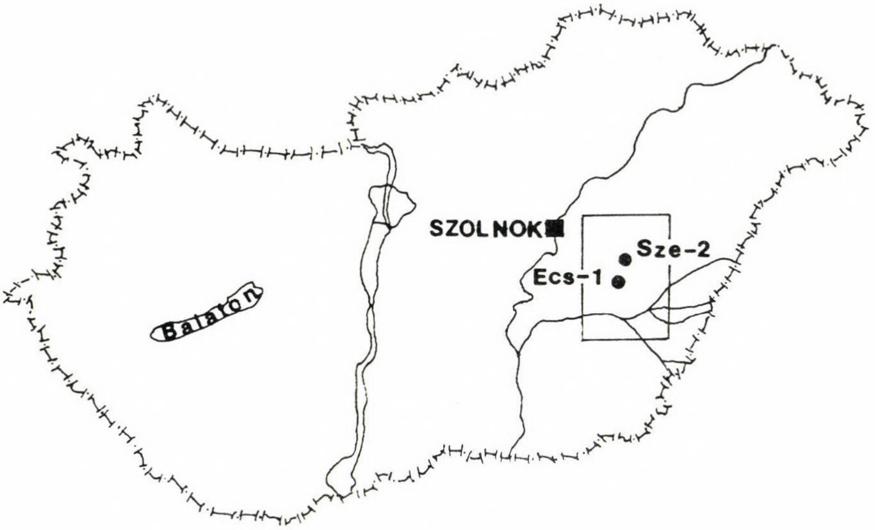

Fig. 1
Location of wells Ecs-1 and Sze-2

testing, but also the geological well log analyses and the sequence analysis of the seismic reflection section. The relation of various facies is demonstrated in Fig. 3. The gas log which was measured by DTA BOX aided the interpretation of geochemical data.

The geological, gas and geochemical logs of the wells are presented in Figs 4 and 5.

Results

The Pannonian deposits which were examined are from the delta slope, the prodelta, and the pelagic areas. The shaly–silty delta slope facies (SL), which contains turbidity and underwater channel fill deposits, was exposed only in well Ecs-1. The rocks in this section of well Ecs-1 are characteristic of fair quality source rock because of their hydrocarbon potential. The organic matter, though, is immature. The upper part of the delta slope, which consists of shale and silt, has five times more HC–POT than the lower part, which consists of psammitic material.

	Upper part 1500–1600 m	Lower part 1600–1800 m
TOC %	1.60–5.25	0.50–1.28
HC–POT mg/g	1.30–6.33	0.41–1.61
T_{max} °C	425–430	424–433
HI	73–113	61–85

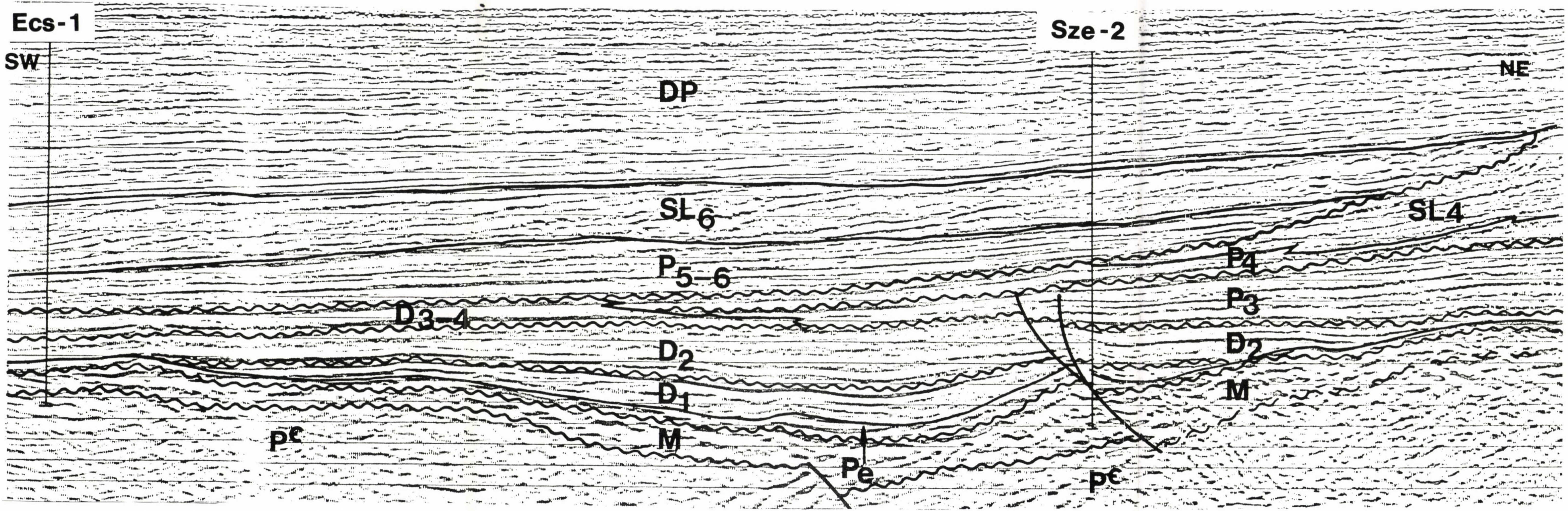

Fig. 3
Reflection section between wells Ecs-1 and Sze-2

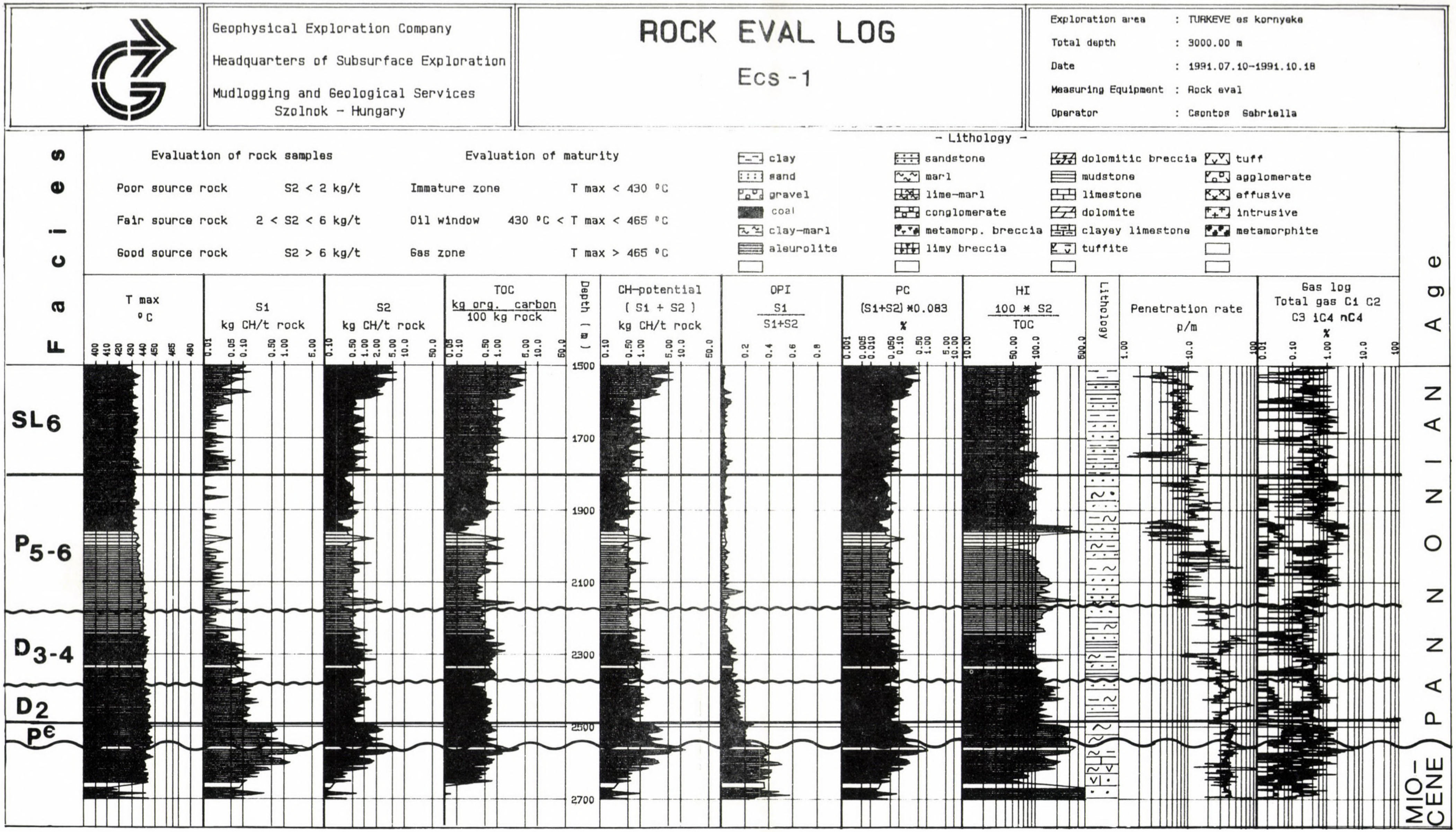

Fig. 4
Geological, gas and geochemical logs of well Ecs-1

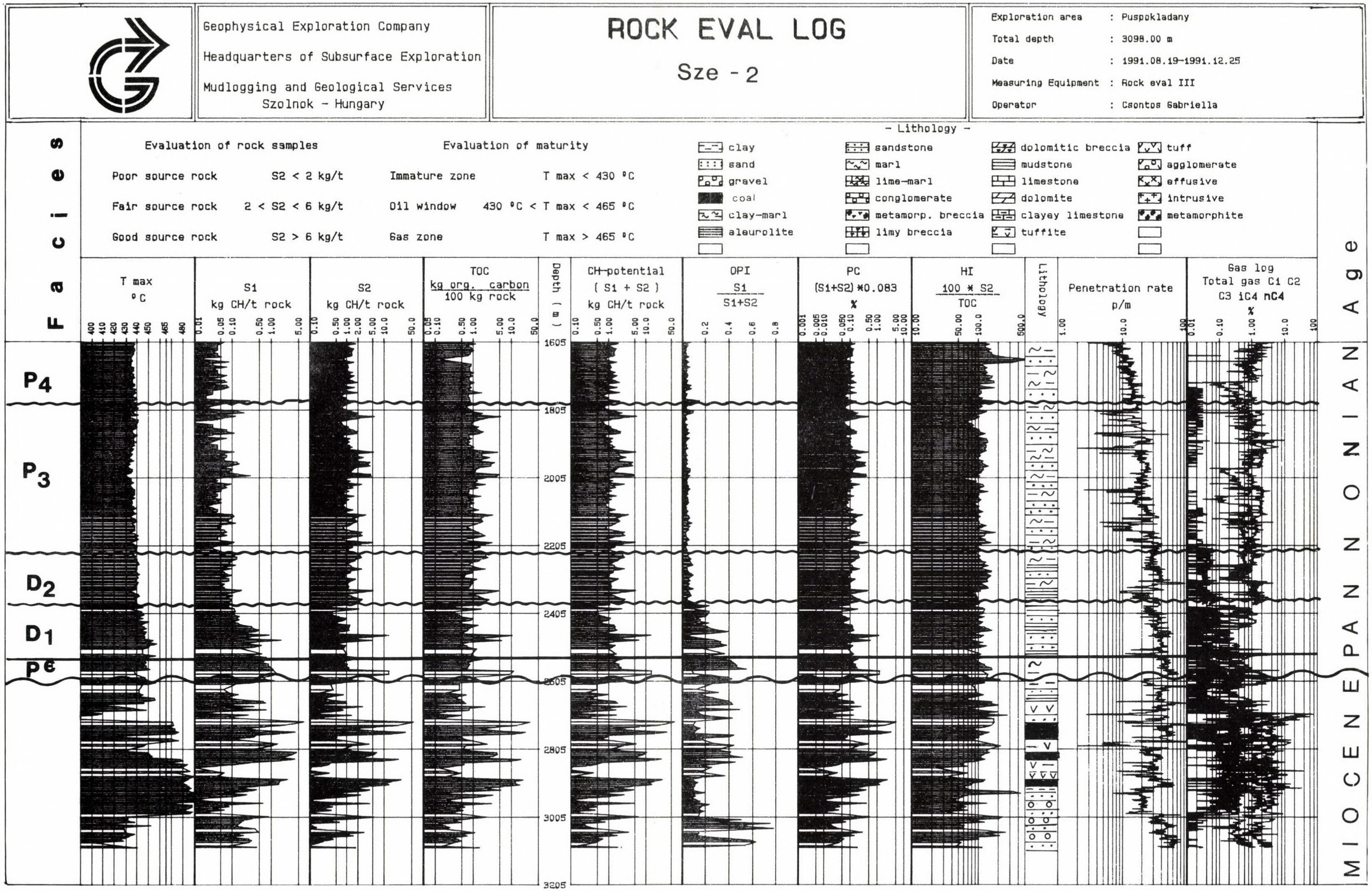

Fig. 5
Geological, gas and geochemical logs of well Sze-2

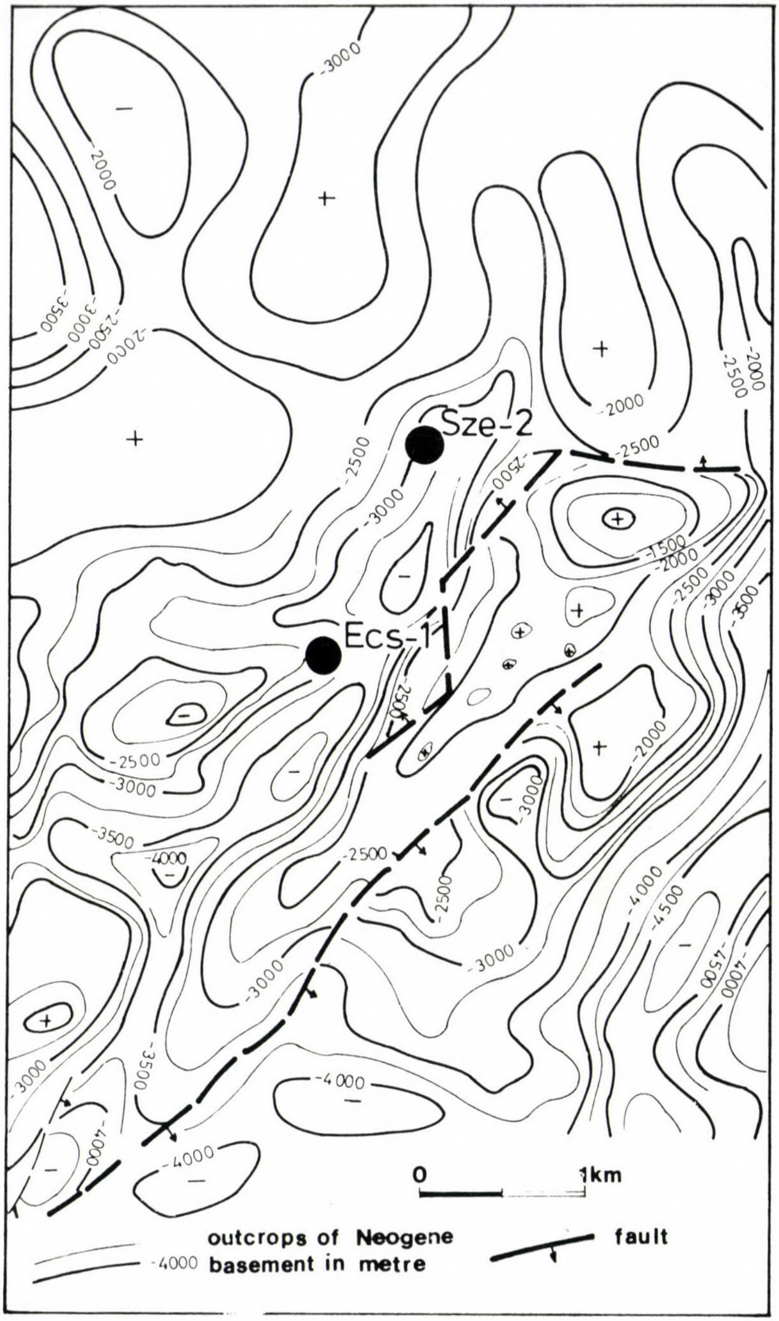

Fig. 2
Map of the Neogene basement around wells Ecs-1 and Sze-2

The prodelta section of the deposits can be divided into proximal and distal parts. The proximal section consists of channel-fill deposits, fans and turbidite sandstones, which are stratified by clay and silt layers. The average T_{max} is the same (435–442 °C) in both wells. The quantities of organic matter in the different parts of the prodelta section which was exposed in the wells differ somewhat. The values, which were measured in the P4 and P3 part of the delta section (in well Sze-2) are approximately about twice as high as those of the P5–6 part of the delta (in well Ecs-1):

	P4	P3	P5–6
TOC %	0.70–1.10	0.7–2.32	0.30–0.60
HC–POT mg/g	0.67–1.90	0.85–2.60	0.20–0.94
T_{max} °C	435–438	435–442	430–436
HI	88–189	80–180	50–140

The high total organic carbon (TOC) values (in the P5–6 part of the delta are 1.13–2.63%; in the part P3 of the delta they are 3.28–5.32%), which probably derive from the ablation of continental organic matter (for example: humic muds of long-lasting floods). The upper part of the proximal turbidite in well Ecs-1 has not only less, but more immature organic matter, which has not produced hydrocarbon yet. The hydrocarbon potential (HC–POT) decreases under 0.5 mg/g, in the turbidite fans, which are situated between 1940–1960 and 2070–2190 m. Three distal turbidites sections were encountered in the wells below the proximal turbidites. Only the middle section (D2) can be found in both drillings. In these sections thin silt, clayey marl and sandstone layers alternate. In these three sections, the organic matter comes from the continent (Type III OM) but somewhere it is of a mixed (II–III) type. The top unit (D3–4), which was found in well Ecs-1, is the furthermost distal facies from the coast of the proximal turbidites (P3–4). This distal turbidite unit (D3–4) is characterized by 442–444 °C T_{max} values and by the first increasing and then permanent values of organic geochemical features (TOC, HC–POT). The trend of the values is the same as in the proximal turbidite (P3–P4.)

In the middle section (D2), which was exposed in wells Ecs-1 and Sze-2, the T_{max} is permanent. However, in well Ecs-1, situated further from the former coast, the T_{max} values are higher. Also, in well Ecs–1, the cuttings have higher free hydrocarbon (S1), and their distribution is more uniform. The oil production index values (OPI) also show the generation of hydrocarbon. In well Sze-2, situated closer to the former coast, the TOC and HC–POT values are higher. However, in this case, the residual organic carbon (S2) is responsible for higher HC–POT values:

	Ecs-1	Sze-2.
TOC %	0.40–0.80	0.50–2.6
S1 mg/g	0.10–0.23	0.06–0.13
S2 mg/g	0.50–1.43	0.77–2.52
HC–POT mg/g	0.60–1.66	0.80–2.65
T_{max} °C	440–444	435–440
HI	100–247	75–158

The lower distal turbidite layer (D1) – which was exposed in well Sze-2 and was deposited farthest from the former coast – is characterized by higher, varying, but increasing geochemical parameters, in comparison with the values in upper layers. In some samples the mature organic matter can be considered of good source rock quality.

TOC %	0.44–1.16
HC–POT mg/g	0.51–6.45 (max: 13.71)
S1 mg/g	0.15–1.76
T_{max} °C	443–451
HI	75–141

The open marine marls (calcareous marls - Pe) differ sharply from the distal turbidites in well Ecs-1. A remarkable increase can be seen in the values of HC – POT. However, the TOC values show no increase. This means that the quality of organic matter has changed and it is suitable for the generation of oil. The hydrogen index values (220–300) indicate oil generating organic matter which formed in marine surroundings. In well Sze-2 there is no obvious dividing line between the distal turbidites (D1) and the open marine marls (Pe). However, the increase of HC–POT (primarily S1) and the change in the quality of organic matter can also be seen. Near the 2850 meter level there are some remarkably high HC–POT values (14 mg/g). The low HI values prove that a shock ablation from the continent occurred. The T_{max} values are 444–451 °C. Analytical datas from the open marine rocks:

	Ecs–1	Sze–2	(Extra values)
TOC %	0.41–0.98	0.46–0.82	(9.15–11.62)
HC–POT mg/g	0.54–3.51	1.65–2.09	(14.36)
S1 mg/g	0.10–0.68	0.64–1.4	
T_{max} °C	444–447	446–451	(438–439)
HI	97–307	114–213	

The development of Miocene layers are different in wells Ecs-1 and Sze-2. The section examined in Ecs-1 consists of clayey marl, siltstone and limestone. These rocks alternate with each other and there are interstratified tuff layers. On the top of the unit, the clayey marl contains a lot of organic matter and its hydrocarbon potential is also high. In the deeper parts, where there are detrital limestone layers, the organic matter and the hydrocarbon potential decreases continuously. The T_{max} value is lower than in the Lower Pannonian open marine marl and lime-marl. In well Sze-2, siltstone, sandstone and tuffaceous mudstone layers are encountered. The tuffaceous mudstone also includes coal layers. The lowest part consists of gravelly sandstone and conglomerate. The geochemical parameters of layers change according to the rocks. In the sandstone, mudstone and tuff layers the average values of TOC and HC–POT are smaller than in the Pannonian open marine section. On the other hand, the parameters of coal layers differ very much from the others. In some places hot tuff caused the organic matter to change into the coking coal state. Lack of cap rock caused the generated hydrocarbons to spread into the air. In these coal layers, the value of HC–POT 60 mg/g is attained and the T_{max} is 484 °C. The first coal layer lies at depths of 2725–2735 meters. The organic matter type is II–III, which means that it was formed in a marine swamp. The average parameters of cutting samples differ from the parameters of coal core samples. So that the difference can be seen, the data of the cuttings sample from 2725 m and the core sample from 2727 to 2728.25 m are illustrated below:

	Cuttings	Core
TOC %	30.7	73.2
HC–POT mg/g	30	121
S1 mg/g	53.90	100.5
T_{max} °C	471	469
HI	175	159

In the deeper coal layers the HC–POT decreases, the T_{max} increases and the organic matter changes to Type III. The deepest part, which consists of conglomerate, has a small TOC value and the T_{max} is only 430–435 °C.

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Comparative study of vitrinite reflectance and T_{max} values of Hungarian Neogene sediments

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Vitrinite reflectance of cuttings and more than 150 core samples from Neogene columns of the Pannonian Basin were compared with the corresponding pyrolysis T_{max} values. As it is known, both parameters increase with the proceeding maturation of organic matter, but systematic differences were found between them. At the beginning of the hydrocarbon generation the T_{max} values indicate frequently higher maturity level than corresponding vitrinite reflectance data. In addition, in the majority of the measured samples, T_{max} values can be applied only in case of immature or moderately matured samples for estimate thermal maturity level, because the T_{max} seems to remain nearly constant although the R_o values increase.

It was found that the most likely reason for this phenomenon is the mineral matrix effect and the presence of some humic acid derivatives which are typical additives for water-base thermostable drilling fluids.

Our results direct attention to the importance of the appropriate and careful sample preparation.

Key words: vitrinite reflectance, pyrolisis, T_{max} values, additives

Introduction

From the point of view of hydrocarbon explorationists, one of the most important aspects of the geological history of a sedimentary basin is its thermal history. Chemical and physical properties can be measured which are known to change as a function of the proceeding diagenesis of rock as well as organic matter.

The most frequently measured maturity parameters are the pyrolysis T_{max} values and the vitrinite reflectance.

T_{max} is defined as the pyrolysis temperature, at which a maximum amount of hydrocarbons is released from the rock sample in consequence of kerogen thermal degradation. T_{max} can be used for the quick estimation of the degree of maturation of organic matter by means of its evaluation from the geochemical log (Espitalié 1987).

Vitrinite reflectance is an optical parameter, measured on polished surface microscope slides. It is known to be an important indicator of the level of diagenesis of organic matter, because it increases in a irreversible way with proceeding thermal maturity (Dow 1977).

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The purpose of this paper is the comparation of the results of these two kinds of measurements. Thus, vitrinite reflectance of cuttings and more than 150 core samples from Neogene columns of Pannonian Basin were compared with the corresponding pyrolysis T_{max} values.

Analysis of the parallel T_{max} and R_o values

In Fig. 1 we graphically compared vitrinite reflectance data of core samples and cuttings from some boreholes with the parallel T_{max} values of Rock-Eval pyrolysis in the range of 0.3–2.0 R_o%. It was proved by Espitalié (1987) that the type of kerogen influences the pyrolysis T_{max} values. Nevertheless, the Neogene sediments of the Pannonian Basin contain predominantly terrestrial organic matter, so we used T_{max} scale suitable for Type III organic matter in our figures. It can be seen in Fig. 1 that R_o and T_{max} do not reflect the very same maturity. In the right-hand corner of the figure the results published by Espitalié (1987) can be seen. It is obvious that the two maturity parameters of the studied Neogene samples cannot be correlated with each other in such a simple way. The difference is better expressed in Fig. 2, where R_o and T_{max} values of cuttings and cores are shown separately as a function of depth.

Under 0.7–0.8% R_o (immature and moderately matured samples), the T_{max} values indicate a somewhat higher thermal maturity level than corresponding vitrinite reflectance data, primarily in the case of cuttings. On the other hand, the T_{max} values are lower than expected above 0.8–0.9% R_o. This effect is not so visible in the cores. Data are more scattered, but the trend is similar.

Discussion

On the basis of publications about pyrolysis methods we tried to find the source of the above-mentioned differences. The effect of the type of organic matter can be neglected because the studied samples contain Type III organic matter uniformly. That heavy oil accumulations cause T_{max} values to decrease (Espitalié 1987) is inconsistent with our observations. So these influences can be neglected.

On the other hand, T_{max} varies with the so-called mineral matrix effect. It means that, depending on rock type and quantity of organic matter, the mineral matrix may interfere with pyrolysis parameters by retaining certain hydrocarbons generated during pyrolysis. Because of this retention, a decrease in the S1 and S2 peaks and an increase in T_{max} values can be observed. These deviations are consistent with the differences noticeable in our data. For example, in Fig. 3 a considerable increase in T_{max} values can be identified in the mud column (which contains immature organic matter) compared with the parallel 0.2–0.5% vitrinite reflectance values.

Thus, the mineral matrix effect may be one of the reasonable causes of the difference between maturities indicated by T_{max} and R_o.

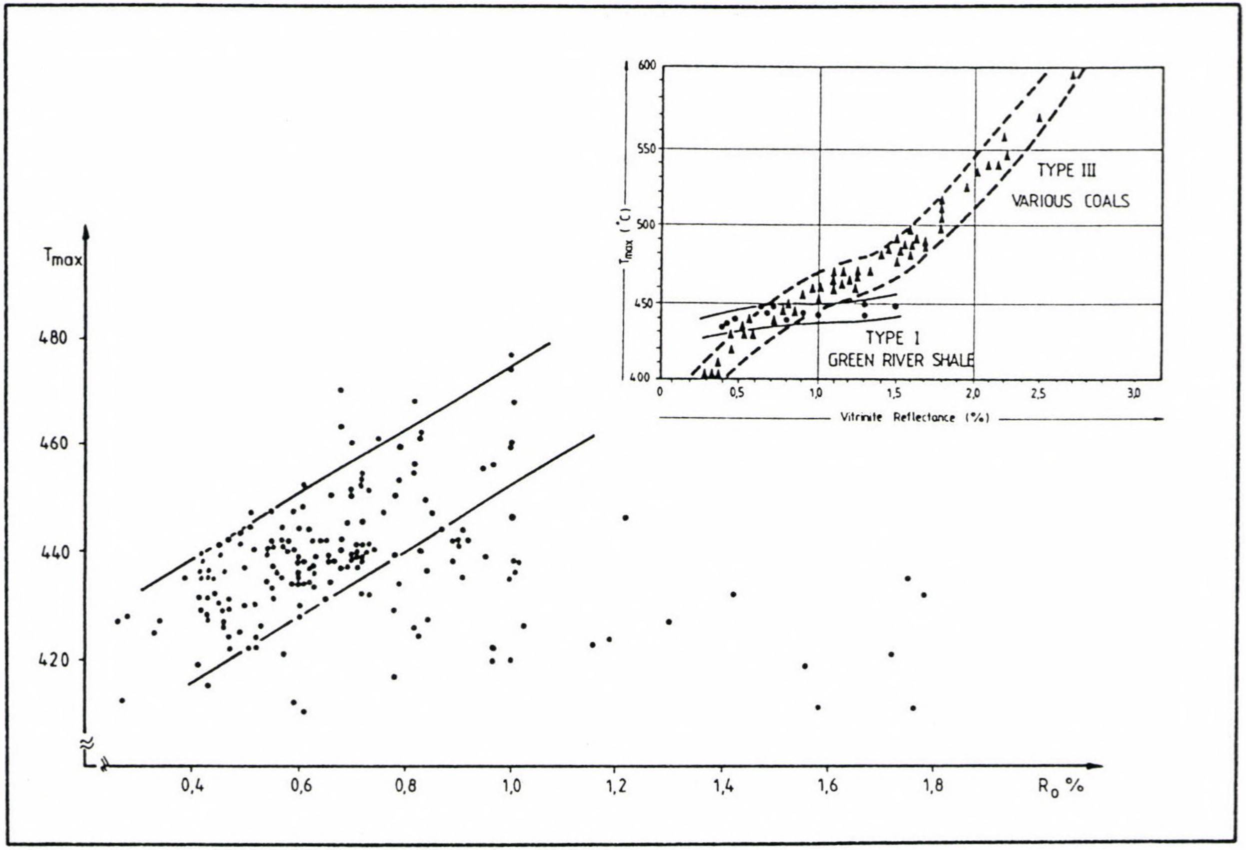

Fig. 1
Correlation between R_o and T_{max} for Neogene samples. In the right-hand upper corner the referred results of Espitalié (1987) can be seen

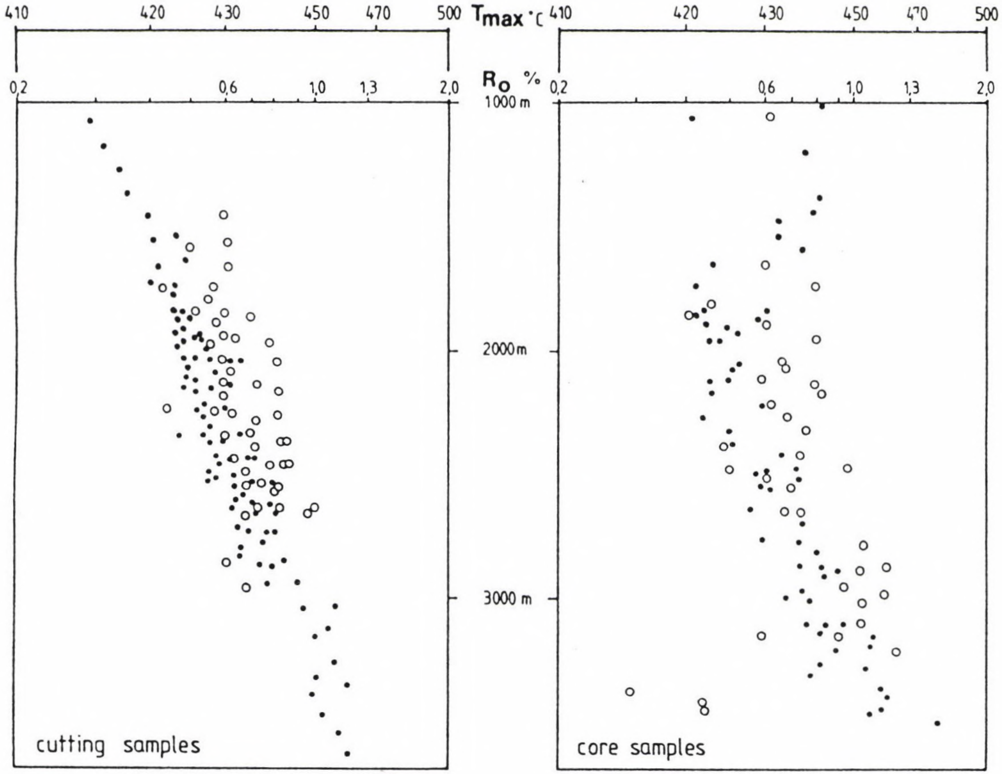

Fig. 2
Maturity parameters of the Neogene core samples and cuttings versus depth. (o) – T_{max}, (.) – Ro

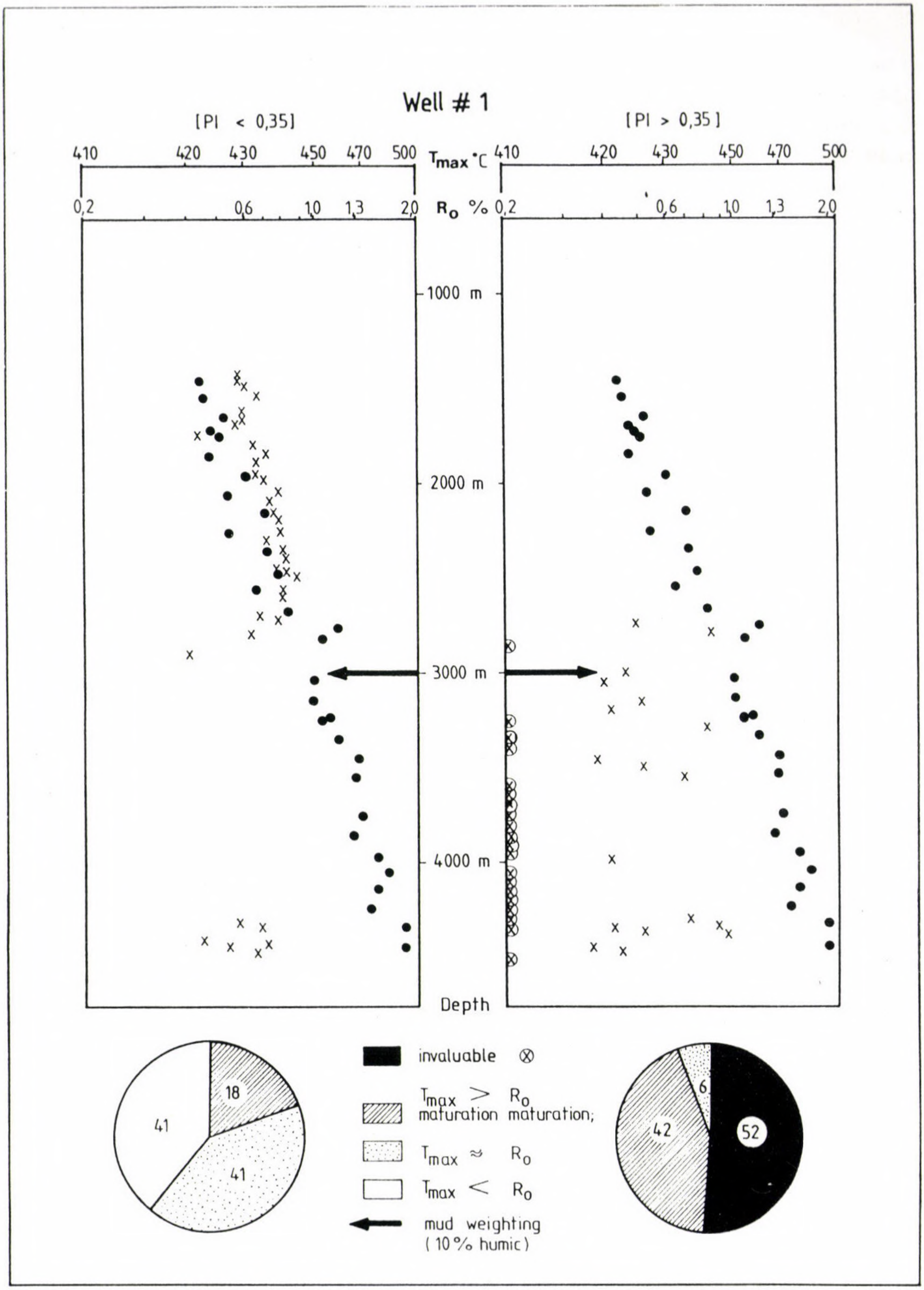

Fig. 3
Extract from geochemical log of a borehole in the centre of Hungarian Plain. (x) – T_{max}, (.) – R_o

The effect of drilling mud additives

We also have found that we must account for the effect for drilling fluid additives, as well. In Fig. 4 a portion of the geochemical log of a borehole in the centre of Hungarian Plain can be seen. It was drilled with thermostable drilling fluid, which contained various additives. The drilling mud also contained humates. The concentration of humates was increased above 5% below 3000 m. It can be seen in the figure that the vitrinite reflectance increases with depth, but the parallel T_{max} values are essentially constant. This uncommon phenomenon attracted our attention to the effects of drilling mud additives.

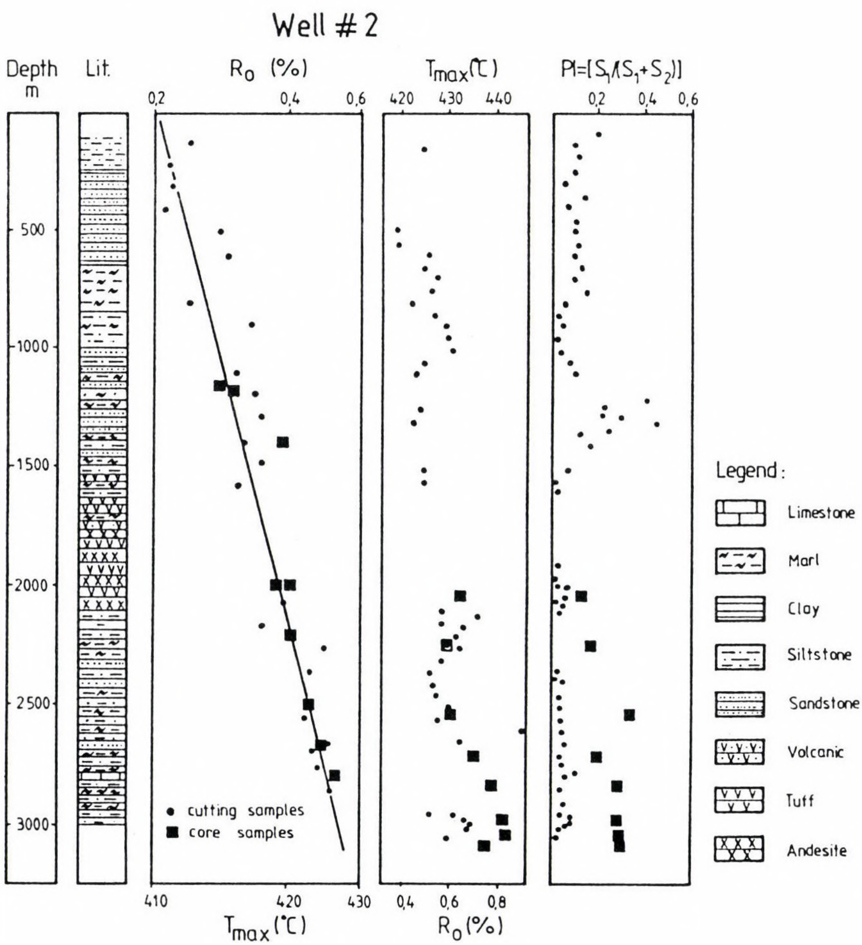

Fig. 4
Deviation between Ro and T_{max} values. Extract from the geochemical log of a North Hungarian borehole

As the inverse thermal gradients in the Pannonian Basin is much higher (0.05 °C/m) than the global average (0.03 °C/m), it is necessary to increase thermal stability of drilling fluids (Völgyi 1977). The humic acid derivatives are typical additives for water-base thermostable drilling fluids (Dormán 1991).

These macromolecular materials are characterized by complex chemical structure. Their molecular weight, chemical and physical properties are not exactly clear yet, but they are especially useful for improvement of thermal and rheological properties of drilling fluids. Unfortunately, the thermal properties of these additives, which were studied in detail by Szöőr et al. (1984), may modify the pyrolysis results, because their thermal degradation is considerable between 300 and 600 °C (Fig. 5).

These additives can be adsorbed on the specific surface of detrital rocks as well, and their presence may modify the results of the Rock-Eval pyrolysis. These may act like an equalizing factor: the lower T_{max} values increase, the higher ones decrease.

Experimental adsorption test

It was supposed that adsorption of humic acid derivatives on the surface of cuttings may be significant and so affect the results of pyrolysis. So a series of measurements was planned under mild circumstances: a cutting sample was crushed and sieved. The 500–2000 μm fraction was used in the experiments. Static adsorption tests were conducted in sealed flasks. 3–3 g of cuttings were immersed in 30 cm^3 water (blank solution), 5–50 g/l Termohumex (sulpho-methylated humic acid, Na–salt) solution (I–III solutions) and weighted, field drilling fluid containing 46.1 g/l humic acid derivatives. After 2 hours the cuttings were sieved, washed with 300 cm^3 water, dried and crushed (particle size: <20 μm). Five hundred mg powdered cuttings were extracted with 3x5 cm^3 NaOH solution of 50 g/l concentration, at 25 °C, then extracts were collected and filtered. The concentration of humic acid derivates in the extracts was measured by UV-spectrophotometer.

Results of the adsorption test

The data in Table 1 include the amounts of the adsorbed humic acid derivatives on cuttings. In spite of moderated conditions (only the Termohumex solution was used in the experiment in reduced concentration without other additives, at 25 °C temperature), the amount of the adsorbed humic acid derivatives is considerable, but fortunately the adsorbed humates can be completely removed from the cuttings by washing with 100 parts of water (Table 1).

After the experiment, treated and untreated samples were pyrolysed in the Rock-Eval instrument. The average pyrolysis results are collected in Table 2.

As can be seen, the T_{max} of the unwashed samples is quite different from the original value, even in the case of the overmature samples. The T_{max} of the less mature or immature samples is not so strongly affected, perhaps because the original T_{max} is close to that of the additives.

Table 1.
Adsorption of humic acid derivatives on fine grained cutting samples

Type of fluid	Adsorbed humic acid derivatives	
	without washing (mg/g)	after washing with 100 parts of water (mg/g)
I. solution 5 g/l Termohumex solution	0.65	< 0.05
II solution 25 g/l Termohumex solution	0.86	< 0.05
III solution 50 g/l Termohumex solution	5.59	< 0.05
K-5 drilling fluid containing 46.1 g/l humates	–	< 0.05

Table 2
Pyrolytic results of original and treated samples

Samples	S_2 (mg HC/g rock)	TOC (%)	HI (mg HC/TOC)	Tmax (°C)	R_o%	Calculated R_o%
Cutting sample	0.20	0.51	19	528	1.6	2 <
"Termohumex" (solid)	4.50	0.86	523	431	0.3–0.4	0.6
Cutting sample treated with I solution	0.25	0.58	25	525	1.6	2 <
Cutting sample treated with III solution	0.60	1.03	58	433	1.6	0.6

Calculated R_o = equivalent to T_{max}

Conclusions

– In the range of 0.4–0.8% R_o, the pyrolysis T_{max} values indicate somewhat higher thermal maturity than vitrinite reflectance.

– One of the reasonable causes of this phenomenon is the mineral matrix effect discussed by Espitalié.

– It was experimentally proven that adsorbed humic acid derivatives, which are common additives of the thermostabile drilling fluids, can also cause similar deviations.

– Measuring of vitrinite reflectance seems to be a more reliable method than T_{max} determination in the evaluation of thermal maturity of organic matter.

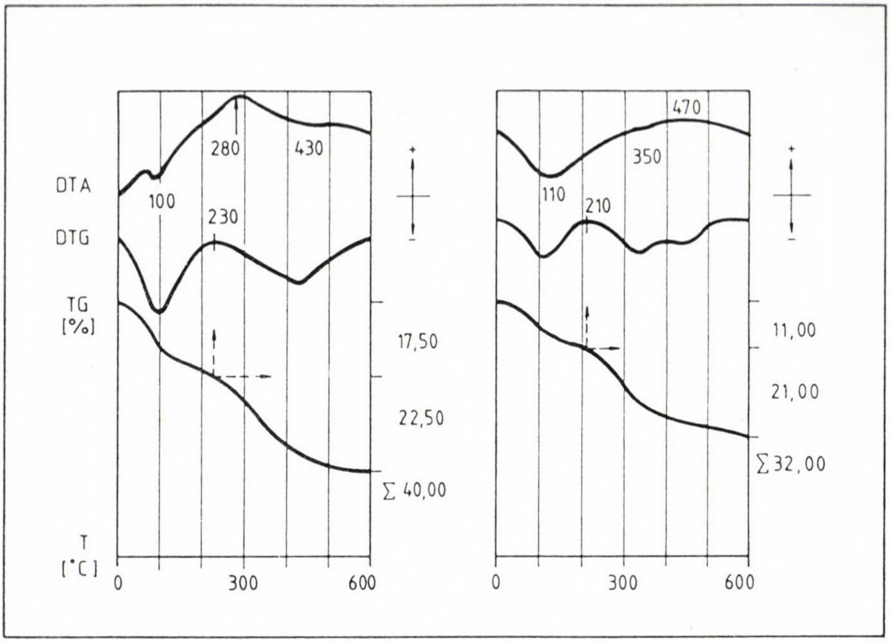

Fig. 5
Thermal stability of a common drilling mud additive (based on derivatographic data) after Szöőr et al. (1984)

Acknowledgement

The authors would like thank I. Hasznos from the Geophysical Exploration Company for the Rock-Eval pyrolysis of humic and treated samples.

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Vitrinite reflectance and smectite content of mixed-layer illite/smectites in Neogene sequences of the Pannonian Basin, Hungary

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Two important parameters of diagenetic transformation, vitrinite reflectance (R_o) and smectite proportion in mixed-layer illite/smectites (S), were measured in Neogene shales of the Pannonian Basin. Nearly 350 R_o and S values were collected from 33 boreholes. Only S values of normal terrigeneous clastic rocks, containing little kaolinite and not affected by weathering, were taken into consideration. The dependence of these parameters on subsurface depth and their interrelation were studied.

R_o varies with depth in a linear manner except in the lowermost portion of the basin below 3400–3700 m. S varies with the depth in a less regular manner, having three distinct zones with smectite contents of 80 to 100 per cent, 20 to 80 per cent and <20 per cent, respectively.

The scattering of S is much higher than that of R_o. Considering the whole Pannonian Basin the scattering is very high for S, due to different depth of transitional zone and to its different length in each part of the basin. In order to see the differences in different regions of the Pannonian Basin, the boreholes were classified in five regional units: (1) Little Hungarian Plain, (2) Zala and Dráva Basins, (3) central shallow areas, (4) North Hungarian basins and (5) Great Hungarian Plain.

In deep partial depressions of the Pannonian Basin, the transition from smectite to illite starts earlier and lasts longer than in shallow and medium-deep areas covered only by few hundreds of meters of Neogene sediments. It is supposed that this is due to shorter duration of the thermal effect in deep subbasins formed by relatively young subsidence. In many shallow and medium-deep areas, on the other hand, erosion removed part of the Upper Pannonian and Quaternary from the top of the sequences, and the zone of diagenetic transition is at present in a relatively elevated position.

Key words: Vitrinite reflectance, illite/smectite, Neogene, Pannonian Basin

Introduction

Vitrinite reflectance and smectite content of illite/smectite mixed-layer clay minerals are useful maturity indicators of sedimentary rocks. Their mutual relationship was studied by a number of authors (see the reviews of Kübler (1984) and Kisch (1987) and more recent investigations by Velde and Espitalié (1989), Francu et al. (1990), Miki et al. (1991)).

According to these studies there is a systematic correlation between the two variables; however, the actual form of the interrelation varies from area to area of study and depends on kinetic factors such as duration and intensity of

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thermal effects, availability of potassium, etc. One important conclusion of these investigations was that vitrinite reacts earlier to increase of temperature than illite/smectite. Kisch (1987) stated that "under the effect of steep geothermal gradient the illitization of smectite... 'lags' behind the progress of coalification" (p. 246). Pollastro (1990, 1992) concluded that it is reasonable to consider two types of burial transformation of illite/smectites, one with short (< million years) duration of the thermal effect and another with sufficiently long heating time (2 to 300 million years). In the first case the transformation of illite/smectites is "delayed" in respect of that of the organic matter.

Vitrinite reflectance data used in the present study were in part previously published and discussed by Laczó (1982) and Laczó and Jámbor (1988a, b) as well as smectite data by Viczián (1982, 1985, 1992). This is the first attempt to correlate these variables in Hungarian part of the Pannonian Basin.

Methods of data collection and evaluation

Vitrinite reflectance (R_o) has been measured in the Hungarian Geological Survey since 1976. Most vitrinite data considered in this study were measured by I. Laczó (until 1988), later by M. Hámor-Vidó.

Analyses of the smectite proportion in mixed-layer illite/smectites were made by I. Viczián in the period 1979 to 1992, essentially by the same method based on the d-position of basal reflections of ethylene glycol-treated samples of the <2 mm grain size fraction (Johns, Kurzweil 1979, Srodon 1980).

Both vitrinite and smectite data were collected for shale samples of Neogene age taken from boreholes in the Hungarian part of the Pannonian Basin. The average distance of sampling for the study of vitrinite was 50 m. The measurements were carried out on core samples, except for the boreholes Doboz-I, Örm-I, Víz-I and Cs-1 where cuttings were used.

The results of smectite analyses were critically reviewed in respect of the geological provenance of the samples and analytical accuracy. Only shales composed of terrigeneous detrital material were considered; samples of clayey rocks produced by in situ kaolinitic weathering, sedimentary rocks with important volcanogeneous contribution and variegated clays of Upper Pannonian and Quaternary age containing iron hydroxides were omitted. Also, samples with authigenic clay minerals in open pore spaces were not considered. Details of this critical evaluation are discussed by Tanács and Viczián (1992).

Analytical data often show a broad range of smectite percent values within the same sample. For a given sample, only the most frequent percentage was considered; very broad intervals (e.g. 40 to 100 per cent) or uncertain results were omitted from the present study.

Results

In studying the variation of vitrinite reflectance (R_o) and smectite (S) values with subsurface depth the boreholes were grouped in 5 regional units (see Fig. 1):

1. Little Hungarian Plain,
2. Zala and Dráva Basins,
3. central shallow areas of the Pannonian Basin,
4. North Hungarian basins and
5. Great Hungarian Plain.

These 5 regions of the Pannonian Basin differ from each other in respect of depth of basement, geothermal gradient, stratigraphy, etc. Very deep (4 to 6 km) depressions are found in regions 1, 2 and 5.

Plots of R_o and S data versus depth for these 5 regions are shown in Figs 2 and 3. The variation of R_o with depth was considered to be linear, and straight lines were computed and drawn for the 5 regional units. A few borehole sections of special interest are indicated in Fig. 2 individually (e.g. Örm-I, Hód-I, Doboz-I in region 5). In borehole Ig-7 (region 3) R_o is remarkably high and its increase with depth is much more rapid than elsewhere. The slope of the line is similar to that of the borehole Ig-7, but the absolute values of R_o are not exceptionally high in the nearby borehole Gf-1. As compared to other boreholes in region 4, there are slightly higher R_o values in borehole Tp-1.

The variation of S with depth cannot be approximated by a linear regression line. Curves for every borehole are shown in the diagram summarizing the 5 regions (Fig. 3). Curves of S represent 3 zones of transformation of smectite into illite:

I. shallow burial zone containing smectite or illite/smectites with high smectite content,

II. transitional illite/smectite zone (S = 20 to 80%) and

III. deep burial zone containing illite/smectites of low smectite content.

Because of geological reasons as well as incompleteness of sampling and omissions according to the criteria discussed above, in many cases only a part of the whole transformation is represented in a particular borehole.

In Fig. 4 R_o is studied as a function of S. As R_o is a linear function of depth, S versus R_o diagrams often have a similar shape as S versus depth plots. The relationship of R_o to S cannot be represented by a single straight line because of the existence of 3 zones of smectite transformation. S versus R_o diagrams differ from S versus depth diagrams in the very limited representation of zone I. This is due to the fact that in the entire shallow burial zone there is only slight variation in the value of R_o (see e.g. short vertical lines near S = 100% and R_o = 0.3% for boreholes Paks-2, Gf-1, Hn-1 in Fig. 4).

Only variations with depth were considered so far. The discussion of the effect of temperature is left for further studies. In the following discussion, however, many aspects of the thermal effect will be briefly mentioned.

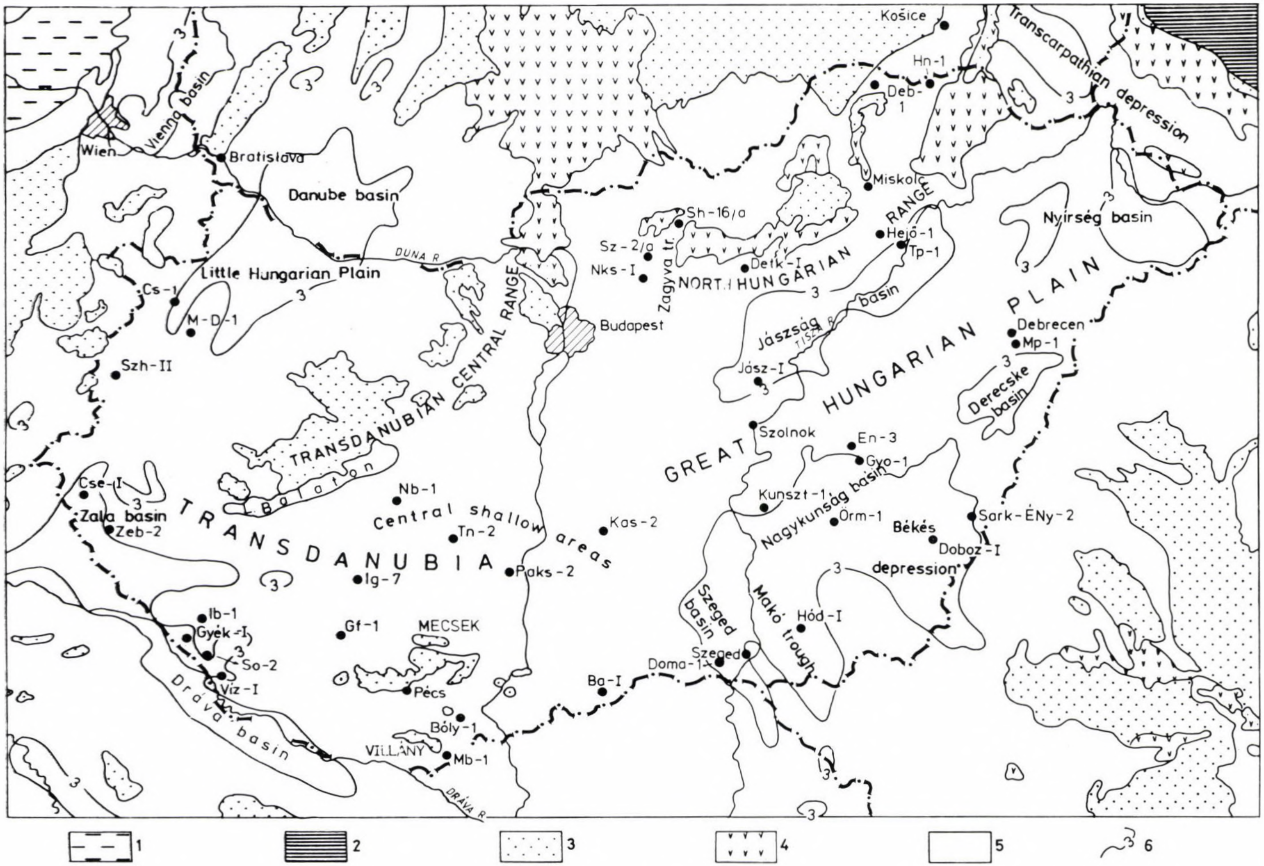

Fig. 1
Location of the boreholes studied in the Pannonian Basin. The geological map was compiled by Horváth (Map 1 in Royden and Horváth (eds) 1988). 1. foredeep; 2. Flysch belt; 3. Inner Alpine, Carpathian and Dinaric units; 4. Late Cenozoic volcanic rocks; 5. Pannonian Basin; 6. 3 km depth contours of the base of Neogene

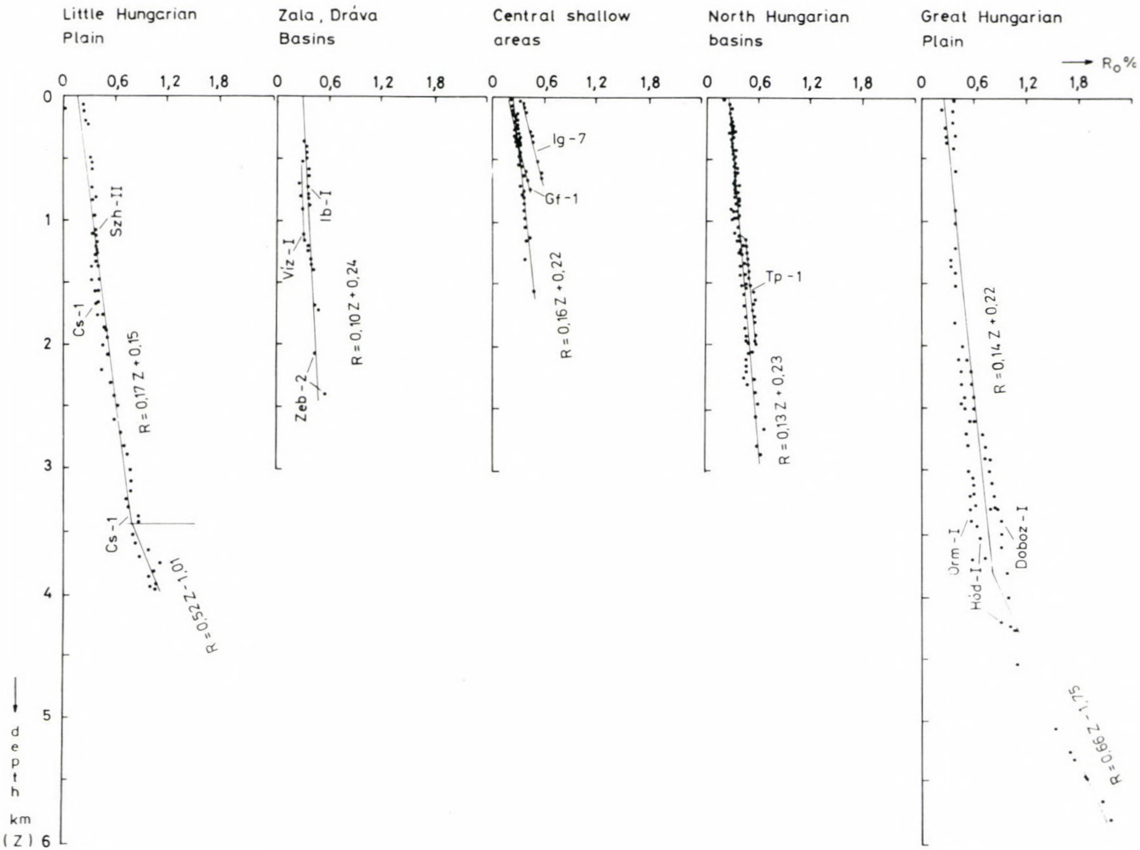

Fig. 2
Variation of vitrinite reflectance (R_o) with depth in 5 regions of the Pannonian Basin. Z. depth. Equations of the linear functions R_o versus Z are given

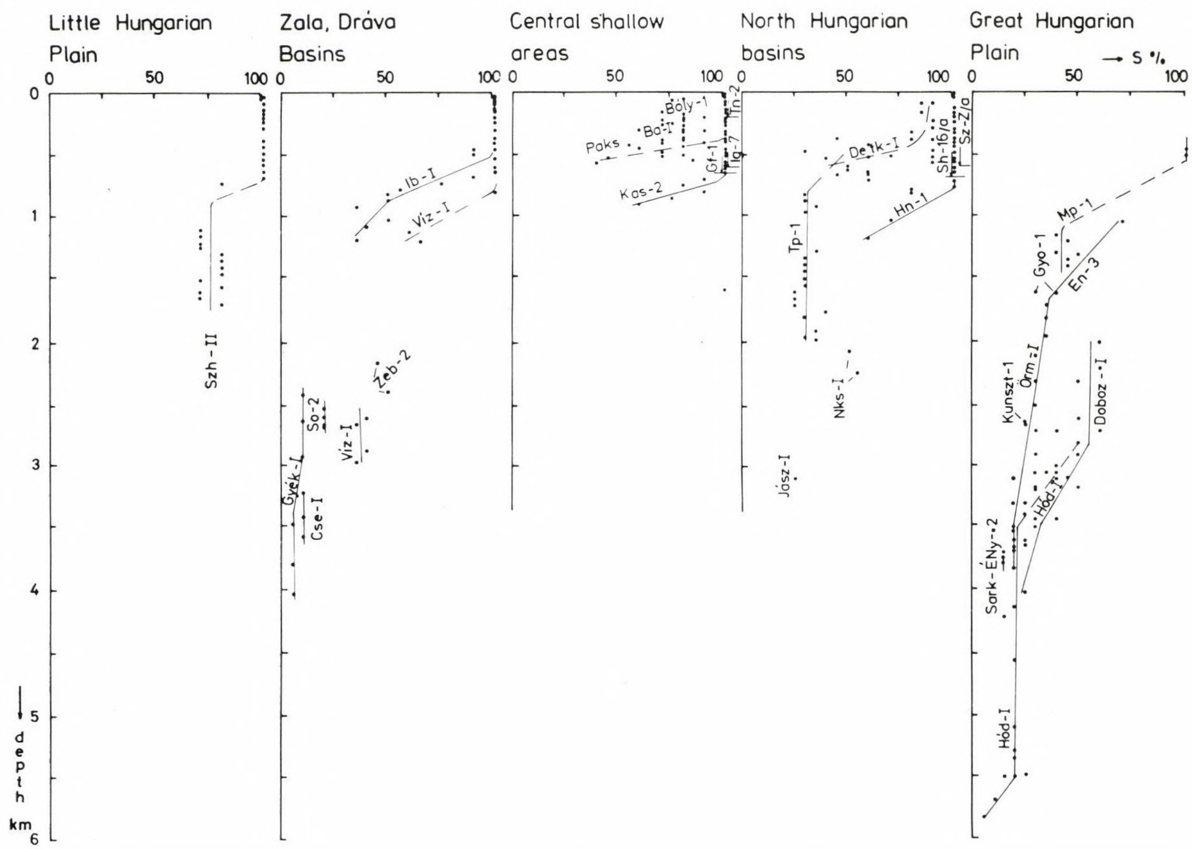

Fig. 3
Variation of smectite content in mixed-layer illite/smectites (S) with depth in 5 regions of the Pannonian Basin

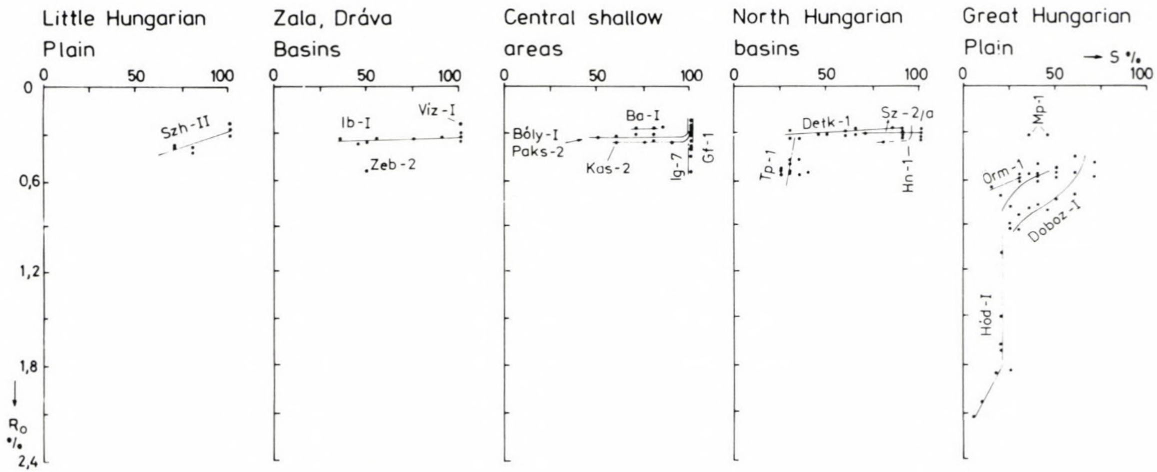

Fig. 4
Interrelationship of vitrinite reflectance (R_o) and smectite content in mixed-layer illite/smectites (S) in 5 regions of the Pannonian Basin

Discussion of the results

A. Vitrinite

The trend lines of vitrinite reflectance are very close to each other in the upper 1500 m in the five regions. Divergence from this trend is detected only in few cases:

– In computing the trend line for the central shallow areas (region 3), Ig-7 borehole samples were left out of consideration because of exceptionally high R_o values. Geological evidence shows that several hundreds of meters of Upper Pannonian sediments were removed here by erosion. Moreover, the rapid growth of R_o with depth points to a local heat source in the neighborhood of this borehole. A local heating effect may be supposed for borehole Gf-1 as well.

– Erosion of Upper Pannonian sediments can be traced in other regions and in other boreholes as well, e.g. in borehole Tp-1 (region 4, North Hungarian basins) where R_o values are higher than the average trend below the depth of 800 m.

The gradient of vitrinite reflectance is the highest in the Little Hungarian Plain. However at shallow depths (0–2200 m) the increase of R_o is lower than that in other regions. This may be due in part to differences in the quality of organic matter.

The Zala and Dráva Basins are the coolest territories according to the least rapid increase of vitrinite reflectance with depth.

At shallow levels the trend found in the Great Hungarian Plain (region 5) is almost identical with that found in the central shallow areas (region 3). At about 3700–3800 m there is a break in the slope of the trend line of the deepest borehole Hód-I. A similar break can be observed in the section of borehole Cs-1 (region 1) at a depth of about 3400 m. The increase or change of R_o with depth seems to be linear even below this break; however, the increase of R_o with depth is higher.

B. Smectite

The scatter of smectite per cent values with depth is relatively high, pointing to the influence of many geological factors (see Fig. 3).

In the shallow burial zone (zone I), most S values are equal or close to 100% (80 to 100%), but there are also S values below 80%. The relatively high scatter in this zone may reflect the presence of reworked material of various provenance and quality. The depth of the lower limit of the shallow zone (zone I) varies considerably (between 0.3 and more than 0.8 km). One important factor of this variation may be the Late Pannonian and Quaternary history of the sequences. In continuously subsiding areas with thick uppermost Pannonian and Quaternary, the shallow burial zone is also thick (e.g. in regions 2, 4, 5, boreholes Víz-I, Nks-I, Hód-I and Doboz-I). On the other hand, in areas of elevation and denudation in Upper Pannonian and Quaternary times, e.g. in

Central Transdanubia and around the mountains in Northern Hungary, the upper part of the shallow burial zone once formed has been eroded, and the present-day lower boundary of the zone is at shallower depth. A special example is borehole Tp-1, for which both smectite and vitrinite data show relatively high maturity below a certain erosional level (about 800 m).

The depth and the thickness of the transitional zone (zone II) vary in a similar manner as those of zone I. Zone II can be characterized by the depth where the transformation curves reach the 50% smectite content (see Fig. 3). This depth of S = 50% was plotted as a function of the depth of the base of the Neogene (Fig. 5). The relatively good relationship shows that in deeper subbasins the transitional zone begins deeper, and is thicker than in shallow subbasins. The highest negative deviation from this trend was found in borehole Tp-1. It is probably due to erosion (see above), while the reason for the positive deviation shown in borehole Doboz-I is unknown.

The composition of the illite-like mixed-layer illite/smectites is relatively constant in the deep burial zone (zone III); their smectite content, however, varies from one borehole to another. In older sequences, S values are lower than in younger ones (see Table I). This is probably due to the slow recrystallization of illite in this zone by the Ostwald ripening process.

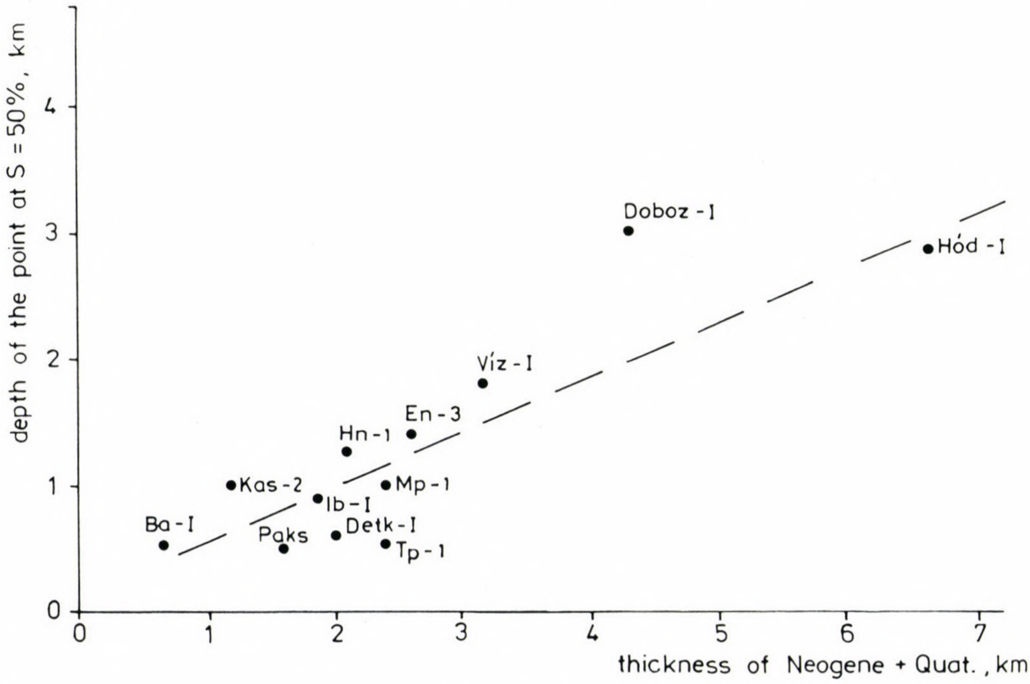

Fig. 5
Depth of the point at S = 50% on trend lines of boreholes in the S versus depth diagrams (see Fig. 4) as a function of the thickness of the infill of the Pannonian Basin

Table 1
Variation of smectite content in the deep burial zone as a function of stratigraphy

Stratigraphy	S%	Boreholes
Pa_2	30-45	Tp-1, Mp-1
Pa_1	20-40	So-2, Víz-I, Tp-1, Örm-I, Doboz-I
Ms + Mb	5-20	Gyék-I, Hód-I
Mb	25	Kunszt-1, Jász-I
Mb + Mk	15	Sark-ÉNy-2
Mb (+tectonic deformation)	10	Cse-I
Mk	5	Gyék-I

Symbols: P – Pannonian; M – Miocene; s – Sarmatian; b – Badenian; k – Karpathian

C. Vitrinite and smectite: a comparison

The conclusions of the diagrams shown in Fig. 4 are summarized in Fig. 6, where the main trend lines represent the various types of regions of the Pannonian Basin. There is a clear difference between the interrelationship of R_o and S in shallow and medium-deep areas and the deepest depressions. The difference is probably connected to the duration of the thermal effect in various subsiding areas. In depressions with thick Late Pannonian and Quaternary cover the sediments subsided into the zone of transition relatively late, as compared with other areas where subsidence of the sequence occurred much earlier. The conclusions of Pollastro (1990, 1992) concerning the two models depending on the duration of thermal effect seem to be applicable here.

In Fig. 6 published data on the relationship of the same pair of variables are included for comparison. The trend line of the Transcarpathian Basin agrees relatively well with that of other deep depressions of the Pannonian Basin (Francu et al. 1990). The Vienna Basin fits well between the two lines of the Pannonian Basin in the low-reflectivity side (Francu et al. 1990). The Paleocene to Quaternary sediments found in borehole Karlsefni H-13 in the offshore of Labrador show a trend similar to those obtained for shales of the shallow and medium-deep regions of the Pannonian Basin: there is a rapid change of S at relatively constant values of R_o near 0.3% (Kübler 1984). In general, Fig. 56 in Kübler (1984) states that, in conditions of normal burial, the zone of rapid transformation of smectites lies in the range of R_o = 0.3% to 0.5%. The data from the Carboniferous of Silesia, Poland (Srodon 1979) fit well with the low S – high R_o side of the diagram for the Pannonian Basin. As a consequence of the relatively old age of these sediments, the relatively constant smectite content of the illite-like phase at R_o = 0.8% to 2.0% is somewhat lower than that of sediments of Tertiary age in the Pannonian Basin (Carboniferous: about 10%, Tertiary: about 20%).

Similar results were obtained by Velde and Espitalié (1989) for the relationship of smectite content and kerogen maturation. In their study, however, kerogen maturation is expressed by T_{max} values obtained by Rock-Eval pyrolysis.

A special case differing from simple burial diagenesis was described by Miki et al. (1991) from Tertiary basins of northern Kyushu, Japan (see curve VIII in Fig. 6). Here 100% smectite content can be found even in samples with relatively high R_o values (up to 0.9%) and the transformation zone of smectite corresponds to R_o values between 0.9% and 1.4%. According to the authors this delay in the transformation of clays in respect of that of the organic matter is a consequence of high paleo-geothermal gradient, and of the special chemical

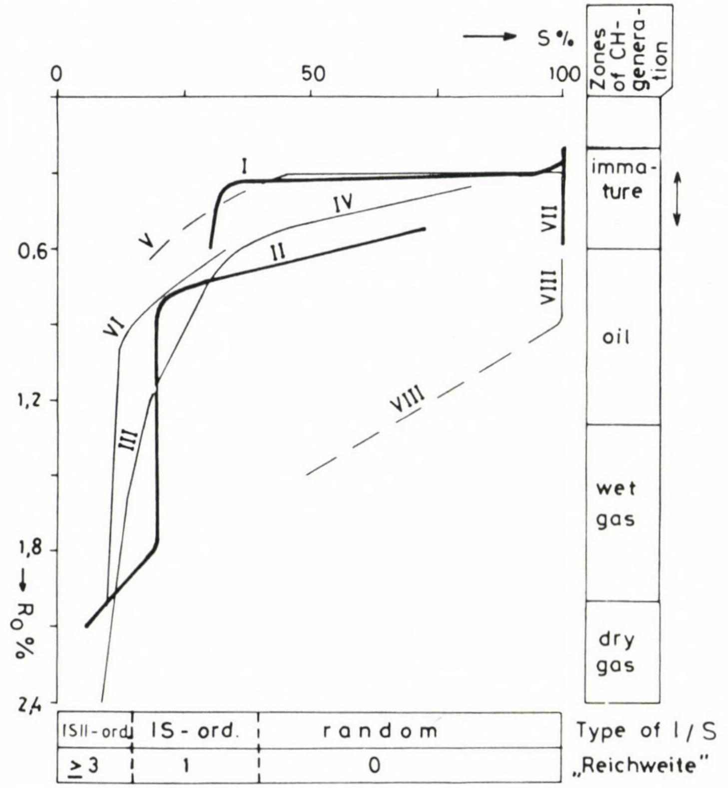

Fig. 6

Types of the interrelation R_o–S in the Pannonian Basin (lines No. I, II and VII). Trend lines from other sedimentary basins are shown for comparison (lines No. III to VI and VIII). R_o data are compared with zones of CH-generation according to Tissot and Welte (1978). S data are compared with type of ordering of the interstratification of illite and smectite (expandable) layers (see Reynolds and Hower 1970). Explanations: I. shallow and medium-deep basins; II deep basins; III. Transcarpathian Basin; IV. Vienna Basin (Francu et al. 1990); V. Tertiary of borehole Karlsefni H-13, off Labrador (Kübler 1984); VI. Carboniferous of Silesia, Poland (Srodon 1979); VII. boreholes Ig-7 and Gf-1; VIII. Tertiary of Kyushu, Japan (Miki et al. 1991). Double arrow on the right side of the diagram shows the R_o interval of transformation of illite/smectites according to Kübler (1984, Fig. 56) composition of pore waters and host sediments. A similar effect could be observed in boreholes Ig-7 and Gf-1, where smectite content remains 100% while R_o increase from 0.2% to 0.6% (see curve VII in Fig. 6).

Different forms of mixed-layer illite/smectites are often described in terms of ordering of component layers. For comparison with these publications type of ordering (Reynolds and Hower 1970) and the so-called "Reichweite" of the structure are shown in the lower side of Fig. 6.

The transformations of clay minerals and kerogen were compared with main zones of CH-generation according to Tissot and Welte (1978). It can be concluded that in the case of normal burial diagenesis, and not too rapid or short-lived heating effect, the bulk of the transformation of smectite occurs prior to oil generation in the zone of immature organic matter.

Conclusions

1. Both vitrinite reflectance (R_o) and smectite content (S) are useful indicators of diagenetic maturity of sediments; however, *no linear relationship exists between* R_o and S. A rapid decrease of S from 100% to about 20% at nearly constant R_o (0.3% to 0.6%) is followed by a subsequent growth of R_o at nearly constant low values of S (10 to 20%).

2. Three types of interrelationship of R_o and S values can be observed in the Pannonian Basin: (1) shallow and medium–deep areas of the basin with "normal" rate of subsidence and duration of heating, (2) deep depressions with relatively rapid subsidence and (3) limited areas of very high geothermal gradient. The differences in the interrelationship of R_o and S are principally due to the fact that clay minerals react much more slowly than organic matter during burial and the transformation of illite/smectite follows the maturation of vitrinite only *with considerable delay.*

3. As both vitrinite and illite/smectite transformations are irreversible changes reflecting the maximum temperature ever attained by the rock, the consideration of *erosional events* in the postdepositional history of the sequence is of high importance in the interpretation of the maturity data. Further investigation of the extent and time of the denudation of young sediments in the Pannonian Basin would be desirable.

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Organic geochemical evaluation of the Makó-3 borehole

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The studied sequence (530–4170 m) of the Makó-3 well contains mostly Type III kerogen. Type III kerogen with low hydrogen index was found both in the psammitic sediments of Nagyalföld Formation and in the mainly pelitic and partly psammitic rocks of the Lower Pannonian Algyő and Szolnok Formations. The high organic carbon content and hydrocarbon potential indicated immature, good gas-prone organic matter in the Nagyalföld Formation, whereas delta front and delta slope depositional facies of the Algyő Formation, as well as prodelta facies of the Szolnok Formation, resulted in poor gas-prone source rocks with low organic carbon content and low hydrocarbon potential. Moderate gas-prone Type III kerogen was identified in the lower part of prodelta sediments (Vásárhely Formation). A high proportion of free hydrocarbons and oil indications showed the presence of migrated hydrocarbons in this sequence. Excellent gas-prone Type III kerogen of relatively high hydrogen index, as well as fair oil-prone Type II–III kerogen, were also detected in the Upper Pannonian delta plain sediments, consisting of psammitic rocks with frequent lignite beds.

Key words: Source rock, depositional facies, hydrocarbon potential, maturity and type of kerogen

Introduction

The Makó-3 well was drilled in SE Hungary, within the Makó–Hódmezővásárhely depression, formed during the Neogene as a part of the Pannonian Basin (Fig. 1). The SSE–NNW-trending depression was filled with 6000–7000 m of Neogene and Quaternary sediments (Körössy 1981). In Hungary, the Neogene sedimentary rocks and in particular the Pannonian formations contain mineral and fossil-fuel resources, including oil and gas reserves (Dank and Jámbor 1987; Szalay and Szentgyörgyi 1988). Organic geochemical investigations showed that the Neogene sediments could be considered as source rocks (Szalay and Koncz 1981; Koncz 1983; Szalay 1988). On the basis of Rock-Eval analyses of more

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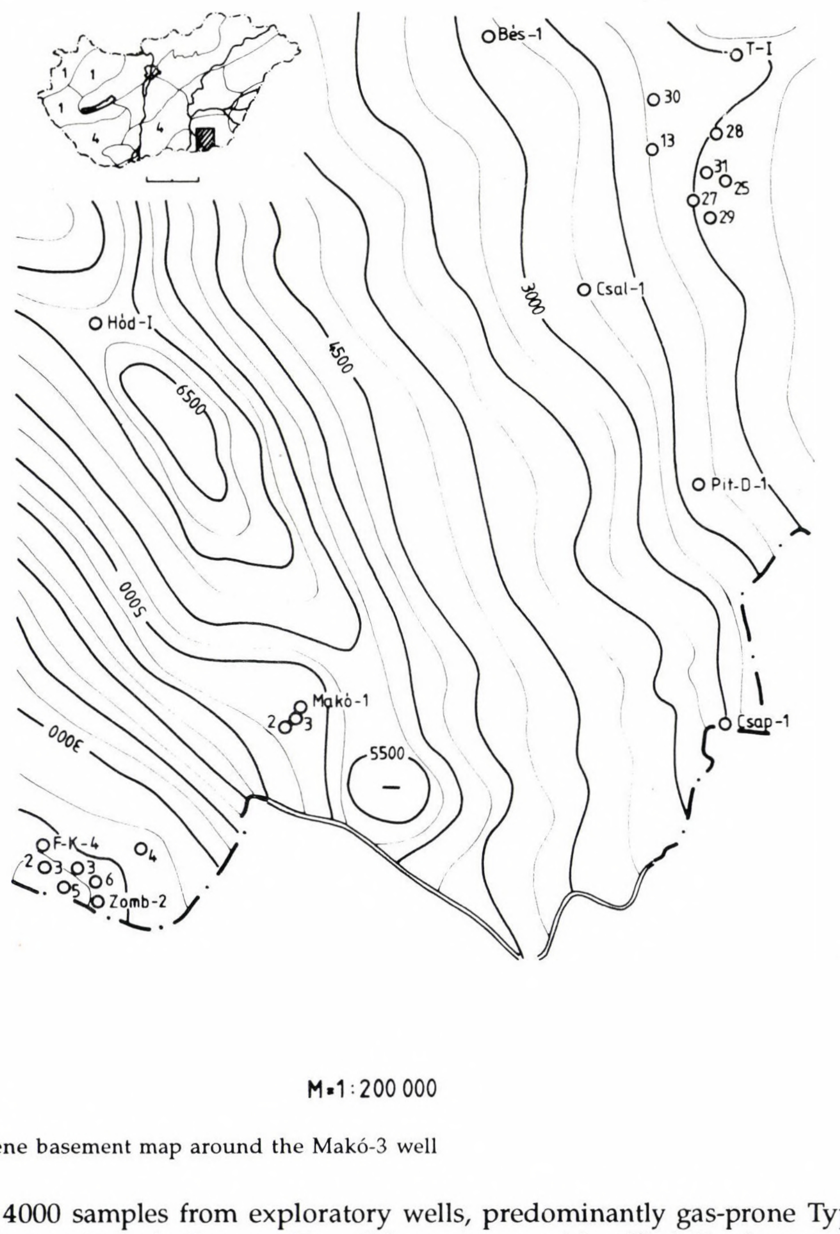

Fig. 1
Neogene basement map around the Makó-3 well

than 4000 samples from exploratory wells, predominantly gas-prone Type III kerogen and good oil-prone Type II kerogen were identified. Furthermore, fair oil-prone Types II–III kerogen was also found (Hetényi 1992). The geochemical analysis of the rock samples from Hód-I (Hódmezővásárhely) and Makó-2 wells (Fig. 1) indicated effective source rocks (Fig. 1). Small oil and gas reserves are known from the Makó-1 borehole, neighbouring the Makó-3 well. One of the

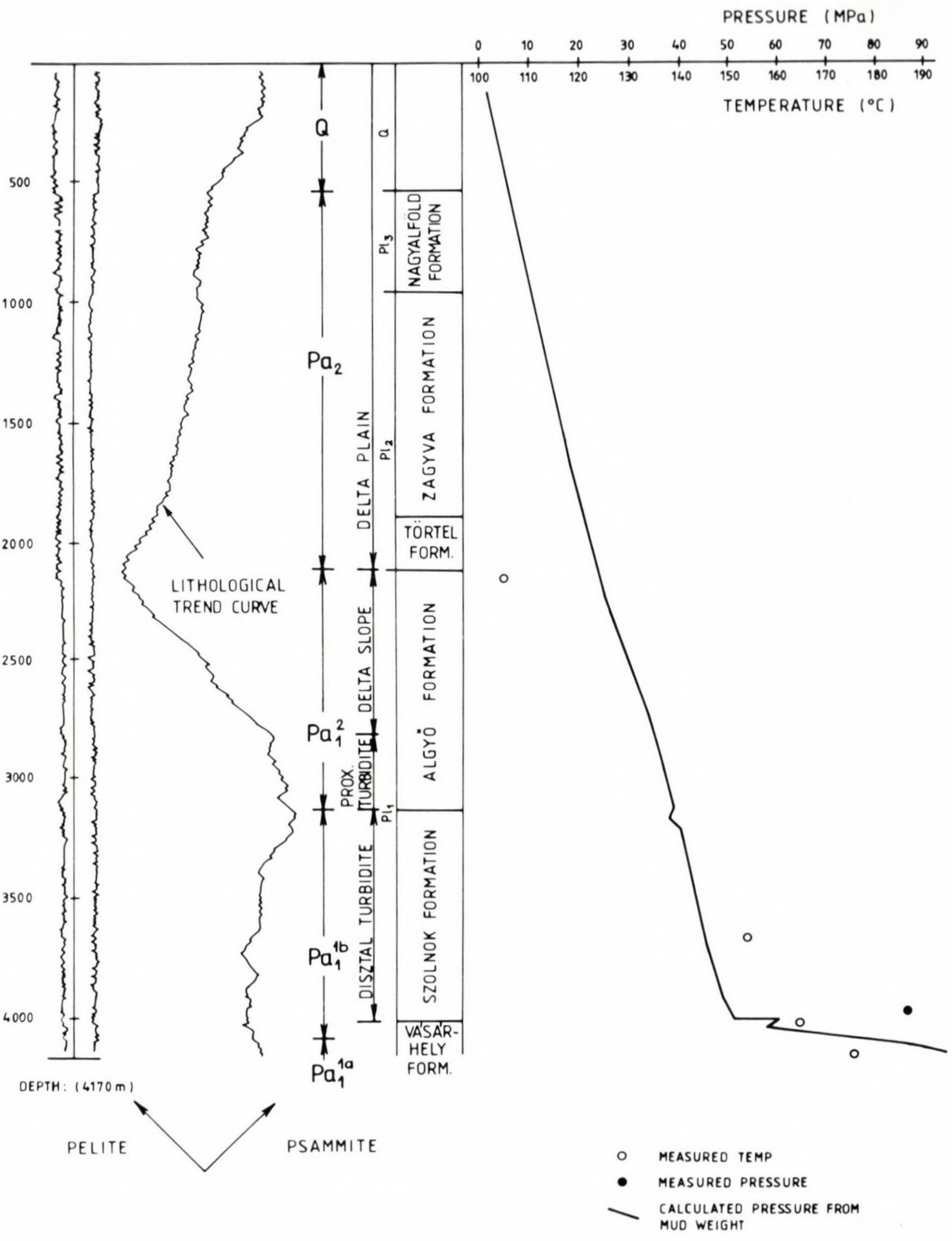

Fig. 2
Lithological trend curve, measured and calculated pressure, temperature of the Makó-3 well

important oil and gas accumulations of Hungary has been discovered in the Upper Pannonian reservoirs (Törtel Formation) at the western margin of the Makó–Hódmezővásárhely depression.

The aim of this work was to study the sedimentary rocks of the Makó-3 borehole as possible source rocks, to determine their organic richness, as well as the maturity and type of their kerogen.

Geological setting

The Makó-3 well was drilled in the western slope of the Makó–Hódmezővásárhely depression (Fig. 1). In this deep depression nearly 1000 m of pelitic sediments were accumulated, which consisted of mainly Badenian and Lower Pannonian basal marls. On the basis of lithological and paleo-geomorphological considerations, continuous sedimentation can be assumed. Neither the presence nor the absence of Sarmatian deposits could be confirmed. The studied sequences of the Makó-3 well (530–4170 m) can be seen in Fig. 2. The Vásárhely Marl Formation contains pelitic rocks, in which significant gas and oil shows have been found at depths of 4135–4144 m. The Lower Pannonian Szolnok Formation, which was determined by Szalay (1979) as a distal turbidite formation of a delta-system, is composed of psammitic and pelitic layers. The Lower Pannonian delta-front, delta-slope deposit (Algyő Formation) consists mainly of pelitic sedimentary rocks. Psammitic sediments, with frequent lignite and brown coal horizons, represent the Upper Pannonian delta plain depositional facies (Törtel and Zagyva Formations). Psammitic deposits were found in the Nagyalföld Formation of fluvio-lacustrine depositional facies.

Experimental

The samples were ground to size d< 0.2 mm. The determination of the organic carbon content was carried out with the Pregl-Dumas method, which is based on the measurement of the quantity of the adsorbed carbon dioxide. After carbonates had been dissolved by hydrochloric acid, the organic carbon was oxidized to carbon dioxide by means of combustion at 900–1000 °C under intense oxygen flow. The desorbed hydrocarbon gases of canned cuttings were analysed with a Carlo-Erba-3250 gas chromatograph on a column packed with Spherosil XOB. (Temperature program from 40 °C to 200 °C at 12 C/min, carrier gas: nitrogen, 25 ml/min.) Organic geochemical features of the kerogen were examined by Rock-Eval II pyroanalyser (Espitalié et al. 1977). Pyrolysis of about 100 mg samples at 300 °C for 4 min. was followed by programmed pyrolysis at 25 °C/min to 550 °C in an atmophere of helium.

Results and discussion

Organic geochemical evaluation of the well was performed by the organic carbon content, Rock-Eval pyrolysis and hydrocarbon gas analyses of 175 canned cutting samples. The change of the maturity of kerogen based on the vitrinite reflectance data can be seen in Fig. 3. The Upper Pannonian delta plain sediments (Zagyva and Törtel Formations) contain immature organic matter

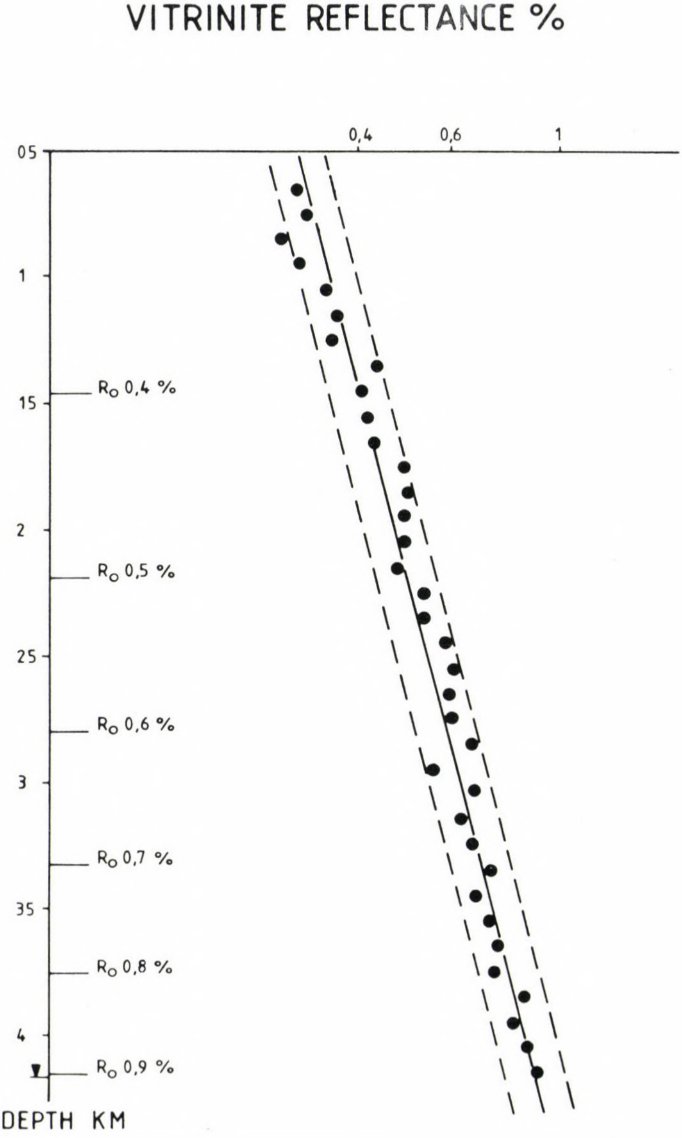

Fig. 3
Vitrinite reflectance vs. depth

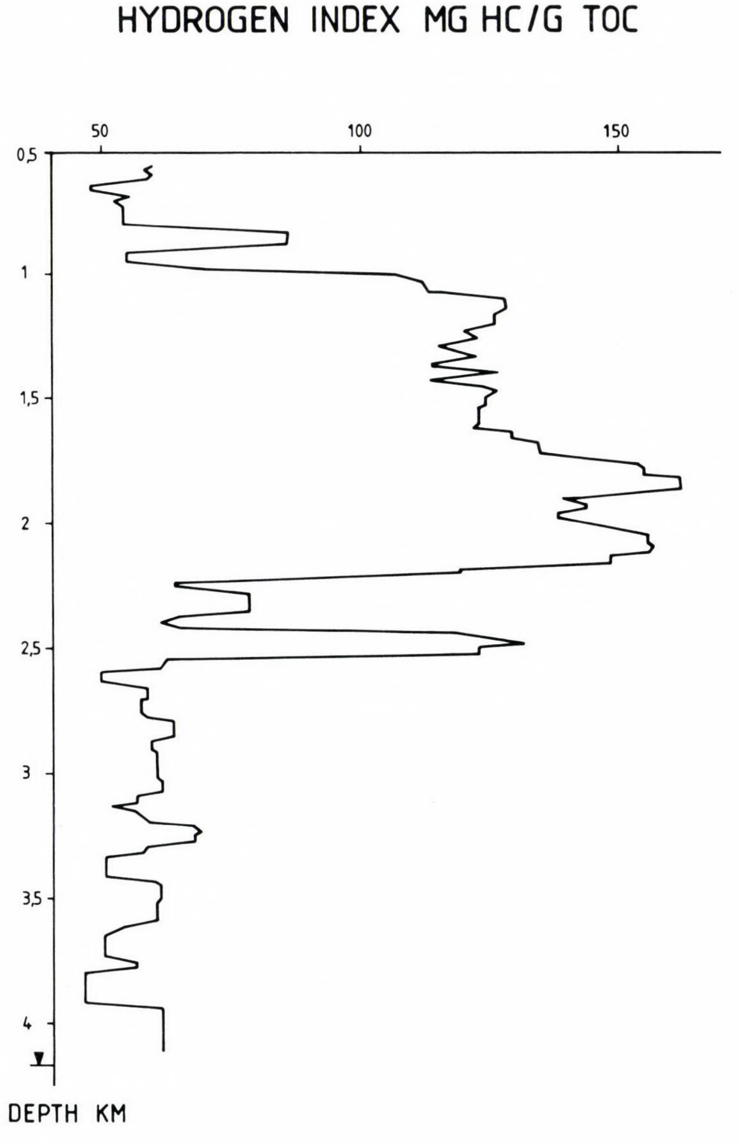

Fig. 4
Hydrogen index vs. depth

(R = 0.5%). The early zone of hydrocarbon generation (R = 0.4%) was reached at a depth of 1450 m (Zagyva Formation). The propane and butane components of the cutting gases first appeared at a depth of 1200 m, indicating the beginning thermal decomposition (Fig. 6). In the main zone of diagenesis (about 1100–2100 m), the rank of evolution can be followed by decrease of oxygen indices (Hetényi 1987–88). Both the results of Rock-Eval pyrolysis and vitrinite reflectance marked the beginning of catagenesis (R = 0.5%) at a depth of 2100–2200 m. During the catagenesis, the progressive thermal evolution of kerogen can be followed by vitrinite reflectance (Fig. 3) and the T_{max} values ranging from 430 °C to 445 °C between 2100 and 4170 m. The main hydrocarbon generation zone began at a depth of 2800 m (R = 0.6%, T_{max} = 435 °C). At the bottom of the well (4170 m), the maturity of the organic matter reached the peak of the oil generation curve (R = 0.9%, T_{max} = 440–450 °C) (Tissot and Welte 1984). In the Hód-I well (Fig. 1), Sajgó (1980) pointed out the main phase of petroleum generation (R = 0.69–1.16%) between 3450 and 5050 m. As can be seen in Fig. 3, the same vitrinite reflectance value (R = 0.7%) was found at depths of 3300 to 3400 m in the Makó-3 well. The studied samples proved to be relatively organic-rich ones. In the bulk of the cuttings, the amount of the organic carbon exceeded the minimum concentration (5 mg/g) accepted for petroleum source rocks (Fig. 5). Owing to lignite and brown coal beds, very high organic carbon contents were measured in the Upper Pannonian delta plain sequences (Törtel and Zagyva Formations), as well as in some strata (730–770, 810–870 m) of the Nagyalföld Formation. More than 10 mg/g rock organic carbon was found in the Vásárhely Marl Formation. The mainly pelitic sediments of the Algyő and Szolnok Formation (delta slope and prodelta depositional facies) contain organic matter of lower quantity (3–8 mg/g rock).

