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Geologic setting of the Pre-Tertiary basement in Vojvodina (Yugoslavia)
Part I: The Tisza Mega-unit of North Vojvodina

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Using the database acquired by lithostratigraphic, biostratigraphic and petrographic analysis of borehole samples, information on horizontal and vertical facial extensions and the recognized general stratigraphic relations in the Fruška Gora and southern margins of the Pannonian Low, three major structural units are distinguished in the Pre-Tertiary Complex of Vojvodina. Their developments are important for the paleogeographic reconstruction of the Pre-Tertiary events:

1) The Tisza Mega-unit of North Vojvodina,
2) the Ophiolitic and Flysch Belts of Central and Southern Vojvodina,
3) the South Carpathian-Balkanian and Serbo-Macedonian Belts of south-eastern Vojvodina.

Paleontological analyses indicate that the complete Mesozoic system is developed in the Pre-Tertiary basement of the Pannonian Basin in Vojvodina.

The Tertiary complex in Vojvodina, together with the Mesozoic formations is also underlain by crystalline schists and granitoids, to date interpreted as of Proterozoic and Paleozoic age, and by the recently recognized Younger Paleozoic rocks affected only by very low-grade metamorphic processes.

This study will consist of two parts, the first dedicated to the geologic formations underlying the Tertiary in North Vojvodina, identified by their lithofacial and structural characteristics as part of the Tisza Mega-unit, the second part will deal with the southern part of Vojvodina.

The Northern part of the Tisza Mega-unit in Vojvodina is made up of the Precambrian(?) or Cambrian Horgoš Series, composed of polymetamorphic paraseries with granitoids. In the southern part of North Vojvodina are developed the Mokrin Series. They are a metavolcanogenic-sedimentary association intruded by granodiorites with migmatites. The crystalline complex is overlain partly by Devonian–Lower Carboniferous metaclastics or by Permian continental clastites and volcanites, followed by Lower Triassic transgressive clastites and Middle and Upper Triassic (generally shallow-water) limestones. On the southern margin of the Tisza Unit, the crystalline basement is partly overlain by clastic-carbonate or flysch deposits of the Campanian–Maastrichtian.

During Alpine tectogenesis, the granite-metamorphite complex with its Triassic cover was subjected to compressional folding and overthrusting in North Vojvodina.

Key words: Vojvodina, Pre-Tertiary basement, wells, geologic formation, stratigraphic analysis, lithostructural units

Introduction

Within the region of Vojvodina, surface outcrops of crystalline and Mesozoic formations are found only on Vršac and Fruška Gora mountains (Fig. 1). Attracting the attention of many researchers, they were subject of investigations

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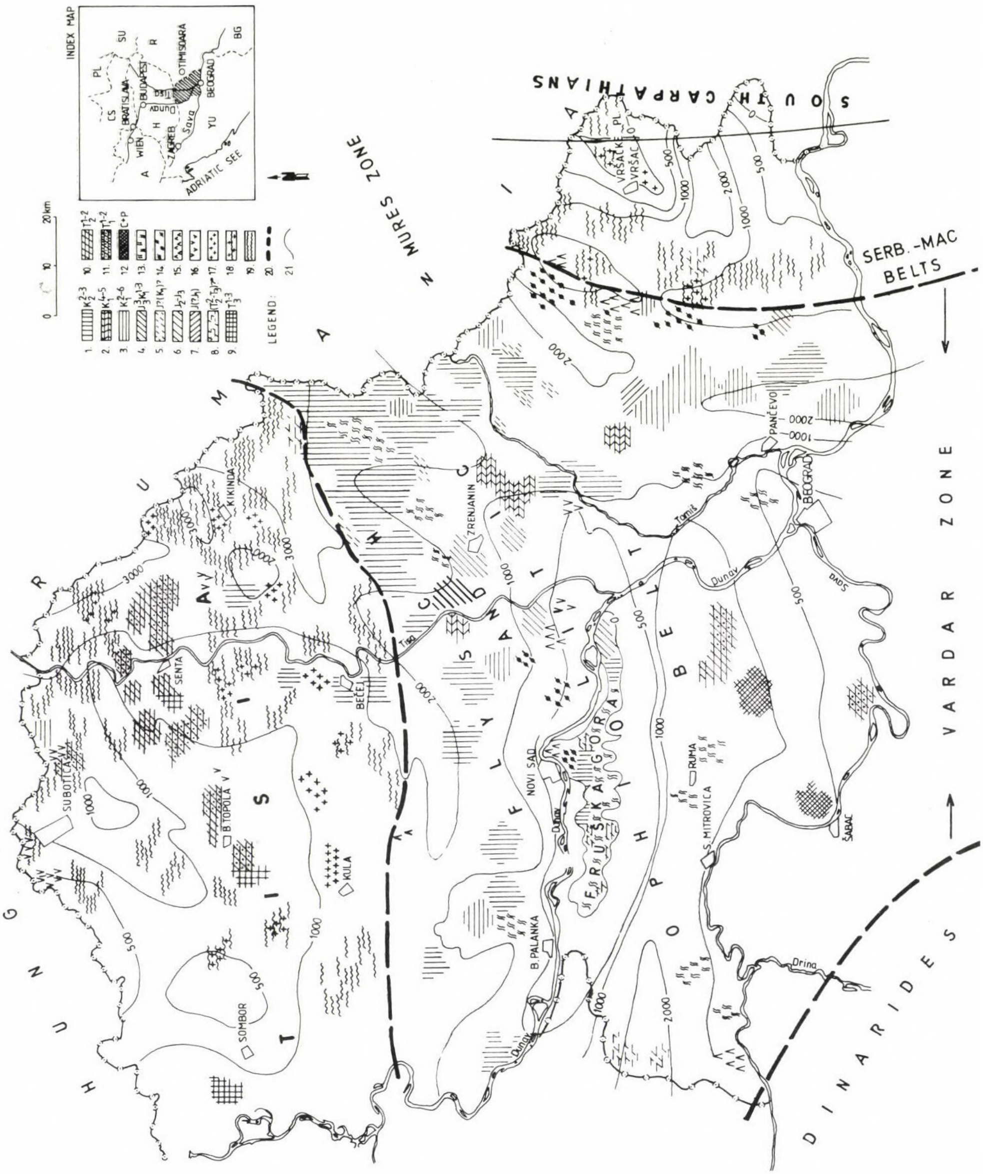
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as far back as the last century. The results obtained to date are presented in publications given in the list of references.

The crystalline schists, magmatites and the Mesozoic formations in the buried portion of the Tertiary basement have become accessible to investigation thanks to the deep wells drilled for hydrocarbon exploration.

The crystalline basement and the Mesozoic deposits lie at varying depths beneath the Tertiary formations. Their burial depths range from several hundreds meters to more than 3300 m. The maximum burial depths of the Mesozoic series occur in Central and North Banat (Fig. 1).

The crystalline basement was penetrated by drilling at apparent thicknesses ranging from 3 to maximum 86 m. There are no available data on actual thicknesses of most Mesozoic and of Pre-Mesozoic units. Some of the wells only reached the coarse clastic sequences of metamorphic material produced by erosion.

The thickness of penetrated Mesozoic series is usually expressed in terms of several tens and less frequently of several hundreds of meters. Only five wells drilled through 1000 to 2000 m of Mesozoic. Six wells penetrated into the crystalline schists in the base of the Mesozoic.

Generally, the objectives of petroleum exploration projects were the accumulations in Tertiary pools and the interest in the underlying formations was rather academic. This partially explains the lack of cores. To aggravate the situation, only a small number of cores recovered during early explorations were preserved and saved. The lack of core material set certain limitations to our investigations. More samples than available would be necessary to define the series poor in paleontological evidence or without sufficiently indicative fossil content. This statement applies principally to the basinal series which have often been drilled in survey area.

Much of the core material was recovered from tectonically disturbed (crushed) series and biostratigraphical data were difficult to acquire.

The infrequent cores did not provide sufficiently reliable paleontological and isotopic geochronological data, which resulted in the lack of accurate stratigraphic interpretations. There is a certain limitation for isotopic analysis application due to the low number of cores and to the intense changes occurring

← Fig. 1

Geologic map of the Pre-Tertiary basement of the Pannonian Basin in Vojvodina (according to well data). From Čanović, M. and Kemenci, R. (1988). 1. Upper Cretaceous, basinal facies (clastic-carbonate and flysch); 2. Lower Cretaceous, shallow-water and reef facies (Urgonian); 3–6. Lower Cretaceous, Upper and Middle Jurassic basin, pelagic and deep-water facies; 7. Lower Jurassic, basin facies; 8. Upper and Middle Triassic, various basin limestones recrystallized, partly marbled, calcschists, silificied, brecciated; 9. Upper Triassic, shallow-water facies (carbonate platform); 10. Middle Triassic, shallow-water and partly semipelagic facies; 11. Lower Triassic, shallow-water facies (clastic-carbonate and evaporite); 12. Permian–Carboniferous, shallow-water facies; 13. serpentinites; 14. gabbros; 15. diabases, spilites; 16. rhyolites; 17. granites, granodiorites; 18. migmatites; 19. crystalline schists (generally); 20. megastructural unit boundary; 21. isobaths, approximated from well data on samples taken from the surface portions of the paleorelief. Some tens of isotopic analyses, mainly related to magmatites, are available at this time (Lovric 1981–86; Árva-Sós et al. 1993).

Besides the petrographic and paleontological investigations, the following methods have been applied on a minor scale so far: X-ray diffactional determination of mineral compositions (Žumberković 1993–94) and a few chemical analyses of magmatic rocks.

The evaluation, or rather approximation of facial stratigraphic and structural features of the buried Pre-Tertiary basement is, however, made possible owing to the large number of wells. The understanding of its development is improved by correlation of the area of interest with the adjacent ones in Hungary, Romania and Yugoslavia, studied by many authors.

Litho-structural units of the Pre-Tertiary basement in Vojvodina

The Pre-Tertiary basement in the part of the Pannonian Basin belonging to Vojvodina comprises several litho-structural units with different geologic compositions. The extents of these units in the Pre-Tertiary base were identified in the first studies published on this area (Aksin and Karamata 1954; Nikolić and Kemenci 1962; Kemenci and Čanović 1975; Kemenci 1977; Čanović 1971–72; Kemenci 1981; Čanović and Kemenci 1988; Kemenci 1991). As the number of wells increased, these extents become clearer.

The subsided Pre-Tertiary basement with local exposed parts in Vojvodina seems to exhibit three regional litho-structural units with highly differing geologic composition (Fig. 1):

1) The Tisza Mega-unit of North Vojvodina,

2) the Ophiolitic and flysch belts of central and South Vojvodina,

3) the South Carpathian-Balkanian and Serbo-Macedonian Belts of south-eastern Vojvodina.

The regional structural units distinguished above are separated by marked dislocations-lineaments associated with extrusions of younger magmatic melts.

Geology of the Tisza Mega-unit of North Vojvodina

According to Kovács (1984–87) the granite-metamorphic complex of Northern Vojvodina together with shallow-water Triassic deposits, are considered to be the southern part of Tisza Mega-unit (Fig. 2). The southern boundary, according to well data, runs in an E–W direction in the central part of Vojvodina. The ophiolitic and flysch belts extend southward (Fig. 1).

The southern part of the Tisza Mega-unit is made up of Precambrian, Paleozoic and Mesozoic formations overlain by thick Tertiary deposits.

The Precambrian(?) or Cambrian Horgoš Series is composed of polymetamorphic paraseries with granitoids reached by wells in North

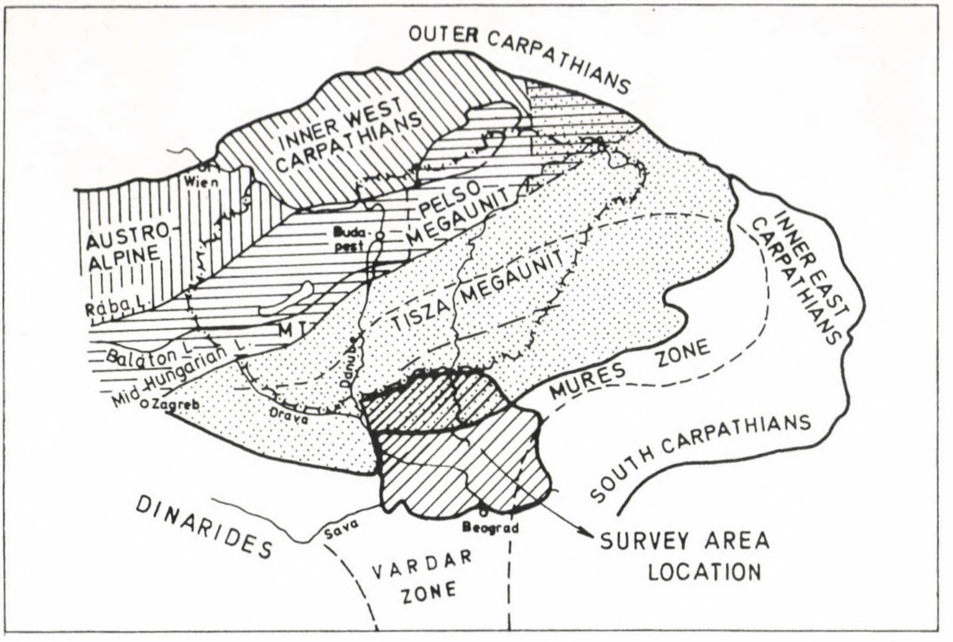


Fig. 2
Survey area location on geotectonic sketch of Carpatho-Pannonian region (after Haas et al. 1995)

Vojvodina, near the Hungarian frontier (Čikeriija, Kelebija, Palić, Subotica, Horgoš, Male Pijace, Kanjiža, Bikovo, Velebit).

These deposits are overlain partly by Devonian–Lower Carboniferous metaclastics or by Permian continental clastites, quartz-porphyry lavas and pyroclastites, followed by Lower Triassic transgressive clastites and evaporites and Middle and Upper Triassic (generally shallow-water) limestones (Fig. 4).

Some of the wells (near Srpski Krstur, Majdan and Banatski Arandelovac) reached only the coarse clastic sequences consisting of metamorphic material produced by erosion, which are covered by Miocene or Pliocene conglomeratic breccias, marlstones or marls.

South of the area with the Horgoš Series are developed the Mokrin Series. Their components are a metavolcanogenic sedimentary association intruded by granodiorites with migmatites. These are covered partly by Devonian–Lower Carboniferous metaclastics, or partly by Miocene–Pliocene transgressive deposits.

In the southern part of the Mokrin Series, the crystalline basement is partly overlain by the conglomerates, detritic organogenic limestones and sandy limestones of the Upper Cretaceous. Their cover is the Miocene–Pliocene complex.

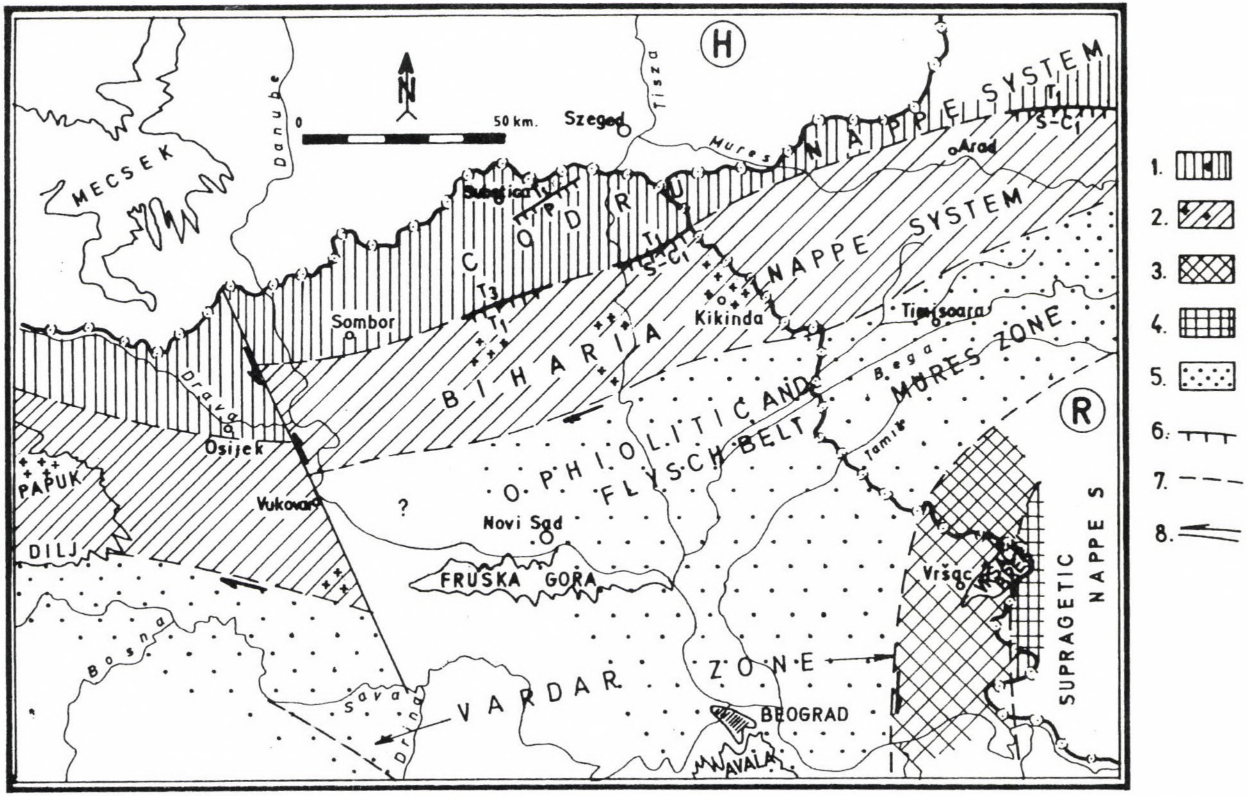


Fig. 3
Approximate extension of structural units in the Pre-Tertiary Basement in Vojvodina and Tisza Mega-unit, granite-metamorphic complexes: 1. Approximate extension of Codru nappe system, diaphthorites-metamorphic rocks of greenschist facies, Caledonian granitoids, Triassic limestones; 2. Approximate extension of Biharia nappe system, metamorphic rocks of greenschist facies, Pauiseni Series, Hercynian granitoids; 3. Serbo-Macedonian units: metamorphic rocks of amphibolite and greenschist facies with Hercynian granitoids and migmatites; 4. South Carpathian-Balkanian belt: Precambrian and Caledonian metamorphic rocks of amphibolite facies; 5. Approximate extension of the ophiolitic and flysch belt, Triassic, Jurassic, Cretaceous to Paleogene; 6. Overthrusts; 7. Unit boundaries of overthrusting type; 8. Zone of horizontal faults

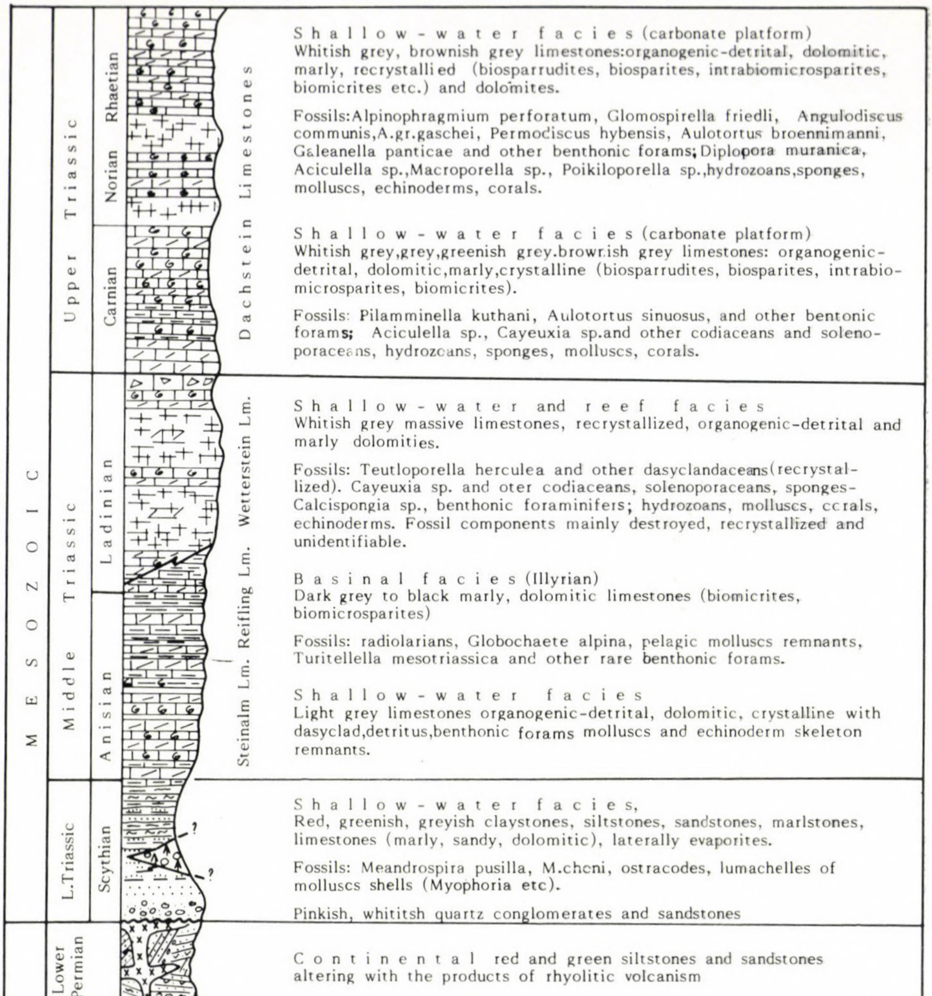


Fig. 4
Schematic column of the development of the Proterozic–Paleozoic and Triassic in North Vojvodina (on well data basis)

Pre-Mesozoic formations of the Tisza Unit in North Vojvodina

Horgoš Series

The crystalline Horgos Series was penetrated by wells only between 3 to a maximum of 86 m; therefore there are no data on its total thicknesses.

Most of the wells penetrating the crystalline Horgoš Series of North Vojvodina drilled through retrograde-metamorphic, cataclastic and mylonitic granitoids, gneisses, micaschists, created within the amphibolite facies and a staurolite–almandine subfacies conditions during the Caledonian Orogeny. Later remobilization during the Hercynian Orogeny caused manifold changes. The Alpine tectonic destruction caused foliation or mm to cm-size crenulations, often cataclization or mylonitization. Syntectonic textures such as helicitic albites, and syntectonic garnet crystals in schists as well as post-tectonic deformation of micas and chloritoids and crenulation reflect these processes (Plates I, II, III, V, VII, VIII).

Today the mineral associations in the metamorphites include:

a) chlorite–garnet–quartz–muscovite,
b) chlorite–garnet–quartz–muscovite–biotite,
c) chlorite–quartz–muscovite–biotite–albite,
d) chlorite–quartz–albite–muscovite–sericite.

The metamorphites and associated granitoids in the northern part of North Vojvodina are lithofacially identical to the Precambrian–Cambrian parametamorphites and granitoids of the Apuseni Mts (Szepesházy 1966, 1978; Dimitrescu 1981) which were metamorphosed and intruded by granites during the Caledonian Orogeny. Lithostratigraphically the Horgoš Series correlate with the oldest metamorphites, which means that some similarities with the Baia de Aries Series (Codru Mountains) may be observed.

The Mokrin Series

The Mokrin Series is constituted by a metamorphosed volcanogenic-sedimentary association intruded by granodiorites with migmatites. It is developed in the southern area of North Vojvodina (Fig. 1). The protoliths of the metavolcanosedimentary association were mafic and felsic volcanites, their tuffs, as well as breccias, conglomerates, sandstones, marlstones and claystones. They were metamorphosed to orthoamphibolites, paraamphibolites, amphibole–epidote schists, quartz–muscovite–chlorite schists, and quartz–biotite–muscovite schists with albite porphyroblasts and metarhyolites (Plates IV, VIII).

This low-grade metamorphic volcanogenic and paraseries is lithofacially identical to the Biharia and Muncel series in the Apuseni Mts (Dimitrescu 1981). The series are transgressively overlain by the Devonian–Lower Carboniferous Paiuseni Series, which is lithofacially identical to the metaconglomerates and breccias from Kikinda and Majdan.

Granitoid rocks from the Mokrin Series consist of coarse cataclastic granites and granodiorites associated with migmatites. Their silica and alkali content places them generally among the granodiorites (Radukić and Kemenci 1989). The mineral composition of the rocks includes K-feldspar, plagioclase, quartz, biotite and phengitic mica, in association with accessory apatite, sphene, rutile and zircon, plus secondary chlorite, sagenite, epidote, siderite, sericite, kaolin and leucoxene. Feldspar grains range from 1 to 2 cm. Plagioclases are commonly albitized, sericitized and/or epidotized. Twinned plagioclase lamellae are often deformed due to postcrystallization processes (Plates V, VII). In contrast to plagioclase microcline is found as coarse unaltered grains. It is cross-hatched with wavy extinction and perthitic intergrowths. The fissures are filled with mylonitic quartz, colourless mica and fine green biotite. Quartz is generally crushed by mechanical action to variably sized grains. Biotite occurs in 1 mm long flakes, and is the only coloured rock constituent. Biotite flakes generally exhibit undulatory extinction and uneven edges. Retrograde metamorphosis has partially altered biotite into chlorite (Fe-Mg chlorite) with sagenite and sphene, the latter partially replaced by leucoxene in peripheral areas. Colourless mica displaying bright interference colours identified with the electron rnicroprobe as phengite mica appears in association with biotite. The granite sample shows that biotite was sometimes blanched into phengite. The presence of phengite indicates retrograde processes during the Alpine Orogeny under significant H_2O pressures.

Migmatites with granodiorites were encountered in wells in the southern part of North Vojvodina (Fig. 1). The metablastic transformation of existing metamorphic rocks in zones of granodioritic masses resulted in the formation of migmatites exhibiting bedded and lensed structure, with more or less preserved original rock structure. The prevailing feldspar forms are metablasts of young microclines (Plate VI, Fig. 1).

These formations were penetrated for the first time in the Bečej area. Aksin and Karamata (1954) identified them as biotite micaschists developed in the mesozone by metamorphosis of clayey and sandy rocks. The same series includes two-mica gneiss with significant microcline and microcline perthite. Karamata in these study believed that the gneiss developed from igneous rocks, as can be surmised by the volume of potassium feldspar absent in the adjacent rocks. He concluded according to chemical test results that the Becej gneiss had an alaskite–aplite composition. Taking into consideration that in the vicinity (well Bč-6). Later granites were drilled, it appears that these schists belong to the migmatite zone.

The metablastic transformation of existing metamorphic rocks in the zones of granodioritic masses also resulted in the formation of large albite metablasts which poikiloblastically buried the previously-formed grains of quartz, mica, and garnet. The proximity of granite magma also had an effect upon the introduction of potassium, thus influencing the development of porphyroblastic flakes of muscovite and resulting in biotitization of schists. A common constituent of these rocks is green-coloured zonal tourmaline (Plate IV, Fig. 2).

The metamorphites with very abundant porphyroblasts of albites were recognized in a series of wells along the southern margin of the Tisza Mega-unit.

The formation time of granitoids has not yet been incontestably determined. Radiometric analysis of granitoids in North Vojvodina has included granite samples taken only at Kikida and Milosevo. According to Árva-Sós et al. (1993), the granite at Kikinda crystallised 306.6±11.6 my ago (K/Ar age of biotite). Radiometric dating of granitoid at Battonya (Hungary) which is on the same structure as the granites of Mokrin and Kikinda, indicates that it developed 386 my ago, during the Hercynian orogenetic cycle. According to the isotopic documentation, Pamić (1986) considers the granites found in the Slavonian mountains to belong to the Hercynian Orogeny. Lovrić (1986) reports only one result of a Ka/Ar-method biotite monomineral concentration measurement, on granite from Miloešvo equal to 131 my, corresponding to a Lower Cretaceous–Portlandian age, which is indicative the effects of rejuvenation during Alpine tectogenesis.

Devonian–Lower Carboniferous(?) metapsephites and metapsamites

The slightly metamorphosed, probably Carboniferous breccias and sandstones occur at Mokrin and Kikinda. These breccias were formed from fragments of old metamorphites and contain gneiss, micaschists and chlorite schists. The cement is sericite and chlorite. According to Szepesházy (1973) these anchimetamorphic breccias, conglomerates and sandstones are comparable to the blastodetrites of Devonian–Carboniferous age of the Paiuseni Series in the Bihor Mountains.

The metabreccia in the Palic well consists of crystalline schists and granitic fragments in black sericite–graphite matrix. The succeeding light grey to white metabreccias are rich in muscovite aggregates and quartz–muscovite epidote schist fragments. Thanks to its lithofacial characteristics the series could be correlated to the Carboniferous d'Arisieni Series of the Bihor Mountain.

Permian clastites

The Carboniferous(?) metabreccias of Palić are overlain by low-grade metamorphic, schistose, reddish brown to greyish green clastites. Identical siltstones and sandstones, always fossil free, were penetrated by a number of wells at Palić. The grains are angular to subrounded, poorly sorted with sericite cement. They alternate with the products of rhyolitic volcanism, which in the adjacent areas in Hungary belong to the Permian (Kassai 1981).

Rhyodacite

Some of the Palić, Kelebija, Stari Grad and Horgoš wells drilled into rhyodacite lavas and tuffs. These volcanics occur generally in base of the Neogene, but also below the Lower Triassic deposits or within the Permian formations. The wells at Kelebija reached total depth in volcanic bodies or tuffs. The volcanites have been traversed only by one well in the Horgoš area, where their thickness is 25 m. This particular well was drilled some 15 m further into the crystalline schists. The structure of the lava is fluidic. It contains coarse quartz and feldspar phenocrystals. The quartz is corroded, fractured and along the fractures sericitized phenocrystals prevail. Feldspar phenocrystals are less abundant. Polysythetic twinned plagioclase with 30 to 32% An have been observed. Sanidine crystals are rare. Biotite is the only coloured constituent. The biotite occurs in large, hexagonal flakes. Their opacity indicates lava flow. The vitric groundmass is generally recrystallized into microcrystals and sericite flakes. In some of the Palić wells, the Campilian sandy-carbonate sediments include reworked rhyodacite fragments.

Mesozoic formations of the Tisza Unit in North Vojvodina

Microfossil evidence (Čanović 1967–86) indicates the presence of Triassic, i.e. Lower (Seisian and Campilian), Middle (Anisian and Ladinian) and Upper Triassic (Carnian, Norian, Rhaethian) and Upper Cretaceous sediments in this particular area.

Lower Triassic

The environmental conditions in the Lower Triassic were similar to those of the Alpine Lower Triassic; clastic and carbonate rocks were deposited along with evaporites in hypersaline lagoons and sabkhas. Lower Triassic evaporites have been penetrated in identical lithofacies of tidal intraformational conglomeratic breccias with anhydrite (Plate IX, Figs 1, 2) in two wells (Bajsa, Crna Bara) approximately 60 km apart from each other. By origin, anhydrite is primary or secondary. Primary anhydrite is contained in mud-pebble fragments resulting from destruction of laminites into micrite–siltstone–anhydrite. Secondary anhydrite is the main component of tidal conglomeratic breccia cement and appears as relatively coarse crystals together with sparry calcite, dolomite and rare gypsum crystals.

The Lower Triassic (Campilian) contains only poor microfossil associations, the most common occurrence being that of *Meandrospira pusilla* and *Myophoria* shells (Plate X, Fig. 1).

Middle Triassic

The Middle Triassic Anisian beds in North Vojvodina were deposited in a shelf environment. This shallow-water facies consists of organogenic, detrital limestones, rather dolomitized, containing detritus of dasycladaceans and benthonic foraminifera (Plate X, Fig. 2).

The basin exhibited a tendency to subside in the Upper Anisian (Illyrian). Sediments of semipelagic type were laid down, consisting of dark grey, marly, microcrystalline limestones with microfossil associations indicating pelagic influences: radiolarians, globochaetae, remnants of pelagic molluscs and rare benthonic foraminifera – *Turitellella mesotriassica*, etc (Plate XI, Fig. 1).

Semipelagic deposition continued into the Lower Ladinian, but during the main part of the Ladinian shallow-water, predominantly reefal and peri-reefal deposition is characteristic. The deposits consist of massive organogenic, detritic limestones, carbonate breccia and dolomites. The microfossil content includes various algae: *Teutloporella herculea*, codiaceans, solenosporaceans, sponges and other metazoic organisms associated with the reefal environment (Plate XI, Fig. 2).

Upper Triassic

Shallow-water deposition, characteristic of carbonate platforms, continued through the Carnian, Norian and Rhaetian stages. In the intertidal, subtidal and reefal zones various limestones (biomicrites, biosparites, intrabiomicrosparites, biosparrudites, etc.) were deposited and subsequently dolomitized (Plate XII, Figs 1,2; Plate XIII, Fig. 2). The Upper Triassic carbonates contain shallow-water, reefal and peri-reefal biocenoses: various benthonic foraminifera, dasycladaceans, codiaceans, solenosporaceans, sponges, ostracodes, hydrozoans, corals, molluscs, echinoderms and other biodetritus (Plate XIII, Fig. 1, Plate XIV, Figs. 1, 2).

The microfacial features of the Triassic deposits in North Vojvodina area correlatable with the Triassic in the Alps, the Dinarides and the Carpathians (Wetterstein, Dachstein, Steinalm, Reifling and other formations).

Upper Cretaceous

On the southern margin of Tisza Mega-unit in Vojvodina three wells drilled into Senonian and Campanian–Maastrichtian deposits which transgressively overlie the crystalline schists (Fig. 1).

As observed in the core material from the first northern well, the deeper horizons contain fossil-free subarkose. Upward, these deposits are succeeded by arkose sandstone with coal. No microfossils were observed in this sequence. Next in the series, detritic organogenic and sandy limestones and breccias occur. The following microfossil content was identified: *Siderolites vidali* Douville, *Marsonella trochus* (Reuss), *Operculina* sp., miliolids, textulariids and coralline algae, *Archaeolithotamnium* sp., dasycladacean detritus, ostracodes, rudist (radiolarian) shell detritus, echinodermata and coral skeletons.

The conclusion to be drawn from the fossil content is that these deposits developed in shallower environments, such as peri-reefal areas and the shelf margin.

In the other two wells on southern margin of the Tisza Mega-unit (Bečej), the deeper horizons contain fossil-free silt and sandstones. Upward, sand, siltstones and marlstones manifest turbiditic features. The following microfossil content was identified: *Globotruncana arca* (Cushman), *Globotruncana linneiana* (d'Orbigny), heterohelicides, globigerionides, etc.

The facial characteristics of these Campanian-Maastrichtian deposits are clastic-carbonate and flysch facies.

Evolution of the survey area

The crystalline schists seem to originate from the presumably thick, Proterozoic geosyncline deposits. According to Dimitrescu (1978), dynamometamorphic processes which affected the Somes and Baia de Aries Series (lithofacially very similar to the schists in North Vojvodina) occurred early in the Caledonian Orogeny.

Szepesházy (1966, 1978) shares the opinion that the schists in the area near the Hungarian border belong to the Caledonian cycle. Krautner and Savu (1978) and Jantsky (1979) believe that the Somes and Baia de Aries-type crystalline schists are even older, and suggest Dalslandian or Baikalian as the more appropriate dating.

In one of the above-mentioned cycles, the Proterozoic greywacke and claystones were subjected to metamorphosis under conditions of the amphibolite facies, or rather the Barrow type (B_{2-1}) staurolite–almandine subfacies.

Later remobilization during the Hercynian Orogeny caused many changes. The Variscan metamorphism influenced Devonian–Lower Carboniferous sediments in a low grade. With this orogenesis were associated granitic intrusions. During the Alpine Orogeny, these rocks were subjected to diaphthoresis to a lesser or greater degree; therefore the most common metamorphites of the complex are diaphthorites from the quartz–albite–muscovite–chlorite (B_{1-1}) to the quartz–albite–epidote–almandine (B_{1-3}) subfacies of the greenschist facies.

In the continental phase of the younger Paleozoic, the existing crystalline schists and granitoids were denuded. Coarse clastic deposits, with clay and coal (now sericite and graphite) as cement probably accumulated in the Carboniferous fluvial cones and taluses. The low-grade metamorphosis of these breccias occurred in one of the Hercynian phases.

The Permian deposits also developed also in continental, fluvial or lacustrine environments as fine-grained sandstone and siltstone facies. These sandstones and siltstones are often tuffaceous and contain rhyodacite extrusions attributable to volcanic activities in the Saalic orogeny. The regional subsidence which began late in the Permian continued into the Lower Triassic.