Comparing the change of the organic carbon content and hydrocarbon potential, as a function of depth, a similar trend can be seen in Fig. 5. Concerning the Nagyalföld, Algyő and Szolnok Formations, a very good correlation can be observed between the average values of data mentioned above (Hetényi 1992). In the Vásárhely Formation, the higher hydrocarbon potential cannot be accounted for merely by the higher quantity of the organic carbon, but it can be explained by the essential proportion of free hydrocarbons (Figs 5, 6). Significant average hydrocarbon potential determined in delta plain sediments can be attributed not only to the very high amount, but to the better quality of the organic matter. The bulk of the studied samples contain Type III kerogen with low hydrogen index (HI<100 mg HC/g TOC, Fig. 4). However, the higher hydrogen indices of the delta plain sediments (average HI = 132 mg HC/g TOC) and especially that of the Törtel Formation (average HI = 149 mg HC/g TOC) indicate Types II–III kerogen (Hetényi 1992). Furthermore, a few samples containing Types II–III kerogen with a hydrogen index of more than 150 mg HC/g TOC were also found (Fig. 4). It is known that coal has a great similarity to Type III kerogen and yields gas rather than oil, but may also generate commercial amounts of crude oil (Bertrand 1984; Bertrand et al. 1986; Durand

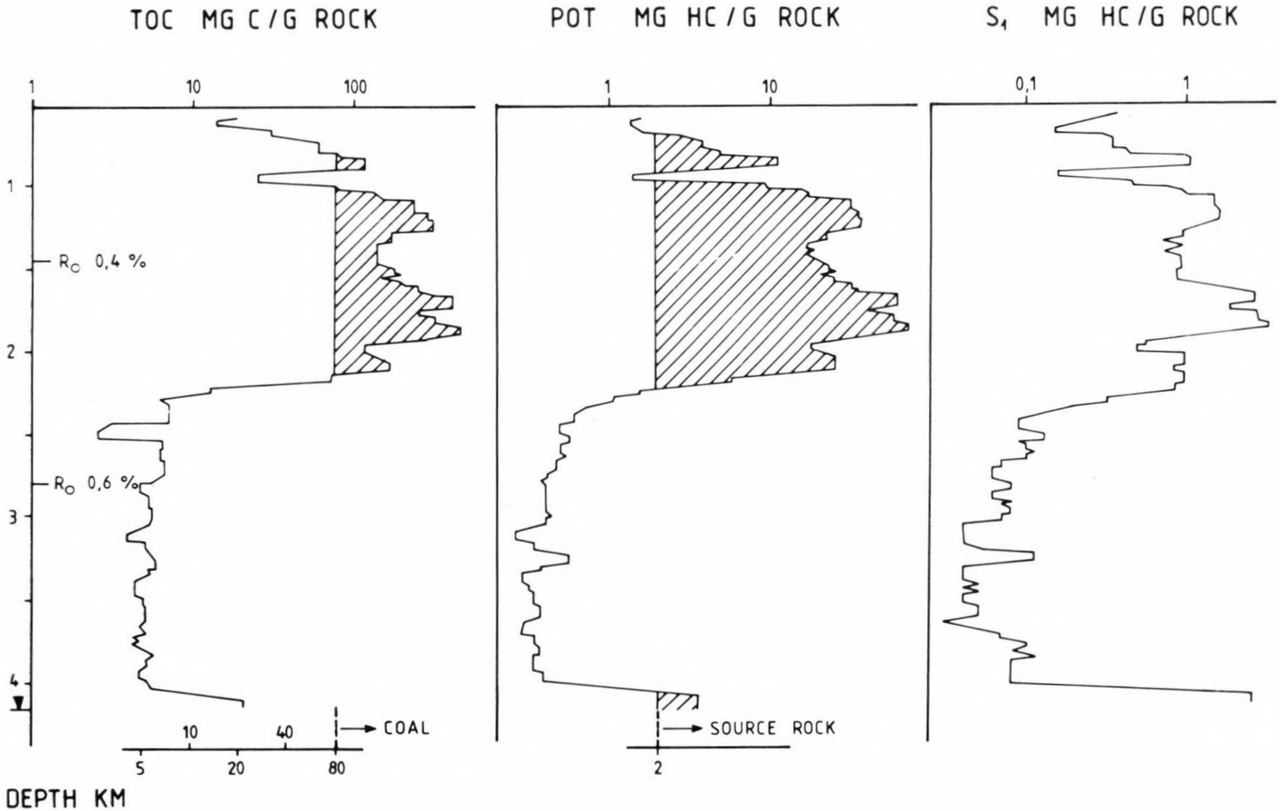

Fig. 5
Total organic carbon, hydrocarbon potential, free hydrocarbons vs. depth

and Paratte 1983). According to Tissot and Welte (1984) the generative potential for higher hydrocarbons is present in certain type of coals, depending on their liptinite content. Experimental thermal degradation of Hungarian Miocene lignites (average HI = 135 mg HC/g TOC) also resulted in liquid hydrocarbons (90–120 mg/g TOC – Hetényi and Sajgó 1990; Hetényi and Kedves 1991). It can be supposed that during their catagenesis, organic-rich sediments with frequent lignite horizons – such as in the Zagyva and Törtel Formations – prove to be good gas source rocks. However, it can be assumed that these sediments can also generate some oil during catagenesis. Unfortunately, the thermal maturation of the Törtel and Zagyva Formations has not reached the main hydrocarbon generation stage (catagenesis). The high production index (PI>0.5) and propane plus butane content of methane to butane hydrocarbons in the cutting gases (>30%) indicate the presence of migrated hydrocarbons at a depth of 2150–2350 m and 4000–4130 m (Fig. 3).

Summary and conclusion

On the basis of the results presented in this paper, the Makó-3 well contains organic matter of relatively high quantity and of fair quality. The average values of some important organic geochemical parameters are summarized in Fig. 7. In the Nagyalföld Formation, as a consequence of the fluvio-lacustrine depositional facies, Type III kerogen with a low hydrogen index (40–80 mg HC/g TOC) was formed. However, its organic carbon content was relatively significant. Good source rock of high hydrocarbon potential (>10 mg/g) was found only between 810 and 890 m, where the amount of the free hydrocarbons was also significant. Upper Pannonian delta plain sediments (Zagyva and Törtel Formations) seemed to be rich in organic matter (TOC $\leq$ 560 mg/g rock), attributed to their lignite and brown coal beds. The very high organic carbon content and the fair quality of kerogen (HI = 96–168 mg HC/g TOC) resulted in excellent hydrocarbon potential up to 67 mg HC/g rock. Both formations may be identified as good gas source rocks. However, it can also be assumed that during their catagenesis they could also generate some oil. Based on oil-source rock correlation and carbon isotope ratio of methane, the hydrocarbons trapped in the Upper Pannonian reservoirs have not been generated by the Upper Pannonian source rocks (Koncz and Etler in press).

In the Lower Pannonian sequences, consisting mainly of pelitic (Algyő Formation) and partly psammitic sediments (Szolnok Formation), small values of organic geochemical data were determined (TOC < 10 mg/g rock, HI < 100 mg HC/g TOC, HC-pot. < 1 mg HC/g rock). These formations can be regarded as poor gas-prone source rocks. However, the upper horizons of the Algyő Formation (about 100 m) contain organic matter of high quantity and better quality. Moderate hydrocarbon potential (4–5 mg HC/g rock), high values of free hydrocarbons (S1>1 mg HC/g rock) and oil indications measured in the Vásárhely Marl Formation, as well as relatively small hydrogen indices, revealed migrated hydrocarbons. All of the measured maturity parameters indicated the beginning of catagenesis at a depth of 2100–2200 m. The upper boundary of the early zone of hydrocarbon gas generation was found at a depth of 1450 m.

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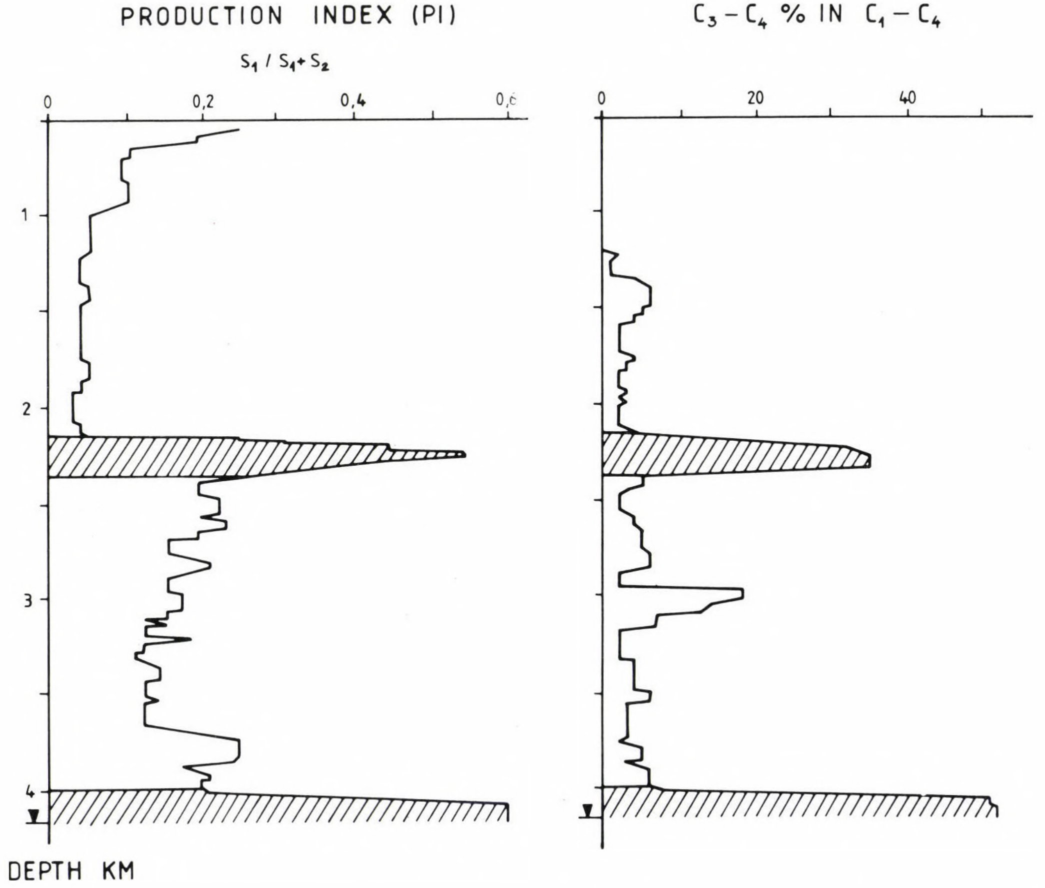

Fig. 6
Production index, propane plus butane components of methane to butane hydrocarbon gases from cuttings vs. depth

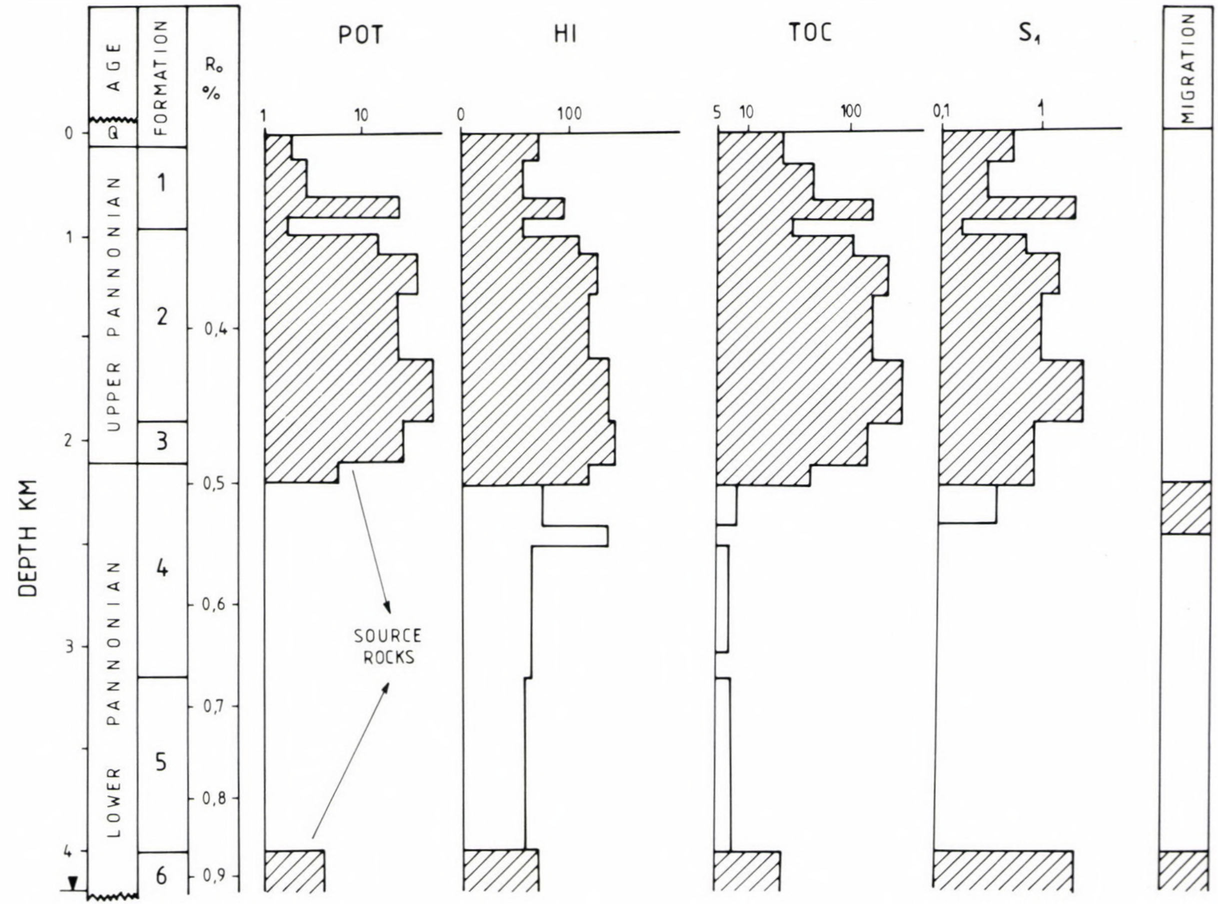

Fig. 7
Geochemical log of the Makó-3 well, obtained average values of cuttings (POT-hydrocarbon potential, HI-hydrogen index, TOC-total organic carbon content S1-free hydrocarbon content. Formations: 1. Nagyalföld; 2. Zagyva; 3. Törtel; 4. Algyő; 5. Szolnok; 6. Vásárhely)

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Relationship of the organic geochemical features of two maar-type Hungarian oil shales

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During the last 15 years several so-called maar-type oil shales have been discovered in Hungary. The craters were formed 3–5 million years ago by the final basalt volcanism in the Pannonian Basin. As a result of volcanism and specific sedimentation, the craters were filled with a special kind of organic matter. Pollen analysis indicates that this organic matter consists of planctonic algae, mainly of *Botryococcus braunii*. In our study the organic geochemical features of two maar-type oil shales (Put-30 and Gét-6 well) are compared. Geologists say that these oil shales were formed under similar geological conditions. The aim of this investigation was to find out whether the biological conditions and environment of sedimentation were also the same.

Total organic carbon content, Rock-Eval pyrolysis of kerogen and complex organic geochemical analysis of extractable organic matter were performed. There are differences in the averages of Rock-Eval pyrolysis data and TOC content between the two kerogens.

Although the average type of kerogen seems to be the same in the two boreholes, the samples of the Put-30 well contain some oil shale layers with kerogen Type I and Types I–II, while the Gét-6 well only contains Type II kerogen.

The feature of IR spectra of soluble organic matter is slightly different in some wavelength ranges in the samples of Put-30 and Gét-6 wells. Significant differences can be found in the group composition of soluble organic matters and the gas chromatograms of non-aromatic hydrocarbon fractions.

The two maar-type oil shales were formed under similar geological conditions; the above-mentioned differences in the organic geochemical features of the extractable organic matters and the kerogens, however, reflect somewhat different biological and sedimentation conditions.

Key words: Oil shale, maar-type, *Botryococcus braunii*, Rock-Eval pyrolysis, phyllocladane, pentacyclic triterpanes

Introduction

During the last 15 years, four alginite- and bentonite-filled maar-type craters have been discovered in Hungary (Fig. 1). Four to five million years ago (during the Pliocene time), in the Pannonian lake system that occupied the basin surrounded by the Carpathian Mountains, and that had already developed into a kind of archipelago, very intense and repeated volcanic eruptions disturbed an otherwise quiet and steady sedimentation. The erupting basic magma gave rise to both stratovolcanoes and the formation of special tuffrings: maars. The

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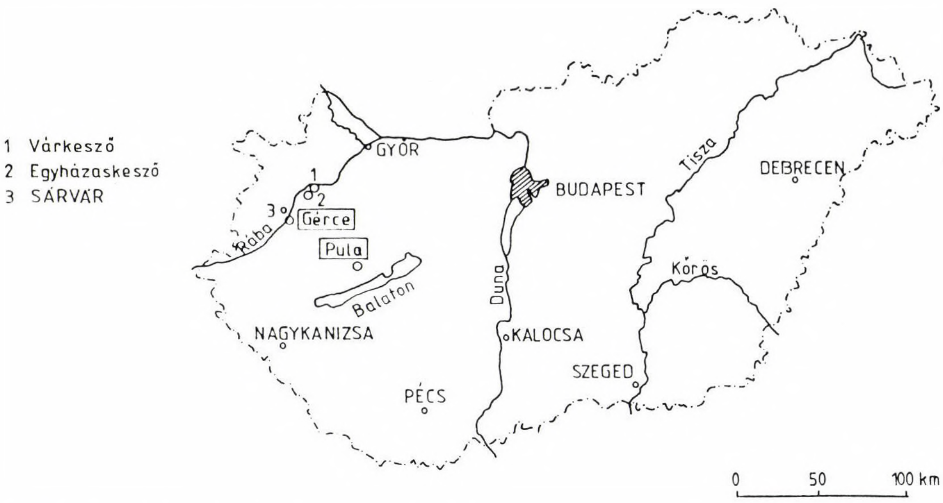

Fig. 1
Maar-type craters filled with alginite in Hungary

explosion of the tuff took place in two phases. Clayey marls and fresh water limestones formed between the two corresponding tuff complexes. After the volcanic activity ceased the crater was filled by the water of Pannonian Lake. A special kind of the sedimentation took place in water with oligohaline character, 0.3% of the maximum value of the salinity, and with a pH of 7.6. Because of the heavy weathering of the volcanic glass in the rock, the water of the crater lake became extremely enriched with macro- and micronutrients. The water in the closed, current-free crater lake may have been warmer than that of the Pannonian Lake. The hot water of postvolcanic geysers considerably heated the lake (Solti 1981). Its temperature exceeded 29 °C, as evidenced by the high aragonite content of its sediments. Very favourable conditions would have existed for planctonic life due to the great amount of nutrients liberated from the weathered crater margin. A very large amount of dead nectonic and planctonic organism was preserved in the anoxic bottom water. This was caused by the absence of benthic organisms; thus, the algae, together with the weathered volcanic material and the precipitated carbonate material, were deposited as a sapropelic mud which became the special kind of oil shale-alginite through diagenesis (Jámbor et al. 1982, Solti and Csirik 1989). As a result of the research by the Hungarian Geological Institute, four volcanic craters containing an alginite deposit are known at present. These are: Pula, Gérce, Egyházaskesző, Várkesző (Fig. 1). The deposits are small in area, 0.3–2.1 km^2, their thickness attaining a maximum of 70 m.

The aim of this study was to compare alginite deposits in maar-type craters. The samples come from the Put-30 (Pula) and Gét-6 (Gérce) wells. The lithological columns of the two wells are slightly different. The geological column of the Put-30 well (Fig. 2) shows broader variety than that of the Gét-6 well. The lithology of Gét-6 is relatively uniform; it consists of mainly laminated alginite.

Experimental

Core samples were ground in a Fritsch ball mill. The extraction of rock powders was carried out with chloroform in a Soxhlet apparatus followed by careful evaporation to provide bitumen. The IR spectra of the extracts were recorded using a Specord IR 75 spectrophotometer using the KBr disc technique, and evaluated by the baseline method. After precipitating asphaltenes with a large excess of petroleum ether b.p. 70 °C, the extracts were separated on a silica gel column by elution with petroleum ether for the saturates, benzene for aromatics and a benzene-methanol (1:1) mixture for the resin fraction. Gas chromatographic analysis of the non-aromatic hydrocarbons was performed on a HP5890 A gas chromatograph fitted with a 25 m x 0.2 mm WCOT fused silica capillary column coated with OV-1, using hydrogen carrier gas. The oven was programmed from 110 to 170 °C at 25 °C/min and 170 to 320 °C at 5 °C/min and the samples were injected in a 20:1 split mode. After removal of inorganic carbonates with dilute HCl, organic carbon contents of the sediments were determined using a LECO carbon analyzer. Pyrolysis of samples was carried out on a Rock-Eval II instrument in a helium gas stream at 300 °C for 4 min, followed by programmed pyrolysis at a rate of 25 °C/min to 550 °C (Espitalié et al. 1977).

Results and discussion

The aim of our study was to find if there is any difference between the organic geochemical characters of maar-type oil shales formed under the same geologic conditions and, if so, what sort of difference it is. The features of chloroform soluble (bitumen) and insoluble (kerogen) organic matter were compared with samples from the Pula-30 (Put-30) and the Gérce-6 (Gét-6) wells.

The total organic carbon (TOC) content of samples from the Put-30 and Gét-6 wells varies between 2 and 41.5% (Table 1) and 2.4 and 9.4% (Table 2), respectively. The corresponding chloroform-soluble bitumens ranges are 0.1–11.5% and 0.6–3.8%. The amount of HCl acid insoluble residue is 56–70% in samples of the Gét-6 well, reflecting the uniformity of lithology. Due to the wide range of lithological components, the amount of the insoluble residue of samples from the Put-30 well varies between 28 and 78%. Because of the different lithological columns of the two wells, only the average values of the laminated alginite sections were taken into consideration (Table 3).

The alginite members of both the maar-type oil shales (Put-30: 10.5–48.5 m and Gét-6: 16.3–65.0 m) seem to contain kerogen of type II. Plotting the average

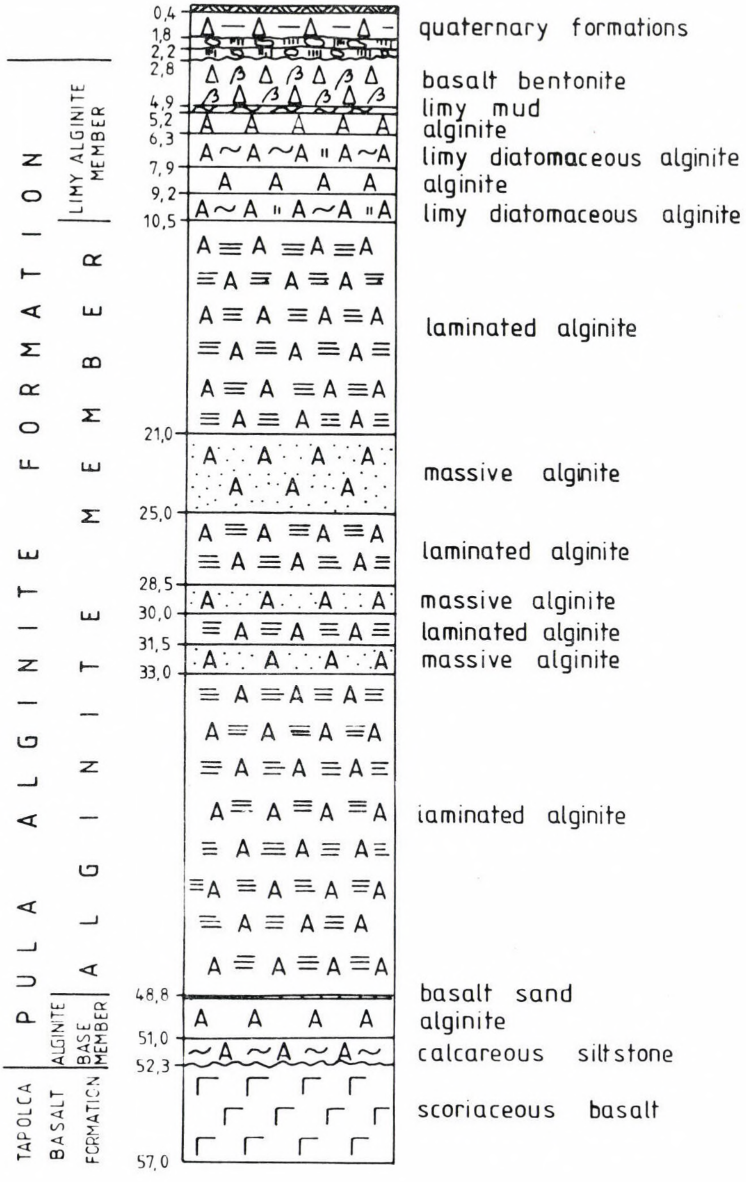

Fig. 2
Geological column of the Put-30 well (Solti et al. 1989)

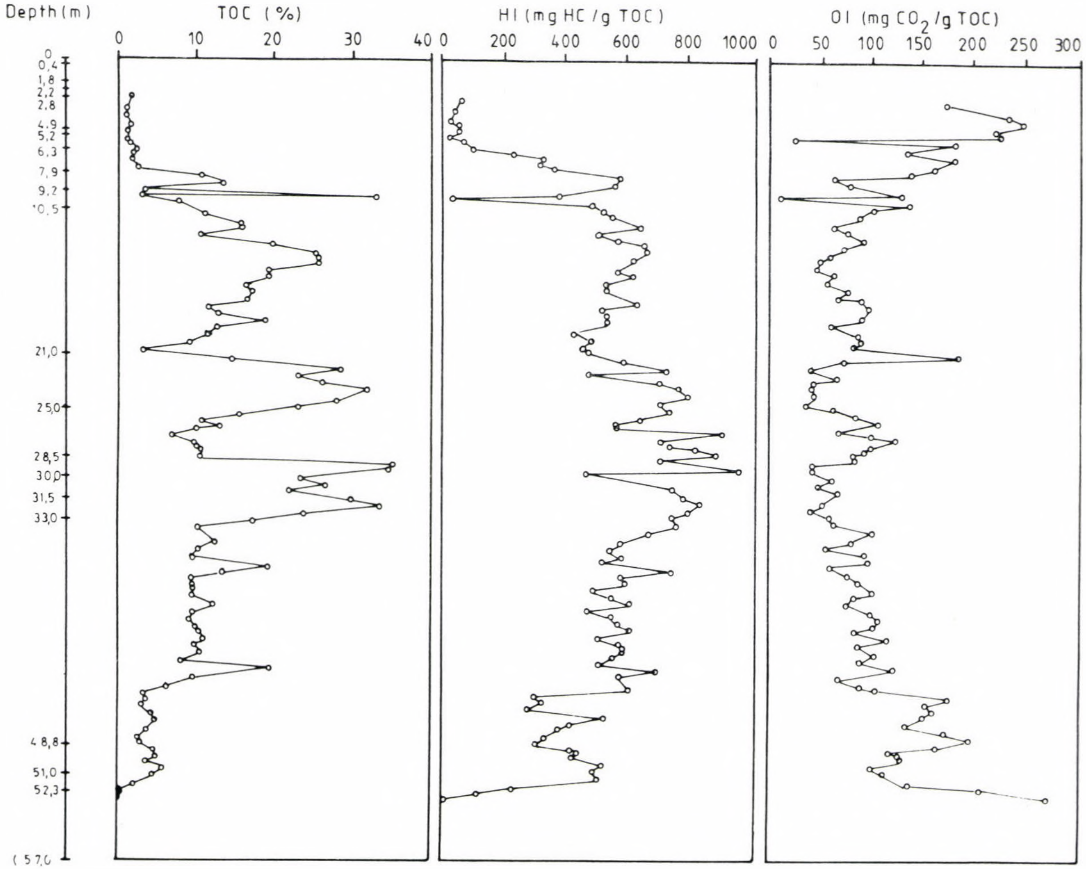

Fig. 3
Quantity and quality of the kerogen in Put-30 well

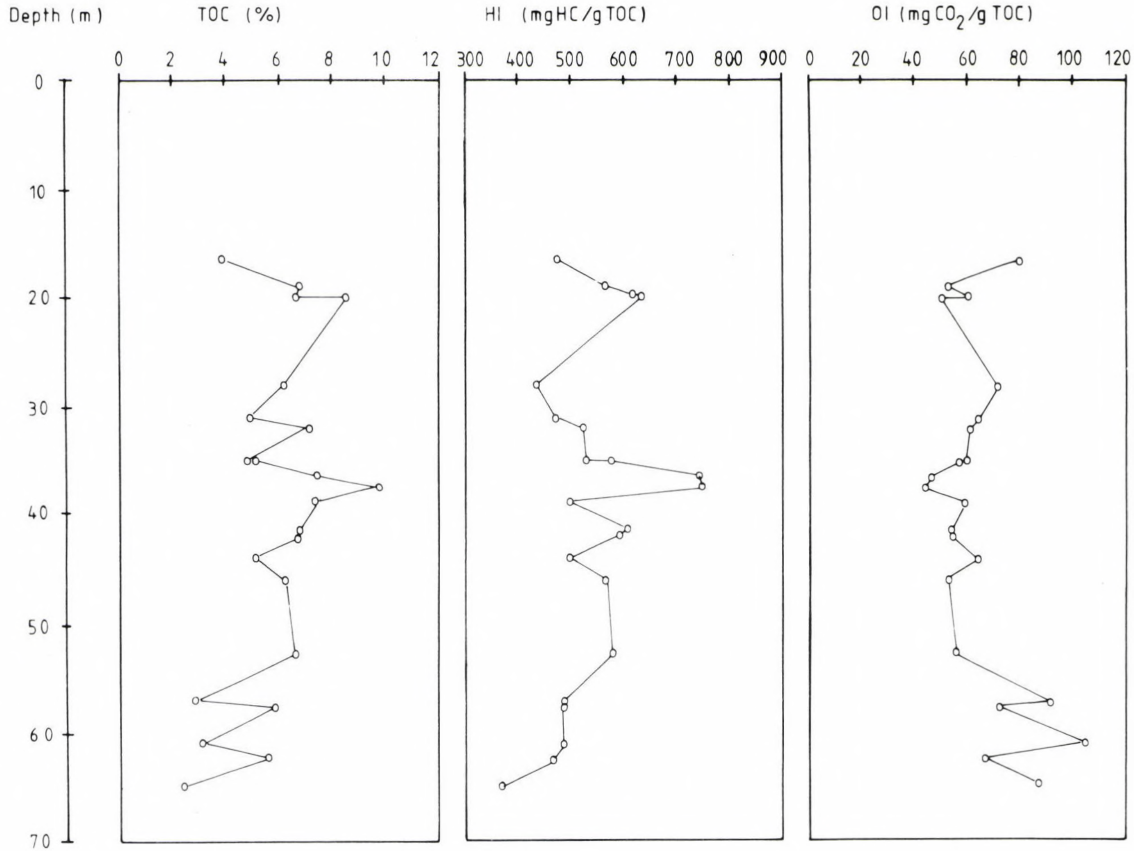

Fig. 4
Quantity and quality of the kerogen in Gét-6 well

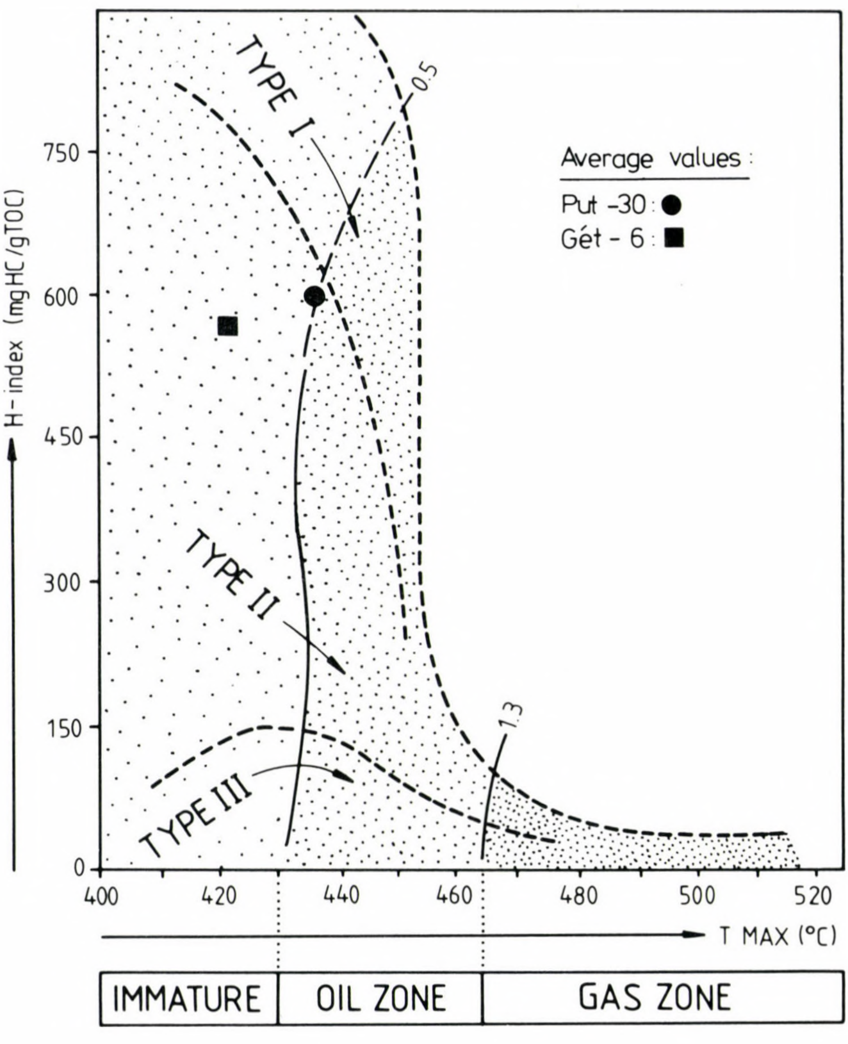

Fig. 5
Classification of the kerogens of alginites on the HI-T_{max} diagram

Table 1
Organic geochemical data of the kerogen in Put-30 maar-type oil shale

Depth (m)	No. of samples	TOC (%)	T_{max} (oC)	S1 mg HC/g rock	S2 mg HC/g rock	HC-pot mg HC/g rock	S3 mg CO2/g rock	S2/S3	HI mg HC/g TOC	OI mg CO2/g TOC	PC/TOC %	Type of kero-gen
2.8– 4.5	4	1.5	440	0.13	0.76	0.89	3.15	0.24	49	214	5	III
4.9– 5.2	1	1.5	442	0.07	0.87	0.94	3.29	0.26	59	223	5	III
5.2– 6.0	3	5.4	444	0.18	1.51	1.69	2.83	0.53	58	112	5	III
6.3– 7.5	4	2.5	439	0.29	7.49	7.78	3.92	1.89	293	155	25	II
7.9– 8.5	2	12.3	440	4.72	68.41	73.08	9.04	8.07	562	72	50	II
9.2–10.0	3	13.4	440	0.98	14.37	15.35	4.49	3.19	300	94	26	II
10.5–20.5*	21	15.8	436	12.07	88.67	100.74	11.19	7.63	549	80	52	II
21.0–24.5	8	27.1	441	13.47	207.42	220.89	13.28	14.63	702	51	63	I–II
25.0–28.0*	9	11.2	439	6.72	84.02	90.74	10.43	7.78	725	95	65	I–II
28.5–29.5	3	31.0	442	9.88	265.54	275.42	15.39	17.10	750	51	73	I–II
30.0–31.0*	3	26.1	441	9.38	201.61	210.99	14.98	13.48	769	58	67	I–II
31.5–32.5	3	25.1	442	9.36	189.77	199.13	13.55	13.74	753	56	66	I–II
33.0–48.5*	32	9.2	438	2.71	36.36	39.07	9.20	5.16	514	97	44	II
48.8–49.0	1	5.0	439	1.37	21.48	22.85	5.86	3.66	429	117	38	II
49.0–50.5	4	5.0	439	1.61	24.82	26.43	5.84	4.18	487	118	43	II
51.0–51.9	3	1.1	384	0.30	2.28	2.58	2.60	0.88	124	237	17	III

Symbols: * – laminated sections

Table 2
Organic geochemical data of the kerogen in Gét-6 maar-type oil shale

Depth m	TOC %	T_{max} °C	S1 mg HC/g rock	S2 mg HC/g rock	HC-pot mg HC/g rock	S3 mg CO2/g rock	S2/S3	HI mg HC/g TOC	OI mg CO2/g TOC	PC/TOC %	Type of kerogen
16.3	3.88	430	2.50	18.45	20.95	3.11	5.93	475	80	45	II
18.8	6.81	428	9.89	38.91	48.80	3.66	10.63	571	53	60	II
19.8	6.74	433	8.10	41.84	49.94	4.13	10.13	620	61	62	II
19.0	8.48	432	9.23	54.02	63.25	4.67	11.56	637	55	62	II
28.0	6.15	402	9.72	27.16	36.88	4.45	6.10	441	72	50	II
31.2	4.90	423	5.69	23.21	28.90	3.23	7.18	473	65	49	II
32.0	7.07	425	8.29	37.21	45.50	4.41	8.43	526	62	54	II
34.8	4.90	426	5.80	26.21	32.01	2.94	8.91	534	60	54	II
35.0	5.29	424	7.08	30.54	37.62	3.10	9.85	577	58	59	II
36.5	7.55	432	10.42	56.39	66.81	3.58	15.75	746	47	74	II
37.3	9.84	431	15.46	73.63	89.09	4.35	16.92	748	44	75	I–II
38.8	7.40	391	14.83	37.21	52.04	4.44	8.38	502	60	59	II
41.5	6.19	430	7.03	37.68	44.71	3.46	10.89	608	55	60	II
42.0	6.16	429	7.59	36.73	44.32	3.42	10.73	596	55	60	II
44.0	5.19	421	7.33	26.01	33.34	3.42	7.60	501	65	53	II
46.0	7.31	408	14.33	41.67	56.00	4.01	10.39	570	54	64	II
52.5	6.65	430	6.55	38.78	45.33	3.80	10.20	583	57	57	II
54.5	5.82	430	12.18	58.61	70.79	4.66	12.57	1007	80	100	I
56.7	2.87	419	3.43	14.00	17.43	2.65	5.28	487	92	51	II
57.3	5.86	411	7.04	28.48	35.52	4.25	6.70	486	72	51	II
61.3	3.18	408	3.34	15.56	18.90	3.36	4.63	489	105	49	II
62.3	5.63	415	7.47	26.50	33.97	3.82	6.93	470	67	50	II
65.0	2.39	412	2.04	8.91	10.95	2.09	4.25	372	87	38	II

hydrogen indices against the average T_{max}-values (Fig. 5), both of the kerogens are in the immature zone. The Gét-6 alginite falls into the evolution field of the good-quality kerogen of type II (HI: 566 mgHC/gTOC, T_{max}: 421 °C), the Put-30 alginite can be found the border of kerogen of type I and II (HI: 598 mg HC/gTOC, T_{max}: 438 °C).

The PC/TOC values (40–80%) indicate well-convertible organic matter in both of the wells, and the S2/S3 ratios reveal a considerable oil potential (Tables 1 and 2).

While the average values of the organic geochemical data show kerogens of similar quality in Put-30 and Gét-6 wells, comparing data measured on individual layers, some differences can be observed between the alginites deposited in the two crater lakes.

Table 3
Average values of the organic geochemical parameters of the soluble organic matter

Locality	Depth (m)	TOC (%)	Insoluble mat. (%)	Bitumen (%)	ΣHC/NSO	HCsat/HCar
Put-30	2.8– 4.4	1.5	66.7	0.076	0.27	1.53
	4.9– 5.2	1.5	56.4	0.064	0.10	1.59
	5.2– 6.0	5.4	72.5	0.106	0.28	0.43
	6.3– 7.5	2.5	37.2	0.170	0.18	0.55
	7.9– 8.5	12.3	67.5	1.494	0.11	0.42
	9.2–10.0	13.4	32.1	0.447	0.09	0.50
	*10.5–20.5	15.8	61.9	4.595	0.08	0.34
	21.0–24.5	27.1	70.7	4.373	0.05	0.91
	25.0–28.0	11.2	46.6	2.202	0.03	0.52
	28.5–29.5	31.0	66.6	4.304	0.03	0.75
	*30.0–31.0	26.1	64.5	2.759	0.04	0.64
	31.5–32.5	25.1	62.8	3.864	0.03	0.54
	*33.0–48.5	9.2	42.8	0.733	0.08	0.29
	48.8–49.0	5.0	33.1	0.237	0.08	0.33
	49.0–50.5	5.0	30.4	0.338	0.08	0.37
	51.0–51.9	1.1	42.4	0.092	–	–
Put-30	laminated	12.4	51.0	2.28	0.07	0.35
Gét-6	alginite	5.8	62.6	1.72	0.04	0.70

* laminated sections

Nearly all of the samples from Gét-6 represent kerogen of Type II; their hydrogen indices vary between 450 and 600 mg HC/gTOC. On the other hand, within the alginite members of Put-30, three main sections can be detected. The kerogen in the lower laminated section and in the upper one is Type of II, the average HI=514 and 549 mg HC/gTOC, respectively. The kerogen in the middle section – which consist of laminated and massive alginites (Fig. 2) –

proved to be an excellent-quality, Types of I–II (HI 700 mg HC/gTOC). This periodicity of the layers and that of their geochemical data cannot be noticed in the Gét-6 well.

The small difference between the quantity and quality of kerogens in two crater lakes, as well as the above-mentioned periodicity of the Pula alginite can be attributed to: 1. the different growth-conditions of *Botryococcus braunii* algae and 2. the different preservation-rate of the organic matter in two maar-type volcanic craters.

1. Though the conditions in both of the crater lakes were favourable for growth of the algae, in the Pula crater their massive growth was enchanced by the carbonic acid water of postvolcanic geysers which periodically heated the lake and resulted in the accumulation in various horizons of good-quality alginite (Solti 1981). The alginites deposited in crater lakes unaffected by hot springs (e.g. Gérce) proved to be of a lower quality (Hetényi 1985).

2. Furthermore, in the Pula crater the biologically inactive depositional environment promoted a good preservation of the organic matter and resulted in kerogen of high hydrogen index. The organic matter in the Gérce oil shale was deposited in a biologically active, probably alkaline environment, and the algal material was reworked by microorganism.

The above-mentioned conclusions can be supported by the quality of kerogens isolated from Pula and Gérce alginites (Put-7 and Gét-2 wells) and by the results of their thermal degradation. The H/C atomic ratio of kerogens isolated from Pula and Gérce alginites were 1.72 and 1.68, respectively (Hetényi 1985). Sixty-five percent of the organic carbon content (Pula kerogen) and 50% of the organic carbon content (Gérce kerogen) could be converted. Their thermal assay yielded 700 (Pula) and 440 (Gérce) mg oil/gTOC (Hetényi et al. 1991).

There are significant losses (L) in bitumen compositions after separation by column chromatography. The losses being retained on the column consist mainly of long chain alkyl esters, acids and ketons as deduced from their IR spectra. This suggests that the retained components belong to the NSO fraction. There are slight differences in the average values of the group compositions between the two wells. The differences are stronger as far as the aromatic HC and asphaltene (A) fractions and losses are concerned (see the table below). This finding is supported by the HC_{sat}/HC_{ar} ratio (Table 3).

	HC_{sat}	HC_{ar}	R	A	L	SHC	NSO
Put-30	1.6	5.1	56.5	18.6	18.0	6.7	74.5
Gét-6	1.3	2.2	58.7	27.6	10.1	3.5	68.9

R: resin

The differences of the basic data and the group compositions of the samples studied suggest a slight variation of the organic facies and sedimentation conditions between the two wells. This assumption seems to be supported by the IR spectra of bitumens.

In the IR spectra of bitumens the long-chained aliphatic compounds containing carbonyl functional groups dominate. A low level of CH_3 groups relative to CH_2 groups is characteristic on the basis of the shape of the spectra in the range of 2960–2850 cm^{-1} and the weak absorption at 1380 cm^{-1} wavenumber. The IR spectra of bitumens from both wells are very similar to that of PRB of outer walls of the alga *Botryococcus braunii* (Largeau et al. 1980, 1984). According to the palynological investigations the organic matter of both wells consists mainly of the alga *Botryococcus braunii*. There are some differences in the site of the occurrence and the shape of certain absorption bands in IR spectra of bitumens between samples of Put-30 and Gét-6 well. The rocking vibration of $(CH_2)_n$ groups, where $n \geq 4$ at 720 cm^{-1} are almost always split in Put-30 samples, while this phenomenon does not occur in Gét-6 samples (Figs 6 and 7). The bands of carbonyl groups are present as ester (1735 cm^{-1}) and acids, ketons, etc. (1710 cm^{-1}) as well in the majority of Put-30 samples (Fig. 6). In samples of the Gét-6 well the bands of the carbonyl groups are present only at 1710 cm^{-1} as acids and ketons (Fig. 7). The bands arising from naphtenic skeletal vibration at 960 cm^{-1} are very weak or not present in IR spectra of samples from the Put-30 well, while in Gét-6 samples the band intensity is slightly stronger, even very strong at the depth of 46 m (Fig. 7).

In the HCsat fraction gas chromatograms of bitumens from both well there is a dominant maximum at n-C_{29}, in some cases at n-C_{31}. There are no significant differences in the maturity parameters calculated from the n-alkane distribution between the two wells. The ratio of n-C_{22}-/n-C_{23}+ (the sum of n-alkanes below n-C_{22}/the sum of n-alkanes above n-C_{23}) is very small (<0.1). The average carbon preference indicates (CPI) calculated over the n-C_{22} and n-C_{32} range is 11.9 for the laminated section of the Put-30 well and 8.5 for the Gét-6 well. As far as the feature of gas chromatograms of HCsat fractions are concerned there are significant differences between samples of the Put-30 and Gét-6 wells. The gas chromatograms of Put-30 samples are very simple and consist mainly of n-alkanes above n-C_{20} (Fig. 8). Some gas chromatograms contain a tetracyclic diterpenoid biological marker compound identified as phyllocladane, which derived from the Coniferospida class of Gymnosperms (Noble et al. 1985, 1986). The presence of this compound in some layers of the Put-30 well indicates the higher terrestrial input to the original organic matter. This compound is absolutely absent from samples of Gét-6. There are several less intense peaks between n-C_{27} and n-C_{34} in gas chromatograms of samples from Gét-6 (Fig. 9). These compounds are most probably members of the 17β (H) 21β (H) hopane series. Ensminger et al. (1974, 1977) suggest these pentacyclic triterpanes originate mainly from bacteria, blue-green algae and protozoa. In our case the presence of these compounds shows somewhat more bacterial contribution to the initial organic matter.

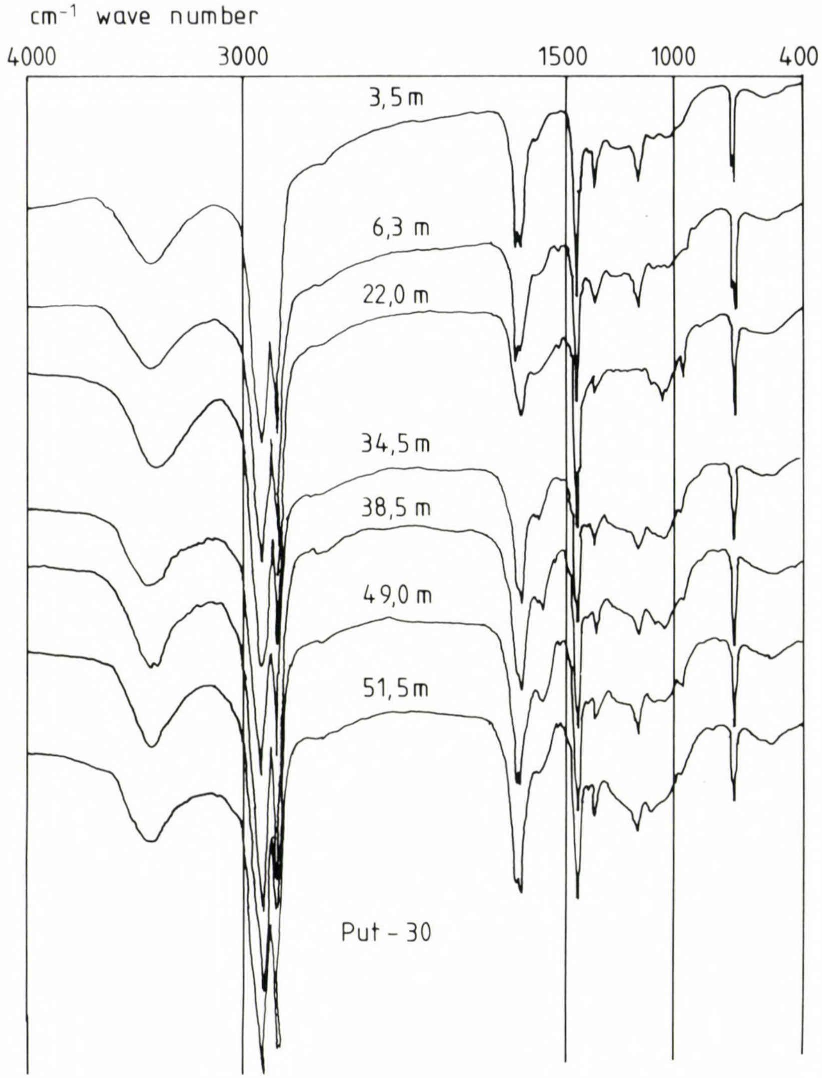

Fig. 6
The IR-spectra of bitumens from alginite of the Put-30 well

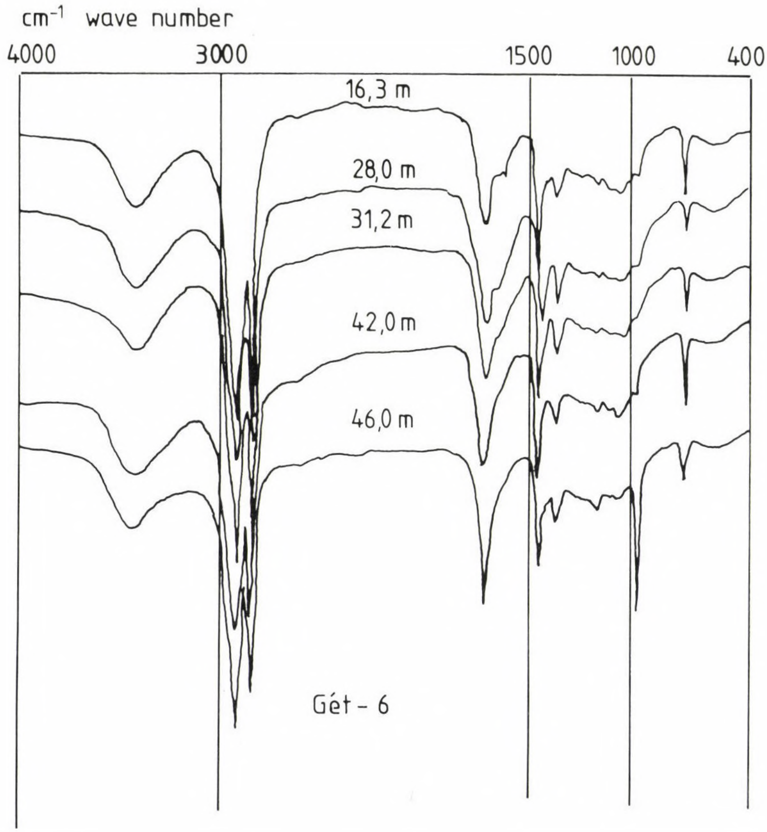

Fig. 7
The IR-spectra of bitumens from alginite of the Gét-6 well

Conclusions

1. The special conditions of the formation and sedimentation of the maar-type oil shales (alginites) of Pula and Gérce are well reflected by all the measured organic geochemical parameters.

2. The special formation conditions of the Pula crater lake are confirmed by the averages of the organic geochemical data referring to each lithological member and the differences of the gas chromatograms (presence of phyllocladane in some layers → higher terrestrial plant input).

3. Comparing the averages of the organic geochemical data of the laminated alginite members in the two wells, the organic matter seems to be different, both in quantity and quality. These differences are also shown by palynological investigations. Though the organic matter of both wells originated from

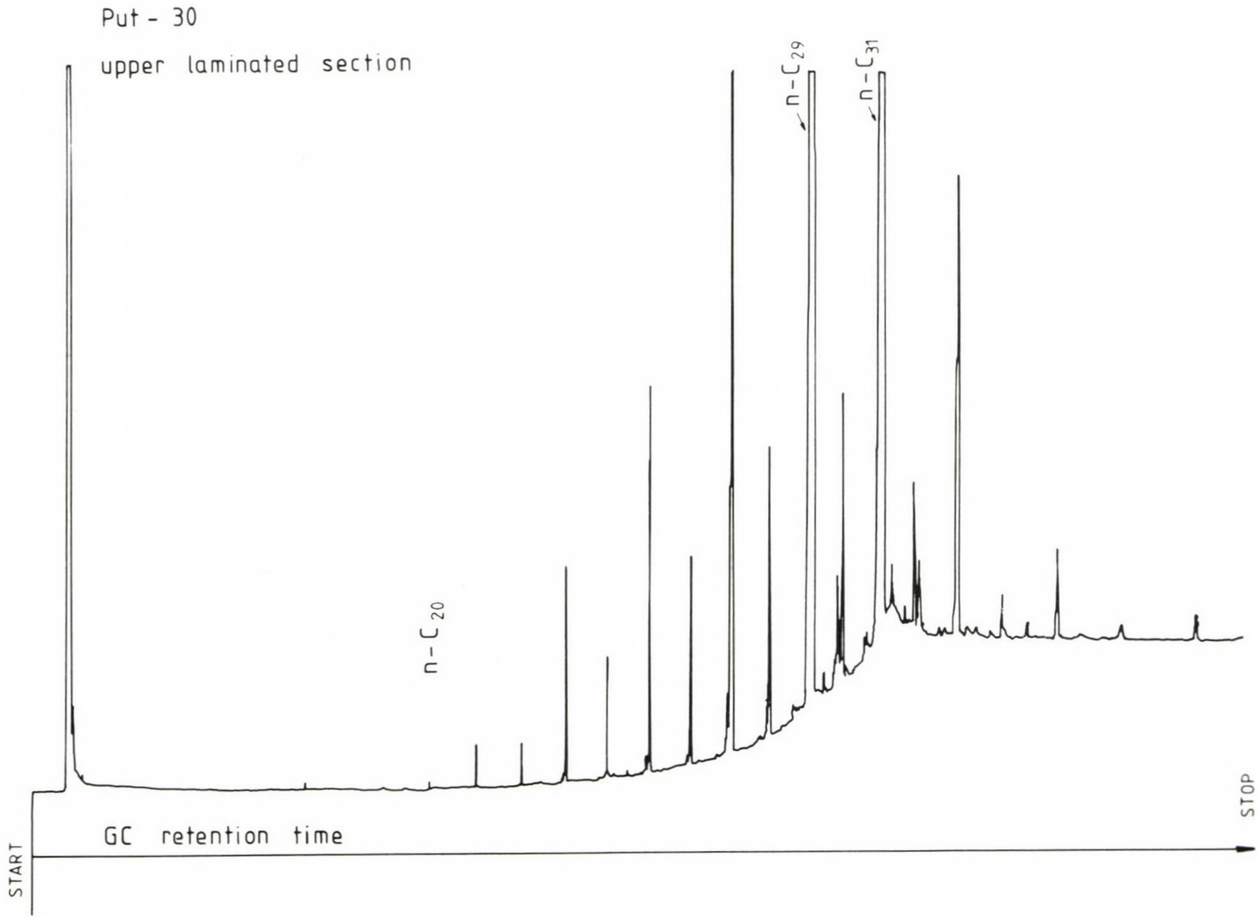

Fig. 8
Gas chromatogram of the saturated hydrocarbon fraction from $CHCl_3$ extract of the upper laminated section of the Put-30 well

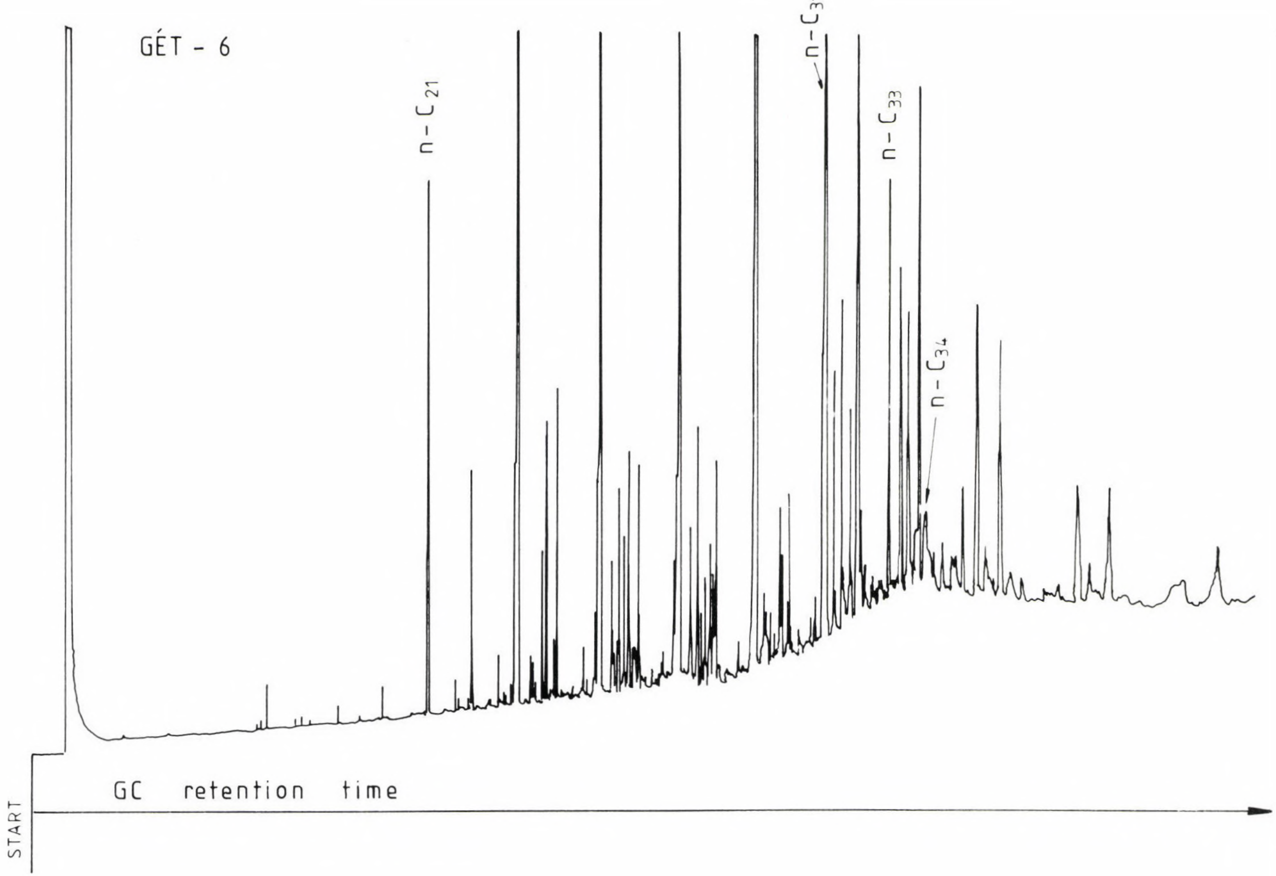

Fig. 9
Gas chromatogram of the saturated hydrocarbon fraction from $CHCl_3$ extract of the Gét-6 well

planctonic algae, mainly of *Botryococcus braunii*, there is a considerable bacterial contribution in the alginite of Gérce (see the range between n-C_{27} and n-C_{34} in gas chromatograms).

4. The two maar-type oil shales were formed under the same geological conditions, but the differences in quantity and quality of the organic matter can be attributed to the somewhat different biological and sedimentation conditions, primarily to the effect of the postvolcanic activity which periodically heated the water of the Pula crater lake and resulted in kerogen of Types I–II.

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Distribution of sulphur in Transdanubian (Hungary) and Middle European Brown Coals

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Distribution of sulphur in brown coal samples derived from the Czech and Slovak Republics, the eastern part of Germany and Poland was compared with that of different Transdanubian brown coals from Hungary. The following order: organic sulphur>pyritic sulphur>sulphate sulphur is characteristic of the majority of coal samples. A disadvantageous feature of the examined Hungarian brown coals is that their total sulphur content (mean 5%) is higher than that of foreign coals; in addition, more than half of its total sulphur content is organic sulphur. In order to use suitable methods and technology to reduce the quantity of sulphur dioxide emitted into the air during combustion, it is important to know the distribution of sulphur in coals.

Key words: Total sulphur content, pyritic sulphur, organic sulphur, sulphate sulphur, brown coal

Introduction

There are two major types of sulphur compounds in coal: inorganic sulphur and organic sulphur. The inorganic sulphur is usually present as iron sulphides (pyrite or marcasite, greigite, pyrrhotine) or some rarely occurring sulphide minerals (chalcopyrite, arsenopyrite, etc.) and sulphate (gypsum, anhydrite, barite etc.).

Sulphides can be categorized as either syngenetic or epigenetic in origin. Syngenetic pyrite was formed during the deposition of peat and during early diagenesis (humification). It is usually small in size (1–10 0μm) and finely dispersed throughout the coal. Epigenetic pyrite was incorporated into the coal at a later time, after partial consolidation and compaction. A number of researchers have described the occurrences of iron sulphides in different bituminous coals. Some of these studies have focused on shapes and sizes of "pyrite" grains and have resulted in schemes or characterization of the sulphide grains (Wiese and Fyfe 1986; Frankie and Hower 1987).

One of the most widely accepted theories for the origin of syngenetic pyrite is that pyrite was incorporated into sediments by the action of bacterial colonies (Kaplan et al. 1963; Berner 1964, 1984; Klesment and Urov 1985).

Isotope measurements give further information about the origin of sulphur occurring in several chemical forms in coal (Smith and Batts 1974; Price and

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Shien 1979; Hackley and Anderson 1985). Sulphur isotope values for peat sulphate, plant sulphur, peat pyrite and peat organic sulphur corroborate the hypothesis that microorganisms reduce sulphate to hydrogen sulphide, which in turn reacts with available ferrous iron or organic matter to produce pyrite and organic sulphur, respectively (Cassagrande 1987).

Neavel (1966) demonstrated that FeS_2 in peat can only form by bacterial activity, since on the basis of reaction kinetics, there is insufficient energy for a purely chemical reduction of sulphates to disulfides. According to the studies a metastable iron monosulfid precipitated from the H_2S and the soluble Fe-ions, which during early diagenesis transforms to pyrite.

Casagrande et al. (1977) studied the origin of sulphur in coals by comparing fresh-water and marine peat. The total amount of sulphur found in coal can be incorporated during the peat-forming stage. Similarly other workers (Teichmüller 1962; Williams and Keith 1963) found that sulphur content of marine peat higher than that of fresh-water peat. The authors concluded that environmental conditions, especially pH, appear to have a dramatic effect on pyrite content. Cecil et al. (1979) stated that high acidity favors the development of low ash and low sulphur coals because of leaching and limited bacterial activity during the peat stage. A high content of calcium leads to a high degree of bacterial degradation of the plant remains and to bacterial reduction of sulphates, resulting in coals with high amounts of collinite and pyrite (Stach et al. 1982).

Sedimentologically, the Provence coal is highly comparable to the Italian Sulcis or the Spanish Berga coals which were deposited in fresh-water, calcium-rich environments with a low rate of deposition. These coals show similar properties as marine-influenced coals. Owing to a high pH, due to calcium, bacterial activity is accelerated, resulting in increased degradation of plant remains, and there is a depletion of soluble iron. The most calcium-rich coals are remarkably high in organic sulphur and syngenetic pyrite. An extreme example of a calcium-rich coal is the Rasa coal of Istria (Yugoslavia), which carries 11% organic sulphur (Boudou et al. 1987).

Szádeczky-Kardoss (1952) had already published the fact that the pH of the water in the swamps would be higher owing to the low humic acid content (1), marine influence (2), and the dissolution of calcium of surrounding limestones (3). In Hungary the third influence is characteristic of many Transdanubian coal occurrences.

H_2S formed from microbial reduction of sulphate reacts with ferrous ion to produce pyrite and with organic matter to form organic sulphur. In their experiments Casagrande et al. (1979) demonstrated that not only is $H_2{}^{35}S$ incorporated in peat as organic sulphur, but elemental sulphur also (Casagrande and Ng 1979). Mango (1983) showed that H_2S reacts with certain carbohydrates yielding a variety of organosulphur compounds, and that sulphate ions oxidize the methyl group to carboxyl in the presence of hydrogen sulphide (Toland 1960).

The organic sulphur compounds found in coals have been categorized according to functionality: sulphide or thio-ether (-S-), disulphide (-S-S-), thiol or mercaptan (-SH) and thiophene.

In British coals the mean organic sulphur content is 0.8% (Wandless 1955). Bhatia (1978) referred to the work by Yancy and Fraser (1929) in which they showed the relative proportions of important sulphur groups present in the coals:

pyrite sulphur: 0.02 to 2.61%; av. = 1,00%
sulphate sulphur: 0.01 to 0.32%; av. = 0.05%
organic sulphur: 0.23 to 7.90%; av. = 1.21%
total sulphur: 0.44 to 9.01%; av. = 2.29%.

According to Attar, about 30–40% of the organic sulphur in lignite is thiolic and the rest is thiophenic. Most of the organic sulphur in coal is in the form of thiophenic structures and aromatic and aliphatic sulphides. The relative abundance of the sulphur groups in bituminous coal is estimated as 50:30:20% (Attar and Corcoran 1977; Attar and Dupuis 1981). Many sulphur heterocycles with 3–6 rings were identified in a coal tar and in coal liquid vacuum residue (Nishioka et al. 1986).

The Provence coal contains thermally labile organic sulphur compounds, such as thiols and aliphatic sulphides, and thermally stable organic sulphur compounds, such as aromatic and high molecular weight compounds. A small amount of organic sulphur would be bound with trace metals (Boudou et al. 1987).

One of the major problems in coal desulphurization is the removal of organic sulphur from coal matrix. Chemical methods, such as hydrogenation or hydrodesulphurization, are effective for organic sulphur removal. However, these methods are very costly, since they operate at high temperatures and pressure. Microbial methods seem to be quite promising for the removal of organic sulphur (Kargi 1984) and of pyrite sulphur from coal (Butler and Kempton 1986).

Samples and analytical methods

Thirty-seven brown coal samples from different Transdanubian mines (Fig. 1) were examined.

The determination of the total sulphur content was carried out with Eschka-mixture (2 parts by weight MgO and 1 part by weight anhydrous Na_2CO_3) at 800 °C. The sulphate precipitated with $BaCl_2$ was determined by gravimetric analysis as $BaSO_4$.

Sulphur content of disulphide in coal samples was reduced by nascent hydrogen to hydrogen sulphide in the presence of Cr(II)-ions. The hydrogen sulphide originated from reduction was bubbled through a cadmium acetate solution and the sulphur content of disulphide was determined by iodometry.

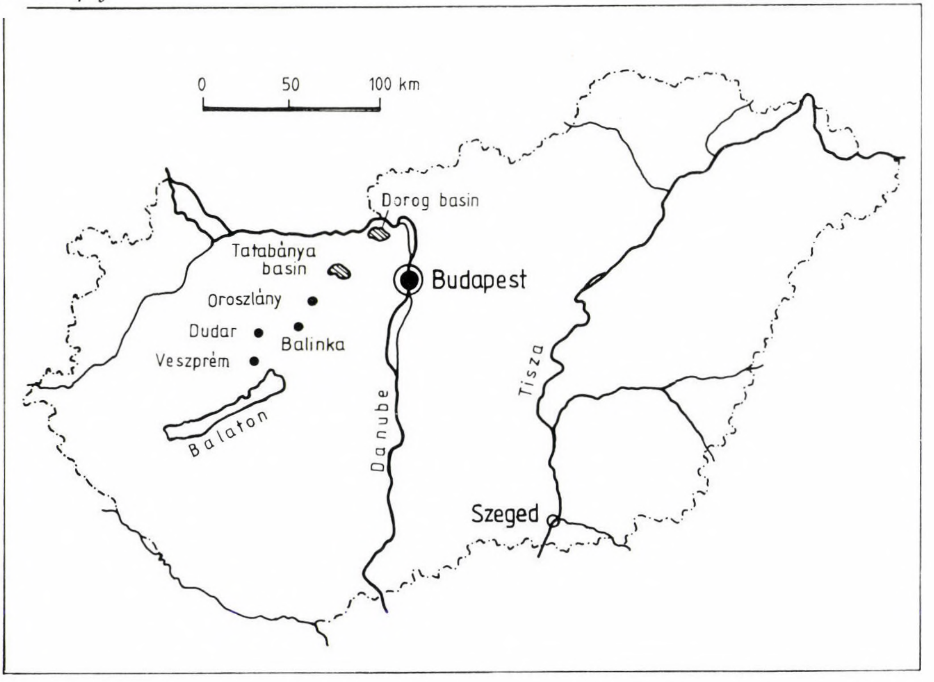


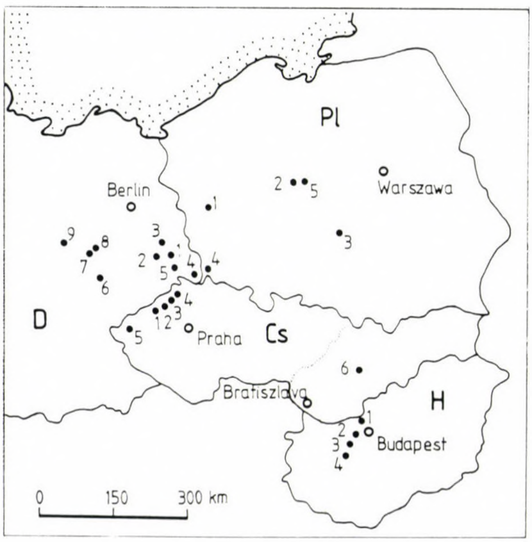
Fig. 1
Location of the Transdanubian brown coal mines

The measuring procedure for the sulphur content of sulphate is as follows: Sulphate in coal samples was extracted by hydrochloric acid and determined by gravimetric analysis as $BaSO_4$.

In earlier years, on the basis of international agreements (among countries of the former Comecom), representative foreign coal samples (which were routinely collected by coal-mine geologists) were analyzed in the Central Mining Searching Institute's Laboratory in Budapest (1985), but data were unpublishable. The method of pyritic and sulphate sulphur determination is the same for the Central European brown coal samples (Fig. 2) as stated above. Determination of total sulphur was carried out by burning the coal in oxygen flow at high temperature (1200 °C) in the presence of iron powder. Sulphur dioxide originated from sulphur compounds of coal samples was bubbled through water containing starch, which was continuously titrated by iodine solution.

Fig. 2 →
Location of the Middle European brown coal mines. **Cs** : 1. M. Gorkij; 2. Merkur; 3. Pres. K. Gottwald; 4. Charbarovice; 5. Juiri Antonin; 6. Handlova; **D:** 1. Spreetal; 2. Klettwitz; 3. Schlabendorf-S; 4. Berzdorf; 5. Puschwitz; 6. Espenhain; 7. Delitzch; 8. Goitsche; 9. Nachterstedt; **Po:** 1. Seniawa; 2. Konin-Patnów; 3. Belchatów; 4 .Turów; 5. Konin-Lubstów; **H:** 1. Dorog basin; 2. Tatabánya basin; 3. Oroszlány area; 4. Veszprém area


0
500
1000 km


Pl
Warszawa
Berlin
D
Praha
Cs
Bratiszlava
H
Budapest
0
150
300 km

Results

Data of different Transdanubian brown coal samples studied by us are shown in Table 1. The distribution of sulphur in foreign brown coals (samples were analyzed in the Central Mining Searching Institute's Laboratory in Budapest) can be found in Table 2 (Kasbohm 1988).

Table 1
Range and means of forms of sulphur in Transdanubian brown coal samples

Coal fields	Mines	Number of samples	Age	Total sulphur	Pyritic sulphur (wt. %)	Sulphate sulphur	Organic sulphur dif.
Dorog basin	Mogyorós-bánya	2	Oligocene	4.4–4.7 4.6*	1.2–2.3 1.8*	0.2–0.6 0.4*	2.2–2.6 2.4*
	Lencsehegy	11	Eocene	2.9–9.2 5.2*	<0.1–2.5 1.0*	<0.1–1.1 0.3*	2.4–5.0 3.9*
Veszprém area	Balinka	4	Eocene	4.7–8.3 6.3*	0.6–4.3 2.2*	0.1–1.2 0.4*	3.4–3.9 3.6*
	Dudar	3	Eocene	5.3–6.7 6.2*	0.4–2.6 1.6*	0.2–0.9 0.4*	3.6–4.5 4.2*
Tatabánya basin	Nagyegyháza	13	Eocene	3.1–6.6 5.1*	0.1–4.3 0.9*	<0.1–0.4 0.1*	1.8–5.2 4.0*
Oroszlány area	Márkus-hegy	4	Eocene	3.3–4.3 3.8*	0.4–1.3 0.8*	0.1–0.2 0.1*	2.2–3.2 2.8*

* average

Conclusion

The total sulphur of Transdanubian brown coals is higher than that of Central European and Australian ones, as well as that of Texas lignites (Pápay 1987–88). On the basis of sulphur distribution it can be established that more than half of the total sulphur content of our coal samples derives from organic sulphur.

A similar tendency can be seen among Middle European brown coals, though a majority of samples are younger than the Hungarian ones. The total sulphur contents of foreign coals consist mainly of pyritic and organic sulphur. The organic sulphur content is approximately the same or higher than that of pyritic sulphur. The sulphate sulphur content of the Hungarian and Central European brown coals is minimal.

From an environmental point of view, the unfavourable property of Hungarian brown coals is dangerous because during combustion of these coals the greatest part of the sulphur content enters the atmosphere as combustible sulphur. The dangerous impact of atmospheric acidification on living organisms is supported among other examples by the dying sessile oaks in Hungary (Jakucs 1986).

Table 2
Range and means of forms of sulphur in Middle European brown coal samples

Country	Mines	Number of samples	Age	Total sulphur	Pyritic sulphur (wt. %)	Sulphate sulphur	Organic sulphur dif.
Czech and Slovak Republic	Juiri Antonin						
	Pres. K. Gottwald						
	M. Gorkij	8	Miocene	0.3–1.8	0.1–1.3	0.1–0.4	0.2–0.6
	Merkur			1.0	0.5	0.2	0.4
	Charbarovice						
	Handlova						
Germany	Spreetal						
	Klettwitz						
	Schlabendorf-S	5	Miocene	0.3–1.4	0.1–0.7	<0.1–0.2	0.2–1.0
	Berzdorf			0.6	0.4	0.1	0.4
	Puschwitz						
	Espenhain	3	Oligocene	1.7–2.1	0.1–0.5	<0.1–0.3	1.0–1.8
	Delitzch			1.9	0.4	0.2	1.3
	Goitsche						
	Nachsterstedt	1	Eocene	0.9	0.1	0.1	0.7
Poland	Seniawa						
	Belchatow	4	Miocene	0.5–1.5	0.1–0.6	0.1–0.2	0.1–0.8
	Konin-Platnów			0.9	0.3	0.1	0.4
	Turów						
	Konin-Lubstów	2	Oligocene	0.1	0.0–0.1	0.0–0.1	0.0–0.1

Hungary joined the 1979 Convention on Long-Range Transboundary Air Pollution on the Reduction of Sulphur Emissions or their Transboundary Fluxes by at least 30%. In our country the sulphur emission will be reduced to that same level by 1993. At the same time, it should be remarked that, according to some authors' calculations, a decrease of 30% in SO_2 emission would not result in 30% improvement of air and of the precipitation quality in Europe, since wet and dry deposition of acid materials would be lowered only by 8 and 19 percent, respectively (Horváth and Möller 1987).

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Source rock potential of the black shale formations of the Ukrainian Carpathians

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The Ukrainian Carpathians form the middle sector of the Carpathian Arc. The flysch sequence contains two main organic-rich facies: Shypot and Spas (Lower Cretaceous) in the base and Menilite (Oligocene–Lower Miocene) in the top of it. The thermal maturity of both of them is controlled by the tectonic history of the thrusted slices in which they occur.

Within the outer tectonic units of the Flysch Belt the Menilite Beds are the main source rock sequence, though the Spas Beds can also contribute to the formation of hydrocarbon deposits.

Within the inner tectonic units, in spite of the variability of the maturation level, both the Oligocene and Lower Cretaceous organic-rich facies in some of the thrust units can be considered as hydrocarbon source rocks.

Key words: organic-rich rocks, thermal maturity, palaeotemperature, hydrocarbon generation, oil window

Introduction

The Ukrainian Carpathians form the middle sector of the Carpathian Arc between the Polish and Romanian Carpathians (Fig. 1). The geological structure and the stratigraphy of the Ukrainian Carpathians have been described in detail (e.g. Vialov 1961; Glushko 1968; Burov et al. 1974; Danys et al. 1974; Gabinet et al. 1976; Burchfiel and Royden 1982). The flysch sequence, which forms the Outer Carpathian Belt and the Boryslav–Pokuttia Unit of the Carpathian Foredeep, corresponds to the stratigraphic interval between Early Cretaceous and Early Miocene and can reach a thickness exceeding 8 km. All of the discovered oil fields occur within the Boryslav–Pokuttia and outer part of Skiba Units. The Oligocene rocks of the Menilite Formation in this area have high oil generative potential (Koltun 1992) and are probably the main source rock sequence. However, the estimation of source rock potential of the Lower Cretaceous organic-rich facies is also important.

The possibility of oil and gas deposits' occurrence in the inner part of the Ukrainian Carpathians Flysch Belt is still not determined. That is why the investigation of the organic-rich facies of this area may contribute to the solution of this problem.

The aim of this study is to estimate the source rock potential of the Lower Cretaceous and Oligocene sequences of the Ukrainian Carpathians and to show the peculiarities of its lateral variations and evolution with the increase in depth.

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Geological setting

The Ukrainian Carpathian Flysch contains two main organic-rich sequences – the Lower Cretaceous and the Oligocene–Lower Miocene.

Lower Cretaceous black shales occur in the lowest part of the flysch sequence and are represented by Shypot and Spas Formations. The Shypot Formation is found in inner tectonic units of the Carpathian Flysch Belt. The Lower Shypot Beds (up to 400 m thick) consist of black shales with minor siltstone and sandstone layers. In the Upper Shypot Beds (nearly 200 m thick) the sandstone layers predominate over black shales. The organic carbon content in the Shypot Formation rocks is up to 8%.

The Spas Formation occurs in the outer part of the Skiba Unit and is considered to be equivalent to the Shypot Formation. It is a more than 200 m thick, predominantly black shale sequence with thick sandstone layers in its middle part. Organic carbon content is up to 5%.

Both the Shypot and Spas Formations are considered to be of Late Barremian–Albian age (Leshtshukh 1988a, b).

The black shale Menilite Formation, of Oligocene–Early Miocene age (Andreeva-Grigorovitsh 1986), is the youngest part of the Carpathian Flysch sequence. It is unconformably covered by the molasse. The Boryslav–Pokuttia and Skiba Units are the main area where the Menilite Formation is found. Its maximum thickness is more than 1500 m. In the Boryslav–Pokuttia and the external part of the Skiba Units, it is represented by Lower, Middle and Upper Menilite Beds. The Lower Menilite Beds are 300 m thick and consist of rhythmical interbeddings of sandstones, siltstones and siliceous black shales. The organic carbon content of the black shales reaches values of more than 18%. The Middle Menilite Beds are up to 200 m thick and differ from those mentioned above by a lower organic carbon content (0.5–2%) in argillites. The Upper Menilite Beds are an up to 1000 m thick sequence of black shales with minor siltstone and sandstone layers. The Middle and Upper Menilite Beds occur in a localized area. That is why the widespread Lower Menilite Beds can be considered as the main object of investigation, especially in respect of hydrocarbon generation. Southwestward, across the Carpathian direction, the thickness of the Lower Menilite Beds decreases. They are gradually replaced from their top by less organic-rich Krosno Formation rocks, and their thickness reaches some tens of metres along the south-western edge of the Skiba Unit.

The important feature of the Carpathian Flysch Belt is its overthrust structure. It is most pronounced in the Boryslav–Pokuttia and outer part of Skiba Units. The slabs here form series of superposed tectonic units, and, as a result, the same formations, in particular Oligocene and Lower Cretaceous organic-rich facies, occur in a wide range of depth. This allows the study of the depth-related evolution of their petroleum potential.

Results and discussion

The Boryslav–Pokuttia and the outer part of Skiba Units are the main areas where oil deposits have been discovered. This is also the area where the Menilite Formation reaches its maximum thickness. Pyrolysis results on Menilite Formation rocks from this area were described in detail earlier (Koltun 1992). According to these data, the Menilite Formation appears to be the main source rock sequence in the studied region. Samples from four wells were investigated: Komarnytska-1, Pidlisivska-1, Pivdenny Monastyrets-5 and Porogy-1 (Fig. 1). The range of the total organic carbon (TOC) content (from 0.91 to 17.83%) covers the range characteristic for whole Menilite Formation values, and therefore the studied samples are rather representative. The values of the S_2 parameter show that the petroleum generative potential of the majority of samples can be

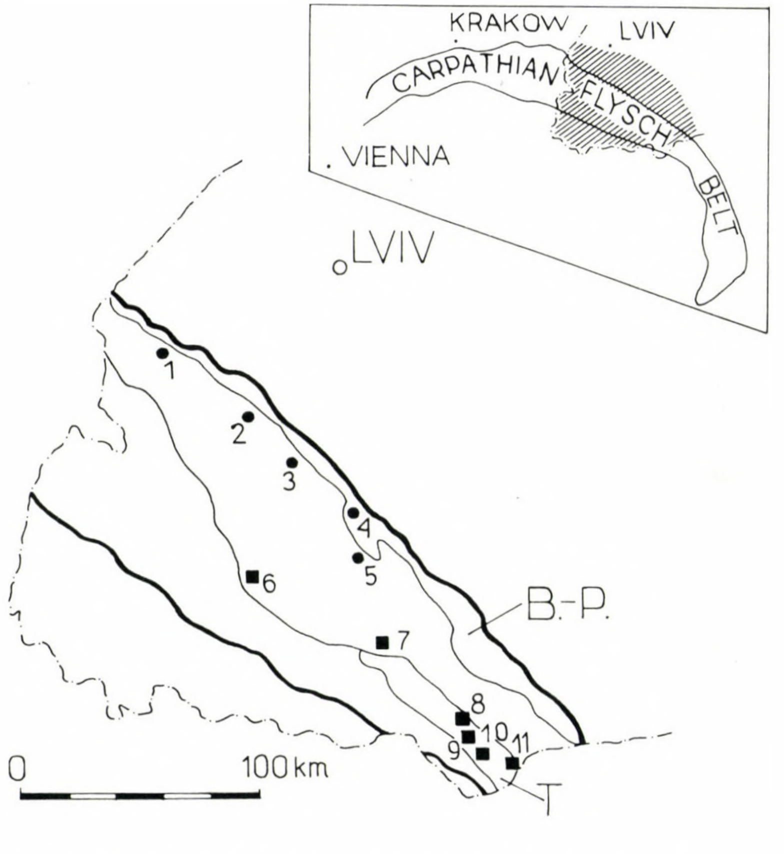

Fig. 1
Location map showing major tectonic units and position of sites under study. Tectonic units: B.–P. – Boryslav–Pokuttia Unit; S – Skiba Unit; T – Tshornogora Unit. Wells (circles): 1. Pivdenny Monastyrets-5; 2. Komarnytska-1; 3. Maxymivka-4; 4. Pidlisivska-1; 5. Porogy-1. Outcrops (squares): 6. Synievir; 7. Yasynia; 8. Gryniava; 9. Goloshyna; 10. Yalovychora; 11. Suchava

considered as good and very good (the interpretation of the pyrolysis parameters is based on the conclusions given by Espitalie et al. 1985a, b, 1986). The HI versus T_{max} diagram (Fig. 2) shows that the studied samples contain mainly Type II kerogen and their thermal maturity corresponds to the end of diagenesis and the beginning of the oil formation zone. The outer part of the Carpathian Flysch Belt is the area where the overthrust structure is the most developed. Thrust slices form several superposed tectonic units. Thus the Menilite Beds, occurring in different slabs, are situated in a wide range of depths. This allows the study of the evolution of the hydrocarbon generating

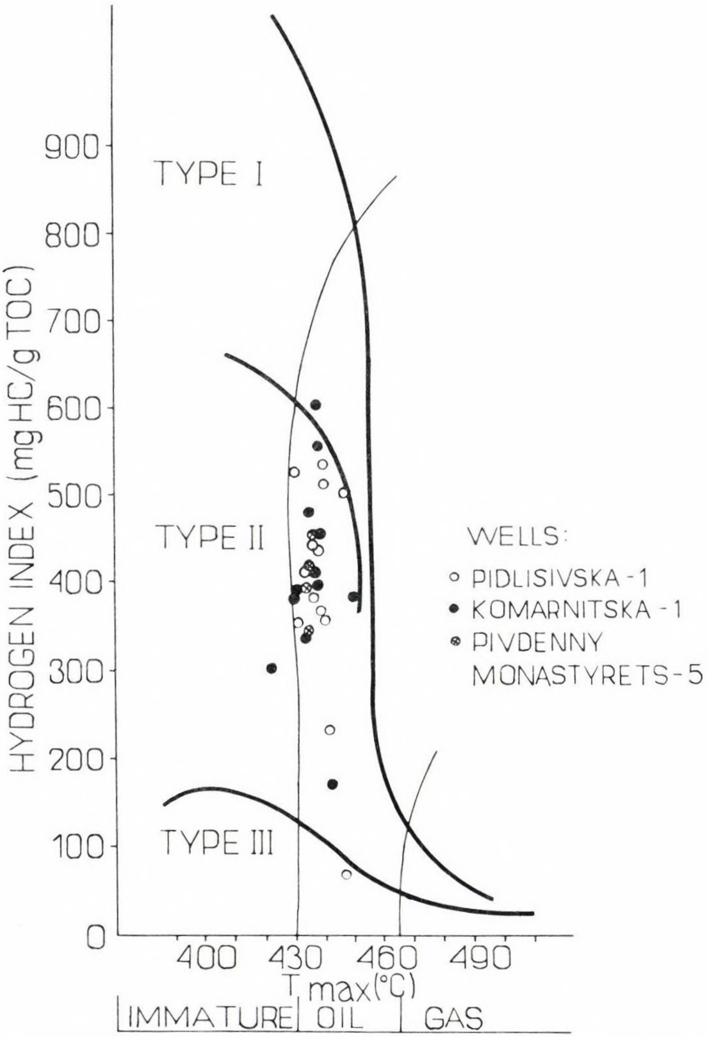

Fig. 2
Plot of Hydrogen Index versus T_{max} for Menilite Formation rocks (Oligocene) from the Boryslav–Pokuttia and outer part of Skiba Units of the Ukrainian Carpathians

potential of these rocks with the increase in depth. The studied wells crossed the Menilite Beds in three thrusted units within depth intervals from 0.8 to 5 km.

Within the upper 4 km of the sequence, the Total Production Index (TPI) ranges from 0.02 to 0.09 (average 0.06) (21 samples), S'_1/TOC – from 10.73 to 34.68 (average 26.47) mg HC/g TOC, showing no tendency of increasing with the increase in depth. At depth range between 4 and 5 km (16 samples) the values of these parameters noticeably increase. TPI ranges from 0.06 to 0.19 (average 0.12), S'_1/TOC – from 15.79 to 78.33 (average 46.78) mg HC/g TOC. We believe that the increase of these parameters at depths over 4 km shows the intensification of hydrocarbon generation related with the increase of thermal maturation level.

Similar conclusions can be made from the vitrinite reflectance data (Koltun 1992). At depths from the surface to almost 4 km, R_o values range from 0.55 to 0.65%, corresponding to palaeotemperatures of 100–128 °C (according to the Ammosov (1987) scale). No noticeable change of R_o values related with the increase in depth is observed. It can be assumed that this is the level of palaeotemperature reached by the Menilite Beds at their pre-overthrusting stage. Only at depths over 4 km, where the present day burial temperatures (average geothermal gradient is nearly 24 °C/km) exceeds the above-mentioned level of palaeotemperatures, R_o gradually increases from 0.58–0.70% at nearly 5 km to 0.84–0.87% at almost 6 km depth, approximately corresponding to the rise of the temperatures they are subjected to at present. That is why we believe that the Menilite Formation rocks within the upper 4 km of the sequence preserved their pre-emplacement level of maturity, while the thermal evolution of the deeper part of the sequence proceeds under the influence of the present day heat flow.

The T_{max} values do not contradict the above-described maturation history. Within the whole studied depth interval, its values range from 421 to 450 °C. The maturation level of the majority of samples corresponds to the end of diagenesis and the beginning of the oil formation zone (T_{max} from 421 to 443 °C). Only at depths more than 4 km do some of the samples show higher T_{max} values.

Petroleum Potential (S_2) values range from 0.77 to 65.34 mg HC/g rock, Hydrogen Index (HI) – from 67 to 595 mg HC/g TOC. Both of these parameters show no clear tendency of decreasing with the increase in depth, and even at a depth of nearly 5 km they retain high values, showing that only a small part of the oil potential of the Menilite Formation rocks has been realized at this depth.

Extrapolating the maturation level of the Menilite Beds according to the present day burial temperatures, it can be assumed that the lower boundary of the oil window can reach the depth of nearly 8 km.

Four studied wells show similar conclusions for both source rock potential and its evolution with the increase in depth. Since they cover a rather great distance along the outer part of the Ukrainian Carpathians Flysch Belt (Fig. 1), it can be assumed that the Menilite Formation rocks show rather similar qualities within this zone.

As the facial change of the Oligocene sequence occurs in the cross-Carpathian direction towards the decrease of the organic-rich rocks thickness, it seems interesting to estimate the source rock potential of the Menilite Formation rocks from the more internal part of the Carpathian Flysch Belt in order to determine the possibility of hydrocarbon generation within this area.

Menilite Formation rocks from the most inner (southwestern) part of the Skiba Unit were studied in two different outcrops: near Yasynia and Synievir villages (Fig. 1). Total organic carbon content in the rocks of this area remains rather high. Four samples from the Yasynia outcrop showed the range of TOC values from 1.71 to 6.39% (average 4.10%), 9 samples from the Synievir outcrop – from 0.47 to 3.97% (average 1.81%). Source rock potential and thermal maturity of the rocks from these two sites appeared to be quite different, though they are situated in an analogous geological and facial setting. Samples from the Yasynia outcrop contain Type II, close to III kerogen (Fig. 3). T_{max} ranges from 434 to 442 °C (average 437 °C), showing that thermal maturity of the rocks corresponds to the beginning of the oil window. The obtained data on the maturation rank is corroborated by the vitrinite reflectance of two samples from the same sequence. The R_o values are 0.58 and 0.60%. TPI values range from 0.07 to 0.29 (average 0.18). HI keeps rather high values – from 136 to 204 (average 184) mg HC/g TOC. Peters (1986) suggested that at thermal maturity corresponding to the beginning of the oil window, rocks with HI between 300 and 150 mg HC/g TOC will produce oil and gas. Thus the Menilite Formation rocks from the Yasynia region can be considered as a potential hydrocarbon source.

Quite a different conclusion can be reached while analysing the pyrolysis data on Menilite Formation rocks from the Synievir outcrop. Both T_{max} and TPI parameters show that they are "overmature". T_{max} ranges from 457 to 500 °C (average 472 °C), TPI – from 0.13 to 0.57 (average 0.38). Low HI values (from 21 to 117 (average 47) mg HC/g TOC) show that the source rock potential is almost completely exhausted. According to Peters (1989, personal communication, cited by Langford and Blanc-Valleron 1990) the rocks with HI between 150 and 50 mg HC/g TOC will produce gas, and those with HI less than 50 are inert. It means that Menilite Formation rocks from Synievir region mostly cannot be considered as a source, either for oil or for gas. The HI versus T_{max} diagram (Fig. 3) shows that the studied samples are concentrated close to the boundary between Type II and Type III kerogen. So there is no significant difference in kerogen type between Menilite Beds from Yasynia and Synievir. Probably these rocks were initially similar, but their subsequent burial history was quite different. We can suppose that Menilite Beds from the Synievir region in the past were deeply overburdened under the frontal part of the adjacent coming from the south-west tectonic unit, but later the overlying thrust slices were eroded. Taking into account the high maturation level of the rocks, the higher palaeoheat flow in this area can also be assumed.

Menilite Beds from the Yasynia area probably were never deeply buried and hence they show a rather low level of maturity.

Lower Cretaceous black shales, being widespread throughout the Ukrainian Carpathian Flysch Belt, are the second important object for the investigation of their source rock potential. Spas Formation rocks are spread in outer tectonic units and were found within a wide range of depths from outcrops to 7 km

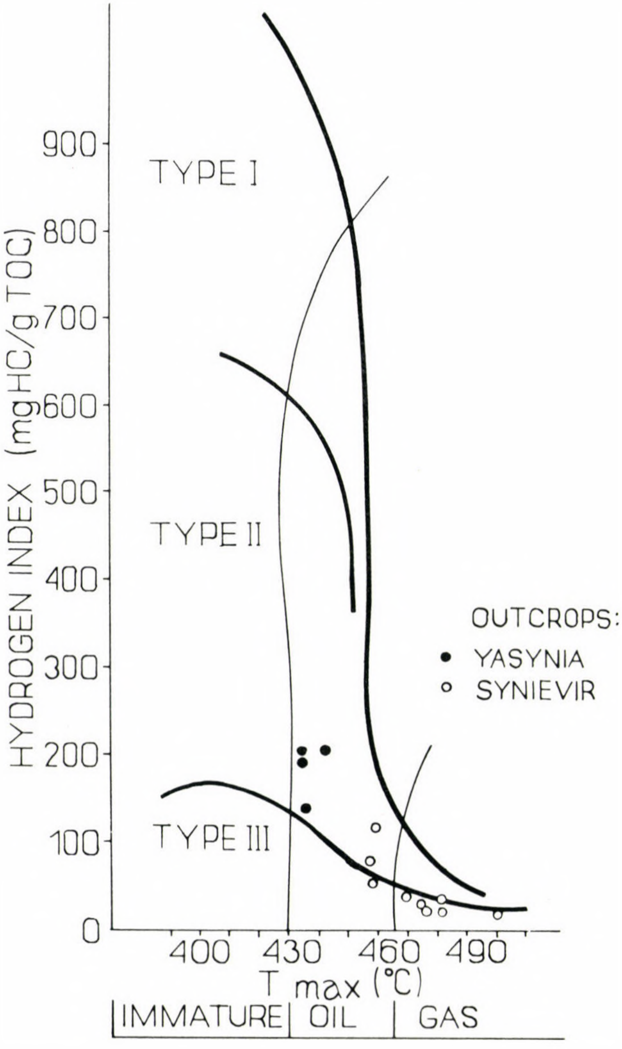

Fig. 3
Plot of Hydrogen Index versus T_{max} for Menilite Formation rocks (Oligocene) from the inner part of Skiba Unit of the Ukrainian Carpathians

depth. Five black shale samples from Maxymivka-4 well (depth interval 4247 to 4433 m) were studied using Rock-Eval pyrolysis. According to these data the rocks contain Type II close to III kerogen (Fig. 4) and can be considered as source rocks with fair to good hydrocarbon generative potential. S_2 values range from 2.17 to 8.82 (average 4.60) mg HC/g rock. HI values range from 107 to 179 (average 147) mg HC/g TOC, showing that these rocks can produce oil and gas. T_{max} ranges from 443 to 449 °C,(average 447 °C), showing that thermal maturity of the rocks corresponds to the oil window. Thus the Spas Formation sequence should also be accounted as a possible source of hydrocarbons in the studied region.

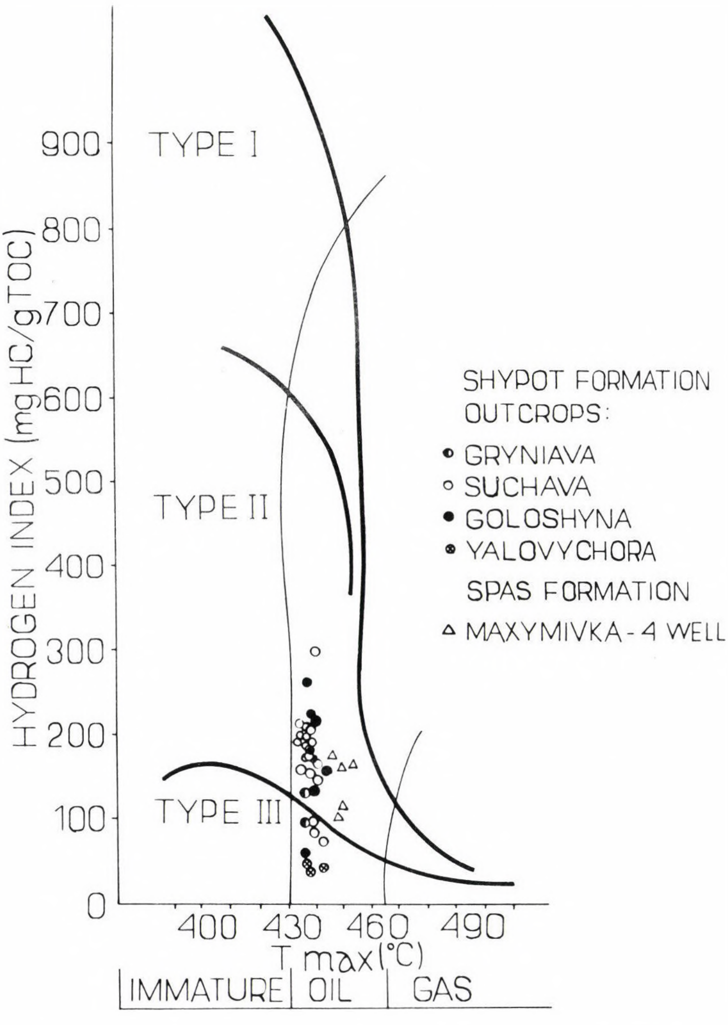

Fig. 4
Plot of Hydrogen Index versus T_{max} for Lower Cretaceous rocks of the Ukrainian Carpathians. Spas Formation rocks (triangles) from the outer part of Skiba Unit. Shypot Formation rocks (circles) from Tshornogora Unit

According to the relatively low T_{max} values of these samples, they can reflect their pre-overthrusting level of maturity, as it was in case of the Menilite Beds from the upper 4 km of the sequence, described above. In the whole area of Spas Formation occurrence the complete flysch sequence up to Oligocene and in some places Lower Miocene occurs. This means that the Lower Cretaceous Spas Beds, being the lowest element of the Carpathian Flysch sequence, were buried under at least 3–4 km of sedimentary cover, until they were drawn into the overthrusting movements. Assuming that the palaeoheat flow was not less, but may be higher, than the present-day one, it can be concluded that the studied Spas Beds from depth 4.2–4.4 km are now exposed to present day burial temperature not exceeding the palaeo one, and their maturation level probably reflects the maximum palaeotemperature before the overthrusting. This is corroborated by the vitrinite reflectance of one sample of the Golovninska Formation, lying just over the Spas Beds, from the outcrop near the town of Tershiv, shown by Gabinet (1985). The R_o value of 0.67% shows a palaeotemperature of near 130 °C (transformation from R_a values given by Gabinet (1985) to R_o and the corresponding palaeotemperatures were taken from the Ammosov (1987) scale); it corresponds to the oil window and is roughly comparable to the above-presented maturation level of the Spas Beds from 4.2–4.4 km depth. If we assume that this is the level of the pre-overthrusting maturity of the Spas Beds of this area, the hydrocarbon generation processes are probably "frozen" in them up to the depth of about 5.5 km, where the present-day temperature reaches nearly 130 °C. At greater depths, maturation level of these rocks probably proceeds under the influence of the present-day heat flow. Gabinet (1985) gives a value of R_o = 1.14% for Spas Beds from 7.2 km depth. This maturation rank corresponds approximately to the present day burial temperature at this depth. It also shows that this depth is already close to the bottom of the oil window for Spas Beds.

The black shale sequence of Lower Cretaceous Shypot Formation, spread in the inner tectonic units of the Flysch Belt, was studied in four outcrops within the Tshornogora Unit (Fig. 1). Lower Shypot Beds were investigated from the Gryniava, Suchava and Goloshyna outcrops, Upper Shypot Beds – from the Yalovychora outcrop. Rock-Eval pyrolysis results show similar conclusions for the Lower Shypot Beds from all of the studied outcrops. They contain Type II and Type III kerogen (Fig. 4) and can be considered as a source rock sequence with fair to good hydrocarbon generative potential. TOC content in the studied samples is from 0.46 to 7.48% (average 3.85%). S_2 ranges from 0.27 to 15.64 (average 6.92) mg HC/g rock. HI values are from 58 to 300 (average 173) mg HC/g TOC, showing that these rocks can produce oil and gas. The thermal maturity of the studied samples is close to the boundary of diagenesis and catagenesis, corresponding in the most cases to the beginning of the oil window. T_{max} values range from 433 to 441 (average 437) °C, TPI values – from 0.04 to 0.21 (average 0.08).

Upper Shypot black shales contain Type III kerogen (Fig. 4) having the same thermal maturity level (T_{max} – from 437 to 441 °C, TPI – from 0.09 to 0.10). They differ from those mentioned above by lower TOC content (from 0.90 to 0.98%) and low HI values (from 40 to 56 mg HC/g TOC) showing that they are inert in respect to hydrocarbon generation.

The low thermal maturity level of the Shypot Beds in the Tshornogora Unit shows that they were never buried under the thick cover of the younger sediments.

The study of the black shale sequences of the inner tectonic units of the Carpathian Flysch Belt shows the great variability of the thermal maturity level of the rocks from different thrust slices, regardless of the similarities in their geologic setting and the stratigraphical position in the sequence. Analogous Oligocene sequences of Yasynia and Synievir showed a very different maturation level, while the Lower Cretaceous rocks of Tshornogora Unit are much less mature than the Synievir and approximately equal to Yasynia Oligocene Beds. The observed variability of the maturation level of the sequences within inner tectonic units of the Ukrainian Carpathians can probably be explained by the different tectonic history of the thrust slices in this area and very complex spatial relationships between them during the overthrusting deformations of the Carpathian Flysch. Variations of the palaeoheat flow in different areas should also be assumed.

Conclusions

Two source rock sequences can be distinguished in the Ukrainian Carpathian Flysch: the Oligocene Menilite Formation and the Lower Cretaceous Spas and Shypot Formations. The thermal maturity of both is controlled by the tectonic history of the thrust slices in which they occur.

The Oligocene Menilite Formation can be considered as the main source rock sequence within the Boryslav–Pokuttia and outer part of the Skiba Units, where the majority of oil fields occurs. The intense generation of oil by these rocks probably begins at a depth of nearly 4 km.

The Lower Cretaceous Spas Formation also can contribute to the hydrocarbon deposits formation in this area. The intensification of hydrocarbon generation processes in these rocks may occur at a depth over 5.5 km.

The lower boundary of the oil window for both the Menilite and Spas Beds within the outer tectonic units can be predicted at depth of nearly 7–8 km.

Within the inner tectonic units of the Carpathian Flysch Belt a variability of maturation level of the organic-rich rocks in different thrust units is observed. However, both Oligocene Menilite Beds and Lower Cretaceous Shypot Beds should be counted as a possible source of hydrocarbons in this area. Thus, in places where the favourable conditions for their accumulation exist, oil and gas deposits can be expected.

Acknowledgements

The author is grateful to Dr. F. Roure (Institut Français du Pétrole), who kindly gave him the pyrolysis results obtained during the common field work. I would like to thank Drs. O. V. Bartashevitsh, V. N. Litvinova and L. D. Burmistrova (NIIGeoInformSystem, Moscow) for pyrolysis of my samples from the Carpathian Foredeep, and Dr. I. Vető (Hungarian Geological Survey) for his review of the manuscript and helpful comments.

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Upper Paleozoic–Mesozoic formations of the Mid-Transdanubian Unit and their relationships

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The Central Transdanubian (Igal) unit has been a key issue in the pre-Neogene basement of the Pannonian Basin, the elucidation of its setting is indispensable from the geodynamic aspect. In the narrow strongly tectonized unit lying between the Central Hungarian and Balaton Lines Late Paleozoic and Triassic formations do not occur on the surface but are known only from structural key boreholes, hydrocarbon and water exploratory wells.

In the northern strip of the tectonic unit Lower Permian clasitc Trogkofel strata and Upper Permian dolomites similar to the Bellerophon-bearing dolomite of the Carnian Alps are known in spots.

Triassic sequences of different formations of the Northwestern, eastern and southwestern parts relate to different tectonic-paleogeographic relations.

In the theoretical sequence of the northwestern area nearly the complete period is represents. The Lower Triassic shallow marine and the Anisian platform carbonates are common. The Ladinian radiolarian tuffitic clastid sequences are known in tectonically strongly disturbed areas. The Upper Triassic platform-marginal back-reef lagoon formations are similar to the corresponding formations of the Southern Caravanca.

The Triassic sequence of the eastern part known only from sporadic data differs both in structure and in fossil assemblage from the corresponding formations of the Transdanubian Central Range, of the Bükk Mountains. Only a few data are available on the Lower Triassic shallow marine carbonate formations. The Anisian intertidal lagoon facies formations are rich in fossils. The Ladinian–Carnian Wetterstein-type platform formations are represented by near-reef and fore-reef slope sediments. These Wetterstein-type reef formations can be found both in the Northern Calcareous Alps, in the Northern Caravanca, in the Dinarides and in the Internal Western Carpathians. The turrispirillina-bearing formations of Igal indicate the enviroment of formation close to the outer platform margin and support the relationship with the Dinarides.

The complex of the southwestern part containing carbonatic, pelitic sediments, acid to intermediate metavolcanics that underwent partly anchizonal metamorphism, as well as serpentinites, displays similarities also to the Dinarides and belongs probably to a lower-situated nappe unit. Its belonging, however, to the Repno complex cannot be ecxluded where the ophiolitic melange is found beneath the Tara-nappe.

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At the level of recent kowledge it is obvious that the pre-Tertiary basement of the Central Transdanubian Unit is foreign to its recent geological setting and is of non-uniform built-up. Part-units got probably close to each other along parallel displacement planes but it is also probable that nappe formation during the Earth's history complicates the tectonic conditions.

Key words: Permian, Triassic, microfacies, Foraminifera, regional relationships

Introduction

The Paleozoic-Mesozoic formations of the narrow, strongly tectonized Central Transdanubian (Igal) structural unit lying between the Central Hungarian and Balaton lines, are hardly known. In Hungary these formations do not outcrop to the surface. They have been explored by a few key boreholes, and a relatively great number of hydrocarbon and hydrological exploratory wells. Location of sites is shown in Fig. 1.

The understanding and evaluation of the relatively few, and not easily interpretable data, as well as their comparison with the formations of similar age of the surrounding regions (primarily with the Sava folds lying in the southwestern continuation of the zone, and with the Dinarides), seem to be one of the key problems of the tectono-evolutionary-geodynamic interpretation of the region.

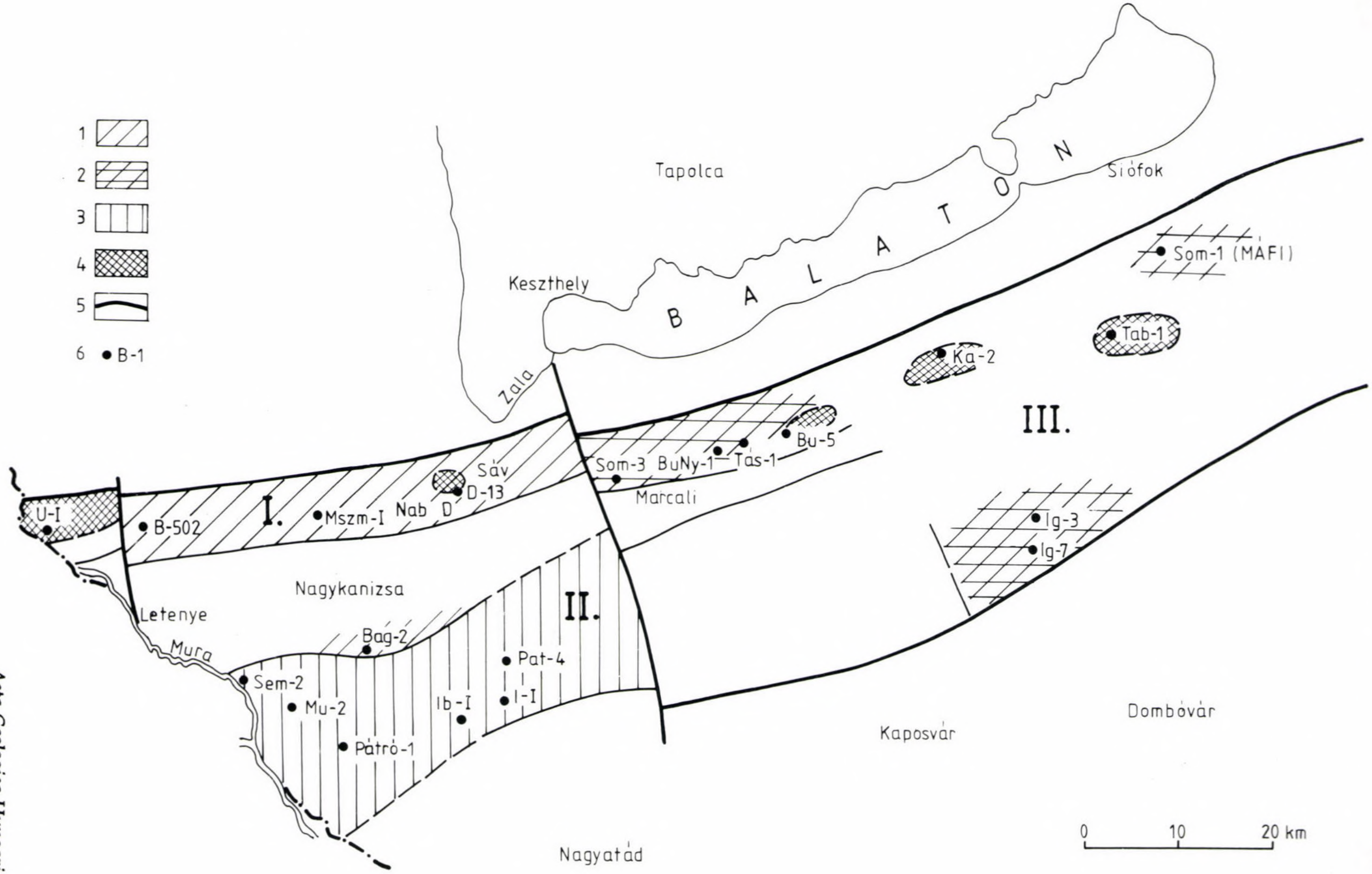
The structural-geological and paleogeographical problems of the region lying in the southern foreground of Lake Balaton (between the Transdanubian Mid-Mountains and the inselbergs of South-Baranya) were brought up in the fifties, partly as a result of the wells drilled in the region of Karád, Buzsák and Igal, and partly due to the paleontological-faciological knowledge obtained from the Bükk Mountains.

In the hydrocarbon exploratory wells of Karád drilled in 1953, fusulina-bearing limestone was found (Karád-1, -2), which was assigned to the Carboniferous (Majzon 1966). Limestones encountered in the Buzsák well were similarly interpreted, and Vadász (1960) described the Upper Carboniferous formations of the Igal–Buzsák–Karád Paleozoic belt.

Almost simultaneously, in the course of explorations in the Bükk Mountains, based primarily on the features and fossils of marine Upper Permian formations, Schréter (1959) came to the conclusion that an "Upper Permian sea branch protruded into the area of the present-day Bükk Mountains from the region of the Julian Alps". According to Balogh (1964) "The Late Paleozoic of the Bükk Mountains is closely related to the Dinarides..."; thus, it is reasonable to assume the direct connection of this region with the syncline of the Bükk Mountains.

On the other hand, Vadász (1960) was of the opinion that the Upper Carboniferous of Igal–Buzsák–Karád is a bay protruding towards the Carnian

Fig. 1→
Structural setting of the Mid-Transdanubian Unit and borehole reached the Upper Paleozoic–Mesozoic basement


1
2
3
4
5
6 B-1
Tapolca
B A L A T O N
Siófok
Keszthely
Zala
Som-1 (MÁFI)
Tab-1
Ka-2
III.
Bu-5
Som-3 BuNy-1 Tas-1
Sáv
D-13
Nab D
Marcali
U-I
B-502
I.
Mszm-I
Ig-3
Ig-7
Nagykanizsa
II.
Letenye
Bag-2
Mura
Pat-4
Sem-2
Mu-2
Ib-I
I-I
Pátró-1
Dombóvár
Kaposvár
0 10 20 km
Nagyatád

Alps "that can be hardly considered to be a sea connected with the Upper Carboniferous of the Bükk Mountains, since the intermediate crystalline basement of the Great Plain was above the sea level at that time".

In spite of the above-mentioned viewpoint, in the seventies the possibility that the basement formations explored in the belt south of Lake Balaton may indicate the lacking links between the Bükk Mountains and the Julian Alps–Dinarides came to be accepted. This idea was most markedly supported by Wein (1969, 1972, 1973), who called the region between the Balaton–Darnó and Zagreb–Kulcs megatectonic lines the Igal-Bükk eugeosyncline. In his later work, influenced by plate tectonic views (1978), he treated this region as a unit which developed in the south Tethyan belt, together with other units northwest of the Zagreb–Kulcs–Hernád line, and which was thrusted on to an oceanic crustal block.

According to the mobilistic view predominating in the eighties, no connection needed to be presumed in the relationship of the Bükk Mountains with the Dinaric formations by means of marine trenches. But concepts have predominated according to which the formations in the Bükk Mountains and in the Igal belt could be situated at the northwestern termination of the Dinaric sedimentary basin during the Late Paleozoic and Mesozoic, and were squeezed out, reaching their recent position during the Alpine orogeny (Kovács 1982, 1983; Kázmér 1984; Kázmér and Kovács 1985; Haas 1987).

In the eighties new wells were drilled, partly aiming at hydrocarbon exploration, partly in the frame of the National Key Section Program, which considerably modified both the stratigraphic concept and the knowledge of the features of the formations. Based on the new information, former data was reviewed. In this paper, besides presenting the new data, an attempt is made to carry out a comparison with the surrounding structural units which may have been in paleogeographic connection with the Mid-Transdanubian unit during the Late Paleozoic–Mesozoic evolutionary phase.

In the review, first the data concerning the Permian (qualified earlier as Carboniferous) formations will be discussed. Triassic formations will be discussed in three parts, describing areal units of different lithostratigraphic patterns.

Permian formations

In the pre-Neogene basement of the northern strip of the Mid-Transdanubian belt, Permian formations were encountered in discontinuous, spotty extension (Fig. 2). Hydrocarbon exploratory wells revealed a detrital sequence of the Trogkofel Formation in three areas (Újfalu, Karád, Buzsák). Upper Permian marine deposits are known at Tab, and an uncertain Upper Permian occurrence was found at Újudvar.

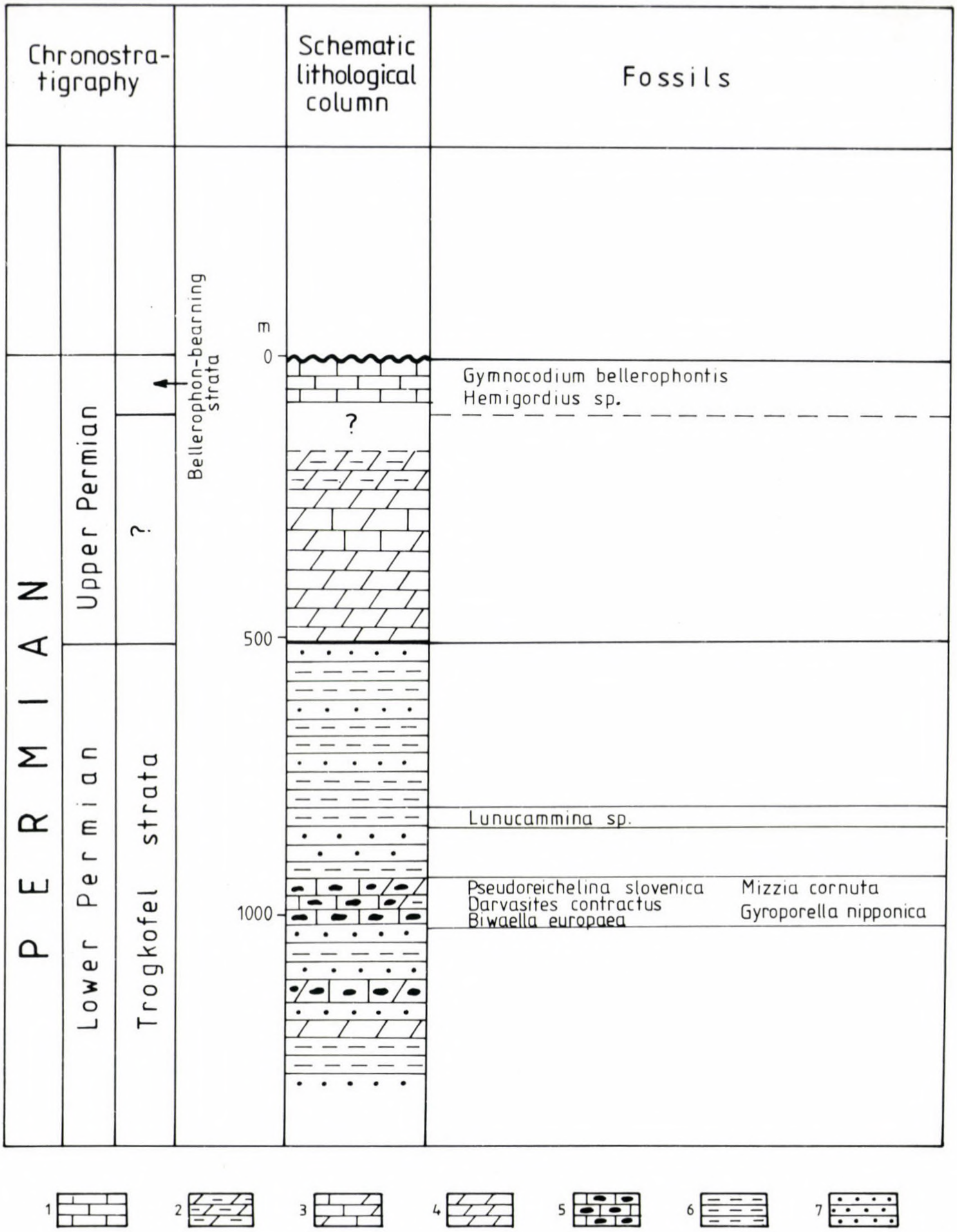


Fig. 2
Stratigraphic column of the Permian formations of the Mid-Transdanubian Unit. 1. grey limestone; 2. grey, argillaceous limestone, 3. light-grey dolomitic limestone; 4. dolomite; 5. dark-grey, grey, brecciated limestone; 6. dark-grey shale; 7. light-grey, fine-grained sandstone

Lower Permian

The most complete Lower Permian sequence was traversed close to the Slovenian frontier by the Újfalu-I borehole, with a thickness of about 600 m (Bérczi-Makk and Kochansky-Devidé 1981). It consists predominantly of fine-detrital rocks (pale-grey fine-grained quartz sandstone, dark-grey silty shale, dark-grey to black silty shale); fossiliferous, dark-grey dolomitic limestone strata and reef-breccia lenses occur only subordinately, containing Algae (*Mizzia cornuta* Kochansky-Devidé et Herak, *Gyroporella nipponica* Endo et Hashimoto) and Foraminifera [Pseudoreichelina slovenica (Kochansky-Devidé), *Schubertella australis* Thompson et Miller, *Schubertella kingi* Dunbar et Skinner, *Schubertella paramelonica* Sulejmanov, *Biwaella europaea* Kochansky-Devidé et Milanovic, *Darvasites contractus* (Schellwien), *Neotuberitina malajvkini* (Mihajlov), *Clima- cammina elegans* (Moelier), *Climacammina* cf. *rugosa* Morczova, *Tetrataxis nana* (Morozova), *Globivalvulina parva* Csernuseva, *Lasiodiscus irregularis* (Miklukho- Maklay)]. The formation at the base of the borehole Buzsák-5 is similar to the development exposed in the Ujfalu well. It can be conditionally fitted into the Újfalu series due to its detrital nature and poorish, Permian foraminifer assemblage (*Globivalvulina vulgaris* Morozova, *Lunucammina* sp., *Pachyphloia* sp.).

The fine-detrital grey sandstones with siliceous cement, marls and yellowish-grey reef-breccia limestones of the strongly tectonized (brecciated, folded, pressed) sequence revealed by the wells Karád-1 and -2 can be fairly well correlated with the Lower Permian clastic Trogkofel strata, on the basis of the petrographic features and of the fossil assemblage [*Rugososchusenella* sp., *Darvasites contractus* (Schellwien), *Lasiodiscus* sp., *Globivalvulina vulgaris* Morozova, *Climacammina* sp., *Tetrataxis* sp.; Bérczi-Makk 1988].

Upper Permian

On the basis of position of the carbonatic sequence (light-grey to grey, fine-crystalline dolomite, dolomitic limestone, clayey dolomite) immediately above the Trogkofel sequence at Újfalu (U-I), it is the oldest formation of the Upper Permian. Because of the lack of fossils it cannot be unambiguously classified. It may be the counterpart of the Bellerophon Dolomite in the Carnian Alps.

In the Újudvar area (D-13), an uncertain datum indicating the upper Bellerophon-horizon was found. Calcareous algae (*Gymnocodium bellerophontis* Rothpletz) and foraminifera (*Hemigordius* sp.) fragments in the traversed grey limestone argue against a classification in the lower Triassic. Due to the few and incomplete data, this lagoonal formation can be only conditionally qualified as Late Paleozoic (Upper Permian).

The lowermost Upper Permian formation in the Tab-1 exploratory well (Szabó 1972) is a strongly tectonized, brecciated, grey to dark-grey oolitic dolomite. In the carbonatic sequence red siltstones and silty dolomites are found as intercalations. In the fossil assemblage, calcareous algae (*Gymnocodium* *bellerophontis* Rothpletz, *Permocalculus* sp.) and, among the foraminifera the Glomospira, predominate.

Triassic formations

Formations of the northwestern area (I)

In the theoretical sequence based on data from hydrocarbon exploration wells (Bérczi-Makk 1988), nearly the whole Triassic period is represented (Fig. 3). The older Triassic formations (Lower Triassic, Anisian) are of extend over the entire area in question (Budafa, Magyarszentmiklós, Sávoly, Újudvar). The younger Triassic strata (Ladinian, Upper Triassic) are missing in the western part. In the eastern margin of the area (Nagybakónak, Sávoly, Újudvar), the Ladinian–Upper Triassic sequences were encountered in tectonically strongly disturbed regions. The polymict, microbrecciated "marker" horizons with radiolarite detritus which occur in several wells in the northwestern area of the Mid-Transdanubian belt, either in Ladinian (D-9) or in Rhaetian (D-9, Sáv-8, -13) platform limestones, can be explained by tectonic movements. The differences in thicknesses of the formations also reflect considerable tectonism.

Lower Triassic

In the Lower Triassic sequence, shallow marine carbonates, as well as siliciclastic formations in the older parts, are known (B-502, -503, Mszm-I, Sáv-7, -10, -26, D-10, -12). Typical rock-types are as follows: dark-grey, locally oolitic limestone, clayey limestone, dolomitic limestone, lime-marl and marl, often with strongly brecciated, in several horizons dark-grey, variegated shale and grey sandstone, as well as with anhydritic dolomarl intercalations. At Sávoly (Sáv-5, -7, -10, -26) and Újudvar (D-9, -13, -14), dark-grey, brecciated dolomite occurs in the topmost part of the Lower Triassic. Samples are generally poor in fossils; the foraminifera, however, unambiguously permit the chronostratigraphic classification: *Ammodiscus incertus* (D'Orbigny), *Glomospira sinensis* (Ho), *Glomospira ammodiscoidea* (Rauser), *Meandrospira pusilla* (Ho).

Middle Triassic

Anisian platform carbonates are known in all exploration areas. The light-grey to grey, locally strongly fractured, dolomite with authigenic breccia (Nab-1, -3; NabÉ-2; Sáv-6, -12; D-9), the dolomarl, dolomite and dolomitic limestone (B-502, -503, -I, -IV; Mszm-I), and limestone (D-10, -12; Sáv-2) strata are unusually rich in foraminifera: *Ammobaculites radstadtensis* (Kristan-Tollmann), *Trochammina almtalensis* Koehn-Zaninetti, *Endothyranella wirzi* (Koehn-Zaninetti), *Earlandia amplimuralis* (Pantic), *Arenovidalina chialingchiangensis* Ho, *Nodosaria* sp., *Meandrospira dinarica* Kochansky-Devidé et Pantic, *Diplotremina astrofimbriata* Kristan-Tollmann. Locally, calcareous algae fragments also occur as accessory elements.

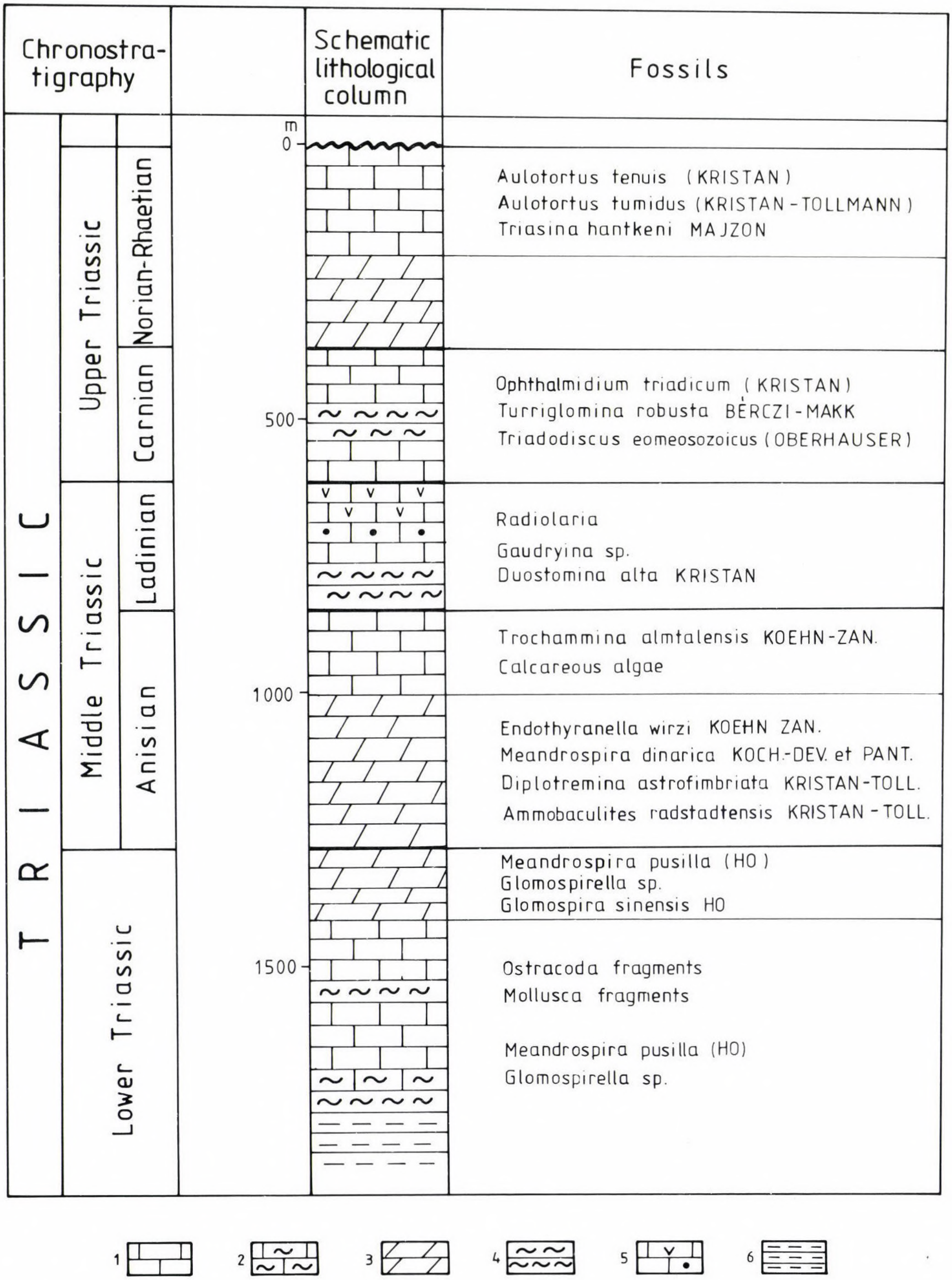


Fig. 3
Stratigraphic column of the Triassic formations in the northwestern area of the Mid-Transdanubian Unit. 1. limestone; 2. lime-marl; 3. dolomite; 4. marl; 5. tuffitic limestone, with radiolarite intercalation; 6. shale

Ladinian formations were found in the eastern, tectonically strongly disturbed part of the area, in the Sávoly (Sáv-9, -12) and Újudvar (D-7, -9) wells. The dark-grey siliceous limestone, marl, clayey limestone series of the Sávoly wells (Sáv-9, -12), with the intercalated radiolarian tuffite breccia and strongly altered porphyrite lapilli, probably represents younger Ladinian horizons. According to investigations of Árkai (1988), the sequence beneath the ankeritic Carnian limestone breccia is metavolcanite, a "shalstein"-like formation in the Sáv-9 well. The problematic sequence of Újudvar consists of light-grey and grey limestone, locally with tuff (D-7). Elsewhere, chert and radiolarite detritus (D-9) is conditionally assigned to the Ladinian. The poorish foraminifer fauna found in the shallow marine limestone does not contradict this standpoint (*Trochammina almtalensis* Koehn-Zaninetti, *Ammobaculites* sp., *Gandryina* sp., *Duostomina* sp.).

Upper Triassic

The oldest Upper Triassic formation, consisting of dark-grey, tectonically brecciated marl, sandy limestone and limestone, was encountered in the Sávoly (Sáv-1, -6, -9) and Újudvar (D-7) wells. The foraminifer fauna, associated with a platform margin, favors the classification into the Carnian: *Ophthalmidium triadicum* (Kristan), *Gaudryina triassica* Trifonova, *Triadodiscus eomesozoicus* (Oberhauser), *Turriglomina robusta* Bérczi-Makk.

The stratigraphic position of the Triassic sequence exposed in the wells Nab-2 and NabÉ-1 is problematic. The upper part of the dark-grey, microcrystalline limestone of locally porcellanitic texture, penetrated in a thickness of about 200 m, is siliceous, cherty. The common calcareous algae fragments suggest a lagoonal facies. The poorish foraminifer association (*Trochammina alpina* Kristan-Tollmann, *Trochammina* sp., *Glomospirella* cf. *expansa* Kristan-Tollmann, *Duostominidae* sp.), allow the unambiguous classification into the Upper Triassic. The uncertain *Glomospirella* cf. *expansa* and the *Duostominidae* sp. sections indicate the younger part of the Upper Triassic (Oravecz-Scheffer 1987).

The light-grey, grey, locally brecciated dolomite, devoid of fossils in the Sáv-1 well, is only conditionally considered to be Upper Triassic.

The youngest formations of the Triassic are the grayish-brown, brownish-grey, brown-shaded light-grey, locally oolitic, authigenic brecciated limestones of the wells Sáv-4, -8, -13 and D-6, -7, -9,- 11 (Újudvar). This stratigraphic setting is proven by foraminifer fauna in the lagoon behind the reef series: *Trochammina alpina* Kristan-Tollmann, *Agathammina austroalpina* Kristan-Tollmann, *Aulotortus friedli* (Kristan-Tollmann), *Aulotortus sinuosus* Weynschenk, *Aulotortus tenuis* (Kristan), *Aulotortus tumidus* Kristan-Tollmann, *Lamelliconus* sp., *Triasina hantkeni* Majzon. In the Sáv-4 core Carnian *Poikiloporella* sp. and *Giroporella* sp. were also determined by O. Piros (pers. com.).

In the Sávoly area, the diagenetized, siliceous, polymict microbreccia, with chert detritus and rounded radiolarite pebbles, encountered among (Sáv-8), and below (Sáv-13) limestone strata of lagoonal facies (uppermost-Triassic) showing synsedimentary brecciated features, is stratigraphically problematic.

Formations of the eastern area (III)

The hydrocarbon exploration wells and key-boreholes drilled in the area provide only sporadic information on the structure of the area, but permit the construction of a nearly complete theoretical strata column (Fig. 4).

Lower Triassic formations were traversed by the hydrocarbon exploration wells of Buzsák and Táska. The lowermost 60 m of the Som-1 key-borehole probably represents the top of Scythian and/or the bottom of the Anisian. The Middle Triassic (Anisian-Ladinian) formations are common in the boreholes drilled in the region, while proven Upper Triassic (Carnian-Norian) strata were traversed only by the wells Som-1 and Igal-7. No data are available so far on Uppermost Triassic (Rhaetian) formations.

Lower Triassic

The Buzsák wells (Bu-2, -4) traversed shallow marine, carbonatic Lower Triassic strata consisting of grey limestone and brecciated marl. The foraminifer assemblage *Glomospirella shengi* Ho, *Gl. facilis* Ho, *Gl. elbursorum* Brönnimann, Zaninetti, Bozorgina, Huber supports the Lower Triassic age determination. The Táska wells (Tás-1, -3) drilled through dark-grey marl, clayey limestone, and limestone with local variegated marl intercalations and with echinoderm and ostracod fragments. The foraminifer assemblage favors an emplacement into the Lower Triassic (*Meandrospira pusilla* (Ho), *Spirorbis phlyctaena* (Brönnimann et Zaninetti).

Middle Triassic

The borehole Som-1 exposed strongly fractured, partly authigenic, brecciated, grey, slightly clayey dolomite of about 60 m thickness. The rocks are strongly recrystallized; the original texture may have been micritic, with intraclasts and shrinkage cracks, with peloidic, occasionally pelletic or algal-field intercalations and with echinoderm and ostracod fragments. Its foraminifer fauna consists of *Glomospira meandrospiroides* Zaninetti et Whittaker and of *Meandrospira gigantea* Farabegoli. This latter species is transitional between the certainly Lower Triassic *Meandrospira pusilla* (Ho) and the Anisian *Meandrospira dinarica* Koch-Dev. et Pantic; thus, this section is believed to represent the transition between the Uppermost Scythian and the Lowermost Anisian.

Above it the Nubecularidaes-Calcitornellas horizon is found, with the species *Earlandia tintinniformis* (Misik), *Calcitornella* sp., *Ammodiscus parapriscus* Ho, *Glomospira tenuifistula* Ho, *Nodosinella* sp., *Glomospirella ammodiscoidea* Rauser, which unambiguously prove the Lower Anisian age of the formation. In the continuation of the sequence, strongly brecciated dolomite and partly dolomitized limestone strata alternate with laminitic dolomite-marl and dolomite of dolomite-marl cement, and with limestone breccia intercalations.

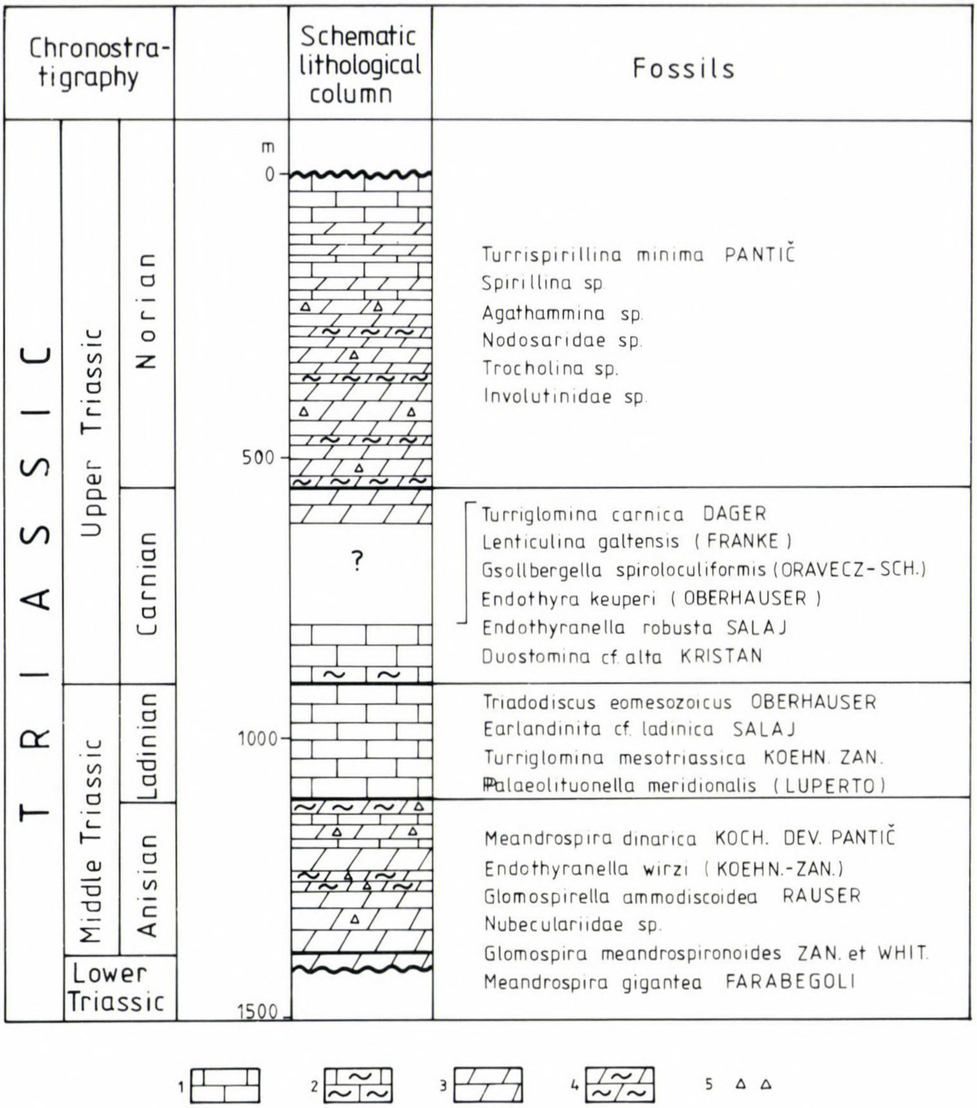


Fig. 4
Stratigraphic column of the Triassic formations in the eastern area of the Mid-Transdanubian Unit. 1. limestone; 2. lime-marl; 3. dolomite; 4. dolomitic-marl; 5. breccia

The foraminifer fauna of this sequence, i.e. *Meandrospira dinarica* Koch.-Dev. et Pantic, *Endothyranella wirzi* (Koehn., Zan.), *Diplotremina astrofimbriata* Kristan, *Glomospirella triphonensis* Baud et al., *Duostomina magna* Trifonova, point to Middle and Upper Anisian age. From the algae the following species were determined by O. Piros (pers. com.): *Physoporella pauciforata pauciforata, Teutloporella peniculiformis, Physoporella pauciforata undulata, Physoporella* *pauciforata sulcata, Oligoporella* sp., *Diplopora hexaster.* The flora indicates also Anisian age.

The Buzsák wells (Bu-5, -8, -13, -15) traversed a light-grey limestone and dolomite sequence, while the Táska wells (Tás-2, -4, -5) exposed brownish-grey limestone, dark-grey, strongly brecciated dolomite and dolomite strata. The fossil assemblage of these consists of calcareous algae, and foraminifera: *Trochammina almtalensis* Koehn-Zan., *Glomospira* sp., *Meandrospira dinarica* Kochansky-Devidé et Pantic, *Endothyranella wirzi* (Koehn-Zaninetti), *Earlandia amplimuralis* (Pantic), *Endothyra* sp., *Ophthalmidium* sp., *Diplotramina astrofimbriata* Kristan-Tollmann. Based on these fossils, the formations are unambiguously assigned to the Anisian. The bio- and lithofacies of the Anisian formations indicate intertidal-lagoonal facies.

Some reef-building organisms, e.g. *Bacinella ordinata* Pantic, *Cryptocoelia* cf. *zitteli* Ott, *Ceotinella* sp., were found in the light- and brownish grey, strongly tectonized, partly authigenic brecciated limestone traversed by the Buzsák (BuNy-1) and Somogysámson (Som-3) wells, to which a Ladinian–Carnian age is assigned. In this sequence, locally (BuNy-1 well) light-grey, slightly schistose tuff or tuffite (?) intercalations are found.

Calcareous algae fragments and foraminifera (*Haplophragmella inflata* Zaninetti et Brönnimann, *Earlandinita oberhauseri* Salaj, *Turriglomina mesotriassica* (Koehn- Zaninetti)), found in the brownish-grey limestone in several wells in the Somogysámson area, point to lagoonal facies and indicate a Ladinian age.

The Ladinian sequence of the borehole Som-1 consists predominantly of authigenic brecciated, light-to-dark-gray and brownish-grey limestone. Radial, calcitic and stromatactis-type fillings are common. Its predominant texture-type is intraclastic-plasticlastic bioplesparite. Common occurrence of algal encrustations, the algal nodules of Tubiphytes type and the "microproblematica" are characteristic of the Wetterstein reef facies. The fossils are *Tubiphytes obscurus* Maslov, *Tubiphytes carinthiacus* Flügel, *Basinella ordinata* Pantic, *Poriferitubus buseri* Sen., Dar., *Baccanella floriformis* Pantic, *Ladinella porata* Ott, *Panormidella aggregata* Sen., Dar. Common foraminifera are *Earlandinita soussi* Salaj, *Meandrospira deformata* Salaj, *Palaeolituonella meridionalis* (Luperto), *Agglutisolenia conica* Sen., Dar. In addition to these, echinoderm, ostracod and mollusc shell fragments and calcispongia sections are found.

Based upon the above-mentioned information, the exposed Ladinian formations indicate a near-reef environment, accumulating mainly on the fore-reef slopes.

Upper Triassic

In the area, the Upper Triassic sequence was traversed only by wells Som-1 and Igal-7. Based on petrological analogies, the light-grey, massive limestone exposed in the upper part of the Triassic sequence in the well BuNy-1 can be assigned to the Upper Triassic, too.

The Ladinian formations of well Som-1 show gradual transition to the Carnian strata, which are also represented by light-to-darker-grey and brownish-grey limestones. The authigenic, brecciated rock type becomes rarer upwards, and karstic cavities filled with calcite or terrigenous material become common. The characteristic microfacies types are biopelmicrite, often nearly completely recrystallized, and the onkoidic microfacies with algal encrustations and algal-mat fragments. It contains Sphinctozoa-type calcarea fragments, some poorly-preserved corals as well as ostracod and echinoderm shell fragments. The foraminifer fauna consists of *Lamelliconus multispirus* (Oberhauser), *Gsollbergella spiroloculiformis* (Oravecz-Scheffer), *Triadodiscus eomesozoiccus* (Oberhauser), *Kollmannita cordevolica* Fuchs, *Austrocolomia marschalli* Oberhauser, *Duostomina alta* Kristan, *Turriglomina carnica* Dager, *Nodosaria ordinata* Trifonova, and this proves the gradual transition from the Middle Triassic to the Carnian. In its uppermost part it contains pelagic fauna, principally pelecypod fragments, in addition to the calcarea fragments.

The sequence of well Igal-7 is built up by strata consisting of strongly tectonized grey dolomite-marl, breccia of dolomitic cement consisting of dolomite and limestone fragments, and brecciated limestone. The characteristic texture types are the algalaminites and intraclastic pelmicrite, or the pelsparite with birdseyes and shrinkage cracks filled by sparite or clayey micrite, and with stromatactis-like cavity-fillings. Subordinately, ooidic-onkoidic pelmicrite and rocks of pelsparite texture can also be observed.

The fossil content is very poor; in addition to ostracod, echinoderm and mollusc shell fragments, it contains a few foraminifera: *Turrispirillina minima* Pantic, *Spirillina* sp., *Agathammina* sp., *Nodosaridae* sp., *Ammodiscidae* sp., *Trocholina* sp., *Involutinidae* sp. These unambiguously prove the Norian age of the sequence but the possibility that the lowermost part is of Late Carnian age cannot be excluded. The texture and the fauna assemblage indicate an intertidal and subtidal shallow marine sedimentary environment. Nevertheless, the Spirillinae are pelagic plankton foraminifera.

Consequently, the Upper Triassic formations explored in the area (Fig. 4) represent shallow marine carbonatic platform facies. At the beginning of the Carnian, the Wetterstein-type platform evolution still continued, with the deposition of fore-reef slope sediments. The nature of the middle part of the Carnian is unknown so far. The Dachstein-type platform evolution probably began in the Upper Carnian, through the formation of dolomitic rock types. In the Norian, Dachstein Limestone is known, although without definite Lofer cyclicity. The locally observable ooidic-onkoidic texture and the pelagic foraminifera indicate an environment close to the outer margin of the platform.

Formations of the southwestern area (II)

The Semjénháza, Murakeresztúr, Pátró, Liszó, Pat and Inke hydrocarbon exploration wells traversed strongly diagenetized, sometimes slightly anchizonal-altered pelites and volcanics, as well as shallow marine carbonates which do not, or hardly, show the imprints of anchimetamorphism (Fig. 5).

The key borehole Iharosberény drilled through formations similar to those above, but here continuous coring clearly showed the relationship between the formations.

It can be seen that, in the continuous sequence between and beneath the grey shale and silty shale, the limestone strata do not, or hardly, show metamorphic features. Thus, in the case of the samples deriving from discontinuous core sampling of the hydrocarbon exploration wells, one must not necessarily assume that the contacts between the penetrated anchimetamorphic formations and those without these features are in all cases tectonic.

Lower and Middle Triassic

The very low-grade metamorphic claystone, dolomite, brecciated dolomite, anhydritic shale, sandstone, quartzite, limestone and gypsum-bearing dolomite strata traversed by the Semjénháza wells (Sem-2, -3) were assigned to the Upper Permian Triassic on the basis of analogies from the Bükk Mountains.

The strata of well Bagola-2 containing *Glomospira* sp., *Trochammina* sp., *Agathammina* sp., Nodosaridae and recrystallized Dasycladacea were emplaced in the upper part of the Middle Triassic (Ladinian), and partly in the lowermost Upper Triassic. In a borehole of Murakeresztur from black radiolarite clasts of a volcanoclastic sequence the following Ladinian radiolarian fauna was found by L. Dosztály (pers. com.): *Baumgartneria* sp., *Falcispongus calcaneum, Gombe- rellus* cf. *hircicornus, Neopylentonema mesotriassica, Pentaspongodiscus ladinicus, Plafkerium(?) nazarovi, Pseudostylosphaera coccostyla, Pseudostylosphaera* cf. *goestlingensis.*

Upper Triassic

Based on petrographic analogies, the very low-grade metamorphic shale, siliceous shale, vulcanite and limestone strata exposed by the other hydrocarbon exploration wells of the region were previously assigned to a range from the Early Paleozoic to the Jurassic.

The hydrocarbon exploratory well Inke-1 penetrated a sequence consisting of very low-grade, deep water black shales, as well as of shallow-water limestone, acidic and intermediary volcanics and serpentinite. Based on the radiolarians found in the black shale, the formations proved to be of Carnian age (Kozur, H. in Chikán et al. 1985).

The Iharosberény-1 structural well traversed, below the Miocene, basal breccia consisting primarily of Triassic limestone detritus, dark-grey shale, siliceous shale, silty shale and limestone strata containing thin, dark-grey radiolarite intercalations, with strongly altered silicified volcanics in the upper part.

In the lower part of the sequence, consisting of carbonatic rocks, biopelmicrite and biopelsparite, as well as clasts of these rock-types cemented with radial calcite spar (intraclastite), are the predominating microfacies types. Intraclasts are frayed or angular, locally slightly elongated (mainly in the clayey parts)

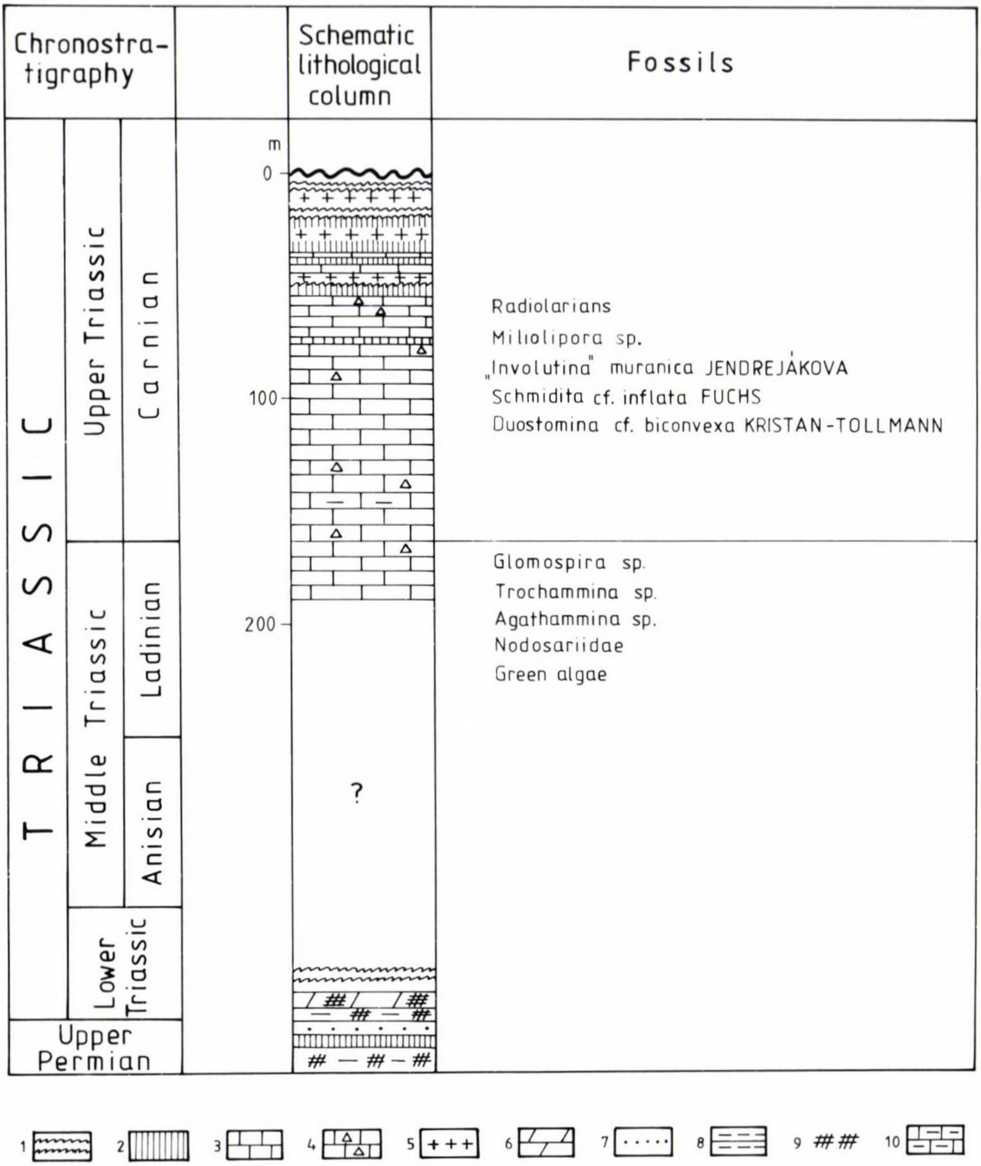


Fig. 5
Stratigraphic column of the Triassic formations in the southwestern of the Mid-Transdanubian Unit. 1. siltstone, shale; 2. quartzite, radiolarite; 3. limestone; 4. authigenic brecciated limestone; 5. volcanite 6. dolomite; 7. sandstone; 8. mudstone, 9. anhidrite, gipsum

due to the very low-grade metamorphism. They often show fenestral lamination (alga-mat breccia). Bioclasts are echinoderm, mollusc, coral, bryozoan mud, calcarea fragments. Algal encrustations, green algae, foraminifera and ostracods are common.

The alteration of the rocks is indicated by the authigenic quartz crystals, the quartz-rimmed dolomite crystals and by the zonal dolomite–rimmed quartz crystals.

In the upper part of the carbonatic sequence slopec brecciated structure is characteristic. The size of the intraclasts and bioclasts is great (exceeds the core diameter), but only a few cm for the most part. Echinoderm fragments, foraminifera, thin and thick-wall pelecypod shell fragments, red algal nodules and coral fragments are found as bioclasts. Radiolarians are also common; moreover, thin radiolarite intercalations can also be observed.

The cement is mixed calcitic (sparite and microsparite) and siliceous. The cavity and fissure fillings are made up predominantly of quartz and chalcedony. The clasts are often surrounded by calcite and quartz crystals of radial-pectinated intergrowth.

Foraminifera of the carbonatic segment, i.e. *"Involutina" muranica* Jendrejakova, *Schmidita* cf. *inflata* Fuchs, *Duostomina* cf. *biconvexa* Kristan-Tollmann, *Oberhauserella mesotriassica* (Ober.), *Miliolipora* sp., *Variostoma* sp., *Lenticulina* sp., *Ophthalmidium* sp., *Palaeospiroplectammina* sp., *Diplotremina* sp., *Endothyra* sp., *Endothyranella* sp., *Pachyphloides* sp. prove the Carnian, and probably also the Upper Ladinian age of the sequence.

Above the carbonate section, the shale-silty shale and siliceous shale levels are found which often have a considerable carbonate content, and underwent very low-grade metamorphism. They are identical to the schists sequence in well Inke-I.

Based on the fossil assemblage and lithofacies of the sequence studied (Fig. 5), it is probable that the lower section was formed on the carbonate platform, as well as in the smaller depressions and lagoons developed on the platform and on the marginal reef slopes. The upper carbonatic section was formed on the proximal belt of the platform margin slope. This is indicated by the large amount of red algal remnants, the mingling of thin and thick shell fragments, the radial fibrous sparitic cavity infillings and by the lithoclast accumulations. The overlying pelitic, siliceous sediments, also containing volcanic intercalations, are the sediments of the relatively deep basin.

Comparison of the Permian-Triassic formations with the sequences of other structural units

The Lower Permian clastic sequences of Újfalu, Buzsák and Karád can be fairly well correlated with the clastic Trogkofel strata of the Karawanken Mountains of Austria. The term Trogkofel Limestone included, for a long time, only the light-grey reef formations of the Carnian Alps and of the Western Karawanken. Based on the studies of the sixties and seventies (Ramovs 1963, 1974; Ramovs and Kochansky-Devidé 1965; Kochansky-Devidé 1973; Flügel 1971), however, it was proven that this reef formation is adjoined by a conglomerate-sandstone sequence in the Eastern Karawanken and in the Velebit Mountains. In its higher parts, biogenic limestone banks and reef-breccia lenses are intercalated in it, which are contemporaneous with the Trogkofel reef limestones of the Carnian Alps and of the Western Karawanken.

The carbonatic sequence of Újfalu can probably be correlated with the dolomite sequence of the Velebit Mountains, initiating in the Upper Permian, though the lack of Neoschwagerina, characteristic of the Velebit sequence, does not permit us to prove this paleontologically. It is quite possible that this dolomitic formation may represent the coeval carbonate facies of the Gröden Sandstone.

The Bellerophon-bearing dolomitic-calcareous formation of Tab can be fairly well correlated with the dolomitic facies of the Bellerophon Formation of the Southern Alps areas.

The possibility of correlation of the uncertain Újudvar Bellerophon limestone exists with both remote (Carnian Alps, Karawanken) and with neighboring (southern foreland of the NE part of the Transdanubian Hills) areas.

The geographically different Triassic formations relate to the different structural-paleogeographic relationships of the subunits. The regional connections of the Lower Triassic and Anisian sequences of intertidal, subtidal, intrashelf basin and carbonate platform formations, revealed in hydrocarbon exploration areas of the northwestern part of the Mid-Transdanubian Unit, cannot be unambiguously determined; this does not, however, exclude the possibility of identification with the Transdanubian Hills (Bérczi-Makk 1976).

Based on few data and poorish microfauna, the radiolaritic, tuffitic limestones of the eastern part of the studied area, assigned to the Ladinian, can be correlated conditionally with similar Buchenstein Formations of the Transdanubian Hills. Based upon the investigations of Árkai (1988, 1990), these formations of the Sávoly unit "underwent diagenetic alterations corresponding to the medium stage of diagenesis. This low degree of alteration agrees with that of the conspicuously low temperature of post-Hercynian formations of the Transdanubian Hills".

The foraminifer fauna found in the platform marginal formations of Upper Triassic (Carnian) age contains the forms characteristic of the Cassian strata. Based on the position of the fossil-free dolomite found in only one well, and on petrological analogies, this formation may perhaps be related to the Hauptdolomit Formation. It is not impossible that the dark-grey limestones of Nagybakónak are representatives of the Kössen Formation. The youngest platform carbonates can be fairly well correlated with the Dachstein Limestone.

In summing up, it can be stated that, in the strip of the northwestern part of the Mid-Transdanubian Unit, directly south of the Balaton Line (i.e. of the metamorphite-granitoid range), Triassic sequences are found which seem to be more similar to the formations of Southern Karawanken than to ones of the Transdanubian Hills.

The formations of the eastern part of the Mid-Transdanubian Unit are dissimilar to the formations of the Transdanubian Hills, and of the Bükk and Mecsek Mountains, in their structure and fossil assemblages.

The development of the Triassic sequence differs considerably from that of the Mecsek Mountains. In the case of the Transdanubian Hills and Bükk Mountains, the lack of Middle Triassic volcanics is the most conspicuous characteristic, and from the paleogeographic point of view it should be noted that an essential difference exists in fossil assemblages.

The Wetterstein-type reef formation traversed by the wells is found in the Northern Calcareous Alps, in the Northern Karawanken, in the Dinarides and in the Inner Western Carpathians.

A similar carbonatic platform facies sequence without volcanics was described from the Tara Mountains of Bosnia (Andelkovic 1976; Pantic et al. 1977), which is a nappe fragment from the southwestern margin of the Drina-Ivanjica Unit.

The foraminifer fauna found in the Norian of Igal also support the relationship with the Dinarides.

The carbonatic, pelitic complex of the southwestern area, containing acidic, intermediary volcanics and serpentinite, and which underwent anchizonal metamorphism, shows features relating to the Dinarides and presumably belongs to a nappe unit of lower position. Nevertheless, it is not excluded that this is the Middle Cretaceous olistrostrome-mélange-like Repno Complex, known in the Ivanscica and Kalnik Mountains, which contains blocks of older Cretaceous and Triassic rocks with basalt, gabbro and serpentinite bodies.

Conclusions

1. In the basement of the Mid-Transdanubian structural unit, lying between the Central Hungarian and the Balaton Lines, exploratory wells revealed Permian-Triassic sequences which differ from the those of the Transdanubian Hills lying in the northern neighborhood, and of the Tisza Unit (Mecsek Subunit) to the south of it, and which also display differences within the area itself, and in the grade of alteration.

Thus, the basement of the Mid-Transdanubian Unit is not akin to its geological environment and is itself of non-uniform buildup. The subunits probably attained their present-day juxtaposed position along displacement planes parallel to the zone strike and it seems to be probable that nappe formation occurred in earlier stages of the structural evolution, also complicating the structural setting.

2. The marine Permian formations, as well as the Triassic formations, indicate that blocks of the Mid-Transdanubian Unit may derive from the area of the western termination of the Carnian Alps, Southern Karawanken and of the Western Dinarides, but could have been parts of different paleogeographical-structural units of the region in the Late Paleozoic to Early Mesozoic period.

Nevertheless, it cannot be excluded that some broken-off fragments of the Transdanubian Hills Unit could have been emplaced in the strip lying in the southern margin of the Balaton Line.

3. The recent, and only rather incompletely known, geological buildup is the result of subsequent (Upper Cretaceous–Paleogene) tectonic movements. Both the more exact reconstruction of the paleogeographic situation, and the determination of the structural evolution, require more coring and thorough investigation of resulting core material.

Plate I

Foraminifers of the Lower Permian Trogkofel layers of the borehole U-I (3880.0–3886.5 m)

1. a. *Schubertella kingi* Dunbar et Skinner, b. *Schubertella australis* Thompson et Miller, c. *Globivalvulina* sp., 50x; 2. a. *Pseudoreichelina slovenica* (Kochansky-Devidé), b. *Darvasites* sp. 40x; 3. *Darvasites contractus* (Schellwien), 8x; 4. *Schubertella* cf. *paramelonica* Sulejmanov, 45x; 5. *Biwaella europaea* Kochansky-Devidé et Milanovic, 45x

Plate II

1. *Darvasites* cf. *contractus* (Schellwien), Karád-1. 12, 956.5–959.5 m 35x; 2. a. *Schubertella* sp. b. *Tetrataxis* sp., Karád-1. 12., 956.5–959.5 m 30x; 3. *Darvasites* cf. *contractus* (Schellwien), Karád-1. 12, 956.5–959.5 m 35x; 4. *Lunucammixna* sp., Buzsák-5. 24, 756.5–762.5 m 50x; 5. *Miliolidae* sp., Buzsák-5. 24, 756.5–762.5 m 50x; 6. *Globivalvulina vulgaris* Morozova, Buzsák-5. 24, 756.5–762.5 m 100x

Plate III

Foraminifers of the Triassic developments of the NW part of the Middle Transdanubian Belt

1-2. *Meandrospira pusilla* (Ho), Újudvar-12. 9, 2143.0–2148.0 m 130x; 3. Foram. indet sp. Budafa-502. 30, 3374.0–3375.5 m 60x; 4. *Nodosaria* sp. Budafa-502. 30, 3374.0–3375.5 m 100x; 5. *Meandrospira dinarica* Kochansky-Devidé et Pantic, Nagybakónak É-2. 4. 2438.0–2440.0 m 70x; 6. *Meandrospiranella samueli* Salaj, Budafa-502. 33, 3476.0–3478.0 m 140x; 7. *Ammobaculites radstadtensis* Kristan- Tollmann, Budafa-502. 30, 3374.0–3375.5 m 60x; 8. *Duostomina alta* Kristan-Tollmann, Újudvar-12. 3, 1939.0–1949.0 m 100x

Plate IV

Foraminifers of the Triassic developments of the NW part of the Middle Transdanubian Belt

1. *Turriglomina robusta* Bérczi-Makk, Sávoly-9. 3, 1406.0–1410.0 m 100x; 2. *Ophthalmidium triadicum* Kristan, Sávoly-9. 3, 1406.0–1410.0 m 100x; 3. *Triadodiscus eomesozoicus* (Oberhauser), Újudvar-7. 19, 2529.0–2530.0 m 60x; 4. *Trochamminidae* sp., Újudvar-7. 19, 2529.0–2530.0 m 100x; 5. *Duostominidae* sp. Nagybakónak-2. 6, 2542.0–2559.0 m 100x; 6. *Duostominidae* sp., Újudvar-7. 19, 2529.0–2530.0 m 100x; 7. *Glomospirella expansa* Kristan-Tollmann, Nagybakónak-2. 7, 2600.0–2608.0 m 100x; 8. *Trochammina almtalensis* Koehn–Zaninetti, Nagybakónak-2. 3, 2460.0–2477.5 m 100x

Plate V

Foraminifers of the Triassic developments of the NW part of the Middle Transdanubian Belt

1. *Trochammina alpina* Kristan-Tollmann, Újudvar-7. 13, 2418.0–2435.0 m 100x; 2. *Aulotortus tenuis* (Kristan), Sávoly-8. 9, 1943.0–1950.0 m 50x; 3. *Aulotortus friedli* (Kristan-Tollmann), Sávoly-8. 6, 1838.0–1844.0 m 50x; 4. *Aulotortus sinuosus* Weynschenk, Sávoly-8. 6, 1838.0–1844.0 m 50x; 5. *Aulotortus tumidus* (Kristan-Tollmann), Újudvar-7. 14, 2435.0–2448.0 m 50x; 6. *Aulotortus* cf. *tenuis* (Kristan), Sávoly-8. 9, 1943?.O–1950.0 m 50x; 7. *Aulotortus sinuosus* Weynschenk, Sávoly-4. 6, 1724.5–1741.0 m 50x; 8. *Aulotortus sinuosus* Weynschenk, Újudvar-7. 14, 2435.0–2448.0 m 50x

Plate VI

Foraminifers of the borehole Som-1

1. *Duostomina* cf. *alta* Kristan-Tollmann, 800.5 m 100x; 2. *Nubecularidae*, 1398.6 m 100x; 3. *Austrocolomia cordevolica* Oberhauser, 833.0 m 45x; 4. *Glomospirella* cf. *triphonensis* Baud et al., 1268.0 m 130x; 5. *Meandrospira dinarica* Koch. Dev.-Pantic, 1268.0 m 130x; 6. *Valvulina azzouzi* Salaj, 829.5 m 75x; 7. *Glomospira* cf. *sinensis* Ho, 1254.3 m 130x; 8. *Endothyranella wirzi* Zaninetti, 1254.3 m 130x; 9. *Glomospirella elbursorum* Brönnimann et al., 1400.2 m 240x; 10. *Reophax* sp., 800.5 m 45x;

Plate VII

Foraminifers of the boreholes Igal-7 and Iharosberény-1

1-4. *Turrispirillina minima* Pantic, Igal-7. 1039.6 m 100x; 5. *Diplotremina* sp,, Iharosberény-1. 1874.0 m 50x; 6. *Miliolipora* ? sp., Iharosberény-1. 1882.6 m 100x; 7. *"Involutina" muranica* Jendrejakova, Iharosberény-1. 1877.5 m 70x; 8. *Schmidita* cf. *inflata* Fuchs, Iharosberény-1. 1883.4 m 100x; 9. *Lenticulina* sp., Iharosberény-1. 1880.0 m 70x; 10. *Variostoma* sp., Iharosberény-1. 1877.5 m 70x; 11. *Duostomina* cf. *biconvexa* Kristan-Tollmann, Iharosberény-1. 1868.3 m 70x

Plate VIII

Microfacies photos from the Triassic sequence of the borehole Som-1

1. Pelagic shell fragments in pelsparites with algal mats (Tubiphytes?), 934.4 m 52x; 2. *Physoporella puciforata sulcata*, 1253.2 m 52x

Plate IX

Microfacies photos from the Triassic sequence of the borehole Som-1

1. Birdseye structures in pelsparites with shell fragments (dolomite), 1364.8 m 52x; 2. Alternating graded spathic and microspathic bands (laminae) containing peloid and pellet grains (dolomite), 1341.8 m 52x

Plate X

Microfacies photos from the Triassic sequence of the borehole Iharosberény-1

1. Authigenic quartz margins on dolomite crystals, 1876.0 m 50x; 2. Strongly silicificated intraclastic sparites with a Solonoporacea section, 1874.0 m 50x; 3. Coral fragment, 1883.8 m 4. Strongly silicificated intraclastic sparites with a Dasycladacea section, 1870.6 m 50x

Plate XI

Microfacies photos from the Triassic sequence of the borehole Igal-7

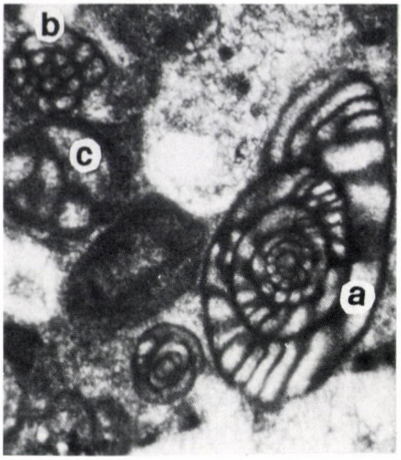
1. Intraclastic oncoidal pelsparites, 1068.7 m 52x; 2. Intraclastic sparites–microsparites with spathic cavity fills and a Dasycladacea section, 970.6 m 52x

Plate XII

Microfacies photos from the Triassic sequence of the borehole Igal-7

1. Intraclastic biopelsparites with an algal sectionand idiomorphic dolomite crystals, 1013.3 m 52x; 2. Laminite, 1378.0 m 52x

Plate I



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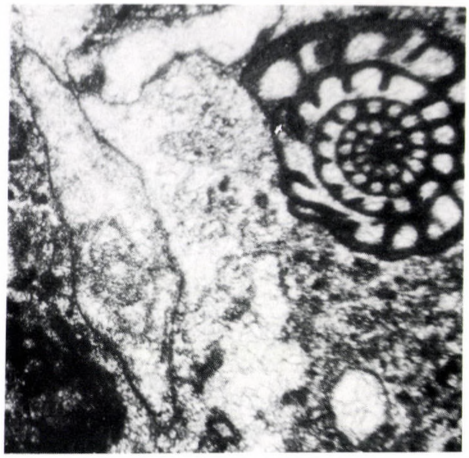

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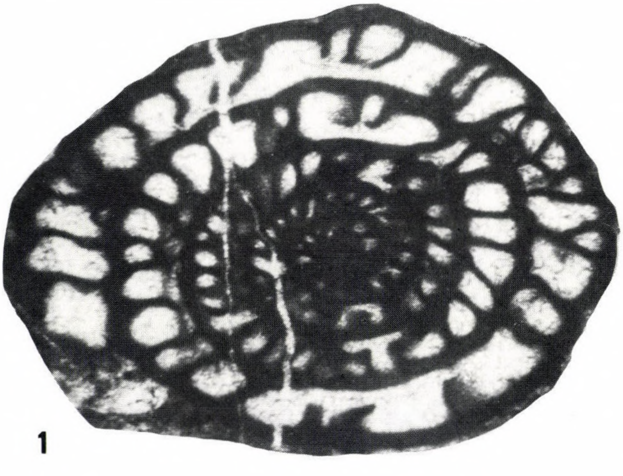

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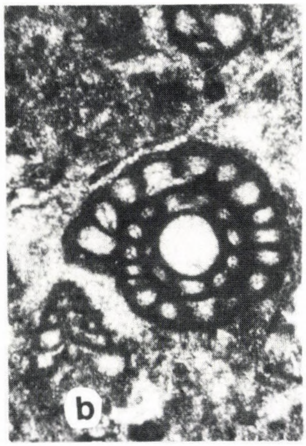

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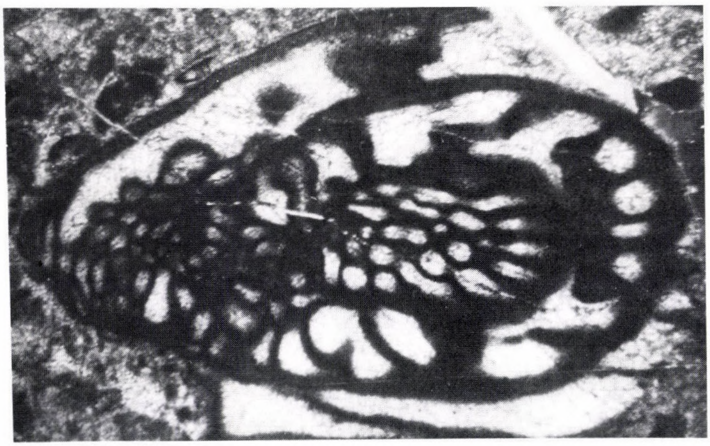

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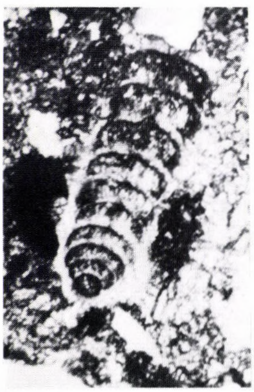

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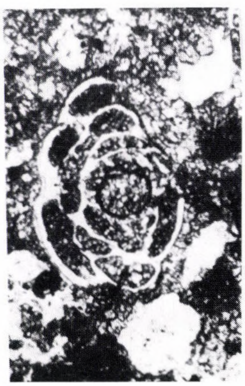

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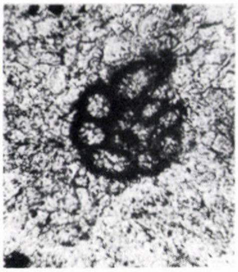

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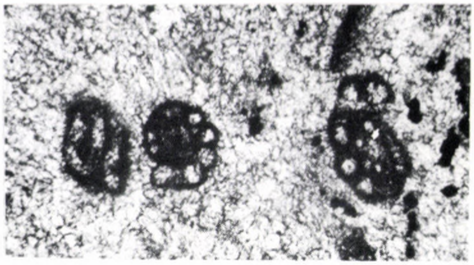

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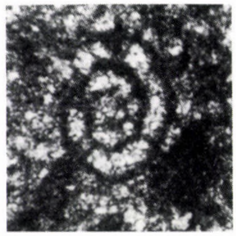

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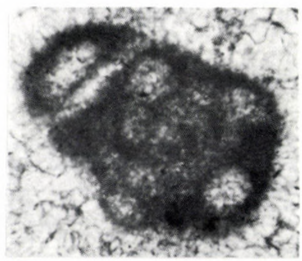

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Plate IV


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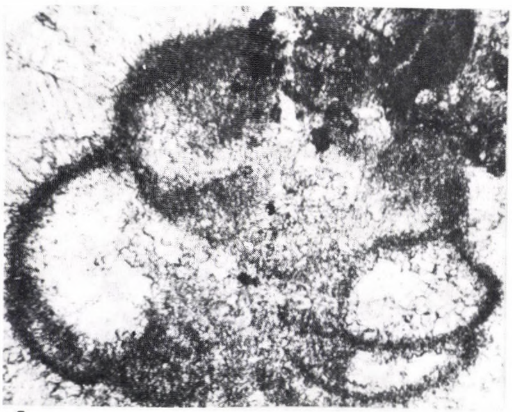
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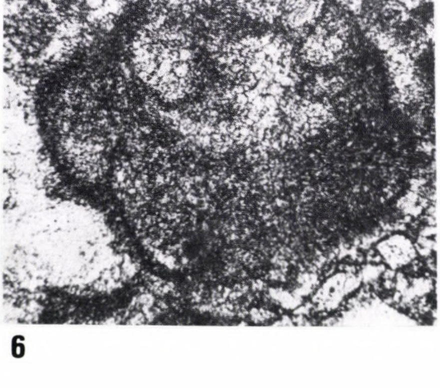
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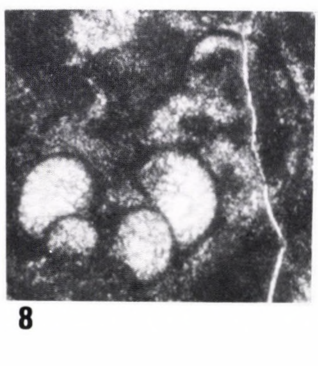
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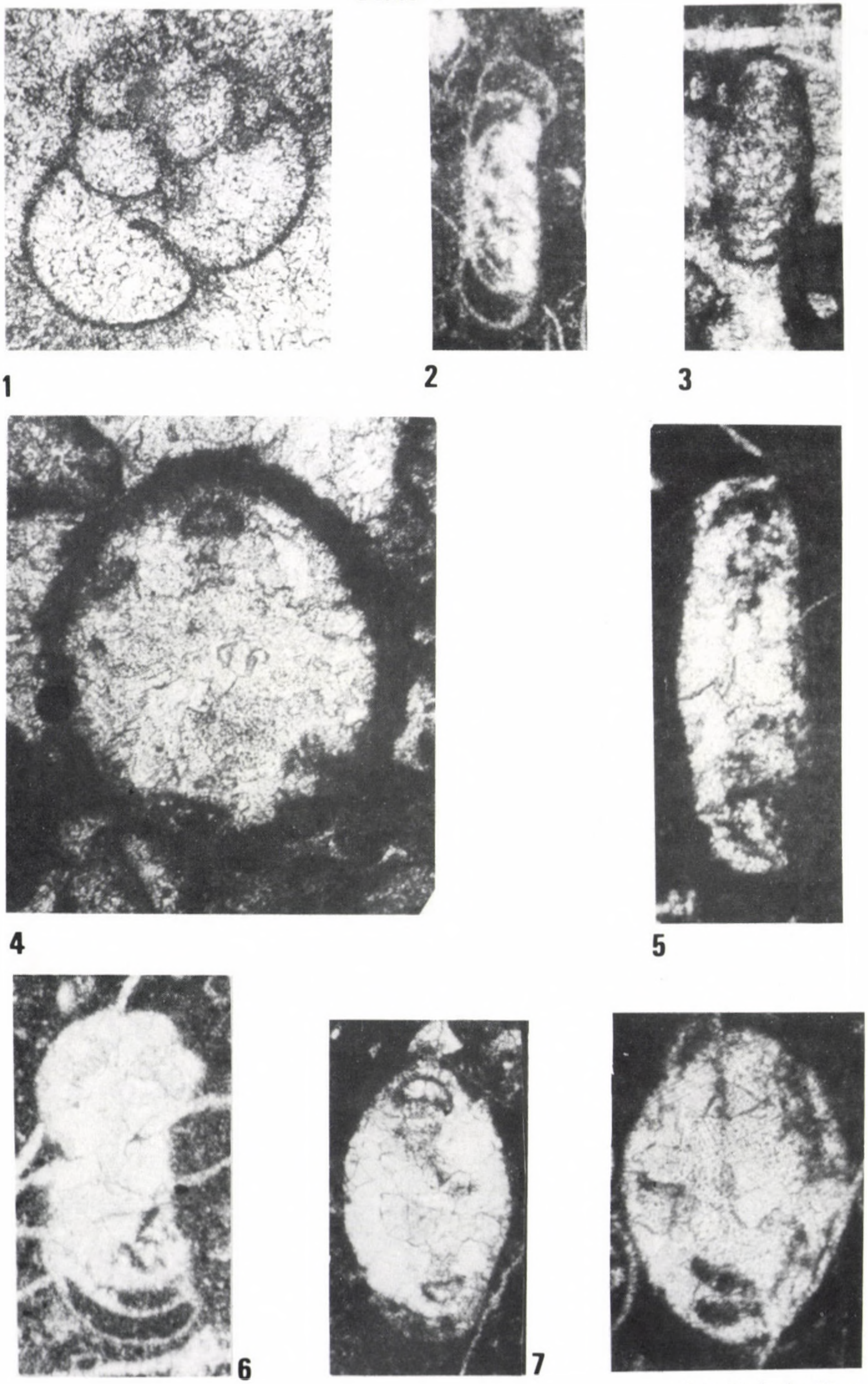

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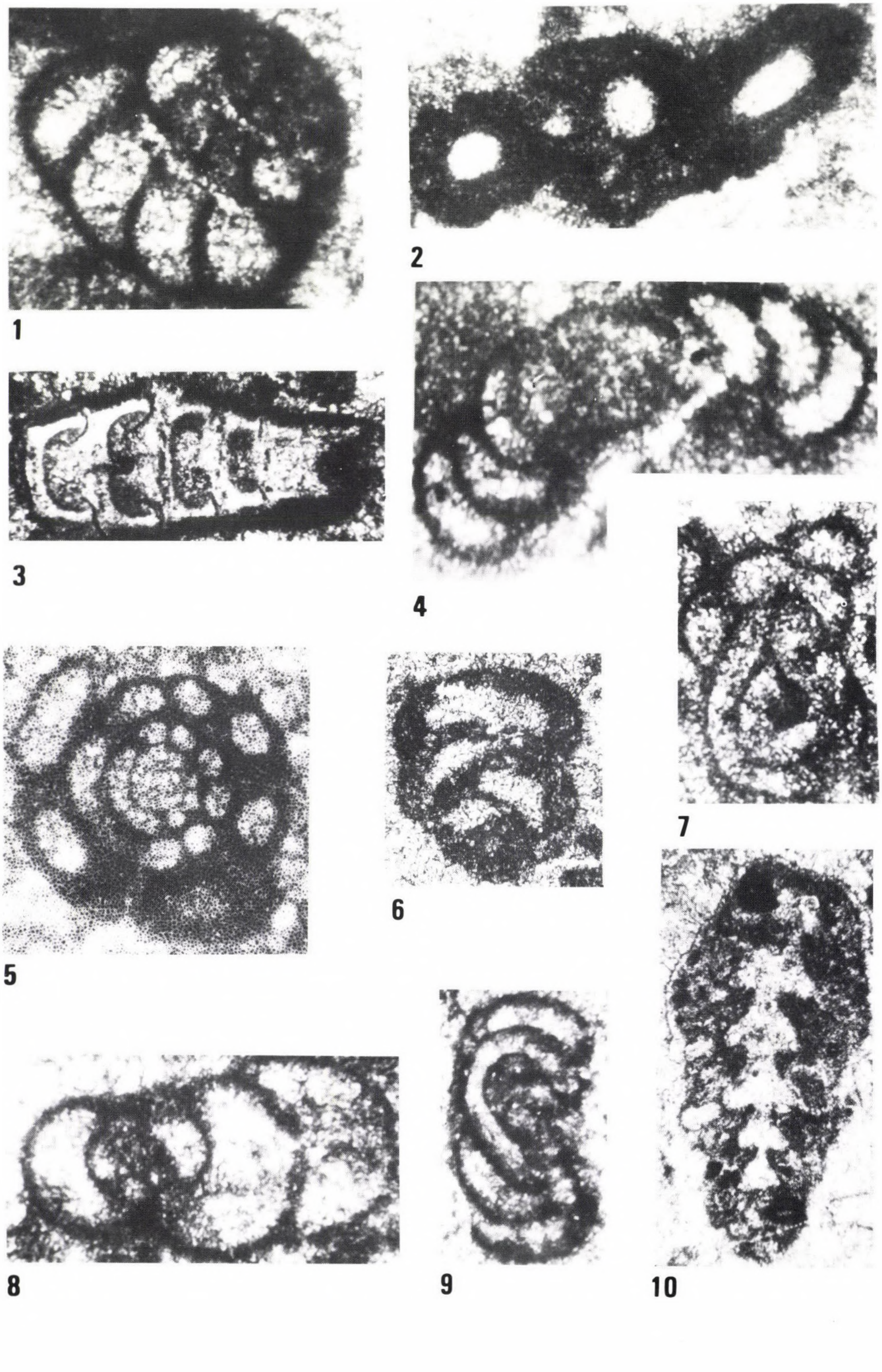

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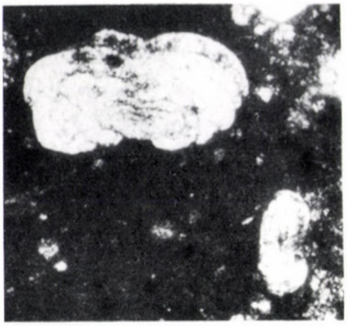

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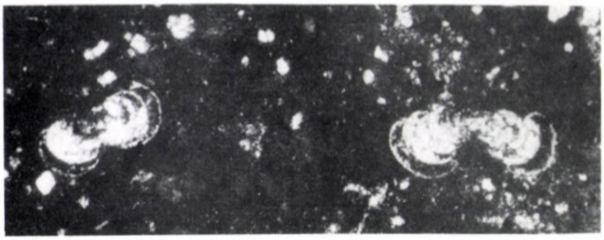

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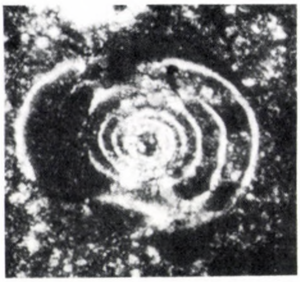

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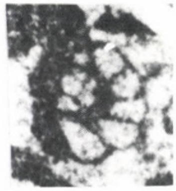

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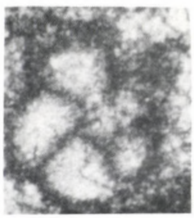

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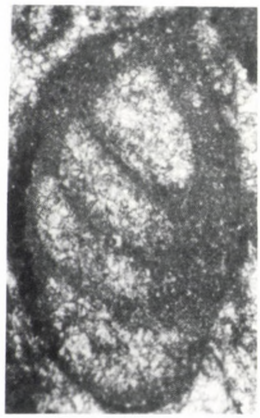

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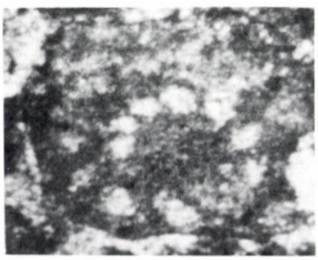

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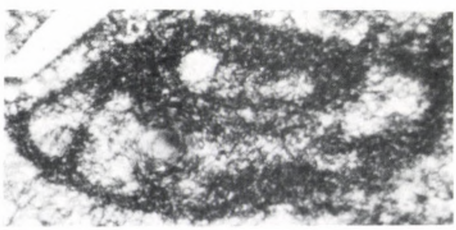

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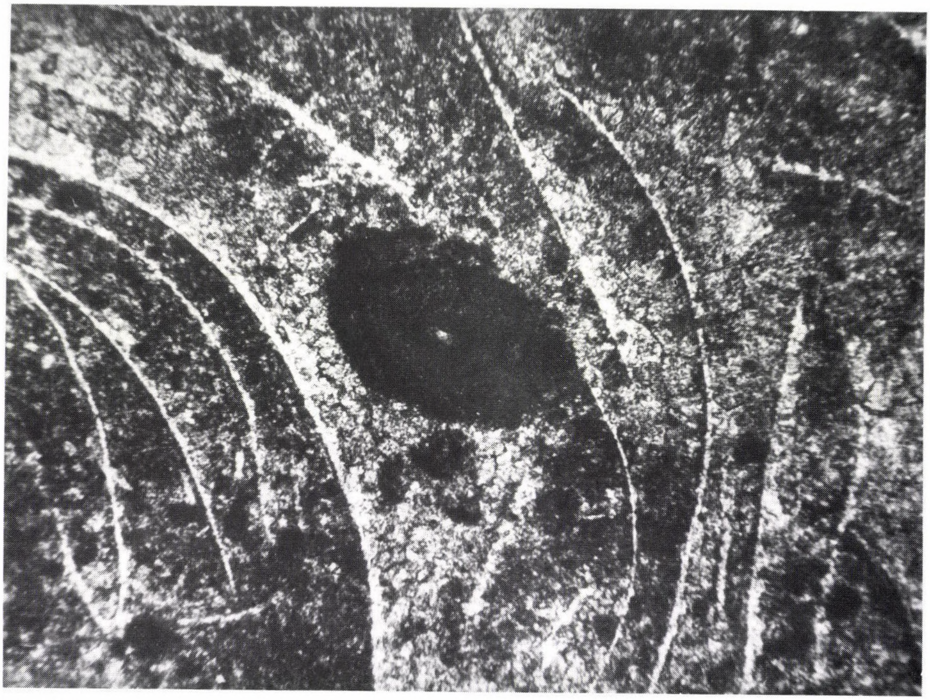

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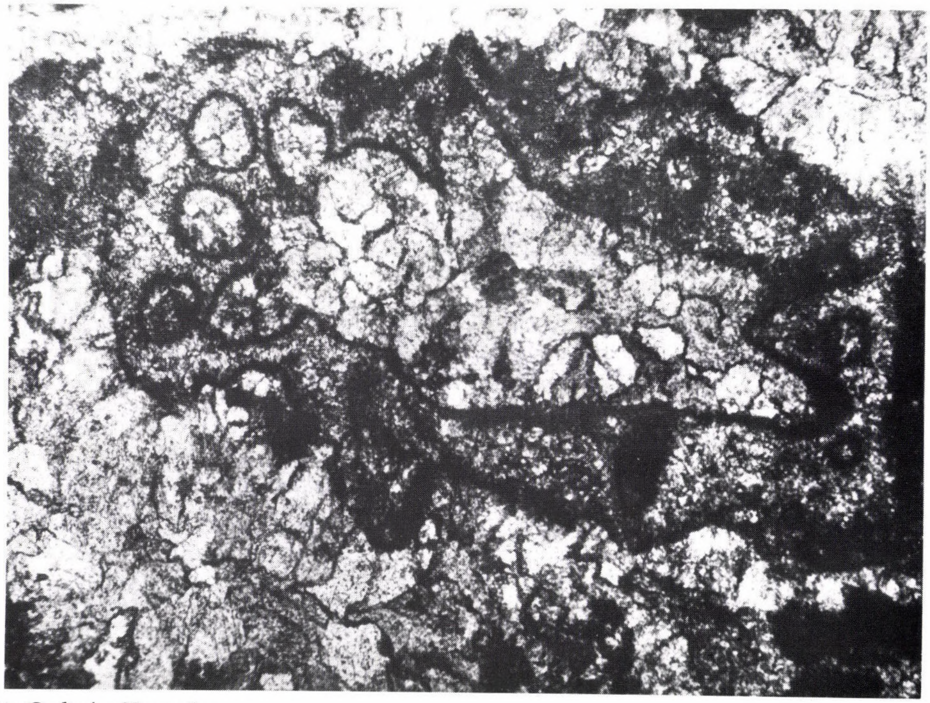

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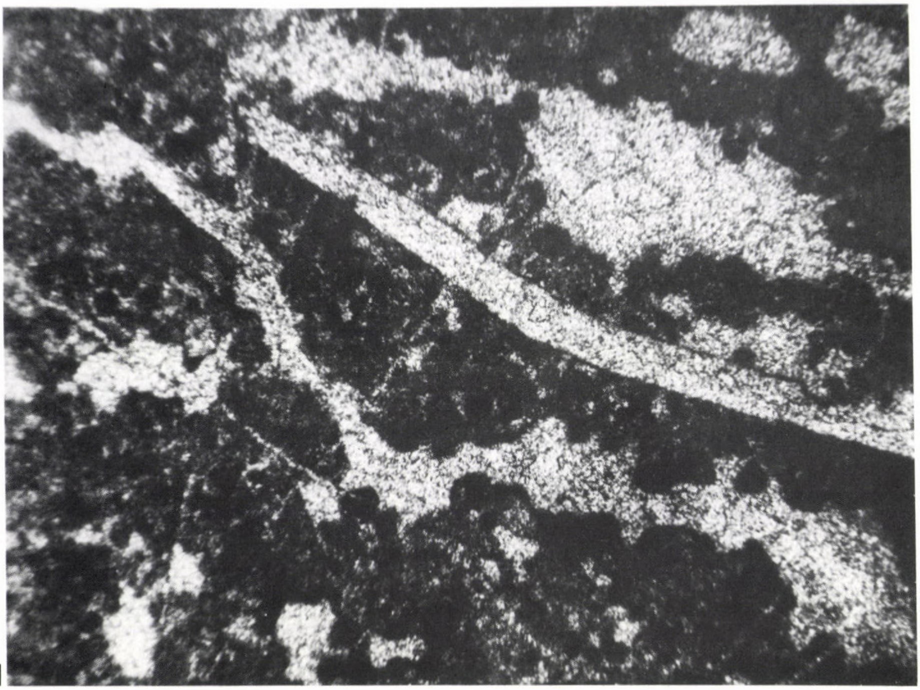

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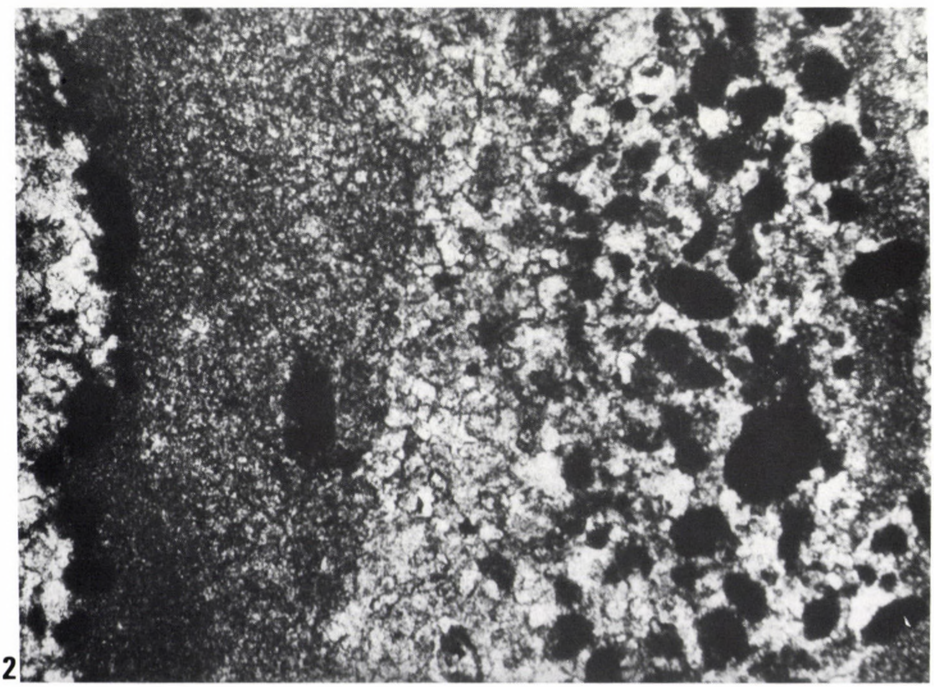

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Plate X

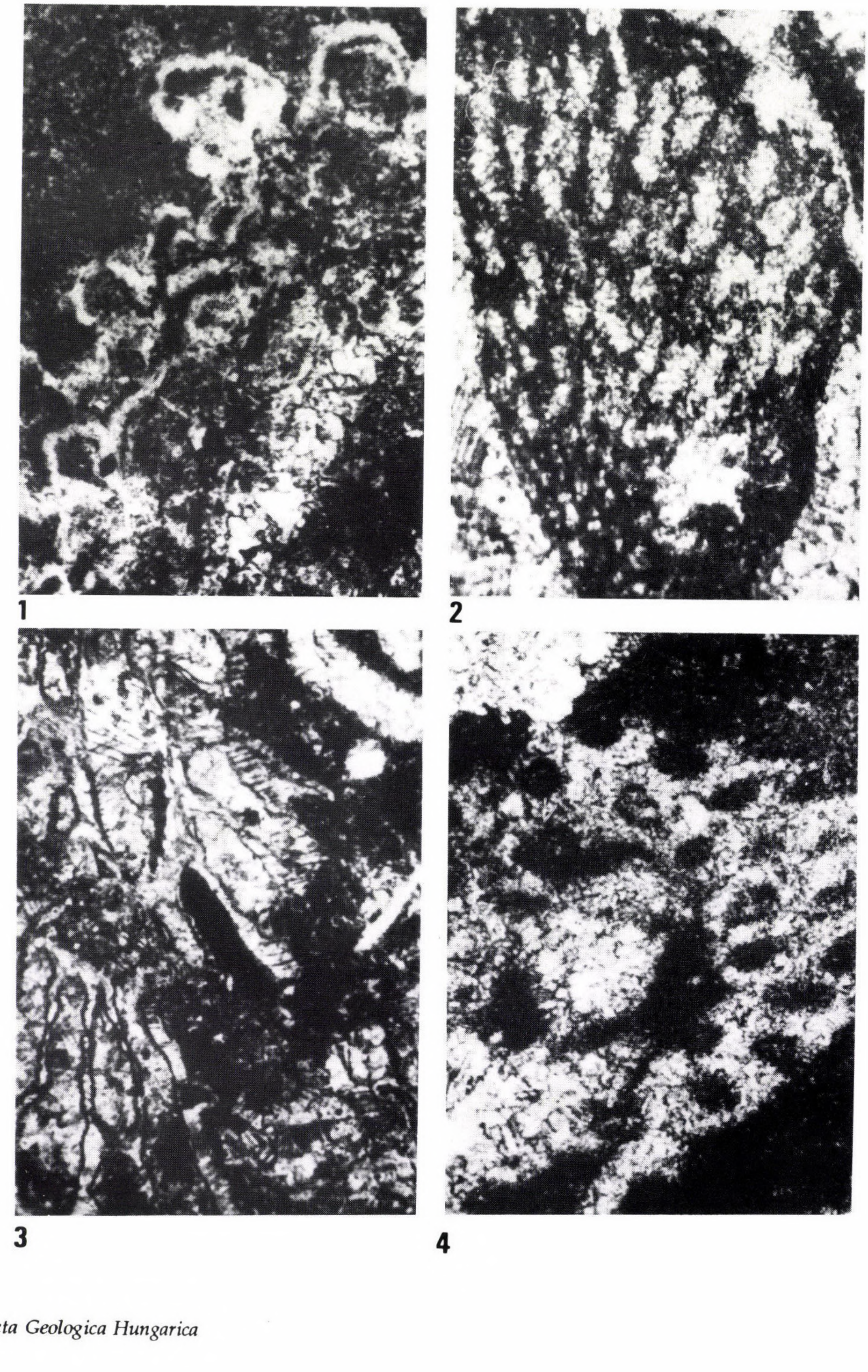
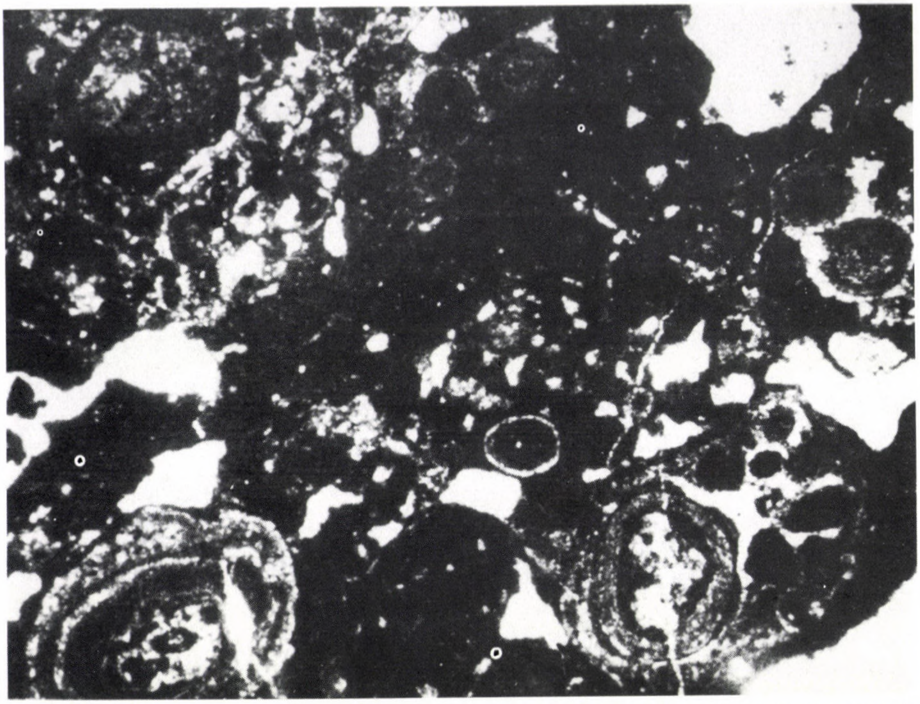


Plate XI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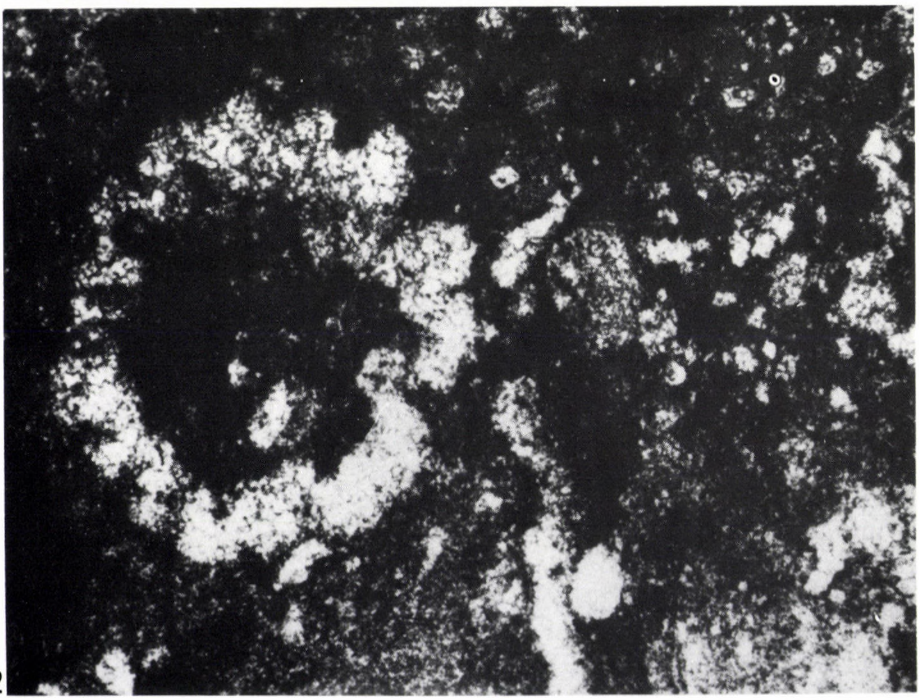
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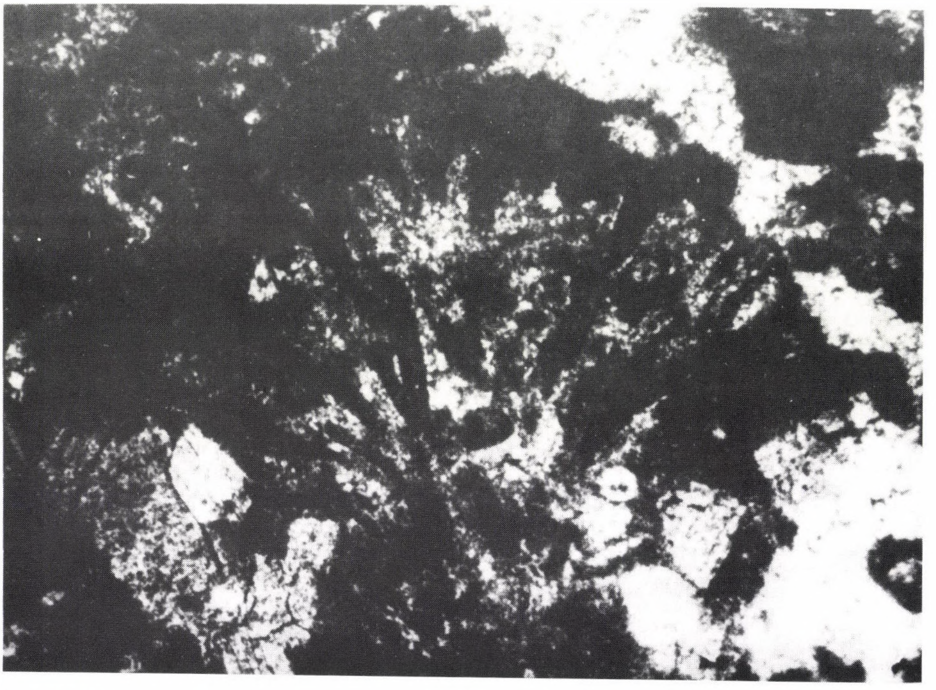
Plate XII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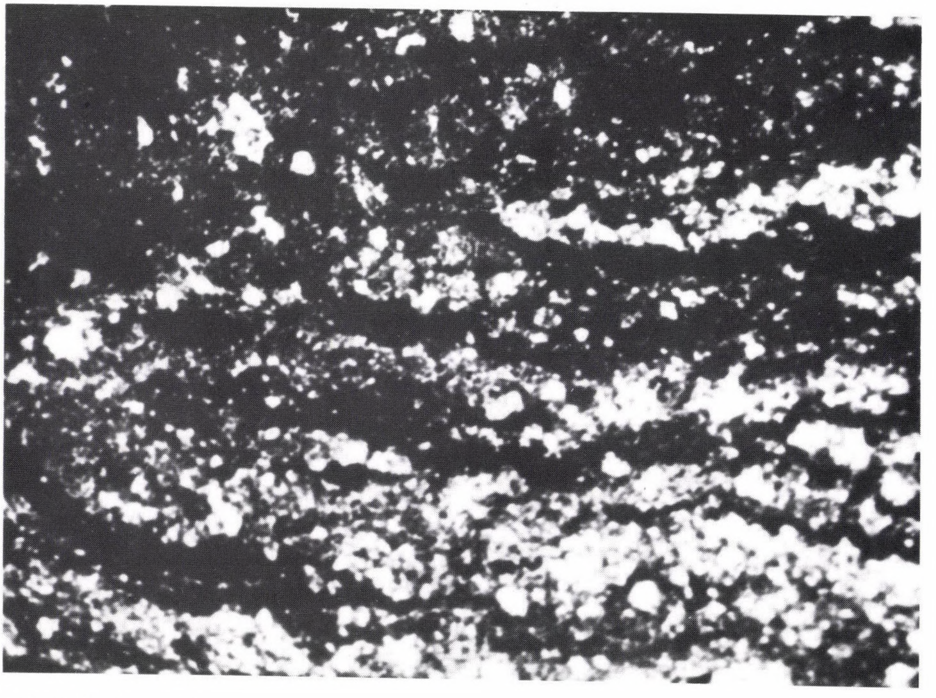
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Turriglomina Zaninetti in Limongi et al. (Foraminifera) species in Triassic formations, Aggtelek–Rudabánya Mts (Northern Hungary)

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A number of different forms of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) and *Turriglomina robusta* nov. sp. have been recovered very frequently from the Ladinian–Carnian pelagic slope (Nádaska Limestone Formation, Derenk Limestone Formation) and near-platform basinal sediments (Reifling Limestone Formation) in the Aggtelek–Rudabánya Mts (Northern Hungary).