The Lower Triassic transgression over the basement made up of crystalline schists and Permian sandstones and conglomerates included extensive areas which today correspond to North Vojvodina according to deep well evidence. Well-rounded and mature conglomeratic sandstones and quartz sandstones were deposited during this Mesozoic transgression (Kassai 1981). The deposits became subarkosic. This sequence fines upwards, sandstone sorting becomes poorer, and the share of carbonates in the Campilian increases.

The Lower Triassic is represented by the shallow-water clastic-carbonate and evaporite facies of Alpine type. Apparently there were no events to interrupt deposition from the Lower to Middle Triassic. The Anisian is developed in the shallow-water, carbonate shelf facies. The microfacial types observed in cores from the Novi Kneževac wells indicate an Upper Anisian (Illyrian) trend to subsidence. This tendency and connection with the open sea probably lasted throughout the Lower Ladinian. Shallow-water deposition continued, however, through the Ladinian well into the Upper Triassic, when a carbonate platform developed in North Vojvodina.

There is not enough evidence to conclude whether there was a break in the depositional sequence at the end of the Upper Triassic, but Jurassic and Lower Cretaceous as yet have not been detected in this area.

On the southern margin of the Tisza Unit in North Vojvodina, according to previous results the crystalline schists are overlain transgressively by Campanian–Maastrichtian clastic-carbonate or partly by flysch deposits.

During the Alpine tectogenesis, the granite–metamorphite complexes with Triassic cover were subjected to compressional folding and overthrusting, in North Vojvodina. The rocks were subjected to regionally intensive retrograde metamorphosis or diaphtoresis. The effects of rejuvenation are noted on the granite–metamorphic complexes of North Vojvodina.

Szepesházy (1978) compares and relates the granite–metamorphic complex of South Hungary and North Vojvodina with crystalline features in the Apuseni Mts (Romania) and observes the extension of the Codru and Bihor overthrusts westward into North Vojvodina. Dimitrescu (1981) introduced a hypothesis on the structural framework of the south-easthern basement of the Pannonian Basin, where, like Szepesházy (1978), he observes the extension of the Bihor and Codru Structural units in the northern part of Vojvodina. In his doctoral thesis, Szederkényi (1984) distinguishes among various lithostratigraphic formations differing both in their lithology and in the degree of metamorphosis and deformation. He has proved the existence of overthrusts identical to those in the Apuseni Mts, suggesting that they are also present farther to the south in basement of the Tertiary in North Vojvodina.

As will be shown later, this proposition is also confirmed by our own observations. Canovic and Kemenci (1988) presented evidence of the Lower

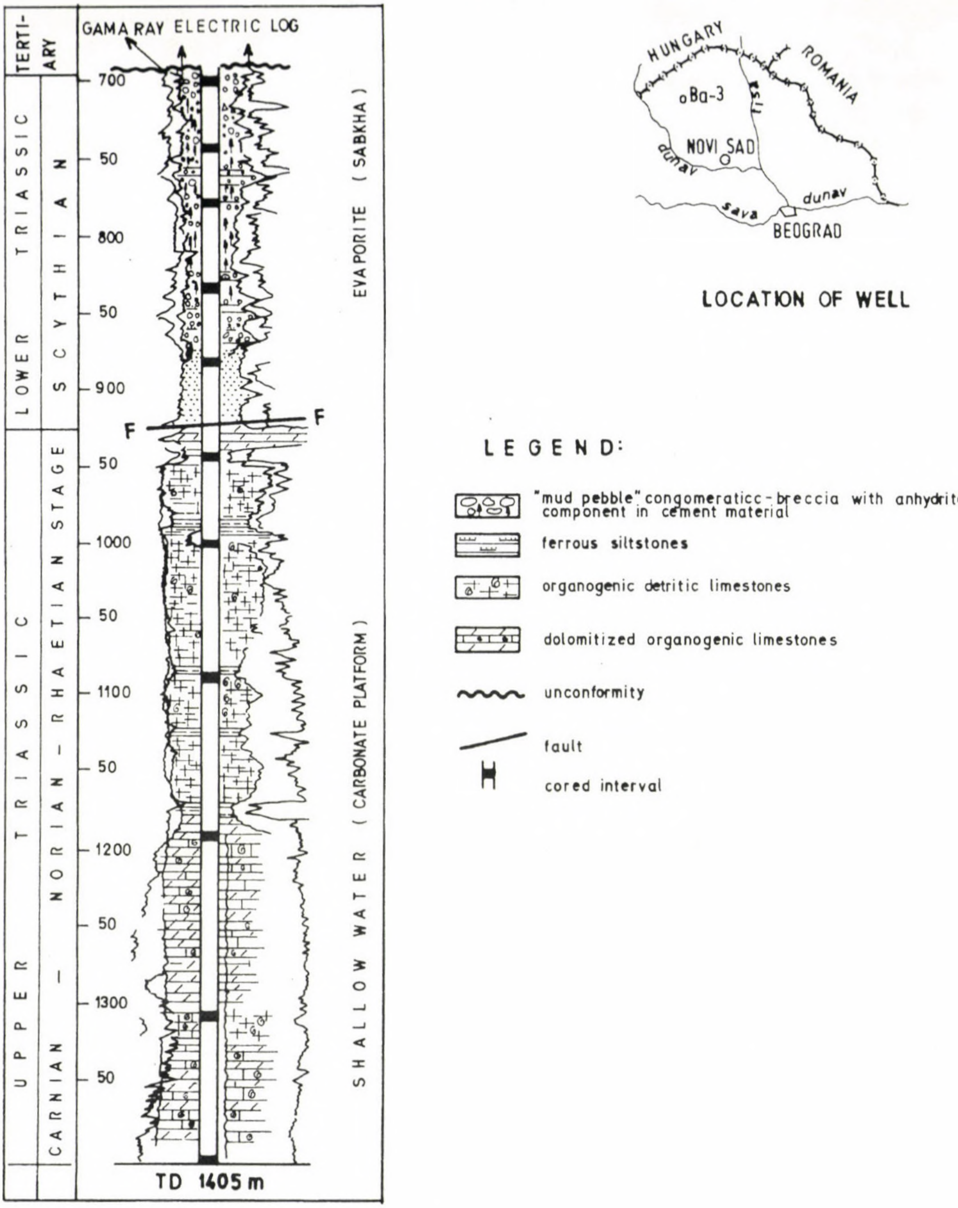


Fig. 5
Part of stratigraphic column in well Bajsa-3

Triassic evaporites overthrusting the Upper Triassic deposits in the Bajša-3 well in North Bačka (Fig. 5). In the Palić-1 and Palić-2 wells, above the paleontologically proven non-metamorphic Lower Triassic deposits, up to 81 m -thick metabreccias were identified, consisting exclusively of schist and granitoid fragments in a metamorphic matrix (Fig. 6). Most likely these breccias are portions of the Szeged-Békés allochtonous zone in the geotectonic unit of southern Hungary (so named by Bérczi-Makk (1985) and Kázmér (1986)), or a part of Codru overthrust, as interpreted by Szepesházy (1978) and Dimitrescu (1981).

According to Kemenci (1991), in the Crna Bara-2 well in northern Banat, the cataclastic granitoids of thicknesses exceeding 100 m are overthrusted onto Lower Triassic conglomeratic sandstones, and most likely represent a part of the Biharia nappe system.

The data published by Pamić (1986) seems to indicate that the granite-metamorphic complexes of the Slavonian Mountains were formed at the same time and under similar conditions. The granite-metamorphic complexes of North Vojvodina are now displaced by horizontal faults northward of the Slavonian Mts (Fig. 3).

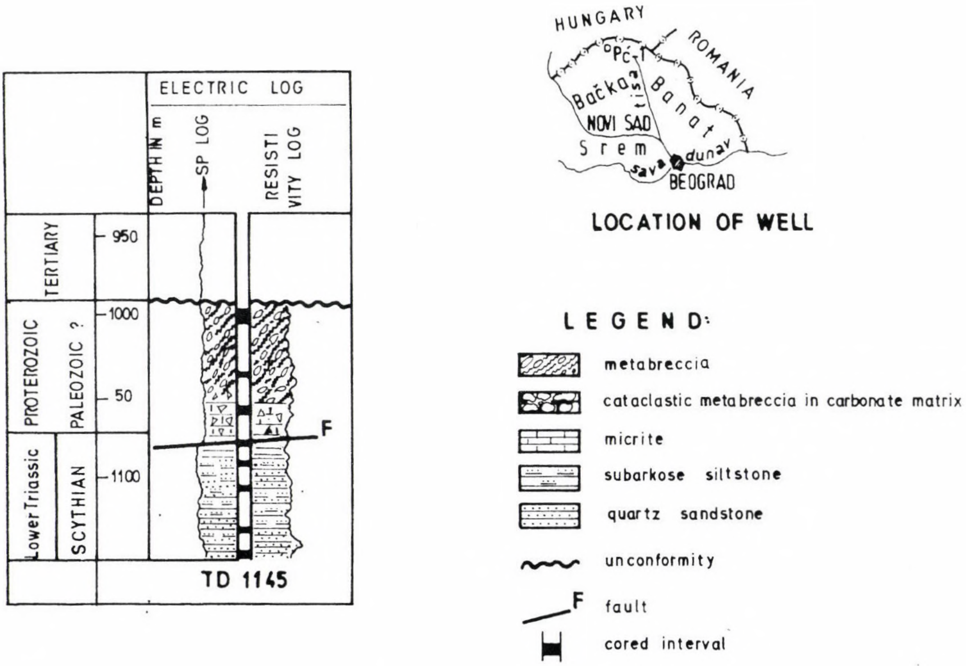


Fig. 6
Part of stratigraphic column in well Palić-1

The regional structures distinguished here are severed by marked dislocations/lineaments, through which younger volcanic rocks extruded.

The presentation of available knowledge of the rocks in subsided Tertiary paleoreliefs seems to confirm the hypothesis of the extension of the Apusenides with the Codru and Bihor Nappes into North Vojvodina. Lithofacial similarities are supported by some paleontological and isotope test results; however, to prove them further corroboration is needed.

Plate I

1. Cataclased crystalline schist, crossed nicols, scale bar 0.2 mm borehole Ho-2, core from 2334–2339 m
2. Cataclased quartz-albite-muscovite schist, crossed nicols, scale bar 0.2mm borehole Kz-l, core from 1403–1405 m
   Horgoš Series. Pre-Cambrian (?) or Cambrian (?)

Plate II

1. Foliated and crenulated quartz-albite-muscovite schist. crossed nicols, scale bar 0.2 mm, borehole Pj-2, core from 1254–1303 m
   Horgoš Complex. Pre-Cambrian (?) or Cambrian (?)
2. Quartz-muscovite-albite schists with helicitic albite containing contorted S_I (post tectonic textures) crossed nicols, scale bar 0.2 mm, borehole Kz-1, core from 1493–1445 m
   Horgoš Series. Pre-Cambrian (?) or Cambrian (?)

Plate III

1. Foliated and crenulated quartz-muscovite garnetiferous schist, crossed nicols, scale bar 0.2 mm, borehole Hop-1, core from 1842–1848 m
2. Helicitic albite porphyroclast in a schistose matrix of quartz and mica, crossed nicols, scale bar 0.2 mm, borehole Ho-1, core from 2145–2151 m
   Horgoš Series. Pre-Cambrian (?) or Cambrian (?)

Plate IV

1. Amphibolite, crossed nicols, scale bar 0.1 mm, borehole Mk-3, core from 2064–2066 m.
   Mokrin Complex. Cambrian (?)
2. Albite-quartz-biotite garnetiferous schist. Compositional growth zoning in tourmaline, crossed nicols, scale bar 0.2 mm, borehole Fi-1, core from 2033–2040 m.
   Mokrin Series Cambrian (?)

Plate V

1. Tectonic fine breccia of granitoid, crossed nicols, scale bar 0.2 mm, borehole KV-66, core from 2001–2004 m.
2. Flaser gneiss. Quartz forms finegrained recrystallized aggregates in matrix around large lenticular porphyroclast of plagioclase, crossed nicols, scale bar 0.2 mm, borehole KV-11, core from 2153–2157 m.
   Mokrin Series. Cambrian (?)

Plate VI

1. Cataclastic augen gneiss with large porphyroclast of microcline, crossed nicols, scale bar 0.2 mm, borehole Mi-1, core from 1894–1896 m.
   Mokrin Complex. Cambrian (?)
2. Cataclastic augen gneiss with large porphyroblast of feldspar, crossed nicols, scale bar 0.2 mm, borehole Ada-5, core from 508–909 m
   Mokrin Series. Cambrian (?)

Plate VII

1. Granite, cataclastic, quartz show undulose extinction, crossed nicols, scale bar 0.2 mm, borehole Kl-2 core from 1051–1052
2. Cataclastic Granite with large biotite flakes and perthitic feldspar, crossed nicols, scale bar 0.2 mm, borehole Fk-1 core from 907–910 m

Plate VIII

1. Paleorhyolite with large quartz phenocrystals is corroded and shows undulose extinction, crossed nicols, scale bar 0.2 mm, borehole KZ-2, core from 2545–2547 m.
2. Ouartz-sericite-chlorite schists with snowball garnet crystals which contains wisps of graphite. Plane-polarized light, scale bar 0.2 mm, borehole VS-3, core from 3332–3337 m

Plate IX

1, 2. Conglomerate-breccia "mud pebbles" in. anhydrite, partly carbonate cement, exhibiting some ptygmatic lamination, Core speciment, borehole Ba-3, core from 741–743.4 m.
Lower Triassic

Plate X

1. *Meandrospira pusilla* (Ho) in micrite, *Meandrospira pusilla* and *Meandrospira cheni* (Ho) in corner, borehole NK-3, core from 1005–1013 (x 96), borehole Gr-1, core from 990.8–996.8 (x 96)
   Lower Triassic
2. *Physoporella pauciforata* Bystricky in intrabiomicrosparite, borehole Gb-1, core from 553.8–994.6 m (x18x)
   Anisian

Plate XI

1. *Turitellella mesotriassica* Koehn-Zaninetti in marlly, microcrystalline limestone (biomicrite), borehole NK-1, core from 1032.6–1041 m (x 57)
   Anisian (Illyrian)?–Lower Ladinian
2. Organogenic, detritic and recrystallized limestonee with dasycladacean remnants, *Teutloporella herculea* Soppani and mollusc detitus, borehole Ve-2, core from 810.3–813.3 m
   Ladinian

Plate XII

1. *Pilamminella cuthani* (Salaj) in itrabiomicrosparite, borehole Ba-3, core from 1400–1405 m
   Upper Triassic (Carnian)
2. *Aciculella* sp. in fractured limestone, borehole Ba-3, core from 1400.6–1405 m
   Upper Triassic (Carnian)

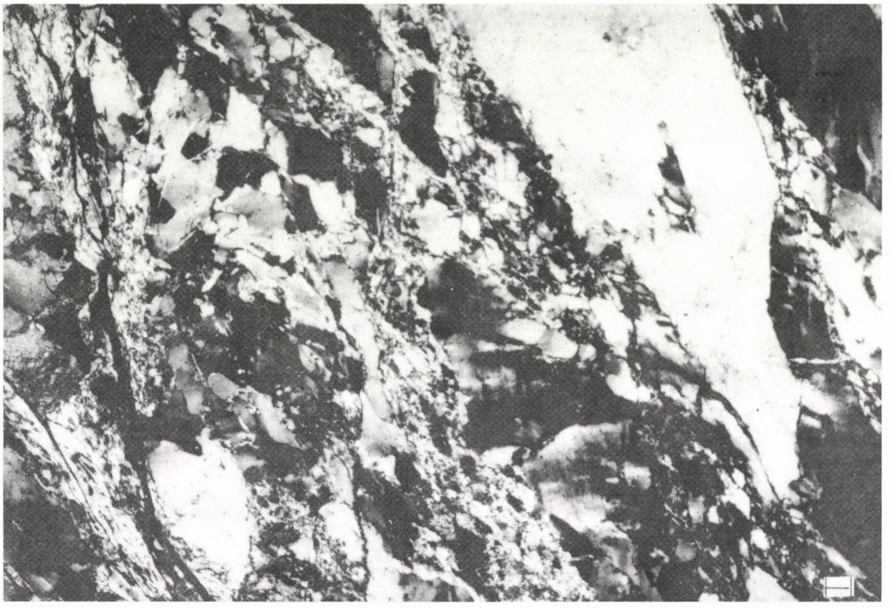
Plate XIII

1. Organogenic, detritic limestone with lithoclasts and bioclasts (solenoporacea, coral, gastropod and other biodetritus), borehole Ba-3, core from 1306.5–1310 m.
   Upper Triassic (Upper Carnian–Norian)
2. *Angulodiscus communis* Kristann micrite, borehole Ba-3, core from 1306.5–1310.4 m
   Upper Triassic (Upper Carnian–Norian)

Plate XIV

1. *Aulotortus sinuosus* Weynschenk (x 38). Upper Triassic limestones (intrabiosparrudites), borehole Ba-3, core from 1306.5–1310 m.
   Upper Triassic (Upper Carnian–Norian)
2. Organogenic, detritic limestones with *Agathammina austroalpina* Kristan-Tollmann–Tollmann, *Sigmoilina* sp. and other detritus, borehole Ba-3, core from 1188.6–1194.6 m.
   Upper Triassic (Norian)

Plate I



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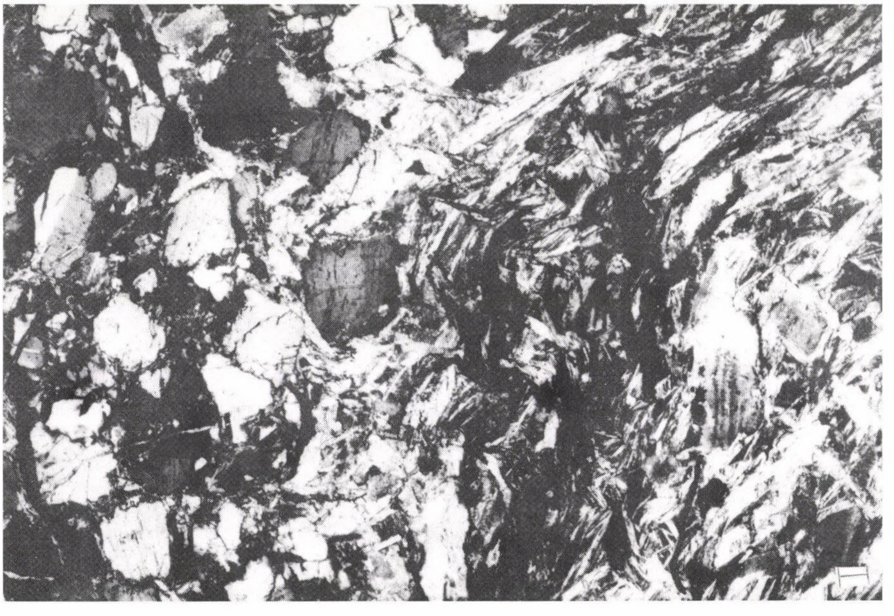

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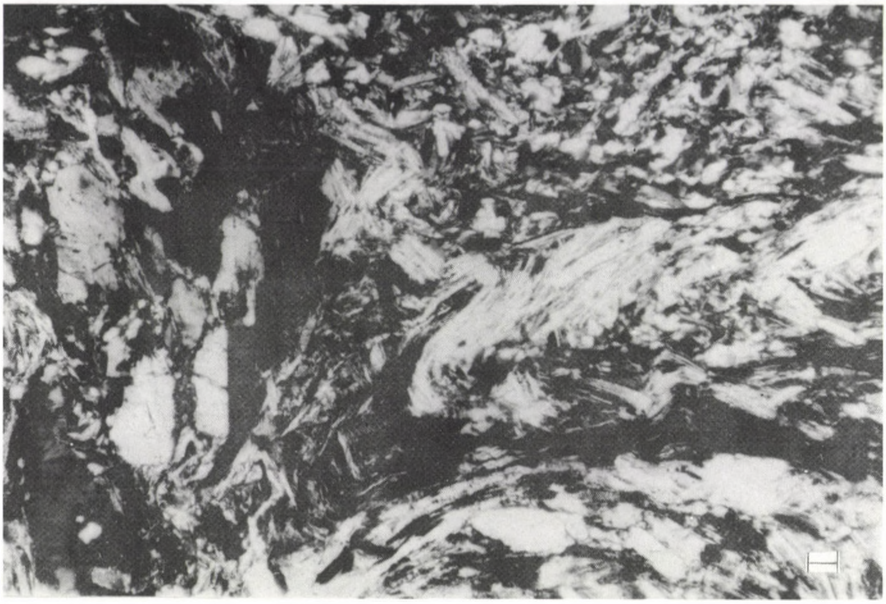

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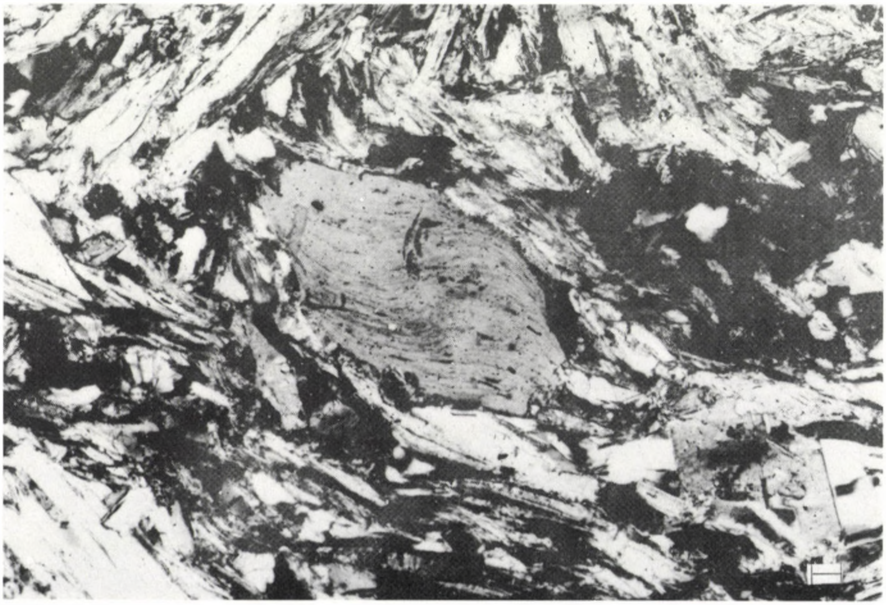

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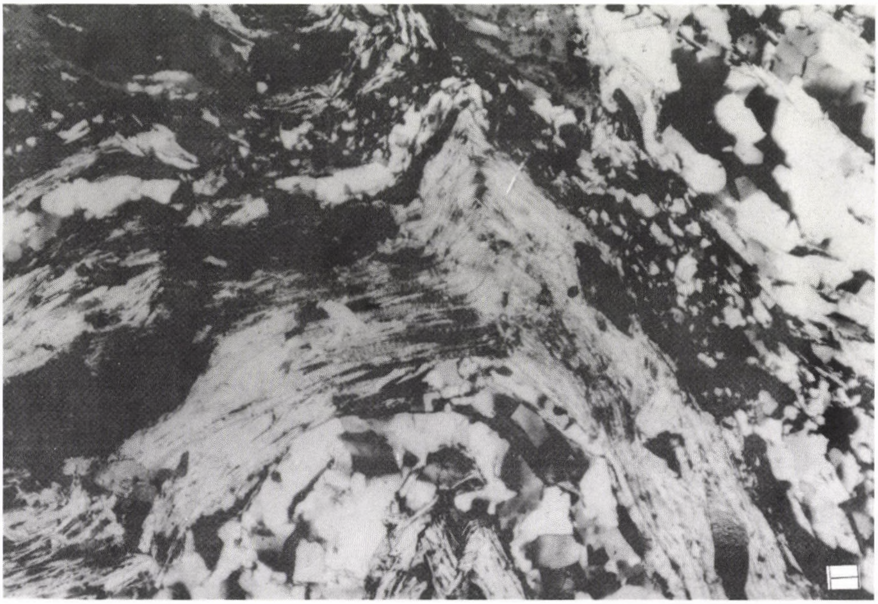

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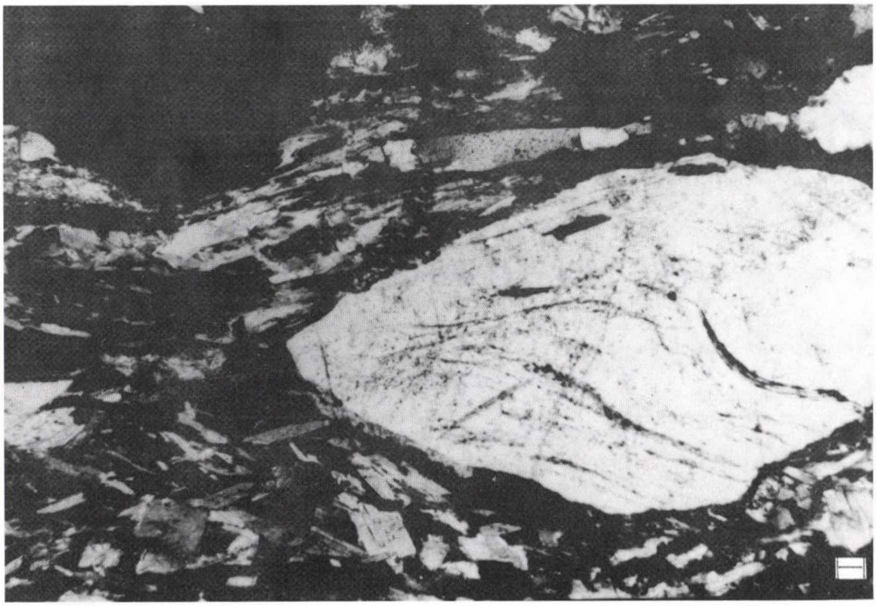

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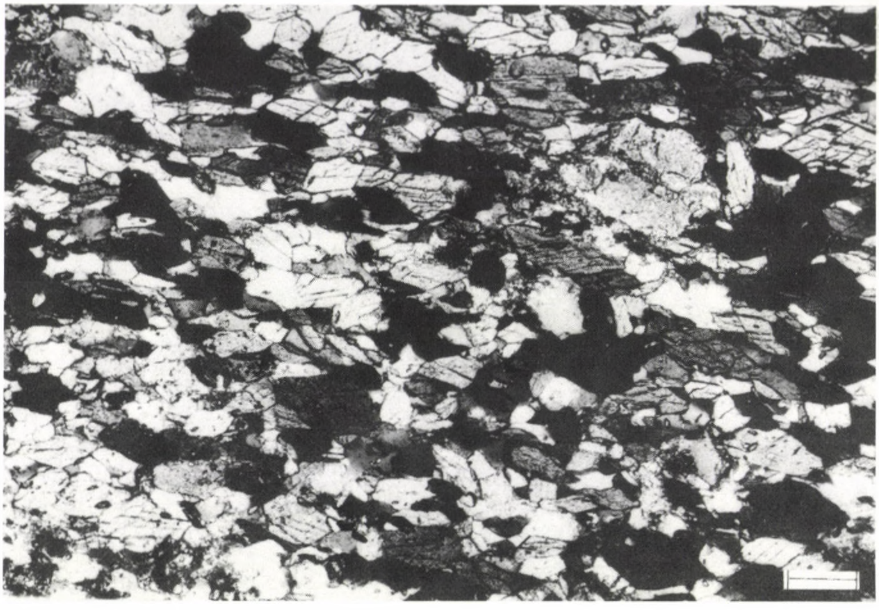

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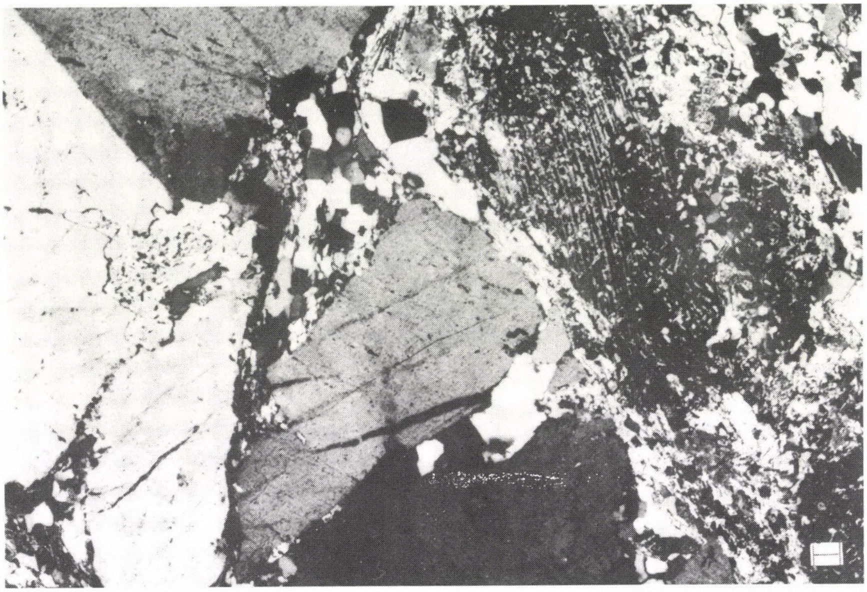

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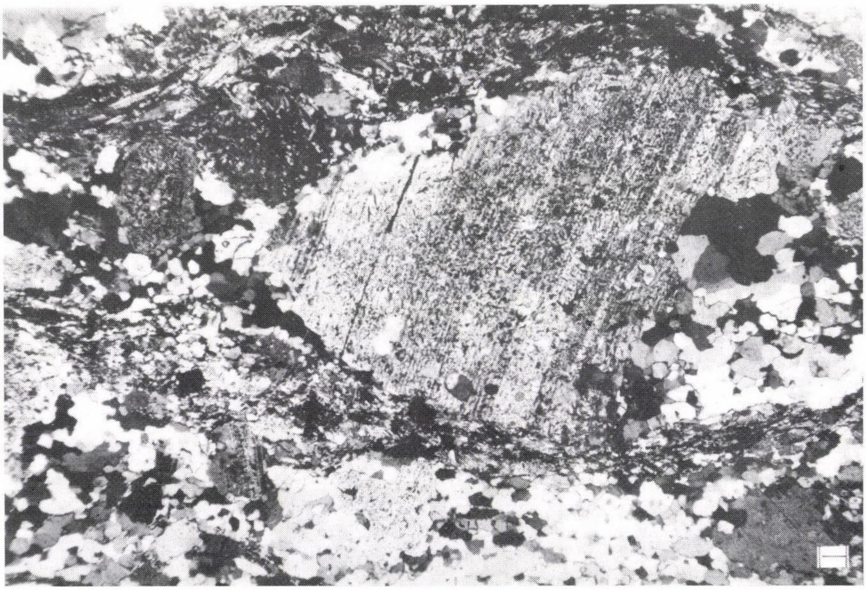

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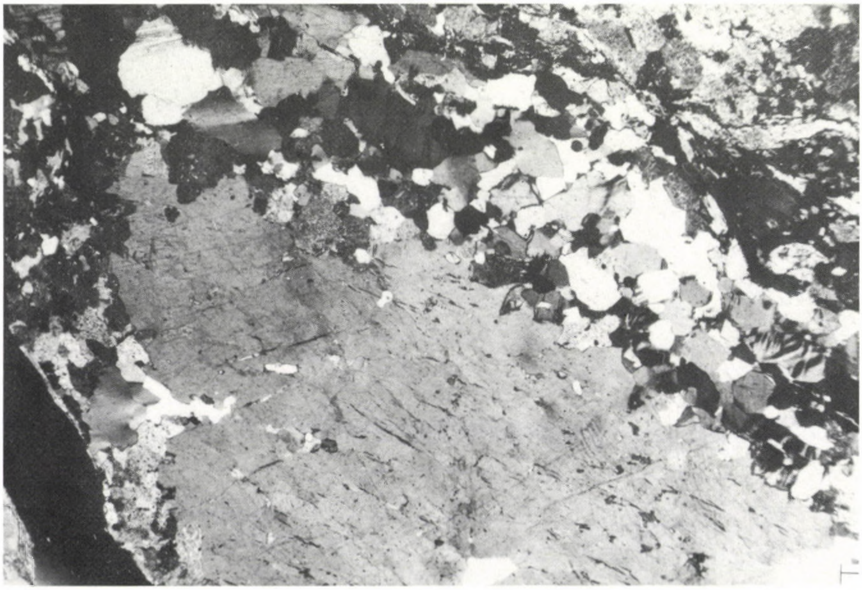

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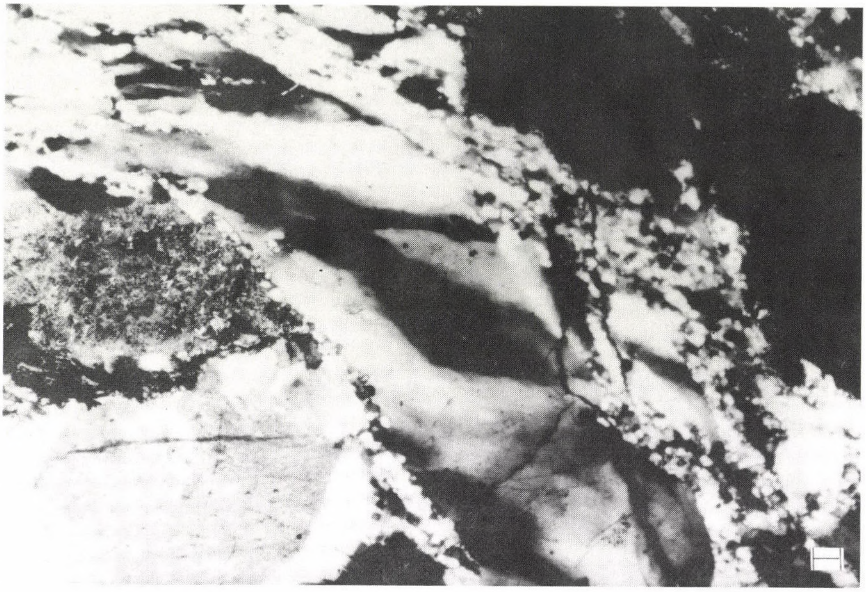

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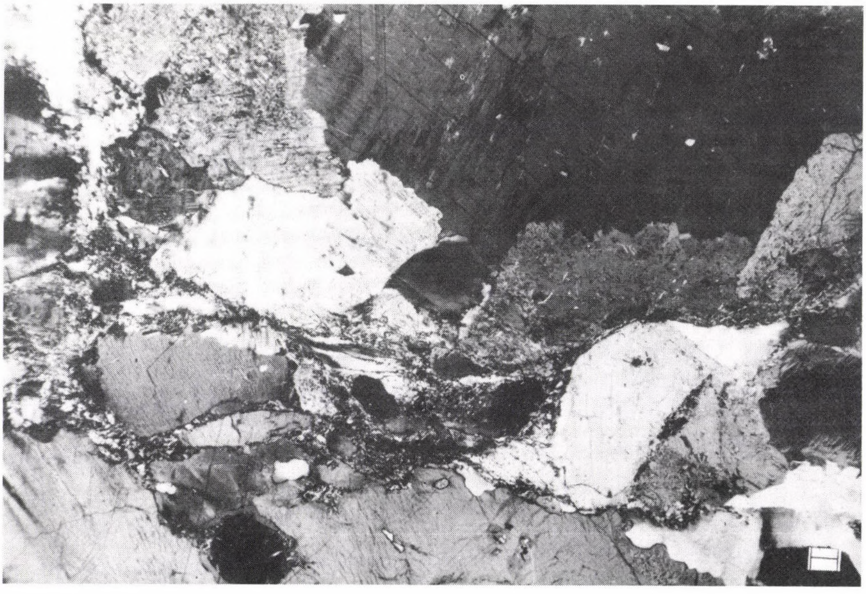