Moving basinward from the carbonate platform frequency, the variability of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) increases with a simultaneous decrease of accompanying benthic foraminifera.

The only foraminifera species in the open marine microfilament sediments is the large-tested *Turriglomina robusta* nov. sp.

Key words: Triassic, biostratigraphy, Foraminifera, Turriglomina, Northern Hungary

Introduction

In processing the foraminiferal fauna (Bérczi-Makk in press) of the Triassic formations of the Aggtelek–Rudabánya Mts (Fig. 1) (Kovács et al. 1988, 1989), forms of the genus Turriglomina were frequently encountered in the Ladinian–Carnian sediments (Fig. 2). Individuals of three species have been identified.

Both in the Ladinian pelagic slope sediment (Nádaska Limestone Formation) and in the Lower Carnian near-platform basinal facies (Reifling Limestone Formation), the species *Turriglomina mesotriasica* (Koehn-Zaninetti) has been found to be predominant, with a spectacular richness of forms. In the platform carbonates of Northern Hungary (Wetterstein Limestone Formation), however, this species is very rarely found.

According to the published data, this form shows a wide areal extension in the Triassic basinal sediments of the Tethys, from the region of the Alps–Carpathians–Dinarides to China via Central and Eastern Asia (Zaninetti 1976; Kristan-Tollmann 1983; Limongi et al. 1987). Correspondingly, it is frequently encountered in the basinal Triassic of Northern Hungary as well.

The species *Turriglomina robusta* nov. sp (Foraminifera), with large test, thick wall and wide columella-axis, is a frequently encountered form in the Derenk

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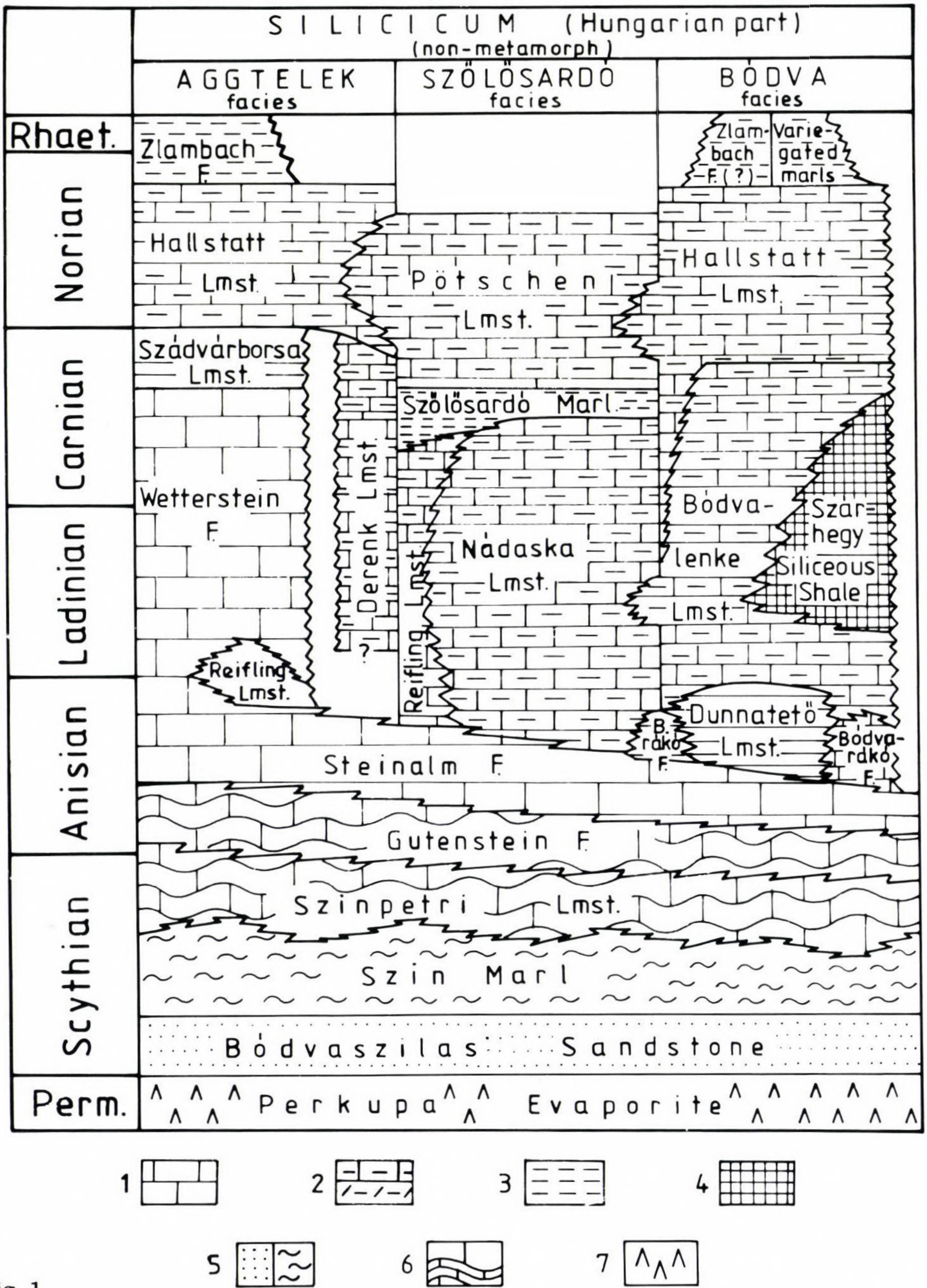


Fig. 1
Triassic formations of the Aggtelek–Rudabánya Mountains (acc. Kovács et al. 1989, simplified). 1. carbonate platform facies; 2. pelagic basinal facies; 3. basinal detrital facies (marly-argillaceous); 4. radiolarite; 5. shallow marine detrital and marly facies; 6. restricted lagoonal carbonate facies; 7. evaporite

Limestone Formation (a Ladinian–Carnian basinal limestone with a syndiagenetic brecciation/fracturing). In one sample of this formation, individuals of *Turriglomina conica* (He) have also been identified.

Turriglomina mesotriasica (Koehn-Zaninetti) in Northern Hungary

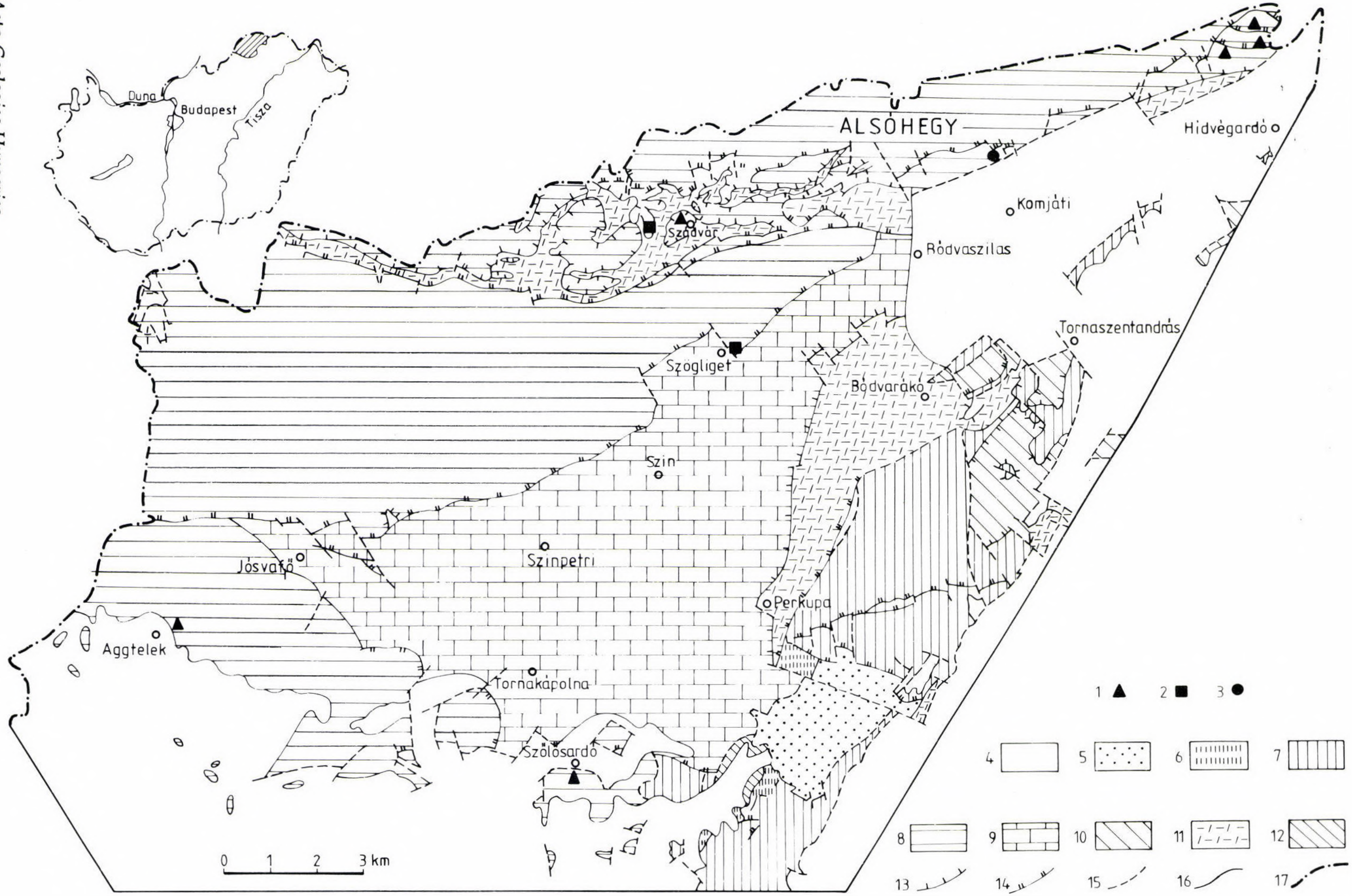
The relationship between the variability in form of the foraminifera species *Turriglomina mesotriasica* (Koehn-Zaninetti) and lithofacies as published by different authors (Limongi et al. 1987; Zaninetti et al. 1987; Hohenegger and Lein 1977) can be clearly studied in the section Alsóhegy-1 (Fig. 3).

Independently of the colour/shade of the host rock, two types of the species can be distinguished (Fig. 4). Type A1 (Plate I: 9a, textfig. 3/13–48 samples), which is more frequent, is slim, with a fairly consistent width (0.05–0.10 mm). Its trochospiral part consists of a number of whorls. The axis of the columella is thin (0.006–0.015 mm). Type A2 (Plate I: 6, 8, 9b, 10, 13; textfig. 3/52b sample) is more compact, and its width increases with the whorls getting younger (0.08–0.10 mm). The number of the whorls is less than that in type A1. The axis of the columella is thicker (0.016–0.022 mm), no meandrospiroid segment has been identified at any of the individuals. The increased width of the columella is supposed to control the width of trochospiral part, i.e. the width of the test (max 0.10 mm) is linearly proportional to the thickening of the columella (textfig. 3/52b sample).

The meandrospiroid segment mentioned above can only be observed in a couple of sections of A1 type forms. Even they are reduced variants of the meandrospiroid segment and generally consist of a spherical proloculum and one whorl of the second, undivided, tubular chamber (Plate I: 7, 12). The absence/sparsity of the meandrospiroid segment, its occurrence independently of the trochospiral segment (Plate I: 12) are not understood as an evolutional irregularity but as a secondary (sedimentary) phenomenon, i.e. a separation of the meandrospiroid segment from the trochospiral one during and/or after fossilization (textfig. 3/52a sample).

A large number of individuals of types A1 and A2 can be observed together in one sample (Sample 52, section Alsóhegy-1, Plate I: 9) from a dark-grey, near-platform Carnian limestone (Reifling Limestone Formation). In the above-mentioned sample (Sample 52, Alsóhegy-1), one individual represents type B of the species *Turriglomina mesotriasica* (Koehn-Zaninetti). This is a small-tested foraminifer (its maximum width is 0.05 mm), consisting of a well-defined meandrospiroid segment and a trochospiral part of few whorls and fast growth rate (textfig. 3/52c sample). Distinguishing type A and B terms as microsphaerical and/or macrosphaerical generation has not been applied intentionally, since the former one is represented by a single B type individual.

It is not reasonable to distinguish types A1 and A2 of Northern Hungary as new species, since it cannot be excluded that type A1 represents an earlier, and type A2 a later (more developed/aged) period of the individuals of the species



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Turriglomina mesotriasica (Koehn-Zaninetti). This "one species" theory is supported by the fact that individuals of type A2 with meandrospiroid segments have not been found yet. There is no doubt that observations made so far are not adequate to draw a final conclusion. In any case, however, the joint occurrence of types A1 and A2 in the same sample (Plate I: 9, textfig. 3/52 a sample) is a fact worth paying attention to.

The combined occurrence of the individuals of the *Turriglomina mesotriasica* (Koehn-Zaninetti), unrelated to the colour/shade of the host rock (Fig. 3), is an observation which contradicts that made in the Southern Appennines (Zaninetti et al. 1987; Limongi et al. 1978). The above-mentioned facts and observations in Northern Hungary may add to the complex problem of solving the nomenclatural and morphological disputes around this species.

Turriglomina robusta nov. sp. in Northern Hungary

The widespread occurrence of a large-tested foraminifer, *Turriglomina robusta* nov. sp., is striking in a Ladinian–Carnian pelagic slope sediment subjected to a syndiagenetic brecciation (Derenk Limestone Formation) (Kovács et al. 1988, 1989; Bérczi-Makk in press). The presence of an initial meandrospiroid segment observed in several individuals (Plate I: 3, 4) has made it unambiguous that this species belongs to the genus Turriglomina. The large test (max. width 0.20–0.25 mm), the thick wall (0.025–0.030 mm) and the very thick columella axis (0.04–0.06 mm) are striking. Rightly these extreme measures, significantly different from those of other species of this particular genus, are its distinctive features (Fig. 5). Until now, individuals of this species have been recovered only from purple-red, red-mottled, light grey and pink limestone of pelagic origin only. The accompanying fossil assemblage is characterized by the total absence of foraminifera, a massive occurrence of microfilaments and a frequent appearance of echinoid fragments.

← Fig. 2
Uncovered geological map of the Aggtelek–Rudabánya Mountains with the locations of Turriglomina species (acc. Grill 1989, simplified). 1. Turriglomina mesotriasica location; 2. Turriglomina robusta location; 3. Turriglomina conica location; 4. sand, clay, lignite freshwater limestone (Upper Miocene); 5. clay-marl (Jurassic); 6. clay-marl, limestone (Upper Triassic–Jurassic), 7. sandstone, limestone, marl (Triassic), Szilice nappe: 8. limestone, dolomite, marl (Middle–Upper Triassic). 9: sandstone, marl, limestone (Lower Triassic). Martonyi nappe: 10. limestone, marl (Middle–Upper Triassic). (Pseudo)autochthon: 11. anhydrite, dolomite, sandstone (Permian–Lower Triassic?) "Uppony-type" unit: 12. siliceous shale, limestone, marl (Paleozoic); 13. nappe border; 14. reverse fault; 15. fault; 16. lithological boundary; 17. national boundary

Paleontology

subordo: Miliolina Delage et Herouard
superfamilia: Cornuspiraceae Schultze
familia: Cornuspiridae Schultze
subfamilia: Meandrospirinae Sajdova
genus: Turriglomina Zaninetti in: Limongi et al. 1987

The test is strongly elongated, consists of a spherical proloculum and tubular second chamber with circular cross section. The initial coiling pattern is glomospiroid (meandrospiroid) which is followed by a segment consisting of a number of whorls with a long spiral axis, thus forming a central columella axis. The wall of the test is calcareous, microgranular. The slot is simple at the end of the tube. The dimorphism is manifested in the measures of the proloculum and the initial, glomospiroid segments as well as in the height of the whorls (the description of the diagnosis of the genus is after Zaninetti).

On the basis of observations made in Northern Hungary it is very probable that even the width of the columella plays a role in distinguishing taxa and generations.

Turriglomina robusta nov. sp.

Plate I, Photo 1–5

Derivato nominis: its large test

Locus typicus: sample No. T-434 from Alsóhegy (Silica Nappe, N. Hungary)

Stratum typicum: Derenk Limestone Formation, Middle–Upper Triassic (Ladinian–Carnian)

Holotype: Micropaleontological Collection of the Hungarian Geological Survey, Budapest

Material: Szádvár, profile AR-2, samples T-434, T-437, collected by Dr. S. Kovács; Szádvár, section collected by Mr. Gy. Don; Szögliget, Ubocs sample L–8408, collected by Mr Gy. Less.

Synonyma: 1983. *Lamelliconus turris* – Salaj; Borza K. et Samuel O. p. 149. pl. 128, Fig. 9.

Description: The elongated, free test consists of a proloculum and a tube-like second chamber, in its initial stage coiled-up like a Meandrospira (Plate I: 3, 4), which is followed by a trochospiral phase. The trochospiral part is produced by the high, closed un-umbilical, helicoidal coiling-up of the second, tube-like chamber (Plate I: 1, 2, 5). The helicoidal coiling-up consists of several (max 10) loops around the axis of the columella. The consecutive loops fit closely together and in an axial profile are found in an alternative pattern in either side of the very wide columella. The wall of the test is calcareous, microgranular and strongly recrystallized.

Dimension of the test: the diameter of the initial, coiled part: 0.160 mm, the greatest width of the trochospiral part: 0.200–0.250 mm, the height of the test:

Chronostratigraphy			Lithostratigraphy: formation	samples	rocks	Sed. facies	Biostratigraphy: Conodont zones
TRIASSIC	carnian	cordevolian	Reifling Limestone Formation	52	dark-grey limestone	basinal facies near the platform	Gondolella polygnathiformis
	ladinian	longobardian	Nádaska Limestone Formation	48	grey limestone	pelagic slope sediment deposited below the wave base	Gondolella inclinata
				47, 46	light-grey limestone		
				27, 26	brownish-red limestone		Gondolella n. sp. D.
		fassanian		17, 14	grey limestone		Gondolella trammeri
				13	reddish-grey limestone		

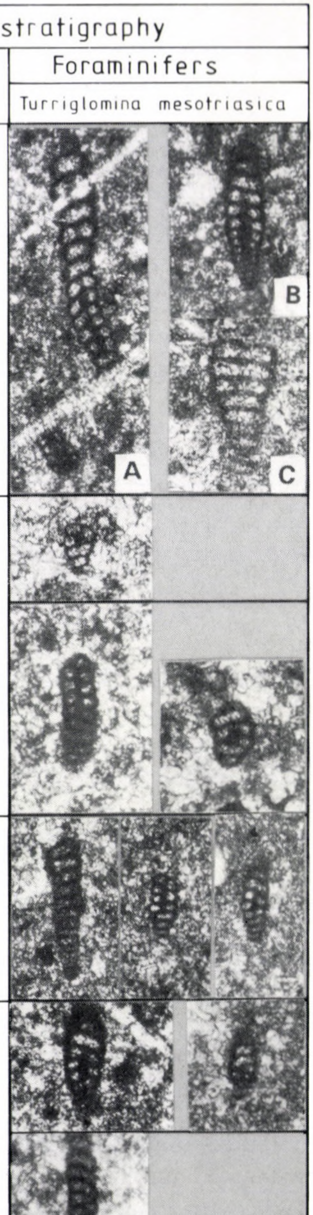


Fig. 3
Distribution of the variants of the Turriglomina mesotriasica in the section Alsóhegy-1 (NE–Alsóhegy)

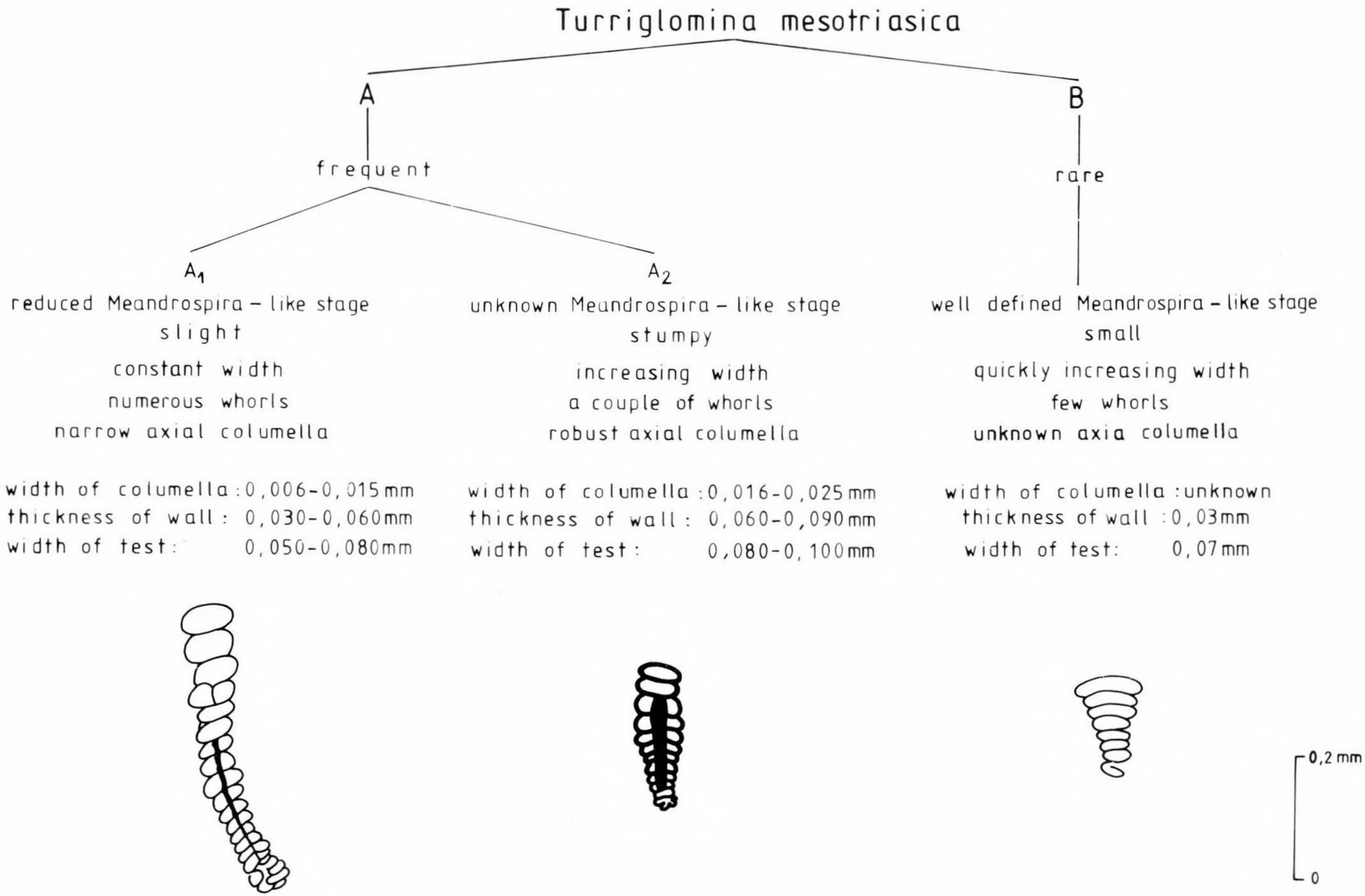


Fig. 4
Turriglomina mesotriasica types recovered from the Reifling Limestone Formation (sample 52) in the section Alsóhegy–1. (NE–Alsóhegy

		trochospira		thickness of wall in mm	width of columella in mm
		width in mm	heigth in mm		
mesotriasica group	T. mesotriasica (KOEHN-ZAN)	0,050–0,100	0,280	0,003–0,006	0,006 – 0,015
	T. carnica (DAGER)	0,042–0,054	0,132 – 0,162	0,006–0,010	0,006 – 0,008
	T. shizishanensis (KRIST.-TOLL.)	0 090	0,520	—	—
	T. lataxis (HE et YUE)	0,060	—	—	0,002
	T. magna (UROSEVIC)	0,170 – 0,190	0,490 – 0,510	—	—
	T. conica (HE)	0, 190	0,150 – 0,430	—	0,043
	T. robusta nov. sp.	0,200 – 0, 250	0,530 – 0,580	0,025	0,040 – 0,060
	T. scandonei (ZANINETTI et al)	0,220	0,800	0,040 – 0,050	0,020

Fig. 5
Characteristic measures of Triassic Turriglomina species

Lithostratigraphy			Derenk Limestone Formation	Reifling Limestone Formation	Nádaska Limestone Formation	Hallstatt Limestone Formation
Sedimentary facies			Basinal sediment affected by repeated cracking in moro or less compact state, in several generations	Basinal facies near the platform	Pelagic slope sediment deposited below the wawe base	Basin facies
Turriglomina specieses / Litho-facies			syndiagenetically brecciated limestone	grey banked limestone	protointraclastic varicoloured limestone	red, pink limestone
T. mesotriasica	A	A_1		X	X	
		A_2		X	+	•
	B			•		
T. conica			•	•		
T. robusta	A		X			
	B		X			X

1. X 2. + 3. •

Fig. 6
Locations with Turriglomina species in the Triassic formations of the Aggtelek–Rudabánya Mountains. 1. massive; 2. frequent; 3. rare

0.530–0.580 mm, the thickness of the wall of the test: 0.025–0.030 mm, the width of the columella: 0.040–0.060 mm.

Fossil assemblage: this is the only foraminifera species in the fossil assemblage of samples from the Derenk Limestone Formation recovered from Szádvár (Alsóhegy) and Ubocs side (Szögliget). The individuals of this species are accompanied by a large amount of microfilaments and fragments of echinoid shells.

Differential diagnosis: This species can be clearly distinguished from the Turriglomina species known so far by its large height, thick wall and wide axis of the columella. Although with its considerable height, the *Turriglomina robusta* nov. sp. is similar to the species *Turriglomina scandonei* Zaninetti et al. by its double wide columella axis, and by the fact that the width of the test is significantly growing during the lifetime of the individuals. This species can be distinguished from the species *Turriglomina scandonei* Zaninetti et al. In most cases the initial, meadrospiroid, coiled up phase is not known. This can be explained, partly by the predominance of oblique sections, partly by the fact that the meandrospiroid part can be easily split off from the trochospiral part during fossilization or thereafter.

Remarks: It is very probable that the individual of *Lamelliconus turris* (Frentzen) shown by Salaj. et al. (1983, Plate 128, Photo 9) belongs to the species *Turriglomina robusta* nov. sp. on the basis of the microgranular wall of the test, and the manner of coiling up. On the other hand, the lamellar appearance of the test wall composed of aragonite in the case of the genus Lamelliconus cannot be detected.

Turriglomina conica (He 1984)

Plate I, Photo 11

1968. *Trocholina procera* – Graciansky, P. Ch. et Lys, M. Plate 1, Photo 6
1984. *Turritellalla conica* – He Y. p. 429, Plate 1, Photos 14–15
1987. *"Turritellella" conica* – Limongi, P. et al (not illustrated)
1987. *Turriglomina conica* – Zaninetti, L. et al. Plate 1, Photos 4–6

Individuals of this species in Northern Hungary are known from the Lower Ladinian–Upper Carnian Derenk Limestone Formation with syndiagenetic brecciation. They were found only in sample KI–69 collected north of the village Komjáti from the Derenk Limestone Formation in the southern foreland of the carbonate platform of Alsóhegy.

On the basis of the measures of the test (the greatest width of the trochospiral part is 0.140 mm) and the wide axis of the columella (0.043 mm) they can be easily identified with the species described from the Anisian Limestone facies in China (Anshun Limestone Group).

In the faunal assemblage of this species microfilaments are the only elements.

Turriglomina mesotriasica (Koehn-Zaninetti 1969)

Plate I. Photos 6–10, 12–13

1945. *Turritellella* sp. – Wirz, A. p. 43. Plate 74, Photos 10–11
1967. *Turritella* sp. nov. – Hirsch, F. Plate 2, Photo 10
1968 *Turritellella mesotriasica* – Koehn-Zaninetti, L. (nomen nudum)
1969. *Turritellella mesotriasica* – Koehn-Zaninetti, L. p. 32, textfig. 4, Plate 3 Photos F, G
1971. *Turritellella mesotriasica* – Premoli-Silva, I. p. 328, Plate 25, Photos 1–2
1972. *Turritellella mesotriasica* – Canovic, M. et Kemenci, R. (not illustrated)
1972. ?*Turritellella mesotriasica* – Pantic, S. et Rampnoux, J.P. Plate 1, Photo 5
1973. *Turritellella mesotriasica* – Gazdzicki, a. et Zawidska, K. Plate 1, Photo 11
1974. *Turritellella mesotriasica* – Efimova, N.A. Plate 1, Photo 18
1974c. *Turritellella mesotriasica* – Pantic, S. Plate 1, Photo 9
1975 *Turritellella mesotriasica* – Gheorgian, D. p. 27, Plate 1, Photos 8–11
1975 *Turritellella mesotriasica* – Patrulius, D.; Gheorgian, D. et E. Plate 1, Photo 2
1976 *"Turritellella" mesotriasica* – Zaninetti, L. p. 105, Plate 4 Photo 1, Plate 8, Photos 13–19
1977. *Turritellella mesotriasica* – Hohenegger, J. et Lein, R. p. 227, textfig. 6, Plate 15 Photos 10–11, Plate 17, Photos 7–8
1977. *Turritellella mesotriasica* – Urosevic, D. p. 228. Plate 1, Photos 1–15
1977. *"Turritellella" mesotriasica* – Dager, Z. p. 50, Plate 1, Photo 8
1978. *Turritellella mesotriasica* – Mirauta, E. et Gheorgian, D. (not illustrated)
1978. *Turritellella mesotriasica* – Oravecz-Scheffer, A. (not illustrated)
1978a. *Turritellella carnica* – Dager, Z. p. 21, Plate 1, Photos 6–8
1978b. *Turritellella carnica* – Dager, Z. p. 50, Plate 1, Photos 15–16
1978. *"Turritellella" carnica* – Zaninetti, L. et Dager, Z. (not illustrated)
1978a. *Turritellella? mesotriasica* – Trifonova, Ek. Plate 2, Photo 11
1878c. *Turritellella mesotriasica* – Trifonova, Ek. Plate 2, Photo 1
1978 *"Turritellella" mesotriasica* – Zaninetti, L. et Dager, Z. (not illustrated)
1979. *"Turritellella" mesotriasica* – Resch, W. Plate 5, Photo 30
1979. *Turritellella* cf. *mesotriasica* – Babic, L., Gusic, I., Krystin, L. et Zupanic J. Plate 1, Photos 2–3
1980. *Turritellella mesotriasica* – He, Y. Plate 73, Photo 6
1980 *Turritellella mesotriasica* – Trifonova, Ek. (not illustrated)
1981. *Turritellella mesotriasica* – Brandner, R. et Resch, W. (not illustrated)
1982. *Turritellella carnica* – Courtin, B., Zaninetti, L., Altiner, D. et Decrouez, D. Plate 4, Photo 5
1982. *Turritellella mesotriasica* – Courtin, B., Zaninetti, L., Altiner, d. et Decrouez, D. Plate 4, Photo 5
1982. *Turritellella mesotriasica* – Trifonova, Ek. et Vapstarova, A. (not illustrated)
1983. *Turritellella mesotriasica* – Oravecz-Scheffer, A. (not illustrated)
1983. *Turritellella shizishanensis* – Kristan-Tollmann, E. p. 295, textfig. 3/18
1983. *"Turritellella" mesotriasica* – Salaj, J., Borza, K. et Samuel, O. p. 70, Plate 16, Photo 1, Plate 142, Photo 3
1983. *Turritellella mesotriasica* – Trifonova, Ek. (not illustrated)
1984. *Turritellella mesotriasica* – He, Y. p. 423., Plate 1, Photos 16–17
1985. *Turritellella mesotriasica* – Bérczi-Makk, A. Plate 2 ,Photo 3
1985. ?*Turritellella mesotriasica* – Dragastan. O., Papamikos, D. et Papanikos, P. Plate 1, Photo 1
1985 Turritellella mesotriasica – Satalov, G. et Trifonova, Ek. (not illustrated)
1985. ?*Turritellella carnica* – Dragastan, O., Papanikos, D. et Papanikos, P. Plate 1, Photos 2–3
1985. *Turritellella mesotriasica* – Trifonova, Ek. et Vaptsarova, A. Plate 1, Photo 5
1987. *Turritellella lataxis* – He, Y. et Yue, Z.L. p. 224., Plate, 2 Photos 3–5, 13–14
1987. *Turritellella mesotriasica* – He, Y. et Yue, Z.L. p. 202., Plate 1, Photos 16–22
1987. *Turritellella* cf. *mesotriasica* – He, Y. et Yue, Z.L. p. 203., Plate 2, Photos 1–2
1987. *Turritellella mesotriasica* Limongi, P. et al. textfig. 2

1987. *"Turritellella" carnica* – Limongi, P. eta al. (not illustrated)
1987. *Turriglomina carnica* – Zaninetti. L. et al. (not illustrated)
1987. *Turritellella carnica* – He, Y. et Yue, Z. L. P. 205., Plate 1 Photo 23
1987. *Turritellella mesotriasica* – Oravecz-Scheffer, A. Plate 16, Photos 2, 7
1987. *"Turritellella" mesotriasica* – Oravecz-Scheffer, A. Plate 28, Photo 1, Plate 31, Photos 1.3.5–8
1987. *Turriglomina mesotriasica* – Zaninetti, L. et al. Plate 1, Photo 2
1988. *Turriglomina mesotriasica* – Agip p. 46 (textfig)
1988. *Turritellella mesotriasica* – Canovic, M. et Kemenci, R. Plate 5, Photo 1
1988. *Turitellella?* sp. – Pirdeni, A. Plate 1, Photo 25
1988. *Turritellella mesotriasica* – Salaj, J., Trifonova, Ek. et Gheorgian, D. (not illustrated)

Concerning the areal and stratigraphic distribution of the species *Turriglomina mesotriasica* (Koehn-Zaninetti), the following points should be stated:

It is common: in the Ladinian Nádaska Limestone Formation (Fassaian–Longobardian), representing pelagic slope sedimentation (Northern Hungary, Geological Section Alsóhegy-1, samples 13, 14, 17, 26, 27, 46, 47, 48, samples Alsóhegy T-377, borehole Szőlősardó-1, 227.35–227.45 m) in the Carnian Reifling Limestone Formation (Cordevolian) representing an open basin near platform facies (Northern Hungary, Geological Section Alsóhegy-1, sample 52, Hidvégardó).

It is scattered: in the Hallstatt Limestone Formation representing basin sediments (sample Alsóhegy T–187 and sample Szádvár-26).

It is sparse: in the platform Wetterstein Limestone Formation (Northern Hungary, sample Aggtelek "J 1399").

Occurrence of the individuals of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) in limestone formations of Northern Hungary as listed above suggests their dominance in pelagic slope sediments deposited below the wave base and in basin sediments near to a platform.

Since the test is very fragile, full, non-fractured individuals could hardly be recovered from the sedimentary rocks involved. Consequently, the number and length of the whorls of the trochospiral part can not be taken as measures specifically characteristic of that particular species. The forams studied have been with no exception tests with oblique sections and of different evolutional stage.

Remarks: In relation to the individuals of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) recovered in Northern Hungary nomenclatural problems are encountered: it is inevitable to consider the possibility of viewing the species *Turriglomina carnica* (Dager), *"Turriglomina" lataxis* (He et Yue) and *"Turriglomina" shizishanensis* (Kristan-Tollmann) as synonyms of the *Turriglomina mesotriasica* (Koehn-Zaninetti). Until this revision takes place, the Turriglomina species mentioned above have been distinguished as belonging to the mesotriasica group (Fig. 5).

The morphology of the species *Turriglomina carnica* (Dager) is fully identical to the measures of *Turriglomina mesotriasica* (Koehn-Zaninetti). The only distinction is that the whorls of the trochospiral part are perpendicular to the axis. The tests recovered in Northern Hungary suggest, however, that the perpendicular or oblique position of the whorl to the axis is dependent upon the orientation of the thin section only.

As for the case of *"Turriglomina" lataxis* (He et Yue), it corresponds fairly to the trochospiral part of the species *Turriglomina carnica* (Dager) and *Turriglomina mesotriasica* (Koehn-Zaninetti). The only distinct feature may be a sparse occurrence of a non-thickened columella axis. Since the meandrospiroid phase of this particular species is not known, the name of the genus is consistently used with quotation mark. The trochospiral part of the test corresponds fairly well to that of the individuals of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) recovered in the Ladinian–Carnian formations of Northern Hungary (textfig. 3/47 sample) evidencing the synonym status suggested above.

Similarly, the meandrospiral phase of the species *"Turriglomina" shizishanensis* (Kristan-Tollmann) is not known either. As a consequence, its status as a species of this particular genus can be questioned. On the other hand, however, on the basis of its trochospiral part it can be identified as a well-developed individual of the species *Turriglomina mesotriasica* (Koehn-Zaninetti).

Conclusions

1. Frequency, areal distribution, measures, and the composition of the accompanying foram fauna of the species of the genus Turriglomina depends on the depositional environment (vicinity of the platform and/or basin) (Fig. 6).

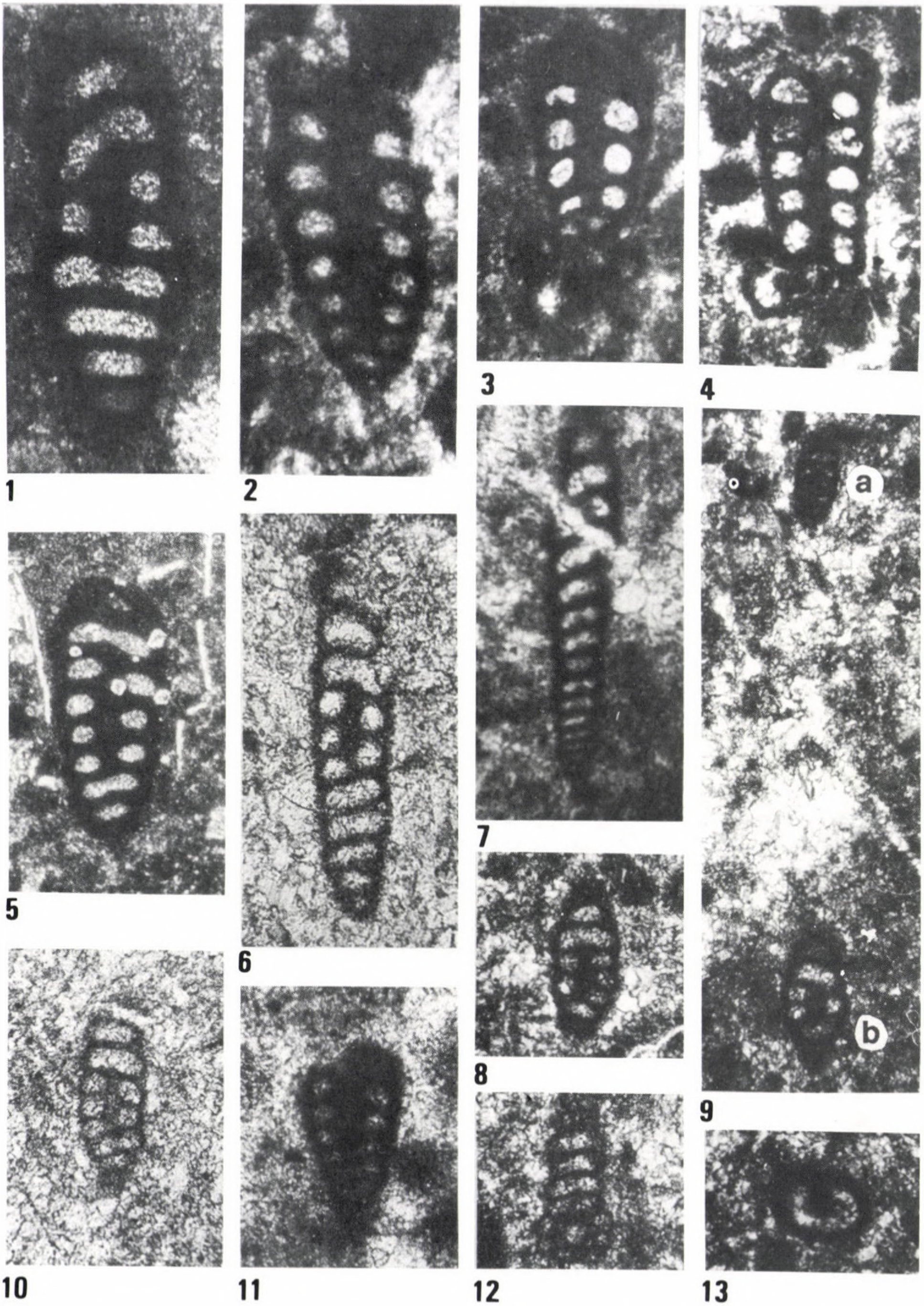
2. Small individuals of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) are very sparse in the platform carbonates of N Hungary (Wetterstein Limestone Formation), which are otherwise rich in benthos forams (Plate I: 7).

3. In the pelagic slope sediments (Nádaska Limestone Formation) the small individuals of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) are frequent (Plate I: 6, textfig. 3/13–48 samples), together with species of the genera Ophtalmidium, Arenovidalina of the family Nodosaridae.

Plate I

1–5. *Turriglomina robusta* nov. sp. from Ladinian–Carnian Derenk Limestone Formation. 1, 5. Szádvár, profile Alsóhegy-2, sample T-434, sampled by S. Kovács, x100, x80; 2,3. profile Szádvár-3, sampled by Gy. Don, 100x; 4. Szögliget–Ubocs side, sample L-8408/1, sampled by Gy. Less, 100x ; 6–10, 12, 13. *Turriglomina mesotriasica* (Koehn-Zaninetti) , 6. from Upper Longobardian Nádaska Limestone Formation, Alsáhegy, sample T-377, sampled by S. Kovács, Type: A_2, x100; 7.from Ladinian Wetterstein Limestone Formation, Aggtelek, sample "J 1399", sampled by L. Roth, Type A_1, x100; 8. from Cordevolian Reifling Limestone Formation, East Alsóhegy, Hidvégardó, sample 80027, sampled by Gy. Less, Type A_2, 100x; 9., 12, 13. from Cordevolian Reifling Limestone Formation, profile Alsóhegy-1, sample 52, sampled by S. Kovács, Type A_1: 9a, 12, A_2: 9b 13 all x100; 10. from Tuvalian Hallstatt Limestone Formation, profile Szádvár (Szv), sample 26, sampled by S. Kovács, Type: A_2, x100; 11. *Turriglomina conica* (He) from Ladinian–Carnian Derenk Limestone Formation, Komjáti (Kl), sample 69, sampled by S. Kovács, x100

Plate I



4. The richness of forms and frequency of the individuals of *Turriglomina mesotriasica* (Koehn-Zaninetti) is observed in the near-platform part of the basin facies (Reifling Limestone Formation, Plate I: 8, 9, 12, 13, textfig. sample 3/52), accompanied by a number of individuals of *Arenovidalina chialingchiangensis* (He). This suggests that moving basinwards from the carbonate platform, the proportion of the individuals of the species *Turriglomina mesotriasica* (Koehn-Zaninetti) grows at the expense of the other benthos forams.

5. In the syndiagenetically brecciated Derenk Limestone Formation (basin limestone with microfilaments) a frequent occurrence of the individuals of *Turriglomina robusta* nov. sp. (Plate I: 1–5) is characteristic, with a few forams of the family Nodosariidae.

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Geology and petrology of Egerian–Eggenburgian andesites from the easternmost parts of the Periadriatic zone in Hrvatsko Zagorje (Croatia)

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In this paper geological, petrological and geochemical data on Egerian–Eggenburgian andesites from the easternmost parts of the Periadriatic zone in the area of Hrvatsko Zagorje, located in the southwesternmost parts of the Pannonian Basin, are presented. Concordant geological and radiometric data indicate that the main volcanic activity took place in uppermost Egerian and lowermost Eggenburgian.

The volcanic activity produced andesites and more abundant pyroclastic roks. The andesites are represented by basic and acid varieties which correspond on average to tholeiitic medium–K basic andesite. Mineral assemblage includes predominant plagioclase, augite, hypersthene and biotite which were analyzed by microprobe. In addition, major and trace element data for the representative rock samples are presented. Geochemical data suggest that andesites could be genetically related to subduction processes. By contrast, field relations and radiometric data indicate that they have post-subduction features.

Key words: Periadriatic Lineament, southwestern Pannonian Basin, Egerian sedimentary country rocks, geological and isotopic ages, basic to acid andesites, pyroclastic rocks, major and trace elements, microprobe analysis, subduction, extension

Introduction

In the northern parts of Hrvatsko Zagorje three interrupted zones with exposures of Tertiary andesites are located. The southern zone stretches as a straight line from the area of Hum on the Sutla River to Varaždinske Toplice. Here, eleven separated masses of andesites and more common andesitic pyroclastic rocks occur along a distance of about 75 km and the same rocks were also penetrated by drilling in the area of Varaždinske Toplice. The northern zone can be traced along strike for about 35 km from Rogaška Slatina to Vinica. In the area of Rogaška Slatina andesites and tuffs are very common whereas in eastern parts of the zone only pyroclastic rocks occur. These two long zones are connected by the shorter transversal one stretching along the southwestern slopes of Mt. Ravna Gora (Fig. 1).

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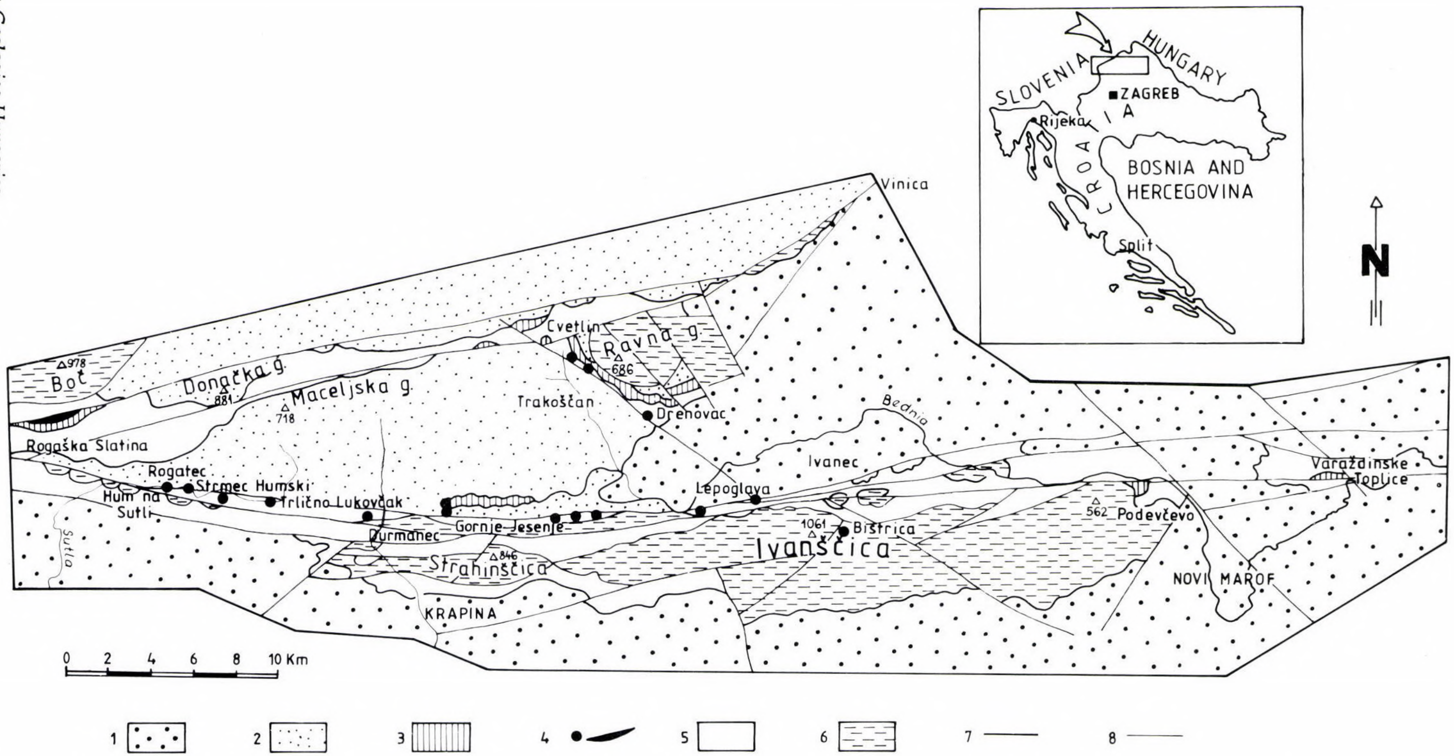

Fig. 1

Schematized geological map of the northwestern part of the area of Hrvatsko Zagorje. 1. Late Badenian–Pontian clastic sedimentary rocks, 2. Eggenburgian–Karpathian clastic sedimentary rocks, 3. Egerian pyroclastic rocks, 4. Egerian andesites, 5. Egerian clastic sedimentary rocks, 6. Mesozoic sedimentary rocks, mostly Triassic dolomites and limestones, 7. contact line, 8. fault

Andesites from separate localities of the northern parts of Hrvatsko Zagorje have been studied during the last 90 years by numerous authors. It is the aim of this paper to give geological, petrological and geochemical data on all the andesites which occur within the three mentioned zones in the northern parts of Hrvatsko Zagorje. The andesites, accompanied with pyroclastic rocks are spatially associated with the Egerian beds. K–Ar crystallization ages of andesites range from 22.8 to 19.7 Ma. Volcanic rocks are represented mostly by hypocrystalline–porphyritic augite andesite and subordinate biotite–augite andesite and rare transitional basaltic andesite. The presence of basaltic xenoliths and the inverse changes in the chemical composition of the rock-forming minerals indicate that the primary andesitic melt was reequilibrated during its solidification.

Literature Data

Gorjanović-Kramberger (1904a and 1904b) first studied andesites from the area of Hrvatsko Zagorje and presumed their Miocene age. He emphasized that these volcanic rocks are related to the "andesitic line" which stretches from eastern Slovenija through Hum on the Sutla River and to Lepoglava. Kišpatić (1909a and 1909b) determined these rocks as hypersthene andesites and dacites. The tectonized zone stretching from Rogaška Slatina to Vinica was named the "Donački fault" by Kossmat (1913).

Anić (1958) presumed Badenian age of the andesites of the area of Hrvatsko Zagorje. Golub and Brajdić (1969) and Crnković et al. (1970) presented additional petrochemical data on the andesites. Šimunic and Hečimovic (1979) pointed out that andesites occur along faults which are parallel to a horst–anticline stretching from the area of Hum on the Sutla River to Varaždinske Toplice. Šimunić et al. (1981) presumed Lower Miocene and Aničic and Jurisa (1985) Egerian age of andesites. Šimunić (1992) emphasized a polyphase character of Tertiary volcanic activity in the area of Hrvatsko Zagorje.

Geology

Andesites from the northern parts of Hrvatsko Zagorje occur in the areas which are made up mostly of Egerian and Eggenburgian sedimentary rocks. Most of the andesite occurrences are found along the fault zone stretching from the area of Hum on the Sutla River to Varaždinske Toplice (Fig. 1). The fault zone longitudinally intersects a large horst–anticline plunging gradually towards the east and reaching the Drava Depression.

The core of the horst–anticline consists of disturbed Paleozoic–Mesozoic formations with Egerian sedimentary rocks preserved in its flanks. These rocks are folded and faulted by reverse and thrust faults.

The Egerian is represented by sands, sandstones and marls with subordinate breccias and conglomerates, in some places interlayered with coal seams. These sediments were deposited in paralic environments on a "wat", i.e. the plain on which exchangeable marine, brackish and even freshwater conditions must have existed. The source area for the detrital material must have been located in the Alps. First traces of the volcanic activity can be recognized in the area of Gornje Jesenje where andesitic tuffs alternate with thick-bedded marls. These marls and the ones comformably underlying andesites of the area of Hum on the Sutla River contain the following Egerian microfossils: *Ammodiscus incertus* d'Orbigny, *Cyclammina acutidorsata* (Hantken), *C. placenta* (Reuss), *Spiroplektammina carinata* d'Orbigny, *Valvulina pectinata* Hantken, *Gladulina laevigata* d'Orbigny and others (Avanić et al. 1990).

The fault zone Hum on the Sutla River–Varaždinske Toplice goes on further to the west to the Smrekovec fault (Mioc 1978) and both of them represent the easternmost parts of the Periadriatic Lineament. In the area of Croatia, this fault was marked by different names. Gorjanović-Kramberger (1904a and 1904b) named it as "Hum–Brda–Željeznica row", Kober (1952) as "Narbe Zone", Sikosek (1958) as "Bordering Zone" etc. The easternmost part of the Periadriatic Lineament plunges under younger Neogene sedimentary rocks which form an anticline. The anticline was also registered below the Drava River alluvion and it can be presumed that it extends in the area of Hungary.

The position of andesite occurrences in the valley of the Bistrica tributory is ambiguous. These occurrences are not spatially related to the three fault zones mentioned above; the andesites are totally surrounded by Triassic dolomites probably as a result of younger horizontal movements.

In the area of Rogăska Slatina, the Periadriatic Lineament branches (Fig. 1) into a shank which was named the Donački fault by Kossmat (1913). The fault represents in fact a system of vertical faults which gradually get a northeast direction. Among them, two faults are most prominant because they controlled the volcanic activity. The Donački fault can be traced from Rogaska Slatina to Vinica and further to the east plunges under Quaternary and younger Neogene formations. It may be presumed that the fault extends further to the northeast as indicated by the occurrences of Egerian andesites which were penetrated by the oil wells Lopatinec and Mačkovec in the Mura Depression (Pamić 1992). The results of detailed field investigations suggest that the Donacki fault zone controls the mentioned horst–anticline whose core is made up of Egerian clastic sediments unconformably overlain by the Eggenburgian Macelj sandstones. In the central part of the core, different rocks in some places occur in the form of tectonic wedges ages of which vary from the Middle Carboniferous to the Eocene (Šimunić 1992). The largest masses of andesites and tuffs are found in the area of Rogaška Slatina which is located on the intersection of the Periadriatic Lineament and Donački fault (Aničić and Juriša 1985). Only tuff occurrences are known further to the northeast. In the area of Cvetlin village, in marls overlying tuffs the same fauna as in sedimentary rocks from the area of Jesenje is found and it can be concluded that the volcanic activity had started along the Donački fault by the end of the Egerian.

In the southwestern parts of Mt. Ravna Gora stretches a fault which connects the Periadriatic Lineament and Donački fault. Along this fault, three smaller andesite occurrences with larger quantities of tuffs occur. In this area, andesite pebbles at the base of the Macelj sandstones are very common. The pebbles, which are 10 to 40 cm in diametre, are well rounded and embeded in tuffaceous matrix. Earlier authors were of the opinion that they represent volcanic bombs but it was found out that they originated by weathering of andesites. Comparatively large size and well roundness of the pebbles indicate to fast erosion of a marked relief and steep shore.

Andesite volcanic activity started under marine environments but afterwards the whole area emerged and a new marine transgression, which had taken place by the beginning of the Eggenburgian, covered the marked relief. The precise thicknesses of flows cannot be determined due to the thick sedimentary cover. However, it is very probable that they are rarely more than 100 m thick. Andesite flows also include pyroclastic rocks which are well exposed in the area of Kameni Vrh in the neighbourhood of Lepoglava.

The interlayering of andesitic tuffs and Egerian marine sedimentary rocks represents evidence for the synsedimentary and submarine character of the volcanic activity. However, it is very probable that surficial parts of larger volcanic bodies might have solidified under subaerial conditions. This is indicated by the fact that Eggenburgian Macelj sandstones and conglomerates, which unconformably overlie the Egerian beds, are made up mostly of andesitic and tuffaceous fragments. Consequently, the Egerian andesitic bodies must have been exposed and strongly eroded during the Eggenburgian.

Age of Tertiary andesitic volcanism of the area of Hrvatsko Zagorje was determined by geological and radiometric methods. Field relations and stratigraphic data suggest that the volcanic activity had commenced by the end of the Egerian but the strongest volcanism took place on the boundary between the Egerian and Eggenburgian.

Radiometric measurement, carried out on three whole-rock samples of andesites, gave the following K–Ar ages; 22.8±7 Ma on an andesite from the Lepoglava Quarry, 19.7±6 Ma on an andesite from the area of Lukovčak and 8.8±3 Ma on an andesite from the area of Rogatec (Pamić et al. 1993). The two older ages obtained from the fresh andesites, which are concordant with the geological age, are considered as reliable age of the rocks; the altered sample with a younger age (8.8 Ma) appears to reflect some Ar loss caused by alteration.

There are data which indicate that the volcanic activity went on during younger Tertiary. According to Šimunić (1992), a weak but explosive volcanic activity, which gave mostly small masses of pyroclastic rocks, was rejuvenated during the Late Eggenburgien, Ottnangian, Karpathian and Badenian. Products of these younger Tertiary phases are not included in this paper.

Petrology

Egerian–Eggenburgian volcanic activity of the area of Hrvatsko Zagorje gave andesites with subordinate basaltic andesites and pyroclastic rocks of andesitic composition.

Petrography and Mineralogy

Andesites

Andesites have fairly uniform structural and textural features and mineral composition.

Structure and texture. Most of andesites are hypocrystalline–porphyritic in texture with a variable proportion of phenocrysts ranging commonly between 1/4 and 1/3 of the whole rock mass. Only in some places, as for example, in the area of Strmec Humski, the phenocrysts predominate over the groundmass. The size of phenocrysts is commonly 2–3 mm but some of them, as for example, the andesites from the Lepoglava Quarry, also contain a finer-grained phenocrysts 0.2 to 0.5 mm in size. The hypocrystalline groundmass includes glass and plagioclase and K–feldspar microlites, commonly mixed with minute grains of augite, hypersthene, and relict biotite.

Basaltic andesites and some andesites from the area of Strmec Humski are porphyritic–ophitic in texture; rare plagioclase phenocrysts, up to 3 mm in the size, are embedded in the fine-grained (0.05–0.2 mm) to medium-grained (0.2–0.5 mm) groundmass.

Most of the andesites are massive in structure. Andesites from the area of Bistrica and Drenovec contain a small quantity of amygdules filled by quartz, chalcedony, chlorite and cloudy clinozoisite. Only the andesites from the abandoned Trlično Quarry are parallel in structure as shown mostly in the lineated prismatic plagioclase phenocrysts. However, some structural varieties are characterized by alternating thin bands made up either of glass or variable proportions of glass and lineated feldspar phenocrysts. In some places these banded varieties include planparallely inserted veinlets filled by a very fine-grained zeolite (?).

Some andesites from the area of Lepoglava and Strmec Humski contain small millimetre xenoliths of darkgreyish fine-grained basaltic rocks.

The mineral assemblage of andesites of the area of Hrvatsko Zagorje includes the following major minerals: predominant plagioclase with subordinate K-feldspar, augite, hypersthene, and biotite. The plagioclase is mostly fresh and in some places contain minute inclusions of clinozoisite and devitrificated glass which is commonly inserted along the cleavage fissures. Femic minerals: hypersthene and more common augite are mainly fresh whereas subordinate biotite is mostly transformed into a hydromica. Accessory constituents are apatite and metallic mineral(s).

A fresh basaltic andesite from the Lepoglava Quarry (an. 2, Table 2) was selected for microprobe analysis and the data obtained are presented in Table 1.

Table 1
Representative microprobe chemical compositions of major rock-forming minerals

	Pl_{pz}		Pl_{pu}-1	Pl_{pu}-2	Pl_g-1	Pl_g-2	Pl_x-1	Pl_x-2
SiO_2	52.28	54.27	53.19	52.61	47.53	52.38	48.04	48.97
Al_2O_3	29.45	28.47	29.18	29.04	31.61	29.14	32.71	31.84
FeO	0.49	0.41	0.43	0.37	0.53	0.60	0.48	0.56
MgO	0.05	0.04	0.08	0.07	0.06	0.07	0.06	0.06
CaO	12.54	11.37	11.29	12.24	15.54	12.50	16.27	15.32
Na_2O	4.11	4.90	4.76	4.33	2.42	4.06	2.19	2.40
K_2O	0.39	0.32	0.37	0.26	0.25	0.63	0.09	0.44
	99.29	99.78	99.30	98.9	97.94	99.39	99.84	99.59
Number of ions on the basis of 8 oxygens								
Si	2.391	2.460	2.424	2.411	2.226	2.397	2.207	2.252
Al	1.588	1.521	1.567	1.569	1.745	1.572	1.771	1.726
Fe	0.017	0.014	0.015	0.013	0.019	0.021	0.017	0.020
Mg	0.004	0.002	0.006	0.005	0.004	0.004	0.004	0.004
Ca	0.614	0.552	0.551	0.601	0.780	0.613	0.801	0.755
Na	0.364	0.430	0.421	0.385	0.220	0.360	0.195	0.214
K	0.023	0.019	0.021	0.015	0.015	0.037	0.005	0.026
An	61.4	55.1	55.5	60.0	76.8	60.7	80.0	75.9
Ab	36.4	43.0	42.3	38.4	21.7	35.7	19.5	21.6
Or	2.3	1.9	2.2	1.5	1.5	3.6	0.5	2.6

	Hy_p	Hy_g	Hy_x	Au_p	Au_g	Au_x
SiO_2	51.64	51.78	52.51	51.50	50.66	51.19
TiO_2	0.14	0.14	0.28	0.29	0.24	0.29
Al_2O_3	0.64	0.74	1.34	1.43	1.27	1.53
Cr_2O_3	0.00	0.00	0.00	0.01	0.01	0.02
FeO	24.67	22.34	19.64	10.70	10.07	10.36
MnO	0.65	0.59	0.50	0.33	0.59	0.31
MgO	20.62	22.08	24.13	13.85	14.24	13.90
CaO	1.18	1.22	1.35	20.79	20.32	20.82
Na_2O	0.35	0.07	0.04	0.29	0.39	0.27
K_2O	nd	nd	nd	nd	nd	nd
	99.89	98.96	99.79	99.19	97.79	98.69
Number of ions on the basis of 6 oxygens						
Si	1.960	1.960	1.943	1.941	1.949	1.947
Ti	0.005	0.005	0.008	0.008	0.007	0.008
AlIV	0.040	0.040	0.057	0.049	0.051	0.053
AlIV	0.009	0.007	0.002	0.015	0.006	0.016
Cr	0.000	0.000	0.000	0.000	0.000	0.001
Fe	0.775	0.707	0.608	0.339	0.324	0.330
Mn	0.021	0.019	0.016	0.011	0.019	0.010
Mg	1.176	1.246	1.331	0.782	0.816	0.788
Ca	0.048	0.049	0.054	0.838	0.837	0.848
Na	0.006	0.005	0.003	0.022	0.029	0.020
K	0.000	0.000	0.000	0.000	0.000	0.000
Mg	0.603	0.638	0.687	0.698	0.638	0.705

Table 2
Major element contents, CIPW norms and petrochemical indices

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	51.0	52.1	55.20	54.48	54.3	54.7	58.4	59.4	60.20	58.44	55.62
TiO_2	0.65	0.75	0.69	0.76	0.75	0.68	0.66	0.69	0.75	0.68	0.71
Al_2O_3	20.37	21.99	18.74	18.93	16.8	17.32	16.41	16.96	17.01	17.20	18.08
Fe_2O_3	2.10	3.26	1.89	3.49	5.50	3.21	4.62	2.58	4.33	3.50	3.45
Feo	4.58	3.30	4.92	3.50	1.51	3.71	1.76	3.10	1.59	3.81	3.18
MnO	0.13	0.07	0.11	0.09	0.11	0.12	0.14	0.12	0.06	nd	0.11
MgO	4.65	3.38	4.03	2.01	4.89	4.47	3.46	1.24	1.24	3.40	3.43
CaO	11.71	10.05	7.57	6.17	4.27	7.90	5.97	5.48	5.16	7.00	7.08
Na_2O	2.58	2.58	4.18	5.49	3.10	2.51	2.43	3.85	3.07	2.80	3.18
K_2O	0.31	0.83	0.86	0.67	2.11	0.56	1.19	1.20	2.67	0.70	1.39
P_2O_5	0.10	0.12	0.09	0.13	0.15	0.14	0.14	0.18	0.20	0.19	0.16
H_2O	2.01	2.62	1.25	3.36	6.41	4.92	4.44	4.59	3.68	2.52	3.62
	110.12	100.06	99.53	99.08	100.00	100.22	99.62	99.75	99.96	100.24	100.01
FeO^x/ MgO	1.4	1.8	1.7	3.0	1.3	1.5	1.7	4.4	4.4	2.0	1.9
					CIPW norms						
qz	4.5	8.7	4.3	4.2	10.5	13.8	21.4	19.0	19.0	17.9	
c	-	-	-	-	2.2	-	0.6	-	0.2	-	
or	1.9	1.9	5.2	4.1	13.4	3.5	7.4	7.5	16.4	4.2	
ab	21.4	22.6	36.0	48.5	28.1	22.4	21.6	34.3	27.0	24.3	
an	44.4	46.3	30.3	26.1	27.1	36.1	30.1	26.8	25.3	33.1	
diwo	5.3	1.9	3.0	2.1	-	1.4	-	0.2	-	0.4	
dien	3.3	1.1	1.8	1.3	-	0.9	-	0.1	-	0.2	
difs	1.7	0.6	1.1	0.6	-	0.5	-	0.1	-	0.1	
hyen	8.6	7.6	8.5	3.9	13.1	10.9	9.0	3.1	3.2	8.5	
hyfs	4.3	4.3	5.5	1.9	5.6	5.8	4.9	3.7	3.5	6.2	
mt	3.1	3.4	2.8	5.3	3.5	3.3	3.3	3.3	3.4	3.2	
il	1.3	1.5	1.3	1.5	1.5	1.3	1.3	1.4	1.5	1.3	
ap	0.2	0.3	0.2	0.3	0.4	0.3	0.3	0.4	0.5	0.5	
norm.P1.	$An_{67.5}$	$An_{67.2}$	$An_{45.8}$	$An_{40.4}$	$An_{43.5}$	$An_{61.7}$	$An_{58.3}$	$An_{43.9}$	$An_{48.3}$	$An_{57.6}$	
					Petrochemical indices						
AI	11.1	10.7	17.1	10.9	40.5	18.2	32.9	23.8	46.5	20.0	
FI	19.6	22.3	40.0	50.0	55.0	28.4	37.7	48.0	52.7	33.6	
MI	59.0	65.6	62.8	77.7	57.8	60.4	63.9	81.9	82.2	67.9	
SI	33.5	27.0	25.7	13.6	29.5	31.6	26.6	10.6	9.9	24.5	
DI	27.8	33.1	45.4	56.9	52.0	39.6	50.4	60.7	62.5	46.4	
VSM	CA	Th	CA	Th	CA	CA	CA	Th	Th	CA	
VSIB	CA	CA	CA	CA	CA	CA	CA	CA	CA	CA	

The chemical composition of major minerals both of the andesite and basaltic inclusion was measured.

Plagioclase and hypersthene show inverse variations in the chemical composition as a result of reequilibration due to assimilation during the solidification of a primary andesitic melt. The most sodic are platy plagioclase phenocrysts which are either zoned (in the nucleus: $An_{55.1}$ and in the rim: $An_{61.4}$) or unzoned showing mainly the same interval of chemical variations (ans. Pl_{pz}, Pl_{pu}-1 and Pl_{pu}-2, Table 1). Minute groundmass plagioclases are more calcic ($An_{60.7-76.8}$) with the average content $An_{66.9}$ (ans. Pl_{g-1} and Pl_{g-2}, Table 1). However, the most calcic are plagioclases from relict basaltic xenoliths showing variations from $An_{75.9-80.0}$ and the average composition amounts to $An_{77.8}$.

In hypersthene, going from phenocrysts to groundmass and to basaltic xenoliths, the contents of Al_2O_3 (0.64–0.74–1.34%) and MgO (20.62–22.08–24.13%) increase and the contents of FeO^x (24.67–22.34–29.64%) and Na_2O (0.35–0.07–0.04%) decrease. However, such variations could not be noticed in more common augite (ans Hy_p, Hy_g, Hy_x, Au_p, Au_g and Au_x, Table 1). In the classification diagram (Poldervaart and Hess 1951), clinopyroxene from the Lepoglava Quarry falls in the upper parts of the augite field but close to the boundary with salite field. The average content of the augite can be correlated with the average augite composition from calc–alkalic basalts of recent island arcs in all major elements except Al_2O_3 and TiO_2 (Leterrier et al. 1982).

Varieties of andesites. Egerian—Eggenburgian volcanic rocks of the area of Hrvatsko Zagorje are represented mostly by hypocrystalline–porphyritic augite andesites with subordinate hypersthene–augite and biotite–augite andesites. All of them are massive in structure except the hypocrystalline–porphyritic augite andesites of the abandoned Trlično Quarry which are characterized by their fluidal structure. The volcanic rock from the area of Donje Jesenje is a porphyritic–ophitic augite basalt which is transitional to andesite (basaltic andesite).

Pyroclastic Rocks

Most of the pyroclastic rocks are represented by tuffs in the area of Donje Jesenje, Strmec Humski, Rogatec and Trlično and subordinate volcanic breccias and agglomerates in the area of Trlično. Smaller masses of pyroclastic rocks are also found in the area of Ivanec, Podevcevo and Varaždinske Toplice.

Symbols for Table 2: 1. porphyric-ophitic andesitic basalt, Donle Jesenje; 2. hypocrystalline-porphyritic hypersthene-augite basaltic andesite, Lepoglava Quarry; 3. hypocrystalline-porphyritic hypersthene basaltic andesite, Drenovac; 4. hypocrystalline-porphyritic biotite (?) basaltic andesite, Bistrica; 5. hypocrystalline-porphyritic augite basaltic andesite, Strimec Humski; 6. hypocrystalline-porphyritic biotite(?)-augite andesite, Rogatec; 7. porphyritic-ophitic augite andesite, Strmec Humski; 8, 9. fluidal hypocrystalline-porphyritic augite andesite, Trlicno Quarry; 10. cryptocrystalline-porphyritic hyperstene-augite andesite, Smrekoves (Hinterlechner-Ravnik and Plenicar 1967); 11. average andesite composition; AI – alkali; FI – felsic; MI – mafic; SI – solidification; DI – differentiation; VSM – Miyashiro (1974); VSIB – Irvine and Barager (1971)

Tuffs are psammitic in texture, medium-grained (0.1–1 mm) to coarse-grained (1–2 mm). The tuffs are represented by crystallovitrophyric and lithocrystallovitrophyric varieties.

Tuffs have an andesitic composition and they are made up mostly of glassy groundmass which is partly but unevenly recrystallized in a fine-flaky aggregate. Detrital fragments are represented mostly by plagioclase, commonly fresh or slightly calcitized, with subordinate quartz and hydromica pseudomorphs after an unpreserved primary femic mineral. Subordinate lithic fragments are represented by hypocrystalline–porphyritic andesite with variable proportions of phenocrysts.

Tuffaceous volcanic breccias and agglomerates are exposed in the area of Lepoglava and in the Trlično Quarry. These rocks are composed of millimetre to centimetre fragments mostly of the varieties of hypocrystalline–porphyritic and spherulitic andesites. In volcanic breccias, these fragments are either embedded in the matrix of crystallovitrophyric tuff or are cemented with calcite.

Geochemistry

The major element contents and CIPW norms of Egerian–Eggenburgian volcanic rocks from the area of Hrvatsko Zagorje are presented in Table 2, which demonstrates the variations which are characteristic for rocks of the andesite association. Based on SiO_2 content, the volcanic rocks vary from basic andesites to acid andesites, and based on their K_2O content, they are mostly low-K and medium-K andesites with an average K_2O content = 1.39; the average andesite composition corresponds to tholeiitic medium-K basic andesite (Gill 1981).

All the chemically analyzed rocks contain normative hypersthene. The chemical variations of andesites are presented on the AFM triangle (Fig. 2) which shows that most of the points fit quite well with the typical calc–alkaline trends of the Asama and Amagi volcanoes in Japan. However, on the FeO^x/MgO versus SiO_2 diagram, all points plot in tholeiite field (Miyashiro, 1974 – see Fig. 3).

The trace element contents are presented in Table 3. They show the variations which are characteristic of andesitic rocks in general. However, these variations fit only for some trace elements of the average tholeiitic medium-K basic andesite (Gill 1981 – see ans 11 and 12, Table 3) which is probably caused by the mentioned reequilibration during the crystallization of andesitic lava.

The range of Sr contents is typical of island arc andesites. In the Rb versus Sr diagram (Fig. 4), all the andesites from the area of Hrvatsko Zagorje fit with typical subduction related andesites. The ratios of Ba:La (12 to 20), La:Th (2.5 to 3) and La:Nb (1.6 to3.3) are also indicative for magmatic arc settings (Gill 1981).

Table 3
Trace element data

	1	2	3	4	5	6	7	8	9
Ba	130	180	610	300	340	510	470	363	850
Ce	17	25	54	42	34	72	66	44	24
Co	25	21	19	21	20	7	6	17	–
Cr	41	36	39	29	68	18	26	37	9
Cu	88	43	170	80	65	200	34	96	46
Ga	16	19	18	17	16	21	22	18	–
La	10	13	30	24	19	38	36	24	11
Li	18	16	37	25	42	13	22	25	–
Nb	5	8	10	10	8	13	11	9	4
Nd	8	14	28	21	17	34	33	22	–
Ni	21	6	9	29	12	12	2	10	6
Pb	22	14	17	17	11	26	22	18	5
Rb	17	23	45	13	25	42	81	34	25
Sc	31	25	26	24	27	27	26	27	16
Sr	340	350	260	420	400	400	280	350	810
Th	4	4	12	9	6	12	12	8	1.3
V	200	190	170	170	200	40	45	145	178
Y	16	19	29	25	21	39	37	27	17
Yb	2	2	4	3	2	5	4	3	1.7
Zn	59	68	120	82	80	180	90	97	74
Zr	70	91	168	140	91	198	186	135	81
K/Rb	151	299	159	427	700	111	122	193	886
Ba/La	13	14	20	12.5	18	13.5	13	15	22
La/Th	2.5	3	2.7	2.7	3.2	3.2	3.0	3.0	8.5
La/Nb	2	1.6	2.4	2.4	2.9	2.9	3.3	2.7	2.7

Symbols: 1. porphyric-ophitic andesitic basalt, Donle Jesenje; 2. hypocrystalline-porphyritic hypersthene-augite basaltic andesite, Lepoglava Quarry; 3. hypocrystalline-porphyritic hypersthene basaltic andesite, Drenovac; 4. hypocrystalline-porphyritic biotite (?) basaltic andesite, Bistrica; 5. hypocrystalline-porphyritic augite basaltic andesite, Strimec Humski; 6. hypocrystalline-porphyritic biotite(?)-augite andesite, Rogatec; 7. porphyritic-ophitic augite andesite, Strmec Humski; 8. fluidal hypocrystalline-porphyritic augite andesite, Trlicno Quarry, 9. average trace element contents of tholeiitic medium-K basic andesite (Gill, 1979)

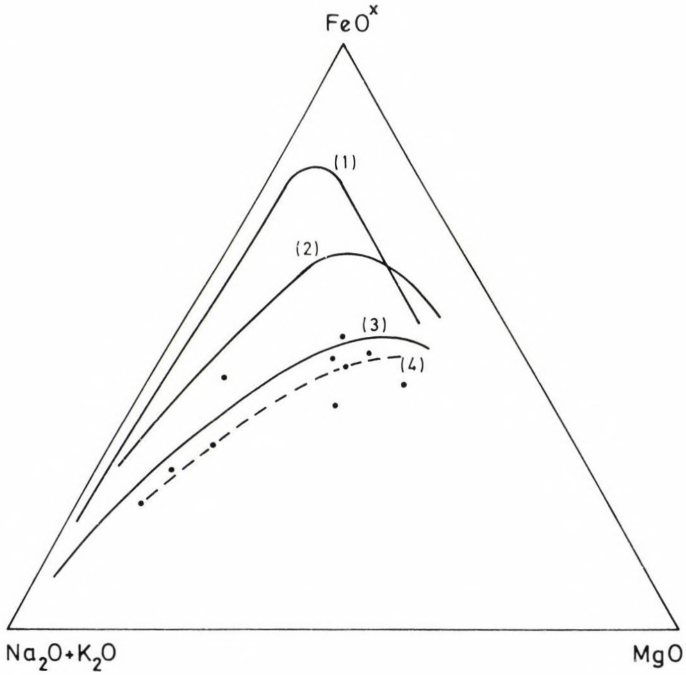

Fig. 2
The AFM diagram for andesites from the area of Hrvatsko Zagorje. 1. and 2. typical tholeiitic rocks series of Skaergaard and Izu–Hakone, 3. the average calc–alkali rocks series of Izu Hakone and Amagi Volcano (Miyashiro 1974), 4. andesites from the area of Hrvatsko Zagorje

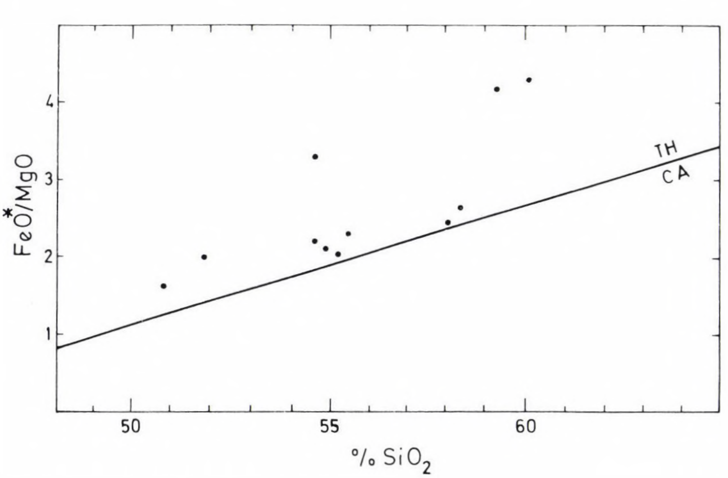

Fig. 3
The FeO^x/MgO versus SiO_2 diagram (Miyashiro 1974)

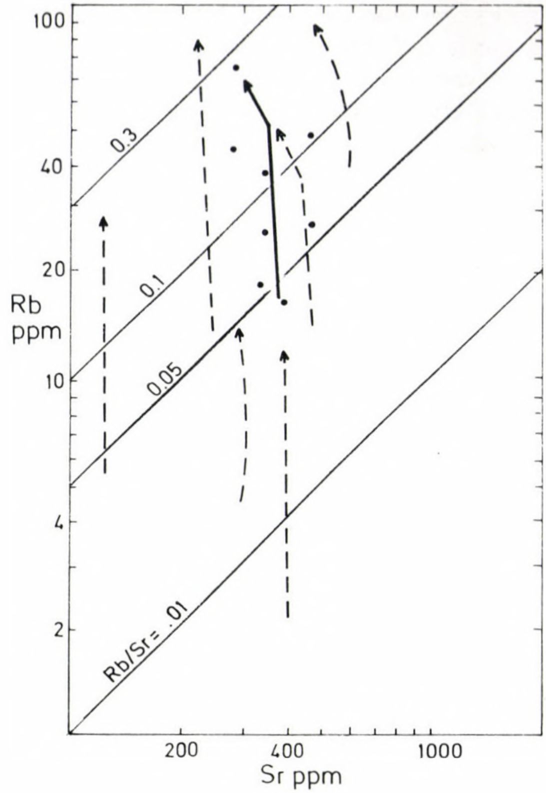

Fig. 4
Rb versus Sr diagram. Dashed lines demonstrate andesites from some recent subduction zone and the full line demonstrates andesites from the area of Hrvatsko Zagorje

Discussion

The largest quantitiy of Egerian–Eggenburgian andesites and pyroclastic rocks are located along the fault zone stretching in a west–east direction from Hum on the Sutla River to Varaždinske Toplice over a distance of about 75 km. The eastern parts of the large fault plung under Neogene sedimentary rocks of the Pannonian Basin of the southern parts of the Drava Depression. In the west, the fault grades into the Smrekovec fault in Slovenija (Mioč 1978) which represents the easternmost parts of the Periadriatic Lineament. Numerous papers (Boegl 1975 and others) have been published on the Periadriatic Lineament and various opinions are presented. Recently, the idea has been generally accepted that the Periadriatic Lineament separates the Southern Alps and Austroalpine crystalline complex and both of them are included into the northern marginal parts of the African plate and Apulian microplate, respectively (Dercourt et al. 1986 and others).

Large masses of Oligocene andesites and pyroclastic rocks, accompanied by the Oligocene "Socka" beds characteristically occur along the Smrekovec Line (Faninger 1966; Hinterlechner-Ravnik 1967; Mioč 1978). Later on, Aničić and Juriša (1985) ascribed Egerian age to most of these rocks. Petrologically, these rocks can be correlated with the andesitic rocks of Hrvatsko Zagorje. However, there is a discrepancy in their geological position.

In this correlation it should be mentioned that igneous and metamorphic rocks correlatable with metamorphic and tonalitic rocks from Mt. Pohorje in Slovenija were penetrated by some oil wells drilled in the surrounding Mura Depression (Pamić 1986). However, these rocks were not drilled east of the Donački fault.

It can be concluded that the area of the fault zone Hum on the Sutla River–Varaždinske Toplice can be considered as the eastern extension of the Periadriatic Lineament which is masked by Oligocene–Miocene sedimentary rocks of the Pannonian Basin. This large fault controlled the geometry of the southwestern parts of the Pannonian Basin during the initial phase of its generation.

The presented geological data indicate that the Egerian–Eggenburgian volcanism of the area of Hrvatsko Zagorje was genetically related to the initial phases of the generation of the Pannonian Basin. According to new geodynamic ideas, the Periadriatic–Vardar fault system had a function of the dextral strike–slip fault since the Oligocene (Laubscher 1971) and separated the Apulia from the Pannonian realm which was pushed towards the east (Royden et al. 1983; Pamić 1993 and others). In fact, such horizontal dextral faults in combination with left–sinistral horizontal faults controlled the E–W extension of the Pannonian Basin.

The presented geochemical and petrological data show that rocks of the andesite association of Hrvatsko Zagorje have distinct subduction features. This opinion is supported by Sr isotope composition measured on an andesite from the Trlično Quarry (an. 7, Table 3); the primary $^{87}Sr : ^{86}Sr$ ratio amounts 0.70743 which is characteristic for crustal andesites (Faure 1986).

This is not in accord with field relations and radiometric data which indicate that the andesitic volcanism of the area of Hrvatsko Zagorje took place after the Pyrenean phase by which must have ceased the main phase of compression and subduction processes. Andesite occurrences of the area of Hrvatsko Zagorje may be related to an orogenic phase which had taken place by the end of the Egerian, in fact, between the Savian and Styrian orogenic phases (Šimunić 1992). This discrepancy could be explained by an idea that some subduction blocks remained in the phase of extension and at a larger depth generated andesitic melt by partial melting. The mentioned comparatively higher initial $^{87}Sr : ^{86}Sr$ ratio indicates that these blocks must have been of crustal origin. The blocks were probably broken off from the overlying parts of the former subduction zone built up of continental crust.

Acknowledgements

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Mineralogy and geochemistry of the rare earth elements in the karstic nickel deposit of Lokris area, Greece

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The mineralogy and geochemistry of the rare earth elements (REE) was studied along vertical sections in two karstic nickel deposits: Marmeiko and Aghios Ioannis. Deposits display a downward enrichment of the REE culminating at the base, where, in the Marmeiko and Nisi deposits, authigenic minerals of the bastnaesite group were found.

Both light REE (LREE) and heavy REE (HREE) were "mobile" and concentrated at the alkaline barrier of the footwall limestone. In the studied deposits, the ratios Σ LREE/Σ HREE and La/Y decrease downwards in the section, showing enrichment of HREE relative to LREE. The same pattern was observed in the karstic bauxite deposits in Bosnia, Montenegro, Serbia and Greece. The same behavior of the REE in the formation of karstic bauxites and karstic nickel deposits is another proof of their genetic relationship.

Key words: rare earth elements, karsric environments, nickel deposit, authigenic mineraLS

Introduction

Karstic bauxites and karstic nickel deposits are connected by deposits of transitional character, which indicate the same genetic conditions for both end types (Maksimovic 1978). Therefore the study of karstic Ni deposits is valuable for a better understanding of karstic bauxite genesis in the same region. The best examples of karstic Ni deposits and their transitions to karstic bauxites occur in Continental Greece, in the Lokris area. The geochemistry and mineralogy of these deposits are better known thanks to a comprehensive study by Rosenberg (1984).

In this paper attention is paid to the mineralogy and distribution of the rare earth elements (REE) along vertical sections in some selected deposits. Deposits display a downward enrichment of the REE culminating at the base, where in

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some deposits authigenic REE minerals of the bastnaesite group have been formed. The same REE pattern is also characteristic for karstic bauxites formed in situ (Maksimovic 1976; Maksimovic and Roaldset 1976; Maksimovic and Pantó 1991).

Geology

The area of Lokris in Continental Greece is part of the Subpelagonian geotectonic unit of the Hellenides. This unit is characterized by outcrops of Triassic dolomites and limestones, deeply eroded and karstified Jurassic limestones, a chert-ophiolite sequence of Upper Jurassic–Lower Cretaceous age, and Cenomanian transgressive limestone and Tertiary flysch. Ophiolitic rocks in the area are represented mainly by serpentinites and weathered serpentinites. Large deposits of Ni–Fe ores are exposed in the Lokris area. Karstic Ni deposits in the area predominate by tonnage over those deposited on serpentinites. Typical examples are the Aghios Ioannis and Marmeiko deposits, which along with the Euboea deposits are exploited for the extraction of nickel. Footwall limestones are of Middle to Upper Jurassic age. The hanging wall, transgressively overlying the Ni–Fe ores, is made up of Cenomanian limestones. From the stratigraphic point of view, these Ni–Fe ores are equivalent to the B_3 bauxitic horizon within the Parnassus–Ghiona geotectonic unit (Albandakis 1974).

The development of the Lower Cretaceous weathering crust and redeposition of the weathering material into karstic depressions was essential for the formation of karstic bauxites and karstic Ni deposits in Central Greece. The Ni content in these deposits increases with the contribution of the weathered material of ultramafic rocks. The REE content, on the other hand, depends on the contribution of more acidic rock types, as well as of various sedimentary and metamorphic rocks. Within the karstic Ni deposits of the Lokris area, a continuous lateral transition from Ni–Fe ores towards bauxitic material can be observed (Rosenberg 1984).

Materials and methods

Samples of the Ni–Fe ores and ferruginous clays were collected along vertical sections in the Marmeiko and Aghios Ioannis karstic deposits. A sample of the black crust in which Rosenberg (1984) discovered bastnaesite was kindly sent by the author for the analyses of the REE in this mineral. Mineralogical composition was studied by means of X-ray diffraction and chemical analyses. Authigenic bastnaesite was analyzed with a JEOL Superprobe JXA–733 in the Laboratory for Geochemical Research of the Hungarian Academy of Sciences, Budapest.

REE were determined by Inductively Coupled Argon Plasma Atomic Emission Spectroscopy (ISP–AES), following a pre-separation of the REE as a group by a two-stage ion exchange column technique. In this method background and spectral line interference corrections, which arise principally from matrix elements, are minimized by chemical separation of the REE prior to measurements. This approach has been discussed by Walsh et al. (1981) and Roelandts (1987), but in this work a superior cation-exchange chromatographic procedure, developed in Oxford by J.W. Arden and M.J. Whitehous, was used. The complete technique has been evaluated by repetitive analyses of the USGS standard rocks BCR-1, GSP-1, MAG-1 and SGR-1. Precisions of less than 5% for each element were routinely obtained and excellent agreement was demonstrated with the consensus values published by Gladney and Roelandts (1988). REE determinations were carried out in the Department of Nuclear Physics of Oxford University, England.

The analyses of the REE from Aghios Ioannis karstic nickel deposit have been carried out by neutron activation analyses (NAA) in the Laboratory of Bondar and Clegg, Montreal, Canada.

Mineralogy

According to Rosenberg (1984), the major constituents in the investigated karstic deposits of Marmeiko, Aghios Ioannis and Nisi are hematite, geothite, kaolinite, chromspinel and chlorite. It should be noted that in the upper part of Marmeiko deposit, this author found boehmite as a major constituent, showing that above the Ni–Fe ores a bauxitic material occurs in this deposit.

We studied in detail one section from Marmeiko deposit. The mineralogical composition of this deposit, derived from all available data, is presented in Table 1 and Fig. 1. Boehmite was not detected in the upper part of the deposit.

Table 1
Mineralogical composition of the Ni–Fe ores and ferruginous clays from a section through Marmeiko deposit, Greece (in %)

Minerals	M1	M2	M3	M4	M5	M6	M7	M8	M9
Kaolinite	30.2	33.0	40.4	43.5	48.0	58.3	58.0	54.5	59.0
Smectite	20.0	22.6	5.0	tr	tr	-	-	-	-
Chlorite	5.0	17.2	13.1	11.4	5.0	tr	-	5.0	-
Hematite	32.6	8.0	38.0	42.0	44.0	39.5	40.0	38.5	39.2
Goethite	-	14.0	-	-	-	-	-	-	-
Chromite	2.1	2.0	1.9	1.8	1.8	0.7	0.9	0.7	0.6
TiO2-minerals	1.0	1.6	1.3	1.3	1.3	1.5	1.2	1.3	·1.2
Calcite	5.7	-	-	-	-	-	-	-	-
Quartz	3.4	-	-	-	-	-	-	-	-
Bastnaesite	tr	1.6	0.3	-	-	-	-	-	-
	100.0	100.0	100.0	100.0	100.1	100.0	100.1	100.0	100.0

Symbols: M1–M5 = samples of Ni–Fe ores; M6–M9 = samples of ferruginous clays in the upper part of the section; for position of samples see Fig. 1

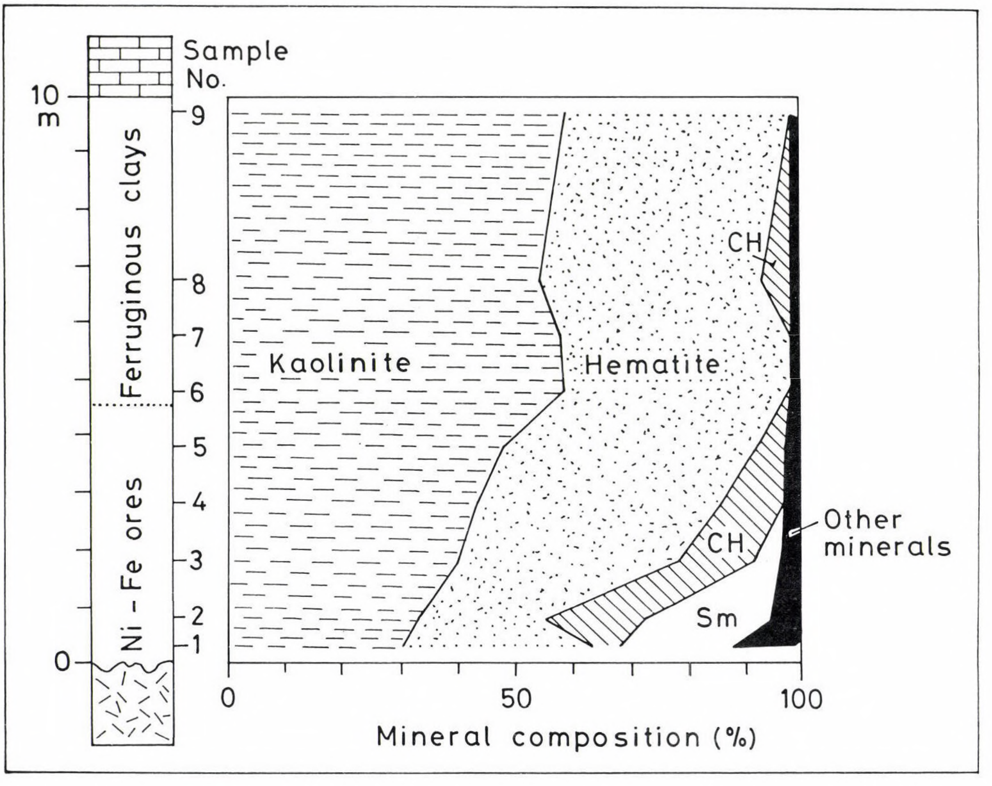

Fig. 1
Mineralogical composition in a section of the Marmeiko karstic nickel deposit. CH – chlorite; Sm – smectite; Other minerals – chromite, TiO_2 – minerals, calcite, quartz, bastnaesite

Kaolinite and hematite are the major constituents, and therefore, in this section, the material above the Ni–Fe ores was presented as ferruginous clays. In the lower part of the section, in the so-called Ni–Fe ores, the content of kaolinite and hematite decreases, but chlorite and smectite increase towards the footwall limestone.

The study of two sections in the Marmeiko deposit indicate lateral changes in the mineralogical composition in the same deposit, which is rather characteristic for karstic deposits. On the other hand, the work of Rosenberg and our study show that, in the vertical section of the Marmeiko deposit, an abrupt change in the mineralogical and chemical composition also exists (Table 1, Fig. 2).

Rosenberg (1984) discovered an authigenic REE mineral, bastnaesite, in the black crust, strongly enriched in Ni, Co, Mn, on the footwall limestone in the Nisi deposit. The analysis of this sample by electron microprobe, which F. Rosenberg kindly sent to Gy. Pantó, has shown a new variety of

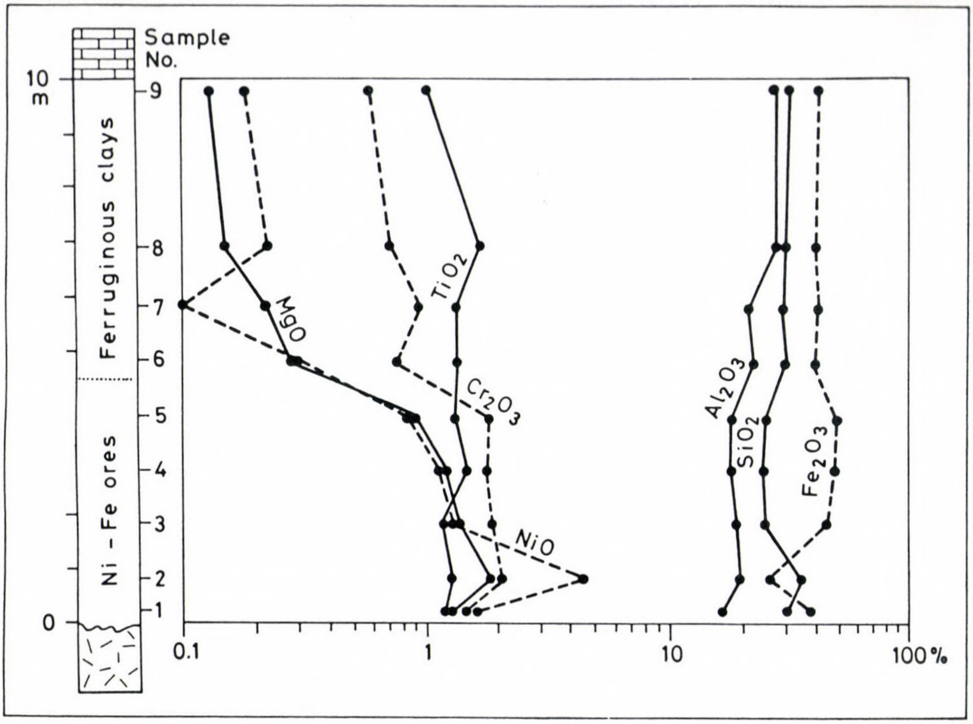

Fig. 2
Chemical variations in a section of the Marmeiko karstic nickel deposit

hydroxylbastnaesite -(Nd), practically without fluorine, with the following structural formula:

$(Nd_{.42}\ La_{.22}\ Pr_{.10}\ Sm_{.06}\ Eu_{.02}\ Gd_{.02}\ Y_{.01}\ Ca_{.17})_{1.02}\ \{(OH)_{.87}\ F_{.02}\}_{.89}\ (CO_3)_{1.03}$

In the lower portion of the Marmeiko deposit, bastnaesite was discovered by X-ray diffraction (Fig. 3). After treatment of this sample with hot 3% HCl, the characteristic reflection of this mineral at 2.871 Å disappeared. Bastnaesite was detected in two more samples (Table 1), indicating an enrichment of the REE in the lowermost part of this deposit.

Geochemistry

The first indication of the enrichment of Y from top to bottom in the Marmeiko and Neon Kokkinon karstic Ni deposits (Maksimovic 1978) was proven through the work of Rosenberg (1984), who analyzed by X-ray fluorescence La, Nd and Y in a great number of samples from karstic Ni deposits in the Lokris area. He found a high enrichment of these elements in the lower part of the Marmeiko deposit. In the Aghios Ioannis, however, concentration of these elements was much lower and a slight enrichment occurred towards the footwall limestone.

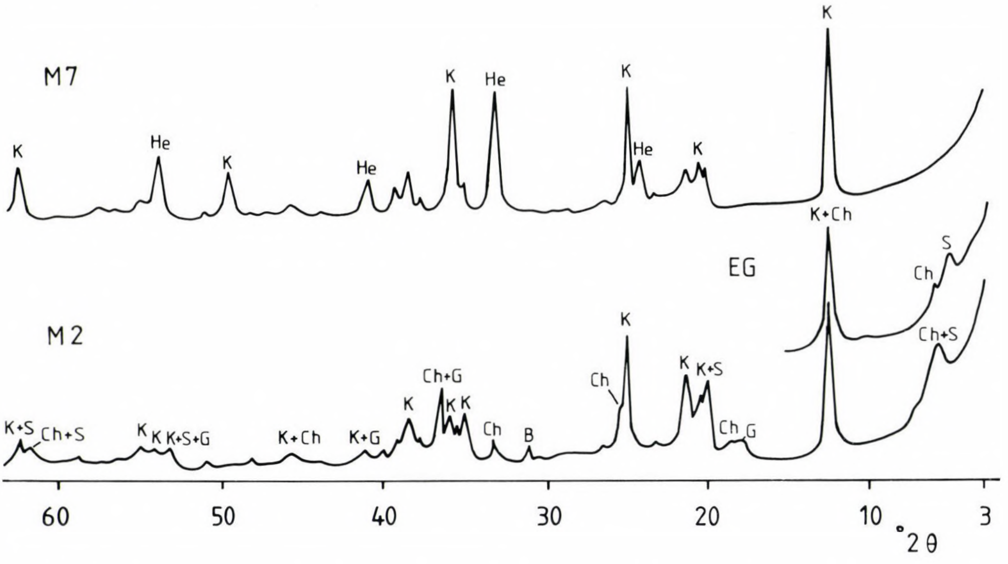

Fig. 3
X-ray diffractograms of a Ni–Fe ore (M2) with bastnaesite, and a ferruginous clay (M7) from Marmeiko deposit. K = kaolinite, S = smectite, Ch = chlorite, He = hematite, G = goethite, B = bastnaesite, EG = treated with ethylene glycol

Table 2
Concentrations of the REE in the Ni–Fe ores and ferruginous clays from a section through Marmeiko nickel deposit, Greece (in ppm)

Element	M1	M2	M3	M4	M5	M6	M7	M8	M9
La	460	573	377	328	294	140	89	90	92
Ce	156	246	221	230	169	164	135	120	122
Pr	60	79	50	44	33	16	11	11	12
ND	243	327	203	175	123	60	41	41	43
Sm	43	62	39	34	27	11	8	8	8
Eu	10	15	10	8	7	2	2	2	2
Gd	55	76	49	42	38	11	9	9	9
Dy	59	75	53	46	44	11	9	10	10
Ho	13	17	12	10	10	3	2	2	2
Er	37	46	33	29	28	7	6	6	6
Yb	32	40	30	26	26	7	5	6	5
Lu	5	6	5	4	4	1	1	1	1
Y	745	890	575	468	485	84	67	77	63

Symbols: M1–M5 = samples of Ni–Fe ores; M6–M9 = samples of ferruginous clays in the upper part of the section; for position of samples see Fig. 1

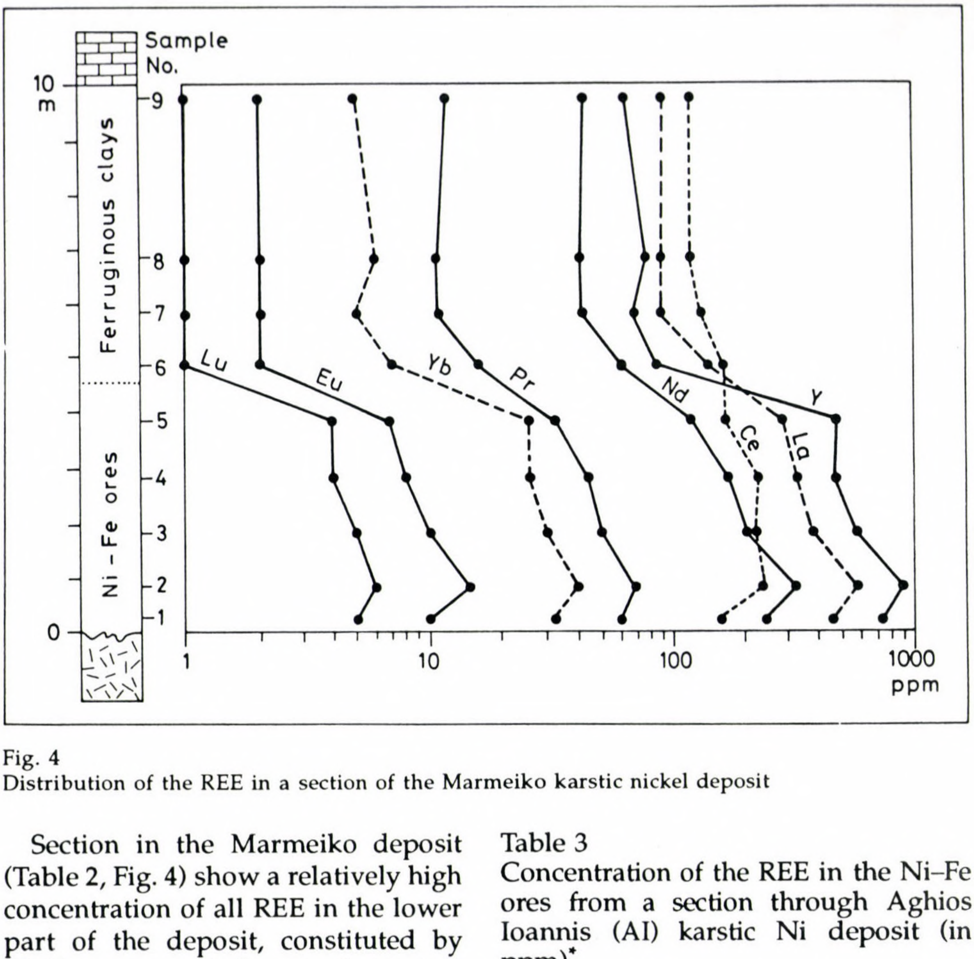

Fig. 4
Distribution of the REE in a section of the Marmeiko karstic nickel deposit

Section in the Marmeiko deposit (Table 2, Fig. 4) show a relatively high concentration of all REE in the lower part of the deposit, constituted by Ni–Fe ores, especially near the contact with the footwall limestone. Thus, the abrupt change in the mineralogy of the deposit is also marked by a sudden increase of the REE downwards. Relatively high concentration of the REE near the bottom of the deposit resulted in the formation of bastnaesite, detected by X-ray diffraction.

Ni–Fe ores from the Aghios Ioannis deposit are much poorer in REE than those from Marmeiko deposit (Table 3). It indicates that source rocks for

Table 3
Concentration of the REE in the Ni–Fe ores from a section through Aghios Ioannis (AI) karstic Ni deposit (in ppm)*

	AI 1	AI 2	AI 3	AI 4	AI 5
La	48	44	8	7	5
Ce	48	65	19	17	7
Nd	28	27	<10	<10	<10
Sm	7.6	7	1	1.1	0.6
Eu	2.4	1.9	<0.5	<0.5	<0.5
Tb	2	1	1	<1	<1
Yb	4	4	<1	<1	<1
Lu	0.8	0.6	<0.2	<0.2	<0.2
Y	39	28	3	1	<1

* Lanthanides determined by NAA, Yttrium by X-ray fluorescence technique; for position of samples see Fig. 5

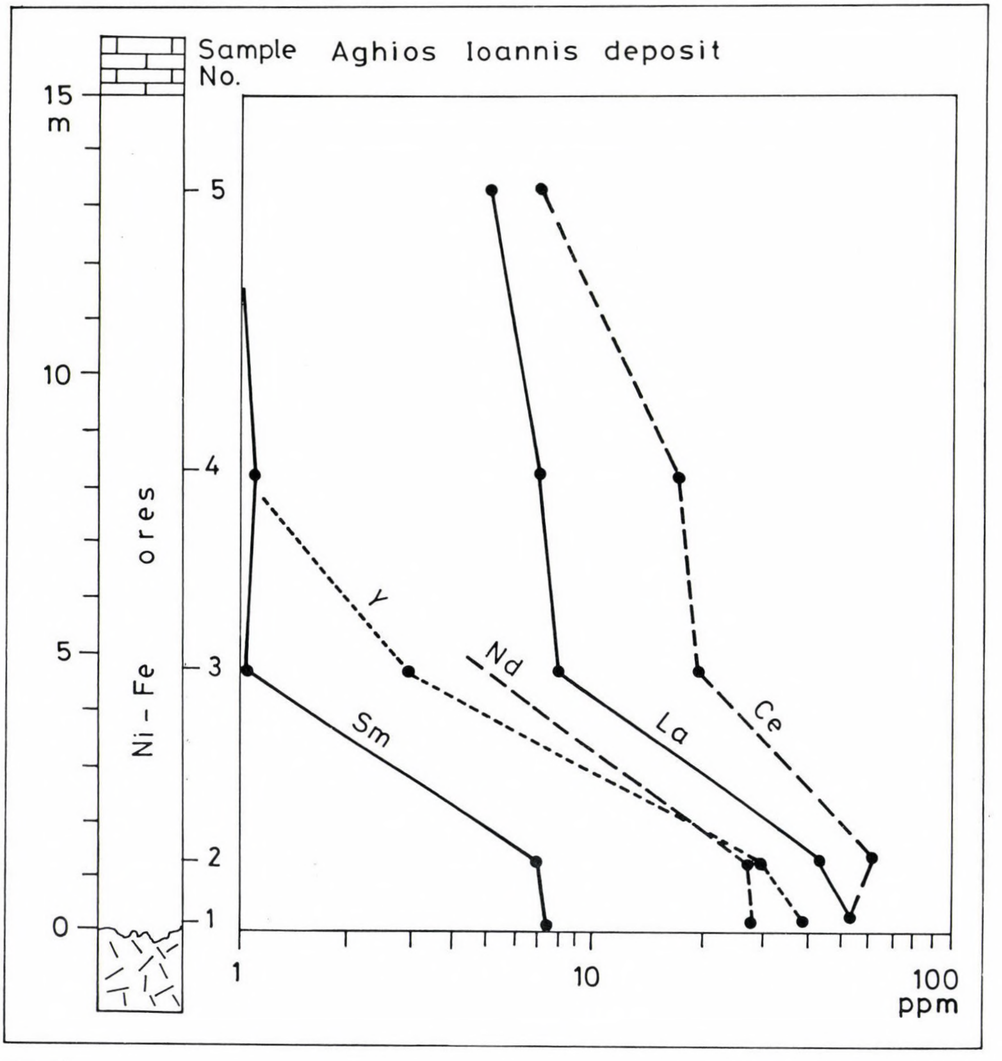

Fig. 5
Distribution of the REE in a section of the Aghios Ioannis karstic Ni deposit, Greece

the Aghios Ioannis deposit were mainly ultramafic rocks, highly depleted in REE. Distribution of the REE along a vertical section (Table 3, Fig. 5) shows a relatively high en- richment of all analyzed REE in the lowermost part of the deposit.

Figure 6 shows the chondrite- normalized REE distribution patterns of four samples from the Marmeiko deposit, two from the Ni–Fe ores (M2, M5) and two samples from the ferruginous clays (M6, M9). It can be seen that these samples differ not only in absolute REE content, but also in the relative distribution of the REE within them. The negative anomaly of Ce is pronounced

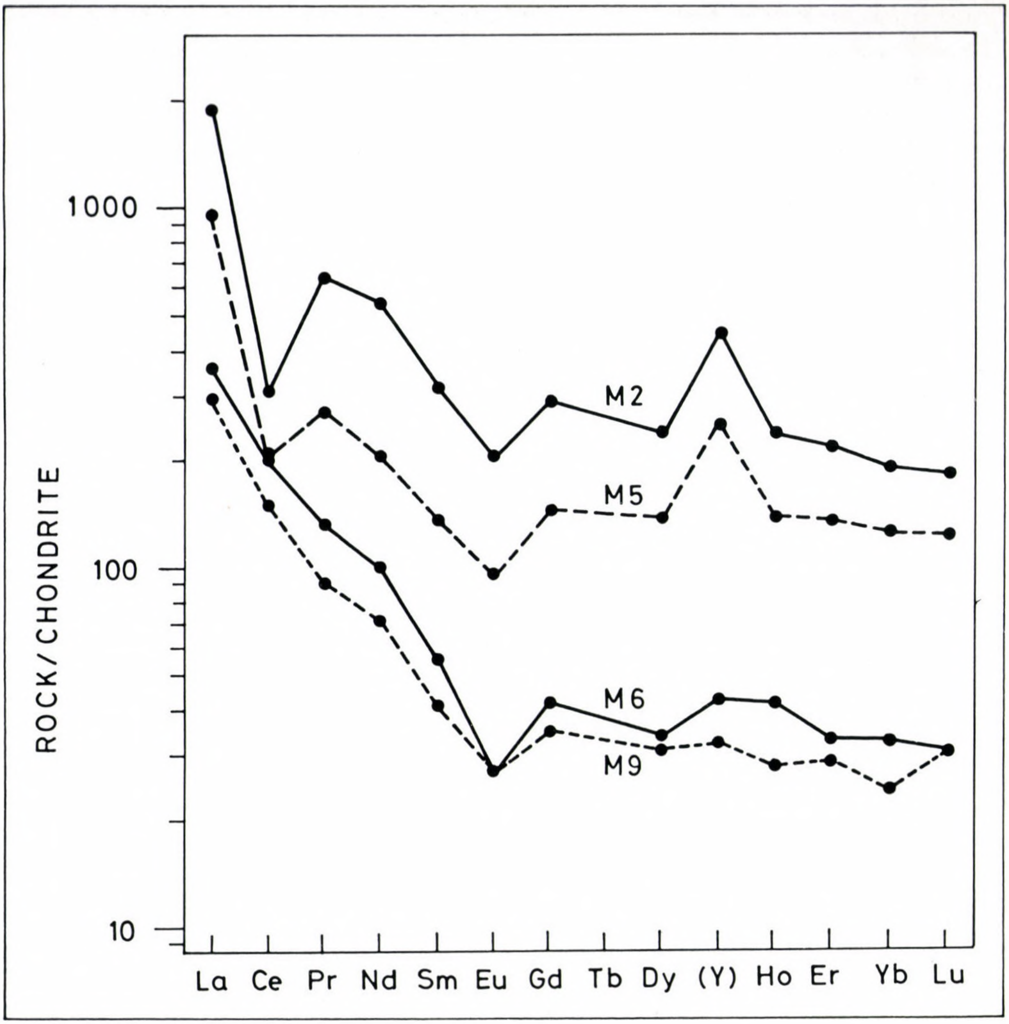

Fig. 6
Chondrite-normalized REE patterns for Ni–Fe ores (M2, M5) and ferruginous clays (M6, M9) from a section in the Marmeiko nickel deposit (Chondrite abundances from Nakamura, 1974)

in two bottom samples. This element behaves differently from the rest of the REE due to the possibility of being oxidized from Ce^{3+} into Ce^{4+} in a relatively strong oxidizing environment. In the Marmeiko deposit Ce does not show a preferential enrichment "per descensum" as other REE do. However, in the Agkhios Ioannis deposit Ce enrichment is evident. A positive Pr, Nd and Y anomaly in two bottom samples of the Marmeiko deposit is the result of high enrichment of these elements during the formation of the deposit, as in the case of some karstic bauxite deposits (Maksimovic and Pantó 1991). On the other hand, a negative Eu anomaly is characteristic for all samples and should be a tracer in searching for source rocks.

Both light REE (LREE) and heavy REE (HREE), show the same concentration trend (Table 4). There is, however, a general decrease of the ratios Σ LREE/Σ HREE and La/Y in samples with depth. In fact, a sudden decrease in these ratios occurs on the boundary between ferruginous clays in the upper part and Ni–Fe ores in the lower portion of the deposit. On the other hand, in the hydroxyl- bastnaesite–(Nd) from the neigh- bouring Nisi deposit, an extremely high enrichment of the LREE relative to HREE was found, as in all karstic bauxite deposits with authigenic REE minerals (Maksimovic and Pantó 1991). In the case of the Marmeiko deposit, the La/Y ratio, in the sense described by Zhuk–Pochekutov et al. (1986), seems not to be valid for the pH-conditions of formation of this deposit, but, rather reveals a basic barrier on the footwall limestone, where authigenic REE minerals have been deposited.

Table 4
REE distribution along a vertical section in the Marmeiko karstic nickel deposit, Greece, including hydroxylbastnaesite–(Nd) from the neighboring Nisi deposit

Samples	[a] Σ REE (ppm)	[b] Σ LREE (ppm)	[c] Σ HREE (ppm)	Σ LREE/Σ HREE	La/Y
M9 Ferruginous clays	375	279	96	2.9	1.5
M8 Ferruginous clays	363	272	111	2.4	1.2
M7 Ferruginous clays	385	266	99	2.7	1.3
M6 Ferruginous clays	517	393	124	3.2	1.7
M5 Ni–Fe ores	1288	653	635	1.0	0.6
M4 Ni–Fe ores	1444	819	625	1.3	0.7
M3 Ni–Fe ores	1657	900	757	1.2	0.65
M2 Ni–Fe ores	2432	1302	1150	1.1	0.64
M1 Ni–Fe ores	1918	972	946	1.0	0.62
Hydroxylbastnesite-(Nd), Nisi deposit	597.541	571.978	25.563	22.3	31.2

Symbols: [a] Σ REE – La–Lu, Y; [b] Σ LREE – La–EU; [c] Σ HREE – Gd–Lu, Y

In the Aghios Ioannis deposit, both LREE and HREE show the same concentration trend as in the Marmeiko deposit (Table 5).

However, in this deposit, the decrease of ratios Σ LREE/Σ HREE and La/Y with depth is much more pronounced.

Table 5
REE distribution along a vertical section in the Aghios Ioannis karstic nickel deposit, Greece

Samples	$^a \Sigma$ REE (ppm)	$^b \Sigma$ LREE (ppm)	$^c \Sigma$HREE (ppm)	— LREE/Σ HREE	La/Y
AI 5 Ni–Fe ores	12.6	12.6	1	12.6	5.0
AI 4 Ni–Fe ores	26.1	25.1	1	25.1	7.0
AI 3 Ni–Fe ores	32	28	4	7.0	2.7
AI 2 Ni–Fe ores	178.5	144.9	33.6	4.3	1.5
AI 1 Ni–Fe ores	179.8	134	45.8	2.9	1.2

Symbols: $^a \Sigma$ REE – La–Lu, Y; $^b \Sigma$ LREE – La–Eu; $^c \Sigma$ HREE – Gd–Lu, Y

Conclusions

1) The behavior of the REE during the formation of karstic Ni deposits is the same as in the case of karstic bauxites formed in situ. REE were readily removed "per descensum" by percolating water and concentrated at the base of the nickel deposits.

2) Both LREE and HREE were "mobile" during the formations of nickel deposits and were concentrated at the alkaline barrier of the carbonate footwall. In two studied Ni deposits, the Σ LREE/Σ HREE and La/Y ratios decrease downwards, showing an enrichment of the HREE relative to LREE, as in the majority of karstic bauxite deposits.

3) A strong fractionation of the REE took place in the formation of authigenic hydrohylbastnaesite–(Nd), which exhibits a very high enrichment of LREE relative to HREE. This fact has been observed in the authigenic REE minerals in karstic bauxites.

4) The concentration of the REE in karstic Ni deposits depends on their content in the source rocks, as in the case of karstic bauxites. Two studied Ni deposits, Marmeiko and Aghios Ioannis, situated close to each other, exhibit quite different REE contents, due to different composition of parent rocks.

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Metamorphic petrology of the Bulfat Area (North-east Iraqi Zagros Thrust Zone)

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The Bulfat post-collision igneous complex, which is largely basic, partly assimilated and thermally metamorphosed the Quandil regional metamorphic rocks. The parental rocks of the Quandil metamorphic group were pelitic, calcareous sediments and basic volcanites. These rocks were regionally metamorphosed, mostly under the conditions of greenschist facies. Due to the intrusion of an igneous complex these rocks were later also thermally metamorphosed. The mineral assemblage which formed during the thermal metamorphism depends on the composition of the parental rocks and PT conditions. The exact termination of the contact effect can be suspected when the first spots appear in phyllite, about 2.5 km away from the igneous body; between 2.5–1.6 km, the rocks recrystallized under the conditions of albite-epidote hornfels (1 Kbar up to 400 °C), between 1.6–0.15 km hornblende-hornfels (1 Kbar 400–520 °C) and between 0.15-0.0 km the pyroxene hornfels facies (1 Kbar 560–580 °C). The rock pedants (xenoliths) assimilated or recrystallized under the conditions of the high-temperature pyroxene hornfels facies (1 Kbar 650–850 °C).

Key words: Iraqi Zagros, thermal metamorphism, forsterite marble, predazzite, diopside-wollastonite hornfels, pelitic hornfels

Introduction

The Bulfat metamorphic-magmatic complex is situated in the Iraqi Zagros thrust zone, as a north-east member of igneous-metamorphic complexes located at the boundary between the Arabian and Iranian plates. The southern complexes (Mawat, Penjwin; Fig. 1), which are Cretaceous ophiolites without thermal contact aureoles, represent mantle and oceanic crust (Masek and Etabi 1973; Buda and Hashimi 1977). The Bulfat igneous complex has a large thermal contact zone and is believed to be a post-collision intrusion, emplaced in the continental crust after the Arabian and Iranian plates collided during late Cretaceous or early Paleogene (Buda et al. 1978; Jassim et al. 1982a; Jassim et al. 1982b). A previous paper dealt only with the petrology of the intrusive complex (Buda 1993), and the present one describes the xenoliths and contact metamorphic aureole of the intrusion.

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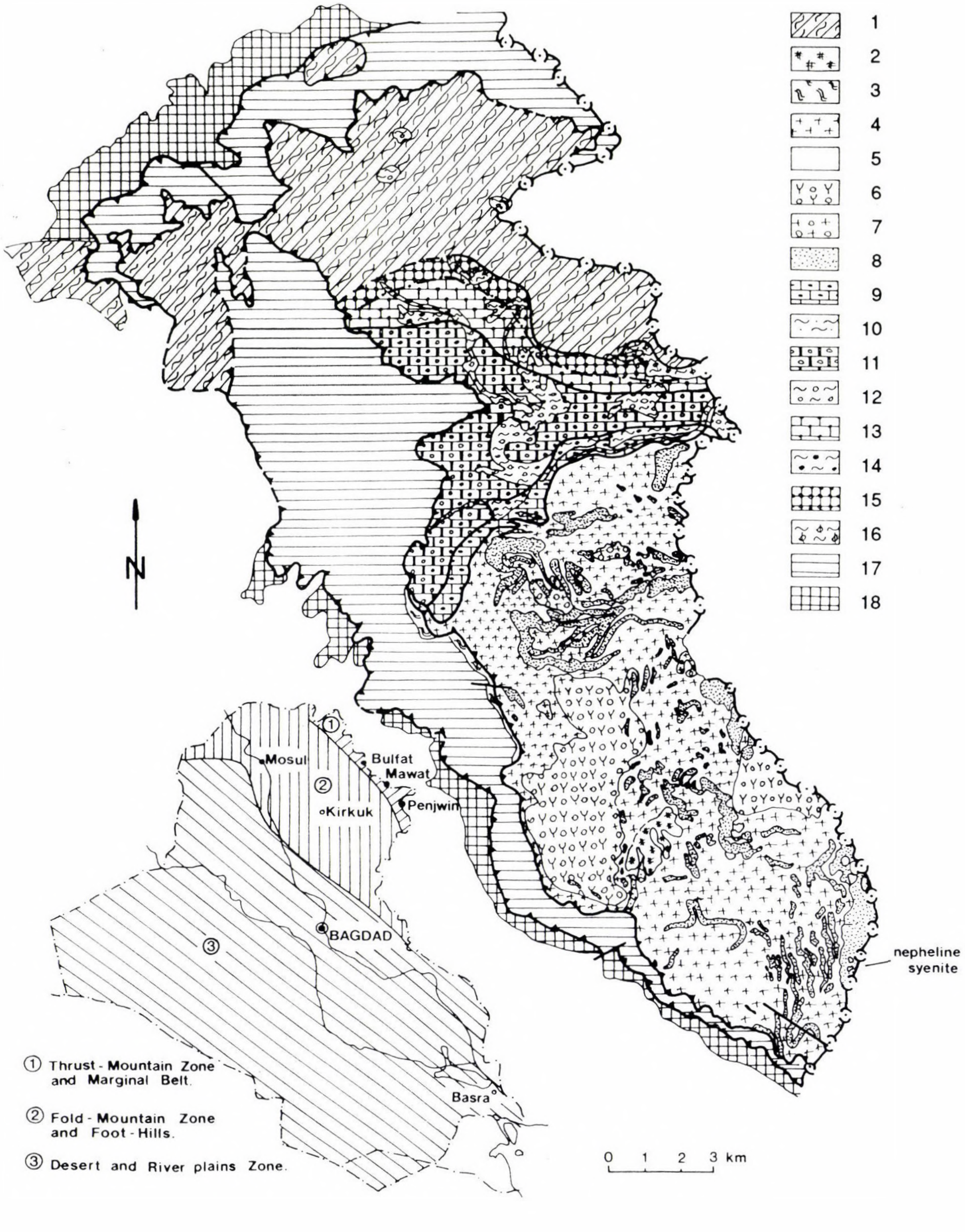
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Mosul
Bulfat
Mawat
Penjwin
Kirkuk
BAGDAD
Basra
① Thrust-Mountain Zone and Marginal Belt.
② Fold-Mountain Zone and Foot-Hills.
③ Desert and River plains Zone.
0 1 2 3 km

Geological setting

The outcrop of plutonic complex is about 15–19 km long in a NNW–SSE direction and up to 5–7 km in width. The complex is tilted towards the east. The exposed thickness is about 1500 m. The lower part consists of probably tectonically-emplaced peridotite, the middle part of gabbro-diorite and in the upper part nepheline syenite occurs just below the contact with skarn rock (Buda 1993). The main pluton was intruded by younger, smaller olivine gabbro-diorite. The earlier intrusive has xenolith (roof-pedants) and a wide thermal contact aureole which extends to the north up to 2.5 km from the last outcrop of the intrusive body (Fig. 1). The current position of the igneous-metamorphic complex is allochthonous, having been transported from the ENE to its present position.

Petrography of the thermal metamorphic rocks

Xenoliths (roof pedants)

The calcareous and basic magmatic rocks – metamorphosed under the conditions of pyroxene-hornfels facies – occur in the gabbroic intrusions as xenoliths.

The strongly sheared amphibole-pyroxene gabbro and diorite contain more xenoliths than the younger olivine gabbro and diorite. The mineral assemblages of these xenoliths are variable, depending on the chemical composition of the parental rocks. Siliceous dolomitic limestone is dominant, but due to the different ratios of Ca, Mg and Si a wide range of mineral assemblages were formed, caused by the thermal metamorphism. The main rock types are: coarse-grained marble, diopside marble, forsterite marble (ophicalcite), predazzite, diopside-wollastonite marble, diopside-wollastonite hornfels, diopside hornfels, melilite–wollastonite marble and hornfels, anthophyllite hornfels and amphibolite.

←Fig. 1
Geological map of igneous and metamorphic rocks of Bulfat thrust block (Northeast Iraq). 1 Qandil regional metamorphic rocks, (green schist facies); 2–5; Old intrusion, 2. ultrabasic rocks, 3. serpentinite (in the Thrust Zone), 4. Amphibole-pyroxene gabbro and diorite, 5. nepheline syienite; 6–7. Young intrusion, 6. olivine-amphibole-pyroxene gabbro, 7. olivine-amphibole- pyroxene diorite; 8–10. Pyroxene hornfels facies, 8. xenoliths in gabbro and diorite, 9. contact metamorphic calcareous rocks, 10. contact metamorphic pelitic rocks; 11–12. Hornblende- hornfels facies, 11. contact metamorphic calcareous rocks, 12. contact metamorphic pelitic rocks; 13–14 Transitional facies, 13. contact metamorphic calcareous rocks, 14. spotted schist; 15–16. albit-epidot facies, 15. contact metamorphic calcareous rocks, 16. spotted slated; 17. Walash volcano-sedimentary series; 18. red beds

a. Coarse-grained marble: contains coarse-grained calcite with polysynthetic deformation twins. The crystal boundaries are well developed and form a typical mosaic texture. Sometimes a slight secondary deformation can be observed causing wavy extinction and bent twin-lamellae. The majority of these rocks contain a very small amount (1–2%) of slightly pleochroic (yellowish-brown) flaky or columnar phlogopite with high Al and Mg and very low Fe content (Table 1). Diopside, feldspar and olivine are rare. Olivine is sometimes entirely altered to serpentine. Diopside has a very low iron, high Ca and Mg content (Table 1). This rock type occurs at the immediate contact with gabbro. The parental rock was magnesian limestone with Si impurities (Table 2 and Fig. 2/A, C).

b. Diopside marble: Two subgroups can be distinguished:
– diopside marble has granoblastic (mosaic) texture. The polysynthetic twins are always present in calcite. Colourless or slightly greenish, subhedral (or hypidioblastic) rounded grains of diopside are frequent. Titanite, muscovite and plagioclase occur as accessory minerals.
– phlogopite diopside marble has a deformed mosaic texture or transition between mosaic and sutured one. Besides calcite (85–95 Vol%) it also contains dolomite. The small colourless rounded grains of diopside are characteristic and phlogopite is abundant. The mica is slightly pleochroic (γ'=light-brown, α'=colourless) and contains lower Si and Al and higher Mg and Fe than phlogopite in the coarse-grained marble (Table 1). The chemical composition of the rock is very similar to the coarse-grained marble (Table 2, Fig. 2/A, C) slightly higher content K and Al due to higher amount of phlogopite.

c. Forsterite marble (or ophicalcite): The dominant mineral is calcite. Three textural types of marble can be distinguished: mosaic, mosaic sutured and sutured. The latter ones always contain dolomite. The forms of the dolomite crystals are variable, mostly rounded or elongated, and show uniform extinction inside the host calcite, suggesting an exsolution origin (Plate I: 1). The polysynthetic twins are always present which show post-contact metamorphic deformation (bent twin-lamellae, wavy-extinction, etc.) due to dynamo-metamorphism. The amount of olivine which forms idioblastic or hypidioblastic crystals is variable (Plate I: 2). Along the cleavage and cracks the olivine is altered to fibrous serpentine. The rate of serpentinization is variable and sometimes the whole crystal is altered to serpentine. Spinel, which occurs as an accessory mineral, forms idioblastic (cubic or octahedral) colourless or pinkish grains. Perovskite is very rare, mostly xenoblastic or hypidioblastic, slightly bluish in colour with polysynthetic twins. It contains a considerable amount of REE (Table 1). About 70 Vol% of the samples have very clear contact with the gabbroic rocks, others are close to the gabbro but the contact is not so obvious. The parental rock was dolomitic limestone with Si impurities (Table 2 and Fig. 2/A ,C).

d. Predazzite: The main constituent is calcite (58–70 Vol%) which shows mosaic or sutured texture with polysynthetic twins. Small rhombohedral grains of

Table 1
Microprobe analyses of representative minerals of calcareous xenolith

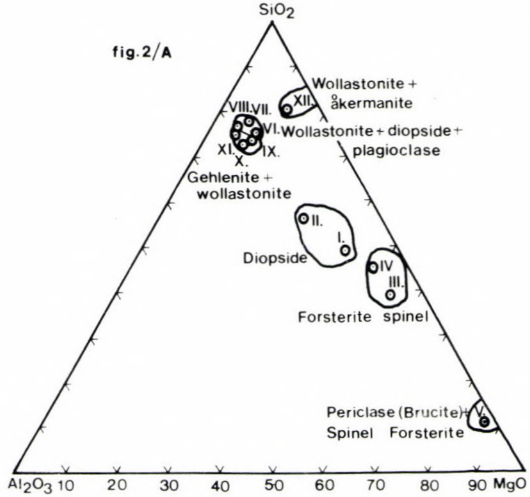
wt %	1	2	3	4	5	6	7	8	9
SiO_2	52.68	41.80	35.46	-	42.90	-	-	37.74	36.23
TiO_2	0.10	0.86	0.95	58.90	-	1.57	-	2.10	0.80
Al_2O_3	1.13	19.50	19.10	0.27	-	68.68	-	16.31	15.11
FeO	0.43	0.70	1.60	0.10	0.24	0.96	0.92	3.20	4.24
MgO	18.80	26.12	28.40	0.07	56.94	28.39	66.45	3.30	2.40
CaO	24.80	0.10	1.90	36.16	0.06	0.03	-	36.30	36.99
Na_2O	0.20	-	-	0.31	-	-	-	0.21	0.07
K_2O	-	9.35	8.20	-	-	-	-	-	-
Σ	98.14	98.43	95.51	95.81	100.14	99.63	67.37	99.16	95.84
				Number of ions on the basis of					
	6 (o)	22 (o)		24 (o)	4 (o)	32 (o)	2 (o)	14 (o)	76 (O, OH)
Si	1.943	5.575	4.970	-	1.004	-	-	3.43	17.120
Al^{IV}		2.425	3.030		-		-		0.880
Al^{VI}	0.050	0.641	0.126	0.059		15.519		1.75	7.513
Ti	0.003	0.086	0.100	8.287	-	0.226	-	0.14	0.283
Fe^{2+}	0.013	0.078	0.188	0.016	0.005	0.154	0.008	0.24	1.671
Mg	1.034	5.191	5.933	0.019	1.986	8.110	0.950	0.45	1.685
Ca	0.980	0.014	0.270	7.248	0.002	0.006	-	3.53	18.678
Na	0.014	-	-	0.112	-	-	-	0.04	0.062
K	-	1.592	1.466	-	-	-	-	-	-
	Mg 51.1				Fo 99.7			Ge 80%	
	Fe 0.4				Fa 0.3			Å 20%	
	Ca 48.5								

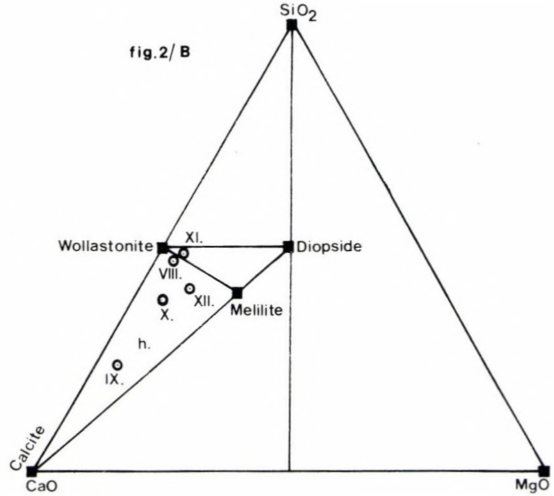
Symbols: 1. Diopside from coarse-grained marble (No. 68); 2. Phlogopite from the same rock; 3. Phlogopite from diopside marble (No. 96); 4. Perovskite from forsterite marble (No. 296, enriched in Ce, La, Y, Nb, Ta only qualitatively measured); 5. Forsterite from predazzite (No. 26A); 6. Spinel from the same rock (No. 26A); 7. Brucite from the same rock (No. 26A); 8. Melilite from melilite-wollastonite marble (No. 50); 9. Vesuvianite from the same rock (No. 50)

Table 2
Chemical composition of calcareous xenoliths

wt%	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
SiO_2	1.71	1.65	4.73	5.85	2.51	10.46	15.23	36.13	16.80	29.36	39.83	38.55
TiO_2	0.12	0.06	0.17	0.20	0.05	0.14	0.27	0.47	0.26	0.47	0.61	0.20
Al_2O_3	0.38	0.47	0.86	0.90	0.45	2.12	3.13	8.58	3.93	7.80	10.22	3.42
Fe_2O_3	0.13	0.12	0.25	0.45	0.01	0.61	0.72	1.65	1.08	1.28	2.64	0.45
FeO	0.11	0.11	0.23	0.09	0.32	0.45	0.84	0.61	0.95	0.87	0.89	0.75
MnO	0.02	0.01	0.01	0.02	0.02	0.06	0.09	0.06	0.03	0.05	0.11	0.02
MgO	1.40	0.83	6.48	6.08	19.05	1.34	1.28	2.35	2.24	3.30	2.95	6.29
CaO	53.00	54.60	47.31	47.11	38.84	49.00	45.98	34.91	45.70	39.82	37.15	42.68
Na_2O	0.39	0.33	0.32	0.33	0.28	0.64	0.83	0.39	1.04	1.12	1.12	0.25
K_2O	0.12	0.21	0.05	0.04	0.05	0.33	0.65	1.97	0.25	0.45	0.09	0.41
P_2O_5	0.05	0.06	0.04	0.07	0.04	0.09	0.11	0.10	0.06	0.16	0.07	0.04
H_2O	0.30	0.14	0.30	0.32	0.33	0.35	0.18	0.43	0.35	0.33	0.38	0.46
H_2O^+	0.06	0.07	0.40	0.08	0.08	0.08	0.14	0.14	0.10	0.05	0.05	0.88
CO_2	42.17	41.19	37.93	38.43	37.95	33.78	30.01	10.30	25.85	13.28	2.47	5.85
Cl	0.01	0.01	0.02	0.01	0.02	0.02	0.02	0.01	0.01	0.01	0.01	-
SO_3	+	+	0.74	+	+	0.50	0.62	1.12	0.92	1.12	0.58	0.06
Σ	99.97	99.86	99.84	99.98	100.00	99.97	100.10	99.22	99.57	99.47	99.17	100.31

Symbols: I. Coarse-grained marble (No. 68); II. Diopside marble (No. 96); III–IV. Forsterite marble (No. 296, 137B); V. Predazzite (26A); VI–VII. Diopside-wollastonite marble (No. 60A, 191); VIII. Diopside-wollastonite hornfels (No. 289); IX–X. Melilite wollastonite marble (No. 2A, 50); XI. Wollastonite-melilite hornfels (No. 130); XII. Melilite wollastonite (garnet) hornfels (31B)





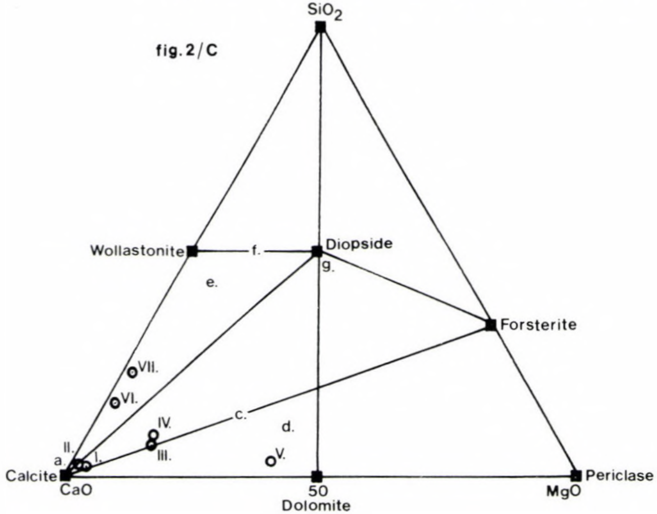


Fig. 2/A
SiO_2, Al_2O_3, MgO plot and index minerals of high-grade pyroxene-hornfels facies of Bulfat calcareous xenoliths. I–XII: chemically analyzed samples

Fig. 2/B
Pyroxene-hornfels facies: derivatives of calcareous shales and marls (xenoliths from gabbro, Bulfat). VIII–XII: chemically analyzed samples

Fig. 2/C
Pyroxene-hornfels facies: derivatives of silica-deficient dolomitic limestone (xenoliths from gabbro, Bulfat) I–VII: chemically analyzed samples

dolomite can be sometimes recognized, too. Brucite is frequent (30 Vol%) showing onion-skin texture (Table 1; Plate I: 3). Sometimes in brucite partly serpentinized olivine and small remnants of periclase occur. Serpentine and brucite have been formed by hydration of olivine and periclase, respectively. This aggregate is surrounded by a thin layer of dolomite. The amount of olivine is variable and it forms idioblastic or rounded crystals. The composition is forsterite (Table 1). Spinel occurs in every sample as an accessory mineral (Table 1). It forms small octahedral colourless crystals. These calcareous rocks were originally calcitic dolomite with Si impurities (Table 2 and Fig. 2/A, C) and occur at the immediate contact with gabbro.

e. Diopside-wollastonite marble: The main constituent is calcite (70–80 Vol%), which forms mosaic texture. The crystals are sometimes strongly deformed, showing wavy extinction and bent polysynthetic twin lamellae. Some samples contain rhombohedral dolomite, formed most probably by exsolution. Wollastonite occurs in columnar or tabular form, grain size is variable, commonly twinned. Diopside occurs (7–10 vol%) in every sample, mostly colourless, rounded and sometimes deformed. Garnet (probably andradite) is always present but in variable amount. It is mostly colourless or slightly pinkish, and forms xenoblastic crystals with poikiloblastic texture. The rare plagioclase is anhedral and the optical sign is positive or negative. The composition is most probably bytownite or labradorite. In some samples scapolite and prehnite occur. They are mostly alteration product of plagioclase. The original rock was argillaceous limestone with Mg impurities (Table 2 and Fig. 2/A, C) and they generally occur near the gabbro.

f. Diopside-wollastonite hornfels: Two subgroups were distinguished:
– medium- to fine-grained, grey, whitish-grey rock with slight schistosity and lamination. The texture is granoblastic, oriented, sometimes porphyroblastic (e.g. wollastonite). The prevailing wollastonite grains are arranged in a parallel direction. Diopside and garnet are evenly distributed or sometimes arranged into streaks. Diopside occurs in nearly every sample, forming xenoblastic, mostly rounded or sometimes columnar crystals. They are slightly zoned, the core is darker than the rim (due to the very fine opaque inclusions). Calcite also occurs in nearly every sample but in smaller amounts, having xenoblastic, sometimes mosaic texture. Polysynthetic twin lamellae are deformed. The plagioclase is xenoblastic, sometimes polysynthetically twinned the composition is labradorite (An_{62}) and bytownite. Garnet is not frequent and does not occur in every sample. They are mostly colourless or sometimes slightly pinkish, isotropic probably andradite. Beside these minerals scapolite, prehnite, altered melilite (vesuvianite), tremolite, quartz, chlorite, zoisite and titanite occur. Some of these minerals are definitely secondary; for example, vesuvianite is an alteration product of melilite. Prehnite always occurs as a vein fillings. This rock type occurs as xenoliths in gabbro.

– fine-grained, laminated, green, greenish-grey rock. The greenish-grey layers alternate with dark-green ones. The mineral composition of this rock is similar to the previous one but the texture is different. The fine-grained garnet, wollastonite, plagioclase, diopside content layer alternates with one containing coarse-grained calcite and diopside, suggesting a pre-contact metamorphic (or pre-metamorphic) microbedding. The parental rock most probably was microbedded calcareous shale and marl (Table 2 and Fig. 2/A, B). Wollastonite, diopside, plagioclase (An_{58}), calcite and titanite occur in every sample. Garnet, scapolite, tremolite zoisite, quartz, chlorite etc. also occur, but not in every sample.

g. Diopside hornfels: Two subgroups were distinguished:

– dark-grey, greenish-grey, coarse-grained rock without any orientation. The dominant constituent is diopside-hedenbergite (80–95 Vol%). It is greenish in colour, showing decussate texture. Besides the (110) cleavage sometimes (001) parting also occurs (malacolite-type). Calcite, titanite, apatite and biotite are accessory minerals. Zoisite, scapolite, prehnite, sericite were formed by alteration of plagioclase. These rocks occur as xenolith in the gabbro.

– dark-grey, greenish-grey, fine-grained, laminated rock. The prevailing constituent is granoblastic (mosaic) colourless diopside. Plagioclase is always present and it also forms mosaic texture. Two generations can be recognized: the first one is altered (cloudy) and the second one, newly formed, is water-clear, unaltered. Polysynthetic twins are not frequent. The composition of the latter one is labradorite (An_{70}). Calcite, titanite, prehnite, zoisite, garnet and tremolite, actinolite occur in accessory amounts.

h. Melilite-wollastonite marble and hornfels: Two subgroups can be distinguished:

– melilite-wollastonite marble: coarse- to medium-grained, grey, greenish-grey rocks, without any preferred orientation. The melilite is grey; when it is altered to vesuvianite the colour becomes greenish-yellow or sometimes pale-green. It forms the groundmass of the rock. In the groundmass reddish garnet spots and white or colourless calcite occur. White columnar wollastonite crystals can be sometimes recognized, too. Calcite forms large tabular crystals with typical mosaic texture and with polysynthetic twins. Melilite is poikiloblastic tabular or columnar, nearly isotropic; the interference colour is dark-grey or anomalous blue, the composition is a gehlenite rich melilite (Table 1, $Ge_{80}Å_{20}$). Vesuvianite was formed by hydration of melilite. It also contains small amount of chlorite. Its chemical composition is nearly same as melilite, with smaller amounts of Si, Al and Mg which is due to Mg-chlorite formation beside vesuvianite (Table 1).

Wollastonite also occurs in every sample. Garnet is poikiloblastic (anhedral), usually isotropic, slightly pinkish, probably andradite. Scapolite, mica, prehnite are secondary minerals. Classification into marble and hornfels was based on the amount of the silica. Whole transition can be found between marble and hornfels.

– melilite-wollastonite hornfels: medium to fine-grained, grey or greenish-grey (melilite) rock with elongated white wollastonite and with pinkish spots of garnet. The main constituents are similar to the previous rock type (calcite, wollastonite, melilite), but besides them diopside and feldspar also occur. The parental rocks were calcareous shale and marls (Table 1, Fig. 2/A, B).

i. Anthophyllite hornfels: It is a very rare rock and occurs at the contact with diorite. It is fine-grained, white and becomes slightly pinkish (garnet), black and coarser-grained towards the contact. The white part consists of calcite, garnet, diopside, plagioclase, titanite, altered melilite (vesuvianite), prehnite and the dark part contains anthophyllite, garnet, phlogopite, calcite sometimes plagioclase (andesine).

j. Amphibolites: are strongly foliated and sometimes folded. They occur in the sheared pyroxene amphibole gabbro and diorite at the south-eastern part of the intrusion surrounded by calcareous metamorphic xenoliths. These were most probably basic igneous rocks which metamorphosed and may be partly assimilated by the gabbroic intrusion.

The prevailing constituent is the strongly pleochroic green hornblende ($\gamma/C=20$). They are mostly prismatic, well-oriented and sometimes show mosaic texture. Plagioclase is also frequent, mostly altered (saussuritized) but sometimes unaltered; newly formed generation can be observed, too. Titanite and apatite occur in accessory amount. Prehnite is in the form of secondary vein filling. Chlorite and zoisite are alteration products of amphibole and plagioclase respectively.

Summary

The basic melt intruded into metasediments which were assimilated or metamorphosed "in situ". This "in situ" position of xenoliths was concluded from the long belts or streaks of metamorphic rocks in the gabbro-dioritic body (the length is sometimes several km), running parallel with the strike of the country rocks. Further evidence is the similar metamorphic mineral assemblages of the xenoliths and contact rocks in the inner aureole (pyroxene-hornfels facies). Obviously the latter one never attained as high temperature as the recrystallization of xenoliths.

The composition of metasediments were variable, reflected in variable assimilation; others had undergone isochemical reactions, forming new mineral assemblages. The majority of pelitic rocks were assimilated due to their low melting point and crystallized as igneous-like rock and it was possible to recognize them, e.g. dark, strongly pleochroic biotite, hypersthene, lower anorthite content of plagioclase, etc. The basic rocks were partially melted, forming amphibolite xenoliths. They originated from basic volcanites.

The calcareous rocks mostly resisted melting but a new high-temperature mineral assemblage formed.

The different mineral formations are due to the variable chemical composition of parental rocks, because in xenoliths, similar PT prevailed, e.g. periclase can form from dolomite at about 700–760 °C (at 500 bars); if the rock contains Si, åkermanite, or with Al gehlenite can crystallize, etc., at the similar physical conditions.

The composition of parental rocks show a wide range: magnesian limestone (with Si impurities), dolomitic limestone, calcitic dolomite, argillaceous limestone, calcareous shale, marl, etc. (Table 2 and Fig. 2/A,B,C).

New mineral assemblages and parental rocks of xenoliths

Mineral assemblages	Parental rocks
1. Calcite and accessory amount of diopside + phlogopite	Magnesian limestone (Si-impurities)
2. Calcite (dolomite)+ forsterite (serpentine)+ spinel+perovskite	Dolomitic limestone (Si-impurities)
3. Calcite (dolomite) + periclase (brucite) + spinel + forsterite (serpentine)	Calcitic dolomite (Si-impurities)
4. Calcite+diopside+wollastonite+ plagioclase +garnet	Argillaceous limestone (Mg imp.)
5. Calcite+melilite (a. gehlenite Ge_{80} $Å_{20}$, b. åkermanite-rich) + wollastonite + garnet	Calcareous shale and marl
6. Diopside+plagioclase+calcite	Basic igneous rock ?
7. Anthophyllite + garnet + melilite (vesuvianite)	Calcareous rock assimilated by gabbroic melt

Reactions caused by high temperature gabbroic melt

1. Diopside: $CaMg(CO_3)_2 + 2SiO_2 = CaMgSi_2O_6 + 2CO_2$

dolomite + Q = diopside

(1 Kbar, 540–700 °C)

Phlogopite:

$3CaMg(CO_3)_2 + KAlSi_3O_8 + H_2O = KMg_3[AlSi_3O_{10}](OH)_2 + 3CaCO_3 + 3CO_2$

dolomite + K-feldspar (detrital) = phlogopite + calcite + CO_2

2. Forsterite: $2CaMg(CO_3)_2 + SiO_2 = 2CaCO_3 + Mg_2SiO_4 + 2CO_2$
dolomite + Q = calcite + forsterite + $2CO_2$
(1 Kbar, 700 °C)

Spinel + forsterite:
$Mg_5Al_2Si_3O_{10}(OH)_8 = MgAl_2O_4 + 2Mg_2SiO_4 + SiO_2 + H_2O$
Mg-chlorite = spinel + forsterite + Q + H_2O

3. Periclase: $CaMg(CO_3)_2 = CaCO_3 + MgO + CO_2$
dolomite = calcite + periclase + CO_2
(1 Kbar, 760–820 °C)
4. Wollastonite: $CaCO_3 + SiO_2 = CaSiO_3 + CO_2$
calcite + Q = wollastonite + CO_2
(1Kbar, 630 °C)
5. Melilite: $CaCO_3 + CaMgSi_2O_6 = Ca_2MgSi_2O_7 + CO_2$
calcite + diopside = åkermanite + CO_2
(1 Kbar, 725 °C)

6. Anthophyllite: reactions mostly in ultrabasic rocks
$9Mg_6(Si_8O_{20})(OH)_4 + 4Mg_2SiO_4 = 5Mg_7Si_8O_{22}(OH)_2 + H_2O$
talc + forsterite = anthophyllite + H_2O
(1 Kbar, 660 °C)

The temperature of the thermal metamorphism was about 600–800 °C at 1000 bars pressure concluded from the investigated mineral assemblages. These temperatures and pressures correspond to the pyroxene-hornfels facies. At the immediate contact higher temperature could occur, but further detailed studies are needed to prove its presence by investigation of the characteristic high temperature minerals. Merwinite, spurrite, rankinite, etc. were not identified because of the lack of suitable samples, but their presence are expected. Secondary reactions at a lower temperature caused by retrograde metamorphism are as follows:
Dolomite exsolution at about 600 °C
Vesuvianite formation from melilite by hydration at 600–700 °C
Brucite formation from periclase by hydration.
Serpentine formation from olivine by hydration

Dolomite rim formation around periclase or brucite:
$3CaCO_3 + Mg(OH)_2 + CO_2 = CaMg(CO_3)_2 + H_2O$

Scapolite formation from plagioclase:
$3CaAl_2Si_2O_8 + CaCO_3 = Ca_4Al_6Si_6O_{24}CO_3$
anorthite = meionite

Contact aureole of the intrusion

The regional metamorphic Quandil group (originally pelitic-, calcareous-sediments and basic volcanites) was thermally metamorphosed in the vicinity of the gabbroic intrusion. From the contact towards the regionally metamorphosed rocks, the mineral assemblages and texture changed because of the decreasing temperature gradient. The hornfelsic texture gradually disappears and the rocks become more foliated. Determination of the exact boundary of the thermal aureole is very difficult. The spotted slates formed due to the remote thermal effect of the intrusion and phyllites are assumed to have formed by regional metamorphism. It is very difficult to determine the real distances from the contact because of the dip, consequently the depth of the contact (with the plutonic body) is unknown. Approximate distances between the contact and outcrops of the hornfelsic rocks can be estimated from the horizontal distances and from the dips of the metamorphic rocks if they are parallel with the contact. The nature of the intrusion (depth, size, etc.) and later tectonic movements can modify these parameters. The remote thermal effect of the gabbro intrusion according to these estimations was up to about 2.5 km. The thickness of the medium-grade one does not extend further than 1.6 km from the contact (Table 5).

A. Contact metamorphic pelitic rock

Three main facies were distinguished according to their mineral assemblages, which were classified into six subgroups based on the texture. The facies are pyroxene-, hornblende-, and albite-epidote hornfels. The last one intermingles with the regional metamorphic rocks.

I. Pyroxene-hornfels facies

a. Pelitic hornfels with granoblastic texture: the rock is medium- to fine-grained, equigranular, slightly foliated. The fresh surface is grey, greyish-brown and the weathered one is brown. The whitish-grey feldspar sometimes forms bands or lenses (due to a slight mobilization). Slightly orientated biotite, grey cordierite and quartz are always present. Every sample contains string perthitic, tabular (or xenoblastic) orthoclase (2V is small). Plagioclase is very frequent, mostly tabular, sometimes with polysynthetic twins (An_{36-43}, Table 3). Myrmekitic texture occurs when it is attached to K-feldspar, suggesting K mobilization. Cordierite is also frequent, mostly tabular or columnar with some inclusions. It has a high FeO content (13.6 wt%). The sectoral twins are common (Plate I: 4). The crystals are pinitized along the margins or cracks and around the zircon inclusions yellow pleochroic halos occur. Quartz shows slight wavy extinction. The biotite is strongly pleochroic (dark-reddish-brown, yellowish-brown or colourless), caused by high TiO_2 content. It sometimes shows intergrowth with sillimanite and contains zircon inclusions with pleochroic halos. Sillimanite occurrences are irregular and were not determined in every sample. It shows slender fibrous aggregates and formed at the expense of biotite. K-feldspar crystallized simultaneously when potassium was released from biotite (Plate II: 1). Poikiloblastic garnet is rare, slightly pinkish in colour, isotropic, most probably almandine. Accessory minerals are: apatite, zircon, tourmaline, andalusite (very rare) and ilmenite. Secondary minerals are: chlorite and sericite.

Hypersthene containing xenoliths occurs in the gabbroic body. This rock type represents the highest temperature mineral assemblage of pelitic rocks which was attained during thermal metamorphism. The hypersthene forms columnar crystals with blasto-ophitic texture ($Mg_{63}Fe_{34}Ca_3$, Table 3), plagioclase has an andesine composition (An_{36}, Table 3). Myrmekitic texture also formed. K-feldspar is tabular, containing plenty of inclusions of plagioclase. Quartz is rare but biotite is always present, showing strong pleochroism (dark-reddish-brown, yellowish-brown) corresponding with the high TiO_2 content (Table 3).

II. Hornblende hornfels facies

b. Pelitic hornfels with transitional texture between granoblastic and poikiloblastic: megascopic appearance of this rock type is very similar to the previous one but slightly finer-grained with weak foliation. The mineral assemblage is nearly the same but orthoclase does not occur; sometimes microcline can be recognized. Cordierite and garnet have a well-developed poikiloblastic texture. The thickness of this unit is between 150 m and 480 m.

c. Pelitic hornfels with poikiloblastic texture: these rocks are brown, greyish-brown, fine-grained and foliated. The main difference from the two previous groups is the first appearance of the poikiloblastic andalusite, occurring together with cordierite. The poikiloblastic texture is widespread (Plate II: 2) (andalusite, cordierite, biotite), with foliation. Sillimanite is rather frequent, it forms fibrous aggregates in close relation with biotite. The distance from the contact is about 1400 m.

d. Pelitic hornfels with transitional texture between poikiloblastic and porphyroblastic: the rock is dark-brown, fine-grained, foliated. The main difference from the previous one is the presence of the large porphyro-poikiloblastic andalusite (chiastolitic; Plate II: 3) and cordierite. In some samples sillimanite also occurs. The pleochroism of tourmaline becomes pale (ε'=colourless, ω'=greenish-brown). Average distance from the contact is about 1600 m.

e. Porphyroblastic pelitic schist (transitional facies between hornblende and albite-epidote-hornfels facies): the rock is dark-grey in colour with silky lustre and fine-grained, with strong foliation. The deformed prismatic or rounded grey porphyroblast of chiastolite is characteristic (Plate II: 4). The margin or sometimes the whole crystal altered to white mica. Quartz is very frequent with wavy extinction. Besides biotite, muscovite also occurs; cordierite and plagioclase are completely absent. This rock-type extends some 2.3 km from the contact with the intrusion.

Table 3
Microprobe analyses of representative minerals of pelitic hornfels

wt%	1	2	3	4	5		6	7	8		9	10
SiO_2	36.53	36.65	37.59	34.11	37.95		48.42	47.42	49.40		60.30	61.40
TiO_2	6.08	3.35	4.30	3.51	1.95		0.25	-	0.11		-	-
Al_2O_3	13.34	18.80	17.54	20.17	18.80		0.80	30.90	29.10		22.35	24.30
FeO	14.78	18.80	18.20	22.52	19.30		23.84	13.60	13.30		0.05	0.10
MnO	0.09	0.17	-	0.12	0.40		0.90	0.45	0.27		-	-
MgO	15.97	8.20	7.10	7.10	9.85		22.45	6.45	7.01		-	-
CaO	0.06	0.10	-	-	-		1.65	-	-		8.30	8.61
Na_2O	0.40	0.40	0.28	0.27	0.60		-	0.45	0.30		7.90	6.30
K_2O	9.90	9.10	9.27	9.35	7.91		-	0.08	-		0.40	0.10
Σ	97.15	95.57	94.28	97.15	96.76		98.49	99.35	99.49		99.30	99.49
	Formula based on		22 (o)				6 (o)		18 (o)			32 (o)
Si	5.40	5.52	5.72	5.18	5.60	Si	1.882	4.982	5.155	Si	10.904	10.841
Al^{IV}	2.32	2.48	2.28	2.82	2.40	Al^{IV}	0.037	1.018	0.845	Al^{IV}	4.763	5.057
Al^{VI}	-	0.86	0.86	0.80	0.88	Al^{VI}	-	2.808	2.734	Fe	0.008	0.015
Ti	0.68	0.38	0.50	0.40	0.22	Ti	0.007	-	0.009	Na	2.769	2.157
Fe	1.82	2.38	2.32	2.86	2.38	Fe	0.772	1.195	1.161	Ca	1.608	1.629
Mg	3.52	1.84	1.62	1.60	2.16	Mg	1.296	1.010	1.090	K	0.090	0.022
Mn	0.02	0.02	-	0.02	0.04	Mn	0.029	0.040	0.024	Ab	62	56.6
						Ca	0.069	-	-	An	36	42.8
Σ	6.04	5.48	5.30	5.68	5.68	Na	-	0.092	0061	Or	2	0.6
K	1.86	1.70	1.80	1.82	1.50	K	-	0.011	-			
Na	0.12	126	0.08	0.08	0.08	Mg	62.9					
Ca	0.02	0.02	-	-	-	Fe	33.9					
Σ	2.00	1.90	1.88	1.90	1.58	Ca	3.2					

Symbols: 1. Biotite from biotite-hypersthene plagioclase hornfels (No, 40B); 2. Biotite from cordierite hornfels (No. 56, 100 m from the contact); 3. Biotite from cordierite hornfels (No. 61, 640 m from the contact); 4. Biotite from cordierite-andalusite-sillimanite hornfels (No. 73, 950 m from the contact); 5. Biotite from andalusite schist (No. 142, 1340 m from the contact); 6. Hypersthene from biotite hypersthene-plagioclase hornfels (No. 40B); 7. Cordierite from cordierite hornfels (No. 56); 8. Cordierite from cordierite hornfels (No. 61); 9. Plagioclase from biotite-hypersthene-plagioclase hornfels (No. 40B); 10. Plagioclase from cordierite hornfels (No. 61).

III. Albite-epidote-hornfels facies

f. Spotted slate: this rock is dark-grey with silky lustre, strongly foliated, very fine-grained with spots. The spots are dark-grey, sometimes brownish, rounded, slightly elongated and contain quartz, very small grains of acicular or flaky sericite and chlorite which has an anomalous blue interference colour. In some samples the spots preserved the prismatic form of andalusite but entirely altered to white mica (sericite). The groundmass contains very fine-grained, strongly oriented muscovite, tourmaline and dust-like opaque material (most probably, graphite). Some samples contain weakly pleochroic biotite (greenish-brown, yellowish-brown) which was sometimes chloritized. Garnet (euhedral), zoisite, titanite and zircon occur in an accessory amount. Approximate distance from the contact is about 2.5 km.

Phyllite: does not belong to the contact series. It occurs in the Quandil formation, alternating with metacalcareous and metavolcanic rocks. The rock is dark-grey with silky lustre, strongly foliated and folded. The main constituent is the very fine-grained, oriented quartz with ill-developed crystal-boundaries. The strongly oriented white mica is always present. Three kinds of opaque minerals were distinguished:

1. Dust-like, very fine-grained uniformly distributed organic material (graphite); 2. Pyrite with cubic crystal habit; 3. Large elongated rounded and oriented grains (most probably ilmenite).

In some samples biotite and idioblastic garnet occur, indicating a slightly higher grade of metamorphism.

Prismatic tourmaline occurs in every sample in an accessory amount (ε=colourless or slightly yellowish, ω=green, brownish-green sometimes with colour zonation).

Summary

The above-described rocks were originally marine clay (Table 4, Fig. 3A). Marine origin was suggested from the uniformly distributed tourmaline and from the chemical composition. This was regionally metamorphosed and later thermally metamorphosed due to the gabbroic intrusion.

The contact metamorphic rocks belong to three main facies.

I. Pyroxene-hornfels. High-grade metamorphism.

Occurrences: xenolith and immediate contact (0–150 m)

Characteristic mineral assemblage: orthoclase-plagioclase-cordierite (sillimanite, hypersthene). Texture: granoblastic (Fig. 4/A, B).

II. Hornblende-hornfels: medium-grade metamorphism.

Occurrences: between 150 and 1600 m from contact.

Characteristic mineral assemblage: cordierite (andalusite, sillimanite Fig. 4/A,B). Texture: poikiloblastic.

Transitional facies: between medium- and low-grade metamorphism

Occurrences: between 1600–2300 m from contact.

Table 4
Chemical composition of pelitic contact rocks

	A	B	C	D	E	F	G	H
SiO_2	66.34	62.10	61.54	66.24	62.99	63.85	63.00	69.78
TiO_2	1.00	0.79	0.82	0.76	0.97	0.86	0.76	0.82
Al_2O_3	17.53	17.35	19.61	16.92	18.67	17.25	17.63	14.84
Fe_2O_3	2.00	0.90	1.22	0.67	1.09	0.77	0.27	0.70
FeO	3.93	4.70	4.30	4.13	4.27	4.10	2.86	2.08
MnO	0.16	0.11	0.10	0.11	0.10	0.14	0.10	0.04
MgO	4.93	1.82	1.65	1.60	1.68	1.47	1.66	0.83
CaO	6.31	1.54	1.40	1.12	1.40	1.96	1.70	1.40
Na_2O	3.14	1.64	0.89	1.42	1.49	1.64	0.77	0.77
K_2O	1.59	3.85	3.76	3.38	3.76	3.01	3.19	3.70
P_2O_5	0.44	0.13	0.18	0.14	0.19	0.11	0.22	0.11
H_2O^-	0.74	0.55	0.47	0.34	0.44	0.39	0.50	0.33
H_2O^+	0.98	3.14	3.01	2.08	2.51	3.23	3.83	3.30
Cl	0.01	0.02	0.02	0.02	0.01	0.02	0.02	0.01
SO_3	trace	1.06	0.58	0.67	trace	0.74	0.80	0.92
Σ	99.70	99.70	99.55	99.60	99.51	99.54	97.31	99.63

Symbols: A. Biotite-hypersthene-plagioclase hornfels (No. 40B); B. Cordierite hornfels (No. 56) C. Cordierite hornfels (No. 61); D. Sillimanite-cordierite hornfels (No. 69); E. Cordierite-andalusite-sillimanite hornfels (No. 73); F. Andalusite schist (No. 142B); G. Porphyroblastic andalusite schist (No. 103); H. Spotted (sericitic) slate (No. 111).

Table 5
Mineral assemblages and estimated prevailing temperatures of recrystallization of Bulfat xenolith and contact rocks

Parental rock type	High temperature pyroxene hornfels facies (xenoliths)	Pyroxene hornfels facies	Hornblende hornfels facies	Albite-epidote hornfels facies
Pelitic	Orthoclase, hypersthene, biotite (Mg, Ti-rich), andesine	cordierite (Fe-rich), biotite, orthoclase microperthite, andesine ± sillimanite, (tourmaline), garnet	cordierite (Mg-rich), biotite (Fe-rich, Ti-poor) andalusite, plagioclase ± sillimanite, garnet, muscovite (tourmaline)	white-mica, chlorite, quartz, (tourmaline)
Calcareous	Periclase, melilite (Ge$_{80}$Å$_{20}$) spinel, forsterite, wollastonite (phlogopite), garnet	wollastonite, diopside ± garnet	diopside, tremolite, quartz	calcite (dolomite), tremolite
Basic	Clinopyroxene (diopside), basic plagioclase, anthophyllite	not identified	green hornblende, plagioclase	actinolite, zoisite, albite
Occurrence	xenoliths in the intrusion	pelitic: 0–150 m, calcareous: max: 270 m	pelitic: 150–1600 m calcareous: 270–1400 m, max: 1600 m	pelitic: 1600–2500 m calcareous: 1400–2000 m, max: 2500 m
Temperature at 1000 bars	Pelitic: 650–850 °C calcareous: 540–630 °C basic: 660 °C, max: 850 °C	pelitic: 560–580 °C calcareous: 540–630 °C, max: 630 °C	pelitic: 400–520 °C calcareous: 450–540 °C, max: 540 °C	pelitic: up to 400 °C calcareous: below 400 °C, max: 400 °C

Fig. 3/A
Chemical composition variation of pelitic hornfelses of Bulfat compared with clays and shales of marine sediments (Winkler 1974)

Fig. 3/B
Pyroxene-hornfels facies (excess SiO_2, quartz, orthoclase, biotite are always present): derivatives of pelitic rocks. A-D: chemically analyzed samples

Fig. 3/C
Hornblende-hornfels facies (excess SiO_2 deficient in K_2O): derivatives of pelitic rocks. E, F: chemically analyzed samples

Fig. 3/D
Albite-epidote-hornfels facies (excess SiO_2): derivatives of pelitic rocks. G, H: chemically analyzed samples

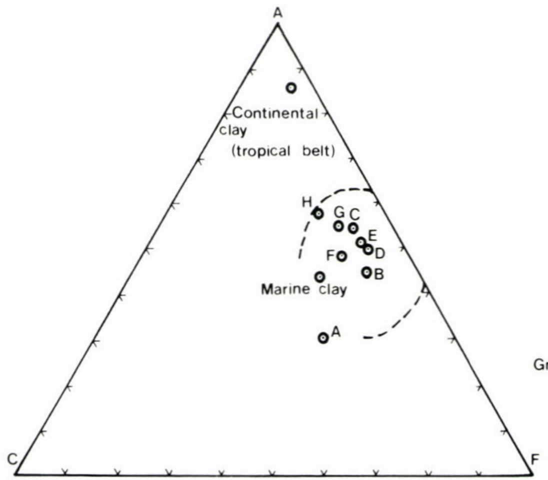
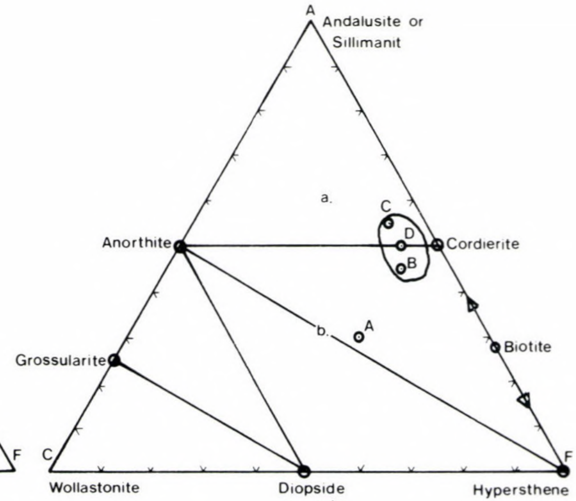


fig. 3/A



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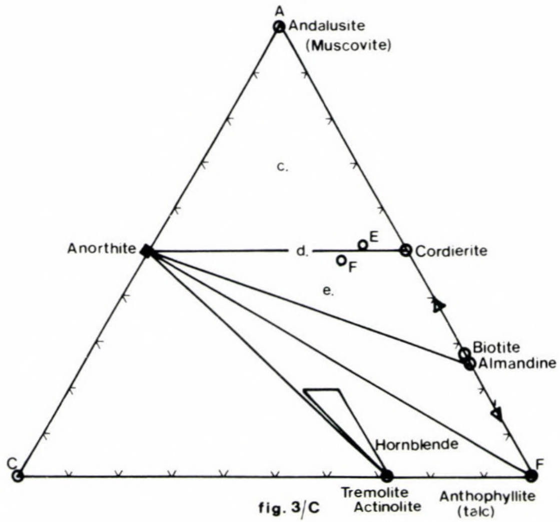

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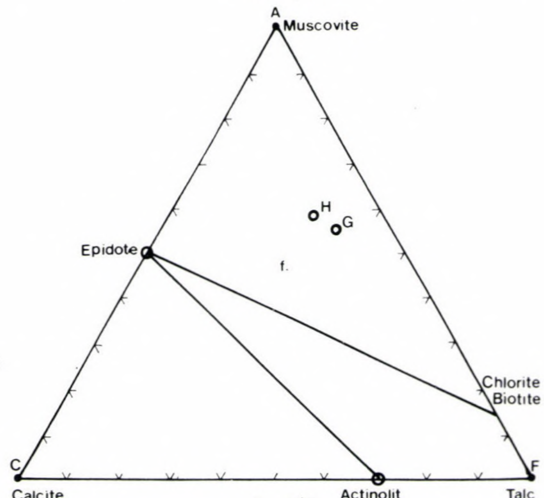


fig. 3/D

Characteristic mineral: andalusite (porphyroblast). Texture: porphyroblastic.
III. Albite-epidote hornfels: low grade metamorphism
Occurrences: between 2300–2500 m.
Characteristic mineral assemblage: quartz, white mica, chlorite (Fig. 4/A, B).
Texture: spotted, foliated.

Chemical composition of the main mineral constituent of pelitic hornfels

The bulk chemical compositions of contact-pelitic rocks are very similar (FeO+MnO/FeO+MnO+MgO ration are between 0.76 and 0.77, Table 4); therefore, the variation of chemical composition of constituent minerals is due to the reactions caused by the variable P, T conditions.

High temperature prevailed in the gabbroic body; therefore the minerals of xenolith represent an assemblage which was stable at this temperature. Biotite is Mg rich and Al, Fe poor. K, Al and Fe were released from the biotite and formed orthoclase and hypersthene, respectively; Mg-biotite occurs as a residuum. At lower-grade conditions pyroxene-hornfels facies cordierite occurs together with biotite. The biotite contains more Fe, Al and less Mg. The released Mg, together with Fe and Al, formed iron-rich cordierite. In the hornblende-hornfels facies the biotite contains more Fe; therefore the cordierite is Fe-poor. The TiO_2 content of biotite decreases with the decreasing temperature. In the xenolith it contains very much, and far from it, very little, TiO_2.

Reactions in the pelitic-hornfels:

Pyroxene-hornfels facies:

a. Xenoliths: hypersthene-orthoclase

Biotite+quartz → K-feldspar+hypersthene and Mg-rich biotite residuum. (1–1.5 Kbars, 650–850 °C)

b. Immediate contact to 150 m: cordierite – K-feldspar

2biotite + 6muscovite + 15quartz = cordierite (iron-rich) + 8K-feldspar + $8H_2O$ (1–1.5 Kbars, 560–580 °C)

Hornblende-hornfels facies:

a. Cordierite - andalusite

Chlorite+muscovite+quartz → Mg-rich cordierite+Fe-rich biotite+andalusite (1–1.5 Kbars, 525 °C)

b. Andalusite-biotite (transitional)

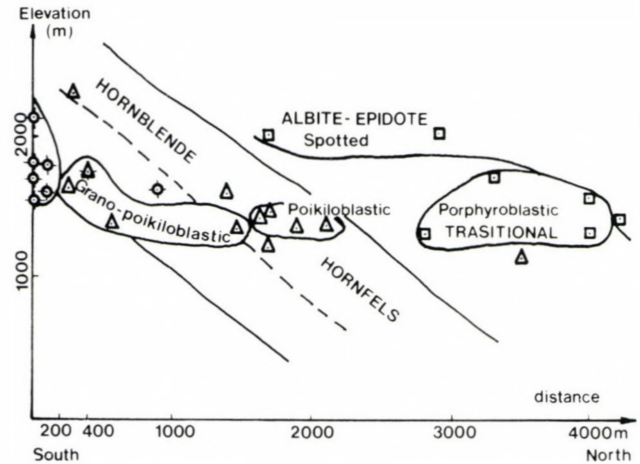
Muscovite+chlorite → andalusite+biotite (1 Kbar, 400 °C)

c. Albite-epidote-hornfels facies

Quartz+white mica+chlorite (1 Kbar, below 400 °C)

Temperature limits of facies are as follows at 1–1.5 kbars.

Pyroxene-hornfels: 850–560 °C
Hornblende-hornfels: 560–400 °C
Albite-epidote-hornfels: 400 °C



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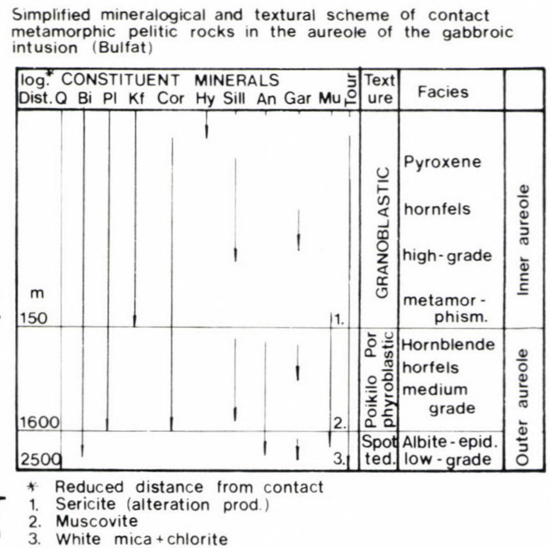


fig. 4/B

Fig. 4/A
Metamorphic facies and texture of pelitic hornfelses plotted elevation versus distance from the contact

Fig. 4/B
Simplified mineralogical and textural scheme of contact metamorphic pelitic rocks in the aureole of the gabbroic intrusion (Bulfat)

At these pressure and temperature conditions the depth of the intrusion was not more than 4 km.

B. Contact metamorphic calcareous and basic rocks

These groups of rock have a very wide compositional range; therefore, tracing the contact zones is more difficult than in the chemically uniform pelitic rocks. The two main groups which were distinguished represent two main parental-rock compositions. The first is rich in calcium- and magnesium-carbonate (parental rocks: limestone, dolomitic limestone and siliceous, dolomitic limestone) and the second one, beside calcium- and magnesium-carbonates, contains larger amount of Si, Fe, Al, (K)... etc. (parental rocks: argillaceous limestone, marl, calcareous chert, basic volcanites... etc.). Each group and subgroup contains a high-temperature mineral assemblage near the contact, and low one far from it. Apparently these rocks metamorphosed under the conditions of pyroxene-, hornblende- and albite-epidote-hornfels facies. Pyroxene-hornfels facies was not identified in the first group, due most probably to a lack of samples.

The two main groups are as follows:

1. Marble and calcsilicate marble
2. Calcsilicate hornfels and schist

1. Marble and calcsilicate marble:

In this group calcite (or sometimes dolomite) is the main constituent (70–90 Vol%). According to the megascopic appearance and amount of calcsilicate minerals four subgroups were distinguished. The subgroups with an increasing amount of calcsilicate minerals are as follows: a) marble; b) yellowish impure marble; c) brownish laminated impure marble; d) grey laminated marble.

Each subgroup was affected by regional and contact metamorphism, resulting in different mineral assemblages due to the different parental composition, although index minerals are usually the same e.g. diopside occurs in each subgroup, representing a higher temperature metamorphism compared with tremolite-bearing ones.

a. Marble: this rock is white or yellowish-white in colour, coarse-grained, without orientation near the contact, medium- to fine-grained, slightly foliated far from the contact. The prevailing mineral is calcite (90–99 Vol%), showing mosaic, mosaic-sutured textures near the contact and sutured ones in the regional metamorphic Quandil formation. Deformation-twins always occur near the contact. The twin-lamellae and crystal boundaries are well developed; far from the contact the crystals are not well separated, and become cloudy, with wavy extinction. Accessory minerals are phlogopite (muscovite), diopside, tremolite, quartz and sometimes plagioclase. The phlogopite was not observed in every sample. It forms very small prismatic or flaky crystal with slight pleochroism. Muscovite also occurs but more frequently in regional metamorphic marble; prismatic or rounded grains of diopside usually occur together with the colourless, slender prismatic tremolite ($\gamma/C=22^o–25^o$). Quartz forms rounded grains which may be detrital in origin, like the rare plagioclase.

b. Yellowish impure marble: white, creamy-yellow, sometimes greenish, medium- to fine-grained, slightly foliated; laminated where calcsilicate rich layers alternate with carbonate rich ones. The presence of scapolite or prehnite and plagioclase suggest a slight aluminium (argillaceous material) enrichment (Al_2O_3 = 0.89 wt%). Calcite is the main mineral (up to 90 vol%). It shows well developed mosaic texture at the contact, but about 2 km away from the outcrops of the contact, sutured texture (or sutured-mosaic) appears with ill-developed crystal boundaries. A later deformation sometimes bent the twin lamellae and caused wavy extinction. Diopside forms small colourless rounded grains. From the contact towards regional metamorphic rocks the extinction angle gradually increases (at the contact $\gamma/C=37^o$ and about 2 km from the contact $\gamma/C=43^o$), which suggests a slight enrichment in iron. Further from the contact (about 3.5 km) in the regional metamorphic marble, diopside is extremely rare. Phlogopite shows slight pleochroism, the larger grains sometimes have poikiloblastic texture. The prismatic (acicular) colourless tremolite ($\gamma/C=19–23^o$) is very rare. Plagioclase is rather frequent; usually it forms rounded, untwinned, small grains. Quartz does not occur at the immediate contact, but after some hundreds of meters very fine rounded quartz grains appear with wavy extinction. Prehnite occurs near the contact (up to 800 m). They are mostly secondary vein fillings. Scapolite is rather rare, it forms poikilo- or idioblastic crystals (the optical sign is negative). Titanite occurs in accessory amounts.

c. Brownish laminated impure marble: carbonate rich layers alternate with calcsilicate rich ones. Calcite prevails (up to 70 Vol%); its texture is mosaic or mosaic-sutured. Deformation twins are always present. Diopside occurs in every sample. It is colourless, mostly rounded, fine-grained, sometimes prismatic ($\gamma/C=38–44^o$). Scapolite was determined near the contact, which sometimes altered to brownish mica. The rounded, fine- to medium-grained quartz is very frequent and occurs in nearly every sample except at the contact. The acicular tremolite is very rare ($\gamma/C=19^o$). Plagioclase is not frequent, mostly untwinned and very similar to quartz. Biotite forms fine- or medium-grained crystals and near the contact they are strongly pleochroic. The rounded small yellowish grain of titanite occurs as accessory mineral. Muscovite, epidote and zoisite are very rare.

In the thinly-laminated variety two types of layers can be distinguished: 1. carbonate-rich: coarse-grained containing calcite, diopside (phlogopite), 2. calcium-silicate rich: contains diopside, quartz, titanite. The parental rock was limestone with Si impurities.

d. Grey laminated and foliated impure marble: this rock is fine- to medium-grained and laminated: carbonate-rich layers alternate with calcsilicate ones. The main constituent is calcite which near the contact is coarse-grained. The grain size decreases towards the regional metamorphic Quandil formation.

The grains are usually deformed, lensoid with wavy extinction and with bent deformation twin-lamellae, which indicate a deformation after the metamorphism. Quartz occurs in nearly every sample, mostly rounded, sometimes it shows wavy extinction and is usually very fine-grained. Diopside is not very frequent and forms small colourless rounded grains (γ/C=38–43°). The tremolite is mostly acicular, sometimes poikiloblastic (γ/C=18–22°). Many samples contain small amounts of muscovite (or white mica); they are sometimes deformed, showing wavy extinction. Scapolite, plagioclase, garnet and titanite were also determined but their occurrences are not widespread.

2. Calcsilicate hornfels and schist:

According to the megascopic appearance and their mineral constituents, two subgroups were distinguished; both were affected by regional and contact metamorphism.

The subgroups are as follows:

a. White, light-brown calcsilicate hornfels

b. Grey, greenish-grey calcsilicate and basic (volcanic) hornfels and schist.

The first group is rich in Ca (beside Si) and poor in Fe indicated by widespread presence of calcite, diopside and wollastonite. The parental rocks were mainly argillaceous limestone or calcareous chert. The second one is rich in Fe (beside Si and Ca) which occurs mainly in green hornblende and actinolite. The parental rocks were most probably basic volcanites with calcareous sediments.

a. White light-brown calcsilicate hornfels: these rocks are laminated and foliated. The dark-grey layers alternate with light-grey (sometimes brownish-white) or whitish-grey (usually wollastonite-rich) ones.

The texture is not typically hornfelsic, showing a foliation which could be the relict of the previous regional metamorphism. These rocks are usually fine-grained. The rounded colourless grains of diopside occur in every sample. The elongated wollastonite is very frequent (sometimes 80 Vol%). It forms oriented layers arranged parallel with foliation. Twinned crystals of wollastonite are common. The plagioclase is mostly rounded and untwinned. At the immediate contact they are more basic than far from the contact, where albitic composition prevails. A small amount of calcite occurs in nearly every sample. It shows a mostly mosaic texture with polysynthetic deformation twins. Xenoblastic, isotropic, colourless small grains of garnet were identified only in few samples, mostly forming layers together with plagioclase, separated from wollastonite content layers. Prehnite is rare and is always a secondary vein filling. The yellowish rounded grains of titanite are nearly always an accessory mineral. Amphibole, chlorite, apatite and scapolite are very rare.

b. Grey, greenish-grey calcsilicate and basic (volcanic) hornfels and schists: these rocks are fine-grained, grey, greenish-grey, thinly laminated and foliated. Sometimes brownish layers (biotite rich) alternate with greenish-grey (diopside, plagioclase rich) ones. The widespread presence of amphibole and quartz are characteristic. At the contact, or near to it usually green-hornblende (γ/C=12°),

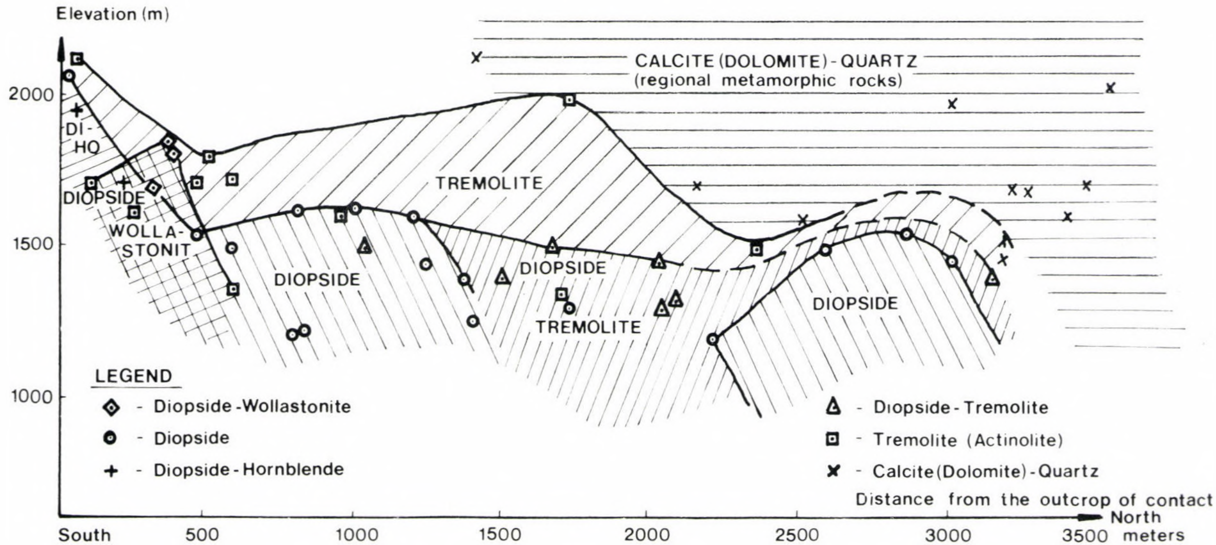


Fig. 5
Index minerals of calcareous metamorphic rocks plotted elevation versus distance from the contact

and further from it acicular actinolite and tremolite are characteristic. Quartz is frequent, showing granoblastic and sutured texture. The larger grains have wavy extinction. Small, strongly pleochroic biotite occurs in those samples which are supposed of pelitic origin, usually concentrated in layers. The occurrence of plagioclase is irregular, sometimes associated with amphibole together with diopside, sometimes only with diopside.

In some samples two plagioclase generations can be distinguished: 1. Large tabular, altered, twinned and zoned relict phenocryst, which indicates igneous origin (most probably basic volcanite). 2. Small rounded untwinned feldspar which was newly-formed as a result of the metamorphism.

The small rounded yellowish grain of titanite is a nearly ubiquitous accessory mineral. Scapolite, zoisite-epidote, chlorite, muscovite, prehnite, tourmaline, garnet and wollastonite were also found, but they are not very frequent, or else an alteration product of other minerals.

Summary

The thermal effect of a gabbro intrusion transformed the mineral assemblage of the regional-metamorphic calcareous and basic (volcanic) rocks of the Quandil formation into the minerals which were stable at the new P, T conditions. The thickness of the different metamorphic facies around the intrusive body is not well known. The high-temperature mineral assemblages, e.g. diopside-wollastonite, occur close to the gabbro. Diopside without wollastonite represents a lower temperature. It appears twice in the N–S section of the area (Fig. 5): 1. Between the diopside-wollastonite and diopside-tremolite-containing rocks (about 600–1400 m away from the contact); 2. Between the diopside-tremolite and tremolite-bearing rocks (2200–2900 m). This late occurrence indicates a thermal dome, most probably the highest position of the igneous body below surface. Tremolite-bearing rocks occur in the whole contact aureole, but close to the outcrop of the intrusion it occurs at a higher, far from it at a lower elevation. Calcite (dolomite) and quartz appear about 1.5 km away from the outcrop of the contact, at a position of high elevation. This belongs to the regional metamorphic Quandil formation.

Metamorphic reactions in the calcareous rocks:

Wollastonite:

Calcite+quartz = wollastonite + CO_2

(P=1 kbar, t=630 ^{0}C)

Diopside:

Tremolite+3calcite = dolomite+4diopside+CO_2+H_2O

(P=1 kbar, t=540 ^{0}C)

Tremolite:

5dolomite+8quartz+H_2O = tremolite+3calcite + CO_2

(P = 1 kbar, t=450 ^{0}C)

P, T conditions of the contact metamorphic facies: Pyroxene-hornfels 630 °C, hornblende-hornfels 540–450 °C, albite-epidote up to 400 °C at 1 kbar.

Summary and Conclusion

The Bulfat metamorphic-igneous thrust block is about 33 km long (NNW–SEE direction) and 12–14 km wide (Fig. 1). The western and southern borders are determined by a thrust zone; to the north the Zahrawa Valley occurs and to the east it stretches beyond into Iranian territory. The block was partly thrusted over the Walash-volcano-sedimentary sequences. The north part of the block consists of regional metamorphic rocks belong to the Quandil group. This sequence gradually passes over into the contact metamorphic rocks towards the gabbro-diorite intrusion. The NNW–SSE extension of the Quandil metamorphic group is about 8 km, the width of the contact aureole is about 2.5 km and the igneous body outcrops in a 15–19 km long and 5–7 km wide area (the outcrop of the igneous body is irregular and contains numerous xenoliths).