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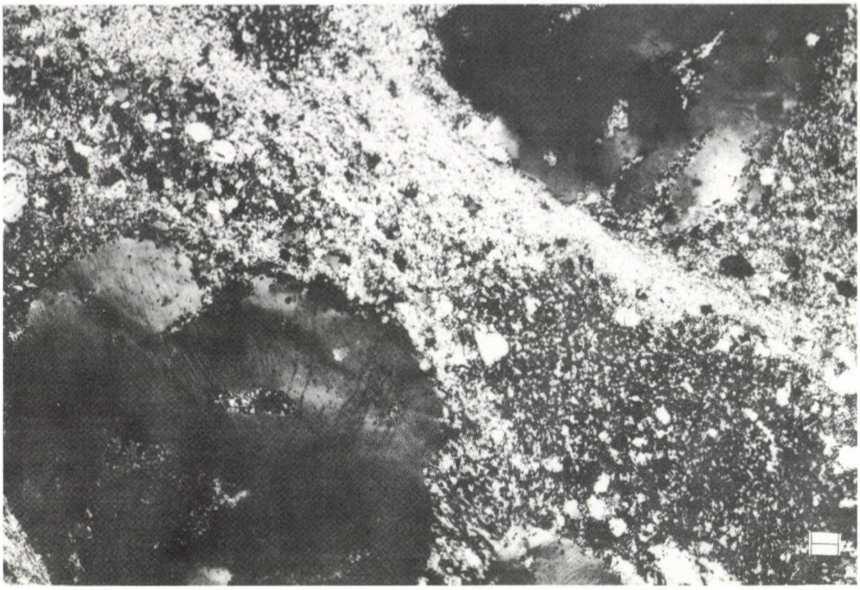

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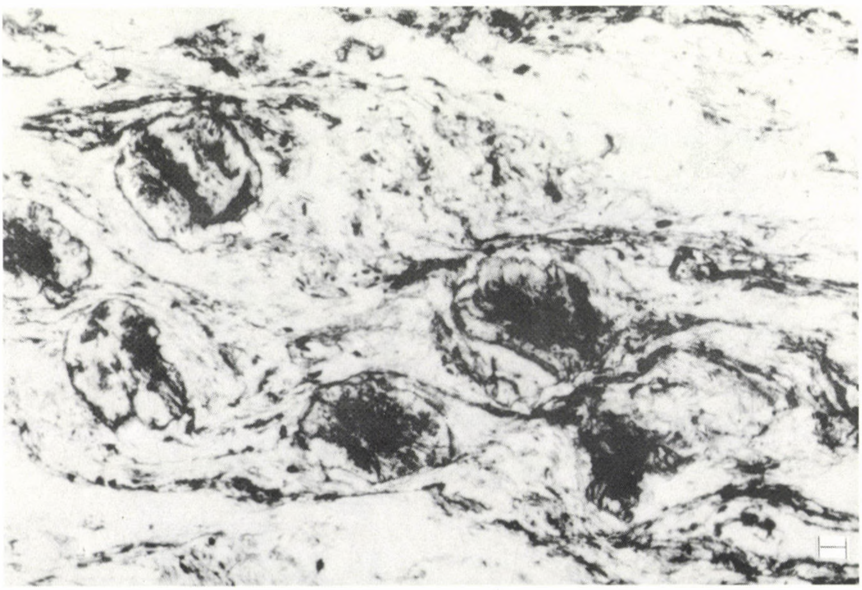

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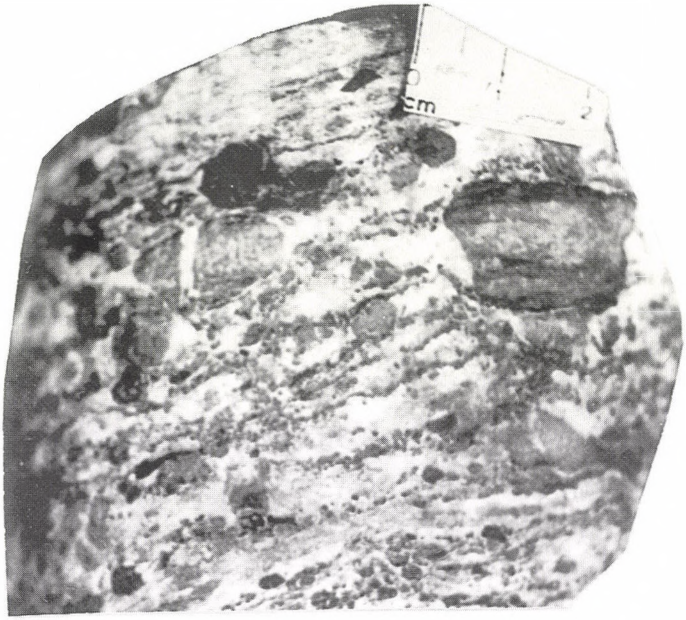

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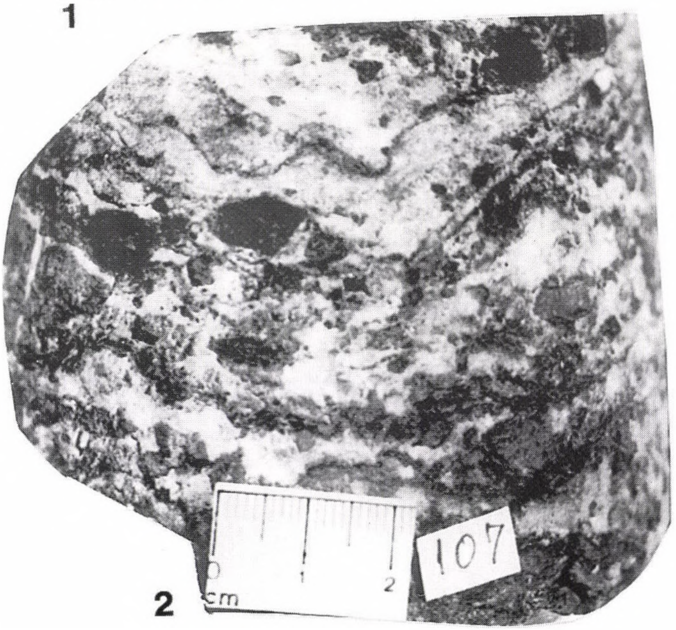

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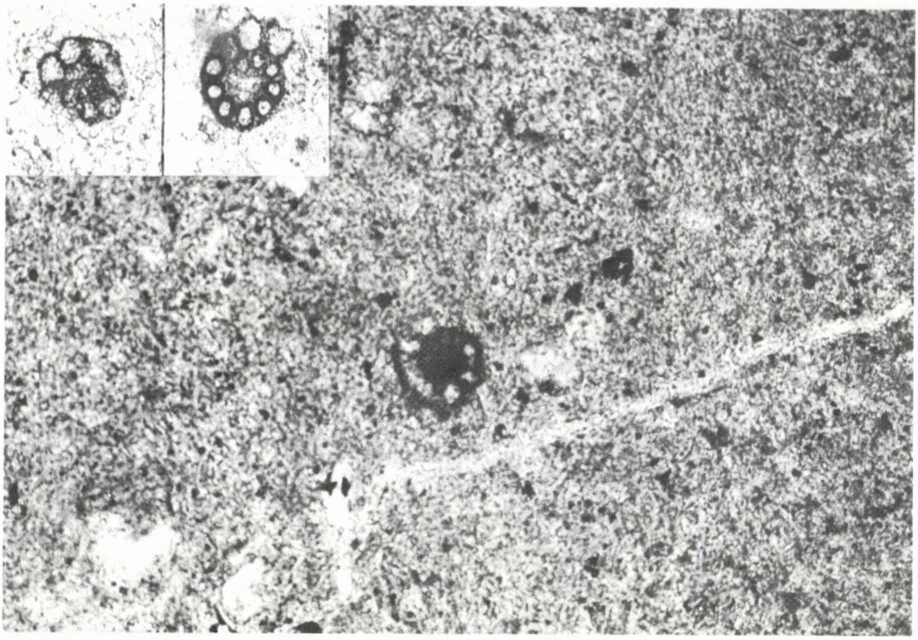
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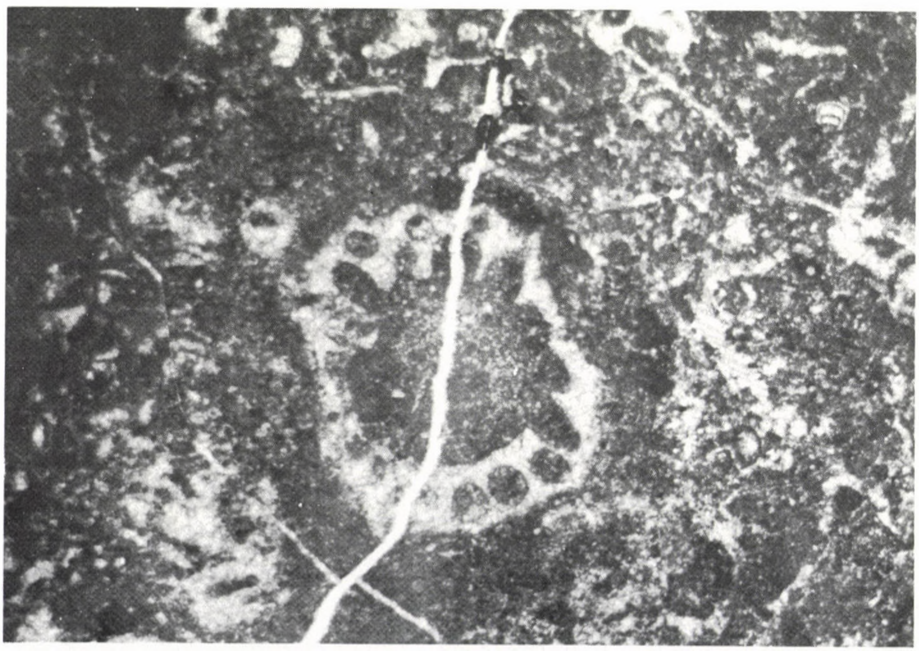
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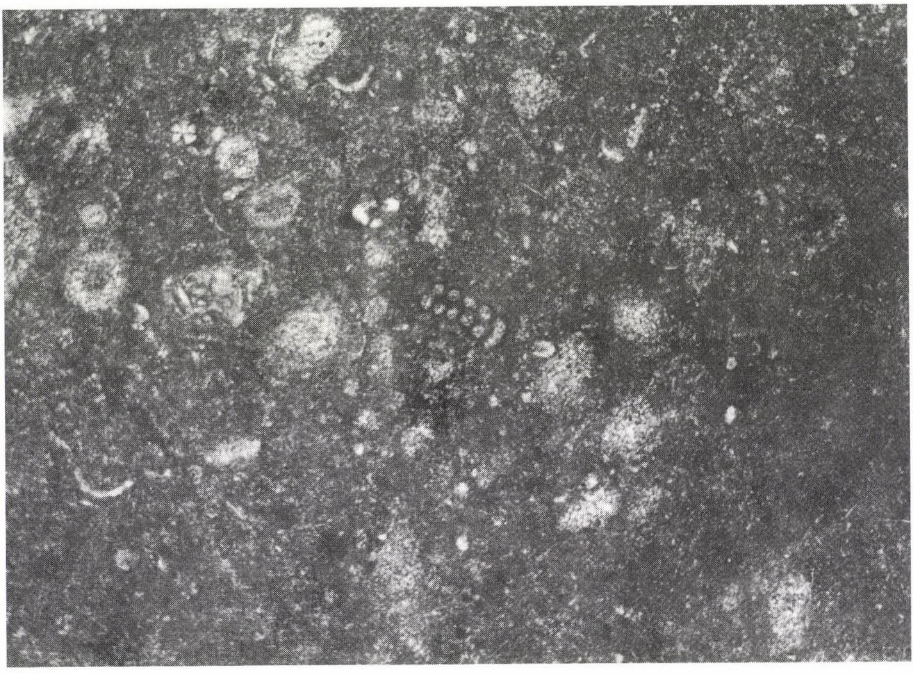

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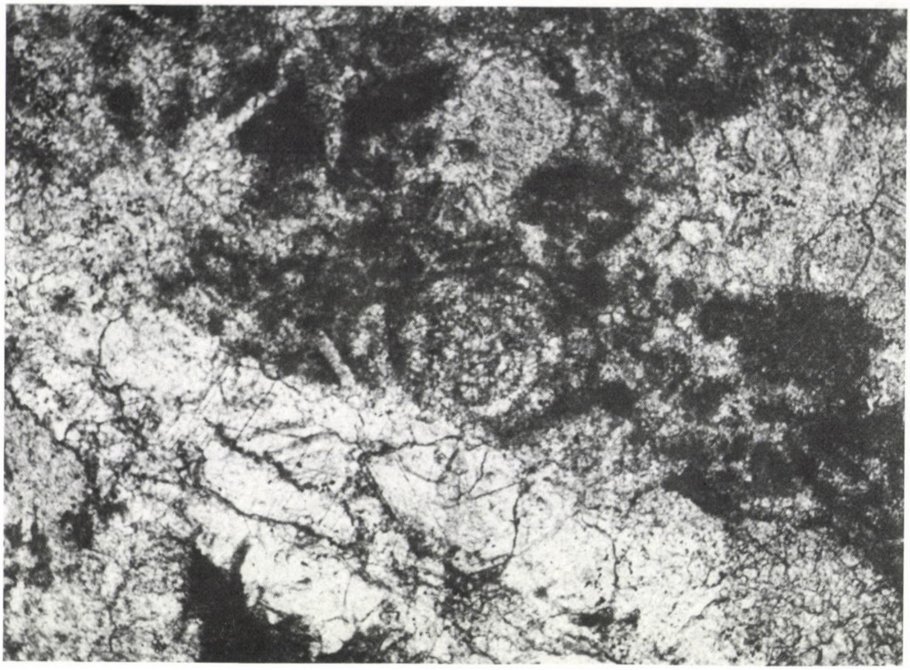

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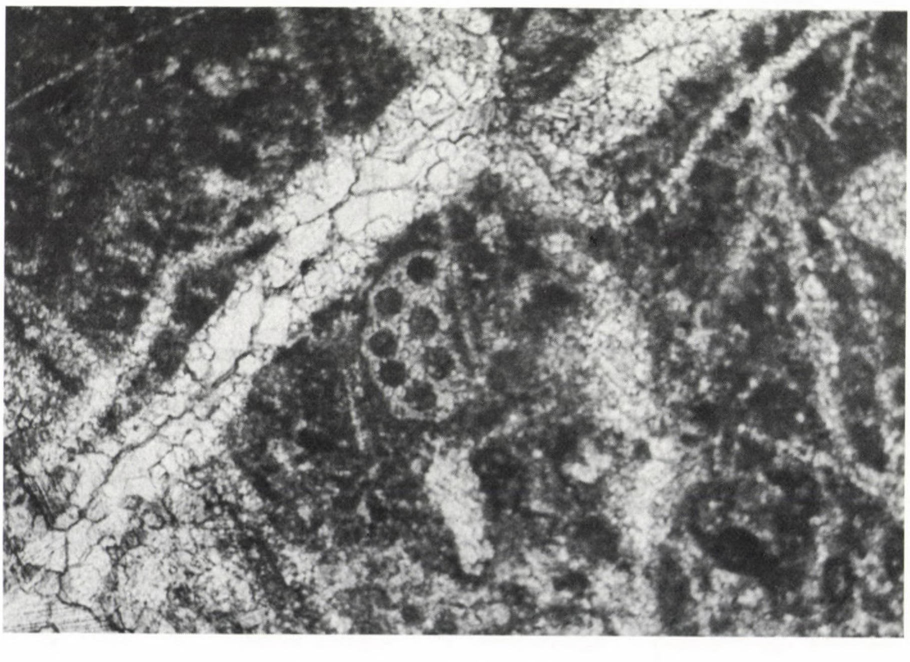

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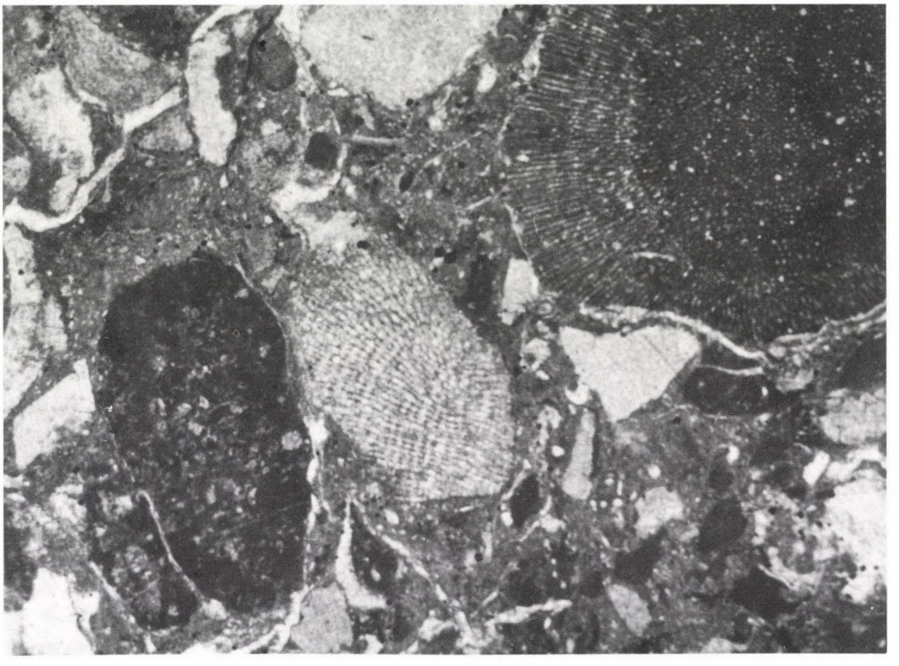

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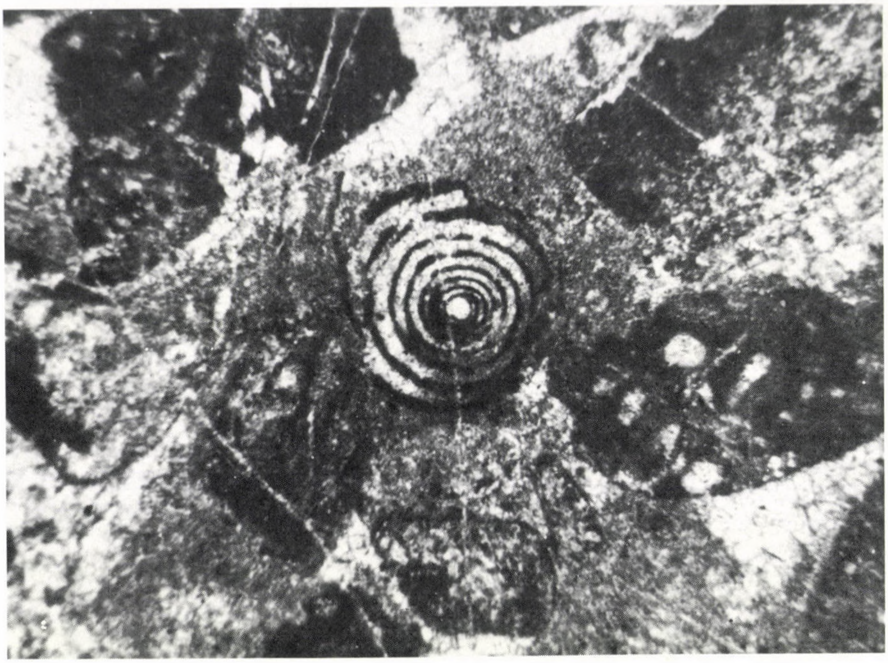

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The northwesternmost outcrops of the Dinaridic ophiolites: a case study of Mt. Kalnik (North Croatia)

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This paper presents geological and petrologic data on allochthonous bodies of Mesozoic ophiolites from Mt. Kalnik, which are correlated with coeval ophiolites from the surrounding mountains of the south-western parts of the Pannonian Basin commonly included in the northwesternmost Dinarides. Basalt is associated mostly with Middle-Upper Triassic to Lower Cretaceous interlayered chert, siliceous shale, mudstone and limestone. Mostly Jurassic olistostrome melange, subsequently strongly tectonized, includes fragments of predominant native graywacke with basalt, diabase, gabbro, amphibolite, serpentinized peridotite, shale, chert, and exotic, stratigraphically varied limestones, which are embedded in a pervasively sheared shaly-silty matrix. Mappable zones of tectonic carbonate megabreccias with fragments of Middle Triassic to Paleocene limestones and dolomites are associated with the tectonized melange. The tectonized melange was probably generated by Eocene final subduction processes. Ophiolites with indigenous sedimentary rocks are overthrust by the large Save nappe stretching into neighbouring Slovenia and further into the Southern Alps.

Rock descriptions, major and trace element data for all ophiolites from Mt. Kalnik are given. These and the surrounding ophiolites from the south-western Pannonian Basin can be correlated in many ways with the ophiolitic complexes of the internal Dinarides and Bükk Mts in Hungary. The correlation and geochemical data obtained suggest that ophiolites from the northernmost Dinarides represent dismembered ophiolites, which probably originated in MORB environments and were subsequently incorporated in the tectonized olistostrome melange.

The present geological position of the allochthonous ophiolites is genetically related to postsubduction extension and transcurrent tectonic movements and subsequent normal faulting which caused the formation of the southern parts of the Pannonian Basin.

Key words: NW Dinarides, basalt, diabase, gabbro, amphibolite, ultramafics, radiolarites, melange, subduction, emplacement, thrusting, geodynamics, correlation, Bükk Mts

Introduction

Mesozoic ophiolitic complexes of the Alpine–Himalayan belt characteristically occur within the internal units of separate mountain ranges (Coleman 1977). However, the Mesozoic ophiolites are not found only in the internal Dinarides but also in the south-western parts of the Pannonian Basin, which represents a typical extensional realm. Here, in the area between the Save and Drava rivers, Mts. Strahinšcica, Ivanščica, Ravna Gora, Medvednica, and Kalnik occur, which are made up mainly of Mesozoic sedimentary rocks (Fig. 1). The ophiolites of this area are of regional importance geologically

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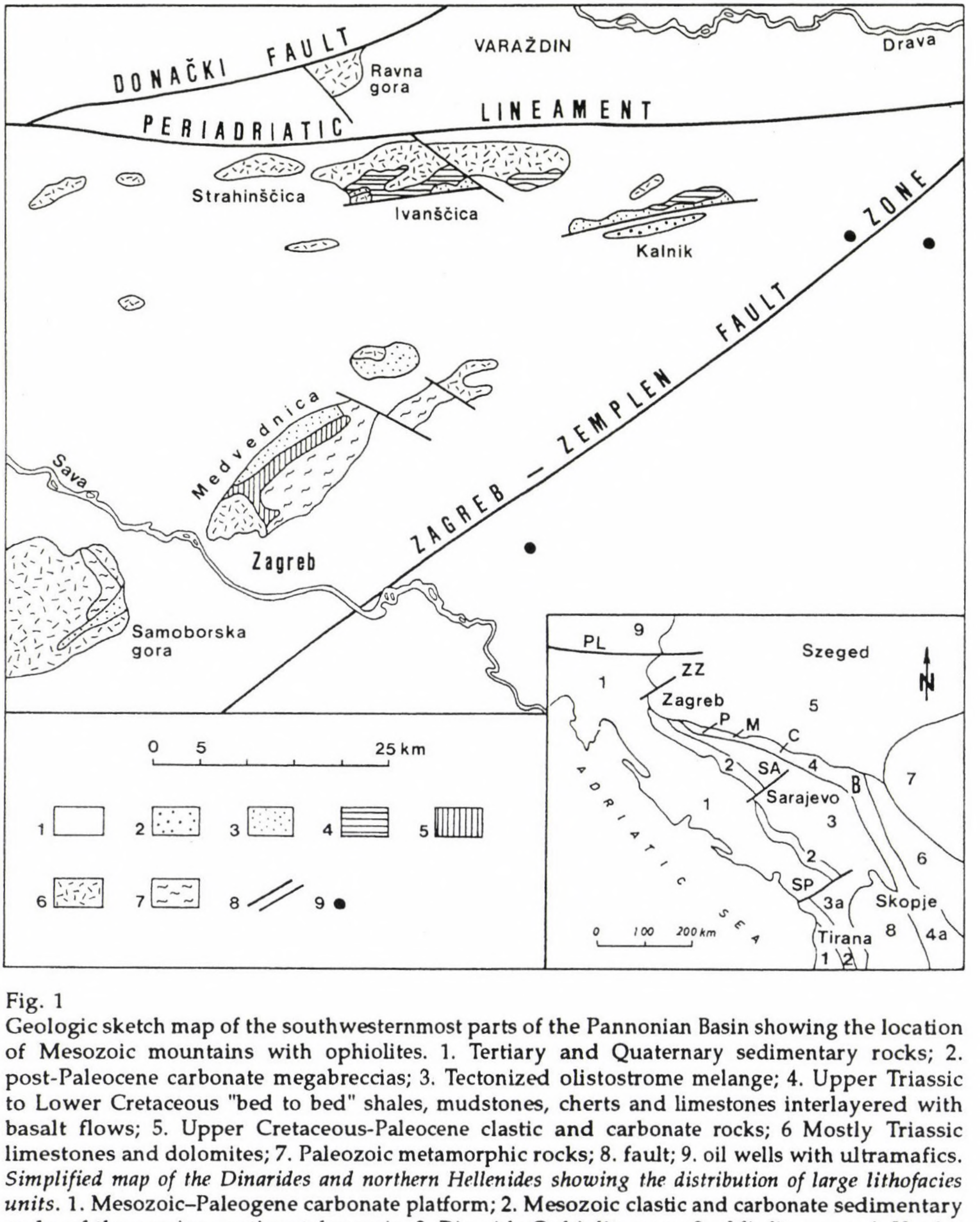

Fig. 1

Geologic sketch map of the southwesternmost parts of the Pannonian Basin showing the location of Mesozoic mountains with ophiolites. 1. Tertiary and Quaternary sedimentary rocks; 2. post-Paleocene carbonate megabreccias; 3. Tectonized olistostrome melange; 4. Upper Triassic to Lower Cretaceous "bed to bed" shales, mudstones, cherts and limestones interlayered with basalt flows; 5. Upper Cretaceous-Paleocene clastic and carbonate rocks; 6 Mostly Triassic limestones and dolomites; 7. Paleozoic metamorphic rocks; 8. fault; 9. oil wells with ultramafics. *Simplified map of the Dinarides and northern Hellenides showing the distribution of large lithofacies units.* 1. Mesozoic–Paleogene carbonate platform; 2. Mesozoic clastic and carbonate sedimentary rocks of the passive continental margin; 3. Dinaride Ophiolite zone; 3a. Mirdita zone; 4. Vardar zone sensu lato; 4a. Vardar zone sensu stricto; 5. Pannonian Basin; 6. Serbo-Macedonian Massif; 7. Carpathians; 8. Pelagonides and Korab; 9. Eastern Alps, *Mountains:* P – Prosara; M – Motajica; C – Cer; B – Bukulja; *Large transverse faults:* ZZ – Zagreb-Zemplen; A – Sarajevo; SP – Skadar-Pec because of its "triple junction" position. These ophiolite outcrops represent in fact the first eastern neighbours of the Apennic ophiolites, the first south-eastern neighbours of the Penninic ophiolites of the Eastern Alps and the first south-western neighbours of the Mid-Transdanubia and the Bükk Mountain ophiolites, i.e. the Meliata.

In some of the previously-mentioned mountains, Mesozoic sedimentary rocks are associated with ophiolites which have so far been petrologically studied in detail only in Mt. Medvednica (Crnković 1963) and geologically in Mts Kalnik, Medvednica, Samoborska Gora, and Ivanščica (Herak 1960; Šikić et al. 1979; Šimunić and Šimunić 1979; Babić et al. 1979 and others). In this area, Alpine ultramafic rocks were penetrated by some oil wells in the basement of the Pannonian Basin (Pamić 1986; Šimunić and Pamić 1989). Not all the ophiolites have been studied within the frame of modern geodynamic ideas as a geologic-petrologic whole.

The aim of this paper is to present the geologic and petrologic characteristics of allochthonous bodies of ophiolites and associated sedimentary rocks from Mt. Kalnik, which will be compared with ophiolites from the surrounding mountains in the south-western parts of the Pannonian Basin. Basalts, which are the most common basic ophiolitic rocks, are mainly interlayered with Middle–Upper Triassic to Lower Cretaceous alternating bedding of shale, mudstone, chert, and limestone. But the basalts are also included as interlayered bodies, blocks and fragments in tectonized olistostrome melange, together with ultramafic rocks, amphibolite, gabbro and diabase, probably representing dismembered parts of an ophiolite sequence. All these mafic and ultramafic rocks can be correlated in many aspects with the ophiolitic complexes of the internal Dinarides and Bükk Mts in north-eastern Hungary. The ophiolitic complex of the investigated area and the Dinaride Ophiolite Zone are characteristically overthrust by the Paleozoic–Triassic Save nappe which continues in the Southern Alps, hinting at a possible subsurface connection (?) with the Ligurian ophiolites.

Basic geological data from Mt. Kalnik

The Mesozoic rocks of Mt. Kalnik and the surrounding Mounts Ivanšcica, Medvednica, and Samoborska Gora have been studied in smaller and separated areas by numerous authors and all these published data are summarised elsewhere (Šikić et al. 1979; Šimunić 1992).

Mt. Kalnik, located in the south-western part of the Pannonian Basin, occurs north-west of the large Zagreb–Zemplen fault zone, within the Mid-Transdanubian unit located between the Tisia and Pelso geotectonic megaunits (Kovács et al. 1989). According to some other opinions, this and surrounding mountains are included in the Supradinaricum (Herak 1986).

Mt. Kalnik can be followed along strike for about 35 km approximately in an east-west direction. This area consists mostly of Neogene sedimentary rocks of the Pannonian Basin, subordinate Middle and Upper Triassic limestones and dolomites and various Mesozoic sedimentary rocks with ophiolites. The Mesozoic rocks can be traced along strike for about 15 km in the northern parts of Mt. Kalnik (Fig. 2) – Šimunić (1992).

The Triassic formations of Mt. Kalnik and the surrounding mountains could be considered as the easternmost tectonically broken and eroded parts of the Save–Julian Savinja nappes (Mioč 1982) which cover the central parts of adjacent Slovenia and extend to Italy toward the northernmost parts of the Southern Alps (Carulli et al. 1990). The nappes, which are bounded by the Periadriatic Lineament in the north and the northern margin of the Mesozoic Dinaridic carbonate platform (the External Dinarides) in the south consist mostly of Triassic carbonate, clastic and volcanic rocks. However, in Slovenia, in tectonic windows and along fault zones within the Save nappe, Jurassic and Cretaceous basinal sedimentary rocks of Dinaridic affinity are preserved (Pamić 1993). In Mt. Ivanščica, Jurassic to Lower Cretaceous basinal sedimentary sequences with ophiolites also occur as tectonic windows below the Triassic nappe (Mioč 1995). However, the Middle Triassic volcanic rocks of the south-western Pannonian Basin, represented mostly by andesite-basalts and genetically related to the carbonate platform, are not included in this paper.

Sedimentary rocks associated with the ophiolites

Mesozoic mafic and ultramafic rocks of Mt. Kalnik occur in two different lithological units which originated in two different environments: (1) basinal sedimentary rocks and (2) olistostrome melange which is mostly tectonized (Fig. 2).

Basinal sedimentary rocks with basalts

This is a volcanic-sedimentary formation characterised by common interlayering of basinal sedimentary rocks and basaltic flows which are best exposed in several active quarries in the Hruškovec Valley in the western part of Mt. Kalnik (Fig. 2).

Sedimentary rocks are represented by alternating bedding of shale, marly shale, chert, mudstone and limestone, both micrite and calcarenite grading into calcareous sandstone. The most characteristic rocks are reddish and greyish radiolarian siliceous shales consisting of crypto- to microcrystalline quartz with subordinate calcite, kaolinite, illite, and limonite. Within this fine-grained mass recrystallized radiolarians are plentiful (Šimunić and Šimunić 1979). By variating modal and mineral compositions, the shales frequently grade into mudstones and cherts, both of them also mostly reddish and greyish in colour.

In radiolarites from Mts Kalnik and Medvednica, most recently Halamic and Goričan (1995) found Middle-Upper Triassic and Jurassic radiolarians. Limestones interlayered in the same basinal sedimentary sequence with

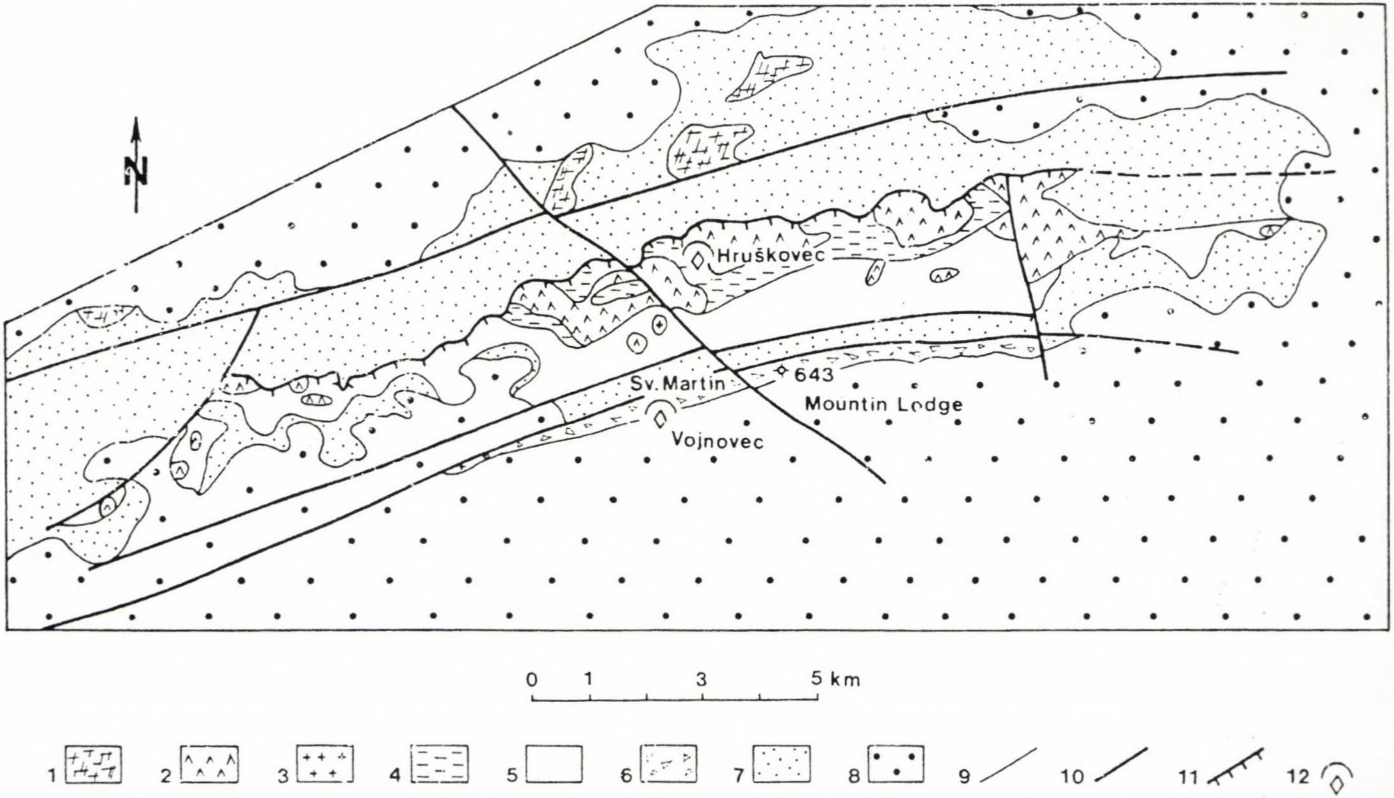

Fig. 2
Geologic map of the central parts of Mt. Kalnik according to Šimunic (1992).
1. Triassic dolomite and limestone; 2. basalt; 3. gabbro and diabase; 4. Upper Triassic to Lower Cretaceous "bed to bed" siliceous shale, mudstone, chert and limestone; 5. Tectonized olistostrome melange; 6. post-Paleocene tectonic melange; 7. Egerian-Eggenburgian sedimentary rocks; 8. Badenian–Pontian sedimentary rocks; 9. contact line; 10. fault; 11. reverse fault; 12. quarry

synsedimentarily basalts from adjacent Mt. Ivanščica contain Albian–Cenomanian microfossils (Šimunić and Šimunić 1979; Babić et al. 1979). Regionally, this sedimentary sequence can be correlated with the radiolarite formation of the central parts of the Dinaride Ophiolite zone which has a large stratigraphic span (Upper Triassic to Lower Cretaceous – Pamić 1982).

Basaltic flows interlayered with these basinal sedimentary rocks indicate the synsedimentary character of the submarine volcanic activity. The flows are commonly a few tens of metres thick, but some of them are more than 100 m thick. Thicker basaltic flows are interlayered with the same basinal sedimentary rocks and tuffs and the interlayers are commonly 1 to 20 m thick.