The Quandil metamorphic group was a volcano-sedimentary sequence containing limestone, dolomite, siliceous limestone (dolomite), pelitic rocks, calcareous and pelitic rocks mixed with basic volcanic tuffs, etc. The whole sequence contains basic (basalt), intermediate (andesite) and subordinate acidic volcanites and subvolcanites or dykes (e.g. microdiorite). This sequence was regionally metamorphosed under the conditions of greenschist or in the lower part of the sequence in the greenschist-amphibolite facies. Due to the regional metamorphism foliation developed and new metamorphic assemblages crystallized: tremolite, actinolite, albite, zoisite, chlorite, epidote, oligoclase, green hornblende, diopside and scapolite. The last four or five minerals indicate a slightly higher grade of metamorphism than the greenschist facies. The volcanites usually preserved their original texture in spite of new mineral formation.

Toward the gabbro-diorite intrusion, new mineral assemblages were formed due to the increasing temperature. The exact termination of the contact effect is not known because of the regional metamorphic rocks which have the same mineral assemblage as the contact rocks in the albite-epidote hornfels facies. The contact effect can be suspected when the first spots appear in phyllite.

The intrusion which caused the wide thermal contact aureole occupies the south-eastern part of the Bulfat block. A nearly complete plutonic differentiation series (older) and a separated basic intrusion (younger) can be distinguished (Buda 1993). The older contains ultrabasic rock (harzburgite), gabbro, diorite and nepheline syenite. The sequence was tectonized, therefore foliation is widespread. They frequently contain xenoliths. The magmatic series belong to the alkalic-calcic suite. The frequent nepheline norm, high calcium and low magnesium and iron are characteristic of the basic member of sequences. The silica deficiencies and high calcium content are due to the assimilation of calcareous rocks. The very small amount of alkali syenite most probably was formed by differentiation and contamination. The olivine gabbro-diorite intrusion is younger. The aluminium content of these rocks are high, therefore plagioclase is very frequent (leucogabbro) suggesting a well-differentiated gabbroic rock with iron rich olivine and kaersutite.

Three main types of parental rocks can be distinguished in the Quandil metamorphic group which were altered by intrusion: pelitic-, calcareous-sediments and basic volcanites. The mineral assemblage which formed during thermal metamorphism is controlled by the composition of parental rock and PT conditions. The pelitic rocks were the most uniform in composition; therefore, the new mineral formations entirely depend on PT conditions. The characteristic minerals are as follows: mica, chlorite, quartz in the albite-epidote-hornfels facies (at 1000 bars max 400 °C). Mg-cordierite, Fe-rich biotite, andalusite, plagioclase (sillimanite) in the hornblende-hornfels facies (1000 bars, 400–520 °C, 150–1600 m from the contact). Fe-cordierite, biotite, orthoclase-microperthite, andesine (sillimanite) in pyroxene-hornfels facies (1000 bars, 560–580 °C, extension of this facies is 150 m away from the contact) and a higher grade of pyroxene-hornfels facies was distinguished in xenoliths. The minerals of this rock are as follows: orthoclase, hypersthene, Mg-rich biotite and andesine (1000 bars, 650–850 °C). The pelitic rocks alternate with calcareous ones; therefore, pressure and temperature conditions were probably the same during thermal metamorphism, since the mineral assemblages which formed under these conditions show differences due to the different chemical compositions. The most important temperature and pressure indicator minerals are tremolite (formed in the albite-epidote and hornblende-hornfels facies), diopside and tremolite (hornblende-hornfels facies: 450–540 °C, 270–1400 m from the contact) wollastonite, diopside, garnet (pyroxene-hornfels facies, 540–630 °C, 270 m from the contact), periclase, melilite, spinel, forsterite and wollastonite (high temperature pyroxene-hornfels facies: 540–820 °C, xenolith in gabbro).

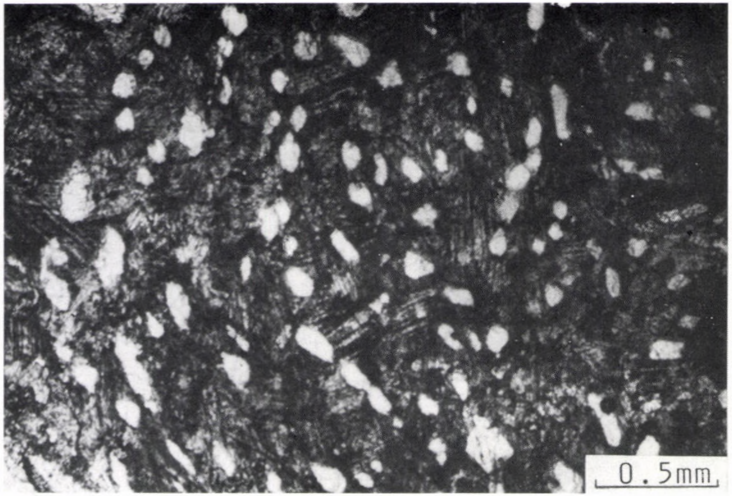
The basic parental rocks were probably volcanics (zoned plagioclase) or tuffs (laminated rocks: layers alternate with calcareous or pelitic ones). Characteristic minerals are: actinolite, zoisite, albite (albite-epidote facies), green hornblende, plagioclase, epidote (hornblende-hornfels facies), clinopyroxene, basicplagioclase (anthophyllite) pyroxene-hornfels (xenoliths in the intrusion, Table 5).

The igneous-metamorphic block was thrusted on the Walash volcano-sedimentary sequence from NE to SW. The movements were aided by serpentinite acting as a basal lubricant. Typical sequence of the thrust zone is serpentinite, amphibolite, pyroxene amphibole diorite (igneous body), andesite tuff, augite-andesite, andesite-tuff (Wallash formation) from the top to the bottom. The Wallash volcano-sedimentary sequence is certainly younger than the Bulfat sedimentary-igneous block. The composition of volcanic rocks are basic and intermediate. The basic rocks are mostly dykes and submarine lava flows. The intermediate rocks (andesite) form volcanic cones according to Buday (1973). The whole sequence shows the characters of the volcanic rocks forming in an island arc setting.

Plate I

1. Dolomite exsolution in ophicalcite, +N, stained with alizarin-red; 2. Hypidioblastic forsterite in calcite, +N; 3. Brucite with onion-skin texture in predazzite, +N; 4. Granoblastic cordierite

Plate I



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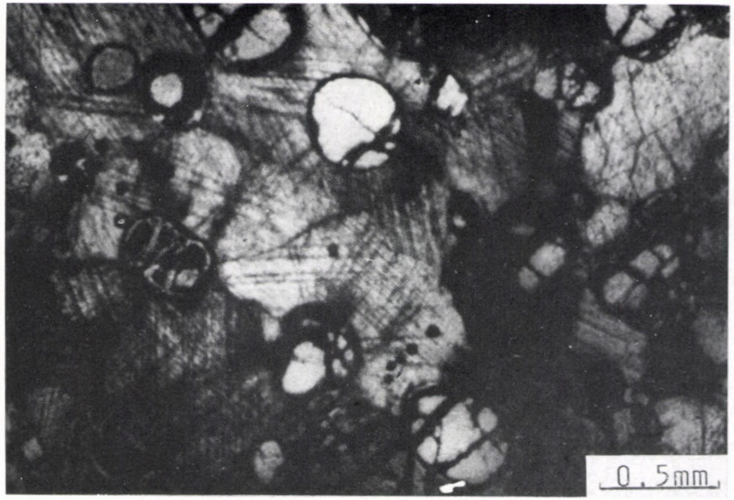

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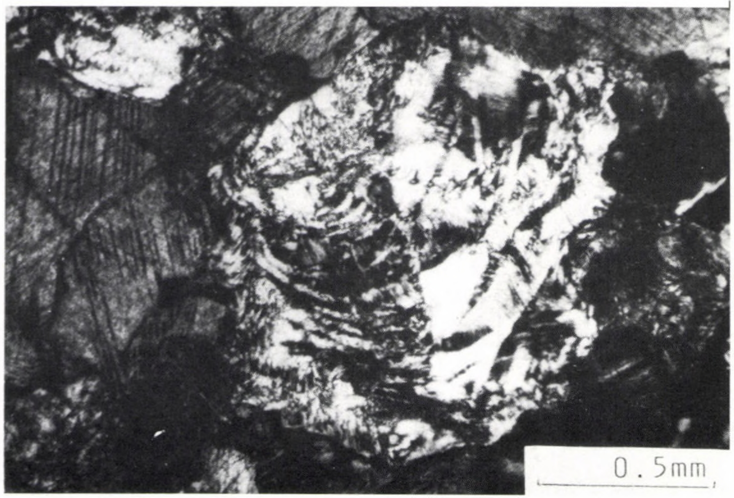

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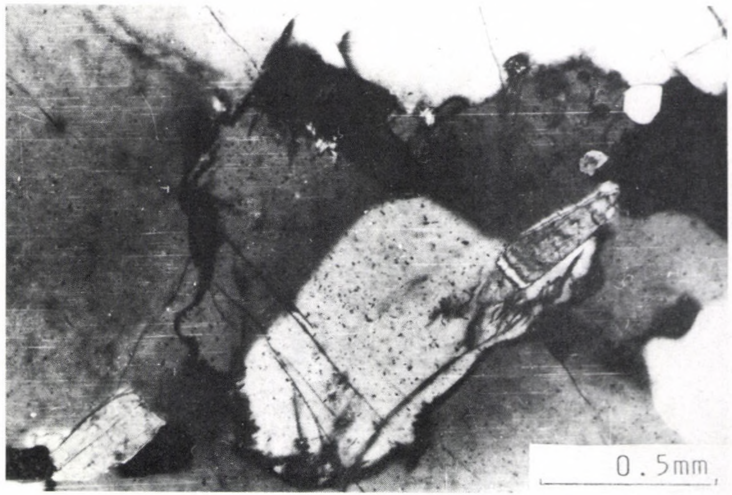

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Plate II

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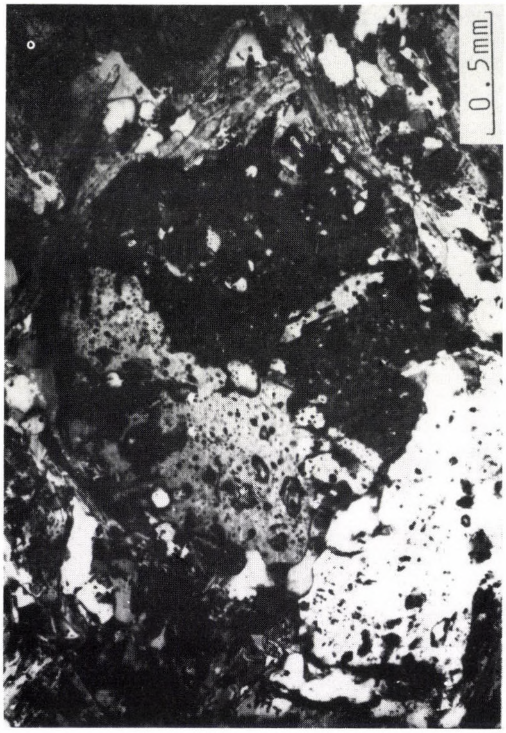

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Plate II

1. Transformation of biotite to sillimanite and K-feldspar, +N; 2. Poikiloblastic cordierite, +N; 3. Porphyro-poikiloblastic andalusite, +N; 4. Deformed porhyroblastic andalusite, +N

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Hydrogeology and oil deposits at Pechelbronn–Soultz–Upper Rhine Graben

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The Upper Rhine Graben at the Pechelbronn–Soultz area is a typical intermontane rift basin which has been subsiding continuously for the last 40 to 50 Ma. The ± 2 km thick sedimentary fill contains areally extensive aquifers and aquitards of Mesozoic and Tertiary age. Oil occurs in Triassic carbonate and Tertiary clastic reservoirs commonly in association with dense faulting and fracturing. Formation waters move from recharge areas in the adjacent mountains toward the valley and discharge preferentially through intensively fractured fault blocks by cross-formational ascension. Their ^{14}C age is 31 000 years or less and they have meteoric type stable-isotope composition. The hydraulic theory of petroleum migration integrates all the observed aspects of the area's petroleum-hydrogeological conditions, including the location of oil fields, vertical distribution of oil types, deep and near-surface temperature anomalies, artesian fluid levels, highly mineralized formation waters and oil seeps, into one coherent set of natural processes and phenomena of a common genetic origin.

Key words: Intermontane basins, regional groundwater flow, petroleum hydrogeology, field manifestation, exploration

Introduction

The paper presents the principal observations and conclusions of a petroleum hydrogeological study conducted in the Upper Rhine Graben at Pechelbronn in NE France recently. The main objectives of the study were to test the validity of the hydraulic theory of petroleum migration and to evaluate its applicability to exploration in intermontane environments. To these ends, basinal groundwater flow was investigated in relation to oil deposits from two different aspects, namely: 1. the possible role that aqueous advective transport may have played in the migration and entrapment of petroleum and 2. some natural phenomena, due or attributable to regional groundwater flow, as possible indicators of actual or potential deposits of hydrocarbons.

The area was selected for its favourable attributes with respect to the project's objectives, including: a simple and well-defined intermontane topography; a relatively simple history of basin evolution; well-known conditions and distribution of hydrocarbon fields; and an uncommonly extensive and accessible, although somewhat outmoded, data base.

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Location and geologic framework

At a surface elevation of 150 to 220 m, the Pechelbronn–Soultz petroliferous basin is located approximately 40 km north of the city of Strasbourg, France, between the 400 to 500 m high plateau of the Vosges, 10 to 15 km to the north-west, and the 900 to 1000 m high Black Forest mountains, 35 km to the south-east (Fig. 1). The two mountain ranges constitute NNE–SSW striking, nearly parallel shoulders of strongly asymmetric altitude of the Alsatian portion of the Upper Rhine Graben. They are detached from the sedimentary basin by the Vosges- and the Black Forest-Rift Faults.

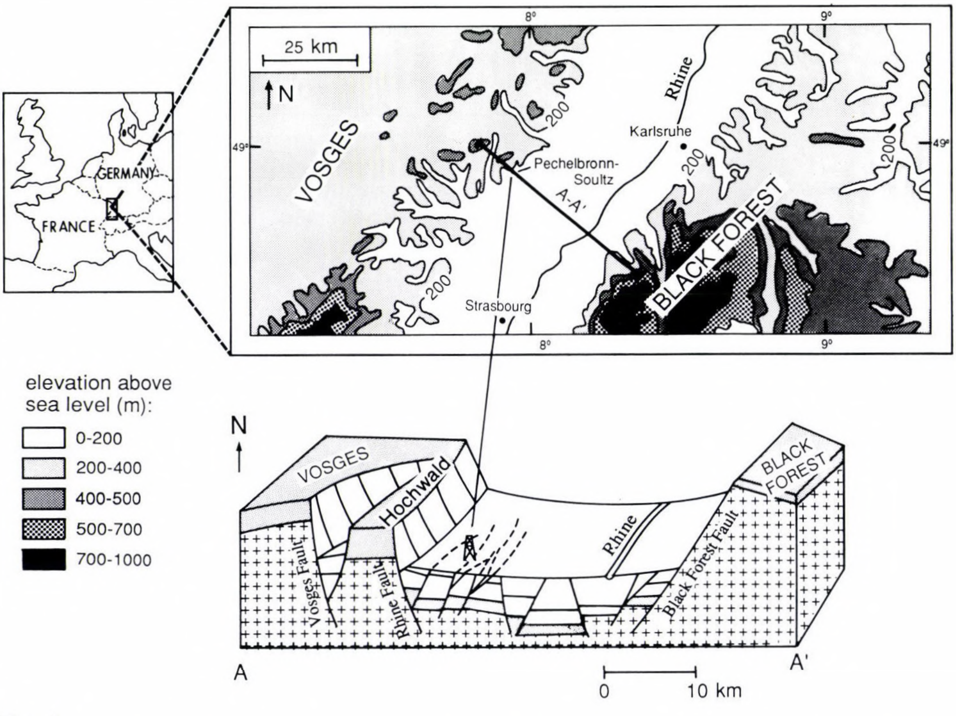

Fig. 1
Location, topography and schematized tectonic structure of the Upper Rhine Graben at the Pechelbronn–Soultz area

In the area the sediments are approximately 2000 m thick. They rest on a pre-Permian granitic basement and comprise, in ascending order (Fig. 2) about 500 m of Triassic sandstones (Buntsandstein), 200 m of fractured lime- and dolostones (Muschelkalk), and 195 m of variegated marls (Keuper); 145 m of Jurassic limestones and marls with organic rich levels and a few meters of arenitic limestones in the upper parts; 100 to 350 m of Tertiary dolomitic and calcareous marls of Late Eocene age, 250 to 400 m of organic rich, lenticular, lagoonal-marine Early Oligocene clastics, and 0 to 20 m of Quaternary loess

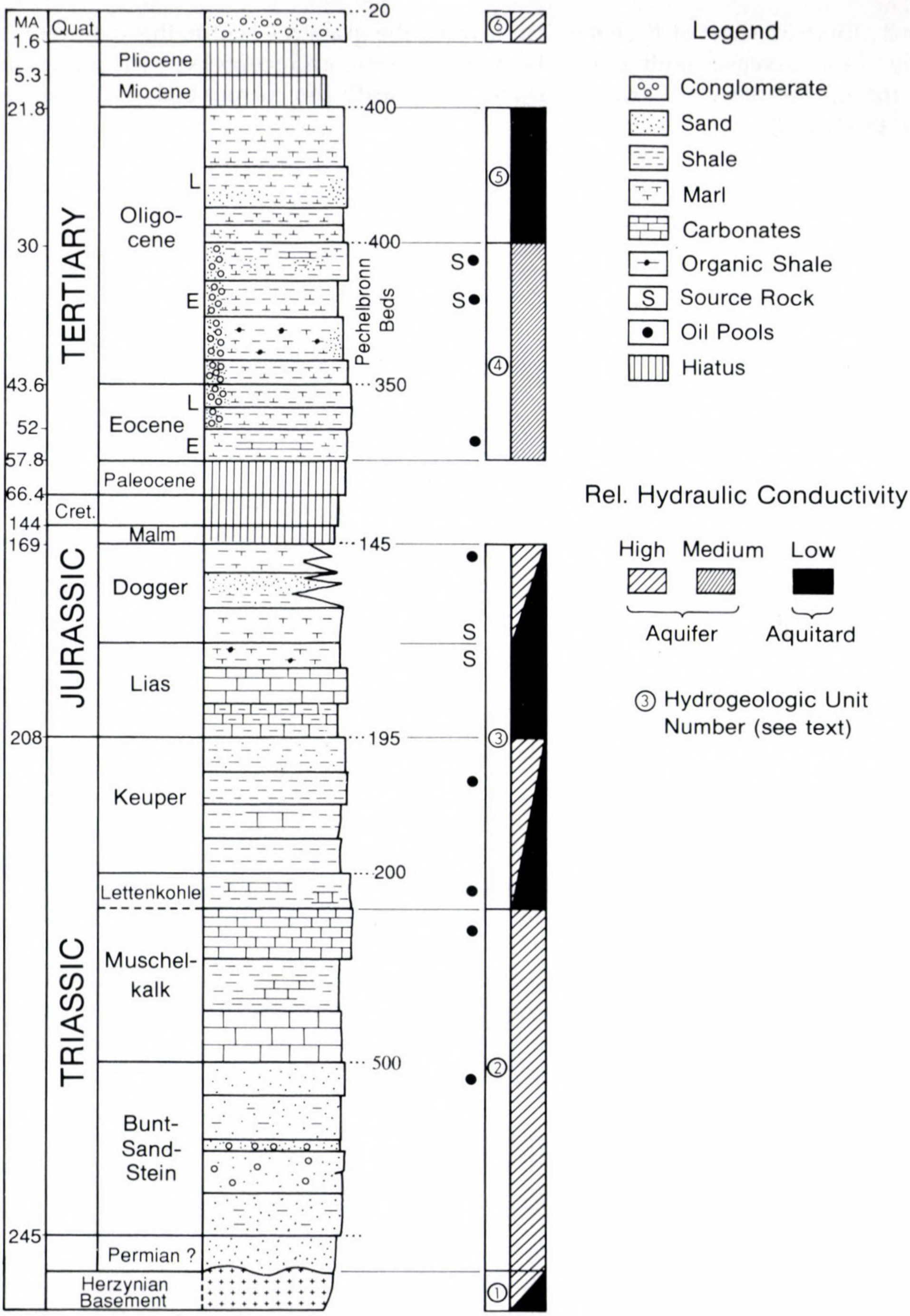

Fig. 2
Lithostratigraphy and hydrogeologic units in the study area

and alluvium. East of Pechelbronn, toward the graben's centre, the sediments' thickness increases both gradually, by deposition, and stepwise, by faulting. It is thought to reach a maximum of 2500 m beneath the current bed of the Rhine River (Fig. 3).

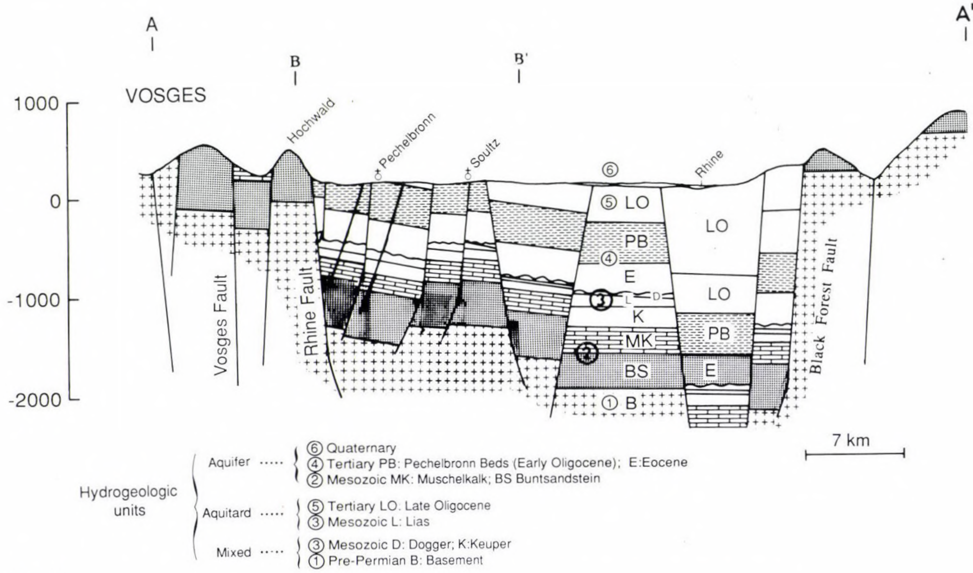

Fig. 3
Schematic representation of regional tectonic setting and hydrostratigraphy

The evolution and history of the graben has been dominated by continuous subsidence which started in the Early Eocene and is still ongoing. The subsidence resulted from northwest–southeast rifting which, subsequent to the initial break-up of the crust in the Early Eocene, had two main rifting phases, namely, one in Late Eocene–Early Oligocene and one, in mid-Pliocene times (Illies 1977).

During these rifting phases old faults were rejuvenated and new ones were generated, giving rise to a complex structure of synthetic and antithetic block faulting. The fault blocks themselves became dissected with swarms of secondary and younger generation faults the density of which may significantly vary from one block to another.

The start of subsidence in Early Eocene was accompanied by a major surge of geothermal heat marked by a high maturity level of organic matter in the Jurassic source rocks as indicated by the Cretaceous non-depositional hiatus (sub-Eocene unconformity). These source rocks had not subsided to the depth required for maturation before initial rifting started. The main rifting phases had their associated events of anomalously increased terrestrial heat flow, with the last event maintaining an above average general heat flux in the Upper Rhine Graben to date.

Petroleum geology

Attention was directed to the presence of oil deposits in this area by oil seeps a long time ago. The town's name, "Pechelbronn", is interpreted to mean "pitch spring". An oil seep, described in 1498, produced a mixture of oil and saline water just west of Pechelbronn. The oil was used by local farmers as lamp fuel and lubricant. In 1735, bituminous sands were discovered, marking the start of industrial oil production in the Pechelbronn–Soultz Basin. During 1888–1916, some 2850 oil wells were drilled, mainly into the Tertiary formations.

The two World Wars have increased the demand for fuel in the years 1917–1953, and prompted the reactivation of mining, the drilling of an additional ± 2000 wells and the discoveries of deeper deposits in the Mesozoic strata. Exploration and exploitation came to a virtual stop in the decade 1953–1963, except for an enhanced recovery project initiated in 1963.

With the drilling for Mesozoic oil, warm and mineralized artesian waters also were found in the Triassic formations which now provide water for two health spas and the heat source for a contemplated geothermal power plant.

Although important at times and locally, the total amount of approximately 3.3 million tons of oil which was produced in the Pechelbronn area over the span of two centuries is small by current global standards. Nevertheless, the geology and, in particular, the hydrogeology of the basin provide instructive examples and potential keys to exploration for petroleum deposits in intermontane graben environments.

The deepest known oil deposits in the area are in the Buntsandstein (Late Triassic) and Muschelkalk (Middle Triassic) and are associated with the basement horst at Soultz (Figs 2, 4 and 5). Both of these formations are major regional aquifers with porosities of 2 to 10% and 20 to 30%, respectively. The few available permeability values suggest a range of 2 to 175 md in the Buntsandstein and 14 to 40 md in the Muschelkalk. Permeabilities can, however, increase considerably due to intensive faulting and fracturing locally. The oil is trapped primarily in fault zones capped by the overlying gypsiferous Keuper. However, the Keuper itself may form reservoirs outside, but in the near vicinity of, the study area.

The Jurassic is only known to contain exploitable deposits slightly (18 km) to the south of the study area proper. The oil is reservoired here in structural and stratigraphic traps.

In the Tertiary, the Eocene contains only minor deposits. The most important oil accumulations in the entire study area are found in sandstone and conglomerate lenses of alluvial fans, floodplains and shallow marine clastics of the so-called "Pechelbronn Beds" of Oligocene age. These are generally rich in organic matter and liquid hydrocarbons. Producible oil is concentrated in the lenticular reservoirs, with noticeably increased permeabilities at locations where the lenses are dissected by subvertical faults and fractures.

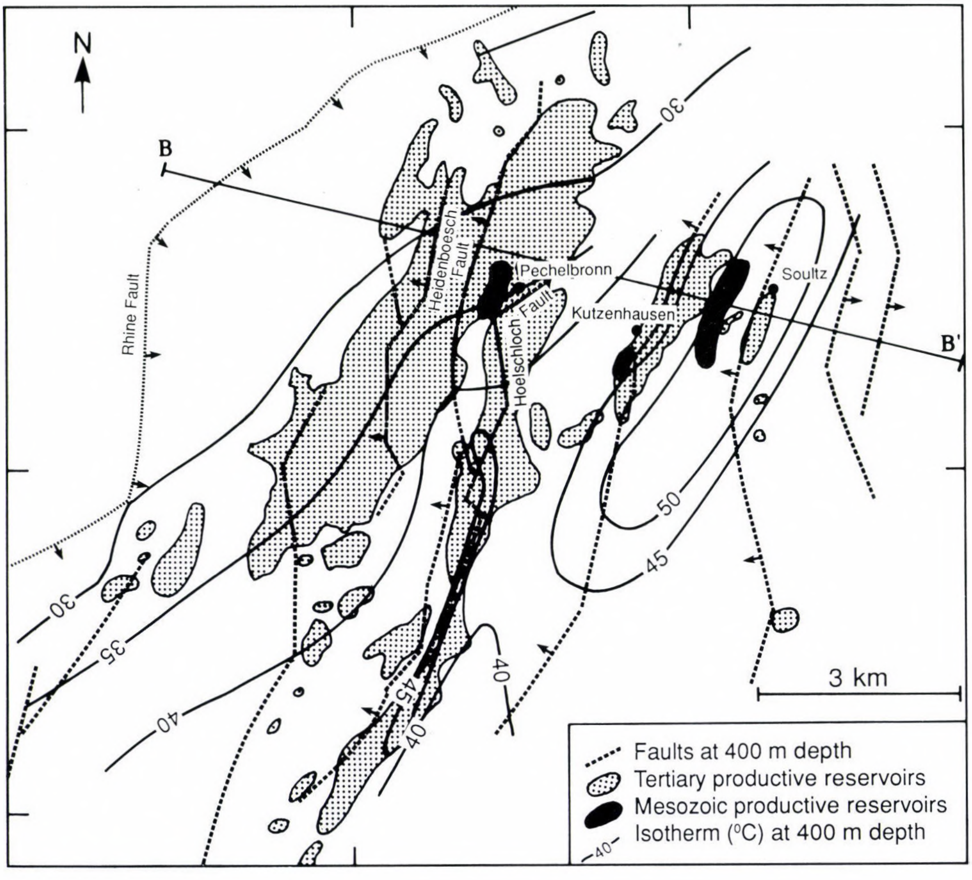

Fig. 4
Increasing subsurface temperatures from west to east with an anomalous high at Soultz and fault-controlled oil pools (after Schnaebele 1948 and Levi 1962)

In general, the geochemical type of the oil changes from paraffinic in the deeper and older (Triassic–Jurassic) reservoirs, to mixed paraffinic–naphthenic or, dominantly naphthenic, in the Oligocene strata.

Two rock groups have been identified as potential sources for the Pechelbronn oil deposits, namely: 1. the bituminous claystones, shales and siltstones of the Jurassic upper Lias and lower Dogger and, 2. the Early Oligocene dark brown foliated carbonaceous shales, with intercalations of dolomitic shales, sandstones and rarely, carbonates (Fig. 2).

In general, the 0.5% vitrinite threshold value indicating the beginning of kerogen catagenesis is reached in northern Alsace at about 800–900 m depth (Robert 1988). Yet, the Jurassic source rocks in the vicinity of Soultz reach maturity already at depths of 650 to 700 m due presumably to the anomalously high (80–100 °C/km) geothermal gradient which is reflected also by a higher

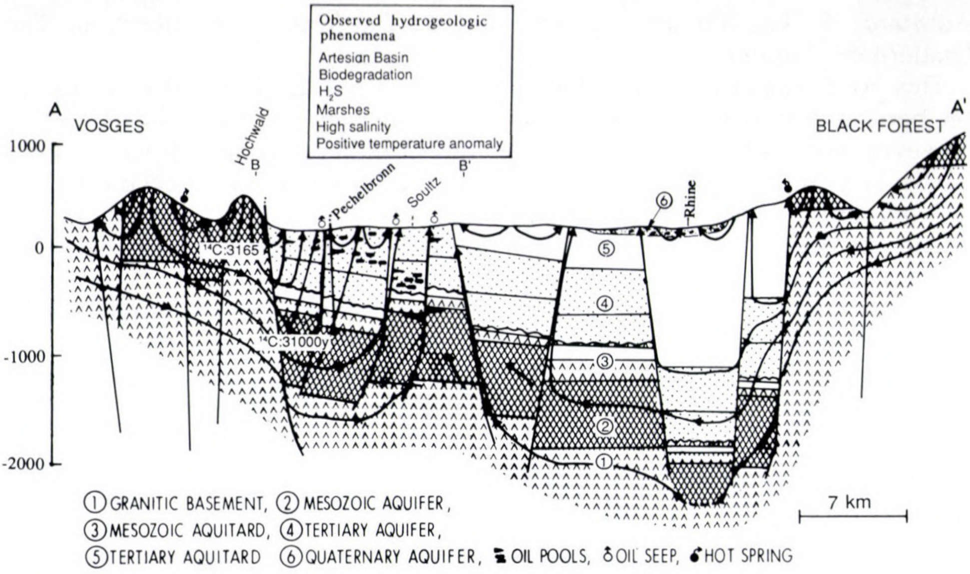

Fig. 5
Schematic representation of regional groundwater flow between the graben's shoulders, with observed phenomena and ^{14}C values

than normal gradient of vitrinite reflectance. The Oligocene source rocks are immature, on the other hand, in the Pechelbronn–Soultz basin, with vitrinite reflectance ranging between 0.2 and 0.4% (Robert 1988), but they have reached maturity toward the graben's centre, i.e., at greater depths with values of R_o over 0.5%. Gerling (1988) concluded that oil deposits in the Pechelbronn–Soultz Basin originated from Jurassic source rocks only.

Hydrogeology

The hydrogeological conditions found in the study area are similar to those often observed in intermontane basins with a mature, i.e., non-compacting, sedimentary fill. The common genetic factor that is manifest by these conditions appears to be regional groundwater flow. Groundwaters move in gravity-driven cross-formational systems from regional recharge areas in the adjacent Vosges and Black Forest mountains to discharge areas in the lower portions of the graben floor. Local basins in the graben, such as the Pechelbronn–Soultz basin, generate local flow systems which are superimposed on, and merge with, the regional systems (Fig. 5).

The flow pattern is modified in detail by the basin's hydrostratigraphy. The main hydrostratigraphic units, in ascending order, are (Figs 2, 3): 1. The pre-Permian granitic basement; 2. The Mesozoic Aquifer; 3. The Mesozoic

Aquitard; 4. The Tertiary Aquifer; and 5. The Tertiary Aquitard. 6. The Quaternary Aquifer.

The pre-Permian granitic basement is considered an aquitard constituting the bottom of the basin. It is somewhat permeable and functions as an aquifer, however, both beneath the mountains and in the basement's upper sections, particulary in the horsts, as suggested by intensive weathering, fracturing and faulting, and indicated by producing water wells.

The Mesozoic Aquifer comprises the Buntsandstein and Muschelkalk (Figs 2 and 3). Together, these two formations constitute a regionally continuous, basal aquifer and provide a ready pathway for meteoric water from the flanking mountains toward the graben's centre beneath the younger sediments.

The Mesozoic Aquitard is formed by rocks chiefly of Late Triassic and Early Jurassic (Keuper and Lias) age. Although these rocks represent a regionally extensive unit of low hydraulic conductivity, the Keuper carbonates are known to be relatively highly permeable locally. The overlying Dogger is absent in the Soultz region. Where it is present further south, it is composed of marl and shale sequences with intercalations of sandstone and its permeability is slightly increased.

The Tertiary Aquifer comprises the evaporites, carbonates and the lenticular and bituminous siltstones and shales of the Eocene and lower Oligocene, including the petroliferous Pechelbronn Beds. The matrix permeability of these rocks is low with the exception of the lenticular bodies of sandstones and conglomerates. Nevertheless, based on their lithologies, the reported flow rates of water and oil to wells, galleries and shafts, the bulk permeability of these beds combined appears to be superior to that of the under- and overlying strata, and therefore to qualify them as an aquifer, albeit of low productivity.

The Tertiary Aquitard consists of the shales, calcareous sandstones and variegated claystones of Late Oligocene age. This aquitard is largely absent from the western part of the valley and the contrast between its permeability and that of the subjacent aquifer is not sharp. Its hydraulic character as an aquitard, eventhough leaky, is therefore not firmly established.

The Quaternary Aquifer consists of loess and alluvial sands and gravels. It is absent from the study area proper and is found only to the south and east of it.

The hydraulic role of the fault zones and fractures cannot be evaluated with respect to local details. Numerous observations of intensive and increased water flows from fault zones penetrated by bore holes suggest high permeabilities at present. Mineralization and hydrocarbon veins in faults and fractures attest to fluid flow at some earlier time. Dry sections, on the other hand, are suggestive of a role as hydraulic seals or barriers for some of the faults, at least locally. It is probable that the hydraulic character of these lithologic discontinuities varies both with time and space: any given fault may be a seal at one place and a conduit at another at the same time, or at any given location in different geologic times. A schematic distribution of the hydrostratigraphic units is given in Fig. 3, with their assumed hydraulic character shown in Fig. 2.

The regional-scale basinal flow pattern (Fig. 5) in the Upper Rhine Graben is inferred from original pressure measurements in bore holes, detailed hydrogeological studies in nearby areas (e.g., internal reports from the Geological Survey of Alsace and Lorraine, SGAL), and numerical modelling (Garven and Person 1987). Potentiometric surface elevations (Fig. 6) derived from formation-pressure measurements show that, on a regional scale, hydraulic heads decline toward the valley's centre and with decreasing depth of measurement below the land surface: flow in the valley is, therefore, toward the river and ascending. Indeed, water levels rise to topographic elevations of 200 m and higher, i.e., often above the valley's floor, from measurement points at –250 m and lower. On the other hand, a detailed hydrogeological study (Simmler 1972) to the southwest and west of the study area proper shows those parts of the Vosges conclusively to act as regions of recharge.

The two principal flow systems generated by the two major highlands, as well as several intermediate and local systems associated with major valleys in the mountains and depressions in the plain are seen in Fig. 5.

The flow distribution which appears relatively uniform at the regional scale becomes complex at a local scale, owing to the heterogeneities in the rock's permeability due particularly to stratification, faulting, and lensing. In detail, the flow field in the Pechelbronn–Soultz area (Fig. 7) has the following salient features: a) flow systems generated by the local topographic relief may reach depths of several hundred meters but probably do not penetrate the basal Oligocene; b) the vertically ascending flow of groundwater is concentrated largely through the faults and fault zones which thus constitute the main pathways for cross-formational fluid flow; c) owing to the discontinuously sealing nature of faults and fault zones, rising fluids may be deflected and forced into highly permeable sections (e.g., lenses) of strata adjacent to the fault planes.

The flow concentrating effect of fault zones appears to be reflected by the observed distribution of subsurface temperatures (Figs 4 and 8). A pronounced positive temperature anomaly has been known in the Soultz area since the early days of exploration and reported, e.g., by Haas and Hoffmann (1929) and Schnaebele (1948), Figs 4 and 8. The area between the Kutzenhausen and Soultz Faults is characterized by intensive faulting, probably related to the "Soultz Horst", and by high water yields, usually with artesian levels, obtained from faults. Based on original temperature measurements, the anomaly appears even more accentuated than it was reported by Schnaebele (op. cit.). Another area, called here the Pechelbronn area, of relatively intensive faulting is located between the Heidenboesch and Hoelschloch Faults (Figs 4 and 8). A temperature anomaly is not as well documented here as in the Soultz area, but the presence of at least one known thermal artesian well (called Les Hélions II) suggests that here, too, above regionally common temperatures are associated with ascending groundwaters. Also, on detailed examination, the isotherm contours

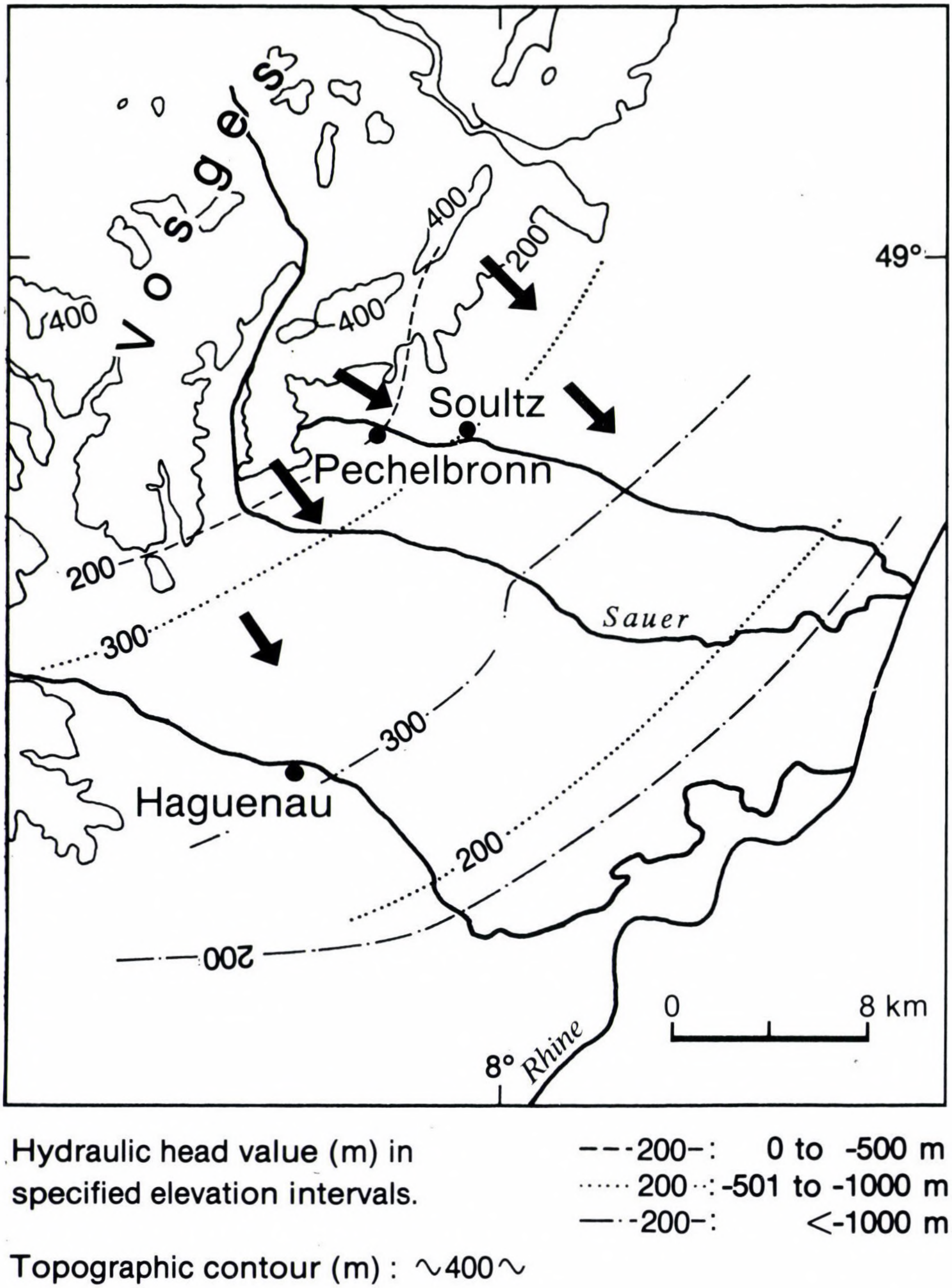

Fig. 6
Dominant hydraulic head values obtained at points of measurements within specified elevation ranges

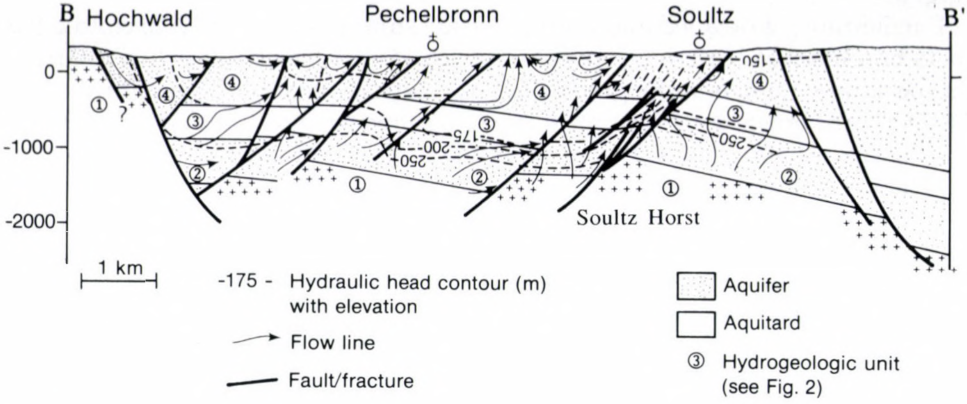

Fig. 7
Local groundwater flow pattern inferred from hydraulic head values, Pechelbronn–Soultz Basin

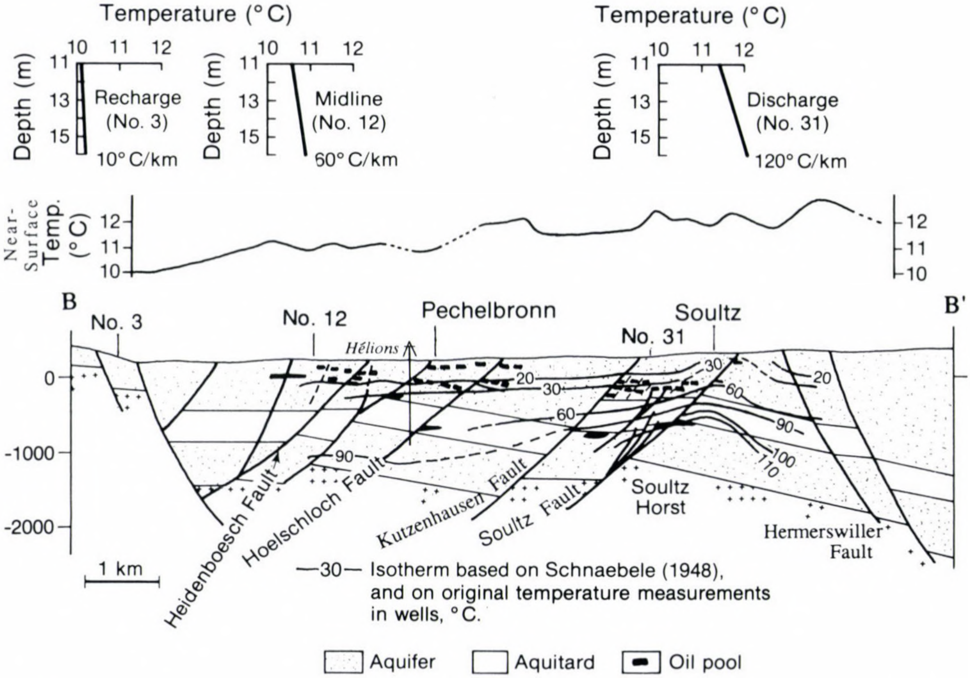

Fig. 8
Near-surface and subsurface temperatures, and oil deposits in the Pechelbronn–Soultz Basin

show distinct stepwise changes in the direction and the proximity of faults (Fig. 8).

Furthermore, a detailed and high-precision survey (Plöthner 1988) conducted in 48 boreholes of 16 m depth revealed a clear correlation between near-surface temperatures, temperature gradients and vertical sense of water flow (Tóth 1987). Near-surface temperatures increase basinward and show local anomalies at outcrops of fault planes (Fig. 8). Also, the gradients of temperature calculated for the depth interval of 11 to 16 m, increase from recharge, to discharge areas as found, e.g., in holes No. 3, 12 and 31, to be 10, 60 and 120 °C/km, respectively (Fig. 8). In summary, both the magnitudes of the overall anomaly and the observed details of the temperature distribution are strongly suggestive of groundwater flow being focussed and intensified by funelling through fault zones, as well as having an observably modifying effect on both the deep and near-surface distribution of geothermal heat.

The mineralization of formation waters exceeds 100 000 mg/l in the deeper faulted zones of the Soultz region, which represents a major salinity anomaly (Fig. 9). This anomaly appears to occur as the final section of a gradually increasing trend from west to east, i.e., in the general direction of the Vosges regional flow system. Again, only a slight increase in mineralization appears to be associated with the Pechelbronn Faults. This situation is consistent with the location of the area located "upstream" from Soultz, i.e., closer to groundwater recharge (Otto and Tóth 1988).

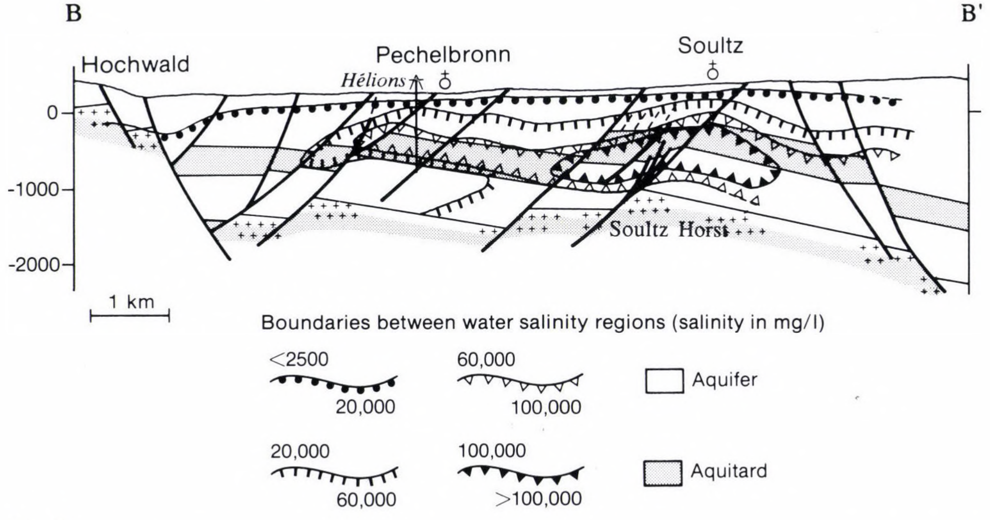

Fig. 9
General increase in formation-water salinity with groundwater flow direction, modified by the channelling-effect of fault zones in the Pechelbronn–Soultz Basin (after Otto 1992)

In addition to the above outlined hydraulic, chemical and thermal conditions, data obtained for the present purposes on ^{14}C, δ^2H and $\delta^{18}O$ provide independent confirmation of the geologically recent age and meteoric origin of the formation waters found even at the depths of, and below, the known oil accumulations.

As the interpretation of these data can best be presented in the context of the complete "Rhine Graben Flow Profile", relevant data have been plotted in the previously introduced Fig. 5. As it can be seen in this Figure, a ^{14}C analysis shows an age of 3165 years for water obtained from a well tapping the shallow "Hochwald Local Flow System". However, radiocarbon ages of 31 000 years were obtained for waters from the Helions artesian thermal well and 26 900 years from the Morsbronn thermal well (not in this section), both completed in the Muschelkalk–Buntsandstein aquifer in the "Vosges Regional Flow System". Assuming that possible distances to the two recharge areas range from 1 to 3 km and, respectively, 10 to 30 km, average groundwater flow velocities in both systems are 0.3 to 1 m/a. These velocities are common for regional groundwater flow and, at the same time result in relatively short throughflow times, i.e., a high rate of groundwater turnover in the basin.

This conclusion is further supported by stable isotope compositions determined for several water samples obtained from wells in the area. As can be seen in Fig. 10, these well-water samples display a close similarity to, and thus probably indicate a genetic kinship with, the δ^2H and $\delta^{18}O$ composition of modern meteoric waters.

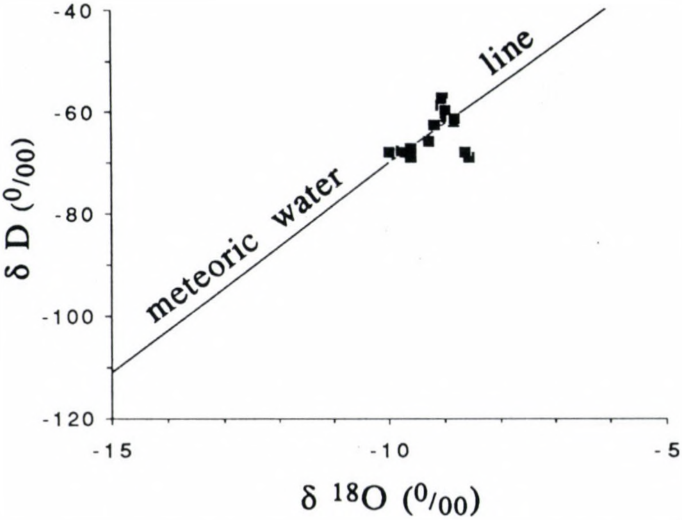

Fig. 10
Stable isotope composition of formation waters sampled from hydrogeologic units 2, 3 and 4 in the Pechelbronn–Soultz area (after Otto 1992)

Petroleum hydrogeological synthesis: ramifications for exploration

The most significant petroleum hydrogeological phenomenon in the study area is the spatial coincidence of oil deposits with locally intensified cross-formational discharge of geologically young groundwaters of meteoric origin (Figs 7 and 8). This phenomenon is suggestive of a genetic role that topography-induced groundwater flow-systems can play in the accumulation of hydrocarbons in intermontane basins. In addition, it also provides various keys to hydrogeologically based exploration techniques that are particulary applicable in these environments. In brief, it corroborates the basic thesis as well as the predicted practical ramifications of the hydraulic theory of petroleum migration, the validation of which was one of the principal objectives of this study.

According to the hydraulic theory (Tóth 1980), geologically mature basins are hydraulically continuous environments in which the relief of the water table, commonly a subdued replica of the land surface, generates interdependent systems of groundwater flow with patterns modified by permeability differences. In these systems, meteoric waters infiltrate and descend in upland recharge areas, migrate laterally under regions of medium elevations, and are discharged in topographic depressions. Where flow systems meet or part, relatively stagnant zones develop and flow directions change abruptly.

Under these conditions, gravity-induced cross-formational flow is the principal agent in the transport and accumulation of hydrocarbons. The mechanism becomes operative after compaction of sediments and the concomitant primary migration cease, and subaerial topographic relief develops. Hydrocarbons from source or carrier beds are then moved along well-defined migration paths toward discharge foci of converging flow systems, and may accumulate en route in hydraulic or hydrodynamic traps. Accordingly, deposits are expected to be associated preferentially with ascending limbs and stagnant zones of flow systems. Hence they are characterized by relative potentiometric minima, downward increase in hydraulic heads possibly reaching artesian conditions, reduced or zero lateral hydraulic gradients, and relatively high groundwater salinity. Continuous flow of meteoric waters imports hydrocarbons into traps until the trap capacity is reached. The excess becomes source material for new accumulations.

The theory predicts also a genetic relation and spatial correlation between oil deposits, on the one hand, and positive geothermal anomalies, chemically reduced chimneys, sulfide ore formation, and marshy and saline surface conditions, on the other. In Fig. 11 a graphical summary of the hydraulic theory is presented in a slightly modified form of the original version (Tóth 1980; Fig. 44). As a comparison, the petroleum hydrogeological conditions as observed in the Upper Rhine Graben in general, and the Pechelbronn–Soultz area in particular, are shown in Fig. 5; a convincing similarity between the theoretical

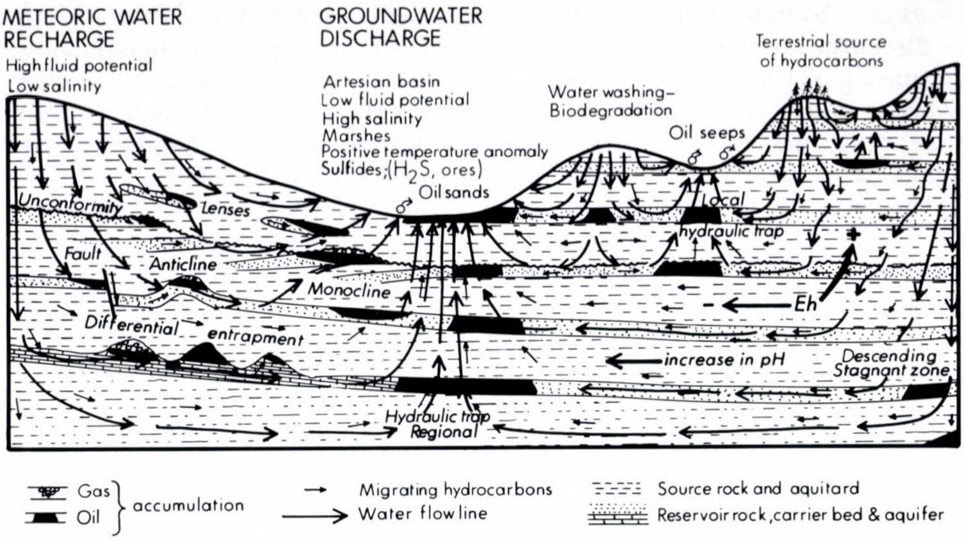

Fig. 11
Graphical summary of the hydraulic theory of petroleum migration and associated natural phenomena (modified after Tóth 1980)

and the observed conditions is evident. It must therefore be concluded, that the hydraulic theory offers a plausible explanation for the genesis of the oil deposits and associated hydrogeological phenomena in the study area and also, in general, in other geologically mature intermontane basins.

Specifically, the relations between petroleum accumulations and hydrogeological processes and phenomena in the study area are envisaged to have developed as follows:

Potential source rocks were deposited in a eugeosynclinal environment during Jurassic times (150–200 Ma). They matured in the Oligocene, ± 20 Ma after the central portion of the basin started to subside in Early Eocene at ± 50 Ma, due to rifting. Ever since rifting began, groundwater has been moving from the graben's shoulders toward its centre, transporting available hydrocarbons in molecular solutions from overlying source rocks partly in the regional, partly in local recharge areas. The same flow systems also transport geothermal heat, which is generally high in the area due to the thinning of the crust. The transported heat accumulates in regions where the laterally moving groundwater is forced to turn upward: primarily through the more heavily fractured fault blocks and along highly conductive major faults. Similary, dissolved mineral matter is transported basinward and accumulates in, and in the vicinity of, areas of intensive discharge.

The amount of water that has been available for the transport of hydrocarbons to date may be estimated in terms of basinal pore volumes. To this end, the time elapsed since organic matter maturation (± 30 Ma) is compared with the average subsurface residence time of the water during its travel from recharge to discharge areas (i.e., the duration of one full cycle of pore-volume flushing). Assuming 30 000 years for average residence time, a "water exchange number" (Kartsev and Vagin 1964) of $N = (30 \times 10^6)/(30 \times 10^3) = 10^3$ is obtained, indicating that approximately 1000 times the basinal rock pore volume of water has been available for hydrocarbon transport in the area since maturation.

The primary area of collection and deposition of the migrated hydrocarbons can be expected to coincide with the region where the oppositely directed regional flow systems from the Vosges and Black Forest meet, i.e., exactly in the Pechelbronn–Soultz area (Fig. 5). In this general discharge area water flow is turned upward and its fluxes increase sharply along those fault planes which are highly permeable, as well as through the intensively fractured fault blocks. Nevertheless, part of the ascending formation fluids will still be forced to flow through the intergranular pores of unfractured rocks. Several entrapment mechanisms of different nature are co-effective under these circumstances to disaccomodate transported hydrocarbons from the carrier waters. At places where they are locally sealed or constricted, fault planes themselves may act as traps, or deflect feed-stock fluids into highly permeable sandstone lenses that they dissect. In both cases, capillary pressure differences and mechanical filtering at, and due to, grain-size boundaries will tend to retain the oil and pass the water. Indeed, most oil fields in the area are intimately associated with faults and adjacent sand lenses particularly in the Pechelbronn Beds, i.e., in immature Tertiary rocks (Fig. 5). Fault controlled traps are found in the Muschelkalk and Keuper also, where low permeability fault planes and cap rocks retain the hydrocarbons from the moving groundwaters.

Groundwater flow-related pressure, thermal and chemical conditions greatly enhance the effectiveness of the lithologic trapping potential in discharge areas. The sudden and rapid decrease in pressure and temperature owing to the upward turn of flow paths in discharging systems causes exsolution of hydrocarbons (Roberts 1980). These effects may be reflected also by the observed vertical distribution of oil types in the study area, namely, by the occurrence of paraffinic based oils at lower levels and mixed paraffinic-naphthenic based ones at higher levels (Bienner and Louis 1953). The distribution is well explained by the "migration-fractionation" model proposed by Illich et al. (1981). It requires, in the direction of flow, "...enrichment of hydrocarbons of intermediate solubility, partial exclusion of hydrocarbons of low solubility, and retention in solution of the more soluble hydrocarbons". The model invokes "...cross-stratigraphic movement of oils in terms of water movement rather than movement of hydrocarbons". Mixing of the paraffinic and naphthenic oils at shallower depths in the study area is probably due to different degrees of fractionation and different migration rates along the different flow paths through faults and fractures, on the one hand, and through unfractured beds, on the other, both leading to common entrapment sites. The water-flow related chemical factor that enhances the entrapment potential in discharge areas is the increase in formation water salinity due to an inverse proportionality between hydrocarbon solubility and water mineralization. As was seen earlier, mineralization of formation water increases toward, and is maximum in, the Pechelbronn–Soultz region, thus contributing potentially to the disaccommodation of petroleum at these sites.

In addition to creating hydrocarbon deposits, regional groundwater flow gives also rise to phenomena that may be used in exploration. Hydrogeologic phenomena that seem to be related to petroleum deposits in the study area through groundwater flow are: positive temperature anomalies both at depth and near the land surface; oil springs; regional anomalies of formation-water salinities; near-surface occurrence of brines (presumably migration up from depth along fractures) and saline springs, often containing H_2S; and localized wet lands which, however, are known in the area only from earlier records because land cultivation and urban settlements have obliterated them. Other hydrogeological indicators of petroleum deposits which can be expected in the study area but have not been examined, include: microseeps and/or anomalous contents of hydrocarbons in soils and groundwater; stressed vegetation growth; anomalies of electrotelluric potential and increased electrical conductivity associated with chemically reduced (i.e., organic rich) rock columns; sulfide ore formation and near-surface anomalies of heavy metals particularly Ni and V and I, Br, and B.

Summary and conclusions

The various aspects of the origin, migration as well as of the fluid-dynamic, thermal and chemical conditions of petroleum occurrence in the Pechelbronn–Soultz area of the Upper Rhine Graben are brought into a genetically linked and unified system of hydrogeological processes and phenomena by the hydraulic theory of petroleum migration (Tóth 1980). Similar conditions are known to occur in other intermontane basins also as, for instance, in Whitney Canyon–Ryckman Creek oil fields in Wyoming, U.S. (Jones and Drozd 1983; Zielinski et al. 1985), Railroad Valley in Nevada, U.S. (Reinsborough 1992), and the Mae Soon Oil Field, Fang Basin in N.W. Thailand. Although the hydraulic theory has been found adequate to explain petroleum occurrences in different environment types, such as, e.g., in the interior type Hungarian Basin (Erdélyi 1985), the platform type Michigan Basin (Vugrinovich 1988), or offshore Persian Gulf (Wells 1988); its predictions can be expected to be most accurate in intermontane basins: it is in this environment that the theoretical boundary conditions of the flow domain most closely approximate the real boundaries.

The most important and relevant properties and consequences of the regional groundwater flow in the Upper Rhine Graben include: well-defined regional recharge areas of high relative elevations which, combined with a slight longitudinal slope of the valley floor, result in a practically two dimensional field of highly intensive flow; a slow and relatively continuous increase in elevation difference between the recharge and discharge areas since Early Eocene, rendering the flow field virtually steady state for the latest approximately 20 to 40 Ma; well defined and localized cross-formational passage ways ("vents", as it were), provided by faulting and fracturing, which focus and direct subvertically the flow of the discharging formation fluids thereby activating and synergizing various lithological, physical, thermal and chemical trapping mechanisms; and the generation of numerous flow-dependent petroleum indicators.

On the basis of the close correlation between these observed properties and petroleum geological consequences of regional groundwater flow in the Upper Rhine Graben, on the one hand, and the predictions of the hydraulic theory of petroleum migration on the other, the theory appears to offer a physically based, many faceted, inexpensive and novel approach to petroleum exploration with a particulary good potential of application to intermontane basin environments.

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Mesozoic tensional basins in the Alpine–Carpathian shelf

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Since the Late Triassic, the Central Western Carpathians were separated from the European shelf by the rising Penninic Rift and transported to the ESE by lateral strike-slip. Middle Cretaceous, they were in juxtaposition with the Outer Carpathians (situated in the eastern sector of the North European shelf). Both elements converged during the Austrian and collided in the Laramian and Styrian phases. Tensional basin systems originated here during the Middle and Late Triassic, the middle Jurassic and the late Early Cretaceous. They were characterized by an expressive subsidence, dominating over the sedimentation rate, at least in their initial stages. The fringe bioherm growth rate along the Ladinian "Reifling" basins was much higher (500 to 700 mm/kyr) than the basinal sedimentation rate (4 to 14 mm/kyr). Since the early Carnian, these basins attained depths of more than one thousand meters. Subsequently, they were filled by the "Lunz humid event" products. Topmost Triassic basins were opened also by synsedimentary tension. Middle Jurassic basins were characterized by an expressive subsidence, inadequately compensated by the sedimentation. Mid-Cretaceous tensional basins formed during advent of the Paleoalpine stresses. Their subsequent filling up by early flysch deposits was conditioned by the tectonic activity, as well.

Key words: Geodynamic evolution, basin subsidence, sedimentation rate, Mesozoic, Western Carpathians

Paleogeographic position of the Central Western Carpathians at the beginning of the Triassic period

According to present paleogeographic knowledge (Kovács 1980; Tricart 1984; Brandner 1984; Michalík and Mišík 1987; Rakús et al. 1988; Michalík 1992; etc.), the Alpine–Carpathian Mesozoic sequences developed on the North European shelf. The arid character of the Triassic climate caused the accumulation of thick neritic carbonate complexes, with only several clastic terrigenous formations indicating humid climatic events (Scythian, Early Carnian, Rhaetian). The sources of the considerable, laterally dispersed clastic terrigeneous sediments are assumed to in the northern "Vindelician Land" (Fig. 1). Moreover, the zonality of both the Outer and Central Carpathians is the same (from terrigeneous deposits in the north to the pure marine facies along their southern margin), indicating an independent position of both segments at least during their Paleoalpine development, controlled by sinistral strike slip between them (Michalík 1992; Michalík et al. 1993).

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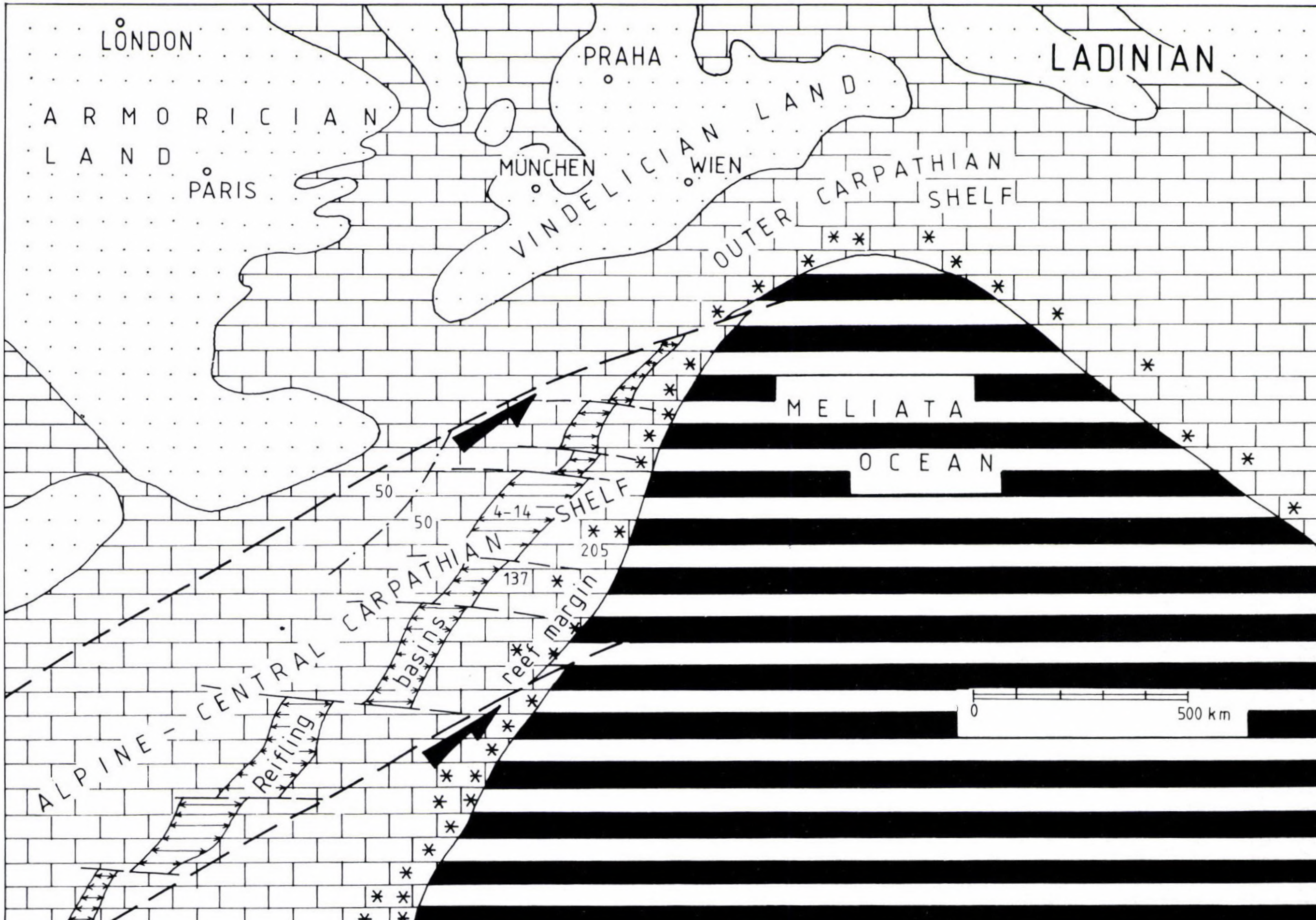

LADINIAN
LONDON
PRAHA
ARMORICIAN
LAND
PARIS
MÜNCHEN
WIEN
VINDELICIAN LAND
OUTER CARPATHIAN
SHELF
MELIATA
OCEAN
ALPINE – CENTRAL CARPATHIAN SHELF
basins
reef margin
Reifling
50
50
4-14
205
137
0
500 km

The volume of the Scythian clastics deposited in the Alpine–Carpathian area was calculated by Michalík (1993) as 75 000–100 000 km^3. This material was derived from 400 000 km^3 of eroded granitoid rocks in an approximately 750 000 km^2 large area (if the erosion rate was not greater than 200 mm/ka). Due to the proximity of the German Basin in the north, such a large southward-drained area could be localized only between the Bohemian and Armorician Massifs. Thus, at that time, the East Alpine–Central West Carpathian shelf segment could be situated to the south of this area (Michalík and Mišík 1987). The subsidence of this shelf was calculated as 5–20 mm/ka in its northern part, but as much as 200 mm/ka along its southern rim.

Middle Triassic carbonate platforms and basins

Terrigeneous deposit in the European shelf ceased during Spathian/Anisian boundary time. The restriction of water circulation between extremely shallow seas and the open ocean limited the dynamics of differentiating salinity and oxygen regime. Such a ramp in the Alpine–Carpathian area was not less than 250–300 km wide, with a volume of 750 000–800 000 km^3. The bioherms starting to evolve along its seaward border formed the transition to the carbonate platform geometry.

The sedimentation rate of the Anisian carbonates was 20–39 mm in the northern zones, but 40–100 mm on the south. Michalík (1993) estimated the tectonic subsidence of this shelf at that time as 40 mm/ka. However, the traces of raised substrate mobility at that time, comprising tsunamite layers and megabreccia infilling of arising rift valleys (Michalík 1979; Michalík et al. 1992), indicate the start of rifting which culminated during the Ladinian (Bechstädt et al. 1978). This rifting originated during left-lateral shear in the area, accompanying the eastward motion of the Apulian Block (Fig. 1). Pull-apart tensional basins, characterized by the deposition of Reifling, Buchenstein or Hallstatt Formations are sometimes accompanied by volcanic products formed in the wide Alpine–Carpathian carbonate platform belt (Michalík 1993). Depositional rate of basinal sediments was enormously reduced (4 to 14 mm/ka) compared to the marginal bioherm structures keeping up with the shelf subsidence (500 to 700 mm/ka). This difference contributed to the rapid deepening of these basins. Their depth reached more than thousand meters at the beginning of Early Carnian, cf. Masaryk et al. (1993). At that time the tension ceased and the intra-shelf basins were rapidly filled up by a debris of prograding reef and by clastic influx accompanying the Lunz climatic event (Michalík 1993).

←Fig. 1
Ladinian paleotectonic sketch of the northern Mediterranean. Dotted: dry-land, bricks: shelf sea, horizontal hatching: tensional basins, thick hatching: oceanic areas. Numbers denote sedimentary rates in the West Carpathian shelf

Late Triassic/Jurassic tensional basins

Boccaletti et al. (1984) argued that the Mesozoic Piedmont–Ligurian Ocean opened along the mega-shear zone between the Teisseyre–Tornquist Line and the Southern Atlas, contemporaneously with the Paleocimmerian shortening of the Paleotethys Ocean. A parallel sinistral strike system caused the opening of new pull-apart basin belts (Fig. 2). The infilling of these basins is represented by the Kössen, Fatra, Hybe, or Zlambach Formations (Michalík 1978; Golebiowski 1990).

Complete separation of the Alpine–Carpathian shelf fragment from the European Tethyan shelf by left-lateral strike during the Early Jurassic was accompanied by progressive tension, block tilting and crustal stretching in the intra-shelf basins. This tension, accompanied by submarine sliding, dyke and breccia origin, has been well documented by Michalík et al. (1994, in print) in the Kadlubek- and Kuchyna Units of the Tatric Superunit in Malé Karpaty Mts. Similar tension also affected the Paleoeuropean foreland and caused an activation of the old NW-SE fault systems (Ziegler 1982, see also Fig. 3).

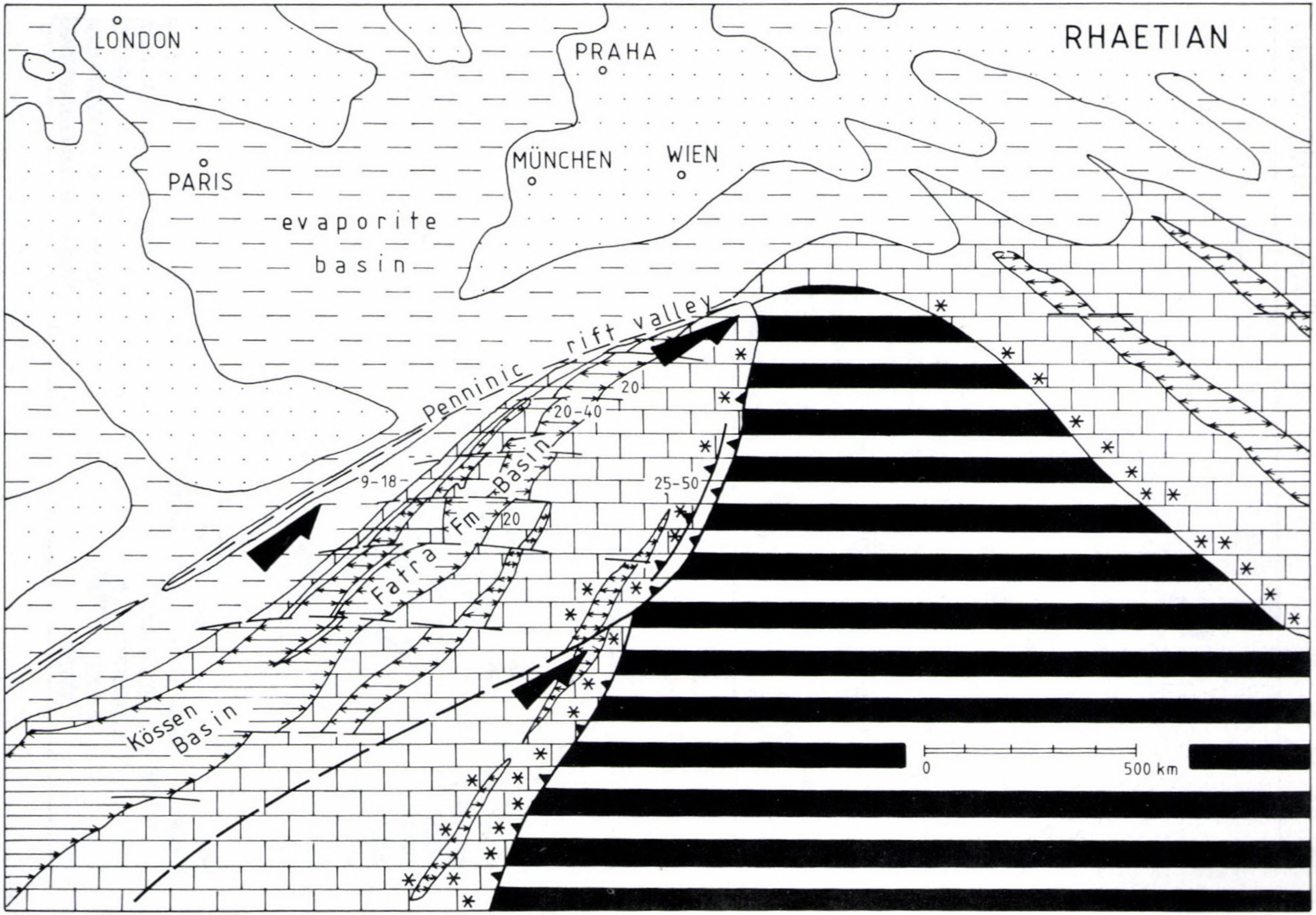
Early Cretaceous tensional basins

The eastward movement of the Alpine-Carpathian and Apulian blocks resulted in Neocimmerian collision. Only indirect effects of these deformations, affecting the southernmost zones, have been recorded in the Central Western Carpathians (Michalík 1991). On the other hand, the Upper Valanginian–Barremian turbidites containing chromium spinel grains, which were deposited in basinal areas, indicate synsedimentary tectonic movements (Reháková in print; Reháková and Michalík 1992, 1994; Jablonský et al. 1993). During the Barremian–Early Albian, carbonate platforms developed on elevated blocks, separated by deep depressions with black shale deposition (Fig. 4). The creation of tilted, elevated and depressed block mosaics were controlled by the third generation of tensional pull-apart mechanisms (Michalík, in print). This evolution was stopped during Middle Albian condensation, followed by the general collapse of the bottom of the Western Carpathian area. Sudden depth increase caused the substitution of carbonate sedimentation by a terrigeneous flysch depositional regime.

Late Cretaceous (Gosau) basins

Late Cenomanian and Turonian compressive movements formed the Austrian nappe structures in the Alpine–Carpathian Belt. The Senonian emergence caused the exposition of the elevated orogen to tropic weathering. On the other

Fig. 2 →
Rhaetian paleotectonic sketch of the northern Mediterranean. Explanations as in Fig. 1


RHAETIAN
LONDON
PRAHA
PARIS
MÜNCHEN
WIEN
evaporite
basin
Penninic rift valley
20
20-40
Fatra Fm Basin
9-18
25-50
20
Kössen Basin
0
500 km

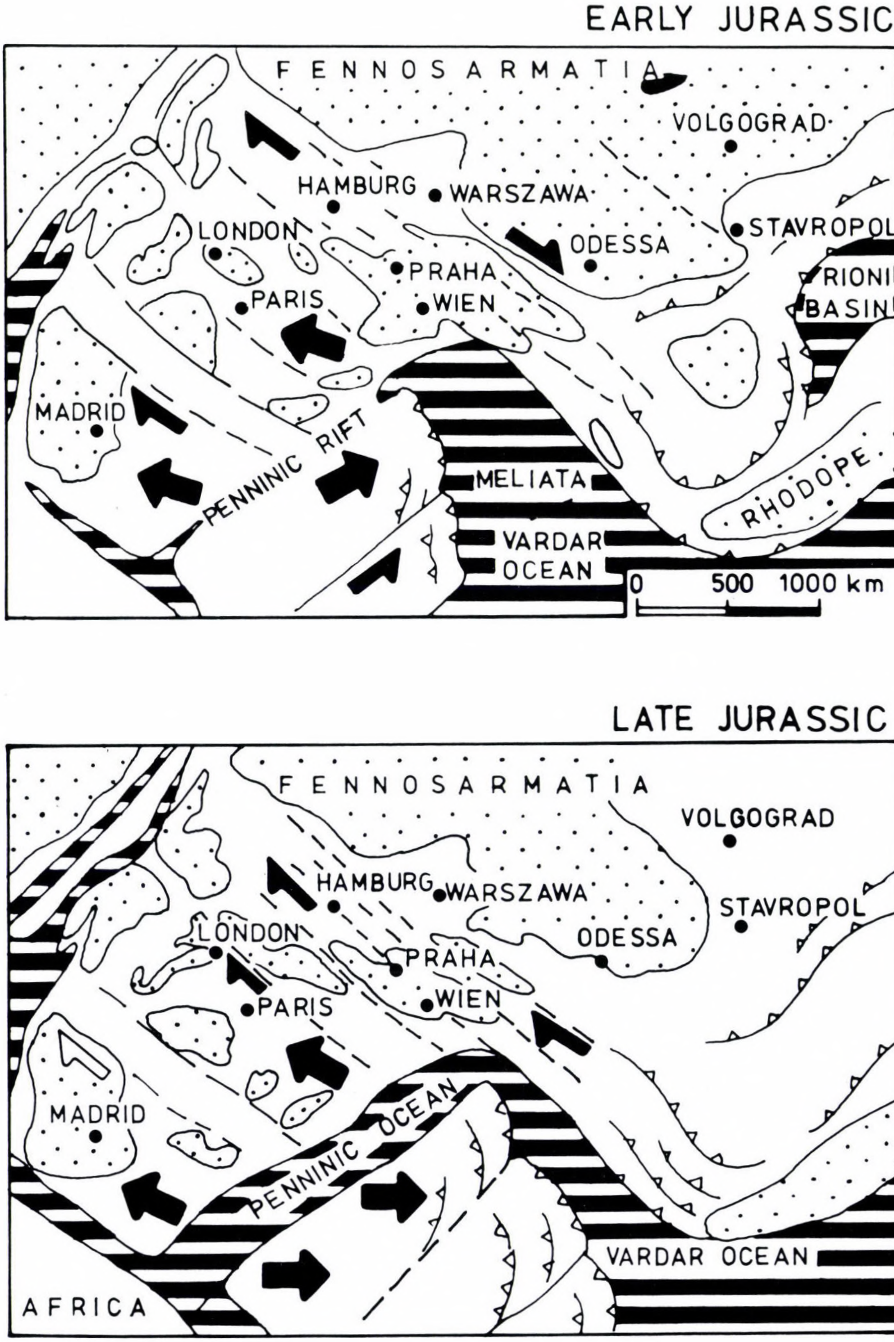
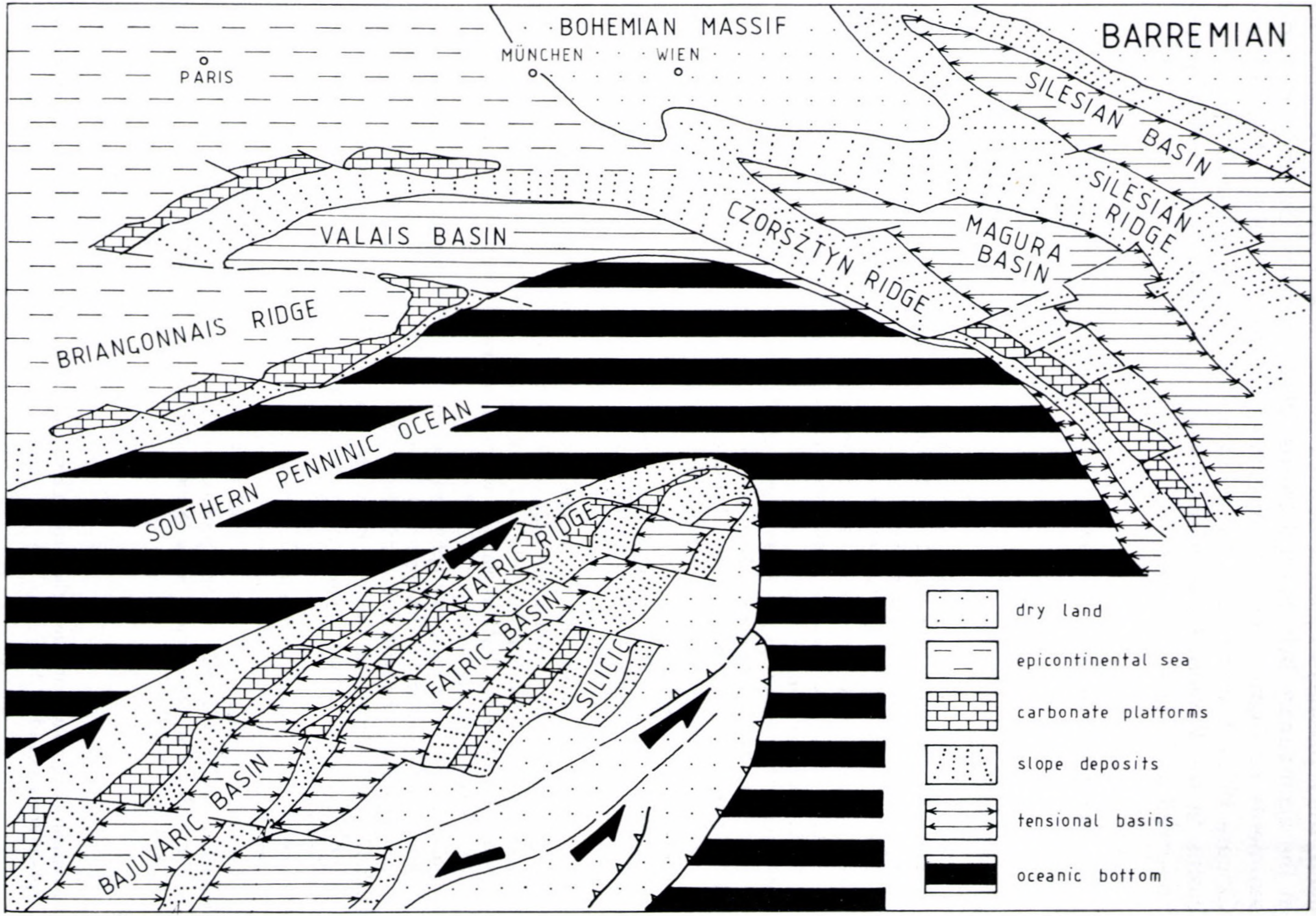


Fig. 3
Jurassic paleotectonics of the Middle Europe. Explanations as in Fig. 1


BARREMIAN
BOHEMIAN MASSIF
PARIS
MÜNCHEN
WIEN
SILESIAN BASIN
SILESIAN RIDGE
MAGURA BASIN
CZORSZTYN RIDGE
VALAIS BASIN
BRIANÇONNAIS RIDGE
SOUTHERN PENNINIC OCEAN
TATRIC RIDGE
FATRIC BASIN
SILICIC
BAJUVARIC BASIN
dry land
epicontinental sea
carbonate platforms
slope deposits
tensional basins
oceanic bottom

hand, thick prisms of flysch sediments were deposited in the externides and in the accretionary belt. Several basins filled by coarse clastics and marine sediments of Gosau type also originated in the centralides (Michalík and Cincura 1992). They represent the youngest generation of Mesozoic pull-apart basins in the Western Carpathians. However, these basins are incompletely preserved and less well known.

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←Fig. 4
Barremian "Urgonian" carbonate platforms–versus euxinic basins pattern in Western Carpathians. Explanations as in Fig. 1

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Palynological revaluation of Lower Triassic and Lower Anisian formations of Southeast Transdanubia

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The first palynological investigations of Lower Triassic and Lower Anisian formations of Southeast Transdanubia (Jakabhegy Sandstone, Patacs Siltstone, Hetvehely Dolomite), together with the underlying Permian formations were published by the author in 1981. The revaluation of these investigations has been proved by several reasons, i.e. in the eighties these formations were revealed in other regions of Southeast Transdanubia and contained well-preserved microflora, further in the past decade several palynological studies were published (mainly from fossiliferous Alpine and German-type regions) that provided novel data to the classification of Lower and Middle Triassic sequences on palynostratigraphic bases.

Triassic formations of Southeast Transdanubia contain sporomorphs from the upper third of the Jakabhegy Sandstone to the bottom of the lower member of the Hetvehely Dolomite (Magyarürög Anhydrite) that can be assigned to three palynological assemblages, from down upwards as follows: *Densoisporites nejburgii, Voltziaceaesporites heteromorphus – Triadispora crassa* and *Triadispora crassa – Stellapollenites thiergartii*. The two older assemblages, indicating the upper part of the Olenekian substage of the Scythian stage, occur only in the Jakabhegy Sandstone Formation, while the assemblage indicating the uppermost, i.e. Anisian stage (*T. crassa–S. thiergartii*) occur in the strata of the overlying Patacs Formations and Magyarürög Anhydrite Member, in addition to the uppermost strata of the Jakabhegy Sandstone.

Thus, the Scythian/Anisian boundary can be drawn in the upper part of the upper third of the Jakabhegy Sandstone Formation, the strata of the microfloristically studied sequences above this boundary belong to the Anisian, those below this boundary to the upper part of the Olenekian substage of the Scythian stage.

Since as to the palynological investigations of 1981 the Permian/Triassic boundary lies in the underlying layer of the Jakabhegy Sandstone, within the upper member of the Kővágószőlős Sandstone, the fossil-free strata of the latter formation above the P/T boundary but below the *D. nejburgi* assemblage of the Jakabhegy Sandstone are of Lower Triassic (Scythian) age.

Key words: Lower Triassic, Middle Triassic, Southeast Transdanubia, Mecsek Mts., Villány Mts., Hungary, palynological assemblages, palynostratigraphy

Introduction

The Jakabhegy Sandstone, the Patacs Siltstone and the Hetvehely Dolomite Formations constitute the oldest, i.e. Lower Triassic and Early Middle Triassic lithostratigraphic units of the Mesozoic of the Tisia tectonic unit of Hungary. These are most thoroughly known in the Southeast Transdanubian part of the Mecsek–Villány Late Paleozoic to Mesozoic belts: in the anticline structure of the Western Mecsek Mountains, in the Villány Mountains and in the covered

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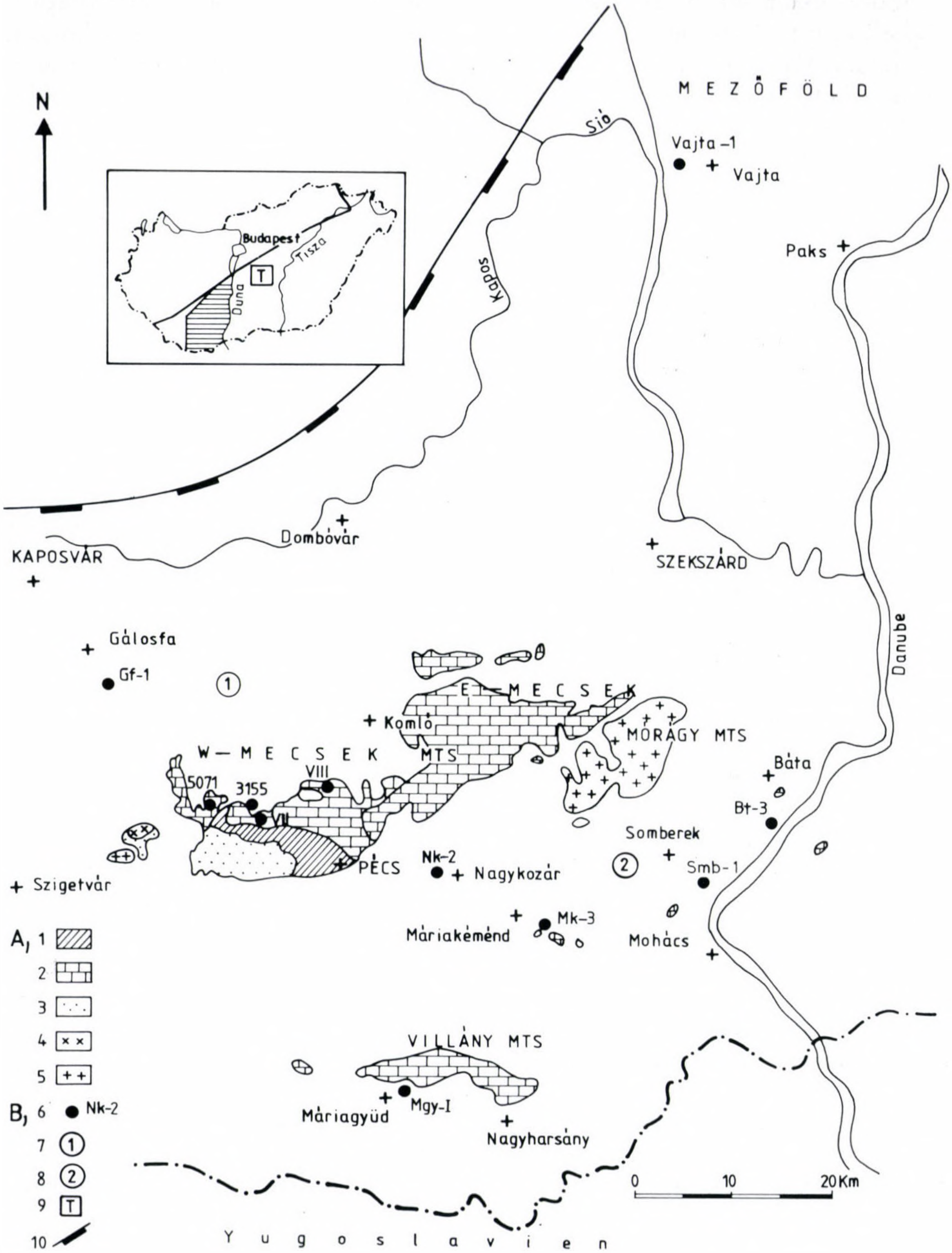
areas surrounding these mountains. Their stratotypes are found on the surface in the Western Mecsek Mountains where their geological exploration started already in the second half of the past century and has been continuous with shorter–longer interruption even to our days. In the past four decades numerous boreholes revealed these formations providing continuous and excellent rock material to their complex processing. These formations have become revealed in the Villány Mountains, in the Mezőföld and in the northern and southern foregrounds of the Mecsek Mountains from boreholes drilled in the past fifteen years, together with their over- and underlying formations (Fig. 1).

Former researches in the Western Mecsek mountains proved that the oldest formations, i.e. the Jakabhegy Sandstone Formation does not contain fauna. The fauna content of the overlying Patacs and Hetvehely Formations is also very poorish. Thus, to improve their temporal classification, I performed the palynological investigation of the formations in question and of the samples deriving from boreholes of the Western Mecsek mountains. Results were published in 1981. Based on the studied microflora of these samples lying rather far from each other (25 to 65 m), as well as on references known up to the seventies and on the stratigraphic evaluation of the fauna of younger formations of that time, I assigned the Jakabhegy Sandstone Formation to the lowermost Lower Triassic, the Patacs Siltstone and Hetvehely Dolomite Formations to the upper part of the Lower Triassic. The palynological assemblage of "transitional character" deriving from the upper strata of the Kővágószőlős Sandstone Formation lying under the Jakabhegy Sandstone Formation and believed as Late Permian, that in addition to typical Upper Permian conifer pollens contained remarkable quantities of Lower Triassic pterydophyte spores, was discussed also in this work.

The palynological revaluation of the three Triassic formations to be discussed below is verified by the fact that in the course of industrial explorations between 1980 and 1985 several boreholes revealed these formations outside the Mecsek Mountains (Fig. 1). In certain boreholes these formations contained more frequently and in greater thickness fine-grained clastic sediments deposited in reductive environment, suitable to preserve the microflora, especially in case of the Jakabhegy Sandstone. In the palynological material found in these formations the assemblages known from the previous investigations were found in much shorter stratigraphic distance. At the same time, older microflora

Fig. 1 →
Generalized geological map of the Mecsek–Villány zones of the Tisza Unit with localities of the samples investigated. A) *Surface formations*: 1. Lower and Early Middle Triassic (palynologically studied formations in this work): Jakabhegy Sandstone, Patacs Siltstone and Hetvehely Dolomite Formations; 2. Middle–Upper Triassic, Jurassic, Cretaceous; 3. Early and Late Upper Permian: Cserdi, Boda Aleurolite, Kővágószőlős Sandstone Formations; 4. Lower Permian: Gyűrűfű Rhyolite Formation; 5. Lower Carboniferous: Mórágy Granite Formation; B) *Other marks:* 6. boreholes; 7. northern foreground of Western Mecsek Mountains; 8. southern foreground of the Mecsek–Mórágy Mountains; 9. Tisza Unit, 10. northern boundary of the Tisza Unit


N
Budapest
Tisza
Duna
T
MEZŐFÖLD
Sió
Vajta-1
Vajta
Paks
Kapos
Dombóvár
KAPOSVÁR
SZEKSZÁRD
Danube
Gálosfa
Gf-1
1
E-MECSEK
Komló
MÓRAGY MTS
W-MECSEK MTS
5071
3155
VIII
VII
Báta
Bt-3
Somberek
PÉCS
Nk-2
Nagykozár
2
Smb-1
Szigetvár
Mk-3
Máriakéménd
Mohács
VILLÁNY MTS
Máriagyüd
Mgy-I
Nagyharsány
0 10 20Km
Yugoslavien
A, 1
2
3
4
5
B, 6 Nk-2
7 1
8 2
9 T
10

assemblages unknown so far in the region, were also determined. Results of the former and recent investigations allowed the more exact determination of temporal extension of the palynological assemblages, the palynostratigraphic classification of the formations and the indication of the Scythian/Anisian boundary. On these bases, their regional parallelization and correlation with known and palynologically studied Lower and Middle Triassic profiles containing for the most part characteristic fauna could be performed with greater accuracy. As a final conclusion, suggestion could be made to the chronostratigraphic classification of the formations.

2. *History of knowledge of the formations*

The history of knowledge of the Mecsek formations was discussed in detail in the paper published in 1981. The temporal change of opinions concerning the age of formations is found in Table 1 of this work. It shows that the Patacs and Hetvehely Formations were assigned by all researchers to the Triassic (Seisian, Campilian), but the age of the fossil-free Jakabhegy Sandstone has been debated. Prior to the palynological investigations (up to 1981) this formation was qualified as Permian in harmony with the lithological similarity to the underlying Kővágószőlős Sandstone, but was assigned to the Lower Triassic when emphasizing the transgressive character of the Jakabhegy Sandstone. To accept the Permian age of the Jakabhegy Sandstone Formation was stressed by the communication in the past century (Peters 1862), i.e. *Claraia clarai* fauna is present in the Patacs Siltstone Formation (discussion see later). Nevertheless, the first palynological investigations supported the Lower Triassic age of the Jakabhegy Sandstone Formation.