Basalt occurs as massive, pillowed and brecciated lavas either filled by variable quantities of fine-grained tuff or well cemented by secondary calcite, quartz, prehnite, epidote, chlorite, and zeolites.

Tectonized olistostrome melange

This is a chaotic lithological unit characterised by a pervasively sheared, generally fine-grained (shaly-silty) matrix containing fragments of predominant native graywacke with other subordinate sedimentary and igneous rocks. Relict interlayering of the shale and graywacke was observed only in a few places. The interlayering of shale and graywacke with basalt is more commonly preserved; the larger basaltic flows are in turn interlayered with the graywacke and shale. Graywackes are most commonly included in the shaly matrix in form of flow-balls, mainly 5–20 cm in diameter, and larger fragments and slumps, a few decimetres to a few metres in diameter.

The olistostrome melange also includes fragments of various rocks: basalt, tuff, both of them at least partly indigenous, peridotite, diabase, gabbro, chert, shale, and exotic blocks of carbonate rocks of different ages originating in different environments. These are Middle and Upper Triassic dolomite and algal stromatolithic limestone, Liassic oncolitic limestone, Malmian siliceous limestone with radiolarians, Tithonian–Valanginian Calpionella limestone and Lower Cretaceous calcarenite (Šimunić and Šimunić 1979).

These exotic blocks must have been tectonically included in the melange whereas the blocks of native graywackes and partly the basalts, originally interbedded with undeformed shales, result from the olistostrome mechanism. Since the native blocks distinctly predominate over the exotic blocks, this chaotic complex must have originally represented an olistostrome which was subsequently reworked and tectonized.

No characteristic fossils have been found to date in graywackes and shales of the Mt. Kalnik olistostrome melange. However, in alternating shales and graywackes from the same tectonized olistostrome melange of neighbouring Mt. Medvednica, comparatively well-preserved angiosperm-type leaves (Jerinić 1994, pers. commun.) are found which indicate that parts of these sediments at least were deposited in the Barremian. The same tectonized olistostrome melange of the Dinaride Ophiolite zone is of a presumed Jurassic age (Dimitrijević and Dimitrijević 1973).

Carbonate megabreccias. In some areas composed of the tectonized olistostrome melange, stratigraphically exotic carbonate megabreccias build up individualised mappable zones. Such breccias occur in the form of an approximately 10 km long and about 100–500 m wide zone along the main ridge of Mt. Kalnik (Fig. 2). This zone is bordered, from both north and south, by narrow but mappable zones of strongly tectonized melange. Smaller zones of carbonate megabreccias are also found in the area of occurrence of the tectonized olistostrome melange on the northern slopes of Mt. Medvednica (Šimunić et al. 1993). Some oil wells in the nearby Drava Depression penetrated the carbonate breccias of the basement of the Pannonian Basin (Šimunić and Pamić 1989) which may be correlated, at least partly, with the Mt. Kalnik carbonate breccias.

The Mt. Kalnik carbonate megabreccias are composed of fragments and blocks of limestone and dolomite, varying in size from a few centimetres to several metres and tens of metres, without cement or cemented with a little calcite. The fragments and blocks are represented mostly by Triassic algal and stromatolithic limestone and dolomite, stromatolithic limestone breccia, Jurassic limestone, and Upper Cretaceous reef limestone with rudists and pelagic Globotruncana limestone. Some of the limestone fragments also contain Paleocene index microfossils (Šimunić and Šimunić 1979).

Since the youngest limestone blocks contain Paleocene index microfossils, these tectonic carbonate megabreccias must be of Upper Paleocene age or younger.

Age of the ophiolites

Most basalts are interlayered with siliceous shale, chert, mudstone, and limestone and therefore they should be also of Middle–Upper Triassic to Lower Cretaceous age.

K–Ar measurements, carried out on whole-rock samples of fresh diabase and ophitic hornblende-augite gabbro, gave fairly concordant ages of 189 ± 6.7 and 185.0 ± 6 Ma (I am grateful to E. McKee from the US Geological Survey in Menlo Park, California who did the radiometric determination). These Early Jurassic K–Ar ages are the same (within analytical errors) and they probably represent the crystallisation (generation) age of the diabase and ophitic gabbro.

The dated diabase and ophitic gabbro form a large fragment (50x60 m) which is tectonically included into the melange. The fragment is composed mostly of coarse-grained diabase or ophitic gabbro intruded by dykes of finer-grained diabases with chilled margins, which means that it may represent the dismembered part of a primary sheeted complex. Smaller diabase blocks are also found in the tectonized olistostrome melange.

Penecontemporaneous to them are the Darnó Hill ophiolites from the northern part of the Pannonian Basin in Hungary. K–Ar ages of the Darnó ophiolites range from 175 to 152 Ma for gabbros and from 110 to 100 Ma for basalts (Árva-Sós and Józsa 1992). Amphibolites interlayered with peridotites of the Dinaride Ophiolite zone yielded K–Ar ages of 170–160 Ma (Lanphere et al. 1975). These Jurassic to Early Cretaceous isotopic ages, though slightly decreased, can be thus correlated (within analytical errors) with K–Ar ages and geological ages obtained from Mt. Kalnik diabase and gabbro. Details on the problem of generation and emplacement ages of the Dinaride ophiolites are presented elsewhere (Pamić 1982).

Correlation with the surrounding Mounts Medvednica, Ivanščica, and Samoborska Gora

Ophiolites also occur in the surrounding Mts Medvednica, Ivanščica, and Samoborska Gora (Fig. 1). Table 1 shows the correlation between the ophiolites and associated sedimentary rocks from Mt. Kalnik and the surrounding mountains.

1) Ophiolites, as defined by Steinmann (1926), occur only in Mts Kalnik and Medvednica, whereas in Mts Ivanščica and Samoborska Gora only basalts have been found to date.

2) In Mts Kalnik, Medvednica and Ivanšica submarine basalt flows are mostly synsedimentarily interstratified in Middle–Upper Triassic to Lower Cretaceous alternate bedding of siliceous shale, chert, mudstone, and limestone.

3) However, in all four of these mountains, the tectonized olistostrome melange is characteristically developed. It generally has the same features as the olistostrome melange of the internal Dinarides (Dimitrijević and Dimitrijević 1973) which is, however, accompanied by much larger masses of ophiolites, particularly ultramafic rocks. Predominantly in Mt. Medvednica smaller and larger basalt bodies are also interlayered with shales and graywackes which are the most common ingredients of the olistostrome melange. As far as ophiolites are concerned, the olistostrome melange includes fragments of basalts and diabases in all four mountains, plus gabbros and ultramafic rocks in the Mts Medvednica and Kalnik.

4) Carbonate megabreccias, which represent a specific "melange facies", are best developed in Mt. Kalnik and in the form of smaller but mappable masses in Mt. Medvednica, whereas they have not been registered in the Mts Ivanščica and Samoborska Gora.

5) As in the Dinaride Ophiolite Zone and particularly in its eastern parts, the melange with ophiolites is accompanied by smaller and larger masses of Triassic platform limestone and dolomite. Their mutual relationship is usually ambiguous but in Mts Samoborska Gora, Ivanščica, and Medvednica Triassic rocks are thrust over the ophiolite melange. Further to the west in neighbouring Slovenia, Liassic to Neocomian basinal sedimentary rocks of the Slovenian

Table 1
Correlative data for ophiolites and associated sedimentary rocks of the mountains from the southwesternmost parts of the Pannonian Basin

Kalnik	Medvednica	Ivanščica	Samoborska Gora
Predominant metabasalts and basalts with tuffs, subordinate diabases and gabbros, rare ultramafic rocks and amphibolites	Predominant metabasalts and basalts, subordinate diabases and gabbros, rare ultramafic rocks	Only metabasalts and basalts	Only metabasalts
Basaltoid rocks are for the most part interlayered in Upper Triassic to Middle Cretaceous alternating siliceous shales, mudstones, cherts and limestones	Cretaceous basinal alternating series is not recorded, but its lithological members are included in the melange	Smaller basalt flows are interlayered in Cretaceous alternating limestones, siliceous shales, mudstones, and cherts which are conformably underlain by Early Cretaceous-Cenomanian turbiditic calcarenites and sandstones	Cretaceous basinal alternating series is not recorded
Tectonically reworked Upper Cretaceous olistostrome melange. The pervasively sheared shaly-silty matrix includes centimetre to tens of metre fragments and blocks of predominant native graywackes with subordinate basalts, diabases, gabbros, shales, cherts and exotic Triassic to Lower Cretaceous limestones originated in different environments	Tectonically reworked Upper Cretaceous olistostrome melange with the same pervasively sheared shaly-silty matrix. Lithology of fragments and blocks included in the matrix is nearly the same as in Mt. Kalnik. However, larger masses of basalts are interlayered with shales and graywackes. In the melange fragments of ultramafic rocks are included	Much smaller outcrops of tectonically reworked Upper Cretaceous olistostrome melange with the same pervasively sheared shaly-silty matrix. Fragments of native graywackes predominate over diabases, basalts and cherts. Exotic limestone fragments are of Upper Triassic, Lower Jurassic and Albian-Cenomanian age	Much smaller outcrops of tectonically reworked olistostrome melange with the same pervasively sheared shaly-silty matrix. Fragments of native graywackes predominate over diabases, basalts and cherts. The fragments are not yet studied in detail
Post-Paleocene tectonic, mostly carbonate melange without matrix or with a poor calcite cement. Tens of metre to hundreds of metre blocks of limestones and dolomites of Triassic to Paleocene age (carbonate megabreccia) predominate over fragments of ultramafic rocks, basalts, cherts and diabases. The melange forms a continuous 10 km long zone along the main ridge of Mt. Kalnik	Post-Senonian carbonate tectonic melange in the form of hundreds of metre to kilometre isolated masses which are included in tectonically reworked olistostrome melange. Predominant limestone blocks, commonly a few metres in size, are Triassic and Jurassic in age whereas poorly preserved carbonate cement contains Senonian (mostly Campanian) microflora	No carbonate tectonic melange yet recorded	No carbonate tectonic melange yet recorded
Triassic platform carbonates are not in direct contact with ophiolites and their sedimentary associates	Smaller masses of Triassic platform carbonates are thrust over tectonically reworked Upper Cretaceous olistostrome melange with ophiolites	Larger masses of Triassic platform carbonates are thrust over tectonically reworked Upper Cretaceous olistostrome melange	Larger masses of Triassic platform carbonates are thrust over Upper Cretaceous olistostrome melange

This table was prepared on the basis of the author's observations and evaluated data published by Herak (1956 and 1960); Golub and Vragovic (1960); Crnković (1963); Golub and Šiftar (1965); Babić and Gušić (1978); Babić et al. (1979); Šimunić and Šimunić (1979); Šimunić et al. (1976); Brajdic and Bukovec (1989); and others.

trough (Cousin 1972), which are lithologically identical to the ones of Mts Kalnik and Ivanšćica, are overthrust by the Save Nappe. The basinal sediments of the Slovenian trough are in some places associated with ophiolites (Mioč 1995). Unfortunately, there are no published detailed data on these ophiolites.

Petrology

Ophiolites are represented by ultramafic rocks, gabbro, diabase, and basalt associated with amphibolite.

Rock description

Ultramafic rocks occur as fragments in the tectonized olistostrome melange of the main ridge of Mt. Kalnik, in its south-western margin surrounded by carbonate megabreccias. The presence of the ultramafic rocks can be also inferred on the basis of the zone of serpentinite debris stretching for several kilometres along the steep southern slope of Mt. Kalnik from Kamišnica Creek in the east to the area of Sveti Martin in the west (Fig. 2). Only two small and poorly preserved outcrops of ultramafic rocks occur at the foot of the Kalnik Mountain Lodge and in the area of Sveti Martin. Small fragments of altered serpentinites can rarely be found along the northern margin of the melange.

All sampled ultramafic rocks are fairly uniform in structure, texture, and in mineral and chemical composition. They are strongly or more commonly completely serpentinized amphibole harzburgite grading into harzburgite. The texture is porphyroblastic and the structure is massive. Mesostasis, which is made up of fine-grained serpentinite mineral(s) and minute magnetite grains, must have been originally represented by olivine. Subordinate porphyrocrysts are represented by laminar enstatite (with an average of 10% ferrosilite as determined by optics) which is frequently aggregated in small lenses. Enstatite is fresh or transformed into bastite and talc and shows effects of tectonic deformation.

One subordinate porphyroblast is a colourless and optically positive amphibole (cummingtonite?). Porphyroblasts are rarely represented by chromium spinel, magnetitized to various degrees.

Serpentinites are commonly crosscut by a network of carbonate and silica minerals, and rarely by chrysotile asbestos.

Accordingly, based upon fabrics and compositional features, the ultramafic rocks from Mt. Kalnik are represented by serpentinized tectonic peridotites.

Amphibolites and amphibole schists. In the tectonic melange of the area of Sveti Martin, fragments of strongly to completely weathered amphibolites and amphibole schists occur rarely.

Amphibolites and amphibole schists are nematogranoblastic in texture, with a grain size of 0.5–2 mm, and massive or parallel in structure due to foliation. The mineral assemblage includes green hornblende and subordinate plagioclase (oligoclase to andesine). Accessory constituents are opaque minerals (apatite and zircon).

Diabases and ophitic gabbros occur commonly as decimetre to tens of metre fragments in the olistostrome melange in the central parts of the main ridge of Mt. Kalnik. The diabases and gabbros are characterised by a fine-grained (0.5–2 mm) to coarse-grained (1 to 4 or 5 mm) ophitic texture and massive structure.

Major minerals are plagioclase, clinopyroxene and amphibole. The plagioclase, commonly represented by labradorite (as optically determined), is slightly to strongly replaced by clinozoisite, prehnite, and in marginal parts of the grains by albite. In some ophitic gabbros grading into diorites, pegmatitic intergrowths of quartz are included in the albite rims. In such samples, ophitic texture grades into hypautomorphic texture.

Clinopyroxene (augite) is an interstitial xenomorphic mineral which is commonly chloritized and uralitized to various degrees. Some samples contain single grains of green hornblende and rarely light-brownish titanaugite. The most common accessory mineral is skeletal ilmenite accompanied by subordinate apatite, zircon, and rutile.

Basalts are the most common ophiolitic rocks. In the area of Mt. Kalnik the main masses of basalts are interlayered as submarine flows with Middle-Upper Triassic to Lower Cretaceous basinal sedimentary rocks, whereas much smaller quantities of basalts are included as fragments in the olistostrome melange. Only the synsedimentarily basalts were petrologically studied in detail.

Fresh basalts are extremely rare and more than 90% of the sampled volcanic rocks on the Mt. Kalnik are represented by metabasalts (spilites). Both basalts and metabasalts have the same structural and textural features. Most of the basalts are ophitic in texture with grain size of 0.1–0.5 mm, whereas coarser-grained varieties and very fine-grained ones are subordinate (Vrkljan 1989).

Pillow basalts and metabasalts are radiated ophitic in texture. The texture is commonly finer-grained (with needles about 0.5 mm long) in the pillows' outer shells and coarser-grained (1.5–2 mm) in the central parts where it frequently grades into a lath-ophitic texture.

Most of the basalts and metabasalts are massive in structure and contain amygdules, commonly 0.1 to 1 mm, rarely up to 3–4 mm in diameter, which are filled by calcite, chlorite, quartz, chalcedony, celadonite, and pumpellyite (?) with rare albite, prehnite, zeolite, and hematite. The same mineral assemblage plus epidote builds up veinlets, lenses and nests which unevenly cut across the metabasalts. The quantity both of amygdules and veinlets decreases from the peripheral to the central parts of pillows.

The major minerals of basalts and metabasalts are feldspar and clinopyroxene. The feldspar is represented by calcic plagioclase (An_{60-65}, normative) in fresh basalts and albite to oligoclase in metabasalts. About 25% of the metabasalts contain fresh or nearly fresh albite which, however, commonly contains moderate to great quantities of minute secondary inclusions. These are clinozoisite, chlorite, pumpellyite (?), calcite, and prehnite.

Clinopyroxene is represented by augite which is interstitially xenomorphic or feather-shaped. In some metabasalts, the augite is fresh but mostly slightly to strongly and rarely completely transformed into chlorite, epidote, and calcite. Fresh light-brownish titanaugite and fine-grained serpentine pseudomorphs after olivine (?) occur very rarely.

Geochemistry

Major and trace element data and CIPW norms for ophiolitic rocks of Mt. Kalnik are presented in Table 2.

Analyses 1 and 2 illustrate Alpine-type, strongly serpentinized amphibole harzburgite and harzburgite characterised by high $MgO:FeO^x$ ratio ranging between 4 and 5. The bulk chemistry of basic rocks is tholeiitic as indicated by the distribution of both major and trace elements. In the AFM diagram (Fig. 3a), the sheeted rocks and amphibolites principally plot close to the line dividing the tholeiitic from the calc-alkaline field. Ophiolitic sequences of the Dinarides show approximately the same trend (Karamata et al. 1980). In the V versus Cr diagram (Miyashiro and Shido 1975), the basic rocks including amphibolites display a mostly tholeiitic affinity (Fig. 3b). This diagram suggests that the original tholeiitic magmas had more or less similar high V contents (250–500 ppm) and that during fractional crystallisation the V content of the residual magma gently increased, or remained constant. Most of these geochemical features presented on the diagrams are brought about by alteration effects both on sheeted rocks and particularly on basalts which are mainly transformed into spilites.

The sheeted rocks and basalts have compositions that are mainly characteristic of oceanic tholeiitic basalts. In the FeO^x/MgO versus TiO_2 diagram (Miyashiro 1973; Ikeda and Yuasa 1989), most of the basic rocks including amphibolites plot in the MORB field (Fig. 4a). In the TiO_2 diagram versus Zr diagram (Pearce et al. 1984) and in the Al_2O_3/TiO_2 versus TiO_2 (Sun et al. 1979), most of the points also plot in the MORB fields (Fig. 4b and c). Some other trace element abundances as for example the Ti:V ratio ranging between 16 and 36, the Sr:Zr ratio ranging between 0.2 and 0.8 and the comparatively high Ni content averaging 91 are also indicative of MOR basalts (Shervais 1982; Ikeda and Yuasa 1989).

Discussion

Ophiolites associated with their sedimentary country rocks occur in Mts Kalnik, Ivanščica, Medvednica, and Samoborska Gora of the northwesternmost Dinarides which are included in the southwesternmost parts of the Pannonian Basin. These rocks are most completely developed in Mt. Kalnik. Basalts are

Table 2
Major and trace element contents of ophiolitic rocks from the Mt. Kalnik

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	35.42	37.81	42.47	44.60	46.53	47.16	53.71	53.47	45.10	47.90	49.62	53.05	42.72
TiO_2	0.11	0.10	1.39	1.35	1.81	1.51	1.57	1.86	1.17	1.39	1.47	1.00	1.75
Al_2O_3	3.57	3.19	14.92	13.01	16.53	16.06	15.50	17.76	16.71	16.37	18.02	14.13	19.23
Cr_2O_3	0.44	0.44	-	-	-	-	-	-	-	-	-	-	-
Fe_2O_3	6.12	5.63	7.04	9.36	4.37	7.71	4.73	3.61	4.69	5.53	5.83	4.18	6.30
FeO	2.10	1.97	6.88	5.40	8.37	4.59	7.78	7.17	5.62	5.44	5.63	4.20	4.04
NiO	0.31	0.31	-	-	-	-	-	-	-	-	-	-	-
MnO	0.09	0.09	0.22	0.23	0.20	0.18	0.19	0.20	0.12	0.16	0.18	0.12	0.12
MgO	29.82	33.95	8.76	8.56	6.43	8.05	4.26	4.37	8.70	7.78	4.47	7.18	3.99
CaO	4.77	2.52	7.87	5.89	7.93	7.20	3.09	1.18	10.88	5.81	6.70	6.12	8.05
NaO	0.05	0.07	2.53	2.88	3.63	3.99	4.87	5.69	2.63	4.86	5.20	4.53	2.89
K_2O	0.02	0.01	0.33	0.45	0.24	0.16	0.06	0.32	0.07	0.73	0.08	0.11	4.42
P_2O_5	0.02	0.02	0.07	0.07	0.11	0.10	0.14	0.16	0.07	0.06	0.11	0.55	0.27
H_2O	17.42	14.24	7.28	7.79	3.85	3.29	4.10	4.21	4.24	3.97	2.70	5.33	6.23
Σ	00.26	00.35	99.76	99.59	00.00	00.00	00.00	00.00	00.00	00.00	00.01	00.00	00.01

FeO/MgO			1.6	1.8	2	1.5	2.9	2.5	1.2	1.4	2.5	1.2	2.5

CIPW norms

qz	-	-	-	2.6	-	-	9.2	7.5	-	-	-	4.2	-
c	-	-	-	-	-	-	2.2	6.6	-	-	-	-	-
or	0.1	0.1	2.1	2.9	1.5	1.0	0.4	2.0	0.4	4.5	0.5	0.7	27.9
an	11.4	9.6	30.7	23.1	29.2	26.3	15.0	5.0	35.1	40.1	26.3	40.5	28.2
ab	0.5	0.7	23.1	26.5	31.9	34.9	43.0	50.3	23.2	21.6	45.2	18.9	4.7
ne	-	-	-	-	-	-	-	-	-	1.5	-	-	11.6
dien	6.1	1.7	3.4	2.9	2.8	3.6	-	-	6.4	2.6	2.1	4.1	4.5
difs	-	-	0.8	0.1	1.6	-	-	-	1.4	0.5	0.6	0.7	-
diwo	7.1	1.9	4.6	3.4	4.6	4.2	-	-	8.7	2.6	3.0	5.4	5.2
hyen	16.9	27.4	10.1	20.3	5.0	11.2	11.1	11.4	3.0	-	4.6	14.8	-
hyfs	-	-	2.3	0.4	2.9	-	8.5	7.8	0.7	-	1.4	2.3	-
fo	46.7	47.9	7.1	-	6.2	4.2	-	-	9.2	12.3	3.3	-	4.3
fa	-	-	1.8	-	3.9	-	-	-	2.2	2.4	1.1	-	-
mt	9.4	8.5	11.0	14.8	6.6	11.4	7.2	5.5	7.1	8.4	8.7	6.4	8.9
hm	1.5	1.9	-	-	-	0.1	-	-	-	-	-	-	0.6
il	0.3	0.2	2.9	2.8	3.6	3.0	3.1	3.7	2.3	2.7	2.9	2.0	3.5
ap	0.1	0.1	0.2	0.2	0.3	0.2	0.3	0.4	0.2	0.1	0.3	0.1	0.7
Norm Pl	-	-	An_{70}	An_{47}	An_{48}	An_{43}	An_{26}	An_9	An_{60}	An_{65}	An_{37}	An_{68}	An_{85}

Trace elements

Ba	22	22	74	61	34	35	77	-	3	-	18	59	87
Co	320	78	58	54	51	51	43	-	52	-	47	43	39
Cr	1345	1483	109	104	121	187	28	-	166	-	165	317	188
Cu	9	13	71	59	42	56	24	-	85	-	60	85	33
Nb	-	-	8	9	11	9	15	-	5	-	9	3	17
Ni	2758	2855	89	65	52	77	23	-	136	-	54	213	84
Sr	4	5	23	18	43	56	64	-	37	-	44	59	103
V	51	60	517	450	445	418	259	-	259	-	312	254	354
Zr	16	12	187	128	87	74	111	-	46	-	198	134	226

1 and 2 serpentinized peridotite; 3 and 4 hornblende and plagioclase-amphibole schist;
5 and 6 dolerite and ophitic gabbro; 7 and 8 gabbro grading into diorite; 9 fresh basalt;
10 to 14 varieties of metabasalt.
Analyses 5 to 13 recalculated to 100% by the subtraction of CaO and CO_2 from secondary calcite.

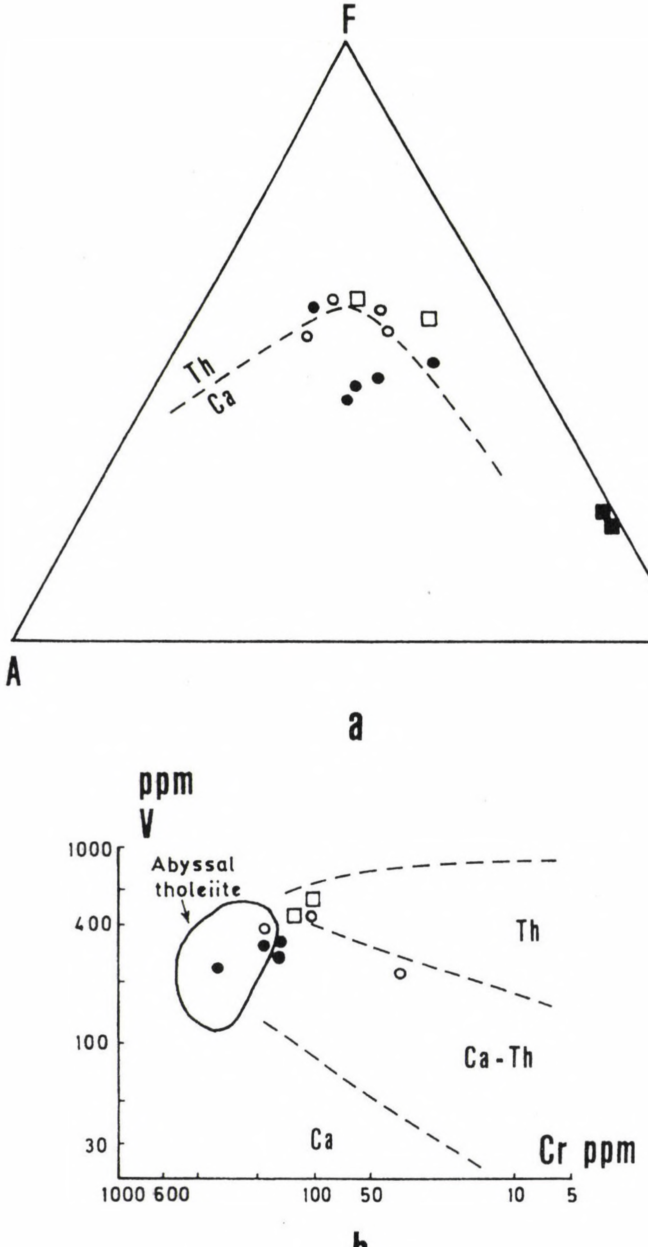

Fig. 3
a) AFM diagram and; b) V versus Cr diagram for the Mt. Kalnik ophiolites; full squares: ultramafic rocks; empty squares: amphibolites; open circles: diabases and gabbros; dots: basalts and metabasalts

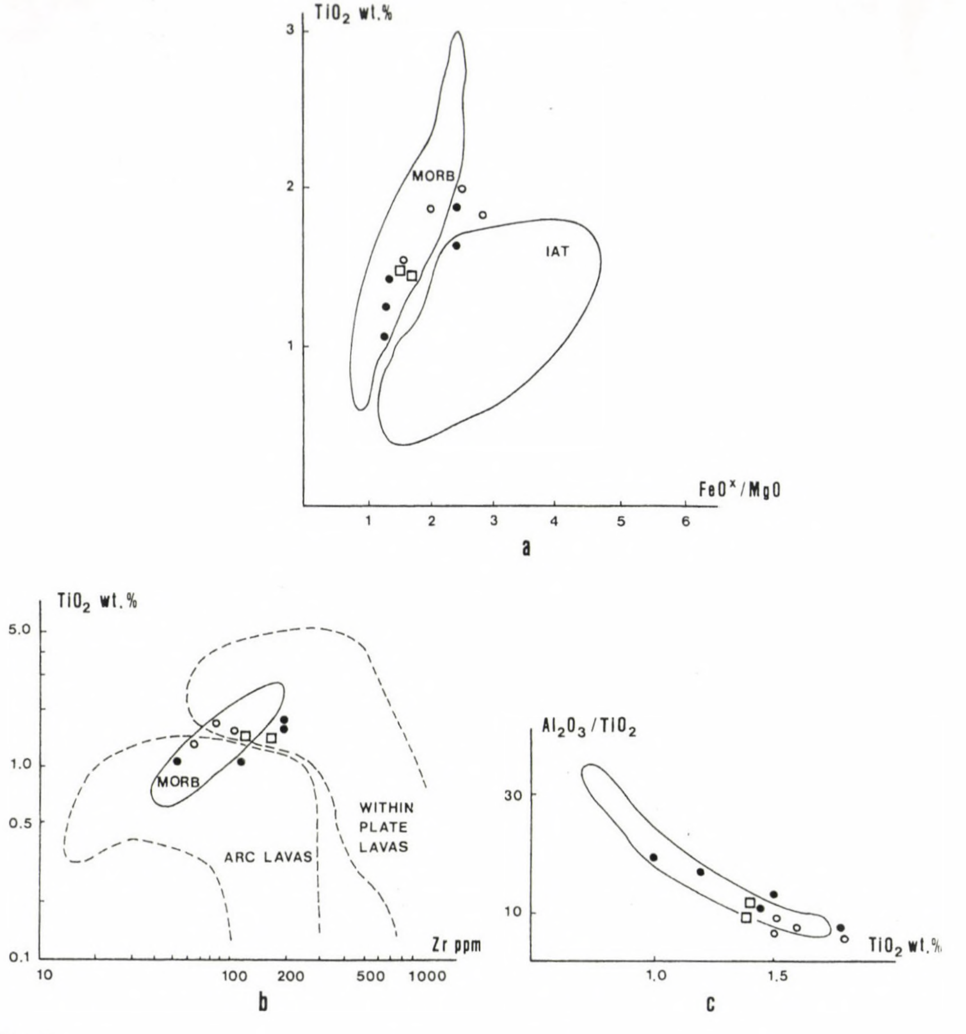

Fig. 4
a) TiO_2 versus FeO^x/MgO diagram for basalts, compiled by Ikeda and Yuasa (1989); b) TiO_2 versus Zr for basalts of different geotectonic settings (Pearce et al. 1984); and c) Plot of Al_2O_3/TiO_2 ratio against TiO_2 abundance for MORB, compiled by Sun et al. (1979)

synsedimentarily interlayered in Middle–Upper Triassic to Lower Cretaceous basinal sedimentary rocks deposited "bed to bed". Originally, this formation could represent a tectonically undisturbed geophysical layer 1 of the open oceanic realm of the Mesozoic Dinaridic Tethys. The basalts are also included, together with ultramafic rocks and amphibolites, diabases and ophitic gabbros of Early Jurassic radiometric age, as fragments and blocks in the tectonized olistostrome melange.

Ophiolites are characteristically overthrust by Triassic sedimentary rocks of the easternmost parts of the Save nappe which continues without a break into adjacent Slovenia where ophiolites are also found in the same tectonic position. The fact that the Save nappe continues into the Southern Alps underlines a problem of a possible subsurface continuation (?) into the Southern Alps of the Dinaridic ophiolitic complexes hidden below the Save nappe, i.e. via the Ligurian and Penninic ophiolites.

Ophiolites associated with the melange complexes characteristically occur within suture zones and close to convergent plate margins (Coleman 1977). They originated by sedimentary (olistostrome) processes and tectonic fragmentation and tectonic mixing genetically related to subduction and/or obduction processes (Gansser 1974). For that reason, melange facies with ophiolites buried and incorporated in ancient orogenic belts can be considered as markers of active continental margins and ancient subduction zones, respectively. Such a geodynamic interpretation has also been proposed for the Dinaride ophiolites (Pamić 1977 and 1982; and others).

Putting aside the differences in the size of outcrops and mutual relative lithological proportions, the ophiolites of Mts. Kalnik, Medvednica, Samoborska Gora and Ivanšcica can be correlated with ophiolites of the internal Dinarides and their country rocks. Dinaride ophiolites are mostly accompanied by an olistostrome melange (Dimitrijević and Dimitrijević 1973) which has the same shaly-silty matrix and predominating native graywacke and exotic clasts, including almost the same stratigraphic range of limestone fragments as the olistostrome melange of Mts Kalnik, Medvednica, Samoborska Gora and Ivanšćica. On the other hand, Middle-Upper Triassic to Lower Cretaceous basinal sedimentary rocks with synsedimentarily basalts of this area are correlative in lithology with the basinal succession of the Dinaride Ophiolite zone which consists of radiolarite, mudstone, shale, and limestone with synsedimentary basalt of Upper Triassic to Lower Cretaceous age (Pamić 1982). In this correlation it must be emphasised that stratigraphically exotic carbonate megabreccias of Mts Kalnik and Medvednica represent an unique formation which to date has not been recognised in the Dinaride Ophiolite zone. However, such a tectonic melange and Triassic to Cretaceous radiolarites are common in the internal units of some other mountain ranges of the Alpine-Himalayan belt as, for example, in the Albanian Hellenides (Shallo 1991) and the Zagros Range of Iran (Gansser 1974; Pamić et al. 1979, and others).

The Dinaridic ophiolites are genetically related to several geologic environments, and rather than being of a single stratigraphic age their evolution spans long portions of the Mesozoic. According to this geodynamic interpretation, the generation of ophiolites, as parts of the oceanic crust of Tethys, took place mainly during the Jurassic and Early Cretaceous, probably along the mid-ocean ridge as indicated by the present geochemical data. Obduction and emplacement of ophiolites began by initial subduction processes at the end of the Jurassic and the beginning of Early Cretaceous and lasted intermittently (?) until their Middle Paleogene termination. The emplacement of ophiolites might have been related to a presumed subduction zone, relics of which can be traced along the Prosara–Motajica–Cer–Bukulja zone of the northwesternmost Dinarides – see index-map in Fig. 1 (Pamić 1982 and 1993). The emplacement of the Save nappe must have been also related to these Lower Cretaceous tectonic events but postdated the first obduction of ophiolites as indicated by field relations and radiometric ages obtained on metamorphosed rocks of the Save nappe (Belak et al. 1995).

On the basis of the correlation presented above, it can be presumed that ophiolites from Mts Kalnik, Medvednica, Samoborska Gora and Ivanščica were originally incorporated in the northwesternmost parts of the present Dinaride Ophiolite zone, probably in the western prolongation of the Prosara–Motajica–Cer–Bukulja zone. Strong tectonization of the original olistostrome melange might have been genetically related to final stages of the subduction processes when stratigraphically exotic blocks of carbonate and other rocks had probably been incorporated in it. After the main compressional event and uplift of the Dinarides (Pyrenean phase), in their northern parts processes of extension and transcurrent faulting began which gave rise to the generation of the Pannonian Basin (Royden et al. 1983). One of the largest NE-trending transcurrent fault systems is the Zagreb–Zemplen fault zone within which Mts Medvednica, Kalnik, Samoborska Gora and Ivanščica are included. It is very probable that along such a NE-trending transcurrent fault, these ophiolites were detached from the northwesternmost parts of the Dinarides, displaced towards the north-east and incorporated in the present structure of Mts Kalnik, Medvednica, Samoborska Gora and Ivanščica.

Within this scenario the north-eastern prolongation of the Dinaridic ophiolites in the Mid-Transdanubian unit and further into the Bükk area remains enigmatic. It is obvious, however, that the ophiolites of the Bükk Mts. in northern Hungary can be correlated with ophiolites both from the Dinarides and the mountains of the southwesternmost parts of the Pannonian Basin. The Darnó Hill and Szarvaskő ophiolites represent Middle to Late Jurassic oceanic fragments consisting of basalt, diabase, gabbro and rare ultramafic rocks. Sedimentary country rocks are represented by pelagic "bed to bed" shale, siliceous shale, radiolarite, and limestone and Middle Jurassic (?) olistostrome melange with fragments of diabase, radiolarite and stratigraphically different limestone (Onuoha 1977; Balla 1984; and others). Geochemical data for these ophiolites indicate that the basic igneous rocks have features characteristic of ocean floor basalts (Balla et al. 1983). Recently Downes et al. (1990) suggested that the basic magmas of the complex were probably originally MORB-like, but during evolution in a shallow magma chamber, they became contaminated by terrigeneous sediments. This positive correlation represents an additional piece of evidence that the Mesozoic complexes of Mt. Bükk bear affinities to the internal Dinarides.

Summarising the criteria of paleotectonic settings, the geochemical data presented above suggest mostly MORB environments for the ophiolites of Mt. Kalnik as do also geochemical data for the Mt. Bükk mafic rocks (Downes et al. 1990). In contrast, available data for the Dinaridic ophiolites suggest marginal sea environments or combined E-type MORB and back-arc environments. However, both the Dinaridic and southwesternmost Pannonian Basin ophiolites differ from the West-Alpine and internal Ligurian ophiolites mainly in that that they lack Fe-Ti-rich gabbros (Pamić 1977; Pamić and Desmons 1989; Lugović et al. 1991).

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Assumed circulation pattern of the Central Paratethys through Badenian (Middle Miocene) times: Quantitative paleoecological analysis of foraminifera from borehole Tengelic-2 (SW Hungary)

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The quantitative analysis of foraminifera from bore hole Tengelic-2, SW Hungary was carried out with the purpose of paleoenvironmental reconstruction. Plankton foraminiferal assemblages were found to be dominated by mixed layer species; plankton/benthos ratios indicated an upper bathyal maximum depth in the Early Badenian. This presumably indicates a deep oligotrophic surface mixed layer involving a deep picnocline, if present. The idea of associating such conditions with an anti-estuarine (lagoonal) circulation is proposed. According to this paleoceanographic model there is a surface flow from the Mediterranean to the Paratethys, while deep water formed in the Paratethys flows in the opposite direction. Further speculations about Middle Badenian evaporites from the Outer Carpathians and the Transylvanian Basin and the resemblance between Central Paratethyan and Mediterranean fauna seem to support the idea of anti-estuarine circulation.

Key words: Badenian, Middle Miocene, Foraminifera, quantitative palaeoecology, Central Paratethys, Hungary, paleoceanographic model, anti-estuarine (lagoonal) circulation

Introduction

Numerous studies were written speculating on connections between the Paratethys and other seas (von Daniels and Ritzkowski 1970; Rögl and Müller 1978; Rögl and Steininger 1983; Steininger and Rögl 1984; Por and Dimentman 1985; Martini 1990). Authors dealing with Central Paratethys data generally agree (Rögl and Müller 1978; Rögl and Steininger 1983; Steininger and Rögl 1984) that during the Early Badenian seaways connected the Central Paratethys to the Mediterranean. The actual sediments of this sea way from the *Orbulina suturalis* Zone are found in present-day Istria (von Daniels and Ritzkowski 1970).

According to Rögl and Müller (1978), Rögl and Steininger (1983), Steininger and Rögl (1984) during the Middle Badenian the connection of the Central Paratethys to the East closed, and in Upper Badenian deposits Indo-Pacific elements were found confirming the break up of the Mediterranean seaway

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and the existence of troughs connecting the Central Paratethys to the East. The Mediterranean character of the Early Badenian fauna is emphasized, while Indo-Pacific relations are supposed for the Late Badenian (Rögl and Müller 1978; Rögl and Steininger 1983; Steininger and Rögl 1984). Others emphasize an Eastern Paratethys connection rather than an Indo-Pacific one during Late Badenian (Kókay 1985).

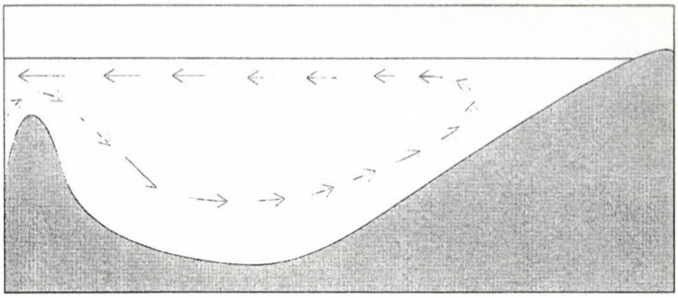
However, hardly any author has ventured to use this paleobiogeographic information to construct a paleoceanographic model except for Kókay (1985). His conclusion is an estuarine circulation pattern for the Late Badenian, which assumes, that hyposaline surface water flowed from the Eastern Paratethys to the Central Paratethys, while normal saline Central Paratethys water formed the deep water flowing back to the East. This model allows currents through the strait in both directions, thus it assumes a sufficient sill depth. His evidence is based on hyposaline and normal saline mollusc faunas at sites very close to each other, where tectonic features between the two sites can be excluded.

Such a paleoceanographic model has not been constructed for the Early Badenian. The Early Badenian foraminiferal data from borehole Tengelic-2 suggest a different surface water structure concerning picnocline and most likely different circulation from the one Kókay (1985) assumed for the Late Badenian and Sarmatian times.

Circulation in enclosed seas

In a basin two types of circulation can evolve depending on the balance between precipitation and inflow (net inflow) versus evaporation (Schopf 1980; Kennett 1982). These are the estuarine and the anti-estuarine circulation patterns. Anti-estuarine circulation is also called Mediterranean, or lagoonal. Certainly there is a strong tectonic control involved in determining the sill depth. Steininger and Rögl (1984) mention "deep-water troughs" connecting the Paratethys to the Mediterranean and Indo-Pacific in the Early Badenian, while Kókay's model (1985) implies the same for the Late Badenian. Thus, let us assume here a strait deep enough to allow the passage of a deep current below the surface current.

An estuarine circulation evolves, if the net inflow of fresh water exceeds evaporation, thus a less saline surface water layer is formed. This water on the surface flows from the continent toward the ocean, while normal saline deep water enters from the open ocean (Fig. 1a). The salt balance of the subbasin is negative compared to the ocean, so the subbasin becomes hyposaline as a consequence. If hyposaline water originating from runoff or precipitation piles up on top of normal saline water over the whole basin, then we might expect a strong, shallow picnocline, possibly hampering vertical circulation. If surface waters are eutrophic, then a dysoxic or anoxic bottom results. This is essentially the model suggested by Kókay (1985) except discussing productivity for the a)



b)

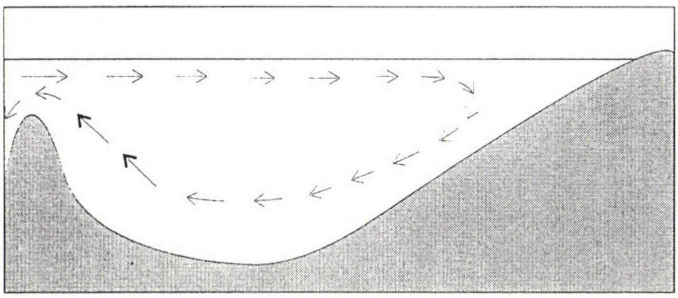


Fig. 1
The possible circulation patterns of an enclosed sea a: estuarine circulation; b: anti-estuarine circulation, after Schopf (1980)

Late Badenian times. The present-day example of the Sea of Marmora is taken as an analogue.

Anti-estuarine circulation develops if evaporation exceeds the net inflow. It involves a surface flow from the world ocean to the basin taking normal saline water, while deep water formed in the subbasin flows out (Fig. 1b). In the case of such circulation system the mass balance is positive with respect to salinity, as the deep water is formed from the normal saline surface water turning hypersaline due to evaporation. One expects its picno-thermocline to be situated deeper than for an estuarine circulation. The reduced density stratification contributes to a well-ventilated bottom. Only this kind of circulation have a positive salt balance making possible the accumulation of evaporites. The Mediterranean Sea (Miller 1983) and the Red Sea (Ross 1983), among other enclosed seas in the present-day oceans have anti-estuarine circulation.

Density stratification of the surface water

The seasonal thermo-picnocline is an important physical boundary, which is reflected in the vertical distribution of foraminifera (Douglas and Savin 1978; Fairbanks and Wiebe 1980; Fairbanks et al. 1982, Ravelo et al. 1990, Hodell and

Vayavananda 1993; Rohling 1994). Above the thermo-picnocline a homogeneous upper layer is found, which is well mixed and is called mixed layer. Here nutrient recycling is effective, thus it is oligotrophic. Just below the thermo-picnocline the nutricline is found. The nutricline can be identified by the DCM (Deep Chlorophyll Maximum Layer), which indicates the phytoplankton maximum. As phytoplankton need light, it cannot follow the picnocline deeper than the minimum light intensity necessary for growth (Fairbanks and Wiebe 1980). When there is a shallow picnocline, the DCM ascends into the photic zone, becoming food resource for the zoo plankton (Rohling 1994). Some planktonic foraminifera species exploit this food resource, thus a typical assemblage evolves indicating a eutrophic environment (Fairbanks and Wiebe 1980; Fairbanks et al. 1982 and others).

If there is a deep picnocline the DCM as a food resource is kept out of reach for the photic zone. The DCM might not develop at all if the picnocline is situated below the level of minimum light necessity. In both cases, without the ample food source the DCM means, the water above the picnocline is going to be nutrient depleted. As a consequence oligotrophic mixed layer species will dominate all the way down to the depth of the picnocline (Ravelo et al. 1990).

Results from borehole Tengelic-2

Introduction

Tengelic-2 is one of the best studied cores in Hungary, as is shown by the fact that a whole volume of Annual Reports of the Geological Institute was devoted to the subject (Halmai et al. 1982; Nagy et al. 1982). The borehole is situated North East of the Mecsek Mts in South West Hungary (Fig. 2), in the vicinity of the Kapos-line, just to the north (Némedi-Varga 1986).

The biostratigraphic frame accepted here is based on nannoplankton and foraminifera (Nagymarosy 1982; Korecz-Laky 1982). The investigated depth interval in the core (847–736 m) starts in NN5, the NN5/NN6 boundary (802 m) can be easily recognized (Nagymarosy 1982). Based on benthic foraminifera stratigraphy the Upper Lagenid Zone (Late Early Badenian) and the Agglutinated Zone can be identified, while Sarmatian/Badenian boundary is at 723 m (Korecz-Laky 1982). Absolute age of biozones are from Steininger (1988, 1990), while in the underlying tuff (872 m) 16±0.7 Ma was determined by the K/Ar method (Halmai et al. 1982) (Fig. 3).

Paleoenvironmental reconstruction of the Badenian based on Tengelic-2 data was far from the focus of attention. Previous authors referred to paleo-environment in Tengelic-2 as "shallow marine" in Early Badenian (Korecz-Laky 1982), without providing much evidence. In spite Bohn-Havas (1982) recognising a deep sublittoral mollusc species (*Amussium denudatum*), the overall picture drawn in the Annual Report by Nagy et al. (1982) is that it is shallower than sublittoral.

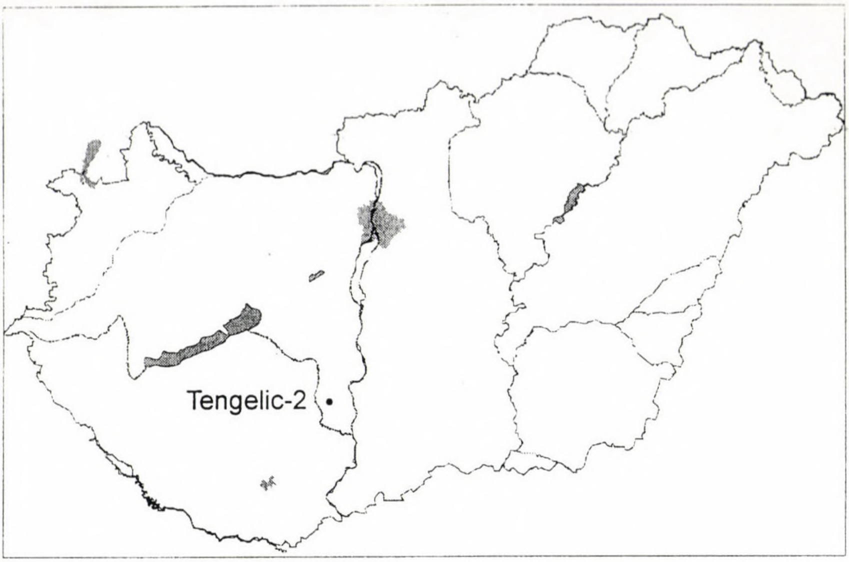


Fig. 2
The location of borehole Tengelic-2 in SW Hungary

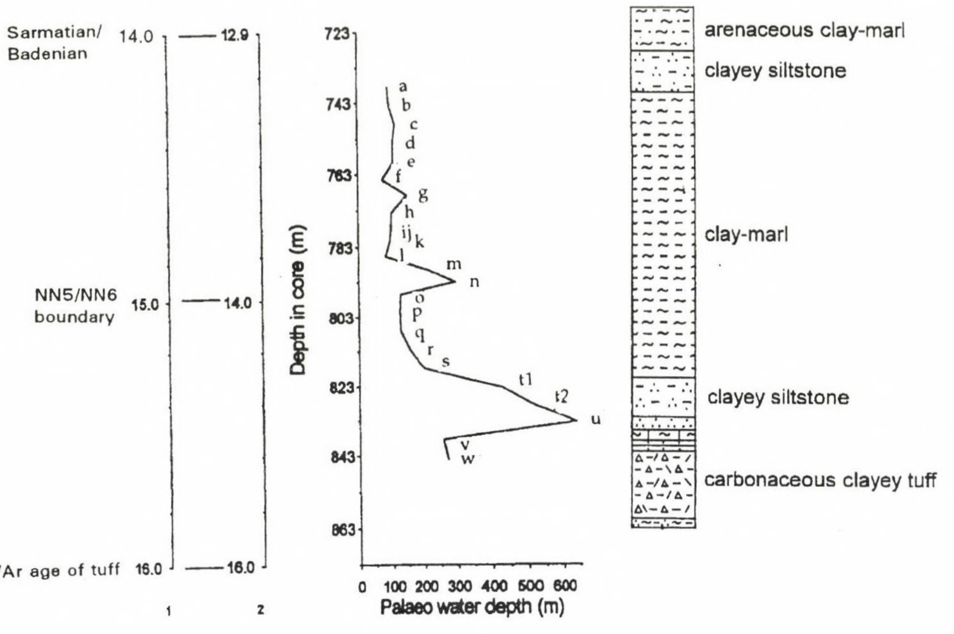
Reconstruction of water depth, an estimation

The benthonic foraminiferal assemblage was quantitatively analyzed (Table 1). Paleowaterdepth was calculated directly from the planktic/benthic ratio, based on the function described by van der Zwaan et al. (1990). In the model used here, paleoproductivity is measured by the number of planktonic foraminifera, while organic flux reaching the bottom is assessed by the number of epibenthic foraminifera. The total number of benthic foraminifera is corrected for the inbenthic species, which does not depend on the organic matter raining down from the surface layers, but has an unlimited food resource in the sediment. Estimated water depth plotted against depth in the core is in Fig. 3. Maximum water depth, approximately 600 m (upper bathyal) was reached in the Late Early Badenian.

There are environments, where the p/b ratio is not valid for estimating paleowaterdepth. Such environments are:

1. Eutrophic environments, in upwelling areas, where a plankton boom can result in high p\b ratios giving the false indication of great depth. Under such conditions the planktic foraminiferal assemblage is strongly dominated by only one or two species. In the case of Tengelic-2 core the planktic foraminifera fauna is far too diverse for such an environment.

2. Oxygen depleted bottoms, anoxic or dysoxic. Anoxia can be excluded based on the presence of benthic life. Dysoxic environments have a typical low



1 Absolute ages based on Steininger et al. 1984
2 Absolute ages based on Steininger et al. 1990

Fig. 3
Estimated paleowaterdepth (Table 1) plotted against depth in core, where samples are marked by letters. Left of it the absolute ages, where pinpoints are. 1. K/Ar age of underlying tuff (Halmai et al. 1982); 2. NN5–NN6 boundary (Nagymarosy 1982); 3. Sarmatian-Badenian boundary (Nagymarosy 1982; Korecz-Laky 1982). Right of it lithologic column after Halmai et al. (1982)

diversity of benthic foraminiferal fauna. In Tengelic-2, although low oxygen tolerant species are present e.g. *Bulimina elongata, Globobulimina pyrula, Valvulineria complanata* (van der Zwaan 1982), the overall high diversity suggests a totally different picture.

3. Environments with strong resedimentation. There are no signs of this in the material. Foraminiferal tests are well preserved, indicating no or not much redepositioning. Resedimentation can never be excluded, but even if present it makes our water depth estimations only deeper, as it can transport only sediments deposited in shallower environments to deeper ones.

The above-mentioned environments, where calculating p/b ratio is not a suitable method can be ruled out, thus these paleobathymetric estimations (Table 1, Fig. 3) have to be considered correct. This implies, that we do not agree with Nagy et al. (1982), but suggest an overall greater water depth, reaching even bathyal depths in the Late Early Badenian.

Density structure of surface water based on planktonic foraminifera of core Tengelic-2

The vertical distribution of planktic foraminifera is well documented by several authors (Fairbanks et al. 1982; Fairbanks and Wiebe 1980; Fairbanks et al. 1979; Douglas and Savin 1978; Ravelo et al. 1990 and others). Douglas and Savin (1978) provide evidence on depth stratification of plankton foraminifera from the Cretaceous times onwards using stable isotope data. Based on the relationship between shell morphology and depth distribution, depth habitats of morphologic categories are given relative to the position of the mixed layer and the thermocline. Here we are concerned only about the few categories present in Tengelic-2 material. According to Douglas and Savin (1978) the globigerinoid, globigerine and orbuline groups live above the thermocline, while the globorotalid group might extend below it.

In Tengelic-2 the planktonic foraminifera fauna was not studied with the same statistical accuracy as benthic foraminifera fauna, since their number studied in the present work did not always reach the statistically significant 200 specimens. Nevertheless the most common species are *Globigerina quinqueloba, G. obesa, Globigerinoides trilobus, G. quadrilobatus*[1]. These species fall into the categories of foraminifera with shallow or intermediate habitat, above the thermocline according to Douglas and Savin (1978). *Globorotalia cf. transylvanica*, an endemic Paratethys species with a rounded periphery without a keel, is present in the material, but only in the few deeper samples, and it is far from dominating in the assemblage (Fig. 4). Based on its endemicity and its primitive morphology compared to later deep dwelling forms (Hodell and Vayavananda 1993) and its scarce number in the material, we might doubt that it could indicate an environment below the picnocline. Thus we can conclude, that plankton foraminifera assemblage in Tengelic-2 is dominated by species living above the thermo-picnocline in the oligotrophic mixed layer.

Discussion

Mixed layer species are dominant throughout the whole core, where the p/b ratio indicates an approximate maximum of 600 m water depth. The mixed layer in present day oceans is generally not deeper then the photic zone, situated at about 200 m depth (Anikouchine and Sternberg 1981). This means, that we must suppose an unusually deep picnocline, if present. This suggests, that the Central Paratethys in Early Badenian time was an enclosed sea characterized by very weak density stratification. On the other hand the model for late Badenian by Kókay (1985) supposes a distinct shallow picnocline.

There are several explanations for changes in strength and depth of a picnocline. One of these explanations is to associate it with changes in the circulation pattern, as it is seen in the well-known debate over reversed

1 W. J. Zachariasse was kind to help me in determining the planctonic species

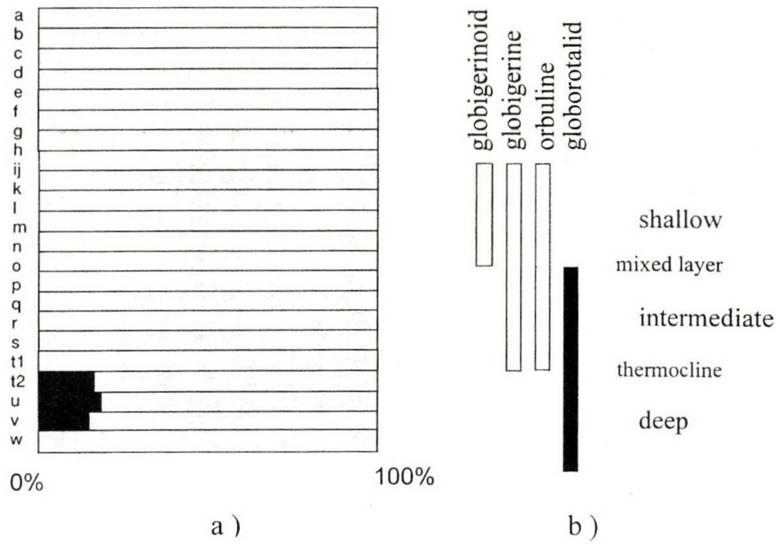


Fig. 4
Composition of planktonic foraminiferal fauna in Tengelic-2 and depth distribution of morphologic categories (Douglas and Savin 1978). a: Total number of planktonic picked is 100%. Light grey: Percentage of morphologic categories of globigerinoid, globigerine and orbuline foraminifera in total plankton. Dark grey: Percentage of morphologic category globorotalid in total plankton. b: Light grey: morphologic categories living only above the thermocline. Dark grey: morphologic category globorotalids with habitat extending below thermocline

circulation in the Mediterranean. Isotopic, sedimentological and faunal evidence suggests, that relative to the present day lagoonal circulation, the circulation was reversed to estuarine during sapropel formation (Howell and Thunell 1992; Thunell and Williams 1989; Buckley and Johnson 1987; Muerdter and Kennett 1984; Ross and Kennett 1983; Stanley et al. 1975 and others). Several authors claim, that sapropel deposition can happen only if there is a low salinity surface layer restricting vertical circulation and/or there is eutrophy in the surface layer (see Howell and Thunell 1992 for further references). The low salinity surface layer means there is a steep picnocline (Rossignol-Strick 1982) associated with the estuarine circulation. Emphasising the existence of such a picnocline suggests by common sense, that the absence of it characterizes the anti-estuarine circulation. This causes us to favour the idea of associating a deep mixed layer with oligotrophic species in the Early Badenian of the Central Paratethys with anti-estuarine circulation.

Speculation on circulation during Middle Badenian time

Middle Badenian sediments are rarely found in Hungary except for the occurrence of normal marine sediments described from bore holes from the metro construction of Budapest (Kókay 1990). Unfortunately there is not much consensus about what we call Middle Badenian. Here I follow Nagymarosy (1982) to call NN6 Middle Badenian. In Tengelic-2 in the timespan concerned, the p/b curve is too shallow to yield evidence about the possible depth of the mixed layer.

In the Carpathian region (Carpathian foredeep, Outer Carpathians and Transylvanian Basin) it is a well-known fact that there are evaporites of Mid-Badenian age. The genesis of these evaporites is a much debated topic (Marinescu and Marunteanu 1990; Peryt and Kasprzyk 1992; Liskowski 1989; Peryt and Jasionowski 1994 and others). Peryt and Jasionowski find evaporites from Southern Poland to be partly in situ and partly redeposited deep-water sediments. Evaporites deposited at great depth may have precipitated from highly saline deep water. The presence of highly saline deep water in the discussed areas can possibly be the result of a chain of enclosed seas with anti-estuarine circulation characterized by increasing salinity towards the East as a result of evaporation, as we will discuss in more detail below.

The evaporites found locally in Hungary have a younger than Middle Badenian age (Late Badenian–Sarmatian) and a totally different origin, as they are deposited in the shallow Zsámbék-basin, due to confinement (Jámbor 1976; Ravasz 1978).

Suggested paleoceanographic model for Early and Middle Badenian

According to the palinspastic reconstruction of Rögl and Steininger (1983), Steininger and Rögl (1984), Rögl and Müller (1978) the Central Paratethys was connected by a seaway to the Mediterranean through Istria and to the Eastern Paratethys. For the Early and Middle Badenian times we are proposing a hydrographic model, where surface water flows from the Mediterranean to the Paratethys with normal marine water. In return, more saline, Paratethyan deep water flows back to the Mediterranean (Fig. 5). The model implies that net evaporation is increasing toward the east, making possible the formation of evaporites in the Transylvanian Basin and the Outer Carpathians, perhaps as far as the Eastern Paratethys while normal marine sediments are deposited in the Pannonian Basin and in shallow parts of the basins, where the surface water dominates. Nevertheless, we cannot exclude the possibility that the chain of subbasins (Mediterranean–Central Paratethys–Eastern Paratethys) might have reversed circulation relative to each other like the present-day Mediterranean (anti-estuarine)–Sea of Marmora (estuarine)–Black Sea (estuarine) system.

Faunal response to change in circulation

Changes in the circulation pattern mean the setting up of a new environmental regime. Distribution of marine fauna is influenced by currents, as currents transport marine organisms to new realms. Buckley and Johnson (1987) explained variation in the planktonic foraminiferal assemblage of the Mediterranean by supposing sluggish or even reversed surface flow at Gibraltar falling short of restocking the Western Mediterranean with Atlantic species.

Table 1
Counting results of benthic foraminiferal taxa from borehole Tengelic-2 and calculations

Columns are samples, rows are recognized benthic foraminiferal taxa. See end of table for explanation of abbreviated names of benthic taxa. Inbenthic species are denoted by asterix, meaning habitat prefered deeper than 1 cm. Calculations are in the last rows: total number of benthic: "sum val", B; number of inbenthic: "num inb", I; number of benthic corrected for inbenthic: "cor ben", B*=B-I; split numerator "spl num", split denominator "split den"; number of planktic foraminifera picked: "pl", P; percentage of planktic "P%", P%= P*100\(P+B); corrected percentage of planktic: "P%*", P%*=P*100\(P+B*); paleowaterdepth estimation after van der Zwaan et al. (1990): "D", D=exp[3.58+(0.035+P%*]

	a	b	c	d	e	f	g	h	ij	k	l	m	n	o	p	q	r	s	t1	t2	u	w
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* Cas obl	2			12	3	7	5	2	6	4	4	7	3	3	7		1		1	5	12	1
Cas sgl	6	4	6	6	2	2				3		1	1	1	5		1		3	14	1	1
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Cib lb1	2	5	8	6	5	1	5	1	5	4	6	12	9	8	7	8	5	2	9	7	9	6
Cib lb2	7	10	3	4	3	2	3	7	11	4	6	2	5	10	14	8	10	2	18	17	18	14
Cib let	4	3	2									2	2		5	8	4		2	1	3	1
Cib ung	5	6	10	17	3	2	8	3	4	3	7	8	8	9	7	3	4	5	15	10	8	19
Cib juv							1															
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Epi sp1				1	3	3	3	9	2	4	3	3	1	2	8	1		2	6			
Epi sp2																			2			
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* Fur acu			1		2		5	4		2	3	7	3	1			1	4	5	1	3	
Gav pra	21	10	14	14	4	3	1	7	10	2	4	3	1	1	2		1	4	9		4	
* Glo pyr					2	2	2		1		1	1	4	1	1	3		2	7			
* Glo ova			2		1			4					2			2					1	
* Glo pup		1	5	2	2		2	4	4	4	1	4	13	2	2	3	3	4	1		2	
Glo ulina										2												
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Gut com	1	1	2			1						1							1			
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Gyr um1			1	1				1	2	1	1	3			1			1			3	1
Gyr um2		7	1	2	2		3	1			3		2	2	2	1		1	3	7	5	2
Gyr sol	6	2	10	2	8	1	4	2	2	6	4	2	21	5	6	1		2	2	5	1	5
Han bou	18	7	8	9	8	17	9	2	11	10	3	7	5	3		2	7	3	11	12	14	2
Het dut	6	7	3	6	2	2	4	11	3	6	8	6	7	16	10	14	19	9	16	11	12	36
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* Mel bar	3	6	5	1	1				4	2	1		2			2			6	4	4	

Table 1 (cont.)

* Mel po	7	9	3	3	6	3	8	2	1	6	4	3	2	3	8	6	10	3	17	3	6	13
Mil Ade										1			1			1						
Mil bil																4		3	1	1	1	1
Mil Qui					1	7	6	1	2		5	2	12	3	4	3	5	1	2	2		
Mil Spi exc																					1	
Mil Spi bad																			2	1	1	2
Mil ids							4						5		1	2	2	4		2		
Nod sp.		1		0.5				0.5					2.5	1.5	0.5	3	5		1	0.5		1
Nod																3	4	4	9	2		2
Nod raph																		8	1			
Non com	1	22	10	43	32	78	44	29	44	27	35	38	10	14	17	5	7	24	16		3	4
Non ella	1															1		2		2		
Ori umb				3	1		1		1	1		2						2	1	2		
Pla aus																						15
* Pul bul	39	26	33	8	15	12	3	9	8	16	13	11	8	15	6	10	17	15	19	8	2	5
* Pul qui								1						2					1	1	2	
Reu spi		1			1																	
Sip ret													1						3	5	4	19
sp1							3											3	3			
* Sph bul	11	11	6	5	1	1			4	3	4	4	2	1	4	8	4	2	2	2		4
Sti ado					0.2					1					1	6	5	1	2		1	3
Tri ang	3		7	1							1									1		1
Tri bra																					1	1
* Uvi his	1	3	1		1																	
* Uvi sem	8	9	16	4	6	2	4	10	15	7	15	14	5	16	10	4	3	5	3		8	
* Uvi pyg											6										1	
* Uvi ven			1	4					4			5		1								
* Uvi acu						11	7	3				3	3	2		15	7	3	1			12
* Uvi acminata																				5	1	23
und uvi		3									8						5	4	1	7	4	8
Vag pedum																			1			
* Val com	9	5	9	2	2	1	1	3		2	2	4	5			2	8	3	24	7	10	
UN DET		2	1	3	4	3	4	3	4	3	4	5	3	3	3	5	5	7	7	11	2	6
Hap sp.	1	2			1	3		4		1	1	1	1	4	1	6	2	1				
Mar com	1.8	2.1	3	1.25	2.2		0.2	0.3	0.2	1	2	1	4	6		2.25	1.8		0.8	0.2	2.1	2.7
Spir cari	15	11	14	3	17	15	6.5	13	5	10	3	5	7	10	4.5	2	3.5	5	12	11	12	8.5
Text sp.					1	3				2	3	1	5			2	5	6	2			1
Sig sp	4		3	2	2	2	1	3	1	8	1	2	1	2				1	2			1
Vul sp.		2			3	2	11		2	5		5	12	4	4	3	7	3	1			
UN DET				2	5	2	2	8	5	4		1	6	10		10	12	11	3	10	8	5
sum val	310	228	243	239	242	259	230	221	246	215	211	235	232	213	202	227	244	215	390	218	228	310
num inb	155	89	101	64	89	71	58	56	64	67	66	90	74	61	46	67	83	62	167	60	72	64
cor ben	155	139	142	175	153	188	172	165	182	148	145	145	158	152	156	160	161	153	223	158	156	246
spl num	1	3	3	5	3	1	3	1	1	1	1	1	1	3	1	7	5	129	1	1	5	1
spl den	512	2048	1024	2048	4096	512	2048	256	512	256	256	256	256	256	512	2048	4096	4096	4096	2048	512	256
pl	58	56	73	80	70	52	128	76	73	58	51	166	239	95	89	95	121	157	520	442	704	378
P%	15.8	19.7	23.1	25.1	22.4	16.7	35.8	25.6	22.9	21.2	19.5	41.4	50.8	30.9	30.6	29.5	33.2	42.2	57.2	67	75.5	55
P%*	27.3	28.7	34	31.4	31.3	21.7	42.7	31.6	28.6	28.2	26	53.4	60.3	38.5	36.4	37.3	43	50.6	70	73.7	81.9	60.6
D	94.7	99.6	120	110	109	77.7	163	110	99.3	97.7	90.6	238	304	141	131	135	165	216	429	489	652	308

Calcareous taxa: All tri: *Allomorphina trigonia*, Am bec: *Ammonia beccarii*, Asg pla: *Asterogerina planorbis*, Ast ita: *Astrononion* cf. *italicum*, Bol ant: *Bolivina antiqua*, Bol heb: *Bolivina hebes*, Bol pli: *Bolivina plicatella*, Bol ret: *Bolivina reticulata*, *Bol sca: Bolivina scalprata* var. *miocenica*, spa spa: *Bolivina spathulata f. spathulata*, spa dil: *Bolivina spathulata f. dilatata*, Bol tra: *Bolivina trajectina*, Bol vie: *Bolivina viennensis*, Bol spl: *Bolivina* sp., Bol und: *Bolivina*, Bul sp.: *Bulimina* sp., Bul acu: *Bulimina aculeata*, Bul cos: *Bulimina costata*, Can aur: *Cancris auriculus*, Cas cra: *Cassidulina crassa*, Cas obl: *Cassidulina oblonga*, Cas sgl: *Cassidulina subglobosa*, Cas lae: *Cassidulina laevigata*, Cas ter: *Cassidulina teretis*, Cer hau: *Ceratocancris hauerina*, Chil sp: *Chilostemella* sp., Cib lb1: *Cibicides lobatulus f. lobatulus*, Cib lb2: *Cibicides lobatulus f. refulgens*, Cib let: *Cibicides lobatulus* f. *letkesiensis*, Cib ung: *Cibicides ungerianus*, Cib juv: *Cibicides juveniles*, Cor sp.: *Coryphostoma* sp., Den tal: *Dentalina*, Den bre: *Dentalina brevis*,Den elg: *Dentalina elegans*, Elp idi: *Elphidium*, Epi spl: *Epistominella* species 1., Epi sp2: *Epistominella* species 2., Fiss sp.: *?Fissurina*, Fur acu: *Fursenkoina acuta*, Gav pra: *Gavelinopsis praegeri*, Globobulimina pyrula, Glo pyr: *Globobulimina pyrula* ssp. pyrula, Glo ova: *Globobulimina pyrula* ssp. *ovata*, Glo pup: *Globobulimina pyrula* ssp. pupoides, Glb uli: *Globulina* sp., Gut aus: *Guttulina austriaca*, Gut com: *Guttulina communis*, Gyr par: *Gyroidina parva*, Gyr uml: *Gyroidina umbonata* morphotype 1, Gyr um2: *Gyroidina umbonata* morphotype 2, Gyr sol: *Gyroidina soldanii*, Han bou: *Hansawaia boueana*,, Het dut: *Heterolepa dutemplei*, Het steg: *Heterostegina* sp., Hoe ele: *Hoeglundina elegans*, Lag ena: *Lagena*, Len cal: *Lenticulina calcar*, Len arc: *Lenticulina arcuata*, Len tic: *Lenticulina*, Mel bar: *Melonis barleeanu*, Mel po: *Melonis pompilioides*, *Miliolides*: Mil Ade: *Adelosina*, Mil Bil: biloculine forms, Mil Qui: *Quinqeloculina*, Mil Spi exc: *Spiroloculina excavata*, Mil Spi bad: *Spiroloculina badenensis*, Mil ids: *Miliolides*, Nod sp.: *Nodosaria* sp., Nod: *Nodosaria*, Nod raph: *Nodosaria raphanistrum*, Non com: *Nonion commune*, Non ella: *Nonionella turgida*, Ori umb: *Oridorsalis umbonatus*, Pla aus: *Planulina* sp., Pul bul: *Pullenia bulloides*, Pul qui: *Pullenia quinqueloba*, Reu spi: *Reusella spinulosa*, Sip ret: *Siphonina reticulata*, Spl: species 1.Sph bul: *Sphaeroidina bulloides*, Sti ado: *Stilostomella adolphina*, Tri ang: *Trifarina angulosa*, Tri bra: *Trifarina bradyi*, Uvi his: *Uvigerina hispida*, Uvi sem: *Uvigerina semiornata*, Uvi pyg: *Uvigerina pygmea*, Uvi ven: *Uvigerina venusta*, Uvi acu: *Uvigerina aculeata*, Uvi acm: *Uvigerina acuminata*, Uvi und: *Uvigerina*, *Vag ped: Vaginulopsis pedum*, Val com: *Valvulineria complanata*, UN DET: Undetermined calcareous specimen, Agglutinated taxa:Hap sp.: *Haplophragmoides* sp., Mar com: *Martinotiella communis*, Spir car: *Spiroplectammina carinata*, Text sp.: *Textularia* sp., Sig sp.: *Sigmoilopsis* sp., Vul sp.: *Textularia ?mexicana*, UN DET: Undetermined agglutinated specimens

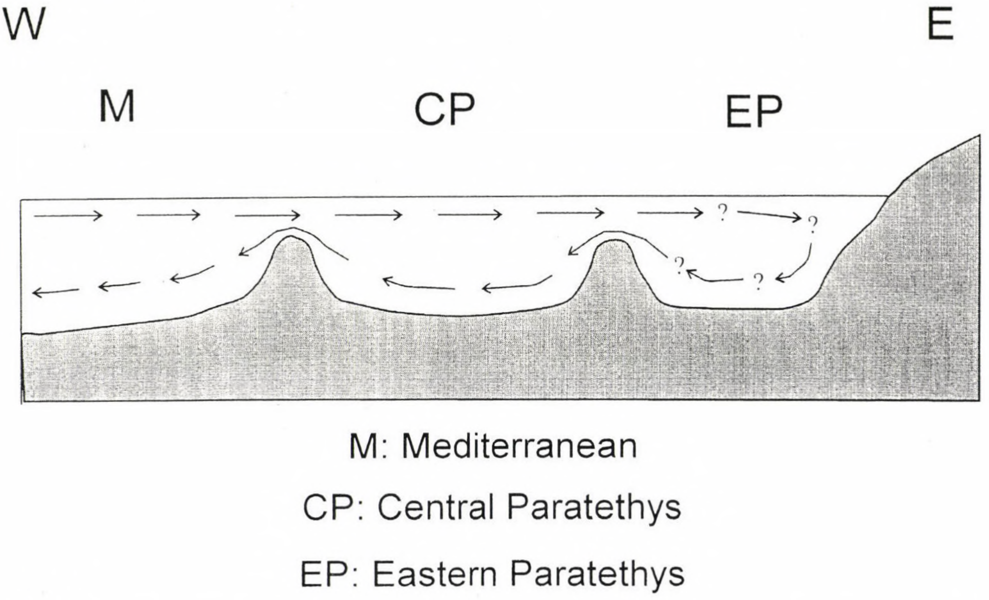

Fig. 5
Assumed antiestuarine (lagoonal) circulation pattern in Early and Mid-Badenian (Middle Miocene). Surface water arrives from the Mediterranean to the Central Paratethys and leaves towards the Eastern Paratethys. Deep water formed in the Paratethys flows back to the Mediterranean

Accepting the hydrographic model for Early and probably for Middle Badenian times (Fig. 5), the marine planktonic fauna must be considered to be carried by the surface currents from the Mediterranean to the Central Paratethys. The majority of benthic organisms, e.g. Mollusca, Crustacea, and other marine invertebrate groups has a planktic life stage, so their direction of spreading might coincide with plankton as well. Even benthic foraminifera lacking a planktic life stage have been reported to occur regularly in plankton samples (John 1987). Thus, we expect to find a similar fauna to the Mediterranean in the Central Paratethys, when the anti-estuarine regime prevailed. This idea is confirmed by Rögl and Müller (1978) finding that Early Badenian fauna in general show Mediterranean character.