Outside the Mecsek region, these formations were identified in boreholes drilled in the environs of the Villány Mountains between 1962 and 1970, but no palynological analyses were made due to the lack of strata suitable to preserve the microflora. The lithology and sedimentology of the Jakabhegy Sandstone, as well as the geology of the Patacs Siltstone and Hetvehely Dolomite Formations found here were discussed by Kassai (1976) and by E. Nagy and I. Nagy (1976), respectively. They put the Jakabhegy Sandstone into the Permian and Scytho–Permian, respectively, while on the basis of their macrofossil content and analogies to the Mecsek Mountains into the Seisian and Campilian substages of the Lower Triassic (Table 1).

The palynological investigations of the formations known from boreholes drilled between 1980 and 1985 in the northern and southern foreground of the Mecsek Mountains, in the Villány Mountains, and in the Mezőföld (Gf-1, Nk-2, Mk-3, Smb-1, Bt-3, Mgy-1, Vajta-3; see Fig. 1), as well as the revaluation results of the boreholes Western-Mecsek VII, VIII and 3155 will be discussed below.

Table 1
Development of opinions concerning the age of the Hetvehely Dolomite, Patacs Siltstone, Jakabhegy Sandstone Formations and their underlying-overlying beds of Mecsek and Villány* Mountains

Böckh, J. 1876	Vadász, E. 1935, 1953	Barabás, A. 1955 Barabás, A. - Kiss, J. 1958	Tőzsér, O. 1961 Kiss, J. 1961	Jámbor, Á. 1964 Nagy, E. 1961, 1968	Barabás-Stuhl, Á. 1962, 1967, 1969, 1975	Kassai, M. 1969, 1971, 1975, 1976a	Balogh, K.- Barabás, A. 1972 Balogh, K.-Barabás, A. -Majoros, Gy. 1973	Kassai, M. 1976b, 1978	Barabás, A. 1977	Barabás-Stuhl, Á. 1981	Nagy, E.-Nagy, I. 1976 (Villány Mts.)*	Recent classification 1991	Age
Muschel-kalk	Anisian			Anisian				Anisian		Middle Triassic	Anisian	Lapis Limestone Formation Rókahegy Dolomite Formation	Middle Triassic (Anisian)
Early Triassic	Campil beds			Campil beds				Campil beds	Lower Triassic	Lower Triassic: Upper part	Campil beds	Hetvehely Formation: Viganvár Limestone Member	
	Werfenian	Seis beds	Early Triassic									Hetvehely Formation: Magyarürög Anhydrite Member	
				Seis beds	Lower Triassic	Lower Triassic	"Seis beds"	Seis beds			Seis beds	Patacs Siltstone Formation	
	Late Permian			Late Permian Upper Permian	Late Permian			"Permo-skyth"		Lower part ?	Late P. Upper P.	Jakabhegy Sandstone Formation	Lower Triassic (Scythian)
Late Permian	Middle Permian	Middle Permian	Late Permian	Middle Permian		Upper Permian	"Zechstein"	Upper Permian	Upper Permian	? Upper Permian		Kővágószőlős Sandstone Formation	Upper Permian
		Early Permian	Middle Permian	Upper Rotliegend		Lower Permian	"Rotliegend"		Middle Permian	Middle Permian		Boda Aleurolite Formation	

3. Geology of the formations

The Jakabhegy Sandstone Formation is the oldest lithostratigraphic unit of the studied sequence. It is built up solely by clastic rocks in a thickness of 100–400 m. In the anticline of the Western Mecsek Mountains, in the western part of its southern foreground and in the northern foreground it overlies with some smaller hiatus and slight angular unconformity the Kővágószőlős Sandstone Formation of Upper Permian to Lower Triassic age. In other areas it is found over different Paleozoic formations with unconformity. In the eastern part of the southern foreground it overlites the Early Upper Permian Cserdi Formation, in the Villány Mountains and in its environs either the Lower Permian Korpád Sandstone or the also Lower Permian Gyűrűfű Rhyolite. In the southwestern and eastern margin of the Mórágy Hills it overlies Lower Carboniferous granite, in the northern foreground of the Eastern Mecsek it is found above the Szalatnak Siliceous Shale Formation. In the Mezőföld it is underlain by red Permian sandstone, the stratigraphic position of which within the Permian cannot be determined. In the syncline structure constituting the eastern part of the Mecsek Mountains the formation in question has been unexplored so far.

The Jakabhegy Sandstone is a rhythmic sequence of predominantly reddish, purplish shade, with green and grey strata in the upper parts, some rock types of which resemble to the rocks of the uppermost so-called Cserkút Red Sandstone Member of its underlying part in the Mecsek Mountains.

The formation is built up by five unnamed lithological members (a, b, c, d and e) that can be traced also horizontally in wide areas. The lowermost unit is the a) "main conglomerate" consisting predominantly of rhyolite and quartz pebbles and displaying unstratified structure, this is overlain by the b) unstratified red gravelly sandstone; this is followed by the c) pale violet, banked gravelly sandstone of finer grain-size, often with cross-bedding and clayey intercalations; and this is overlain by d) red-brown, fine-grained, unstratified aleurolitic sandstone strata. These fine-grained sediments are overlain by the e) slightly stratified clastic rocks showing transition to the overlying formation. In this last member the typical violet Jakabhegy Sandstones of cross-bedding are rarely found, but the sandstone, aleurolite and intraformational breccia strata of varied colour (red, green, grey) are more frequent (Fig. 2).

The unit d) is lacking in the borehole Vajta–3 and also in the Jakabhegy Sandstone Formation revealed by boreholes in the northern foreground of the Mecsek Mountains.

No faunal fossils are found in the Jakabhegy Sandstone, some macroflora remains, however, were found in the uppermost rock unit containing also grey strata. These were determined by G. Andreánszky as *Voltzia heterophylla* BRGT. and Equisetites sp. (in Kassai 1976).

In lack of faunal fossils, the facies of the Jakabhegy Sandstone Formation can be hardly defined. Nevertheless, on the basis of temporal changes of

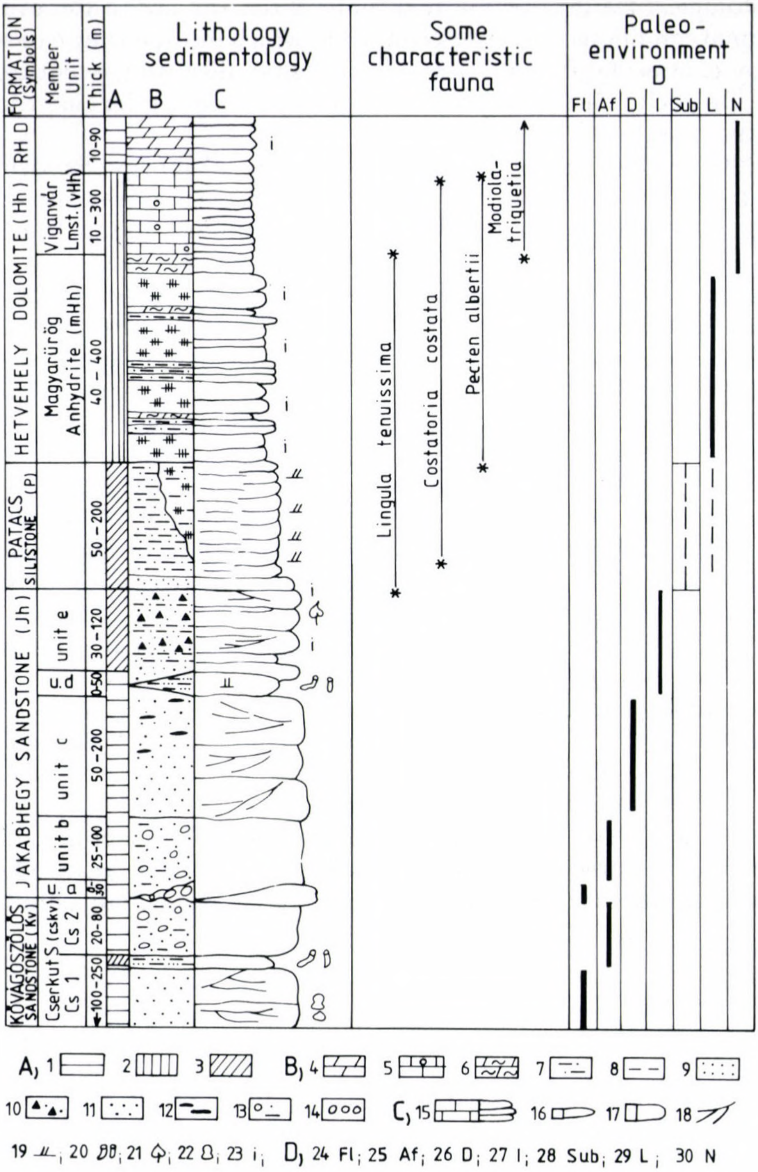


Fig. 2
Idealized profile of the Jakabhegy Sandstone, Patacs Siltstone and Hetvehely Dolomite Formations. A) *Colour:* 1. red; 2. grey; 3. varying colour (red, green, grey); B) *Lithology*: 4. dolomite; 5. nodular limestone; 6. dolomite–marl; 7. mudstone and aleurolite; 8. gypsum and anhydrite; 9. fine-grained sandstone; 10. intraclastic sandstone; 11. coarse-grained sandstone; 12. mudstone pebble in sandstone; 13. gravelly sandstone; 14. conglomerate; C) *Sedimentology*: 15. lamination; medium-thick bedding; 16. middle-thick-bedding; 17. thick-bedding or unstratified; 18. cross-bedding; 19. micro-crosslamination; 20. vertical/horizontal burrows; 21. plant remains; 22. carbonate concretion; 23. intraclasts; D) *Paleoenvironment:* 24. fluviatile; 25. alluvial fan; 26. deltaic; 27. intertidal; 28. subtidal; 29. lagoonal; 30. neritic

sedimentological features of the rock units discussed above the evolution of depositional conditions can be determined in which the deposition of sediments started by continental fluviatile environments (a), then underwent the alluvial fan (b), the delta facies (c) and finally continued under intertidal plain conditions (d and e).

The Patacs Siltstone Formation overlies the Jakabhegy Sandstone with gradual transition, the boundary is marked by the lack of sandstone strata of Jakabhegy-type and by the predominance of aleurolitic strata. The formation is 50 to 120 m thick, it is built up by differently stratified sandstone, aleurolite and claystone strata of red colour at the bottom and varied colour upwards, with some dolomarl intercalations in the upper part. In the northern and southern foregrounds of the Mecsek Mountains the evaporites (gypsum, anhydrite) already occur in this formation, that are characteristic of the overlying formation. This means that the shallow marine, littoral clastic sedimentation is gradually replaced by the deposition of chemical sediments that become predominating in the anhydrite, dolomite and limestone bearing members of the Hetvehely Dolomite Formation and concordant Patacs Formation that do not contain red sediments. The latter formation is 50 to 400 m thick. Its Magyarürög Anhydrite Member of lagoonal facies is built up by grey anhydrite, gypsum, dolomite, dolomarl, foliated claystone while the overlying 10 to 300 m thick Viganvár Limestone Member of shallow marine deposition consists of dark-grey micro-nodular, thin-stratified limestone strata with marl intercalations and with breccia at the bottom (Fig. 2).

Above the Viganvár Limestone Member thin-stratified red dolomite of brecciated texture (Rókahegy Dolomite Formation) follows. This formation, however, is out of our recent field of study.

The Patacs and Hetvehely Formations contain macrofauna both in the surface outcrops and in boreholes, with changing frequency. The main elements in harmony with the comprehensive work of E. Nagy (1968) are as follows:

Patacs Siltstone Formation:

Isaura albertii (Voltz.)
Lingula tenuissima Bronn.
Costatoria costata Zenk.
Myacites fassaänsis Wissm.

Hetvehely Dolomite Formation, Magyarürög Anhydrite Member:

Pseudomonotis sp.
Costatoria costata Zenk.
Velopecten cf. *albertii* Goldf.
Lingula tenuissima Bronn.

Hetvehely Dolomite Formation, Viganvár Limestone Member:

Lingula tenuissima Bronn.
Pecten albertii Goldf.
Modiola triquetra Seeb.
Modiola gibba Alb.
Gervilleia modiola Frech.

Gervilleia goldfussi Stromb.
Costatoria costata Zenk.
Myophora elegans Dunker
Naticella costata Münst.

As regards the fauna it is to be noted that some authors put the Jakabhegy Sandstone Formation into the Upper Permian on the basis of stratigraphic evidence after Peters (1862), i.e. in the overlying strata of the formation *Posidonomya clarai* (*Claraia clarai*) indicating Early Triassic can be found. As to our recent knowledge the occurrence of this species is rather doubtful since in the course of recent thorough researches no traces of it were found. Peters determined his find as *P. clarai* being in embryonal state, so it can be presumed that it might have been a phyllopod, frequent in the Patacs Formation.

Among the formations discussed above, within the Southeast Transdanubian area differences occur only in their thickness and in the frequency of occurrence of their rock types. E.g. the thickness of all the three formations is greatest in the Western Mecsek Mountains and in its foregrounds, it is much thinner in the other localities. Further, evaporite strata occur already in the Patacs Formation in the northern and southern foreground of the Western Mecsek Mountains. In these localities, the intercalation of fine-grained aleurolitic claystone intercalations are more frequent in the Magyarürög Anhydrite than in other parts of the region. It is to be noted too, that in the Jakabhegy Sandstone Formation revealed by the borehole Vajta–3 the unit d) does not occur, here the unit c) is overlain with conformity by the unit e).

As to the recently generally accepted view, on the basis of their formation the Jakabhegy Sandstone, the Patacs Siltstone and the Hetvehely Dolomite Formations show relations to the German Triassic facies and similarly, these were generated in the northern (European) margin of the Paleotethys.

4. *Palynological characterization of the Lower and Early Middle Triassic*

In the past decades many excellent studies were published on the microflora content and palynostratigraphic evaluation of the sedimentary formations deposited in the Lower and Middle Triassic in the Tethyan realm and surrounding epicontinental and continental regions as well as in the Gondwana continent. Based on the studies dealing with the microflora and palynostratigraphy of the Lower and Early Middle Triassic formations of Alpine and German facies of the surrounding regions (Antonescu 1976; Brugman 1986; Góczán et al. 1986; Orlowska-Zwolinska 1984, 1985; Broglio Loriga et al. 1990) as well as on the "experimental" (manuscript) comprehensive work of Brugman (1983) summarizing the results of Permian–Triassic palynological studies published so far, the Lower Triassic (Scythian) and Early Middle Triassic (Anisian) can be palynologically characterized as follows:

1. The lowermost beds of Lower Triassic is characterized both locally and globally by the increased significance of spores that played subordinate role in

the Permian only. Locally, already in the youngest Permian and oldest Triassic strata new genera and species of spores occur. Certain spore species reach their acme in about the middle of the Lower Triassic (e.g. *Densoisporites nejburgii*), then their frequency suddenly decreases and in the upper part of the Lower Triassic as well as in the Early Middle Triassic they are found in the microflora in small individual number. New spore species occur with small species and individual number also in the upper part of the Lower Triassic. It is also characteristic that in the lower part of the Lower Triassic the marine facies are very abundant in organic microplankton remains.

In the Lower Triassic microflora several species of bisaccate conifer pollen predominating in the Permian exist in addition to the proliferating pterydophyte spores, but the index species of the Upper Permian, i.e. *Lueckisporites virkkiae* is found either in traces or is lacking. Bisaccate pollens of Paleozoic character coming from the Permian become gradually subordinate during the Lower Triassic, later these disappear. Their disappearance coincides roughly with the cease of predominance of pterydophyte spore species mentioned above.

2. Contemporaneously with and subsequently to this event, in the upper part of the Lower Triassic a locally and globally detectable change follows in the microfloral composition: new bisaccate pollen species of Mesozoic type occur that flourish in the Middle Triassic.

Based on the changes of microflora discussed above and related to the changes of climate and paleogeographic conditions two important points of evaluation can be determined that may serve as a basis (in case of palynological samples of sufficient density) to mark the boundaries of chronostratigraphic value in fauna-free continental sediments. This procedure is especially correct in the case when the fauna-free microflora can be correlated with strata containing fauna (in favourable case index fossils) in addition to sporomorphs.

5. *Palynological investigation of the Jakabhegy Sandstone, Patacs Siltstone and Hetvehely Dolomite Formations*

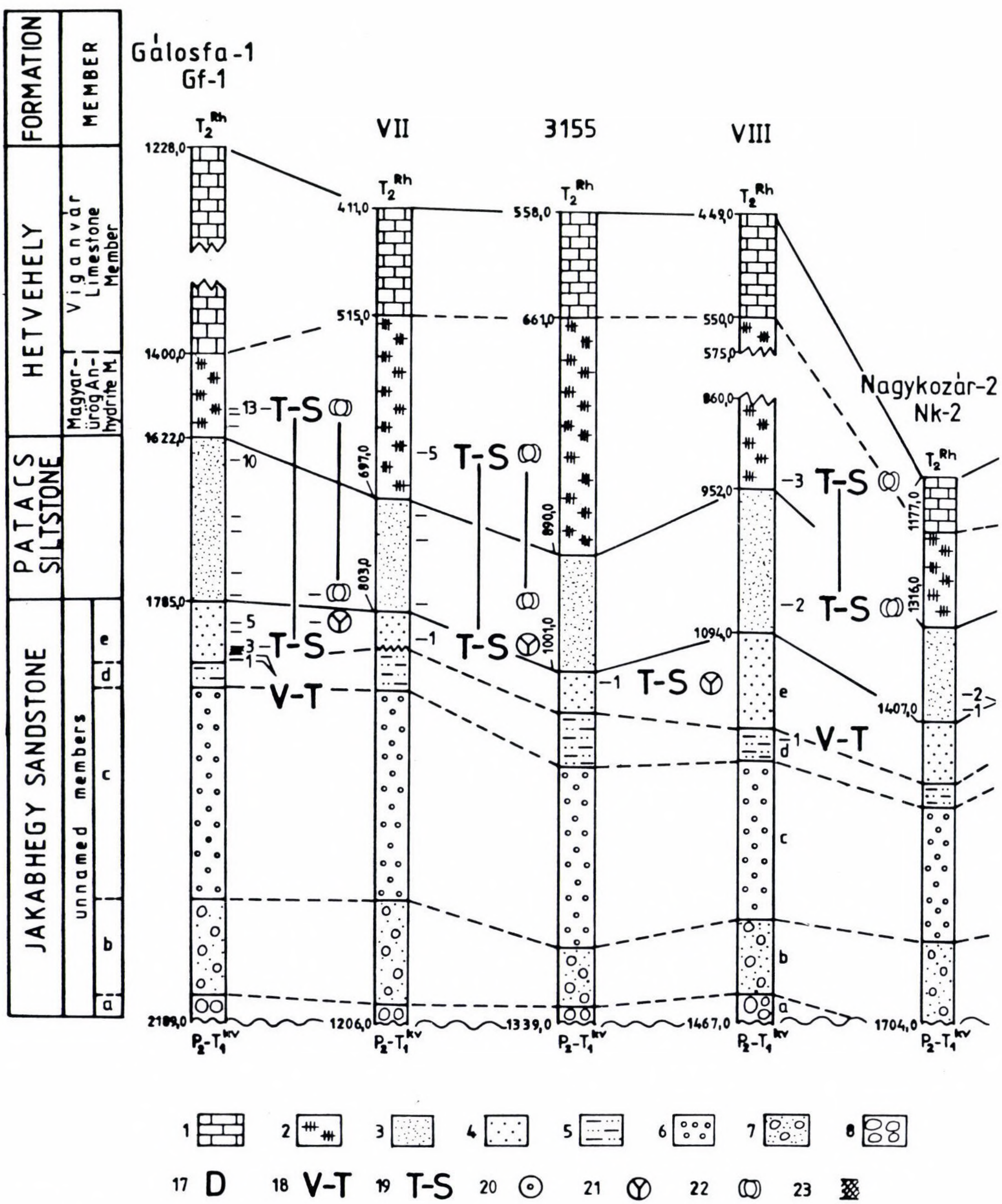
Palynological investigation of the formations was carried out in all cases on samples deriving from boreholes. Borehole localities are presented. in Fig. 1; out of the boreholes only the palynological investigation of Western Mecsek VII, VIII and No. 3155 was published in the first paper mentioned in the introduction (Barabás-Stuhl 1981). The investigation of microflora of the Lower and Early Middle Triassic sequence of the other boreholes drilled outside the Western Mecsek is presented first in this paper.

All data of sporomorph containing samples are listed in Table 2 and in Fig. 3. In the Table the data of two samples of the borehole No. 5071 deriving from the underlying strata of the Jakabhegy Sandstone Formation (Kővágószőlős Sandstone Formation) and indispensable to evaluate the Permian/Triassic boundary are also presented.

Table 2
List of the microflora samples

Borehole	Depth (m)	Sample number original/ in study		Lithology of samples	Symbol of lithostrat
Gálosfa-1	1601.0	Po 2107/	13	dark grey siltstone	mHh
(Gf-1)	1606.0	2106/	12	dark grey siltstone	P
	1617.0	2108/	11	dark grey siltstone	P
	1645.0	2111/	10	dark grey siltstone	P
	1700.0	2117/	9	dark grey siltstone	P
	1709.0	2148/	8	dark grey siltstone	P
	1753.8	2149/	7	dark grey siltstone	P
	1778.7	2131/	6	dark grey siltstone	P
	1809.2	2173/	5	grey siltstone	Jh/e
	1817.5	2134/	4	grey siltstone	Jh/e
	1835.6	2132/	3	grey siltstone	Jh/e
	1846.5	2135/	2	grey siltstone	Jh/e
	1857.6	2213/	1	grey siltstone	Jh/e
VII	647.0	Po 1347/	5	dark grey siltstone	mHh
	704.0	1686/	4	dark grey siltstone	P
	732.0	1689/	3	dark grey siltstone	P
	798.0	1691/	2	dark grey siltstone	P
	833.0	1362/	1	dark grey siltstone	Jh/e
3155	1013.0	Po 1491/	1	dark grey siltstone	Jh/e
VIII	941.0	Po 1692/	3	dark grey siltstone	mHh
	1063.0	1697/	2	dark grey siltstone	P
	1200.0	2154/	1	grey siltstone	Jh/d
5071	1071.0	Po 1805/	2	grey sandstone	cs Kv
	1122.0	1810/	1	grey sandstone	cs Kv
Nagykozár-2 (Nk-2)	1384.5	Po 2507/	2	dark grey siltstone	P
	1397.0	2506/	1	dark grey siltstone	P
Máriakéménd-3 (Mk-3)	998.0	Po 2305/	1	dark grey siltstone	P
Somberek-1 (Smb-1)	1510.0	Po 2585/	1	dark grey siltstone	P
Báta-3 (Bt-3)	712.0	Po 2694/	1	dark grey siltstone	P
Máriagyüd-I (Mgy-I)	1174.2	Po 2619/	3	grey fine sandstone	Jh/e
	1174.5	2672/	2	grey fine sandstone	Jh/e
	1188.6	2673/	1	grey fine sandstone	Jh/e
Vajta-3	1049.0	Po 2538/	5	dark grey siltstone	mHh
	1121.0	2636/	4	dark grey siltstone	P
	1155.0	2667/	3	dark grey siltstone	Jh/e
	1167.0	2669/	2	dark grey siltstone	Jh/e/c
	1169.0	2670/	1	dark grey siltstone	Jh/e/c

Fig. 3


FORMATION
MEMBER
HETVEHELY
Viganvár Limestone Member
Magyarürög Anhydrite M.
PATACS SILTSTONE
JAKABHEGY SANDSTONE
unnamed members
a
b
c
d
e
Gálosfa-1
Gf-1
VII
3155
VIII
Nagykozár-2
Nk-2
1228,0
1400,0
1622,0
1785,0
2189,0
411,0
515,0
697,0
803,0
1206,0
558,0
661,0
890,0
1001,0
1339,0
449,0
550,0
575,0
860,0
952,0
1094,0
1467,0
1177,0
1316,0
1407,0
1704,0
T-S
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1
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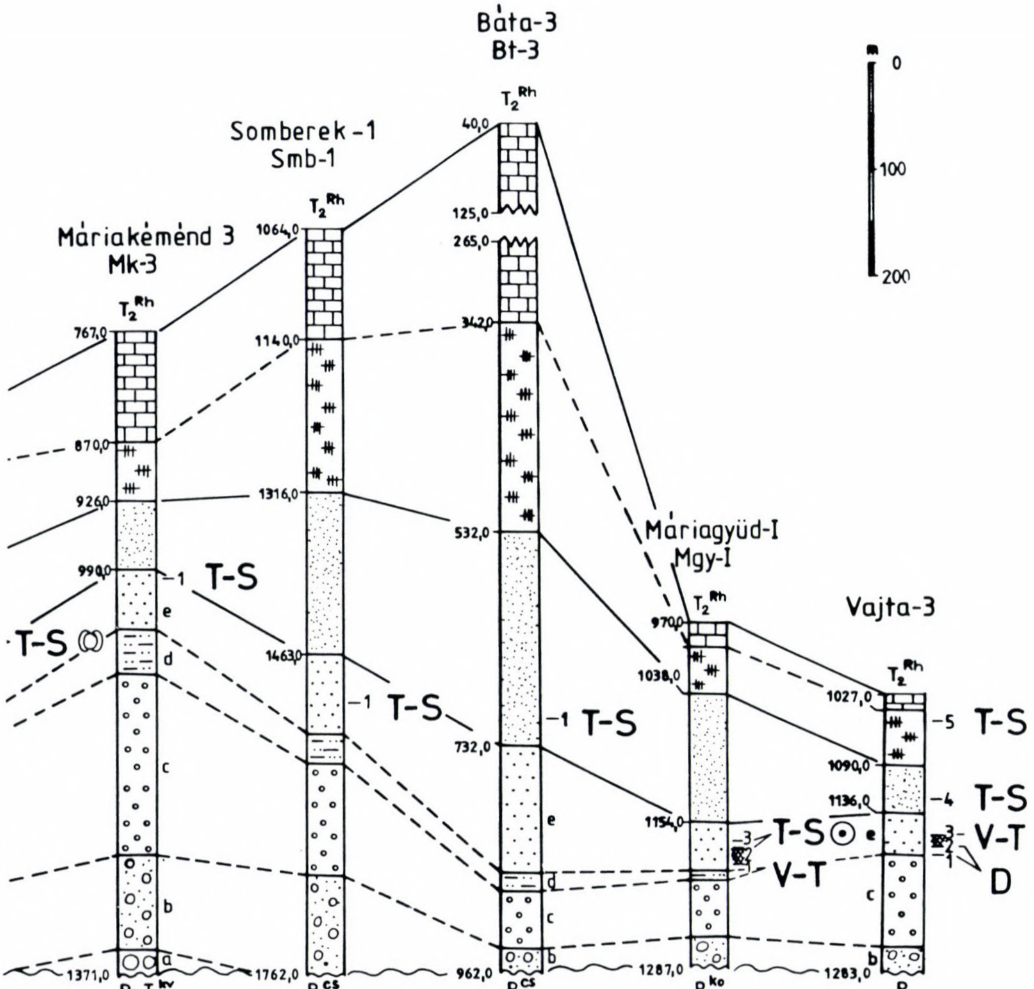
Fig. 3 (cont.)



9 T_2^{Rh} 10 P_2-T_1^{kv} 11 P_2^{cs} 12 P_1^{ko} 13 P 14 15 16 –1

In the table and figure cited it is seen that microflora-containing samples display different numbers by boreholes and by regions, their number is low in general and in many cases lie far from one another. From the aspect of palynological investigation the formations revealed by the borehole Gálosfa–1 (Gf–1) are most favourable that were not involved in the paper of 1981. The same is the situation in case of the boreholes Vajta–3 of Mezőföld and Máriagyüd–I (Mgy–I) of the Villány Mountains in which different palynological assemblages are present though in smaller number but close to one another in samples.

The Jakabhegy Sandstone built up predominantly by red clastic sediments contains strata suitable to preserve the microflora only in its upper parts (units c, d and e), but samples deriving from these units proved to be barren with respect to palynology. The same can be said on the relatively frequent grey fine-grained strata of the Patacs Siltstone and Hetvehely Formations in which – especially in the Magyarürög Anhydrite and Viganvár Limestone Members – the barren parts were more frequent than the palynologically productive ones. Though the samples of the anhydritic and limestone members contained microflora in many cases, their state was unsuitable to determination (translucent, light, structure–free specimens). In the formations revealed by the Somberek–1 (Smb–1) and Báta–3 (Bt–3) boreholes, the thermal effect of the subsequently intruded Cretaceous diabase veinlets caused high-grade coalification in the microflora material that grains could not be used to determination even after strong oxidation. Otherwise, the state or preservation of spore and pollen grains of the assemblages to be discussed was medium, in some cases of very good quality.

The studied microflora contained a lot of spore and pollen genera and species. Morphologically the spores belong for the most part to the categories Azonotriletes and Zonotriletes while pollens to those of Disaccitriletes, Disacciatriletes and Multitaeniate.

Based on the predominance of each species and/or on the joint occurrence of several species and/or on the individual taxa three assemblages can be

← Fig. 3

Palynologically investigated boreholes in SE–Transdanubia. 1. Viganvár Limestone Member; 2. Magyarürög Anhydrite Member; 3. Patacs Siltstone Formation; 4. Jakabhegy Sandstone Formation; e) unit; 5. Jakabhegy Sandstone Formation, d) unit; 6. Jakabhegy Sandstone Formation, c) unit; 7. Jakabhegy Sandstone Formation, b) unit; 8. Jakabhegy Sandstone Formation, a) unit; 9. Rókahegy Dolomite Formation, Middle Triassic; 10. Kővágószőlős Sandstone Formation, Late Upper Permian–Lower Triassic; 11. Cserdi Formation, Early Upper Permian; 12. Korpád Sandstone Formation, Lower Permian; 13. Permian, in general; 14. unconformity boundary; 15. tectonic boundary; 16. number and locality of palynological samples; 17. *Densoisporites nejburgii* palynological assemblage; 18. *Voltziaceaesporites heteromorphus–Triadispora crassa* palynological assemblage; 19. *Triadispora crassa–Stellapollenites thiergartii* palynological assemblage; 20. *Villanyipollis hungaricus* predominating; 21 *Guttatisporites* sp. div. predominating; 22. Alisporites/Falcisporites/ Colpectopollis/ Paravesicaspora predominating; 23. interval of the Scythian/Anisian boundary determined that were named after the quantitatively predominating and first occurring taxa. These are as follows:

The oldest assemblage found in the studied formations is that of the *Densoisporites nejburgii* that derives from the boundary zone of the units c) and e) of the Jakabhegy Sandstone. In its only locality (borehole Vajta–3) the unit d) did not develop. Its lower boundary cannot be determined due to the underlying very thick red palynologically barren rocks.

The assemblage *Voltziaceaesporites heteromorphus–Triadispora crassa* is the second one that is present in the unit d) and in the lower strata of unit e) of the Jakabhegy Sandstone Formation.

The youngest *Triadispora crassa–Stellapollenites thiergartii* assemblage is found in the upper strata of unit e) of the Jakabhegy Sandstone Formation, as well as the older strata of the Patacs Siltstone Formation and of the Magyarürög Anhydrite Member. The upper boundary of its extension cannot be determined in lack of evaluable microflora.

The two older assemblages were unknown at the time of the first publication (1981), the youngest one was termed then as "Palynological Assemblage III".

5.1 Assemblage Densoisporites nejburgii

The palynological assemblage of the Lower Triassic of Southeast Transdanubia known as oldest one so far is found in the sample Nos 1 and 2 of the Jakabhegy Sandstone Formation revealed by the borehole Vajta–3 (Fig. 3 and Table 3).

The assemblage is characterized by the great quantity of spores. Among them the *Densoisporites nejburgii* (Schulz) Balme, the *Densoisporites playfordi* (Balme) Dettman, and *Punctatisporites triassicus* Schulz are most frequent, while the quantities of *Lundbladispora* sp. and *Cyclogranisporites* sp. are smaller. Pollen is represented by disaccites forms: the proportion of *Jugasporites* sp. coming from the Permian and of *Limitisporites moersensis* (Grebe) Klaus is remarkable in the microflora. New and young forms are the *Alisporites cymbatus* Venkatachala, Beju et Kar and *Voltziaceaesporites heteromorphus* Klaus the latter becoming of leading role in the younger assemblages. In traces the *Endosporites* sp. and *Cycadospites* sp. also occur (Table 3).

Comparison and correlation: the *D. nejburgii* assemblage of Southeast Transdanubia is similar to the upper part of the palynozone 14. ("bisaccat-reductum–nejburgii) and/or to the overlying palynozone 15 ("heteromorphus-nejburgii") determined from the Csopak Marl Formation of the Transdanubian Central Range of Hungary containing the transitional forms of *Tirolites cassianus* (Góczán et al. 1986, Broglio Loriga et al. 1990). There is a slight difference between in quantity of the characteristic sporomorphs of the cited assemblages and palynozones of the two regions, i.e. the Early Triassic of the German type Southeast Transdanubia and the Alpine Transdanubian Central Range. An other difference is that in Southeast Transdanubia the assemblage *D. nejburgii* does

Table 3
Qualitative and quantitative distribution of microflora in the Vajta-3 borehole

Species	Symbols of lithostratigraphy units				
	Jh/c/e		Jh/e	P	mHh
	Palynological assemblages				
	D		V–T	T–S	
	sample numbers				
	1	2	3	4	5
	%				
Calamospora sp.	+	+	+	+	+
Cyclogranosporites sp.	5	3	2	2	1
Punctatisporites triassicus	30	22	8	+	1
Densoisporites nejburgii	20	22	3	2	–
Densoisporites playfordi	20	14	+	+	–
Lundbladispora sp.	5	4	+	+	–
Endosporite sp.	+	+	–	–	–
Jugasporites sp.	13	16	19	+	–
Limitisporites moersensis	2	4	+	+	–
Alisporites cymbatus	5	9	22	4	–
Cycadopites sp.	+	+	+	3	+
Voltziaceaesporites heteromorphus	+	6	32	12	33
Triadospora sp. div.	–	–	1	43	22
Angustisulcites klausii	–	–	+	3	2
Angustisulcites gorpii	–	–	+	1	4
Alisporites progradiens	–	–	14	4	2
Alisporites grauvogeli	–	–	–	5	8
Alisporites cf. *ovatus*	–	–	–	6	2
Stellapollenites thiergartii	–	–	–	3	1
Lunatisporites jonkeri	–	–	–	6	2
Lunatisporites multiplex	–	–	–	5	+
Verrucosisporites sp.	–	–	–	1	–
Colpectopollis ellipsoideus	–	–	–	–	4
Alisporites microreticulatus	–	–	–	–	7
Vitreisporites pallidus	–	–	–	–	9
Chordasporites sp.	–	–	–	–	1

Symbols: Jh – Jakabhegy Sandstone Formation, c, d, e units (unnamed members) of Jakabhegy Sandstone Formation; P – Patacs Siltstone Formation; mHh – Hetvehely Dolomite Formation, Magyarürög Anhydrite Member; D – *Densoisporites nejburgii* assemblage; V–T – *Voltziaceaesporites heteromorphus–Triadospora crassa* assemblage; T–S – *Triadospora crassa–Stellapollenites thiergartii* assemblage not contain the *Cyclotriletes presselensis* Schulz spore form and the plankton characteristic of the marine formations and present also in the Triassic of the Transdanubian Central Range, are missing. Authors cited above mark the age of palynozones 14 and 15 determined in the Csopak Marl Formation as Olenekian (Spathian; see Fig. 4).

The palynological composition of the Southeast Transdanubian assemblage is very similar to the microflora of the middle (PII) subzone of *Densoisporites nejburgii* assemblages zone determined by Orlowska-Zwolinska (1984, 1985) in the Middle Buntsandstein strata of Poland, that she put into the Olenikian substage (Fig. 4), where *Gervilleia murchisoni* is the characteristic fauna.

Similar assemblage is reported from Roumania by Antonescu et al. (1976) from the area of the Moesian Platform: from the sandy-argillaceous strata of presumed Spathian age lying in the lower part of the "Facies de Buntsandstein" formation, as well as from the Spathian "Werfener Kalk" series of the "Serie transylvanie" of the Eastern Carpathians containing Tirolitidae, *Eumorphotis telleri* and *Costatoria costata* fauna (Fig. 4).

In this work dealing with the palynostratigraphy of the Late Scythian and Middle Triassic formations of the Alpine-type regions, Brugman (1986) determined six characteristic phases. The *Densoisporites nejburgii* assemblage of Southeast Transdanubia is similar to the phase "heteromorphus–conmilvinus" (upper part of Spathian) of Brugman (Fig. 4) with the only difference that in the Southeast Transdanubian assemblage (and from those discussed below) the pollen Jugasporites conmilvinus Klaus is missing.

In general it can be stated that the assemblages being characterized by the predominance of *D. nejburgii* are well-known in several Spathian series. E.g. similar assemblage is found in Germany, in the Germany-type Middle Buntsandstein (Schulz 1964, 1966), in the Columbites/Tirolites containing strata of Ireland (Visscher 1971), in the European part of the Soviet Union (Bogacheva–Vinogradova 1973 and Yaroshenko–Golubeva 1981) as well as in Pakistan in the Subcolumbites containing strata of the Salt Range series (Balme 1970).

Based on the genera Densoisporites and Lundbladispora the character of the microflora II of Triassic age reported from the Arctic Canada (Fischer 1979) is also similar to the assemblage of Southeast Transdanubia. The age of the Canadian assemblage is defined by the author as Dienerian.

5.2 Assemblage Voltziaceaesporites heteromorphus – Triadispora crassa

The assemblage is found in the d) and e) rock units of the Jakabhegy Sandstone Formation revealed by four boreholes in the Mezőföld, in the Western Mecsek Mountains and in its northern foreground and in the Villány Mountains (Vajta–3, Gf–1, VIII, Mgy–I) (Fig. 3; Tables 3–6). It is found together with the older assemblage discussed above in one profile only in the borehole Vajta–3, in sample No. 3, 12.0 m above the former one.

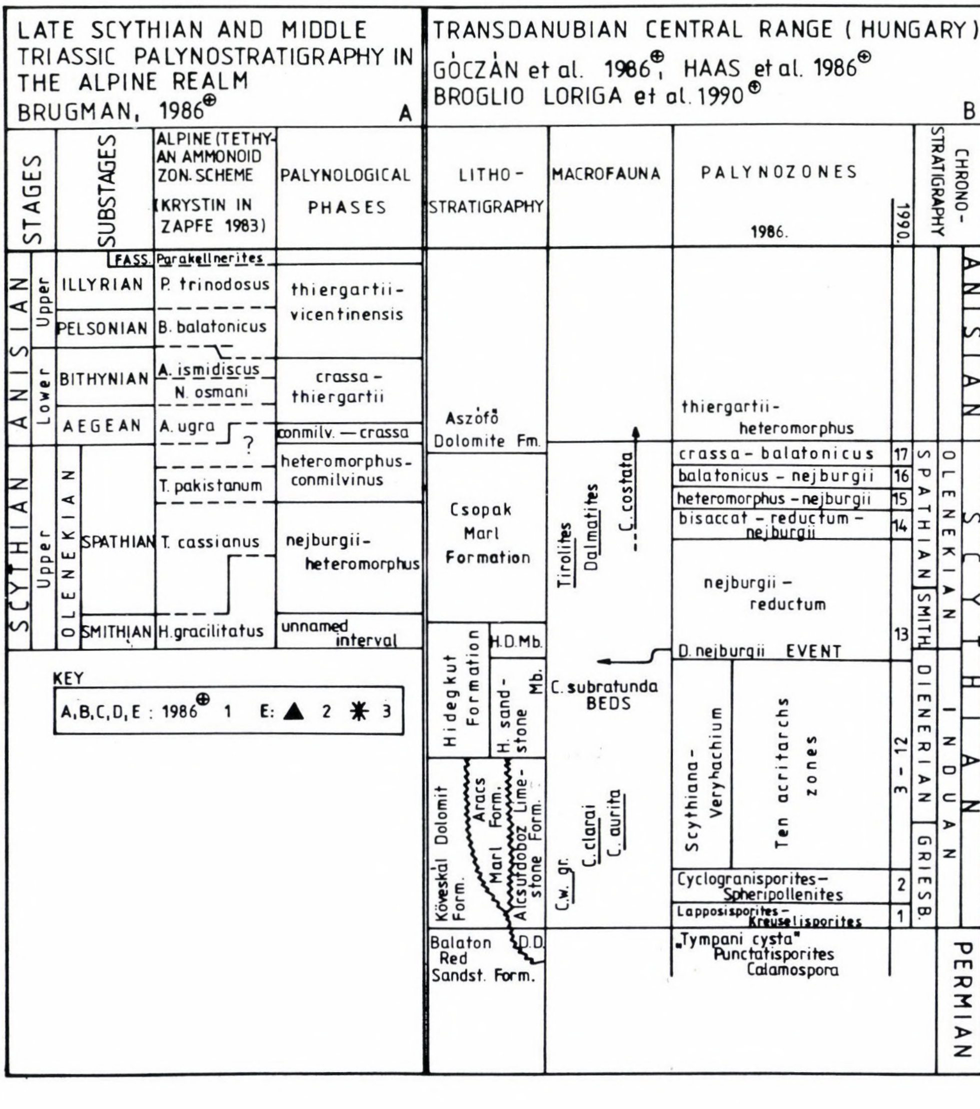


Fig. 4
Correlation of Lower Triassic and Anisian palynological zones from the Alpine Realm, Transdanubian Central Range, SE-Transdanubia (Mecsek–Villány Mountains), Polish epicontinental Triassic and Roumania. A), B), C), D), E) 1. details from the study; 2. *Densoisporites nejburgii*; 3. Triadispora; H.D.Mb. – Hidegkút Dolomite Member; D.D. – Dinnyés Dolomite Formation; Nk.D. – Nádaskút Dolomite Member; C.w. – gr. – *Claraia wangi – griesbachi*

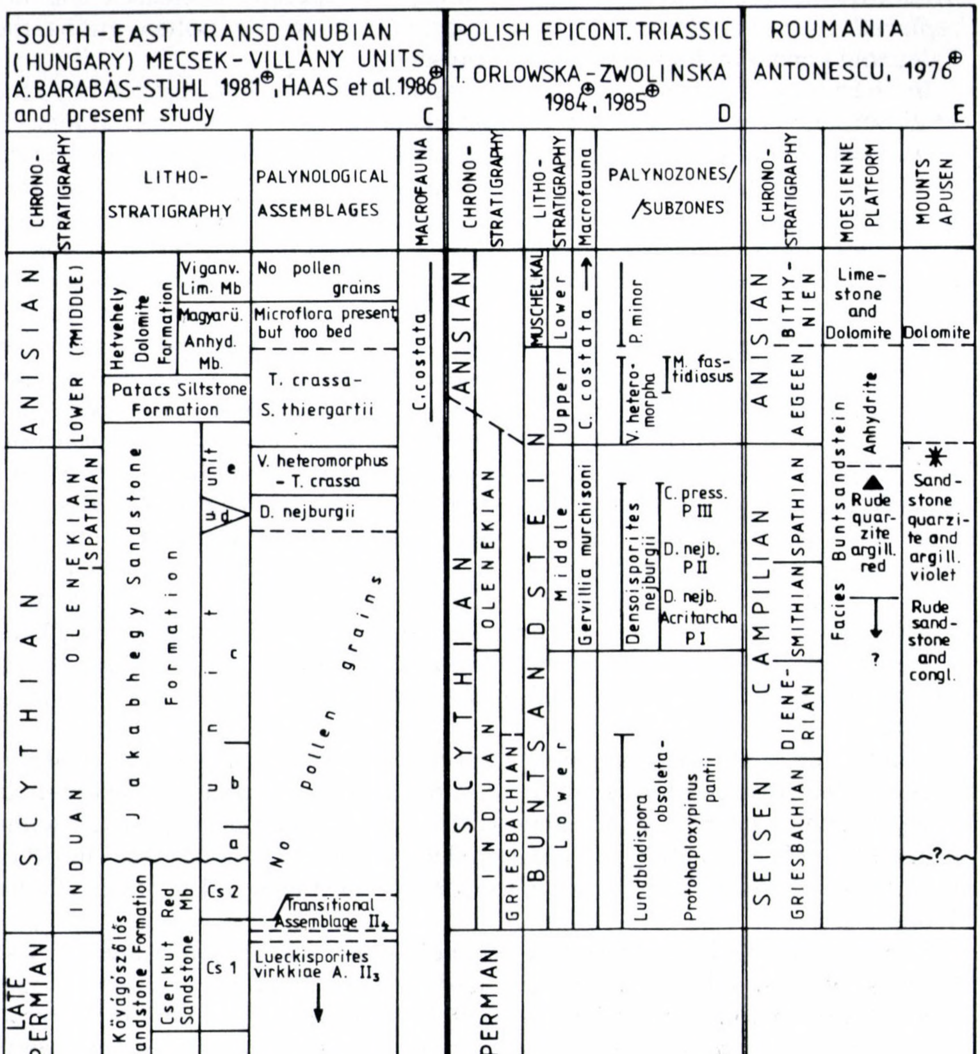

SOUTH-EAST TRANSDANUBIAN (HUNGARY) MECSEK-VILLÁNY UNITS Á. BARABÁS-STUHL 1981⊕, HAAS et al. 1986⊕ and present study
C
POLISH EPICONT. TRIASSIC T. ORLOWSKA-ZWOLINSKA 1984⊕, 1985⊕
D
ROUMANIA ANTONESCU, 1976⊕
E
CHRONO-STRATIGRAPHY
LITHO-STRATIGRAPHY
PALYNOLOGICAL ASSEMBLAGES
MACROFAUNA
CHRONO-STRATIGRAPHY
LITHO-STRATIGRAPHY
Macrofauna
PALYNOZONES/ /SUBZONES
CHRONO-STRATIGRAPHY
MOESIENNE PLATFORM
MOUNTS APUSEN
ANISIAN
LOWER (?MIDDLE)
Hetvehely Dolomite Formation
Viganv. Lim. Mb
Magyarü. Anhyd. Mb.
No pollen grains
Microflora present but too bed
Patacs Siltstone Formation
T. crassa- S. thiergartii
C. costata
SCYTHIAN
OLENEKIAN SPATHIAN
INDUAN
Jakabhegy Sandstone Formation
unit e
d
c
b
a
V. heteromorphus - T. crassa
D. nejburgii
No pollen grains
Cs 2
Cs 1
Transitional Assemblage II4
Lueckisporites virkkiae A. II3
LATE PERMIAN
Kővágószőlős Sandstone Formation
Cserkut Sandstone Red Mb
ANISIAN
SCYTHIAN
OLENEKIAN
INDUAN
GRIESBACHIAN
PERMIAN
MUSCHELKALK
BUNTSANDSTEIN
Lower
Upper
Middle
Lower
C. costata
Gervillia murchisoni
P. minor
V. hetero-morpha
M. fas-tidiosus
Densoisporites nejburgii
C. press. P III
D. nejb. P II
D. nejb. Acritarcha P I
Lundbladispora obsoleta-Protohaploxypinus pantii
ANISIAN
BITHY-NIEN
AEGEEN
CAMPILIAN
SPATHIAN
SMITHIAN
DIENE-RIAN
SEISEN
GRIESBACHIAN
Lime-stone and Dolomite
Dolomite
Anhydrite
Facies Buntsandstein
Rude quar-zite argill. red
?
Sand-stone quarzi-te and argill. violet
Rude sand-stone and congl.
?

Fig. 4 (cont.)

Remarkable change can be experienced in the composition of the *V. heteromorphus–T. crassa* microflora assemblage. The significance of spores that play predominant role in the former assemblages is strongly decreased and are replaced by novel disaccate pollen species, but among pollen grains the Paleozoic genera and species are also present.

In harmony with the term, the assemblage is characterized by the predominance of *Voltziaceasporites heteromorphus* and by the appearance of *Triadispora crassa*. Spores are partly represented by genera and species known from the *D. nejburgii* assemblage: *Densoisporites nejburgii, D. playfordi, Punctatisporites triassicus, Cyclogranisporites* sp., *Lundbladispora* sp. and *Endosporites* sp. New spore forms appearing here are: *Calamospora* sp., *Cyclotriletes oligogranifer* Clarke, *Verrucosisporites* sp., *Kraeuselisporites hoofddijkensis* Visscher. The spore quantity in the boreholes mentioned above is similar: between 6 and 13%.

A part of disaccites pollen found in the palynological spectrum of the assemblage has been present also in the older assemblage. These are: *Jugasporites* sp. and *Limitisporites moersensis*. Diasaccites pollen forms that were subordinate in the former assemblage, now predominate: *Voltziaceaesporites heteromorphus* and *Alisporites cymbatus*. Out of the elements being important in the subsequent Triassic ages, especially in the Middle Triassic the following appear first: the disaccitriletes *Triadispora crassa* Klaus, *Angustisulcites klausii* Freudenthal, *Angustisulcites gorpii* Visscher, out of the Multitaeniate forms the *Lunatisporites* sp., *Lunatisporites jonkeri* Visscher and *Lunatisporites multiplex* Visscher as well as the disaccitriletes *Alisporites progrediens* Klaus, *Alisporites grauvogeli* Klaus, *Alisporites* cf. *ovatus* Jansonius, *Paravesicaspora planderovae* Visscher, *Falcisporites snopkovae* Visscher, *Colpectopollis ellipsoideus* Visscher, *Alisporites microreticulatus* Reinhardt.

In addition to the disaccites pollen species *V. heteromorphus* and *A. cymbatus* only the Multitaeniate forms are present, but these are lacking in the borehole Vajta–3. The quantity of *T. crassa* amounts to several percent in general, while is of extreme high quantity in the assemblage of the borehole Mgy–I. The other new conifer pollens are either in traces or amount to several percents (Tables 3–6).

The assemblage determined from boreholes far from each other are believed to be similar in spite of the experienced smaller differences, i.e. low spore content, the predominance of *V. heteromorphus* and first appearance of *T. crassa*.

Comparison and correlation: the microflora composition of the assemblage is similar to the spectra of palynozones 16. ("balatonicus–nejburgii) and 17. ("crassa-balatonicus") of small vertical extension, determined in the Alpine-type Triassic formations of the Transdanubian Central Range in Hungary (Góczán et al. 1986; Broglio Loriga et al. 1990). The main difference between the compared assemblages is that in the palynozones of the Transdanubian Central Range there are less Triadispora and the *Neojugasporites balatonicus* (?*Jugasporites conmilvinus* Klaus) is missing in the assemblage of Southeast Transdanubia. The palynozones of the Transdanubian Central Range assigned by authors to the

Table 4
Qualitative and quantitative distribution of microflora in the Gálosfa-1 (Gf-1) borehole

Species	Symbols of lithostratigraphy units												
	Jh/e					P					mHh		
	Palynological assemblages												
	V–T		T–S										
	sample number												
	1	2	3	4	5	6	7	8	9	10	11	12	13
	%												
Krauselisporites hoofddijkensis	+	+	–	–	–	–	–	–	–	–	–	–	–
Jugasporites sp.	3	3	–	–	–	–	–	–	–	–	–	–	–
Punctatisporites triassicus	+	1	–	–	–	–	–	–	–	–	–	–	–
Punctatisporites sp.	7	8	+	–	3	–	–	–	–	–	–	–	–
Lundbladispora sp.	+	+	+	+	+	–	–	–	–	–	–	–	–
Alisporites cymbatus	11	2	2	1	+	+	3	–	–	–	–	–	–
Angustisulcites klausii	+	+	3	2	2	1	1	–	–	–	–	–	–
Limitisporites moersensis	1	1	1	1	+	+	+	–	–	–	–	–	–
Cyclogranisporites arenosus	3	1	+	+	+	+	1	–	–	–	–	–	–
Densoisporites nejburgii	3	4	+	+	+	+	1	+	2	–	–	–	–
Verrucosisporites sp.	+	1	+	+	+	+	+	+	+	–	–	–	–
Calamospora sp. div.	3	6	+	1	+	+	+	3	+	4	+	+	+
Lunatisporites multiplex	8	9	5	3	3	1	1	4	1	1	+	+	+
Lunatisporites jonkeri	10	8	3	3	1	2	1	6	+	+	1	+	+
Triadispora sp. div.	5	11	39	43	12	49	46	42	26	42	39	40	39
Voltziaceaesporites heteromorphus	19	16	20	16	17	10	8	6	6	7	3	3	3
Alisporites cf. ovatus	15	14	3	2	2	5	2	3	8	8	7	4	4
Alisporites grauvogeli	2	1	3	5	3	3	14	14	31	21	28	30	30
Colpectopollis ellipsoideus	+	1	8	6	4	3	3	+	2	1	1	3	3
Falcisporites snopkovei	6	2	1	1	2	+	3	4	2	1	3	1	1
Paravesciaspora planderovae	4	3	1	2	+	+	1	+	2	2	1	1	2
Alisporites progrediens	+	+	+	+	5	1	3	2	2	4	6	6	6
Cycadopites coxii	+	1	1	1	+	1	1	+	8	2	4	2	1
Lunatisporites sp.	–	–	1	+	–	+	–	–	–	–	–	–	–
Stellapollenites thiergartii	–	–	4	6	8	5	2	2	4	1	2	2	2
Alisporites microreticulatus	–	–	1	2	+	1	1	+	2	1	–	–	–
Striatoabietites aytugii	–	–	+	–	+	–	–	–	–	–	–	–	–
Apiculatisporites plicatus	–	–	+	+	+	–	–	–	–	–	–	–	–
Guttatisporites guttatus	–	–	+	+	18	+	+	–	–	–	–	–	–
Guttatisporites microechinatus	–	–	+	+	8	+	+	–	–	–	–	–	–
Retusotriletes sp.	–	–	+	+	2	–	–	–	–	–	–	–	–
Angustisulcites gorpii	–	–	+	+	2	1	1	–	–	–	–	–	–
Chordasporites sp	–	–	–	–	–	4	6	12	1	1	4	3	3
Chordasporites magnus	–	–	–	–	–	+	1	+	+	+	1	4	4
Concentricisporites nevesi	–	–	–	–	–	+	–	–	–	–	–	–	–
Lueckisporites junior	–	–	–	–	–	–	–	–	–	–	–	–	+
Succintisporites grandior	–	–	–	–	–	–	–	–	–	–	–	–	+

Symbols: see Table 3

Table 5
Qualitative and quantitative distribution of microflora in the VIII borehole

Species	Symbols of lithostratigraphy units		
	Jh/a	P	mHh
	Palynological assemblages		
	V–T	T–S	
	sample number		
	1	2	3
	%		
Calamospora sp.	4	1	+
Punctatisporites triassicus	4	+	+
Cyclogranisporites sp.	4	–	–
Endosporites sp.	4	–	–
Lunatisporites sp.	17	–	–
Jugasporites sp.	12	–	–
Limitisporites moersensis	3	1	–
Voltziaceasporites heteromorphus	13	11	10
Alisporites cymbatus	25	+	–
Alisporites progrediens	12	+	+
Triadispora sp. div.	1	41	36
Cycadopites coxii	+	2	1
Leschikisporites adunctus	–	2	–
Guttatisporites guttatus	–	+	–
Concentricisporites nevesi	–	+	–
Concentricisporites plurianulatus	–	+	–
Stellapollenites thiergartii	–	8	4
Angustisulcites klausii	–	4	2
Angustisulcites gorpii	–	12	4
Alisporites cf. ovatus	–	1	1
Alisporites grauvogeli	–	+	18
Alisporites microreticulatus	–	3	5
Paravesicaspora planderovae	–	2	2
Colpectopollis ellipsoideus	–	2	6
Chordasporites magnus	–	+	4
Lunatisporites multiplex	–	5	3
Lunatisporites jonkeri	–	3	3

Symbols: see Table 3

Table 6
Qualitative and quantitative distribution of microflora in the Máriagyüd-I (Mgy-I) borehole

Species	Symbol of lithostratigraphy unit Jh/e		
	Palynological assemblages		
	V–T	T–S	
	sample numbers		
	1	2	3
	%		
Calamospora sp.	+	+	+
Punctatisporites triassicus	3	–	–
Densoisporites nejburgii	1	–	–
Densoisporites sp.	+	–	–
Limitisporites moersensis	6	1	–
Jugasporites sp.	5	1	+
Triadispora sp. div.	41	11	16
Stellapollenites thiergartii	+	14	14
Lunatisporites jonkeri	4	2	1
Lunatisporites multiplex	9	3	4
Angustisulcites klausii	+	1	1
Angustisulcites gorpii	+	3	3
Voltziaceaesporites heteromorphus	12	34	31
Alisporites cymbatus	6	–	–
Alisporites grauvogeli	+	8	7
Alisporites microreticulatus	2	+	+
Colpectopollis ellipsoideus	4	2	+
Cycadopites sp.	6	6	7
Villanyipollis hungaricus	–	14	16

Symbols: see Table 3

Olenikian substage are involved by the upper part of the Csopak Marl Formation containing *Costatoria costata* as well as *Meandrospira ammodiscoidea* and *M. gigantea* foraminifers and by the basal strata of the overlying Aszófő Dolomite Formation (Fig. 4).

The comparison with the assemblages of Bundsandstein of Poland is uncertain due to the differences in the microflora. It is our opinion that the spore-pollen compositioin of the lower part of the "*Voltziaceaesporites heteromorpha*" assemblage zone described by Orlowska-Zwolinska (1984) from the Upper Buntsandstein of *Costatoria costata* content and characterized by the lack of *Stellapollenites thiergartii* (syn. *Hexasaccites muelleri)* resembles to the assemblage of Southeast Transdanubia. Author put the Polish assemblage conditionally to the Anisian (Fig. 4).

No microflora suitable to correlation can be recognized in the work by Antonescu et al. (1976) dealing with the Triassic palynology of Roumania. Assemblages describe as "Triadispora Association" are similar rather to the *Triadispora crassa–Stellapollenites thiergartii* assemblage to be discussed in the next part.

The *V. heteromorphus–T. crassa* assemblage of Southeast Transdanubia can be compared with the "conmilvinus-crassa" phase out of the palynological assemblage determined by Brugman (1986) from the Late Scythian and Middle Triassic formations of the Alpine region. The main difference is that the *Jugasporites conmilvinus* Klaus is missing in the assemblage of Southeast Transdanubia but is not present also in the Early Triassic assemblages of the Buntsandstein of Poland and of other German type Triassic areas. Thus, it can be presumed that this form occurs only in the Alpine formations (see also the comparison with the palynozones 16 and 17 of the Lower Triassic of the Transdanubian Central Range). The "crassa-conmilvinus" phase was assigned by Brugman (1986) as a "transitional" phase to the Upper Scythian and to the lower part of the Lower Anisian (Fig. 4).

5.3 Assemblage Triadispora crassa – Stellapollenites thiergartii

The assemblage determined in all boreholes occurred in the unit e) of the Jakabhegy Sandstone Formation and/or in the samples deriving from the Patacs Siltstone and from the Magyarürög Anhydrite Member of the Hetvehely Dolomite Formations (Fig. 3 and Tables 3–8).

It is seen in the figure that the assemblage is contained by the boreholes Gf–1 and VII within the greatest stratigraphic distance. Together with the older *V. heteromorphus–T. crassa* assemblage it was found only in four borehole in one section, the distance between them is 34.0 m (Vajta–3), 30.0 m (Gf–1), 137.0 m (VIII) and 14.0 m (Mgy–I borehole in the Villány Mountains). These distances are, of course, determined by the occurrence of palynologically positive samples. Its extension upwards can be traced up to the lower part of the Magyarürög Anhydrite Member. Upwards, i.e. in the upper part of the member above and in the Viganvár Limestone Member no evaluable microflora is found, thus the upper boundary of the *T. crassa – S. thiergartii* cannot be determined.

Characterization of the assemblage: it is characterized by the predominance of *Triadispora crassa* and the appearance of *Stellapollenites thiergartii* that play important role up to the end of the Anisian, as well as by the disappearance of Paleozoic disaccites pollen forms.

Spores are represented by forms known from the older assemblages: *Densoisporites nejburgii*, *Punctatisporites triassicus*, *Verrucosisporites* sp., *Calamospora* sp. and *Cyclogranisporites* sp. Relatively much spores appear in the assemblage though in restricted number of individuals: *Punctatisporites* sp., *Apiculatisporites plicatus* (Visscher) Orlowska-Zwolinska, *Cyclotriletes microgranifer* Mädler, *Cyclotriletes arenosus* Mädler, *Leschikisporis aduncus*

Table 7
Qualitative and quantitative distribution of microflora in the VII borehole

Species	Jh/e	P			mHh
	Palynological assemblages T–S				
	sample numbers				
	1	2	3	4	5
	%				
Calamospora sp.	2	1	4	4	+
Punctatisporites sp.	1	–	–	–	–
Punctatisporites triassicus	6	–	–	–	–
Leschikisporites aduncus	3	2	2	2	+
Apiculatisporites plicatus	–	–	–	–	–
Verrucosisporites sp.	1	–	–	–	–
Densoisporites nejburgii	1	1	1	1	–
Guttatisporites guttatus	12	+	–	–	–
Guttatisporites microechinatus	8	–	–	–	–
Limitisporites moersensis	3	1	–	–	–
Triadispora sp. div.	10	37	30	33	36
Stellapollenites thiergartii	6	16	2	2	6
Lunatisporites jonkeri	5	4	3	1	4
Lunatisporites multiplex	7	5	3	3	2
Lunatisporites sp.	1	+	+	–	–
Angustisulcites klausii	3	2	1	3	2
Paravesicaspora planderovae	4	2	+	+	2
Alisporite cf. ovatus	+	1	+	1	1
Voltziaceaesporites heteromorphus	14	8	10	9	11
Cycadopites coxii	2	2	2	2	1
Cyclotriletes sp.	+	+	+	+	+
Retitriletes jenensis	–	+	+	+	+
Concentricisporites nevesi	–	+	–	–	–
Concentricisporites plurianulatus	–	+	–	–	–
Angustisulcites gorpii	–	10	8	10	2
Alisporites grauvogeli	–	+	17	16	15
Alisporites microreticulatus	–	3	4	4	5
Colpectopollis ellipsoideus	–	4	10	6	8
Chordasporites sp.	–	+	2	2	4
Succintisporites grandior	–	–	+	+	+
Vitreisporites pallidus	–	–	+	3	+
Lueckisporites junior	–	–	+	–	+
Saturnisporites praevius	–	–	–	–	+

Symbols: see Table 3

Table 8
Qualitative and quantitative distribution of microflora one or two samples containing in boreholes

Species	Symbols of lithostratigraphy units					
	Jh/e			P		
	Palynological assemblages T–S					
	Boreholes and sample numbers					
	3155	Mk-3	Smb-1	Nk-2		Bt-3
	1	1	1	1	2	1
	%					
Calamospora sp.	4	2	+	+	1	+
Punctatisporites sp.	1	–	–	+	+	+
Punctatisporites triassicus	2	1	–	–	–	–
Leschikisporis aduncus	3	–	–	–	–	–
Cyclotriletes micrograni fer	2	–	–	+	–	–
Cyclogranisporites sp.	+	+	+	–	–	2
Apiculatisporites plicatus	2	–	–	–	–	–
Verrucosisporites sp.	1	–	–	–	–	–
Densoisporites nejburgii	1	1	+	–	–	–
Guttatisporites guttatus	22	+	3	+	1	3
Guttatisporites microechinatus	12	–	–	–	–	–
Limitisporites moersensis	1	3	3	–	–	2
Triadispora sp. div.	5	37	41	35	30	32
Stellapollenites thiergartii	10	12	16	3	7	10
Lunatisporites jonkeri	4	6	3	10	9	4
Lunatisporites multiplex	6	4	7	15	10	7
Angustisulcites klausii	5	2	–	1	2	1
Angustisulcites gorpii	–	3	3	3	5	3
Voltziaceaesporites heteromorphus	10	15	11	6	4	15
Alisporites progrediens	2	–	–	3	3	–
Alisporites microreticulatus	–	–	–	3	2	+
Alisporites grauvogeli	2	7	6	4	4	4
Alisporites cf. ovatus	2	+	–	5	4	2
Colpectopollis ellipsoideus	–	4	3	3	8	9
Paravesicaspora planderovae	2	–	–	–	–	–
Falcisporites snopkovei	–	–	–	4	+	3
Chordasporites magnus	–	–	–	3	5	–
Cycadopites coxii	1	2	3	2	2	3

Symbols: see Table 3

(Leschik) Potonie, *Saturnisporites praevius* Visscher, *Saturnisporites* sp., *Verrucosisporites morulae* Klaus, *Guttatisporites microechinatus* Visscher, *Guttatisporites guttatus* Visscher, *Concentricisporites plurianulatus* Antonescu, *Concentricisporites nevesi* Antonescu. The quantity of each spore genera and species varies between 0.1 and 5 percent, except the species *G. guttatus* and *G. microechinatus* that occur in the older samples of the boreholes VII, 3155 and Gf–1 in quantities between 20 and 30%, i.e. in very high proportion (Tables 4, 7, 8). In younger samples these are present only in traces, later disappear.

Concentricisporites nevesi and *C. plurianulatus* are to be mentioned out of the spore species appearing as new forms of the assemblage, these are important elements of the Anisian microflora of the German Middle Triassic. These species were found only in traces in the samples of boreholes VIII/2, VII/2 and Gf–1/7.

Among the disaccites pollens the *Triadispora crassa* predominates, there are less *Triadispora plicata* Klaus, the quantities of *Triadispora epigona* Klaus and *Triadispora staplini* Klaus are only several percents. In the lowermost sample of the borehole Gf–1 involving the assemblage within greates stratigraphic distance the total quantity of Triadispora species is 12%, above this sample 35–45% (Table 4). This predominating value is characteristic also of the Triadispora content of the assemblages determined in other boreholes.

The other predominating disaccites pollen of the assemblage appearing first here, the *Stellapollenites thiergartii* (Mädler) Clement–Westerhof et al. can be found in all samples between 1 and 10%. In the Gálosfa section its quantity shows decreasing tendency upwards (Table 4). In the assemblage of some samples it shows extreme high quantities: 16% in the sample 2 of borehole VII, 14% in that of 2 of borehole Mgy–I and 16% in that of 1 of borehole Smb-1.

Other disaccites pollen forms being present in the former assemblage are: *Lunatisporites jonkeri, Lunatisporites multiplex, Limitisporites moersensis, Angustisulcites gorpii, A. klausii, Alisporites gorpii, A. klausii, Alisporites progrediens, A. grauvogeli, Alisporites cf. ovatus, A. microreticulatus, Falcisporites snopkovei, Colpectopollis ellipsoideus, Paravesicaspora planderovae, Voltziaceaesporites heteromorphus.*

Cycadopites sp. is present also in the microflora and the monocolpate pollen species *Cycadopites coxii* Visscher also appears though in a quantity of less than 5%.

The monosaccate *Villanyipollis hungaricus* nov. gen. and sp. is a new pollen appearing in the assemblage, that can be found only samples 2 and 3 of the Mgy–I borehole (Table 6) though in high proportion (16%). The new disaccites pollens appearing here are represented by *Striatoabietites aytugii* Visscher, *Lunatisporites* sp., *Succintisporites grandior* Leschik, *Lueckisporites junior* Klaus, *Chordasporites magnus* Visscher. *Vitreisporites pallidus* (Reissinger) Nilson occurs in the assemblage of the upper samples of boreholes Vajta–3 and VII, in a quantity of 7 to 9%.

The quantity of *Multitaeniate disaccites* pollen forms is more (8 to 10%) in the older and much less (1 to 6%) in the younger samples. The role of

V. heteromorphus plays predominant role: its quantity is 10–20% in the older and 1–2% in the younger samples in the samples series of the borehole Gf–1 (Table 4).

The feature of microflora composition is worthy of mention, i.e. the quantities of Alisporites species out of the Disacciatriletes group, of the *Colpectopollis ellipsoideus, Falcisporites snopkovei* and *Paravesicaspora planderovae* are rather high in the younger strata, between 20 and 40% (samples 7–13, Gf–1; see Table 5). The accumulation of smaller degree of the same forms can be observed in the boreholes VII, VIII and Nk–2 (Tables 5, 7, 8).

Chordasporites magnus and *Chordasporites* sp. occur only in the younger samples (samples 7–13 in Gf–1, Table 4; sample 3 of 8 and samples 4–5 of 7, Tables 5–8), while *Succinisporites grandior* and *Lueskisporites junior* were found only in the uppermost sample of the borehole Gf–1.

In relation with the *Villanyipollis hungaricus* nov. gen. and sp. known in the assemblage *T. crassa–S. thiergartii* of the borehole Mgy–I of the Villány Mountains but lacking in others of the area, it is to be noted that this monosaccate pollen genus is mentioned by some authors only from the region of the German Basin, from the Upper Buntsandstein. E.g. Orlowska-Zwolinska mentioned (1984) from Western Poland from the "*Voltziaceaesporites heteromorpha*" assemblage zone and indicated as *Accintisporites* sp. in the middle of the Rőt (OTYN IG–1 profile, samples 53 and 56, PL 20/6). Here it occurs together with *S. thiergartii* (syn. *Hexapollenites muelleri)* but is missing in the younger samples, i.e. it is a shortlife form, Schulz (1965) described the form *Accintisporites diversus* Leschik from the strata of the Triassic Upper Buntsandstein of Thüringen (Tafel XII/8). The *Enzonalasporites leschiki* Mädler, described by Mädler from the Rőt-strata of NW-Germany is a form similar to the *Accintisporites* sp. that is missing in the younger strata (Muschelkalk, Keuper) demonstrated by author's table. The *Villanyipollis hungaricus* defined in the borehole Mgy–I is similar to the forms listed above, but surely represents a new sporomorph since it is much greater than those and its saccat is narrower (Plates V, VI, XIV). Its description as new genus and species can be found in this study.

Comparison and correlation: the assemblage can be fairly well correlated with the palynological zone "crassa-thiergartii" described from the Aszófő Dolomite and overlying Iszkahegy Limestone Formations of the Alpine-type Transdanubian Central Range in Hungary, the age of which was determined as Anisian by Góczán et al. (1986) (Fig. 4). In some sections the Aszófő Dolomite contains *Costatoria costata* macrofauna, as well as *Glomospirella senghi, Glomospirella* div. sp. and *Meandrospira gigantea* foraminifers (Góczán et al. 1986, Broglio Loriga et al. 1990).

The *T. crassa–S. thiergartii* assemblage of Southeast Transdanubia can be fairly well correlated with the "crassa–thiergartii" phase described by Brugman in his monography on the Permian–Triassic palynology and published in 1986 on the palynology of the Alpine Late Scythian and Middle Triassic, the age of which was indicated as Lower Anisian (Aegean–Bithynian) by the author (Fig. 4).

The *T. crassa–S. thiergartii* assemblages is very similar to the *"Voltziaceaesporites heteromorpha"* assemblage zone determined in the Rőt-strata characterized by *Costatoria costata* fauna, deriving from the uppermost lithological unit of the Buntsandstein profiles of Western Poland, and palynologically processes by Orlowska-Zwolinska (1984, 1985). Nevertheless, in the assemblage of Southeast Transdanubia known so far the *Microcachrydites fastidiosus* (Jansonius) Klaus, *Microcachrydites doubingeri* Klaus and *Microcachrydites sittleri* Klaus disaccites pollen forms being important in the upper part of the *"Voltziaceaesporites heteromorpha"* assemblage of Poland, are unknown. Orlowska–Zwolinska defined the age of the *"Voltziaceaesporites heteromorpha"* assemblage zone conditionally as Anisian (1984), later in her work on the epicontinental Polish Triassic (1985) she indicated an alternative to the Scythian/Anisian boundary: the first lies between the *Densoisporites nejburgii* and *Voltziaceaesporites heteromorpha* zones, the second lies in the upper part of the *V. heteromorpha* zone where *Microcachrydites fastidiosus* appears (Fig. 4).

The microflora composition described by Antonescu et al. (1976) from Roumania as "Triadispora Association" is very similar to the assemblage of Southeast Transdanubia. The "Triadospora Association" was described from the Mounts Apuseni (Bihor Autochthon, sandstone series),(Fig. 4), as well as from calcareous sandy series of Anisian age with *Gervilleai modiola* and *Costatoria costata* of serie bucovinienne and serie de Brasov in Carpathes Orientales.

Similar assemblages are found in several European countries, mainly in the regions of German type: in Germany (Schulz 1965, 1966; Mädler 1964, 1968; Reinhardt and Schmitz 1965; Doubinger and Büchman, 1981), in the Netherlands (Freudenthal 1964; Visscher 1966), in France (Adloff and Doubinger 1969), in England (Warrington 1973). Klaus (1964) described similar assemblages from the Alpine Werfen shales, outside the German regions.

Concerning the regions outside Europe, Fisher (1979) reported similar assemblage from the Arctic Canada as assemblage IV of the microflora of the Triassic of that region, the age was indicated in harmony with the stratigraphic scale of Tozer (1969) as Spathian.

Brugman, in his work of 1983 noted that the *Triadospora* sp. div. was a very important form of general extension. Triadispora species were described in North Africa, in the USA, in Australia and in Pakistan in addition to those mentioned above.

6. Stratigraphic conclusions

Many stratigraphers are of the opinion that the chronostratigraphic classification of lithological units containing only microflora cannot be performed only on palynostratigraphic bases. Nevertheless, the chronological classification of the fauna-free or fauna-poor Early Triassic formations that have sporomorph remains is favourized by the fact that from the Late Permian to the Middle Triassic rapid and considerable changes followed in the vegetation both in regional and on global scale on the basis of which attempts can be made to mark the boundaries between systems and stages and the series can be chronologically classified even in lack of index fossils. When solving this task, however, the correlation with lithological units has to be searched that contain both fauna and microflora.

The remarkable changes of vegetation mentioned above are reflected also in the microflora of the Triassic formations of Southeast Transdanubia though the possibility of absolutely correct evaluation is affected by the objective condition that the recognized palynological assemblages including the youngest Permian ones are not found in one profile. This is why special attention was paid to the comparisons and correlations when making the chronostratigraphic classification. These were discussed in detail in the previous chapters.

The biostratigraphic and chronostratigraphic profiles providing the most complete correlation possibilities on the Lower Triassic and Early Middle Triassic are compared with the palynological data of the assemblages of Southeast Transdanubia in Fig. 4. Among these the study dealing with the biostratigraphic zones of Lower Triassic of the Transdanubian Central Range of Hungary (Góczán et al. 1986) is of special importance that provided aid to subdivide the Scythian stage into substages, to mark the P/Tr boundary and to draw the Scythian/Anisian boundary on the basis of cores sampled by one meter in boreholes and of the correct biostratigraphic evaluation of the micro- and macrofauna and palynological zones.

The other work used to the correlation is the profile of Brugman (1986) on the Late Scythian and Middle Triassic palynostratigraphy of the Alpine regions, in which the scheme of Ammonoid zones of Krystyn (in Zapfe 1983) was also demonstrated in addition to the palynological phases.

Important correlation data were provided by the works of Orlowska-Zwolinska (1984, 1985) on the Triassic palynological zones of Poland of German type, being rather similar to the Triassic of Southeast Transdanubia.

Prior to the circumstantial stratigraphic evaluation it is to be noted that based on faunal data the Scythian is subdivided into three or four substages, i.e. Griesbachian, Dienerian, Smithian and Spathian. On the contrary, in the Transdanubian Central Range the Lower Triassic is bipartite. In its lower part the organic-walled microplankton and poorish terrestrial sporomorph assemblage dominate, in which the *D. nejburgii* appears suddenly at about the upper boundary of the Hidegkút Sandstone Member indicating the remarkable change of terrestrial vegetation. This is why Góczán et al. (1986) suggested to subdivide the Lower Triassic series into two chronostratigraphic units: Induan and Olenekian substages.

The oldest assemblage known in the Triassic of Southeast Transdanubia is that of the *Densoisporites nejburgii*, contained by the uppermost lithological units (boundary zones of the units c and e) of the Jakabhegy Sandstone Formation (borehole Vajta–3, Fig. 3). Based on *D. nejburgii* predominating in the microflora composition of the assemblage and on other in mass occurring pterydophytes spores and surviving Paleozoic and appearing new Mesozoic disaccites pollen grains, this assemblage belongs to the base of the upper part of the Olenekian substage of the Scythian, in case of multipartite classification to the upper third of the Spathian, according to the correlations performed under chapter 5 (Figs 4 and 5).

Since the *D. nejburgii* assemblage was found in Southeast Transdanubia only in two samples of the borehole Vajta–3 very close to each other and the red strata below the samples are palynologically barren thus it is impossible to determine the palynological bipartiteness of the Lower Triassic, i.e. to determine the Induan/Olenekian boundary.

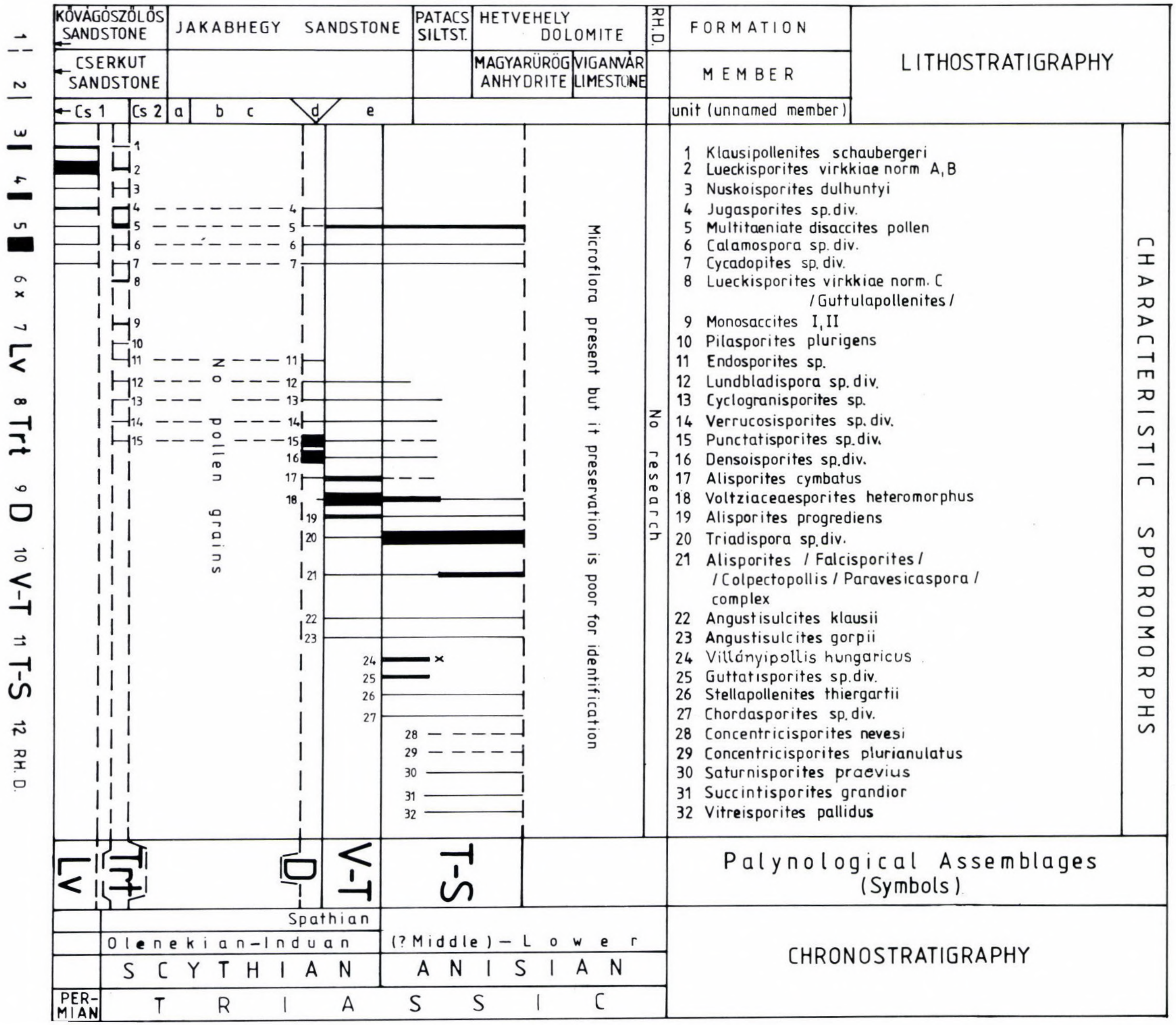
The *Voltziaceaesporites heteromorphus–Triadispora crassa* assemblage lying upwards is contained also by the upper units of the Jakabhegy Sandstone Formation (Figs 3 and 5). Its palynological composition considerably differs from that of the *D. nejburgii* assemblage. It contains less spores, the young disaccites pollen species are predominating but the Paleozoic-type pollen species are also present. As to the correlations (chapter 5.2) this assemblage belongs to the Olenekian substage or to the uppermost Spathian, i.e. it represents the youngest palynological assemblage of the Scythian.

The assign the assemblages *D. nejburgii* and *V. heteromorphus–T. crassa* to the upper part of the Olenekian (Spathian) is supported orthostratigraphically by the fact that both assemblages could be correlated with palynological assemblages that derive from strata containing *Tirolites cassianus* (Alps) or the transitional forms of this species (Transdanubian Central Range, Hungary) or from strata above this horizon.

The youngest *Triadispora crassa–Stellapollenites thiergartii* assemblage can be identified in all the studied boreholes, but together with the older assemblage/assemblages it is present only in one section (Fig. 3). In boreholes Gf–1 and VII it can be found in several samples within great stratigraphic distances, this relates either to the stabilization of the vegetation or to the diachron character of the formations, to prove this latter presumption, however, no sufficiant fossils are available.

In this assemblage the Paleozoic element are completely missing, the young forms predominate, out of them the *S. thiergartii* and Concentricisporites genera as well as the *Villanyipollis hungaricus* appear first here (Fig. 5). Góczán et al. (1986) stated that in the fossiliferous Triassic strata the *S. theirgartii* disaccites pollen species occurs from the Early Anisian up its end in the Alpine and German type formations, indicating a change in evolution that can be used to mark the Scythian/Anisian boundary.

In correlation regions (chapter 5.3) similar assemblages are considered as oldest palynozones of the Anisian and in Alpine formations it is correlated also with Ammonoid zones (Fig. 4). On this basis the *Triadispora crassa–Stellapollenites thiergartii* assemblage represents the oldest palynological


KŐVÁGÓSZŐLŐS SANDSTONE
JAKABHEGY SANDSTONE
PATACS SILTST.
HETVEHELY DOLOMITE
RH.D.
FORMATION
CSERKUT SANDSTONE
MAGYARÜRÖG ANHYDRITE
VIGANVÁR LIMESTONE
MEMBER
LITHOSTRATIGRAPHY
Cs 1
Cs 2
a b c d e
unit (unnamed member)
No pollen grains
Microflora present but it preservation is poor for identification
No research
1 Klausipollenites schaubergeri
2 Lueckisporites virkkiae norm A, B
3 Nuskoisporites dulhuntyi
4 Jugasporites sp.div.
5 Multitaeniate disaccites pollen
6 Calamospora sp. div.
7 Cycadopites sp. div.
8 Lueckisporites virkkiae norm. C /Guttulapollenites/
9 Monosaccites I, II
10 Pilasporites plurigens
11 Endosporites sp.
12 Lundbladispora sp. div.
13 Cyclogranisporites sp.
14 Verrucosisporites sp. div.
15 Punctatisporites sp. div.
16 Densoisporites sp. div.
17 Alisporites cymbatus
18 Voltziaceaesporites heteromorphus
19 Alisporites progrediens
20 Triadispora sp. div.
21 Alisporites /Falcisporites/ /Colpectopollis/ Paravesicaspora/ complex
22 Angustisulcites klausii
23 Angustisulcites gorpii
24 Villányipollis hungaricus
25 Guttatisporites sp. div.
26 Stellapollenites thiergartii
27 Chordasporites sp. div.
28 Concentricisporites nevesi
29 Concentricisporites plurianulatus
30 Saturnisporites praevius
31 Succintisporites grandior
32 Vitreisporites pallidus
CHARACTERISTIC SPOROMORPHS
Lv
Trt
D
V-T
T-S
Palynological Assemblages (Symbols)
Spathian
Olenekian-Induan
(?Middle)-Lower
SCYTHIAN
ANISIAN
PERMIAN
TRIASSIC
CHRONOSTRATIGRAPHY
1
2
3
4
5
6 x
7 Lv
8 Trt
9 D
10 V-T
11 T-S
12 RH.D.

assemblage of the Middle Triassic in Southeast Transdanubia. Nevertheless, the Scythian/Anisian boundary can be marked only in sections where the *V. heteromorphus–T. crassa* indicating the end of Scythian and the *T. crassa–S. thiergartii* indicating the start of Anisian are present not far from each other.

In Southeast Transdanubia there two profiles of this type, in the boreholes Gf–1 and Mgy–I where the Scythian/Anisian boundary can be marked within the uppermost e) lithological unit of the Jakabhegy Sandstone Formation between the assemblages V. *heteromorphus–T. crassa* and *T. crassa–S. thiergartii* lying in a distance of 11.0 and 14.0 m from each other (Fig. 3). Consequently, the strata of the Jakabhegy Sandstone above this boundary as well as the Magyarürög Anhydrite Member of the Hetvehely Dolomite Formation and the Patacs Siltstone Formation belong to the Lower Anisian, up to the appearance of palynological data. Nevertheless, due to the lack of sporomorph species indicating the upper part of the Anisian (e.g. *Dyupetalium vicentinensis* Brugman) the presence of Middle Anisian can be excluded neither since this form can be found in other areas only in small number of individuals and only in fragments. At the same time, the *Concentricisporites* species appearing in the Triassic of Southeast Transdanubia, together with the predominance of Triadispora may indicate the Middle Anisian, too, further the *S. thiergartii and T. crassa* occur together throughout the Anisian (Figs 4 and 5).

Concerning the sections studied in Southeast Transdanubia, in the boreholes VIII and Vajta–3 the distance between the two assemblages determining the Scythian/Anisian boundary is rather great, so it can be stated only that this boundary lies between the assemblages *V. heteromorphus–T. crassa* of the e) unit of the Jakabhegy Sandstone and *T. crassa–S. thiergartii* of the Patacs Siltstone Formation (Fig. 3).

In the sections where due to the great number of palynologically barren samples only the *T. crassa – S. thiergartii* assemblage was found (VII, 3155, Nk–2, Mk–3, Smb–1, Bt–3) it can be only stated that the Scythian/Anisian boundary is found deeper than the sample containing this assemblage. Since in three of these boreholes (3155, VII and Smb–1) the Anisian assemblage is present in the e) unit of the Jakabhegy Sandstone Formation, it can be presumed that in the Triassic of the whole Southeast Transdanubia the Scythian/Anisian boundary can be marked within the uppermost unit of this formation.

In harmony with the facts discussed above, the oldest studied formation, the upper units of the Jakabhegy Sandstone belong partly to the upper part of the Lower Triassic Olenekian (Spathian) substage, partly to the Early Middle

←Fig. 5
Late Upper Permian, Lower Triassic and Early Anisian palynological assemblages of SE–Transdanubia (generalized on the basis of analysis of 36 samples). 1. present, 0.5%; 2. rare, 1–5%; 3. frequent, 6–15%; 4. common, 16–20%; 5. abundant, 25%; 6. higher on no data; 7. *L. virkkiae* assemblage (II_2–II_3; 1981); 8. "transitional" assemblage (II_4; 1981); 9. *D. nejburgii* assemblage; 10. *V. heteromorphus–T. crassa* assemblage; 11. *T. crassa–S. thiergartii* assemblage; 12. Rókahegy Dolomite

Triassic, i.e. to the Lower Anisian. The chronostratigraphic classification of the older and paleontologically barren units of this formation is supported by information on the results concerning the first palynological studies of the Kővágószőlős Sandstone Formation (Barabás-Stuhl 1981). In the Western Mecsek Mountains (borehole 5071), in the upper part of the so-called Cserkút Member, the youngest member of the Kővágószőlős Sandstone, below the basal strata of the Jakabhegy Sandstone Formation by 60 m, a "transitional palynological assemblage" occurred in a thin green fine-grained sandstone layer that considerably differed from the assemblages characteristic of the typical Upper Permian Kővágószőlős Sandstone (Barabás-Stuhl 1981, II_2–II_3 P.A., pp. 64–65). This is characterized, in addition to the cease of predominance of Upper Permian disaccites pollen species — mainly of *Lueckisporites virkkiae* Pot. et Klaus – the appearance of large amounts (25%) of Lower Triassic pterydophyte spores (Barabás-Stuhl 1981, II_4 P.A. pp. 65–67 and this study Fig. 5).

The spores of the "transitional" assemblage, the taxa of Multitaeniate pollens (e.g. *Protohaploxypinus* sp.) are index forms of the oldest microflora assemblages of the Lower Triassic. Such assemblages – with traces or lack of *L. virkkiae* characteristic of the Upper Permian – are known from the Griesbachian of the Lower Triassic: in West Canada (Jansonius 1962), in East Greenland (Balme 1979) as "Protohaploxypinus Association" around strata with Otoceras boreale, in Western Poland in the Lower Buntsandstein as "*Lundbladispora obsoleta – Protohaploxypinus pantii*" assemblage zone (Orlowska-Zwolinska 1984, 1985) and in North China in a continuous Permian–Triassic sequence of continental facies, as the microflora of the so-called "transitional strata" in the Goudikeng Formation (Yang Juduan et al. 1984). From our aspect the "Lapposisporites–Kraeuselisporites" assemblage zone is very important (Góczán et al. 1986, Haas et al. 1988, Broglio Loriga 1990) demonstrated from the Alcsútdoboz Limestone Formation and from basal strata of its heteropic facies (Köveskál Dolomite Formation) in the Triassic of the Transdanubian Central Range of Hungary and being rather similar to the "transitional" assemblage of Southeast Transdanubia, which is considered by authors as the oldest palynozone (1) of the Lower Triassic (Fig. 4).

The great amount of pterydophyte spores appearing suddenly in the "transitional" palynological assemblage of Southeast Transdanubia indicate the sudden change of vegetation that, together with correlation data, is suitable to mark the boundary between the Permian and Triassic systems. Accepting this, in the units of the studied area where the Jakabhegy Sandstone overlies the Kővágószőlős Sandstone Formation (Western Mecsek and its northern and southern foreground), the Permian/Triassic boundary lies at the upper part of the Cserkút Member of the Formation mentioned above (in the boundary zone of the Cs–1 and Cs–2 units, Fig. 5), thus the uppermost Cs–2 unit of the Cserkut Member and the slightly unconformly overlying Jakabhegy Formation, i.e. its lower paleontologically barren strata belong to the Induan and older Olenekian substages of the Lower Triassic (Scythian), and to the substages older than Spathian, or even to the Lower Spathian (Figs 4 and 5).

In relation with the marking of the P/Tr boundary within the Cserkút Member on palynostratigraphic bases it is to be noted that the Member is fairly well associated with the rapid changes followed in the litho- and tectonofacies of the sedimentary sequence. That is, above fluviatile (channel – flood plain – paludal) sediments of rapid lateral–vertical changes and characteristic of the lower strata of the Kővágószőlős Sandstone, uniformly developed and lithologically very characteristic alluvial fan sediments are found throughout the whole Upper Permian basin (Cs–2). This is followed with a small hiatus, then follows with erosion unconformity the Jakabhegy Sandstone Formation, the transgression of which is most significant geohistorical event at about the Permian/Triassic boundary. Based on the diastorphic change that can be experienced in the recent practical geological works, the P/Tr boundary is drawn.

The sudden disappearance of pterydophyte spores appearing in the Late Permian and fluorishing in lower part of the Triassic as well as the appearance of the new-type bisaccate pollens can be related to the cease of the terrestrial clastic environs and to the predominance of the marine environment in the region studied.

7. *Description of the new taxa*

Villanyipollis nov. gen.

Derivatio nominis: occurrence in Villány Mts, S. Hungary

Genus typus: Villanyipollis hungaricus nov. gen. et sp.

Genus diagnosis: Protomonosaccat alete pollen grains, consisting of central body and protomonosaccus. The grains are discus-like in E-plain. The outline of the exine is roundish or oval. The central body is covered entirely by the protomonosaccus. C-axis of exine is short. Unilateral compression during diagenesis led to the formation of one or more secondary folds, ranning along the outline. The intexine of the central body is mostly contracted. Sculpture of ektexine is irregular network-like and fine granulate or punctate. Sculpture of the body is infragranulate or infrapunctate.

Differential diagnosis: Villanyipollis nov. gen. similar to genera Enzolanasporites, Vallasporites and Patinasporites Leschik 1956, classified by him as alete Monosaccites. Villanyispollis nov. gen. differs from the above listed genera in the sculpture of protomonosaccus and in the connection to the body.

Remarks: The genus diagnosis of Accintisporites Leschik 1956 (p. 48) is very close to Villanyipollis nov. gen., but not to the genus typus *Accintisporites ligatus* Leschik 1956 (p. 49, Table 6, Fig. 17). Most similarity can be recognized between genus types of Villanyipollis and *Accintisporites angustus* Leschik 1956. Based on this similarity it seems that proposition of a new genus is not justified. But

Scheering in 1974 (pp. 207–209) certified undoubtedly the identity of genus typus of Accintisporites with *Lunatisporites acutus* Leschik 19956. So validity of Accintisporites is over. Therefore proposition of Villanyipollis nov. gen. is necessary for our specimens. At the same time we propose a new combination too: *Villanyispollis angustus* (Leschik 1956) nov. comb.

Villanyipollis hungaricus nov. gen. et nov. sp.

Plate V, Photos: 1–2
Plate VI, Photos: 1–4
Derivatrio nominis: occurrence in Hungary
Holotypus: grain in slide No Po–2619/OX; 21.0–117.6, Plate V: 1–2.
Locus typicus: Borehole Máriagyüd–I. (S. Hungary)
Stratum typicum: 1175.2 m; dark gray siltstone, Patacs Formation; Lower Anisian
Diagnosis: identical with genus diagnosis.
Sizes of holotype: 137 µm in diameter;
thickness of saccus: 10 µm
thickness of body: about 3 µm
breadth of lumina of network: 1.5–3 µm
Remarks: Size of specimens in diameter varies between 102 to 150 µm. Greatest frequency of *Villanyipollis hungaricus* nov. gen. et sp. is in the stratotype.

Plate I

1. *Calamospora tener* (Leschik 1955) Mädler 1964, VII borehole 798.0 m, x 600; 2. *Calamospora keuperiana* Mädler 1964, VII borehole 732.0 m, x 750; 3. *Calamospora* sp., Gf-1 borehole 1846.5 m, x 750; 4–6. *Punctatosporites triassicus* Schulz 1965, 4, 5 Gf-1 borehole 1846.5 m x 750, 6. Vajta-3 borehole 1169.0 m, x 600.7–8. *Punctatisporites* sp., Vajta-3 borehole 1167.0 m, x 600; 9. *Verrucosiporites* sp., Gf-1 borehole 1846.5 m. x 750

Plate II

1. *Guttatisporites microechinatus* Visscher 1966, 3155 borehole, 1013.0 m; 2. *Guttatisporites guttatus* Visscher 1966, 3155 borehole 1013.0 m; 3. *Cyclotriletes microgranifer* Mädler 1964, 3155 borehole 1013.0 m; 4–5. *Cyclogranisporites arenosus* Mädler 1964, 4. Gf-1 borehole 1817.5 m, 5. Gf-1 borehole 1846.5m; 6–7. *Cyclogranisporites* sp., 6. Vajta-3 borehole 1155.0 m, 7. VIII borehole 1200.0 m. All x 750.

Plate III

1. *Guttatisporites guttatus* Visscher 1966, 3155 borehole 1013.0 m; 2. *Endosporites papillatus* Jansonius 1962, VIII borehole 1200.0 m; 3–5. *Densoisporites nejburgii* (Schulz 1964) Balme 1970, 3. Vajta-3 borehole 1167.0 m, 4–5. VII borehole 732.0 m; 6–9. *Densoisporites playfordi* (Balme) Dettman 1963, 6. Vajta-3 borehole 1167.0 m, 7, 9. Gf-1 borehole 1817.5 m, 8. Gf-1 borehole 1778.7 m; 10–12. *Lundbladispora* sp., 10. Vajta-3 borehole 1167.0 m, 11–12. Gf-1 borehole 1817.5 m. All x 750.

Plate IV

1. *Kraeuselisporites hoofddijkensis* Visscher 1966, Gf-1 borehole 1846.5 m; 2. *Concentricisporites plurianulatus* Antonescu 1969, VII borehole 798.0 m; 3. *Concentricisporites nevesi* Antonescu 1970, VIII borehole 1063.0 m; 4. *Saturnisporites praevius* Visscher 1966, VII borehole 647.0 m; 5–6. *Saturnisporites* sp., Gf-1 borehole 1601.0 m. All x 750.