On the other hand, in Late Badenian times, due to estuarine circulation the Paratethys did not receive surface water from the Mediterranean. The existence of a seaway making this possible is also doubted (Rögl and Müller 1978; Rögl and Steininger 1983; Steininger and Rögl 1984). During this time surface currents carried faunal elements from the East to the Central Paratethys. The Eastern relationship of the Central Paratethyan fauna is supported by Rögl and Müller's (1985) finding of Late Badenian assemblages partly Indo-Pacific origin and by Kókay's (1985) observation of a supposedly Eastern Paratethys connection.

It is worthwhile mentioning here the ecozones for the Neogene set up by Kordos et al. (1987) based on corals, diatoms, rodents, macroflora and pollen. The boundary of ecozone B and C falls at 15 Ma, coinciding with Middle Badenian–Late Badenian boundary (Vass et al. 1987), the proposed time of the switch in circulation. Keeping in mind the fauna-forming potential of circulation and its relation to land climate as discussed above, we are inclined to consider the ecozone boundary as a possible result of a reversal in circulation.

Conclusions

The quantitative analyses of foraminifera made possible the reconstruction of a deep oligotrophic mixed layer without density stratification in Early Badenian times in the region of Tengelic, Central Paratethys. The planktic benthic ratio showed as deep as upper bathyal depths, where planktic foraminiferal assemblages were dominated by mixed layer species. Such a deep mixed layer is considered possibly to indicate anti-estuarine (lagoonal) circulation. The idea is brought up, that anti-estuarine conditions also prevailed in Middle Badenian times and in the whole Paratethys (Fig. 5).

The independent method to confirm the suggested idea on circulation and surface water density structure is stable isotope (O^{18} and C^{13}) measurements. One would expect, that such investigations could identify the Mediterranean character of surface water, and would show that deep saline water was indeed formed in the Paratethys. Unfortunately, for the studied time interval, we have no knowledge of successful measurements. Two factors are suspected to play a role in this, one is the bad preservation of sediments for isotope analyses and the other is financial.

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Dinoflagellate stratigraphy of the Senonian formations of the Transdanubian Range

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The dinoflagellate stratigraphy presented here for the first time assigns the Senonian marine formations of the Transdanubian Range to two assemblage zones and, within them, to six subzones:
– *Odontochitina operculata* Assemblage Zone: Subzone-1, *Apteodinium deflandrei*, *Tarsisphaeridium geminiporatum*, and *Dinogymnium digitus* subzones;
– *Pyxidinopsis bakonyensis* Assemblage Zone: *Manumiella* div. sp. and *Pterodinium cingulatum–Isabelidinium bakeri* subzones.

On the basis of the correlation with the nannoplankton zonation (CC16–CC22c), the marine formations were deposited in the time interval between the Late Santonian and Late Campanian.

The common boundary of the two dinoflagellate assemblage zones coincides with the boundary of nannozones CC21–CC22, which may provide a reliable basis for further correlations.

Based upon the investigation of the dinoflagellate associations, the deposition of the Senonian marine formations took place in shallow marine, outer and inner shelf environments as well as in an environment of oceanic influence – simultaneously with the gradual expansion of Tethys.

Key words: dinoflagellate stratigraphy, Transdanubian Range, correlation, nannoplankton zones, Late Santonian–Late Campanian, paleoenvironment, shallow marine–oceanic

Geologic environment

In Hungary, four Upper Cretaceous areas of different facies are known (Fig. 1).

Upper Cretaceous formations of the Transdanubian Range were deposited predominantly on Upper Triassic, Jurassic or Lower Cretaceous carbonate surfaces. In the sequences, the marine formations develop continuously from the fluvial–lacustrine Csehbánya and the lacustrine–paludal Ajka Coal Formations, respectively (Fig. 2).

The oldest marine formation is the Jákó Marl Formation, whose initial layers of brackish-water, lagoonal facies are included in the Csingervölgy Member with corals and molluscs (Haas 1983). In the younger part of the formation, the carbonate content shows a tendency to increase. The average thickness of the formation is 60–80 m, and can even reach 100 m in some places.

The Ugod Limestone Formation of rudist facies is partly of the same age as the Jákó, and partly of the Polány Marl Formation. Its thickness may attain 200 m.

Both horizontally and vertically, the Polány Marl Formation is the most extended Senonian formation. In the SW foreland of the Bakony Mts, it may reach 800 m in thickness; however, its complete thickness is unknown due to subsequent erosion. In the lower part of the formation, the Rendek Member

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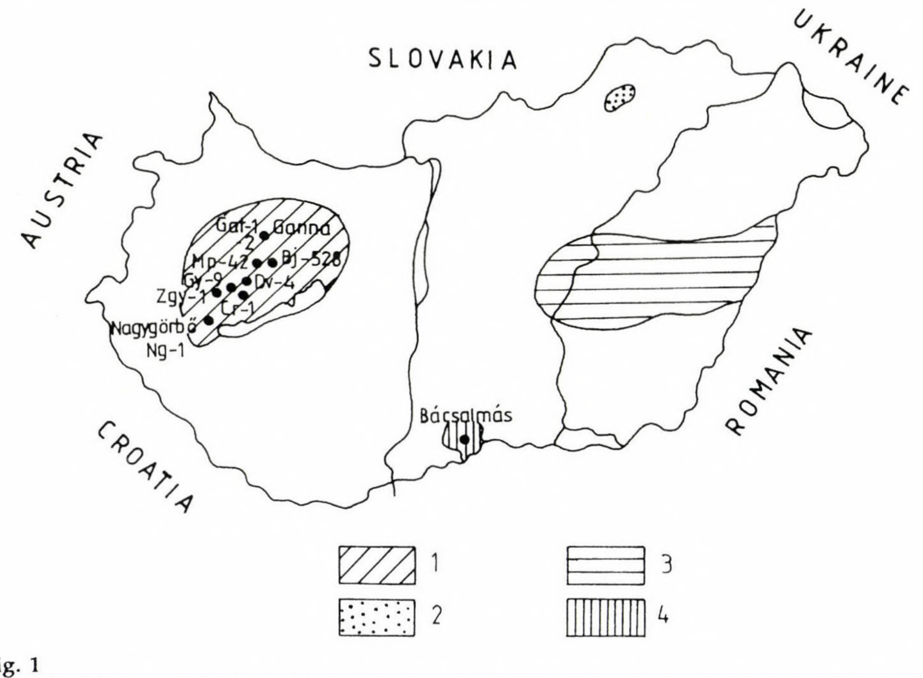


Fig. 1
Extend of Senonian facies units in Hungary. 1. Central Range-type (Bakony and Keszthely Mountains); 2. Gosau-type (Uppony Mountains); 3. Flysch-type (Trans-Tisza region); 4. Bácsalmás-type

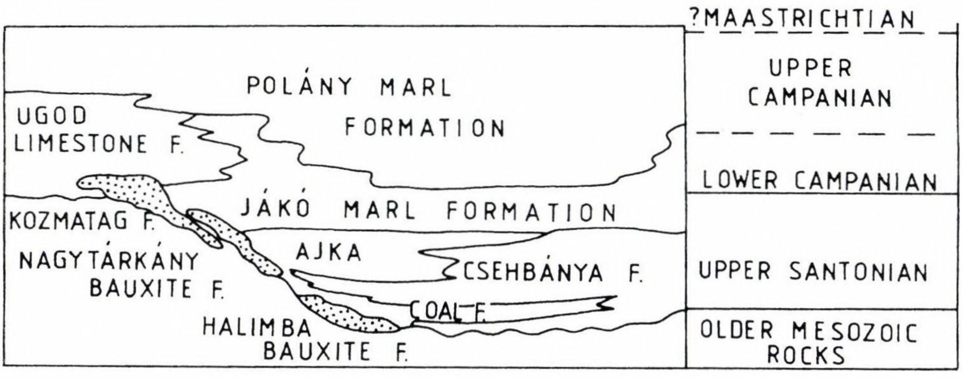


Fig. 2
Simplified section to illustrate relationships Senonian formations

with higher carbonate content (Haas et al. 1985), followed by the Jákó-hegy Breccia (indicating intermittent slumping), and in its upper part, the strongly sandy Ganna Member (Haas 1983) can be distinguished (Fig. 5).

Most of the boreholes (Bakonyjákó Bj-528, Csabrendek Cr-l, Devecser Dv-4, Gyepükaján Gy-9, Magyarpolány Mp-42, Zalagyömörő Zgy-1) investigated in order to establish the dinoflagellate stratigraphy were drilled in the territory of the Bakony Mts, and borehole Nagygörbő Ng-1 in the Keszthely Mts (Fig. 3).

Borehole Ng-1 exposed only the Polány Marl Formation, while borehole Gy-9 also encountered the Jákó Marl and polány Marl in addition to the Ugod Limestone Formation, as did the other boreholes in the Bakony Mts. In addition to the above-listed boreholes, this paper will provide reference data from some non-systematically studied sections (Celldömölk Cell-1, Ganna-1, -2, Magyarpolány Mp-37, Pápa-2, Vinár-1), as well.

In selecting the sequences for dinoflagellate investigations, the most important requirements were as follows:

– to represent both the marginal and inner-basin areas;

– to obtain as complete as possible a Senonian sequence during the correlation (Fig. 4);

– to compare, within the parts of the basins of different evolutionary history, the initial phases of sedimentation;

– to attempt to correlate with the nannoplankton zonation;

– to correlate with the formations of borehole Bj-528, investigated by integrated stratigraphic methods (magnetostratigraphy and biostratigraphy), and to pinpoint the Santonian–Campanian boundary.

Historical overview

From the territory of Tethys, regionally effective dinoflagellate zonations were published by Robaszynski et al. (1980), Decommer (1982), Schumacher- Lambry (1977), Foucher in Robaszynski et al. (1985), Herngreen et al. (1986), Kirsch (1991), etc.

The dinoflagellate stratigraphy carried out on the basis of the Upper Cretaceous formations of the Helveticum and Ultrahelveticum tectonic zones of Upper Bavaria can be regarded as being the most complete. The Turonian–Maastrichtian formations, which were exposed in a non-continuous sequence, but were correlatable, could be assigned to six interval zones and, within them, to four subzones (Kirsch 1991). The dinoflagellate zones were correlated by Kirsch with the foraminifer zones of Robaszynski et al. (1984). The demonstrated dinoflagellate association contains several elements in common with the assemblages known from the territory of the Transdanubian Range.

The comprehensive work of Williams and Bujak (1985), extended also to the territory of Tethys, and is of reference value.

The first not yet systematical Hungarian dinoflagellate investigations are linked with the name of Góczán (1962). He assigned the species determined by himself to the Upper Campanian–Upper Maastrichtian. In his palynological standard, one of the eponyms of the Upper Maastrichtian palynozone

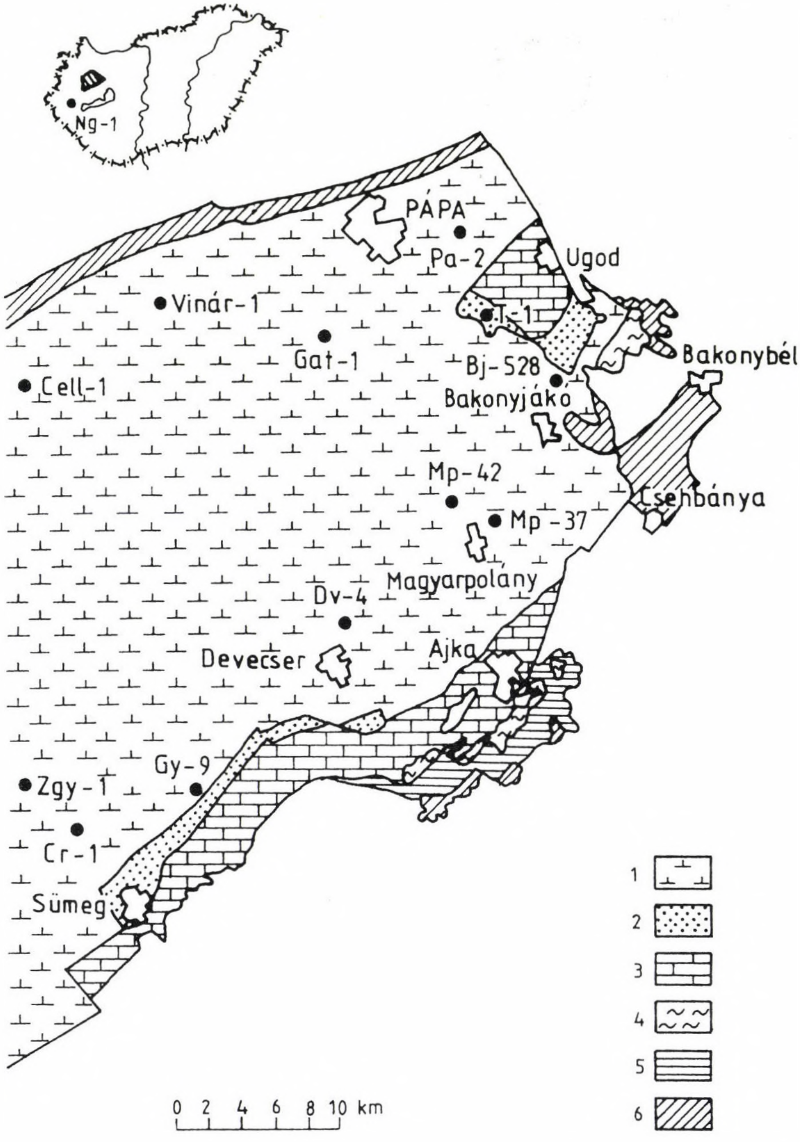


Fig. 3
Map of the Senonian formations of the Transdanubian Central Range with indication of the locations of the important key drill sections. 1. Polány Marl Formation; 2. transition between the Ugod and the Polány Marl Formation; 3. Ugod Limestone Formation; 4. Jákó Marl Formation; 5. Ajka Coal Formation; 6. Csehbánya Formation. (after Haas 1983)

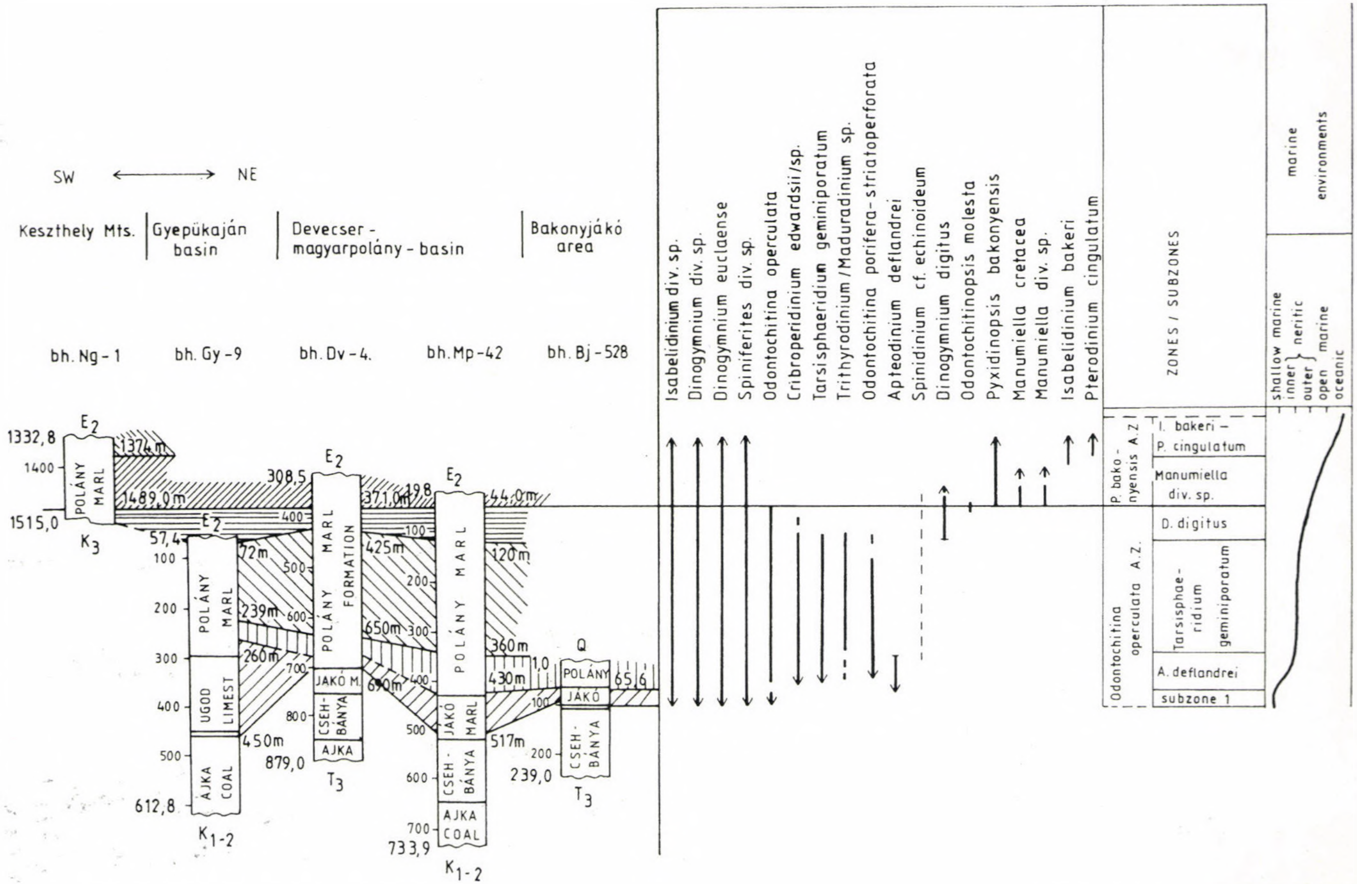


Fig. 4
Dinoflagellata-zones and environments of the Senonian formations in Transdanubian Central Range (Hungary)

Pyxidinopsis bakonyensis (Góczán 1962) Stover and Evitt (1978), is a dinoflagellate species characteristic of the assemblage zone (Góczán 1973).

A systematic dinoflagellate investigation of stratigraphic character of the Senonian formations of the Transdanubian Range was begun in recent times (Siegl-Farkas 1995; Siegl-Farkas and Wagreich 1994, 1996).

According to the investigations, the Jákó Marl is still characterised by poorish, and the Polány Marl by rich dinoflagellate associations. Exceptions are constituted by beds of high carbonate content of the latter formation: the Jákóhegy Breccia and Rendek Members. Dinoflagellates have not been determined in the Ugod Limestone Formation.

The flora list and the occurrences of stratigraphically more important species of the associations from boreholes of the Gyepükaján Basin (Gy-9, Cr-1, Zgy-1) as well as boreholes Bj-528 and Ng-1 can be studied in the works of Siegl-Farkas (1995), Siegl-Farkas and Wagreich (1994) and Lantos et al. (1996).

In this paper, the zonal index species as well as the assemblages characterising the biozones will be presented.

Description of dinoflagellate zones

Odontochitina operculata Assemblage Zone

Eponymous species: *Odontochitina operculata* (O. Wetzel 1933) Deflandre et Cookson 1955, Pl. I, Figs 1–2.

The *Odontochitina operculata* Assemblage Zone was established by Williams (1975) for all of the eastern Canadian Campanian formations. It was proposed by Williams and Bujak (1985) as a global ("world-wide") zone, although they did not have sufficient data on the area of Tethys.

Areal occurrence and age classification of the species: Australia, Canada, Europe, New Guinea: Lower Cretaceous–Maastrichtian (e.g. Wetzel 1933: Germany, Senonian; Deflandre and Cookson 1955: Australia, New Guinea, Lower Cretaceous; Cookson and Eisenack 1958: Australia, Albian–Lower Turonian; Góczán 1962: Hungary, Upper Campanian; Gorka 1963: Poland, Cenomanian–Campanian; Clarke and Verdier 1967: England, Cenomanian–Campanian; Millioud 1969: Germany, France, Upper Hauterivian; Davey and Verdier 1971: France, Albian; Corradini 1973: Italy, Upper Cretaceous; Antonescu 1972: Romania, Cenomanian–Santonian; Yun 1981: Germany, Lower Santonian; Robaszynski et al. 1985: the Netherlands, Maastrichtian; Ioannides 1986: Canada, Upper Cretaceous; Kirsch 1991: Bavaria, Coniacian–Campanian).

In the formations of the Transdanubian Range, the occurrence of *Odontochitina operculata* (O. Wetzel 1933) Deflandre and Cookson 1955 is as follows: Bj-528: 51.6 m, Cell-1: 2579.5 m, Dv-4: 371.0–635.1 m, Gy-9: 145.6–189.7 m, Mp-42: 46.5–331.0 m, Ng-1: 1489.0 m, Zgy-1: 236.7 m.

On the basis of the data listed above, the species can appear in the deepest part of the Jákó Marl Formation, but becomes a species of regular occurrence only in the associations of the Polány Marl Formation, and disappears from the associations in the upper third of the known thickness of the latter unit.

By its last appearance, it designates the main correlation plane of the Senonian marine formations of the Transdanubian Range (Fig. 3).

Within the zone, in its middle part, the species *Odontochitina porifera* and *O. striatoperforata* can be found in a much greater number than the eponymous species.

Age of the zone: Upper Santonian–Middle Campanian (Nannozones CC17–CC21).

In the zone, four subzones can be distinguished:

Subzone 1

The dinoflagellate association is of a more poorish composition. It is easier to find organic foraminifer tests than phytoplanktons. Species of the genera *Dinogymnium, Isabelidinium,* and *Spiniferites,* occurring regularly in the sequence, appear here.

Hopefully, later investigations will provide an opportunity for assigning the subzone a name.

Older beds of the Jákó Marl Formation belong to this subzone.

Apteodinium deflandrei subzone

Eponymous species: *Apteodinium deflandrei* (Clarke et Verdier 1967) Lucas-Clark 1987, Pl. I, Fig. 6, Pl. II, Figs 3, 5.

Areal occurrence and age classification of the species: Canada, Europe, Turonian–Middle Campanian (e.g.: Clarke and Verdier 1967: England, Senonian; Foucher 1972, 1974: France, Turonian–Coniacian; Williams 1975: Canada, Campanian; Yun 1981: Germany, Lower Santonian; Robaszynsky et al. 1983: Belgium, Middle Campanian, the Netherlands, Lower Campanian; Kirsch 1991: Bavaria, Coniacian–Middle Campanian).

In his zonation, established for the Canadian formations, Williams (1975) marks the final occurrence of the species in the *Odontochitina operculata* zone, covering the Campanian.

Simultaneously with the expansion of the marine environment, the number of species and specimens of dinoflagellates increased suddenly in the territory of the Transdanubian Range. Besides the regular occurrence of the eponymous species, some species of the genus *Trithyrodinium* (also occurring regularly), *Odontochitina porifera* and *O. striatoperforata, Cribroperidinium edwardsii* and *Tarsisphaeridium geminiporatum* appear among others in this subzone.

The upper part of the Jákó Marl and the deepest part of the Polány Marl Formation can be assigned here.

Age of the subzone: Lower Campanian (Nannozones CC17–CC18).

Tarsisphaeridium geminiporatum subzone

Eponymous species: *Tarsisphaeridium geminiporatum* Riegel 1974 (Prasinophyta), Pl. II, Figs 1, 4.

Areal occurrence and age classification of the species: Riegel (1974) described the taxon from Spanish Upper Cretaceous (Senonian?) formations, while Kirsch (1991) described its occurrence from Upper Bavarian Santonian–Campanian formations.

The lower boundary of the subzone is marked by the absence of the species *Apteodinium deflandrei*. The common occurrence of the eponymous species and *Cribroperidinium edwardsii* is characteristic. In the upper part, *Trithyrodinium* and *Odontochitina* species can be found only sporadically. The middle part of the Polány Marl Formation can be assigned here.

Age of the subzone: Lower–Middle Campanian (Nannozones CC18–CC20).

Dinogymnium digitus subzone

Eponymous species: *Dinogymnium digitus* (Deflandre 1935) Evitt, Clarke et Verdier 1967, Pl. IV. Fig. 4, Pl. V. Fig. 5.

Areal occurrence and age classification of the species: Africa, America, Australia, Europe, India, Turonian–Maastrichtian (e.g. Deflandre 1935: France, Senonian; Eisenack 1961: Australia (Rough Range), Upper Cretaceous; Vozzhennikova 1967: USSR (Kazakhstan), Turonian; Jain et al. 1975: India (Assam), Maastrichtian; Rauscher and Dubinger 1982: Morocco, Maastrichtian; Firth 1987: USA (Georgia), Maastrichtian; Schrank 1987: Egypt, Campanian–Maastrichtian).

The subzone was emplaced in the uppermost part of the assemblage zone, on the basis of the absence of the species listed above and the regular occurrence of the eponymous species. In its upper part, the index species *Odontochitinopsis molesta* appears first. The association is of poorish composition, probably due to an increase in carbonate content.

The lower part of the upper third of the Polány Marl Formation belongs here.

Age of the subzone: upper part of the Middle Campanian (Nannozones CC21).

Pyxidinopsis bakonyensis Assemblage Zone

Eponymous species: *Pyxidinopsis bakonyensis* (Góczán 1962) Stover et Evitt 1978, Pl. VI, Figs 1–2.

Areal occurrence and age classification of the species: Góczán (1962) described the taxon from the borehole Bakonypölöske-1 of the Bakony Mountains, and determined its age to be Upper Maastrichtian. Later, (Góczán 1973) he identified it as one of the eponymous species of the youngest palynozone in his palynologic standard. Since then, it has been found in several boreholes in the territory of the Transdanubian Range. From other areas of Tethys, its occurrence has not been mentioned until recently. In the association of the Tercis profile (France), designated as the type section of the Campanian–Maastrichtian (Brussels, 1995), I determined it with several specimens. Thus, on the basis of its occurrence here, the zone is already of regional value.

The fitting of the assemblage zone into the global zonation is also made easier by the very characteristic, subzonal index species *Manumiella*, appearing at its lower boundary.

In the investigated boreholes of the Transdanubian Range, the occurrence of *Pyxidinopsis bakonyensis* (Góczán 1962) Stover et Evitt 1978 is as follows: Dv-4: 308.5–341.8 m, Mp-42: 30.1–34.0 m, Ng-1: 1332.8–1476.0 m.

Both in numbers of species and specimens, the dinoflagellate association is of poorer composition than in the assemblages of the deeper formations. The eponymous species is of regular and frequent occurrence. With it appear some new genera and species, respectively, which justified the establishment of two subzones. The zone is most developed in borehole Ng-1 (Siegl-Farkas and Wagreich 1994), but the upper part of the Ganna Member, exposed in boreholes Dv-4 and Mp-42, can also be assigned here.

Age of the zone: Late Campanian (Nannozone CC22abc).

Manumiella div. sp. subzone

Eponymous species: *Manumiella cretacea* (Cookson 1956) Bujak et Davies 1983, and *M.* div sp., pl. VII, Figs 2–5, 7–8.

Areal occurrence and age classification of the species: Australia, Canada, Germany, USA, Santonian–Maastrichtian (e.g. Cookson 1956: Australia, Upper Cretaceous; Drugg 1967: California, Late Cretaceous, Paleocene, C/T boundary; Bujak and Williams 1978: Canada, Maastrichtian; Helby et al. 1987: Australia, Santonian–Lower Maastrichtian; Kirsch 1991: Bavaria, Middle Campanian.

Helby et al. (1987) assigned the Santonian Australian formations to the *M. cretaceum* zone.

Among the boreholes investigated in detail, the association of the subzone can be encountered in the uppermost part of the Polány Marl Formation of boreholes Dv-4, Mp-42 as well as in the deepest part of borehole Ng-1. The species, characteristic here, can also be traced in the associations of sections Ganna-1 and Ganna-2.

Manumiella cretacea is of frequent occurrence; the species *Manumiella seelandica* and *Manumiella hungarica* n. sp., as well as *Cannosphaeropsis utinensis, Tanyosphaeridium* sp., and *Veryhachium reductum* are represented by a single specimen each.

The genus represents an important link in the chain of the evolutionary history of Deflandreaceae towards the Paleogene Wetzellielloideae (Stover and Williams 1987).

Age of the subzone: Late Campanian (Nannozone CC22ab).

Isabelidinium bakeri – Pterodinium cingulatum subzone

Eponymous species: *Isabelidinium bakeri* (Deflandre et Cookson 1955) Lentin et Williams 1977, Pl. VII, Fig. 1, *Pterodinium cingulatum* (O. Wetzel 1933) Below 1981, Pl. IV, Figs 1, 6.

Areal occurrence and age classification of the species: Australia, Canada, Europe, Lower Santonian–Eocene (e.g. Deflandre and Cookson 1955: Australia, Paleocene, Eocene; Kjellström 1973: Sweden, Maastrichtian; Yun 1981: Germany, Lower Santonian; Ioannides 1986: Canada, Upper Cretaceous; Kirsch 1991: Bavaria, Maastrichtian).

P. cingulatum: Africa, Australia, Europe, Cenomanian–Lower Tertiary (e.g. Wetzel 1933: Baltic countries, Upper Cretaceous; Davey and Williams in Davey et al. 1966: England, Cenomanian; Clarke and Verdier 1967: England, Cenomanian, Santonian; Davey and Verdier 1971: France, Albian; Kjellström 1973: Sweden, Maastrichtian; Foucher and Taugourdeau 1975: France, Cenomanian; Norvick 1976: Australia, Cenomanian; Below 1981: Morocco, Lower Cretaceous; Yun 1981: Germany, Lower Santonian; Robaszynski et al. 1985: the Netherlands, Lower Campanian; Ioannides 1986: Canada, Upper Cretaceous–Lower Tertiary; Kirsch 1991: Bavaria, Santonian–Lower Maastrichtian; Marheinecke 1992: N Germany, Maastrichtian (sumensis zone)).

The subzone was designated on the basis of the association of the uppermost part of borehole Ng-1; however, some elements of the assemblage are known from the surroundings of Pápa, Magyarpolány, and Vinár. In addition to the eponymous species, some taxa are determined only from here: *Canninginopsis denticulata*, *Paleocystodinium* sp., and *Fromea amphora*. The above-listed species represent the youngest part of the Polány Marl Formation known so far.

Age of the subzone: Late Campanian (Nannozone CC22bc).

Remarks: Some specimens of *Odontochitinopsis molesta* (Deflandre 1937) Eisenack 1961 have been determined only at the common boundary of the *Odontochitina operculata–Pyxidinopsis bakonyensis* Assemblage Zones (Dv-4: 369.0 m, Mp-42: 46.5 m, Ng-1: 1489.0 m).

Deflandre described this genus or species, respectively, based on a single specimen, from the Senonian of the Paris Basin in 1937. According to our knowledge, since then no newer specimen has been found.

The occurrence of the species in such a narrow time interval confirms the justification of the main correlation plane established for the territory of the basins of the Keszthely and Bakony Mountains; thus, it qualifies as an index dinoflagellate.

Its age: Late Campanian (boundary of Nannozones CC21–CC22).

Correlation of dinoflagellate and nannozones. Problems of age classification

The Senonian formations of the Transdanubian Range were classified differently by the various paleontological investigations (Góczán 1964, 1973; Sidó 1983; Czabalay 1983).

The foraminifer investigations proposed the Coniacian–Upper Maastrichtian as time of deposition of the entire sequence in the Transdanubian Range, and subdivided the marine section into four foraminifer zones (*G. concavata*, *G. arca–G. stuartiformis*, *G. conica–G. stuarti*, and *G. contusa–G. mayaroensis*; Sidó 1983).

The later results of the investigations of borehole Magyarpolány Mp-42 justify the age classification of Sidó (Budai 1983), already applying, however, the 1985 foraminifer zonation of Charon (Bodrogi 1993).

Palynostratigraphy suggested the deposition of the formations in the time interval of the Upper Santonian–Upper Maastrichtian (Góczán 1964, 1973; Góczán and Siegl-Farkas 1989, 1990; Siegl-Farkas 1993).

The method of palynostratigraphy left the possibility of the first correlation with ammonites – *Placenticeras polyopsis* (Dujardin), (Summesberger in Partényi, 1986) – out of consideration.

In order to eliminate the uncertainty of age classification, a problem for more than 30 years, correlation using the global nannoplankton zonation, accepted for the Mediterranean areas of Tethys, was employed Wagreich in Siegl-Farkas and Wagreich (1996) (Fig. 5).

The first possibility of this kind was provided by the comparison and the correlation with the nannoplankton zonation of the sporomorph associations of the marine Gosau-type layers of the Gams Basin (Austria) and the basal Upper Cretaceous formations of the Transdanubian Range (Siegl-Farkas and Wagreich 1996). According to this, the deposition of the Senonian formations of the Transdanubian Range began in the upper part of the *Oculopollis–Complexiopollis* Dominance Zone and at the time of Nannozone CC16, respectively.

It was followed by the integrated stratigraphical investigation of borehole Bj-528, which emplaced the Santonian–Campanian boundary in the Jákó Marl Formation, at the boundary of Chronozones C33r–C34n, in Nannozone CC17b, in the *Hungaropollis* Dominance Zone and in the lower part of Subzone *Apteodinium deflandrei* of the *Odontochitina operculata* Assemblage Zone (Lantos et al. 1996). On this basis, deposition of the marine layers began at the end of the Late Santonian.

On the basis of the correlation of dinoflagellate and nannoplankton zones of boreholes Gat-1, Mp-42 and Ng-1, covering the entire Senonian profile of the Transdanubian Range, the formations were deposited between the early stages of the Upper Santonian and the late stages of the Upper Campanian (Siegl-Farkas and Wagreich 1996). Within it, the marine formations were formed

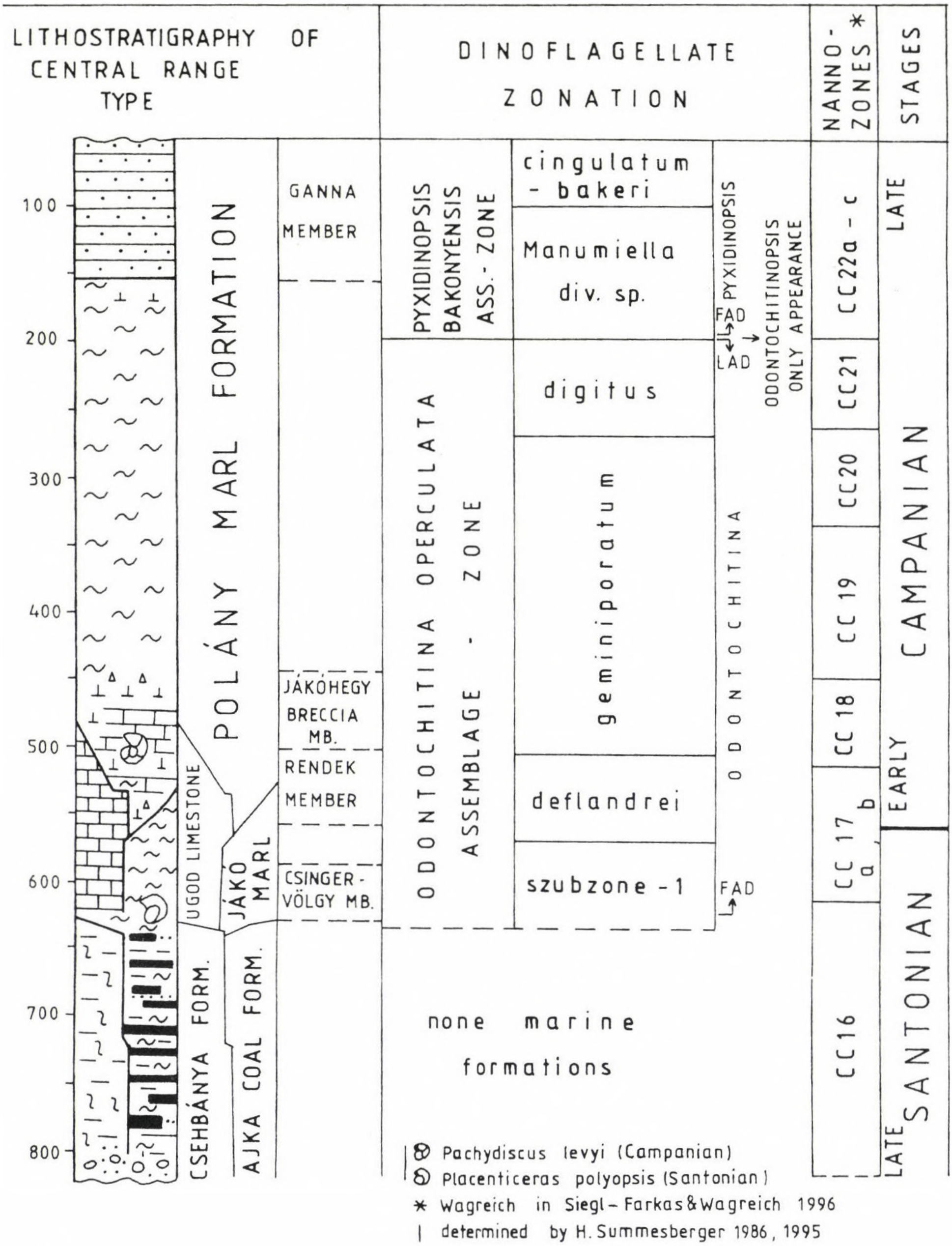


Fig. 5
Correlation of Dinoflagellata- and Nannozones in Transdanubian Central Range (Hungary)

at the time of the *Odontochitina operculata* and *Pyxidinopsis bakonyensis* Assemblage Zones, which is during the period between the late stage of Nannozone CC16 and the zone CC22c, or at the end of the Late Santonian and during almost the entire Campanian (about 9 Ma).

In the upper part of the Polány Marl, the boundaries of the two established assemblage zones coincide with the boundary of the *Palaeostomocystis bakonyensis–Pseudopapillopollis praesubhercynicus* (Góczán 1973) palynozone as well as (to judge by borehole Ng-1) with the boundary of nannozones CC21–CC22. It gives a reliable basis for further correlations.

According to the revisionary investigations of Summesberger (1995, pers. comm. Summersberger and Siegl-Farkas 1997 in prep.), the ammonite specimen found in the lower third of the Polány Marl, determined by Noszky (1944; Museum of the Hungarian Geological Institute: K8645) as *Pachydiscus neubergicus* (Hauer), and misleading the parastratigraphic investigations, is *Pachydiscus levyi* De Grossouvre 1894 Pl. VIII. Figs 1–3. Its occurrence is characteristic of the Lower Campanian (Fig. 5).

The Congress of Cretaceous Stage Boundaries, held in 1995 in Brussels, decided to designate the boundary of the Campanian–Maastrichtian at the first appearance of *Pachydiscus neubergicus* (Hauer) or, in its absence, in the biozones correlated with it. I think that until the results of the investigations concerning this question are published, Maastrichtian "near-boundary" formations cannot be traced anywhere in the territory of the Transdanubian Range, unless possibly in the sequence of borehole Gat-1.

Paleoenvironmental conclusions

The Senonian formations of the Transdanubian Range represent a classic transgressional sequence, in which the marine formations develop gradually from the fresh-water ones.

On the basis of the dinoflagellate associations, five significant changes in environment can be recorded as a function of the expansion of the transgression (three at the time of the *Odontochitina operculata* Zone, and two at the time of the *Pyxidinopsis bakonyensis* Zone):

– the depositional area (Subzone-1) of the Jákó Marl Formation is shallow marine, which is indicated by the scarce association and the genus *Odontochitina*, appearing here and enduring even changing salinity (Lister and Batten 1988).

– the lower and middle parts of the Polány Marl Formation were deposited in an inner neritic environment at the time of the *Apteodinium deflandrei* and *Tarsisphaeridium geminiporatum* Subzones. *Cribroperidinium*, occurring frequently and together with the eponyms of the subzones (e.g. on the basis of the investigations of May (1980), Brinkhuis and Zachariasse (1988), Marschall and Batten (1988), Wilpshaar and Leereveld (1995)), is characteristic of the inner neritic and the pronouncedly marine environment.

– the beds of the time of the *Dinogymnium digitus* Subzone (lower part of the Ganna Member) are considered to be sediments of a neritic–outer neritic environment. The scarce occurrence of phytoplanktons may be due to the change in lithofacies.

– the species *Veryhachium reductum*, *Tanyosphaeridium* sp., and *Cannosphaeropsis utinensis* with a long process, determined in the formations of the *Manumiella* div. sp. subzone of the *Pyxidinopsis bakonyesis* Assemblage Zone, and represented by only one specimen each, indicate open marine influence (Ganna Member, Ng-1).

– the association of the *Pterodinium cingulatum–Isabelidinium bakeri* subzone (*Fromea amphora, Canninginopsis denticulata*, and *Palaeocystodinium* sp., with the increasingly frequent *Pyxidinopsis bakonyensis*) already indicates oceanic influence (youngest part of the Ganna Member, Ng-1). The oceanic environmental demand of the genus *Pterodinium* is referred to by Harland (1983) and Wilpshaar and Leereveld (1995).

Lentin and Williams (1980) distinguished three global climate provinces (tropical–subtropical, warm, boreal) for the Campanian. On the basis of the ecological demands of the genera designated by them (*Alterbidinium, Chatangiella, Isabelidinium, Spinidinium*, and *Trihyrodinium*), all of which are of regular, sometimes frequent occurrence in the associations of the Transdanubian Range, the area of the northern margin of Tethys studied here belonged to the warm climate province.

The above-mentioned facts represent the evolution of dinoflagellate assemblages indicating a warm climate, of the areas coming moving toward an oceanic influence from a shallow marine, one from the end of the Santonian to the end of the Campanian; also indicated are the changes of the basin in time and space, as a function of the gradual south-west to north-east expansion of Tethys, in the area of the basin of the Transdanubian Range (Fig. 4).

Oceanic influence can only be encountered in the youngest associations of boreholes Ng-1, Mp-37, Pápa-2 and Vinár-1.

The rich dinoflagellate assemblages are connected with formations of relatively high organic material content, while in formations of high Ca content, no (or only a few) phytoplankton can be found.

Acknowledgements

I would like to express my thanks to F. Góczán for assigning to me the photo material on the Upper Cretaceous dinoflagellates of the Transdanubian Range, collected over many years, to H. Summersberger for revision of ammonite species, to D. J. Batten, J.-C. Foucher, K.-H. Kirsch and H. Leereveld for helping me to collect the necessary literature as well as to I. Bátori, I. Laky and E. Oszlánczi (Hungarian Geological Institute) for the conscientious laboratory work.

Plates

All magnification is 1000 X except for Pl. V. fig. 2. and Pl. VI. fig. 5. which are 500 x

Plate I

1–2. *Odontochitina operculata* (O. Wetzel 1933) Deflandre et Cookson 1955 1.: bh. Dv-4. 580.8 m, 2.: bh. Cell-1. 2579.5 m.
3. *Dinogymnium euclaense* Cookson et Eisenack 1970, bh. Ng-1. 1452.0 m.
4. *Dinogymnium acuminatum* Evitt et al. 1967, bh. Bj-528. 58.5 m
5. *Spiniferites ramosus* var. *granosus* Davey et Williams 1966, bh. Gy-9. 216.6 m
6. *Apteodinium deflandrei* (Clarke et Verdier 1967) Lucas-Clark 1987, bh. Dv-4. 644.2 m

Plate II

1, 4. *Tarsisphaeridium geminiporatum* Riegel 1974, 1: bh. Gy-9. 71.9 m, 4: bh. Dv-3. 598.0 m
2. *Odontochitina striatoperforata* Cookson et Eisenack 1962 bh. Gy-9. 95.7 m
3, 5. *Apteodinium deflandrei* (Clarke et Verdier 1967) Lucas-Clark 1987, 3: bh. Dv-4. 644.0 m, 5: bh. Gy-9. 268.0 m

Plate III

1. *Dinogymnium* cf. *euclaense* Cookson et Eisenack 1970, bh. Bj-528, 92.5 m
2. *Cribroperidinium* cf. *edwardsii* (Cookson et Eisenack 1958) Davey 1969, bh. Gy-9 196.0 m
3. *Spiniferites ramosus* var. *ramosus* (Davey et Williams 1966) Lentin et Williams 1973, bh. Gy-9. 233.2 m
4. *Canninginopsis denticulata* Cookson et Eisenack 1962, bh. Ng-1. 1368.0 m
5. *Trithyrodinium* cf. evitti Drugg 1967, bh. Mp-42. 386.2 m
6. *Odontochitina operculata* (O. Wetz. 1933) Deflandre et Cookson 1955, bh. Gy-9. 188.7 m

Plate IV

1, 6. *Pterodinium cingulatum* ssp. *conterminatus* Marnheinecke 1992, 1: bh. Ng-1. 1350.0 m, 6: bh. Ng-1. 1350.0 m
2, 5. *Cymatiosphaera eupeplos* (Valensi 1948) Deflandre 1954, bh. Mp-42. 405.7 m
3. *Odontochitina striatoperforata* Cookson et Eisenack 1962, bh. Gy-9. 76.5 m
4. *Dinogymnium digitus* (Deflandre 1935) Evitt et al. 1967, bh. Gy-9. 76.5 m

Plate V

1, 2. *Odontochitinopsis molesta* (Deflandre 1937) Eisenack 1961, 1: bh. Dv-4. 369.0 m, 2: bh. Ng-1. 1489.0 m(500x)
3. *Fromea amphora* Cookson et Eisenack 1958, bh. Ng-1. 1501.6 m
4. *Pterodinium* cf. *aliferum* Eisenack 1958, bh. Mp-42. 22.1 m
5. *Dinogymnium digitus* (Deflandre 1935) Evitt et al. 1967, bh. Bj-528. 63.5 m
6. *Odontochitina operculata* (Wetzel 1933) Deflandre et Cookson 1955, bh. Gy-9. 146.5 m
7. *Pterospermella australiensis* (Deflandre et Cookson 1955) Eisenack 1972, bh. Dv-4. 672.4 m

Plate VI

1, 2. *Pyxidinopsis bakonyensis* (Góczán 1962) Stover et Evitt 1978, 1: bh. Ng-1. 1393.0 m, 2: bh. Ng-1. 1358.0 m
3. *Palaeohystrichophora infusorioides* Deflandre 1935, bh. Gy-9, 244.3 m
4. *Nelsoniella aceras* Cookson et Eisenack 1960, bh. Gy-9. 169.0 m
5. *Cannosphaeropsis utinensis* O. Wetzel 1933, bh. Ng-1. 1434.0 m (500x)
6. *Spinidinium* sp., bh. Bj-528. 63.5 m

Plate VII

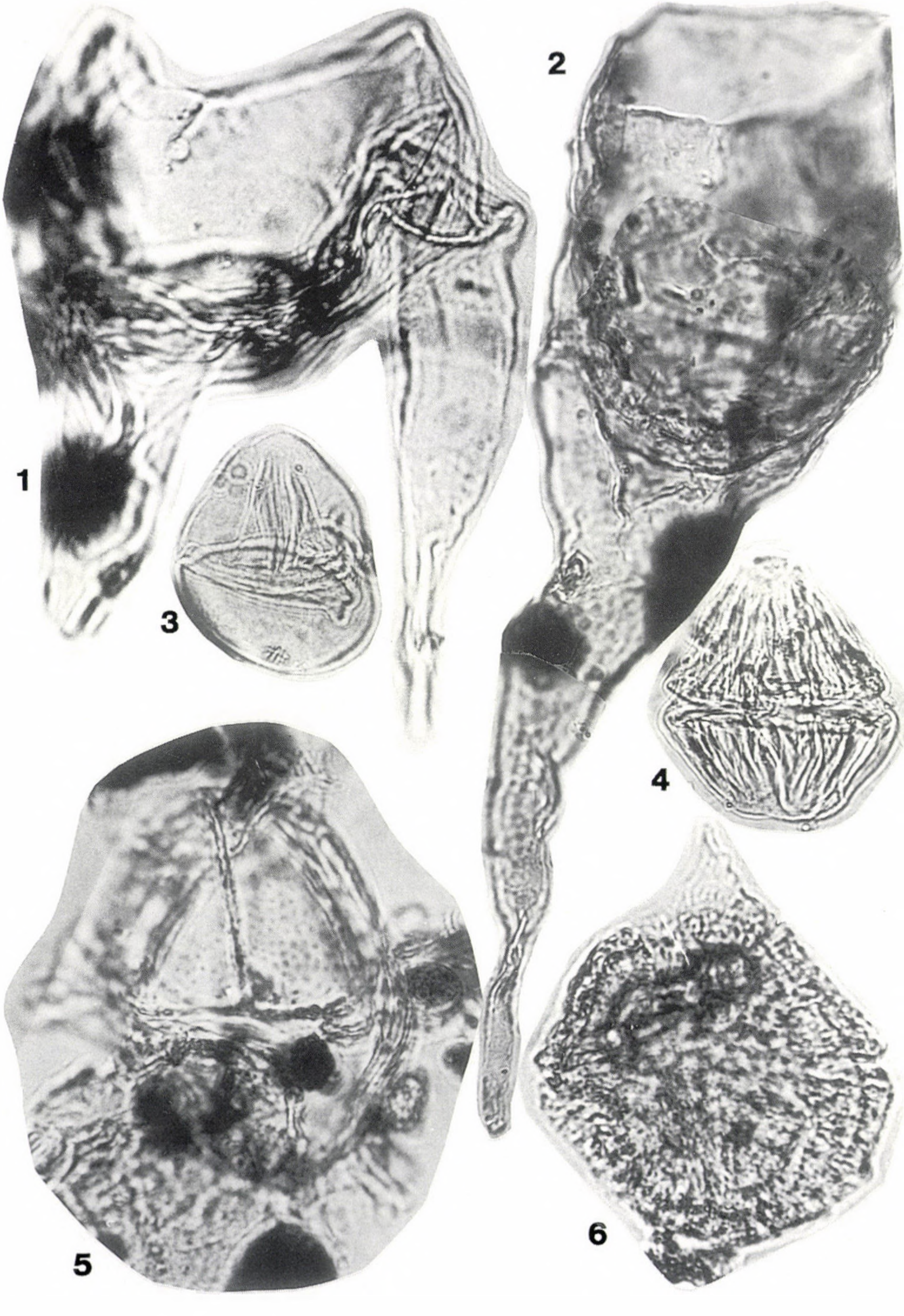
1. *Isabelidinium bakeri* (Deflandre et Cookson 1955) Lentin et Williams 1977, bh. Ng-1. 1337.7 m

2, 4,
8. *Manumiella ?cretacea* (Cookson 1956) Bujak et Davis 1983, 2: bh. Mp-42. 27,1 m, 4: bh. Mp-42. 30.1 m,8: bh. Ng-1. 1473 m
3. cf *Cladopyxidium seaptum* (Morgenroth 1968) Stover et Evitt 1978, bh. Dv-4. 699.2 m
5. *Manumiella* cf. *sealandica* (Lange 1969) Bujak et Davies 1983, bh. Mp-42. 27.1 m
6. *Dinogymnium euclaense* Cookson et Eisenack 1970, bh. Gy-9. 235.0 m
7. *Manumiella hungarica* n. sp. bh. Ng-1. 1473.0 m

Plate VIII

1–3. *Pachydiscus levyi* De Grossouvre 1894. Det: H. Summersberger 1995. Photo: I. Laky

Plate I



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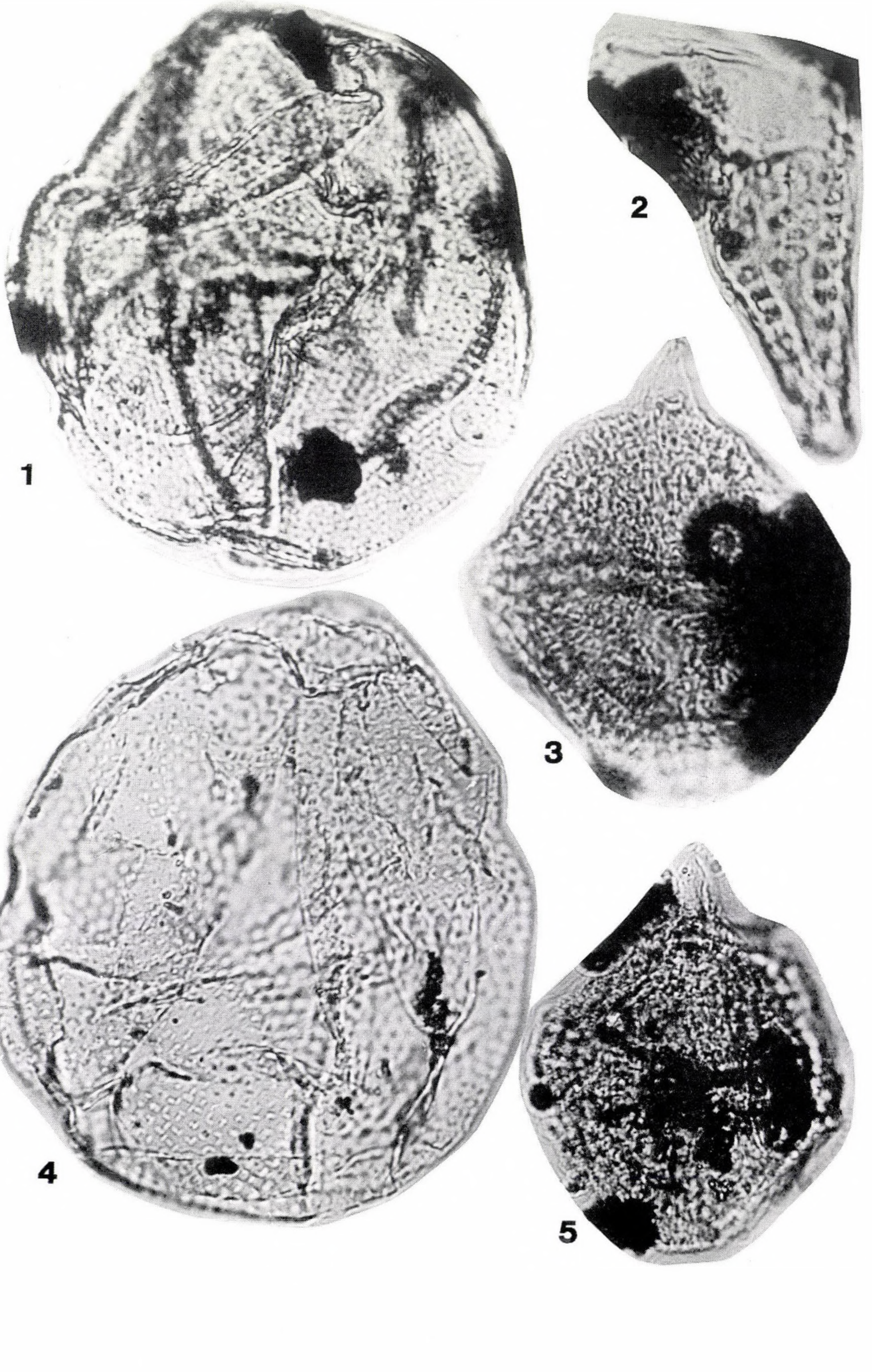

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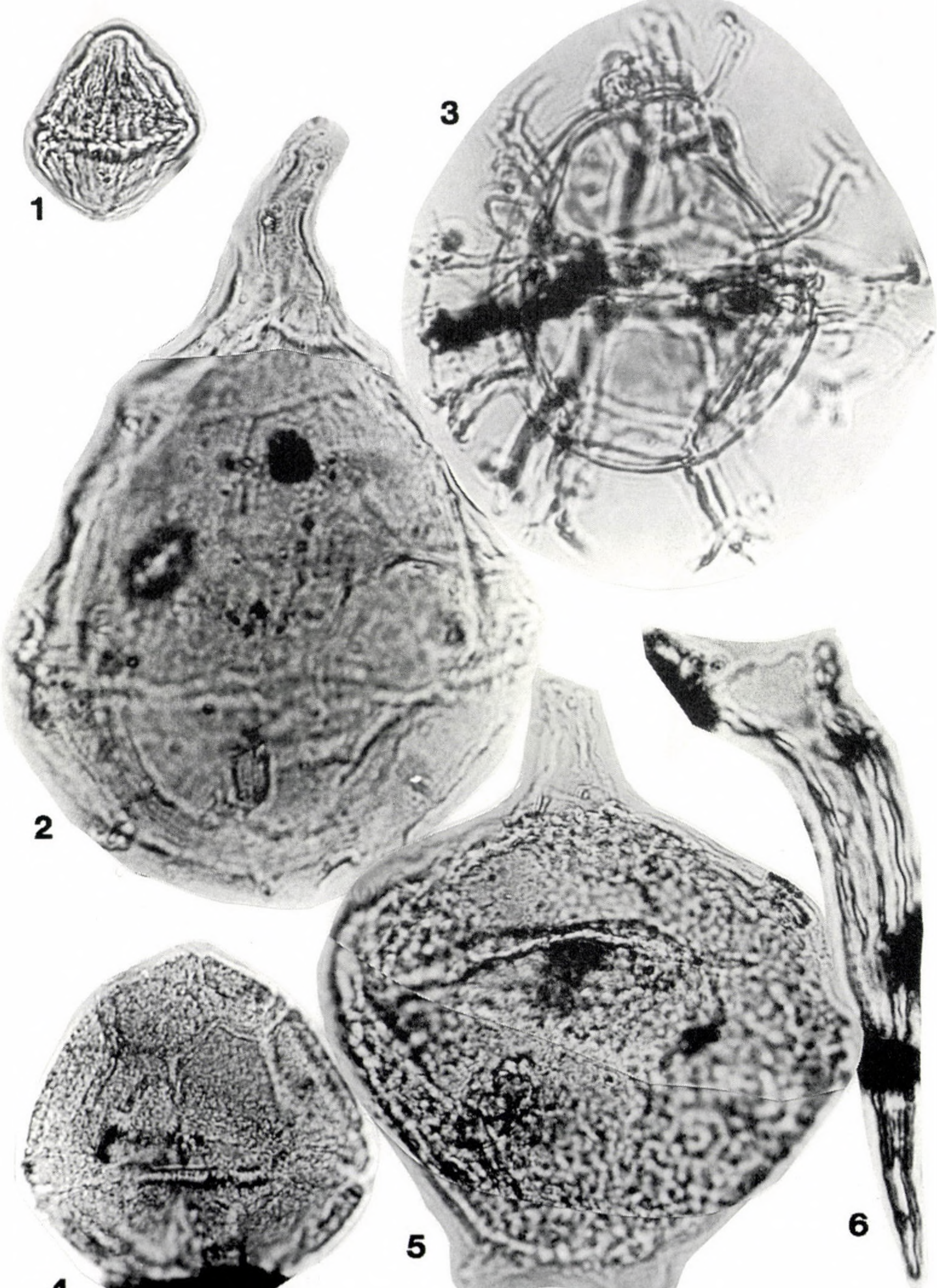

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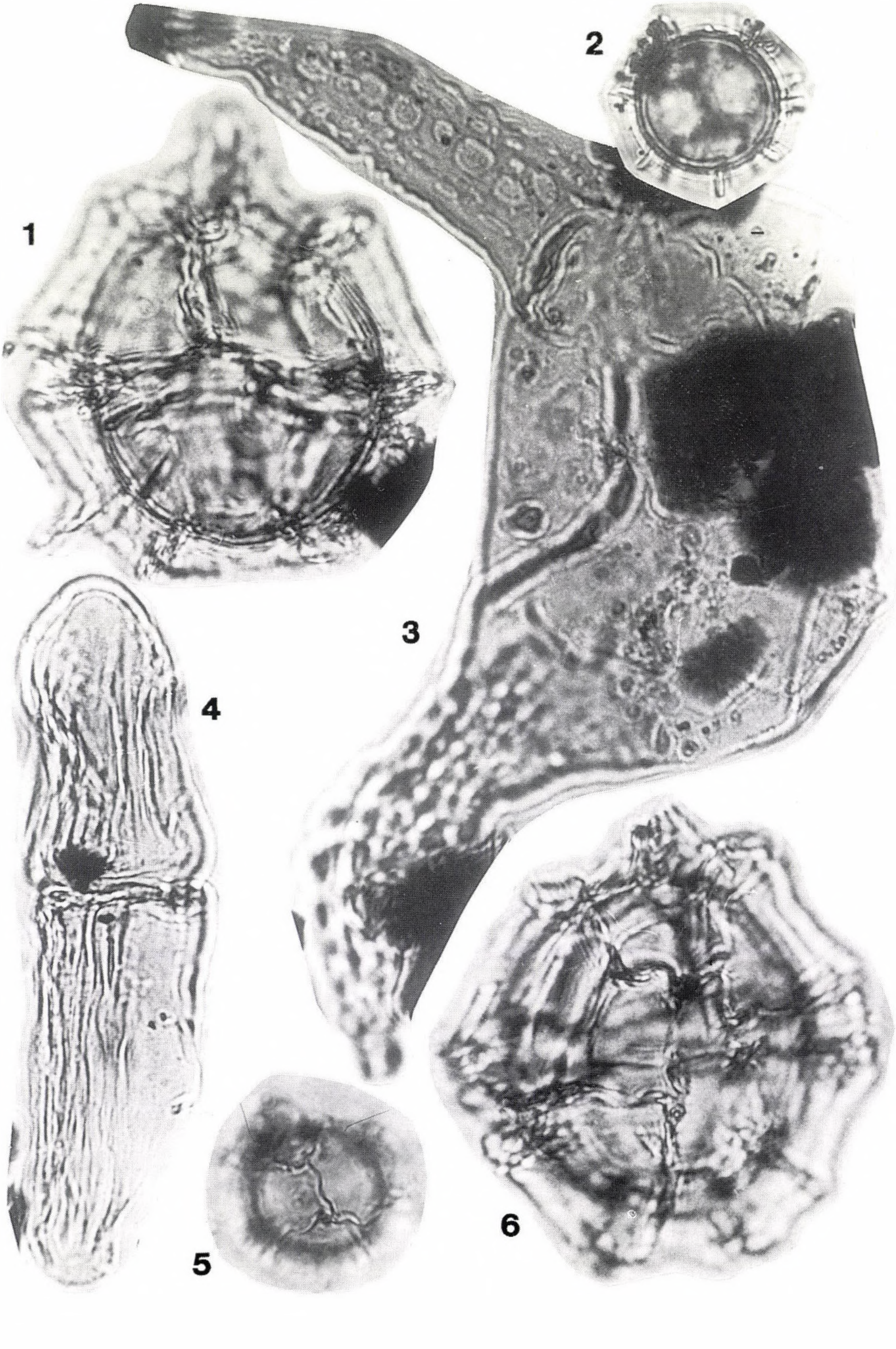


Plate V

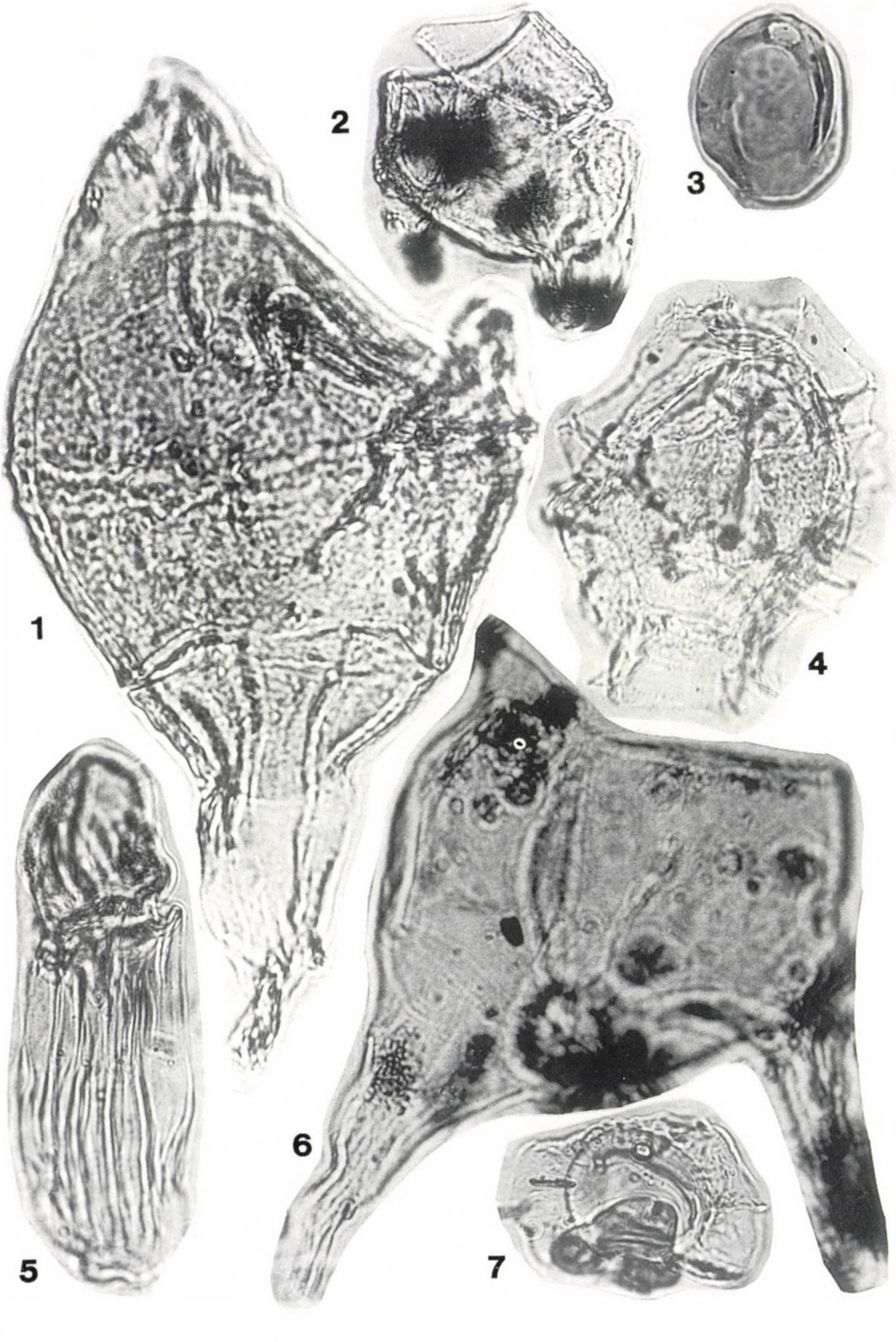


Plate VI

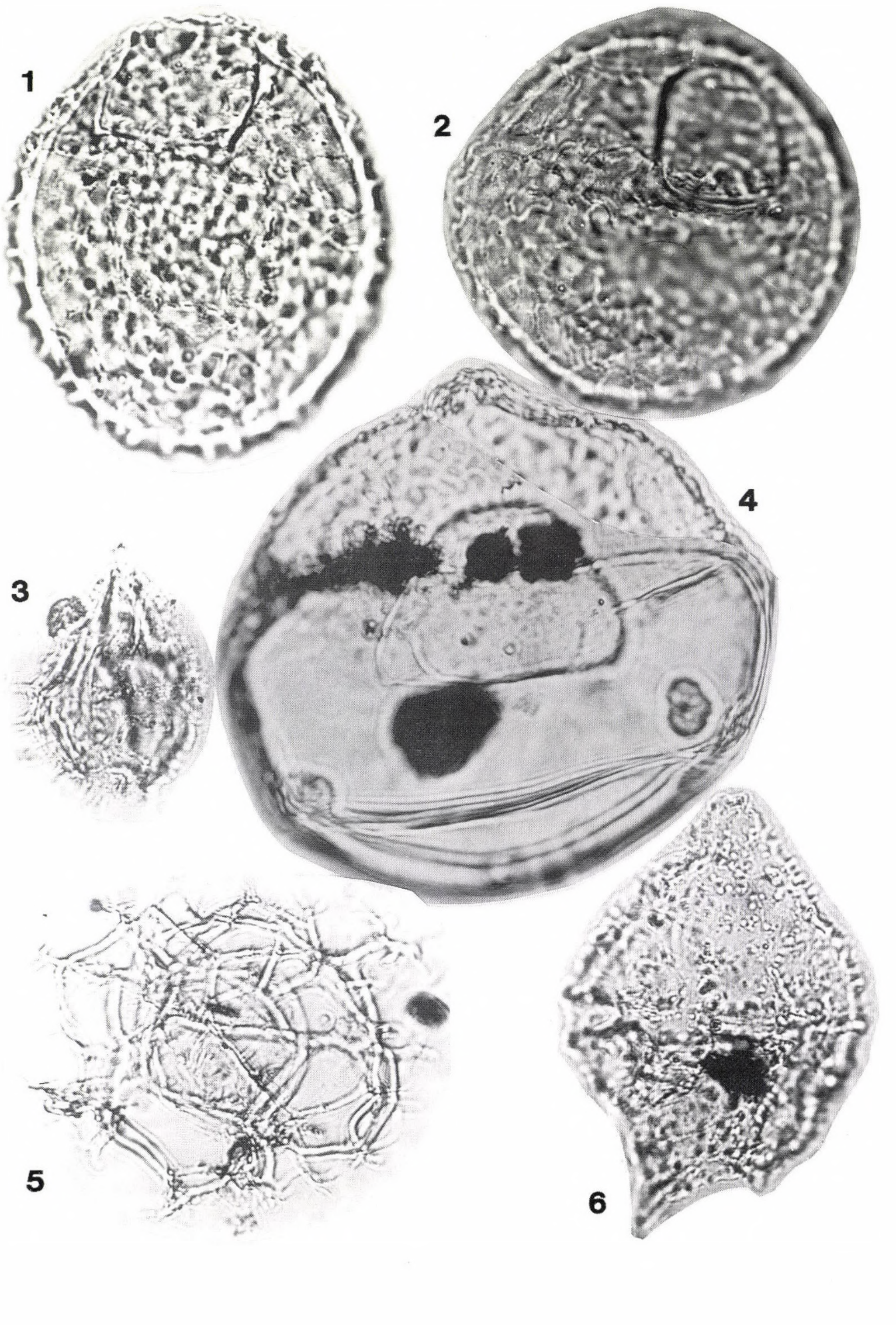
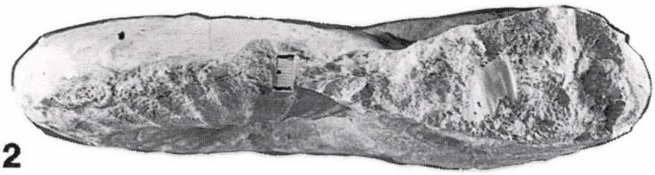


Plate VII

Plate VIII

1



2

3

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Climate curve construction from palynological data of the Pannonian section in borehole Berhida-3 (Hungary)

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In terms of mean annual temperature, a climate curve for the Late Miocene part of the Pannonian (s.l.) time was constructed from the pollen spectra found in borehole Berhida-3 (Hungary). The temperature for each paleo-assemblage was calculated as an expert-tuned weighted average of temperatures typical of the climate represented by the occurring sporomorphs groups. The curve correlates well with paleoclimate interpretations by other authors. At the same time it demonstrates important small-scale changes in the local climate during the Pannonian. The general trend is a slow decrease of temperature, i.e. in the Pannonian the climate was warmer in Hungary then today. The estimated difference is 2 °C (on average).

Key words: palaeoclimate, temperature, Pannonian palynology, weighted average, Hungary

Earlier climate reconstructions for the Late Miocene in Hungary based on palynology

The close connection between climate and vegetation is evident, although a precise description thereof requires extended scientific analyses. A much greater problem is to clarify the connections of palaeoclimate with palaeovegetation, as available data on both are sparse and indirect.