Plate V

1–2. *Villanyipollis hungaricus* nov. gen. et sp., Máriagyüd-I borehole 1174.2 m, 1. holotype, x 500, 2. detail, showing intrareticulat and punctat structure and the protosaccus, x 100

Plate VI

1–4. *Villanyippollis hungaricus* nov. gen. et sp., Máriagyüd-L borehole 1174.2 m, 1–4. paratypes, x 500

Plate VII

1. *Limitisporites moersensis* (Grebe) Klaus 1963, VIII borehole 1200.0 m; 2. *Jugasporites* sp., VIII borehole 1200.0 m; 3–7. *Triadispora crassa* Klaus 1964, 3–4, 6–7. VII borehole 798.0 m 5. VIII borehole 1063.0 m; 8. *Triadispora plicata* Klaus 1964, VIII borehole 1063.0 m. All x 750

Plate VIII

1. *Triadispora staplini* Klaus 1964, VII borehole 798.0 m; 2. *Triadispora epigona* Klaus 1964, VII borehole 798.0 m; 3–7. *Stellapollenites thiergartii* (Mädler 1964) Clement-Weesterhof et al. 1974, 3, 7. Gf-1 borehole 1817.5 m, 4–6 VIII borehole 1063.0 m. All x 750

Plate IX

1–2. *Angustisulcites klausii* Freudenthal 1964, 1. Gf-1 borehole 1817.5 m, 2. VIII borehole 1063.0 m; 3. *Angustisulcites gorpii* Visscher 1966, VIII borehole 1063.0 m; 4. *Lunatisporites* sp., Gf-1 borehole 1817.5 m; 5. *Lunatisporites multiplex* Visscher 1966, VII borehole 798.0 m. All x 750

Plate X

1. *Lunatisporites* sp., VIII borehole 1200.0 m; *Lunatisporites jonkeri* Visscher 1966, Gf-1 borehole 1817.5 m; *Lunatisporites multiplex* Visscher 1966, Gf-1 borehole 1817.5 m; *Striatoabietites aytugii* Visscher 1966, Gf-1 borehole 1817.5 m. All x 750

Plate XI

1. *Alisporites grauvogeli* Klaus 1964, Gf-1 borehole 1817.5 m; 2. *Alisporites microechinatus* Reinhardt 1966, Gf-1 borehole 1778.7 m; 3–4. *Succintisporites grandior* Leschik 1955, 3. Gf-1 borehole 1778.7 m, 4. VII borehole 798.0 m; 5. *Chordasporites magnus* Klaus 1964, Gf-1 borehole 1817.5 m; 6–7. *Cycadopites coxii* Visscher 1966, 6. VII borehole 704.0 m, 7. VII borehole 798.0 m. All x 750

Plate XII

1. *Paravesicaspora planderovae* Visscher 1966, Gf-1 borehole 1846.5 m, x 750; 2–3. *Falcisporites snopkovae* Visscher 1966, 2. Gf-1 borehole 1778.7 m, x 650, 3. Gf-1 borehole 1817.5 m, x 750; 4–5. *Colpectopollis ellipsoideus* Visscher 1966, VII borehole 798.0 m, x 750; 6. *Alisporites* cf. *ovatus* Jansonius 1962, VII borehole 798.0 m, x 750

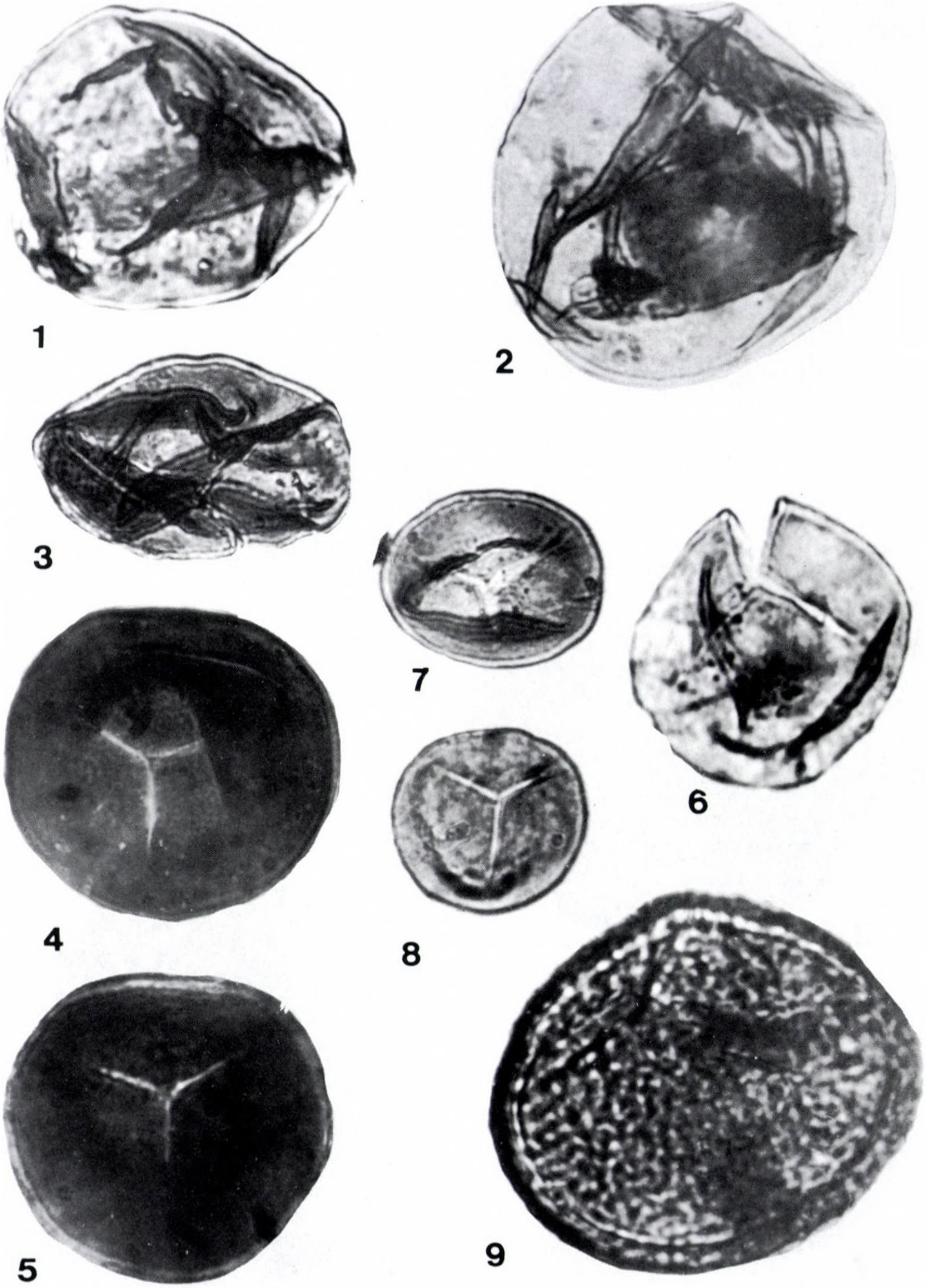
Plate XIII

1. *Alisporites grauvogeli* Klaus 1964, Gf-1 borehole 1778.7 m; 2–3. *Alisporites cymbatus* Venkatachala Beju et Kar 1968, 2. VIII borehole 1200.0 m, 3. Gf-1 borehole 1835,6 m; 4–6. *Voltziaceaesporites heteromorphus* Klaus 1964, 4. Gf-1 borehole 1778.7 m, 5. Gf-1 borehole 1817.5 m, 6. VIII borehole 1063.0 m. All x 750

Plate XIV

1. Anisian sporomorph assemblage containing *Villanyipollis hungaricus* nov. gen. et sp. (1–3), with other bisaccat grains [*Alisporites cymbatus* (4), *Voltziaceaesporites heteromorphus* (5), *Stellapollenites thiergartii* (6–7)], x 80

Plate I



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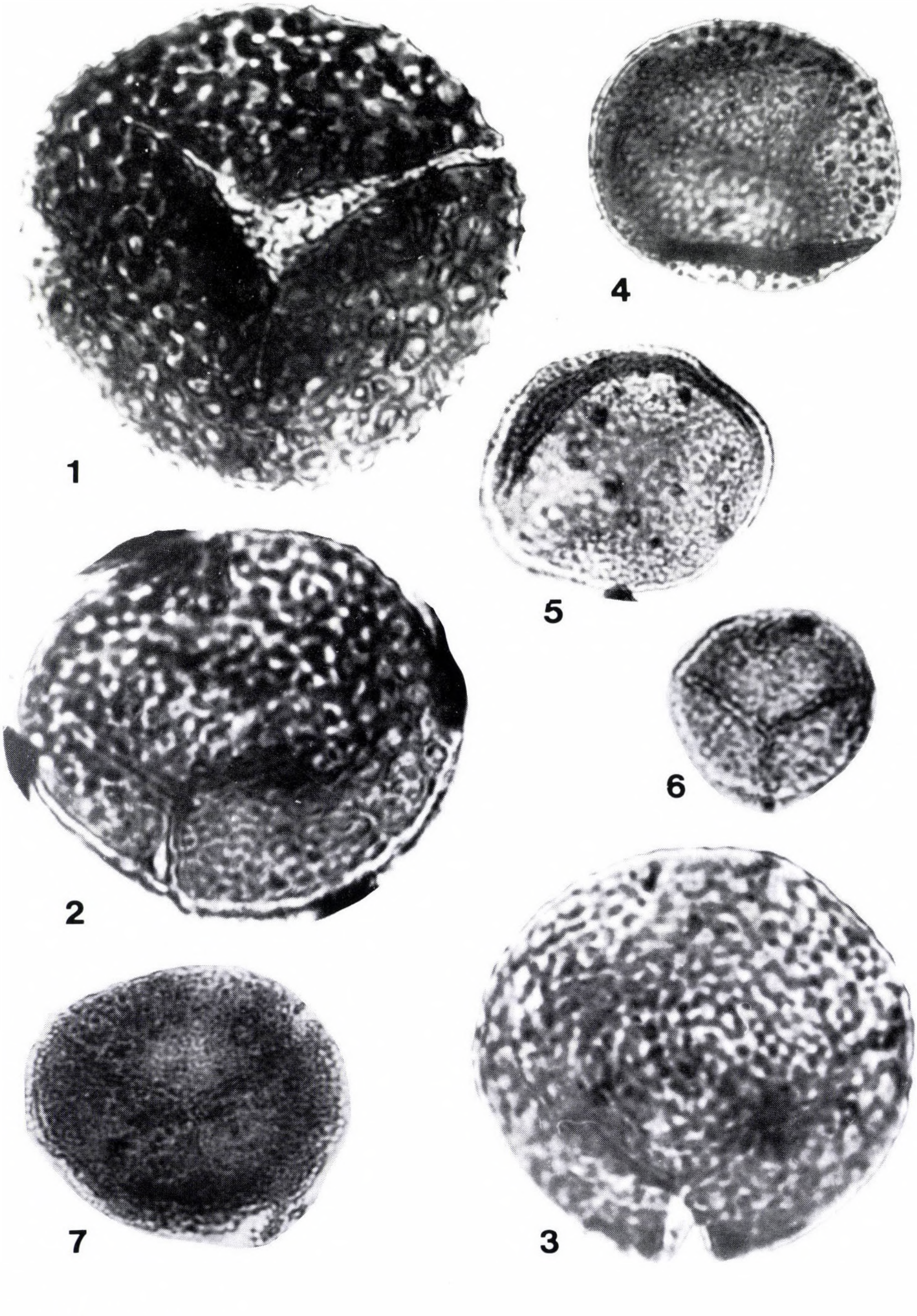

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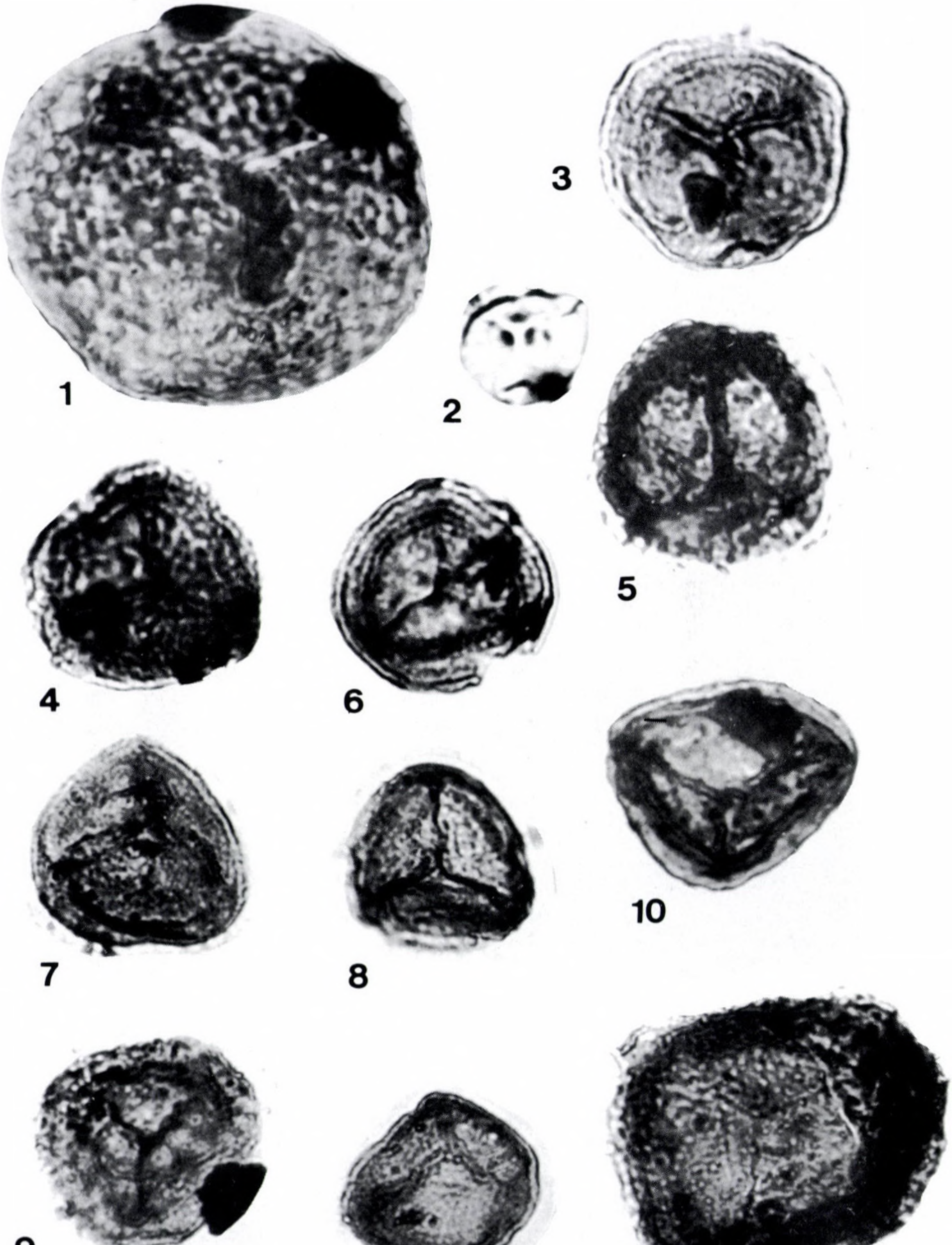

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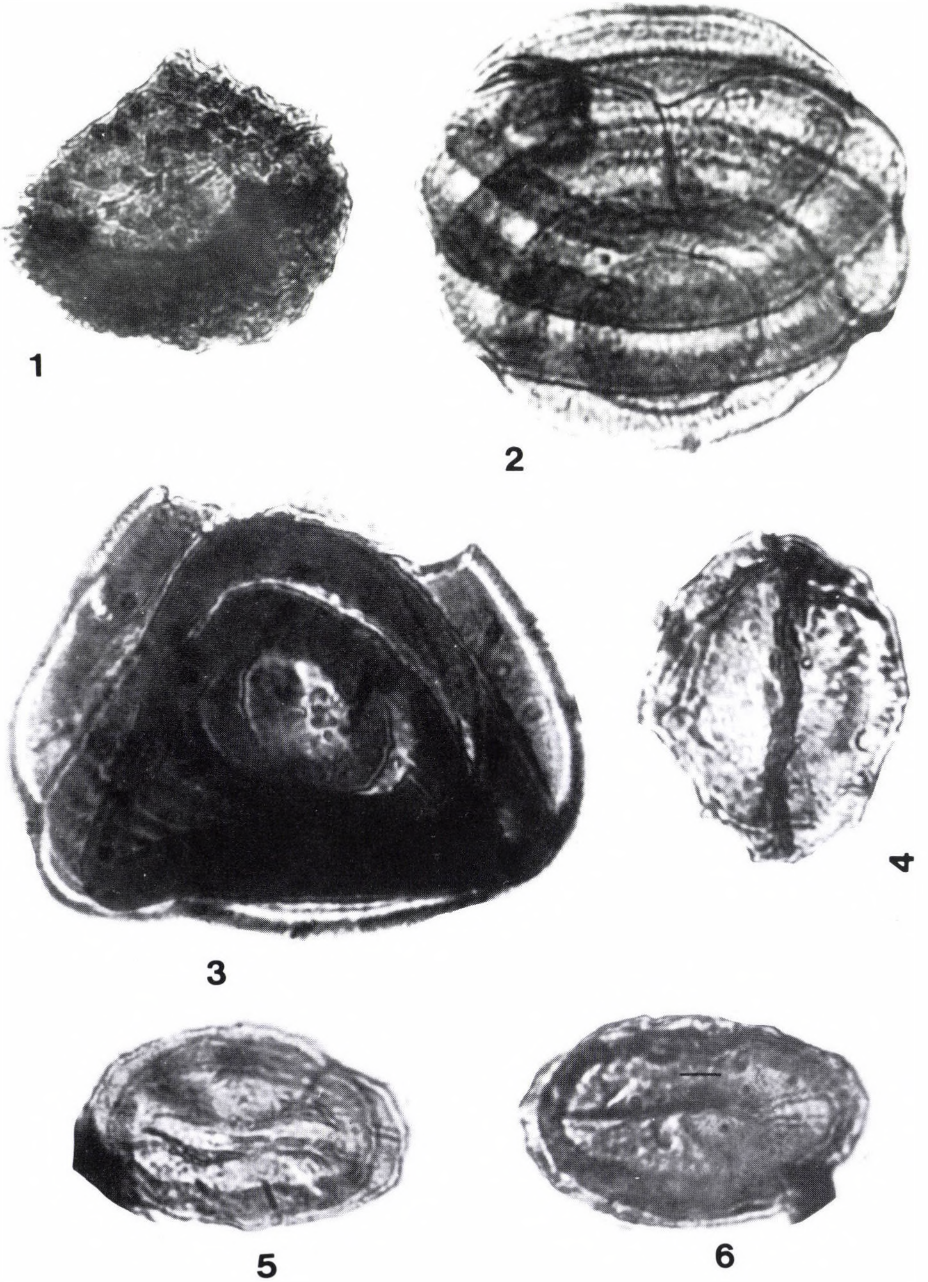

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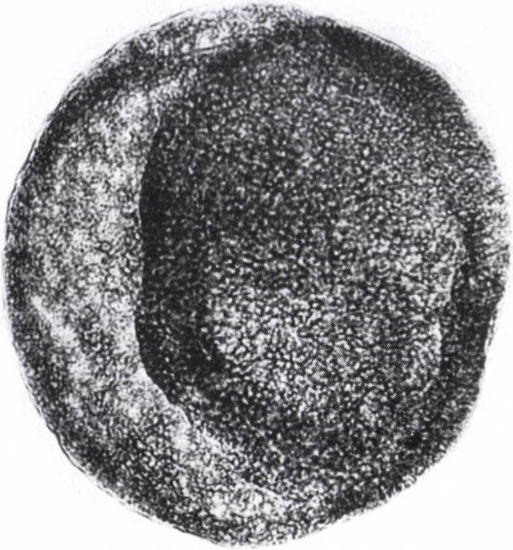

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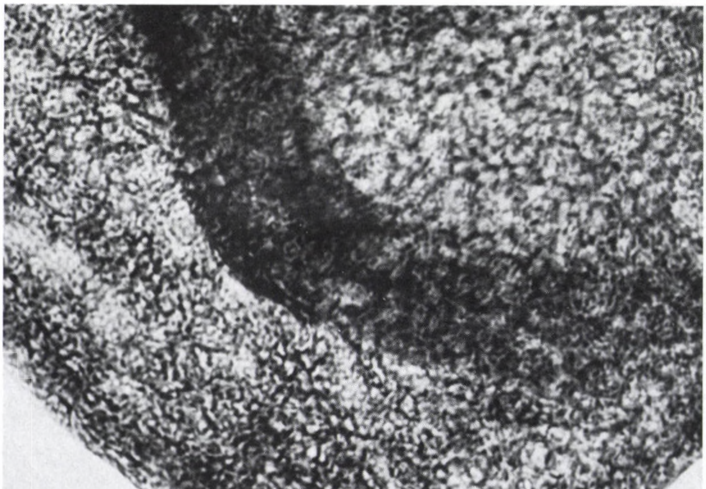

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Plate VI

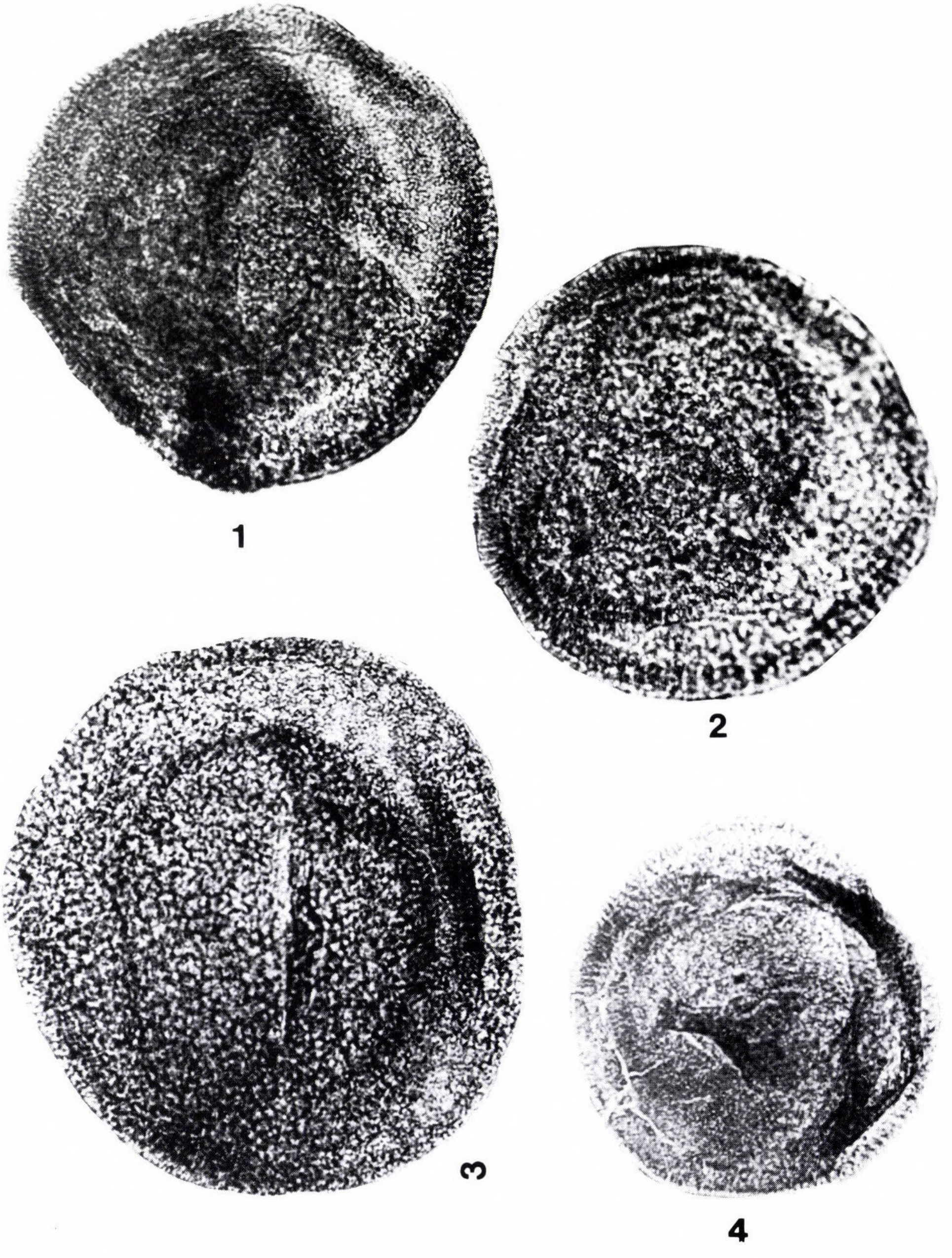
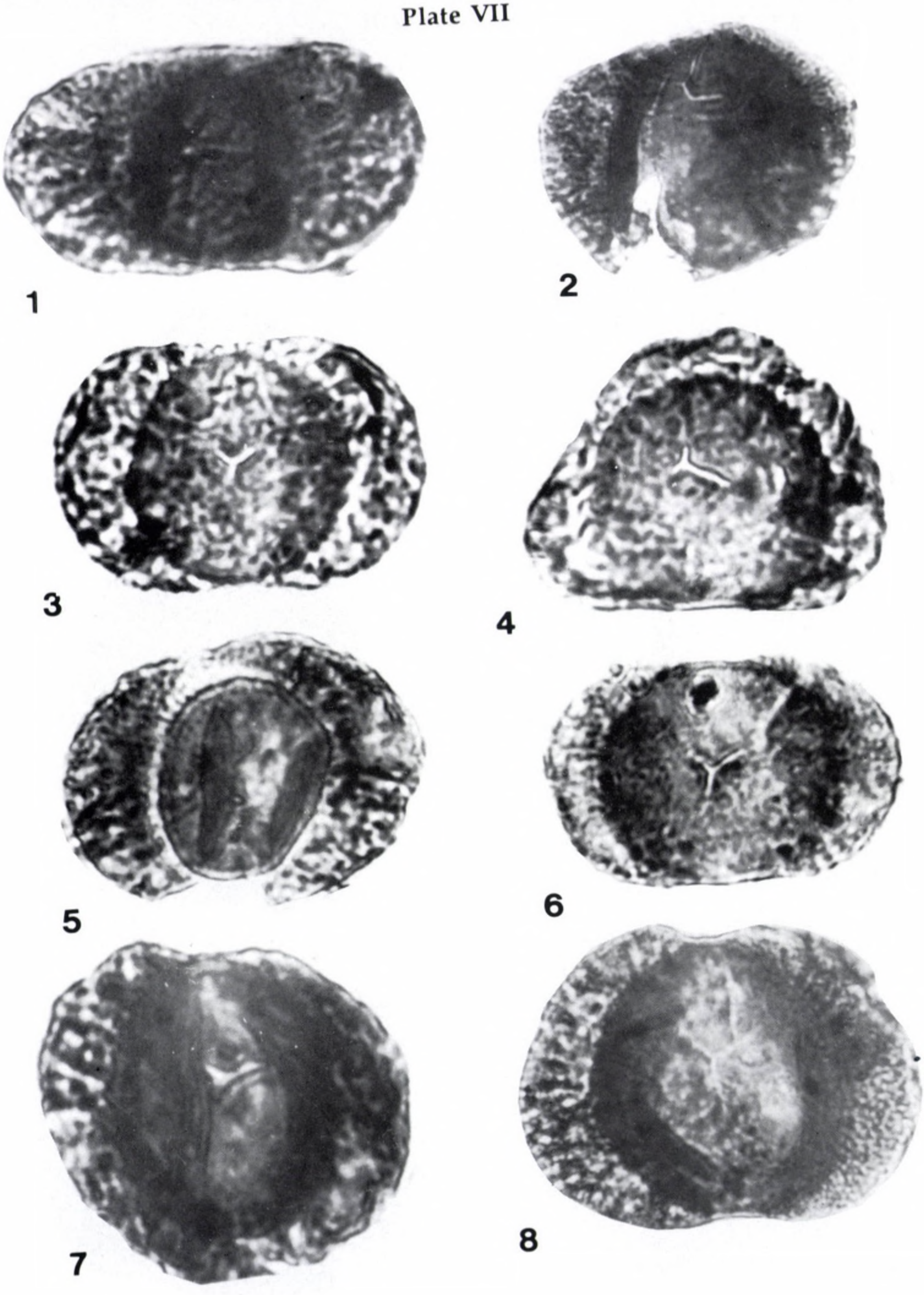


Plate VII



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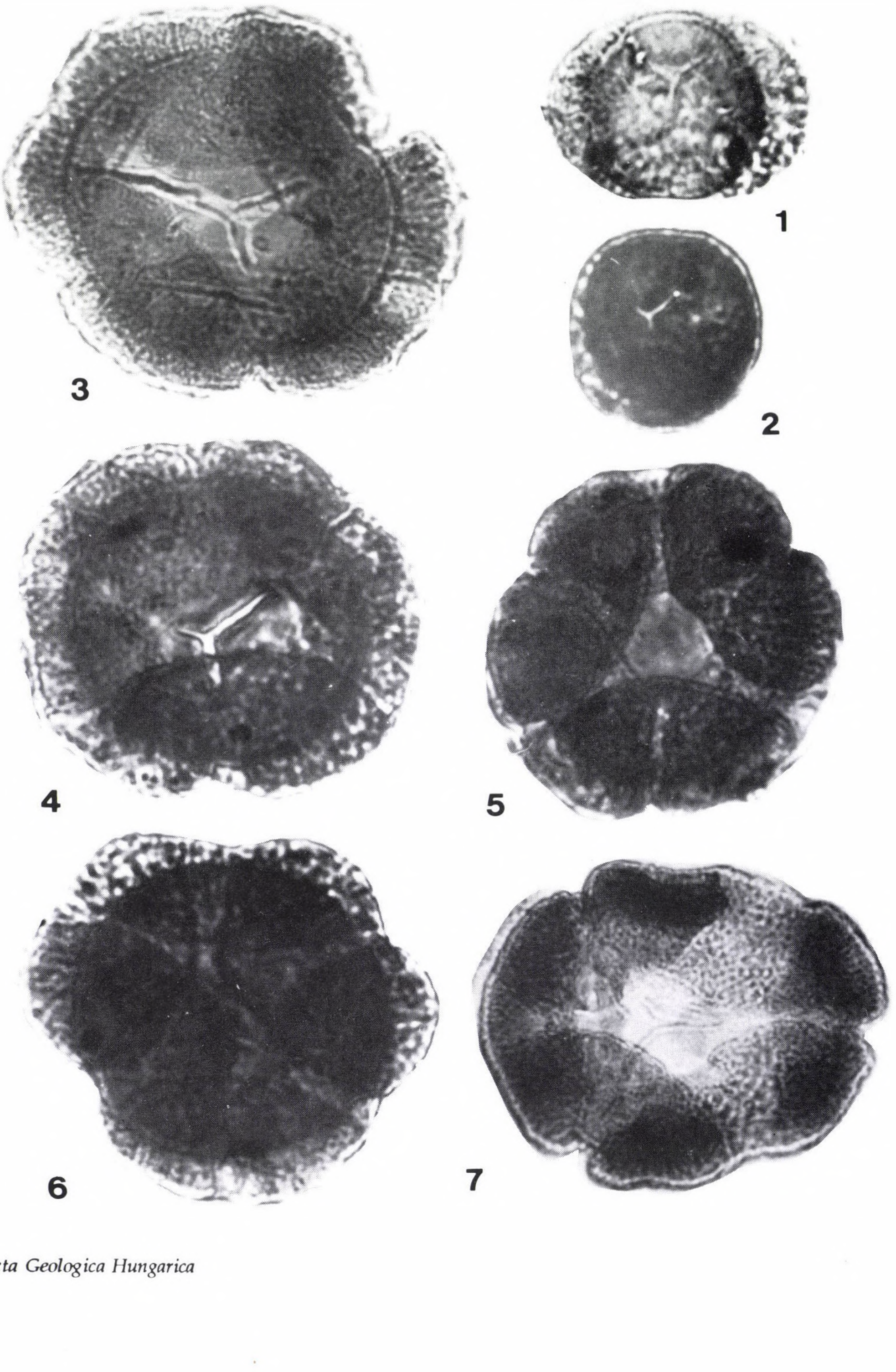

Plate IX

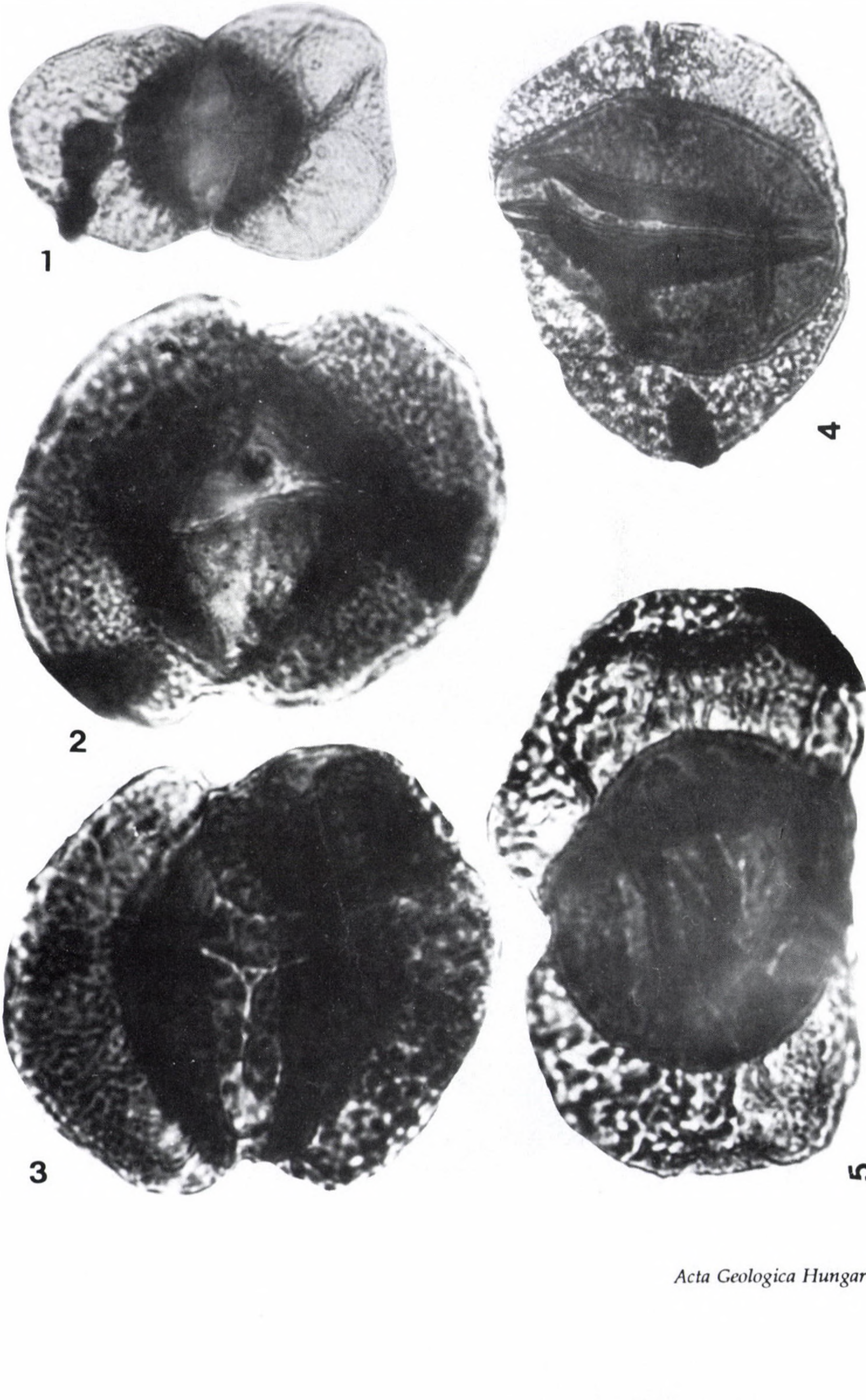

Plate X

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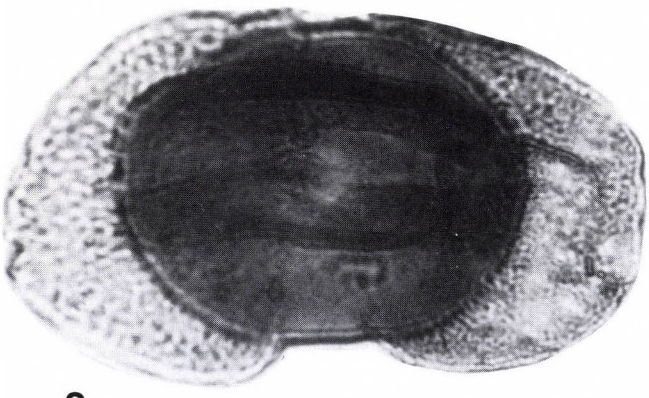

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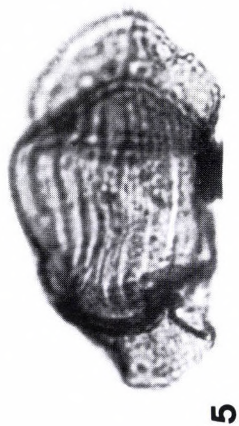

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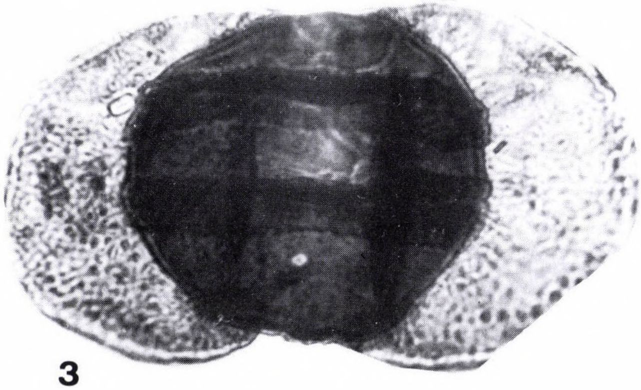

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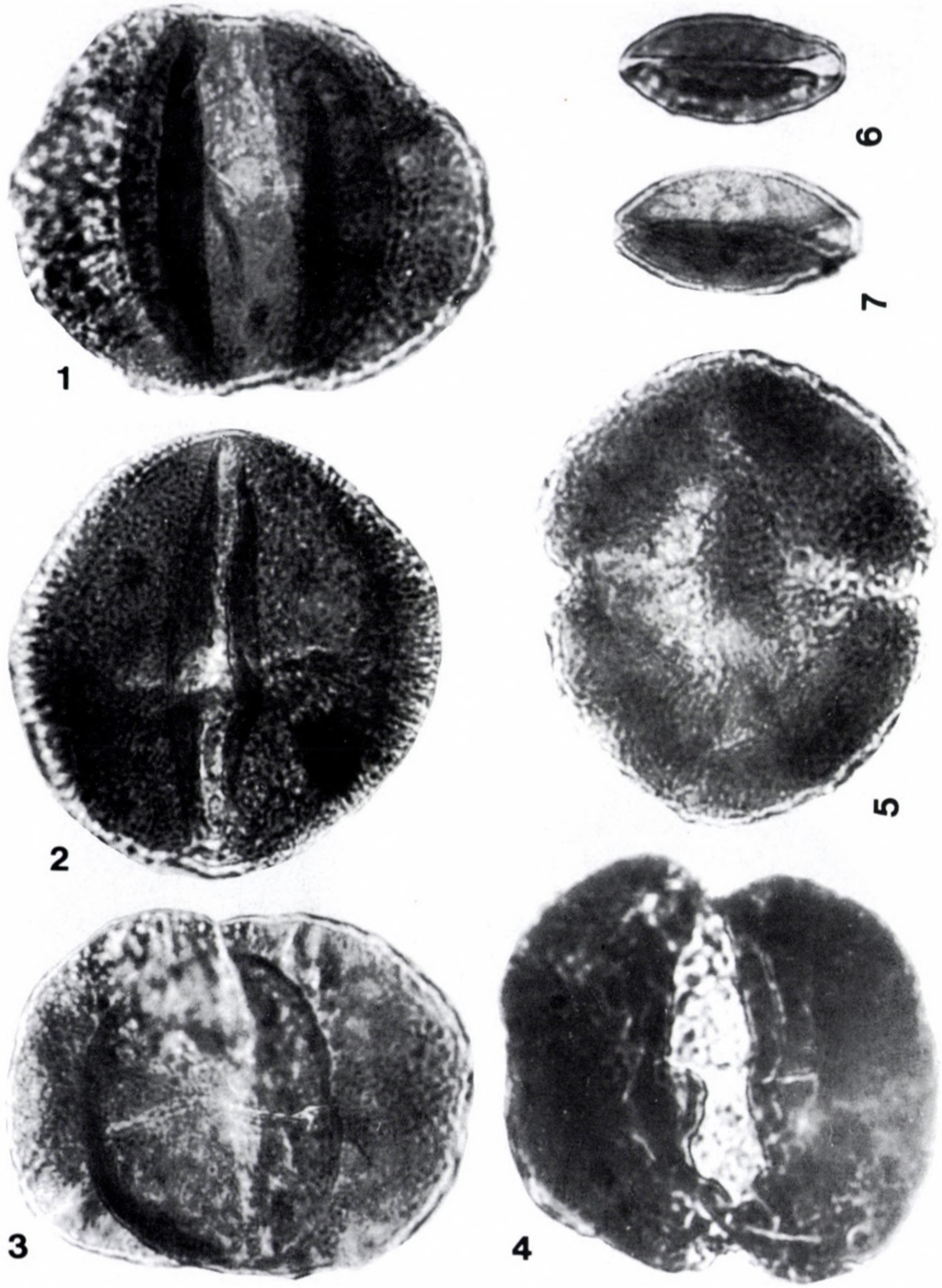
Plate XI



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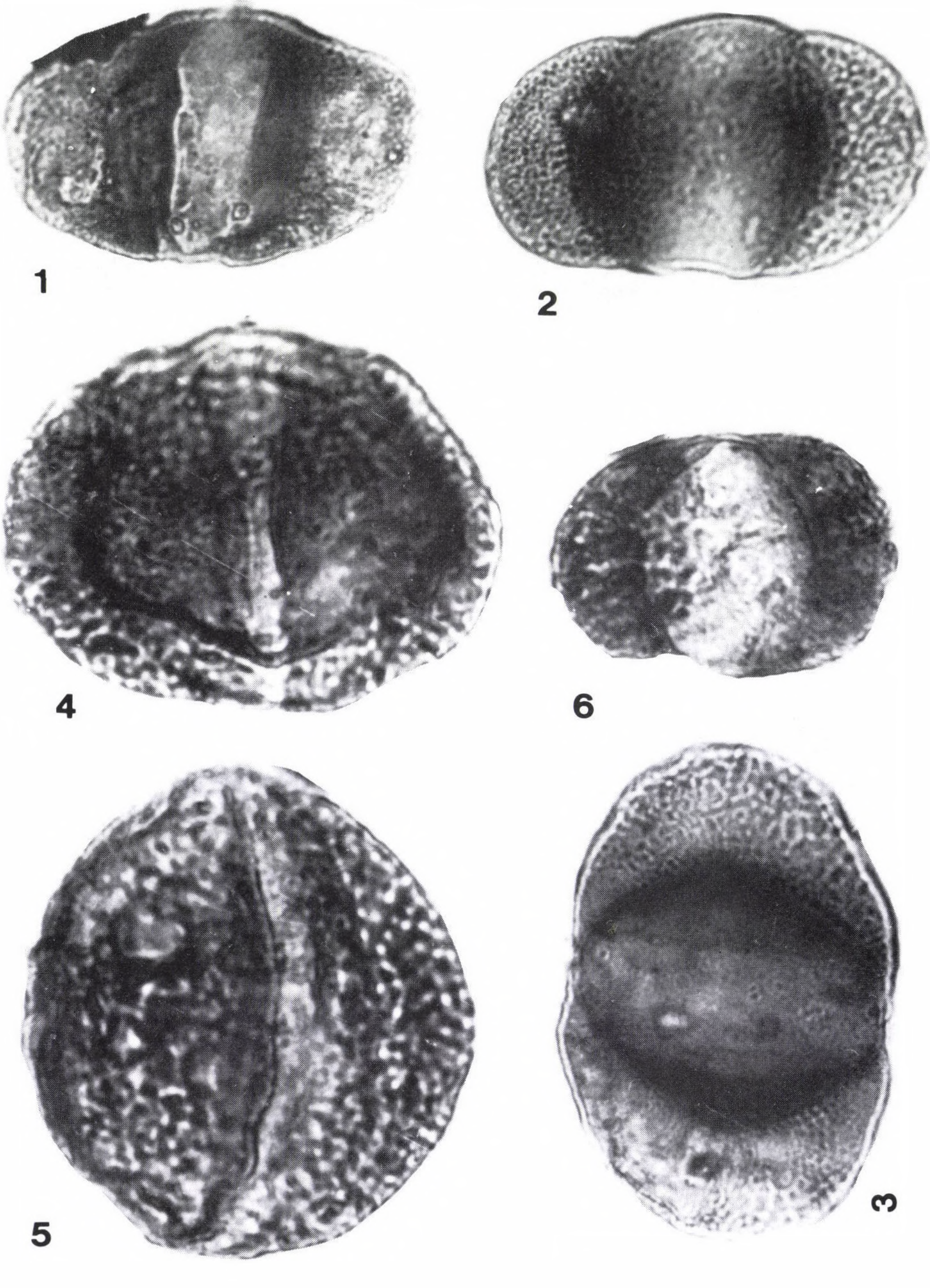

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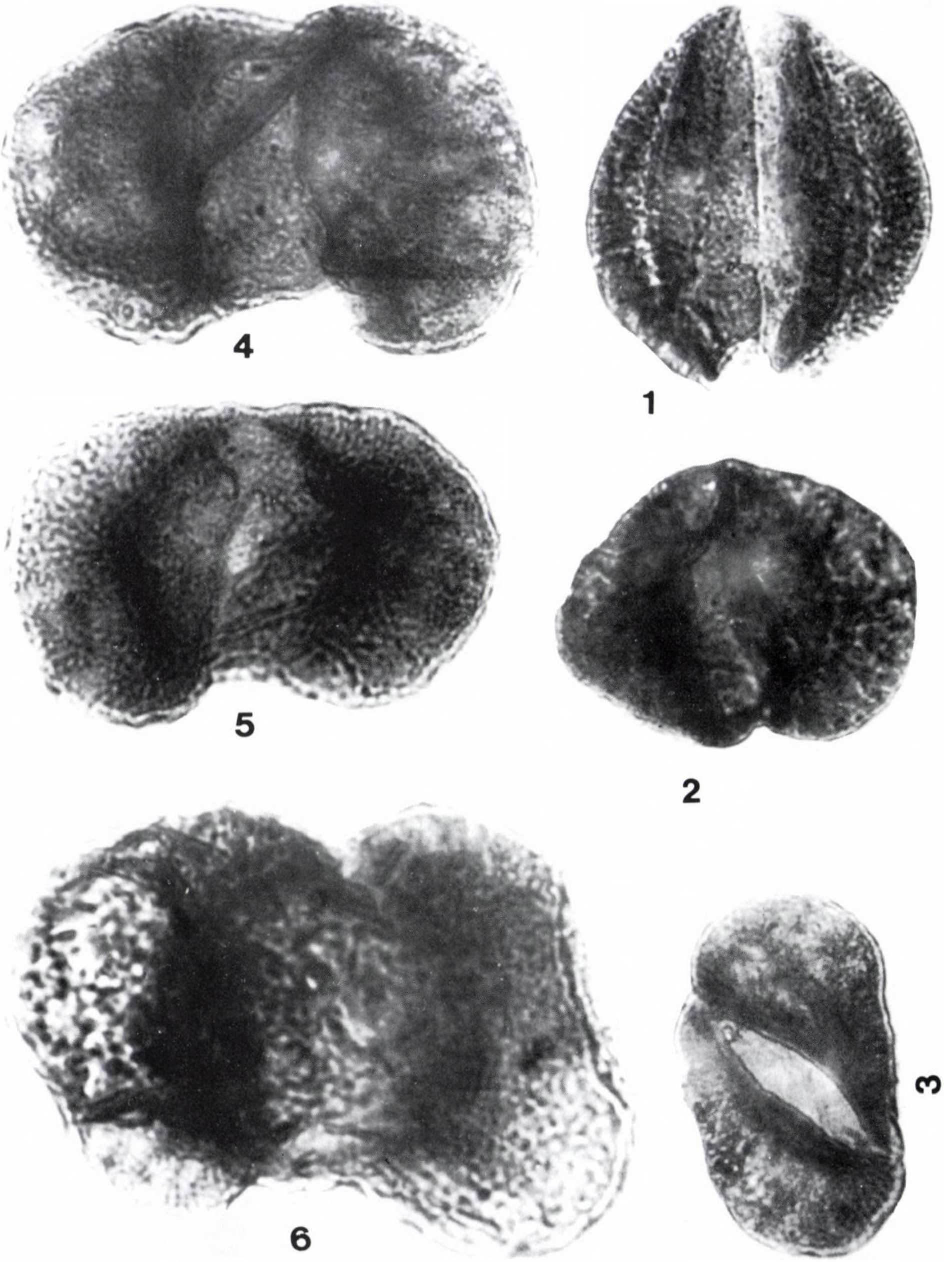
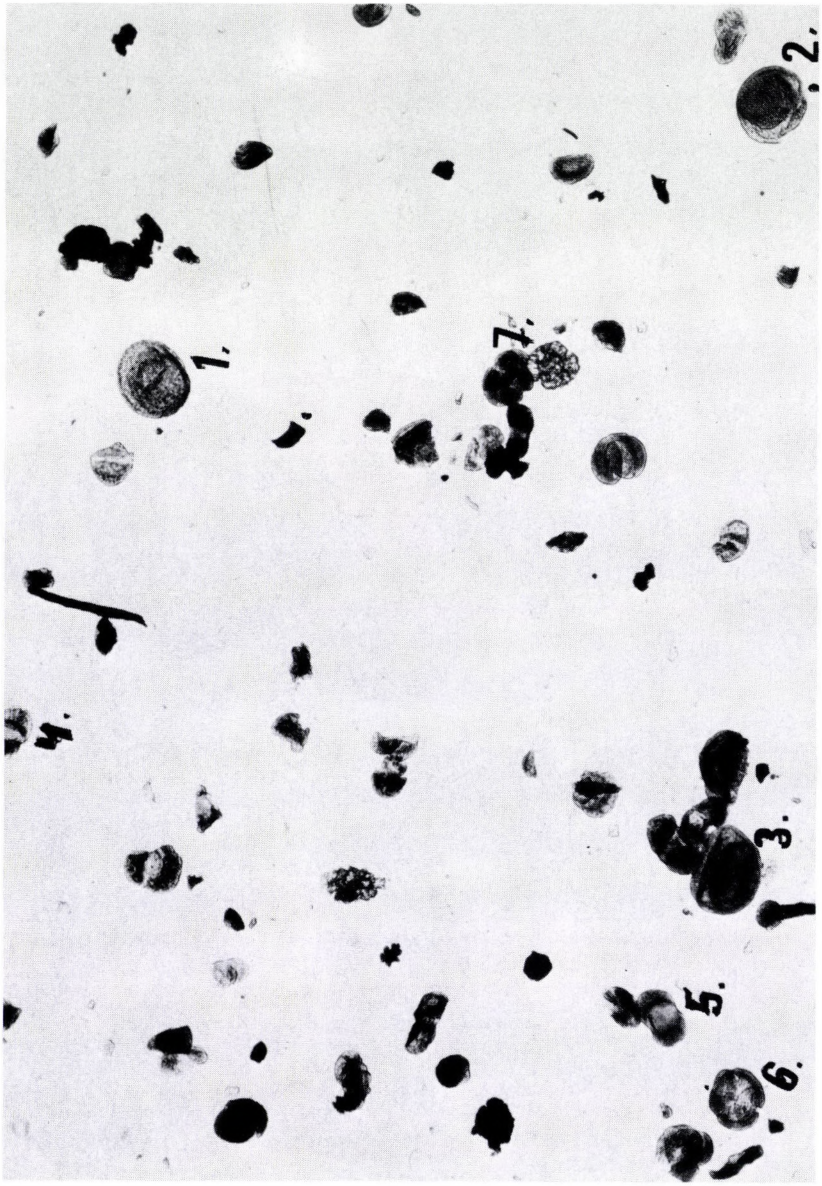


Plate XIV



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Geochemical exploration for Blind Bauxite Ore bodies in Jajce, Central Bosnia

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Unfavorable production commodities in the Jajce underground bauxite mine necessitate introduction of geochemical exploration methods in search for blind ore bodies. Investigation was focused on primary and secondary dispersion halos and their size around the bauxite ore bodies. Concentration of trace elements Cd, Cu, Zn, Ni, Mn, Sr and Li in the foot wall and hanging wall host rocks, as well as in the paleokarst weathering products, taken on their contact, have been determined by AAS and Hg by CVAAS after digestion of the rock samples by aqua regia. The applied statistical methods, elementary statistics, clustering and discriminant analysis determined two distinctive trace element populations in the footwall and hanging wall rocks. Multivariate linear regression created an efficient mathematical model for prediction of the distance (D) between the blind ore body and missed (ore barren) drill-hole.

Key words: Bauxite, trace elements, geochemical exploration, primary dispersion halo, host rocks, paleokarst weathering products

Introduction

Bauxite exploration in the Jajce producing area has been performed by geological and geophysical methods nearly a half of the century. The outcropping bauxite ore bodies have been already mined out and the exploitation progresses ever deeper under the hanging wall rocks. Costs of deep-drilling exploration have risen due to decreasing effectiveness. Unfavorable production commodities necessitate introduction of new geophysical and geochemical methods, which may use geological information obtained even on the missed drill holes, and could lead to more effective prediction of the blind bauxite ore bodies or assessment of the area potential.

Having in mind, that the bauxite as a weathering product, formed during emergence phases, taking place intermittently within the long-lasting Mesozoic

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and Paleogene marine sedimentation, represents "a foreign body" in the calcareous environment, attention has been paid to its geochemical influence on the surrounding rocks.

Beside the secondary trace element dispersion around the bauxite body which might have resulted by diagenetic and epigenetic processes, caused by migration of connate water, or circulation of meteoric water, one may expect a primary synsedimentary trace element distribution in the paleokarst weathering crusts and the hanging wall rocks, due to weathering and sedimentary processes on the diversified karst topography, which collected bauxite material in the paleokarst depressions.

Geochemical analyses of trace elements in the bauxites and the host or possible parent rocks have been performed mainly for studies of bauxite genesis in particular by Dudich (1972); Šinkovec (1976); Schroll (1979); Dudich and Mindszenty (1985); Maksimovic (1976, 1988); Šinkovec et al. (1989); Maksimovic et al. (1991), and others.

Trace element distribution in the bauxite ore bodies, within an ore-bearing field has been studied by Horvath and Peter (1985). Data on the primary or secondary dispersion halos of trace elements around the bauxite ore body, however, are scanty (Miko 1987; Palinkaš et al. 1989) or absent as to our knowledge.

In order to get geochemical data for the statistical characterization of the hanging and foot wall rocks and the contact material between them (paleokarst weathering products), the rock samples have been collected from drilling cores and some bed rock exposures, and then analyzed on Hg, Cd, Cu, Zn, Ni, Mn, Sr and Li. Data analysis, using elementary statistics, multivariate linear regression, cluster and discriminant analysis provided a better insight into the regularity of elemental distribution, what may be applied in search for the blind bauxite ore-bodies.

Geology

The investigation took place in the Jajce bauxite-bearing area at the localities of Poljane, Seoce and Crvene Stijene, a few kilometers northeastward from the town of Jajce (Fig. 1).

The local geological structure comprises the Lower and Upper Cretaceous carbonate sediments. The bauxite deposits are situated in the paleokarst depressions formed over different lithotypes of the Aptian and Albian limestones (micrites, biomicrites, intrasparites, etc.). They originated in shallow, low energy water of the Dinaric carbonate platform shelf (Dragičević 1981; Polšak 1981; Polšak et al. 1982; Tomić 1983; Dragičević, 1987).

The Late Albian, Turonian and the Early Senonian regression gave rise to a long-term emergence phase. Dry-lands, extending along the carbonate platform underwent severe karstification, bauxitization of weathering crusts, and washing out of bauxite material into the newly formed karst depressions.

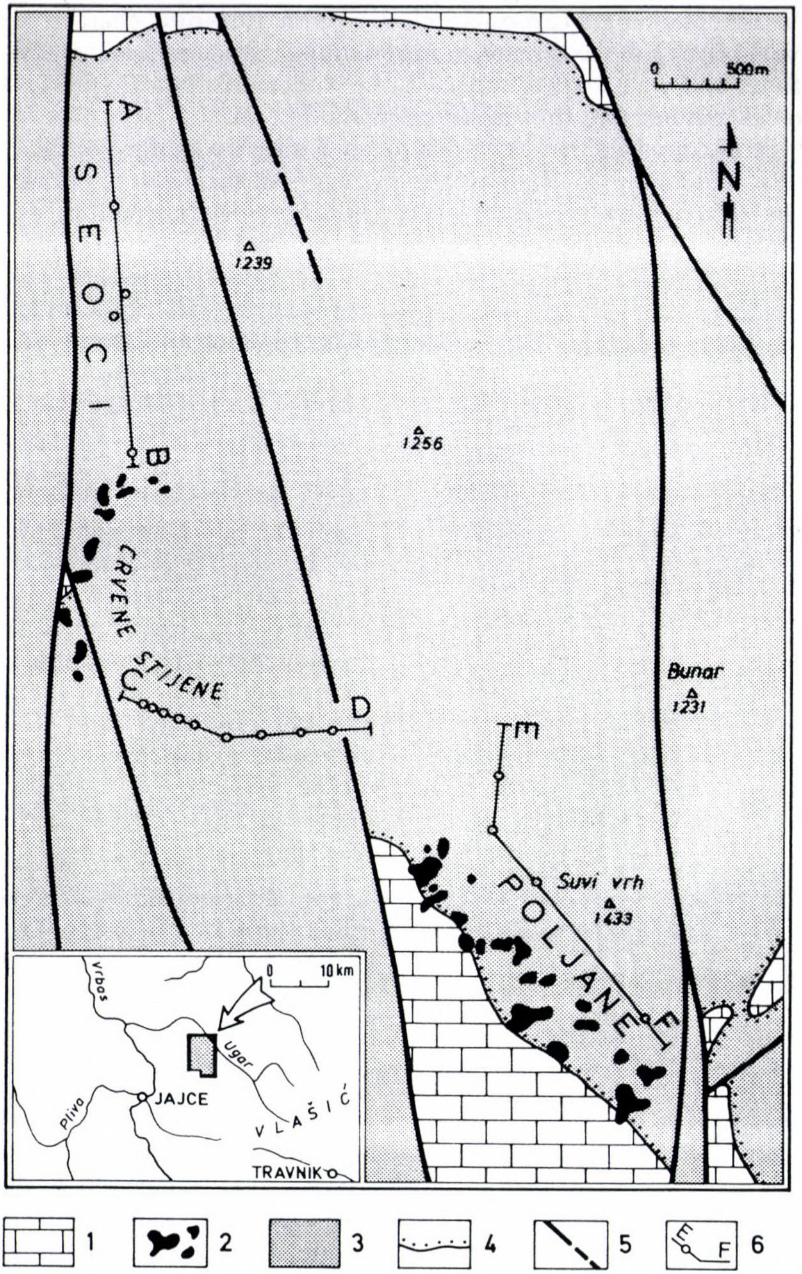


Fig. 1
Geological situations at the bauxite-bearing terrain of Jajce. 1. limestones, Aptian–Albian, the footwall rocks of the bauxite; 2. the bauxite deposits; 3. calcareous clastics, Santonian–Maastrichtian ("flysch") the hanging wall rocks of the bauxite; 4. the main tectonic-erosion unconformity; 5. the main faults; 6. geological cross-sections with drill-holes

Intensive tectonics at the beginning of Santonian brought about transgression with sedimentation of carbonate clastics with turbidite characteristics. The Senonian flyshoid sediments, surmounting transgressively the Albian limestones and the bauxites, consist of the basal rudist biocalcirudites followed by bio-calcarenites and thin layers of marls (Fig. 2).

Part of the deposits had been destroyed during deepening of the carbonate margin as evidenced by the bauxite fragments in the Santonian breccias. The deposits covered by the rudist breccias and those formed in a back-reef environment had a better chance not to be affected by the erosion.

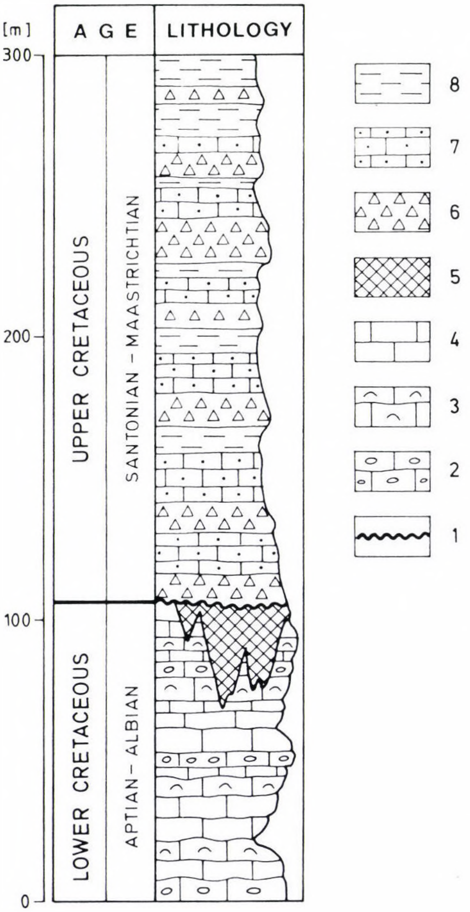


Fig. 2
Stratigraphic column of the bauxite-bearing sedimentary sequence. 1. unconformity; 2. intrasparite, 3. biomicrite; 4. micrite; 5. bauxite; 6. biocalcirudite; 7. biocalcarenite; 8. marl

The bauxite deposits vary in size between a few thousands to over 300 000 tons and occur in shape as regular, channel-like lenses and pot-like bodies (Dragičević 1981).

Average bauxite of the region has the following mineralogical composition in percents: boehmite, 66.5, kaolinite, 3.9, hematite, 29.9, anatase, 0.7, calcite, 1.5 and an undetermined rest 1.4. Microscopic examination revealed cryptocrystalline kaolinite-boehmite matrix, engulfing rounded, deformed and fragmented ooids. Detrital, rounded and fragmented limestone grains are present as well (Miko 1987):

Sampling

Sampling procedure was performed on 20 exploration drill holes, at three ore fields (Poljane, Seoce and Crvene stijene) and three rock samples were taken from the outcropping rocks (Surjan, locality). Each exploration drill has been sampled on the core of the foot wall and hanging wall rocks and the paleokarst material from their contact. The hanging wall rocks have been sampled 1 m from the contact, while the foot wall rocks were taken at a random vertical distance from it (Fig. 3). The ore bodies were discovered in four drill holes.

Analytical procedure

The rock samples were crashed and ground to the powder. In order to get optimum applicability, within the constraints of cost and time "hot extractable" or "near-total" extraction method has been chosen (Rose et al. 1979). Five grams of sample powder was digested by a 15 ml mixture of acids, HCl–HNO_3, 3:1, at 70° C (to keep mercury in solution) for 3 hours on a bath. The resulting solution was diluted by deionised water in a 50 ml volumetric flask. The procedure attains a very good reproducibility (±5%) and recovers great deal of trace elements from carbonate and non-carbonate fraction of the limestones.

Determination of Cd, Cu, Zn, Ni, Mn, Sr and Li was accomplished by standard AAS techniques, while Hg by cold vapor AAS. Mercury vapor was liberated by 10% $SnCl_2$ and 20% hydroxylamin-hydrochloride solution in a glass flask, driven through a quartz glass tube (Mesaric 1974) and measured by Pye Unicam spectro-photometer. Precision of the measurements, expressed as standard deviation was far bellow the standard deviations of the trace element concentrations in the rocks (two orders of magnitude, Table 1).

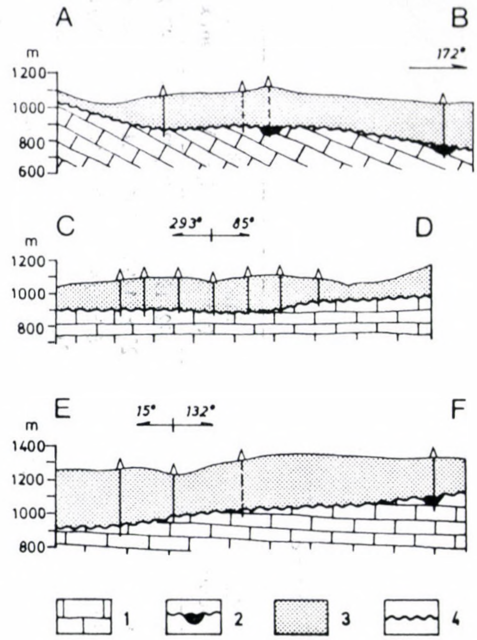


Fig. 3
Simplified geological cross-sections through the ore-bearing sedimentary sequence with position of the drill-holes. 1. limestones, Aptian–Albian, the footwall of the bauxite; 2. the bauxite deposits; 3. calcareous clastics, Santonian–Maastrichtian ("flysch"); 4. tectonic erosion unconformity

Statistical analysis

The aim of the investigation was to create a simple and inexpensive statistical model able to recognize the barren from the ore-bearing host rocks, and to make a regression equation that would define relation between the distance from the ore body and concentration of the relevant trace elements in the barren host rock. The problem has been approached by elementary statistics, and multivariate linear regression, cluster and discriminant analysis.

All the rock samples (50)-group A, have been subdivided into 6 smaller groups: the hanging wall rocks (16)-group H, the foot wall rocks (34)-group F,

Table 1
Mean concentration/standard deviation, trace elements in the rock groups

Rock group	Number of samples	Hg (ppb)	Cd (ppb)	Cu (ppm)	Zn (ppm)	Ni (ppm)	Mn (ppm)	Sr (ppm)	Li (ppm x 100)
A	50	82/53	130/99	1.4/1.3	5.3/4.1	12/15	12/19	70/82	97/7
H	16	102/57	131/84	1.4/0.9	8.7/7.6	7/8	25 ± 24/96	156 ± 24/96	98/6
F	34	73/48	141/104	1.5/1.4	5.4/3.1	17/20	6 ± 1/4	29 ± 4/20	94/7
BH	12	96/59	109/64	1.2/0.8	5.3/4.5	12/15	21± 5/18	159 ± 31/106	100/6
BF	13	68/38	149/139	1.7/1.3	5.8/3.2	14/14	5 ± 1/2	36 ± 32/64	93/6
BxH	4	119/54	198/113	1.8/1.0	6.5/6.8	11/15	36 ± 26/53	156 ± 32/64	93/3
BxF	6	101/82	147/103	2.7/2.3	4.1/2.8	23/35	7 ± 3/7	17 ± 4/11	95/7
C	5	66/45	75/43	1.5/2.2	5.7/1.6	15/7	20/14	184/125	100/8
Mean ratio: H/F		1.40	0.92	0.93	1.06	1.42	4.17 ± 1.4	5.34 ± 9.1	1.04
BH/BF		1.41	0.73	0.71	0.89	1.27	4.20 ± 1.3	6.12 ± 1.4	1.08
BxH/BxF		1.52	0.71	1.04	1.19	1.53	5.14 ± 4.2	9.12 ± 2.9	1.00
BxH/BH		1.29	1.81	1.50	0.71	1.64	1.71	0.98	0.93
BxF/BF		1.49	0.99	1.59	0.88	1.36	1.40	0.65	1.02

Symbols: A – all rock samples; H – hanging wall rocks; F – footwall rocks; BH – barren hanging wall rocks; BF – barren footwall rocks; BxH – bauxite-bearing hanging wall rocks; BxF – bauxite-bearing footwall rocks; C – contact: hanging wall–footwall rocks; ± – stand. error the barren hanging wall rocks (12)-group BH, the barren foot wall rocks (13)-group BF, the bauxite-bearing hanging wall rocks (4)-group BxH, the bauxite-bearing foot wall rocks (6)-group BxF, and a special group C (5), made of the contact material (paleokarst weathering crust), Table 1.

Reason for the subdivision arises from the lithological diversities. Difference in the rock lithology makes probable geochemical difference among the groups which by itself might be used as a recognition indicator.

A good ground for that has been justified by inspection of the frequency diagrams and probability plots, and especially by applying multivariate linear regression, which was much more efficient by the treatment of the separate groups.

The Table 1 presents average trace element concentration and standard deviations of the aforementioned sample groups. Statistics of each elemental distribution shows certain individuality. Judging by the probability plots one may draw following conclusion:

Cd, Cu and Zn: log-normal distribution (Fig. 4a, b, c);

Li: normal distribution, slightly truncated at the highest values (Fig. 4d), this is a complex distribution of the sample group H and F, as testified, hypothesis H_0: $\mu_H=\mu_F$, rejected (t=2.26705, sign. level α =0.02793).

Sr: log-normal distribution seemingly (Fig. 4e), justified existence of two-population distribution (Fig. 4f) of the group H (Fig. 4g) and the group F (Fig. 4h),

H_0: $\mu_H=\mu_F$, t=7.588, rejected at α =0.05;

H : $\delta^2_H = \delta^2_F$, F=22.1978, DF 33,15, not rejected at α =0.05;

High ratios H/F = 5.34 ± 1.10, BH/BF=6.12 ± 1.41 and BxH/BxF = 9.12 ± 2.86 (values are standard errors of the ratios, Findly and Kichener, 1953) can be used as a simple recognition indicator for the hanging and foot wall rocks (Table 1).

Mn: does not fit well into normal or log-normal distribution as testified by χ^2 = test. The t-test and F-test proved probable existence of the group H and F (Fig. 4i), the same ratios as for Sr can also be used as a recognition indicator (Table 1);

Ni: a combined two-log-normal distribution (Fig. 5a), seemingly (Sinclair 1976), divided at the inflection point (0.8 log ppm), but the same tendency observed in the group H and F (Fig. 5b, c);

Hg: Similar behavior like Ni, two-population or even more complex distribution (Fig. 5d, e, f, i). Subdivision persisted even into the group BF and the group BH (Fig. 5g, h). As a peculiarity, the t-test and F-test did not reject hypothesis of the unique source population for the group F and H. The frequency diagram (Fig. 5i), suggests existence of the distinctive Hg populations not only for the group F and H but also for different ore bearing fields.

An overall review of the distribution characteristics for the group F and H, presented by the "box and whisker" display (Statgraphics 1988) is given on Fig. 6.

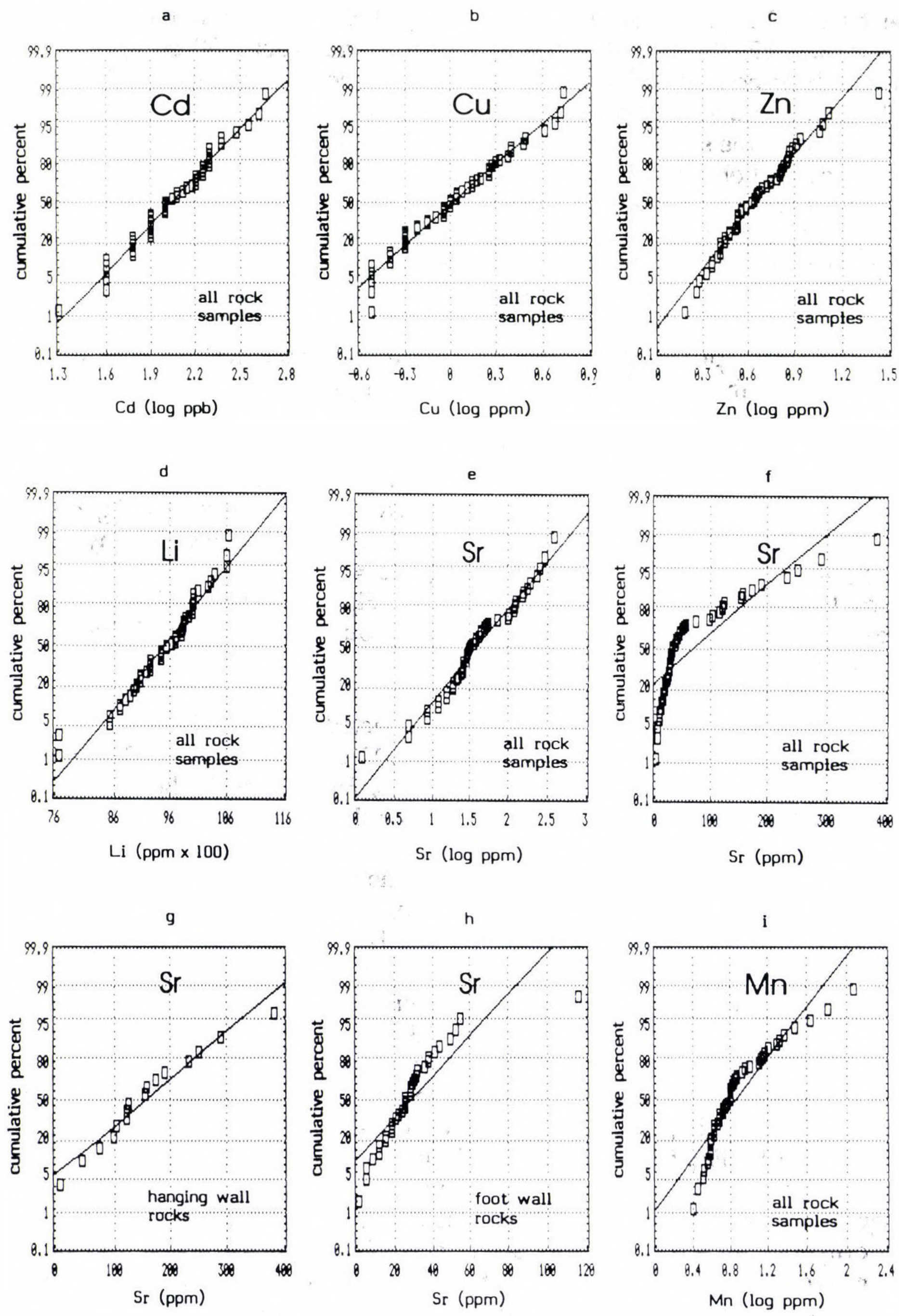


Fig. 4
Probability plots and frequency distribution diagrams

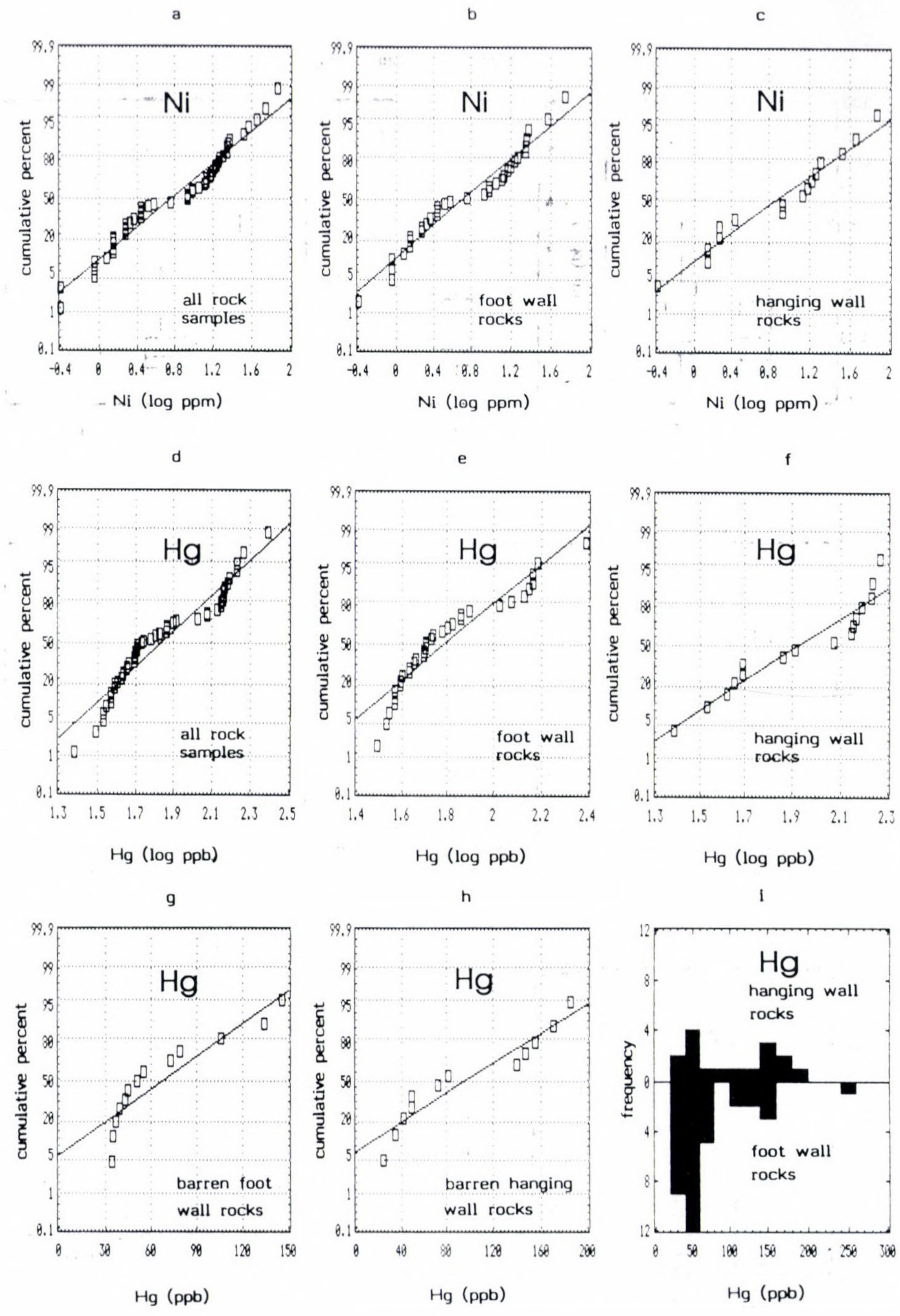


Fig. 5
Probability plots

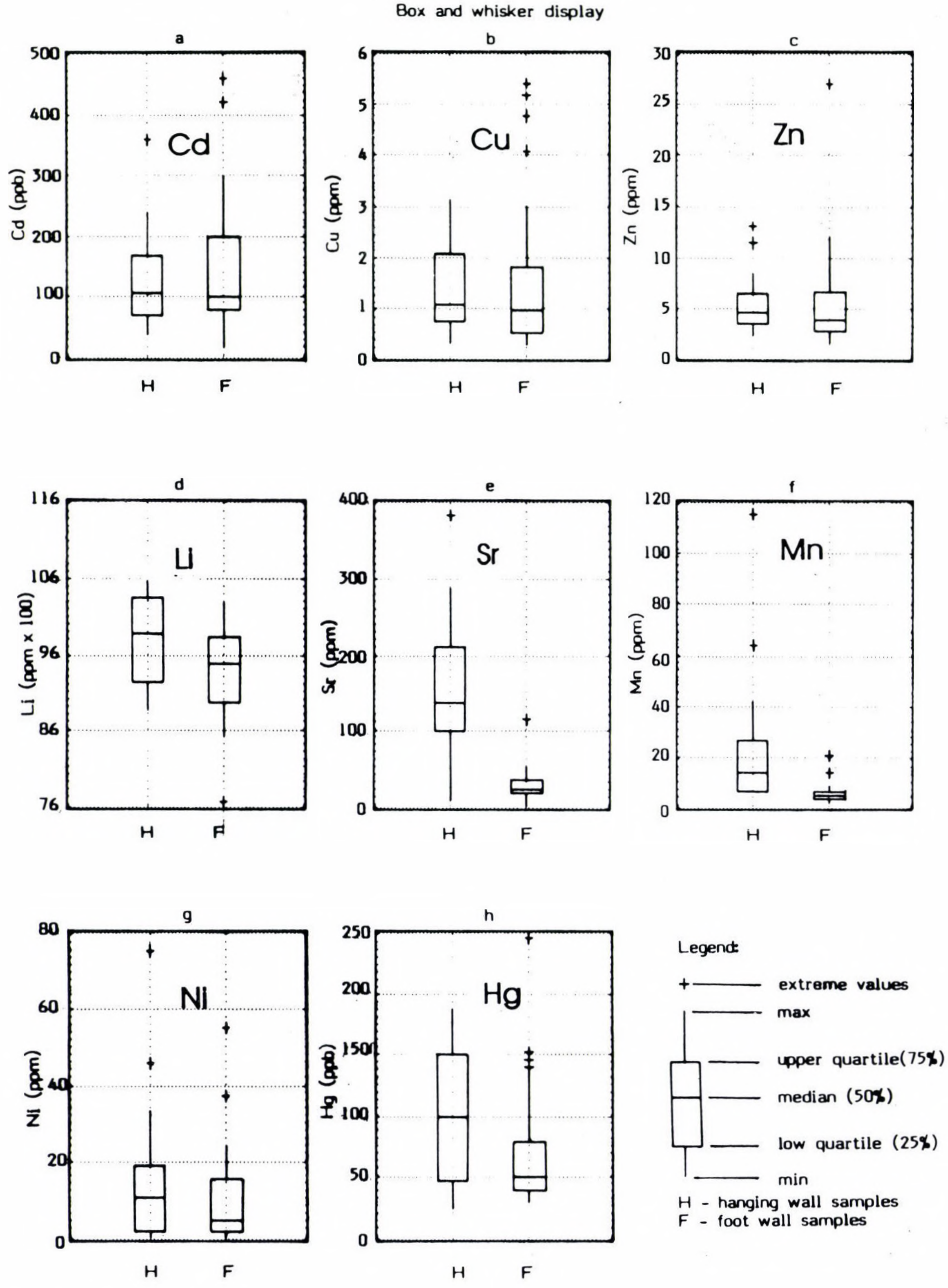


Fig. 6
Box and whisker displays of the distributions

Statistics with applied testing definitely proved at least two populations, represented by the groups F and H, with their own characteristics. Since the subdivision persisted even into the group BF and BH (Ni, Hg) there must be a certain difference in the target populations of different ore-fields. Each element shows certain individuality, seeking for more detailed geochemical explanation based upon additional research.

Clustering

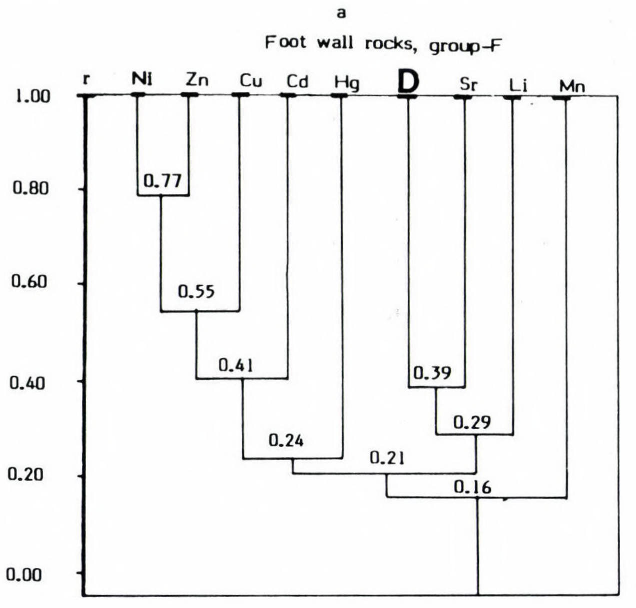
Construction of the correlation matrices (Table 2) preceded clustering. There is a pronounced negative correlation between the distance (D) from the missed drill hole and the nearest ore body and all trace elements in the group H. Most of the r–s are very close to the level of significance, $\alpha = 0.05$, but only for the pair D-Sr ($r=-0.4974$, $\alpha = 0.0500$) and D-Cu ($r=0.5279$, $\alpha = 0.0348$) can be accepted as significant. Cluster dendrograms were constructed by pairing of r without averaging, to avoid cophenetic distortion (Davis, 1973). The observed relationship is less obvious in all F groups (Fig. 7a, b).

Table 2
Correlation matrice

	Footwall rocks								
	D	Hg	Cd	Cu	Zn	Ni	Mn	Sr	Li
D	1.00	**-0.37**	0.09	-0.13	0.21	0.02	-0.14	**0.39**	-0.12
	0.00	**0.03**	0.95	0.46	0.24	0.91	0.43	**0.02**	0.49
Hg	-0.36	1.00	-0.21	0.24	-0.06	0.03	0.08	-0.32	0.09
	0.17	0.00	0.25	0.17	0.72	0.85	0.66	0.06	0.60
Cd	-0.34	0.30	1.00	**0.38**	**0.41**	0.19	0.08	0.11	-0.07
	0.20	0.26	0.00	**0.04**	**0.02**	0.31	0.67	0.57	0.70
Cu	**-0.53**	-0.04	0.18	1.00	**0.55**	**0.37**	0.16	0.01	0.10
	0.03	0.88	0.50	0.00	**0.00**	**0.03**	0.38	0.94	0.58
Zn	-0.25	0.15	-0.20	0.23	1.00	**0.77**	-0.03	0.03	0.07
	0.35	0.58	0.46	0.37	0.00	**0.00**	0.87	0.85	0.71
Ni	-0.36	-0.01	-0.82	**0.67**	0.36	1.00	0.11	0.01	0.09
	0.17	0.96	0.29	**0.00**	0.17	0.00	0.55	0.95	0.60
Mn	-0.46	-0.22	-0.28	**0.68**	**0.49**	**0.88**	1.00	0.03	0.00
	0.08	0.40	0.29	**0.00**	**0.05**	**0.00**	0.00	0.88	0.98
Sr	**-0.50**	-0.24	-0.21	**0.55**	0.47	0.43	**0.66**	1.00	0.27
	0.05	0.37	0.43	**0.03**	0.06	0.09	**0.01**	0.00	0.13
Li	-0.03	-0.37	**-0.52**	0.23	0.42	0.11	0.28	**0.71**	1.00
	0.92	0.16	**0.04**	0.38	0.38	0.69	0.29	**0.00**	0.00

Symbols: boldfaced significant correlation coefficients

Corr. coef.
sig. level



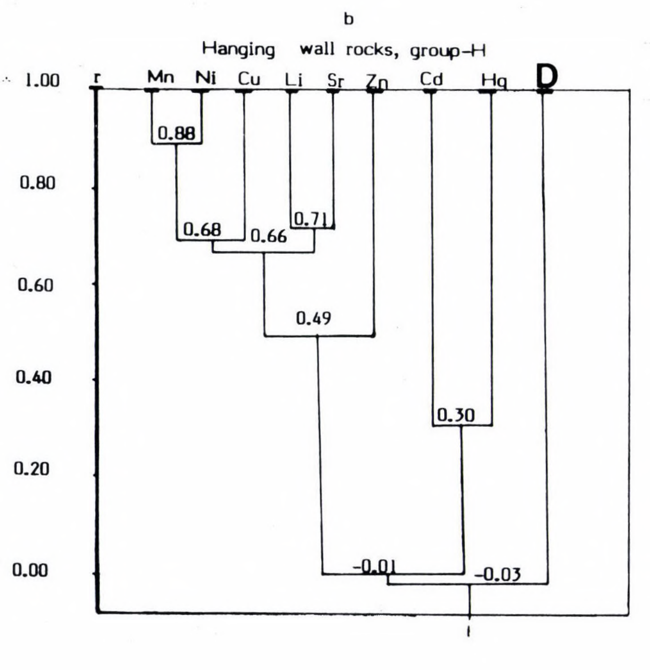


Fig. 7
Cluster dendograms

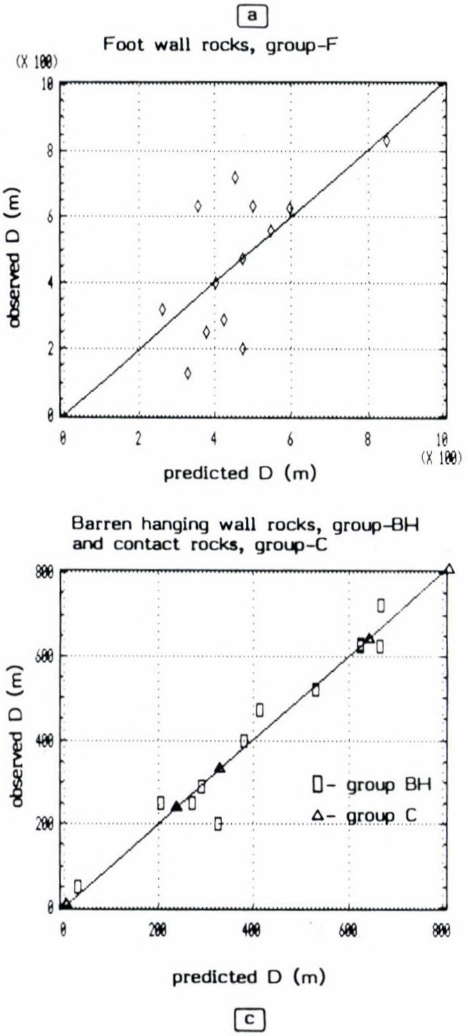


Fig. 8
Multivariate linear regression efficiency

Multivariate linear regression

Multivariate linear regression aimed to get stochastic relation between the distance D (m) and trace element concentration. Clustering pointed out to the groups H with all negative correlation coefficients between D and the trace elements. Model fitting results for the group H are given in the Table 3, case a, with appropriate sign. levels for every partial regression coefficient, R^2, stand. errors and mean absolute error of the regression. Segregation of the highest t-values of the par. reg. coefficients, reduces number of independent variables from 8 to 4 and efficiency of the prediction rose as observed on the increased multiple correlation coefficient $R^2 = 0.4442$ to $R^2 = 0.5659$, and t-values for part. reg. coefficients (Table 3 case b).

Even a better prediction model is gained by the group BH (Table 3, case c), with high $R^2 = 0.8269$. Reduction from 8 to 5 variables, with the highest t-values, increased R^2 to 0.8983 (Table 3, case d). The best fit of the results on the plot; observed vs. predicted values by the multiple regression for the group BH shows the prediction possibility of the method.

The contact rocks, the group C , seems to be the best sampling media for the same purpose, and the multivariate regression analysis gave rise to an extraordinary efficient model, although based on only 5 observations (R^2 = 0.9999, stand. error or estim. = 0.0, Table 3, case e, Fig. 8a, b, c, Bennet and Franklin 1967).

Discriminant analysis

Discriminant function analysis was provided to distinguish the ore-bearing from the barren foot wall rocks and the ore-bearing from the barren hanging wall rocks. Although BxF/BF quotients in the Table 1 do not suggest convincingly distinction between the barren and ore bearing rocks, discriminant analysis provided a successful way of doing it especially for the group BxH/BH.

Characteristics of the discriminant functions are given in the Table 4 with appropriate values of discriminant index R_o, projections of the multivariate means R_1 and R_2, F-test of significance for discrimination function, Mahalanobis distance D^2, relative contribution of each variable to D^2 in % (E), and finally the positive scoring (S) in %.

Discrimination between the groups BxH–BH (Table 4, case a) is successful with 100% scoring in both groups but not significant since Ho: D^2=0 is not rejected (F= 2.4792, with 8 and 6 deg. freedom).

Reducing number of variables to Li, Sr and Cu, due to their highest relative contribution to D^2, discrimination becomes significant (F = 6.6874 with 3 and 11 deg. freedom, Table 4, case b) and keeps scoring at 100%.

Less efficient is discrimination between the groups BxF–BF (Table 4, case c). Small F = 1.2080 (8 and 26 deg. freedom) makes it non-significant although positive scoring for the group BF is 93%, but for the group BxF, however, is low 63%.

Table 3
Multivariate linear regression

Independent variables	coefficient	std. error	t-value	sig. level
A. Group H – hanging wall rocks, D (m) – dependent variable, 16 samples, R^2=0.4442, stand. error=194 m				
Constant	2048.38	2261.65	0.90	0.39
Hg	-299	1.30	-2.28	0.05
Cd	-1.62	1.31	-1.23	0.25
Cu	59.89	139.94	0.42	0.68
Zn	30.04	26.31	1.14	0.29
Ni	4.43	7.55	0.58	0.57
Mn	-9.99	6.77	-1.47	0.18
Sr	-0.71	1.30	-0.54	0.60
Li	-11.99	23.53	-0.50	0.62
B. Group H – reduced number of indep. var. D(m) – dep. variable, 16 samples, R^2=0.5659, stand. error = 171 m				
Constant	949.15	144.68	6.56	0.00
Hg	-1.93	0.82	-2.34	0.03
Cd	-1.27	0.56	-2.25	0.04
Mn	-3.42	2.06	-1.65	0.12
Sr	-1.18	0.61	-1.91	0.08
C. Group BH – barren hanging wall rocks, D(m) – dep. var., 12 samples, R^2=0.8269, stand. error = 88m				
Constant	3833.43	1316.95	2.91	0.06
Hg	-2.71	1.12	-2.40	0.09
Cd	-0.68	0.62	-1.10	0.34
Cu	82.15	75.24	1.09	0.35
Zn	9.77	18.78	0.52	0.63
Ni	1.56	4.55	0.34	0.75
Mn	-5.14	5.08	-1.01	0.38
Sr	-0.18	0.71	-0.25	0.81
Li	-31.33	13.88	-2.25	0.10
D. Group BH – reduced number of indep. vars, D(m) – dep. var., 12 samples, R^2=0.8983, stand. error = 67 m				
Constant	3700.46	609.48	6.07	0.00
Hg	-2.18	0.44	-4.86	0.00
Cd	-0.72	0.37	-1.91	0.10
Cu	68.24	40.00	1.70	0.13
Mn	-4.48	1.86	-2.40	0.05
Li	-29.92	6.18	-4.83	0.00
E. Group C – reduced number of indep. vars, D(m) – dep. var., 5 samples, R^2=0.99, stand. error = 0 m				
D(m) = -1.70 Hg + 4.09 Cd -192.48 Cu + 10.85 Mn + 9.22 Li				

Table 4
Discriminant analysis

Discriminated groups	Mahalanobis D^2	R_{BxH}	R_o	R_{BH}	Variables	Constants	E %	Positive scoring % BxH	BH	F degrees of freedom
Case a	12.9	190.8	183.3	177.9	Hg	-0.12	2.67	100	100	2.47
BxH–BH					Cd	0.39	-4.08			8.6
					Cu	2.67	13.05			
					Zn	0.66	-1.43			
					Ni	0.07	-9.61			
					Mn	0.02	-3.37			
					Sr	-0.18	23.15			
					Li	2.04	79.61			
Case b	7.1	91.5	87.9	84.4	Cu	0.01	9.57	100	100	6.68
BxH–BH					Sr	-0.08	18.60			3.11
					Li	1.01	71.82			
		R_{BxF}	R_o	R_{BF}				BxF	BF	
Case c	2.0				Hg	0.00	2.57			1.20
BxF–BF		12.87	11.88	10.89	Cd	0.01	3.99	63	93	8.26
					Cu	0.85	42.17			
					Zn	-0.39	0.56			
					Ni	0.95	19.03			
					Mn	0.16	12.14			
					Sr	-0.01	0.08			
					Li	0.11	19.47			

Conclusion

Since interest of the investigation has been focused to geostatistics, in order to find an effective exploration tool in looking for the blind ore bodies, geochemical explanation of trace elements behavior in the different rock groups may only be guessed by limited lithogeochemical data.

Trace element behavior in the carbonate media is governed by diversified processes in syndiagenesis, anadiagenesis and epidiagenesis. Variation in their concentration, on the other hand, is also controlled by quantity of oxyhydroxides (Mn), clayey fraction in the insoluble residue (Li) or by carbonate fraction being effected by diagenetic conditions (Sr). Clustering in the hanging wall rocks has gathered Mn, Sr and Li with other elements, while the distance D has persistently negative correlation with all of them (Fig. 7a). Sedimentary processes, proceeding over diversified paleorelief, accumulated bauxites in the negative forms of topography, and obviously shaped primary geochemical patterns in the superficial weathering material and the first, juxtaposed, transgressive, carbonate layer. The geochemical patterns and particularly their size of more than 1000 m with still undetermined margins cannot be attributed to a secondary halo around the bauxite ore body.

Clustering in the foot wall rocks does not confirm the same regularity. They were formed under uniform sedimentary geochemical conditions what gave rise to the homogeneous distribution of Sr (type of diagenesis) and Mn (Eh–pH, indicator) with much smaller stand. deviations and means than in the former (Fig. 6e, f, box-whiskers, Statgraphics 1988).

Statistics with applied testing proved definitely at least two target populations, represented by two sample groups, H and F. For some reason subdivision is extended into the groups BH and BF (Ni, Hg), probably due to spatial variation among the ore fields. Recognition of the separate lithogeochemical groups directed multivariate analysis in a more efficient way.

Clustering after calculation of the correlation coefficients, in the groups H points out to the firm negative stochastic relationship between distance (D) and trace element concentration.

Multivariate linear regression analysis introduced an efficient model for prediction of the blind ore body on the basis of trace element concentrations in the barren hanging wall rocks, the group BH. In addition, promising media for a successful regression model are weathering products at the paleokarst contact between the hanging and foot wall rocks, the group C. It is worth to be examined on a larger number of samples in continuation of the investigation.

Discriminant analysis enables efficient recognition of the barren and ore bearing hanging wall rocks on the basis of trace element concentrations, although appropriate ratio BxH/BH for mean values of Sr and Mn does not suggests that. Reduced number of trace elements (Li, Sr, Cu) enhances discrimination to the significance ($\alpha = 0.05$) as determined by F-test.

Results of applied statistical analysis allow significant constraints to be placed on such models. It is enough promising to be continued on a larger number of the hanging wall rocks and particularly of the contact material. Additional geochemical characteristics as well as other numerical variables can be introduced into the model to seek for further improvements.

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The authors are indebted to many people from the mine, who assisted in their own way during collection of the samples and performance of the investigation. Unfortunately, many of them will never be able to accept our acknowledgment due to tragic events that stroke Bosnian lands. We hope that our research might benefit operation of the Jajce bauxite mine at some better times.

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