For the Neogene of Hungary, climatological interpretations appeared in connection with different research tasks. In the Upper Pannonian Lignite Research Project (Nagy 1958), temperature range charts were constructed for the determined pollen material where it could be matched with recent taxa. Climatological data were taken from the nearest meteorological station to their biotopes.

In the material of the Neogene of the Mecsek Mts. (Nagy 1969), from the percentages of tropical, subtropical and temperate climate taxa, a stacked percentage diagram was made in order to understand the palaeoclimate.

In studies on the Neogene of Hungary (Nagy 1990, 1992), a joint plot of three curves representing the absolute amounts of sporomorphs indicating tropical, subtropical, or temperate climate was made. Changes of the curves were interpreted as climatic changes during the Hungarian Neogene.

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Present material

Berhida-3 was drilled in 1986 as a borehole of key importance intended to explore a Miocene lignite basin in western Hungary (Fig. 1). The borehole, reaching a depth of 690 m, cut through a number of formations representing a period of time ranging from the Permian up to the Holocene. A detailed geologic study on the core was presented by Kókay et al. (1991).

In borehole Berhida-3 the Pannonian (s.l.) is represented only by its Upper Miocene part. From the recently investigated paleontological material of this Pannonian (s.l.) section, the reconstruction of palaeoclimate is also based on palynology. The structure of the sampled section in terms of lithostratigraphic units, some short lithologic descriptions and the numbers of samples taken from the formations drilled are given in Table 1.

The samples were treated by traditional extraction methods, using HCl, HF, and a heavy liquid ($ZnCl_2$ at 2.2%). The determination and counting of sporomorphs were performed under light microscope on slides with cover glass of 20x20 mm^2. Altogether 200 samples were examined, 97 of which contained sporomorphs.

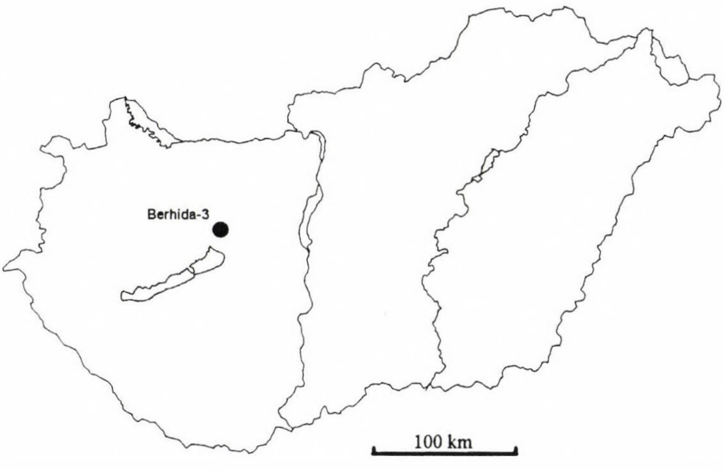

Fig. 1
Location of borehole Berhida-3 (in Hungary)

Quantitative data evaluation

The quantitative data processing aimed at revealing the climate change during the Pannonian as represented by the section studied. The climate was described in terms of present climatic belts characterized by mean annual temperature. The mean annual temperatures of the relevant climate types used in calculations were 10 °C for the temperate, 18 °C for the subtropical, and 25 °C for the tropical climate (Walter and Lieth 1960).

Table 1
Concise description of the Pannonian (s.l.) part of borehole Berhida-3 (after Kókay et al. 1991) with the distribution of samples

Depth interval (m) and age data	Formation	Lithology	No. of samples taken	No. of samples containing sporomorphs
30.6–5.5	Nagyvázsony Limestone	calcareous marl, limestone, lime-marl	9	1
114.6–30.6	Tihany	silty clay-marl, silt with fine sand, limonitic clay, clayey silt, calcareous sandstone, sand	98	34
153.8*–114.6 *10–8 My B.P.	Somló	clay-marl with Limnocardium, clay-marl with fine sand, silty fine sand, silty clay-marl, sand	36	33
167.9–153.8	Csór Silt	silty fine sand, silty marl, clay-marl with shells of molluscs and ostracods	7	7
190.3–167.9	Csákvár Clay-marl	marl, silty clay-marl, silty clay, lime-marl, coquina, silt, calcareous sand, gravel	23	20
222.6*–190.3 *12.3 My B.P.	Ősi Variegated Clay	dacit tuff, limonite-spotted lime-marl (calcareous marl), silty clay-marl with shells of molluscs and ostracods	27	2

Essentially, the calculations correspond to a formalized expert estimation of the probable temperature based on the analyzed taxon composition. For the calculations the taxon composition was defined as follows. The determined sporomorphs were evaluated from the climatic point of view and divided into three main types: taxa indicating tropical, subtropical, or temperate climate. Within the temperate-climate taxa *Pinuspollenites labdacus* and *Sparganiumpollenites* were separated as special subgroups. The total specimen abundances of these five categories were taken as input data for the calculations (Table 2). Taxa known or supposed to be insensible to climate were excluded from the work database.

In each sample, the temperature was estimated as the weighted average of temperatures of the given climate types. The weights were proportional to the taxon abundances in the categories of tropical-climate and subtropical-climate fossils. Within the temperate-climate taxa, *Pinuspollenites labdacus* and *Sparganiumpollenites* were considered to be over-represented from the point of view of their climate-indicating importance. Therefore, their actual numbers were reduced for calculations to one-fifth and one-third, respectively, based on an expert judgement. As a result the following formula was used:

$$T_{est} = \frac{A_{trop}\ T_{trop} + A_{subt}\ T_{subt} + (A_{temp} - 4/5\ A_{Pinu} - 2/3\ A_{Spar})\ T_{temp}}{A_{trop} + A_{subt} + A_{temp} - 4/5\ A_{Pinu} - 2/3\ A_{Spar}}$$

Table 2
Sporomorphs abundances used in calculations (for explanation see text)

	number of specimens						number of specimens				
depth, m	tropical	subtropical	temperate	Pinuspollenites labdacus	Sparganium-pollenites	depth, m	tropical	subtropical	temperate	Pinuspollenites labdacus	Sparganium-pollenites
24.90			1			135.20	5	87	373	170	
36.10			3			136.30	1	16	117		
42.70			1			136.80		14	71		
44.80			1			137.20	1	14	114		
45.60			3			138.20		4	47		
47.10			2			139.30		8	97		
53.10			1			141.00	2	18	248		79
53.20			1			142.10	2	2	88		
54.00			2			142.90	2	16	99		
54.90			1			143.10	3	5	82		
54.95			3			144.10	2	12	163		31
55.60			1			144.90	2	15	174	51	
55.90			1			146.00	1	20	225	76	
58.80		3	23			147.80	3	11	146		
64.40			1			148.70	3	12	102	42	
68.40			1			149.20	3	20	280	84	70
76.20			2			150.10		10	40		
80.50			25			152.20	1	4	68		
82.70		87	447	163		153.10	4	25	145	57	
83.20		55	354	235		154.20		13	42		
84.80		17	111	58		155.00	1	9	110	64	
85.40	2	31	151			156.10		10	25		
89.40			1			157.10		2	18		
92.70		60	260	99		157.30		2	8		
93.30		41	192	76		165.20		8	33		
94.30			2			167.10	4	129	356	163	
94.40		1	5			168.10	1	23	147	56	
95.40			3			169.10	3	38	130	60	
97.30			12			170.90	2	3	48		
98.80			17			171.60		14	83		
109.90		8	94			172.30		6	43		
110.40	1	39	226		92	175.40		8	19		
111.90			5			176.60		6	25		
112.80		4	30			177.40	1	1	2		
113.20	4	68	479	108		177.60	1	2	61		
120.00		14	56			178.60		1	29		
121.20	4	61	403	170		179.10		3	8		
121.40		9	88			180.10	1	4	12		
122.20	4	96	743	297		181.10	5	6	25		
124.10	5	42	348	81		182.50		1	1		
124.80	1	54	354	115		182.80		1	1		
125.80		6	55			185.90	1	1	17		
126.80	1	10	166			187.70			1		
128.20		3	32			187.80		155	589	500	
129.00		70	587	350		188.00	4	55	128	88	
129.80	5	150	1051	297		189.50			3		
130.20	4	137	819	288		191.50			1		
132.00	4	90	500	285		192.60		4	30		
133.10	2	96	410	161							

where T_{est} is the estimated temperature, A_{trop}, A_{subt}, A_{temp}, A_{Pinu} and A_{Spar} are the total specimen abundance of the tropical, subtropical, temperate, (and within the latter) *Pinuspollenites labdacus* and *Sparganiumpollenites* taxa, respectively, T_{trop}, T_{subt} and T_{temp} are the mean annual temperature of the respective climate type (see above).

Although the climate curve calculated with the above formula (Fig. 2a) is based on factual data and hence ready for comparison, correlation, etc., due to the uneven sampling and to lithological reasons it is rather erratic. For a less detailed correlation a smoothed version thereof (Fig. 2b) might be more appropriate. In order to give the climate change trend in the simplest terms, the straight line fitted to the data (Fig. 2c) can be used.

Palynological interpretation

The calculated curves picture the local palaeoclimatological conditions characteristic for the Pannonian part of borehole Berhida-3. In general, the curve is well comparable with the presumed Pannonian climatic conditions (Kázmér 1990; Pantić 1990; Planderová 1990; New Enc. Brit. 1993). The general trend is a slow decrease of temperature, i.e. in the Pannonian the climate was warmer in Hungary then today (Fig. 2c). Comparing this curve with the taxa of the pollen spectra we find the following:

– The Ősi Formation is very poor in fossilized sporomorphs. The assemblage found at the depth of 192.6 m indicates a warm-temperate climate. Elements of subtropical origin (*Cedrus*, Oleacea) are sparse.

– In the Csákvár Formation the palaeovegetation represents a warm-temperate climate close to the temperate-subtropical transition. Most of the taxa require a warm-temperate climate, but many of them are of subtropical origin. Leafy trees are significant. Some are insect-pollinated species (*Symplocos*, Oleaceae, *Rhus*). Consequently, their pollen production is lower, and their presence has a signal value. There are also a few tropical elements; they must have been chiefly an undergrowth, e.g. the *Sabal* palm from the *Taxodium* swamp forest.

– In the Csór Silt Formation the curve shows a gradual decrease which is the result of an increase in temperate elements. The climate could also have been equivalent to warm-temperate with some marine influence (as indicated by plankton organisms).

– In the Somló Formation the tendency of the curve is warm to temperate, due to the large quantity of hillside conifers. In intermediate times we always find tropical elements. In the lower part of the formation forms of the Sapotaceae family occur regularly, although in low abundances. At a depth of about 140 m *Reevesia* pollen grains (Sterculiaceae) appear and they occur as far as the top of the formation. In the unit some *Symplocos*, *Magnolia* and pollen grains of Araliaceae also support the idea of a warmer climate. The subtropical influence is also verified by the *Taxodium* swamps.

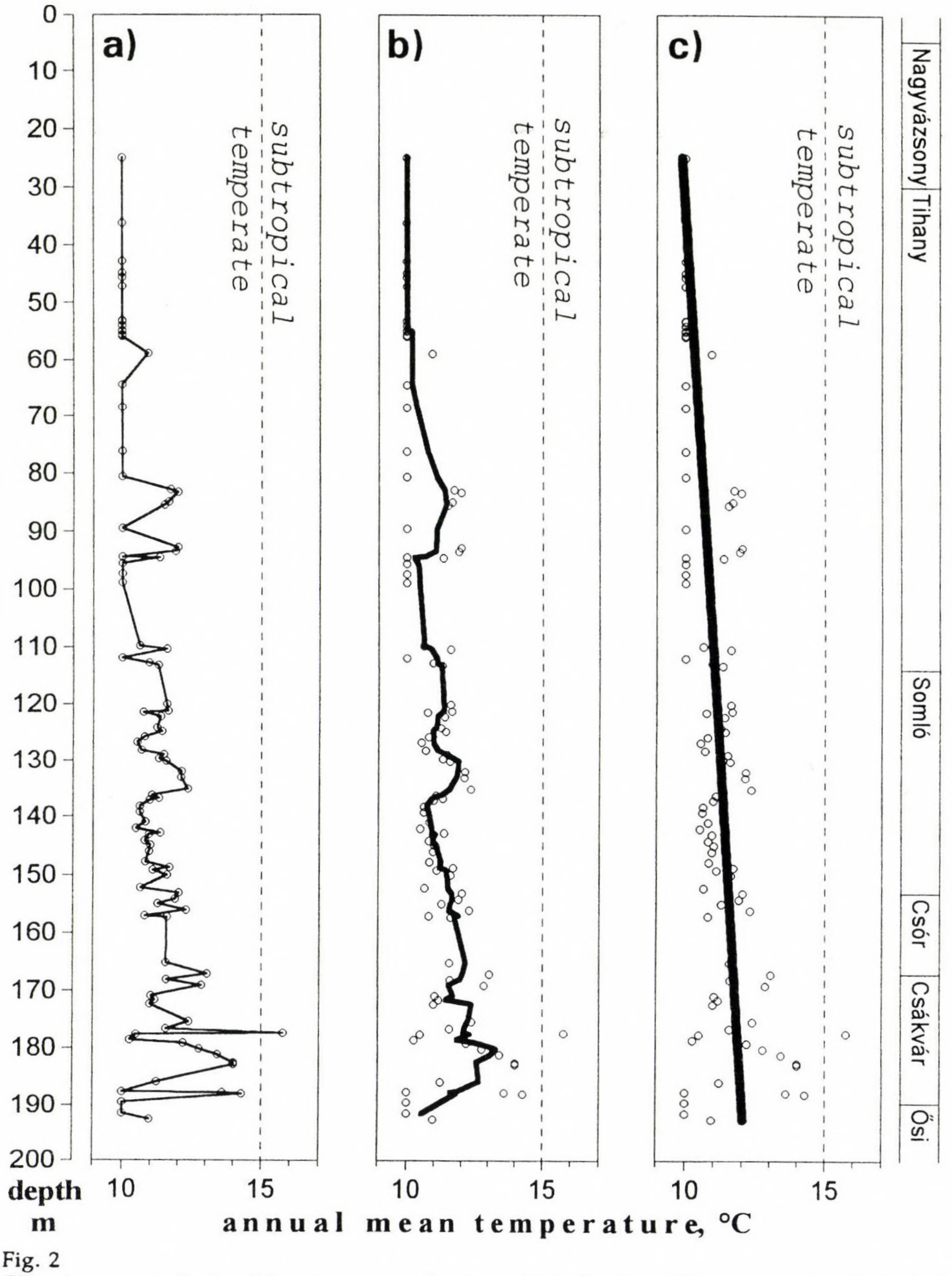

Fig. 2

Climate curves calculated from sporomorphs data with indication of the temperate–subtropical climate boundary, and the lithostratigraphic formations concerned. a) Raw curve connecting data points; b) Smoothed curve showing principal changes; c) Linear trend in the data

– From the Tihany Formation only a few samples could be evaluated for lithological reasons (the samples are limonitic, or composed of limestone or fine sand). The bottom of the formation is a continuation of the Somló Formation from a palaeobotanical and palaeoclimatological point of view. Though the *Reevesia* pollen grains vanish, the *Symplocos* and *Taxodium* swamp-forest with the *Sabal* palm indicate the warmer biotopes. The sample from 109.9 m has a drier character certified by a decrease in the number of fresh-water plants and the appearance of *Ephedra*. The samples from 93.3 and 92.7 m prove an eastern Mediterranean character of the climate with *Cedrus, Zelkova,* Oleaceae and *Rhus* pollen paleo-assemblages. The samples from 85.4, 84.8, and 83.2 m demonstrate the equalizing effects of the inland sea (indicated by dinoflagellates), and an eastern Mediterranean character as well. The last two samples include no more tropical elements. The warm-temperate climate is marked by the *Taxodium* swamp-forest, the eastern Mediterranean influence by *Cedrus* and *Zelkova* pollen grains, and the Southeast European influence is registered by *Pinus omorica*.

– From among the samples of the Nagyvázsony Limestone Formation the only productive sample (24.9 m) contains the pollen of *Pinus silvestris*.

Conclusions

1. Despite the lithological constraints and the resulting uneven sampling, the palynological material is informative and interpretable.

2. The applied weighted average technique is an efficient tool in synthesizing a large amount of quantitative palynological data for a simple climate reconstruction. It assures a uniform (and hence objective) manner of data evaluation along a section, but also may incorporate expert's judgements concerning the informativeness of special sporomorphs groups.

3. The climate curve obtained or one of its smoothed versions can be conveniently used for comparison with similar representations from other boreholes or areas.

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Anniversaires

Sándor Koch Centennial held in Szeged

The last great representative of "classic" mineralogy, professor of Attila József University in Szeged, founder of the Acta Mineralogica Petrographica (University of Szeged), "honoris causa" member of the Hungarian Geological Society and member of numerous scientific associations abroad, Sándor Koch was born one hundred years ago. This anniversary was celebrated in the form of a two-day session held in the University and in the "House of Science and Technology" of Csongrád County in mid-October, with the participation of the Koch family.

The first part of the session was taken up by commemorative speeches lauding Sándor Koch as a scientist, a teacher, a university leader, a mineralophile, and a propagator of popular science, respectively, in the presence of over one hundred warm admirers, given in the festive hall of Attila József University. Silver coins and bronze plaquettes with Koch's profile were awarded to the family and the University by the Sándor Koch Foundation. Following these presentations, a marble memorial tablet with Koch's image in bronze was unveiled at the entrance of the Department of Mineralogy, Geochemistry and Petrology. The final event of this day was the opening of the "Sándor Koch Mineral Collection", the contents of which had been reclassified.

On the next day, an impressive range of scientific lectures was delivered on the topics of mineralogy, petrology, geochemistry and raw material prospecting, chosen from among favourite subjects of Sándor Koch's scientific activity.

This heartfelt centennial programme was organized by the Department of Mineralogy, Geochemistry and Petrology of Attila József University (Szeged), the Great Hungarian Plain Branch of the Hungarian Geological Society, the "Sándor Koch Popular Science Association" (TIT) of Csongrád County, and the "Sándor Koch Foundation".

Tibor Szederkényi

Centenary Celebration of Professor Miklós Vendel

Born in the West Hungarian town of Sopron (8 October 1896), M. Vendel was Professor of Mineralogy, Geology and Mineral Deposits at the Academy (later University) of Mining, Metallurgy and Forestry of Sopron from 1923 through 1959.

His scientific research activity encompassed a wide range of subjects, from mineralogy and igneous petrography through regional geology (mainly of the surroundings of Sopron and the easternmost edge of the Alpine crystalline belt) to ore geology and hydrogeology. Moreover, he was one of the pioneers of geochemistry in Hungary.

An ordinary member of the Hungarian Academy of Sciences, Professor Vendel was awarded an impressive array of scientific honours, including the highest one, the Kossuth Prize, honorary membership of the Hungarian Geological Society, the Austrian Geological Society and the Austrian Mineralogical Society, as well as the honorary freemanship of his beloved native city of Sopron.

On 16 October 1996 a commemorative session was organized by the Sopron municipality, the University, and a number of scientific societies and institutions. It was attended by 110 people, including six from Austria. A bronze memorial tablet was unveiled at the house M. Vendel had lived in, and flowers were deposited on his tomb (he passed away in 1977).

The honours were acknowledged with thanks by Professor Vendel's family which had authorized silver and bronze commemorative medals to be made for this occasion.

Endre Dudich

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GUIDELINES FOR AUTHORS

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

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In North Hungary east of the Danube, small outcrops of the Mesozoic basement are known. They are generally considered to belong to the Transdanubian Range Unit. The basement block in the vicinity of Csővár is made up of Triassic carbonates of platform and basin facies, respectively. The 1200 m deep Csővár (Csv)-1 well exposed a significant part of the basin facies, and entered into strongly tectonized dolomites beneath it. The succession exposed by the Pokolvölgy quarry completed the sequence upward and offered an excellent opportunity for detailed sedimentological observations.

Re-investigation of the Csv-1 core led to the conclusion that the approximately 600 m thick lower part of the sequence, consisting of dolomites inserted by thin slices of extremely varied lithology and age (Triassic to Cretaceous) is not a part of the normal stratigraphic succession.

The upper part of the Csv-1 core and the section of the Pokolvölgy quarry exposed an Upper Carnian–Rhaetian succession of basin and toe-of-slope facies. The latter is indicated by gravity-flow deposits and redeposited fossils of carbonate platform origin, and in the large amount of remnants of Rhaetian terrestrial plants.

Key words: stratigraphy, sedimentology, basin facies, toe-of-slope facies, foraminifera, sporomorphs, organic microfacies, Upper Triassic, North Hungary

Introduction

In North Hungary, east of the Danube, in the vicinity of the town of Vác, small, fault-bounded, basement blocks are located (Fig. 1). They are made up of Upper Triassic platform carbonates, akin to those in the Buda Hills. However, in the Csővár block, located between Csővár and Nézsa, thin-bedded cherty limestones also appear. Problems of stratigraphic setting, facies interpretation and of the relationship of this formation with the surrounding platform carbonates have attracted the interest of geologists for a long time. The Csővár

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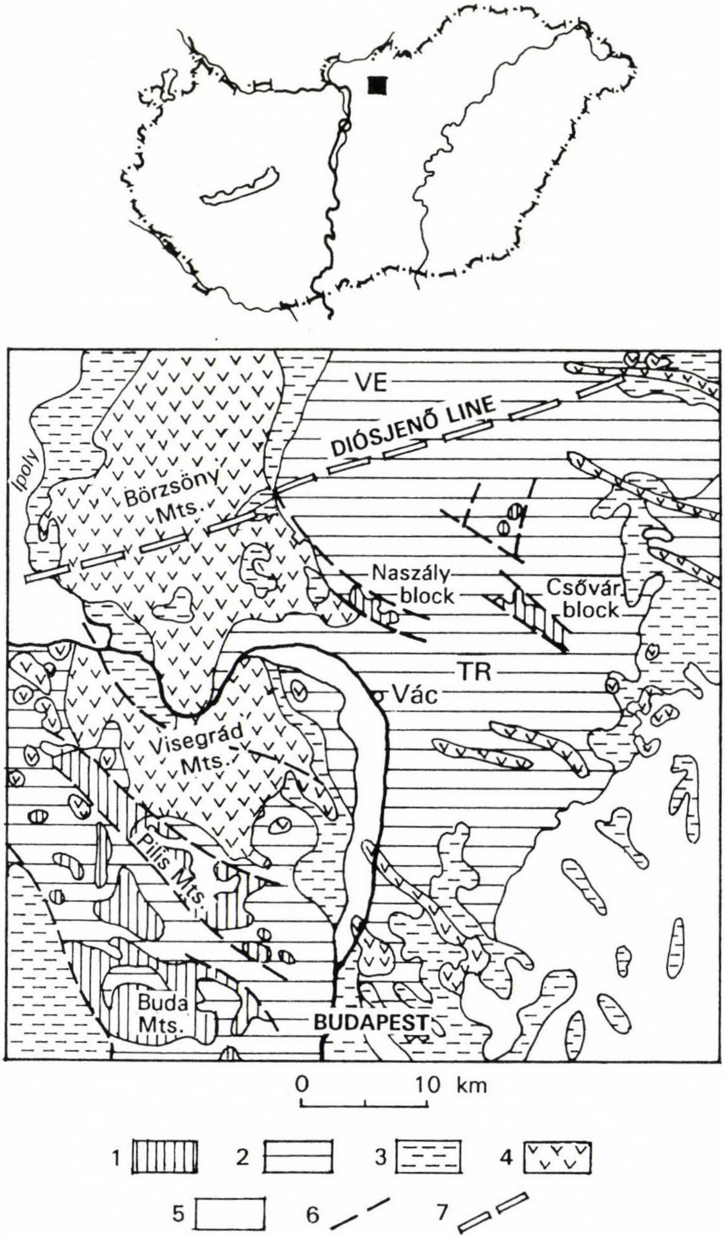


Fig. 1
Geologic map showing the location of the Csővár block (after Fülöp et al. 1984). 1. Mesozoic formation (mainly Upper Triassic carbonates); 2. Paleogene formations; 3. Neogene sedimentary formations; 4. Neogene volcanic rocks; 5. Quaternary formations; 6. faults; 7. major fault-zone (lineament) in the basement; TR – Transdanubian Range Unit; VE – Vepor Unit

block is the easternmost outcrop of the Mesozoic of the Transdanubian Range and, according to our present-day concept it may have been originally located near the shelf margin, in the most internal part of the Transdanubian Range segment of the Tethys shelf.

In 1968, in order to better understand the stratigraphy and structural setting of the area, a continuous core well was drilled by the Mecsek Ore Company (MÉV) in the Pokol (or Kecskés) valley, where the best outcrops and quarry exposures of the cherty limestones were located. The 1200 m deep Csővár Csv-1 borehole traversed the cherty limestone formation reaching dolomites below it. Although Csv-1 core yielded extremely valuable information on the Triassic formations making up the Csővár block, it seemed plausible that with the advances in the fields of micropaleontology and sedimentology since the early 70s, a re-investigation of the core, supplemented by a detailed study of the quarry section in the Pokol Valley, could provide a significant contribution to the knowledge of this sequence of critical importance.

The aim of the present paper is to present and discuss the results of the re-investigations carried out in the last couple of years, and to draw conclusions on the Triassic paleogeography of the region.

Previous studies

József Szabó (1860) first described Mesozoic formations in the study area. He mentioned Liassic (?) "brown shales" from Csővár, considering them the oldest Mesozoic formation in the area. It is worth mentioning that he found an ammonite imprint in these layers.

In the report on his mapping activity, Stache (1868) described dark and light grey and brownish limestones from Csővár, and assigned them a Jurassic age.

In 1908, in a session of the Hungarian Geological Society, Vadász classified the dark cherty limestones at Csővár into the Triassic. In the discussion on his report, Schafarzik mentioned a *Harpoceras* which had been found in Csővár, which made him support a Jurassic age for the formation.

In his monograph on the Mesozoic basement blocks east of the Danube, Vadász (1910) presented a detailed description of the "Upper Triassic grey cherty limestone" formation. He also described a fossil assemblage collected from Csővár: a few foraminifera, crinoids, brachiopods, a relatively large number of bivalve species, gastropods, and a few ammonites.

Based on these fossils, he classified the formation into "the lower part of the Raibl horizon of the Carnian stage", assuming that it is overlain by the Norian Main Dolomite ("ődolomit") and, above it, the Rhaetian Dachstein Limestone. He considered the dolomites cropping out in the Csővár block (on Vas Hill) as part of the Main Dolomite.

Oravecz (1962, 1963) supposed the stratigraphic position of dolomites of the Vas Hill to be beneath the Carnian cherty limestones. He also proposed a gradual lateral transition between the cherty limestones and the Dachstein Limestone.

The Csv-1 well was drilled in 1968–69. The upper part of the core was documented by Áron Jámbor, geologist of the Mecsek Ore Company. Recognising the importance of the core from the point of view of Triassic stratigraphy, he requested Elemér Nagy, a geologist of the Hungarian Geological Survey, to take part in the work. Subsequently, Nagy took over the description of the core. Later on, Csaba Detre was also involved into the documentation. He took the samples for the laboratory tests and identified the megafossils found in the core.

Based on new fossil findings in the outcrops and the Csv-1 core, Detre (1969, 1970, 1971) re-evaluated the stratigraphy of the Mesozoic rocks of the Csővár block. From the fossils, collected in the Pokolvölgy quarry, he determined ammonites (*Badiotites eryx* (Münster), *Apleuroceras* sp. ex. gr. *sturi* (Mojs.) and an Orthoceratid form called *Michelinoceras* cf. *politum* (Klipstein)) which seemed to support the Carnian age determination of Vadász. According to Detre's opinion, bituminous, dolomitic limestones at the foot of Vár Hill constitute the oldest member of the succession, which was emplaced in the upper part of the Lower Carnian. They gradually pass upward into bituminous, cherty, dolomitic limestones. Reef limestones at the northern part of the Csővár block were classified by him as Upper Carnian in age.

In the Csv-1 core, between 347–352 m, Halobias were found. They were determined by Detre (1971) as *Halobia styriaca* Mojsisovics and considered as indicators of the lower part of the Carnian. At 564.3 m, a massive occurrence of *Daonella pichleri* Mojsisovics and *Halobia cassiana* (Mojsisovics) was reported by Detre which was thought to indicate the uppermost Ladinian or the Ladinian/Carnian boundary interval.

In 1970, Detre requested H. Kozur to make micropaleontological investigations on a few samples from the *Halobia*-bearing interval of the Csv-1 core. Later on, these investigations were complemented by field observations and microfacies studies of the outcrops and quarries at Csővár. Kozur and Mostler (1973) reported graded bedding and allochthonous biogenic detritus of reef origin in the carbonate sequence and concluded that these layers were deposited in a near-reef basin.

In contrast to Detre's opinion, based on foraminifera, holothurian sclerites and conodonts, they classified the quarry section in the Pokol (Kecskés) valley as uppermost Norian (Upper Sevatian). On the other hand they found conodonts in the *Halobia styriaca*-bearing interval of the Csv-1 core (352.4 m). The association with *Epigondolella mostleri* Kozur and *Epigondolella diebeli diebeli* (Kozur et Mostler) constrained this segment of the core section to a Lower Carnian (Cordevolian) age.

Influenced by the results of the microfossil studies of Kozur and Mostler (1973), Detre (1981) re-evaluated his previous statements on the age of the sequence. *Clionitites nicetae* Diener, found by Kozur and identified by Krystyn, appeared to support the Upper Norian age determination of the sequence of the Pokolvölgy quarry. In this paper, however, *Halobia styriaca* Mojsisovics was declared as a characteristic species of the lowermost Norian, although this statement was incompatible with Kozur and Mostler's (1973) Cordevolian conodonts from the same layer.

According to Detre's proposal, the entire sequence exposed by the Csv-1 core and the quarry comprised the total Carnian and Norian interval.

In 1981, Balogh summarized the available stratigraphic data on the Csővár block. He proposed a terminology for the recognised lithostratigraphic units: "Vas hegy Hauptdolomite" – for dolomites exposed on the Vas Hill; "Pokolvölgy Dolomite" – for cherty dolomites and dolomites exposed in the lower part of the Csv-1 core and "Csővár Limestone Formation" – for cherty limestones of basinal facies exposed in outcrops and quarries in the surroundings of Csővár and in the upper part of the Csv-1 core (above 522 m).

In 1986, specimens of *Choristoceras nobile* (Mojsisovics) were found by Herman in the Pokolvölgy quarry, and based on this finding Detre et al. (1986) suggested that the succession passes up into the Rhaetian.

Kozur and Mock (1991) found conodonts indicative of the Upper Rhaetian in the samples taken from the Pokolvölgy quarry. They also described a new conodont species (*Neohindeodella detrei*) from the topmost part of the quarry, noting that the layers yielding these fossils had probably already been formed in the Liassic.

Initially, Kozur and Mock (1991), and then Kozur (1993), proposed to divide the Csővár Limestone Formation into two formations. The lower unit, consisting of dark, bituminous, often resedimented and graded limestones, cherty limestones and marls would be the "Csővár Limestone Formation", whereas the name "Várhegy Cherty Limestone Formation" was proposed for the upper unit, consisting of bedded light-yellowish to light-brownish micritic limestones and cherty limestones. The former unit was classified by Kozur (1993) as Norian to Upper Rhaetian, and the latter one as Hettangian and Sinemurian. He declared the conodont species *Neohindeodella detrei* Kozur et Mock, found in the basal part of the "Várhegy Formation" to be lowermost Jurassic (Hettangian), i.e. the youngest conodont species in the world.

Direct antecedents of the present work

In 1988, J. Haas and J. Oravecz studied the conserved samples of the Csv-1 core and took samples for micropaleontological and sedimentological investigations. Micropaleontological investigation of these samples was carried out by Oravecz-Scheffer (foraminifera) and Góczán (sporomorphs), the microfacies studies were made by Haas and Tardy-Filácz.

The thin-section studies led to a surprising result, in that within the 600 m-thick lower part of the cored section, consisting predominantly of dolomites, insertions made up of very strange and heterogeneous rock-types of various ages occur. This segment of the core being in tectonic contact with the upper parts, does not fit into the normal stratigraphic succession. Therefore, the characteristics of dolomites and inserted rock-types exposed in the lower part of the Csv-1 core are discussed in a separate paper of the present volume.

To complement the core section, Haas and Dosztály carried out detailed measurements and sampling of sections in the Pokolvölgy quarry in 1985.

Lithology and facies

Located above the lower dolomite unit, the 522.0–622.0 m interval of the Csv-1 core is made up of dark-grey, brownish-grey, thin-bedded, usually bituminous and cherty dolomites (Figs 3, 4). Small intraclasts, 0.2–1 cm in size, could also be observed.

Although both the upper and the lower boundaries of the unit are tectonically controlled, at the upper boundary the limestone-dolomite transition appears to be gradual.

According to the thin-section studies the characteristic texture type is xenotopic – A dolosparite with fine to medium crystal-size (25–50 μm). Relict texture elements are common. Vague lamination, intraclastic texture and moulds of radiolarians and rarely filaments were observed. These relict features refer to pelagic basinal facies, and suggest that precursor carbonates of this unit do not differ significantly from limestones in the lower part of the overlying limestone unit. Consequently this unit fits well into the stratigraphic succession. Although biostratigraphic evidence for the age of this unit has been missing, it is probably older than the overlying sequence.

From the top of the core to 522 m, the sequence is made up of brownish grey, drab or grey limestones, cherty limestones, in some horizons with coquinas and intrabreccia interlayers.

Based on macroscopic and microscopic features, the upper limestone unit can be subdivided into three subunits (see Fig 3).

The lower subunit (522–345 m) consists of dark or light brown, drab, grey limestones, dolomitic limestones (calcareous dolomites in the few meter thick basal part) with chert lenses, nodules, or silicified patches, and bioturbated intervals (Fig. 4). Intraclasts were also observed in a few samples.

In the upper part of the subunit *Halobia–Daonella* coquinas were found in two horizons (Detre 1971). Radiolaria, filament or radiolaria-filament wackestones (Plate I) are the characteristic texture types. Bioclasts of platform origin are very rare, and were found only in a few samples (see: Fig. 4).

These features clearly indicate a relatively deep marine pelagic depositional environment, relatively far from the shallow platforms. The *Halobia–Daonella* assemblage also indicates a pelagic basinal environment. However, conodonts found in the upper *Halobia* horizon (Kozur and Mostler 1973) suggest a rather restricted environment, since elements of the typical pelagic *Gladigondolella*, with the exception of *Prioniodina venusta*, are missing (Kovács S. pers. comm.).

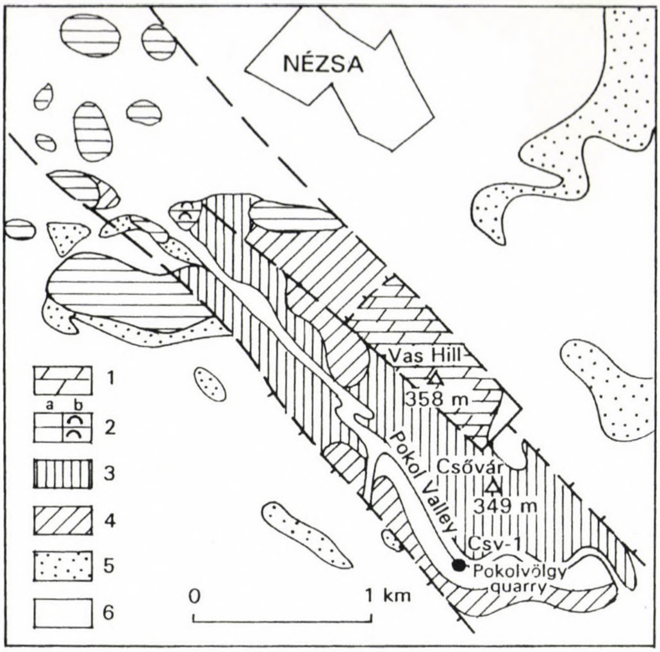


Fig. 2
Geologic map of the study area (after Detre 1970). 1. "Vashegy Dolomite"; 2. Dachstein Limestone (Upper Triassic), a) oncoidal facies, b) reefal facies; 3. Csővár Limestone Formation (Upper Triassic–Lower Jurassic); 4. Szépvölgy Limestone Formation (Eocene); 5. Hárshegy Sandstone Formation (Oligocene); 6. Quaternary formations

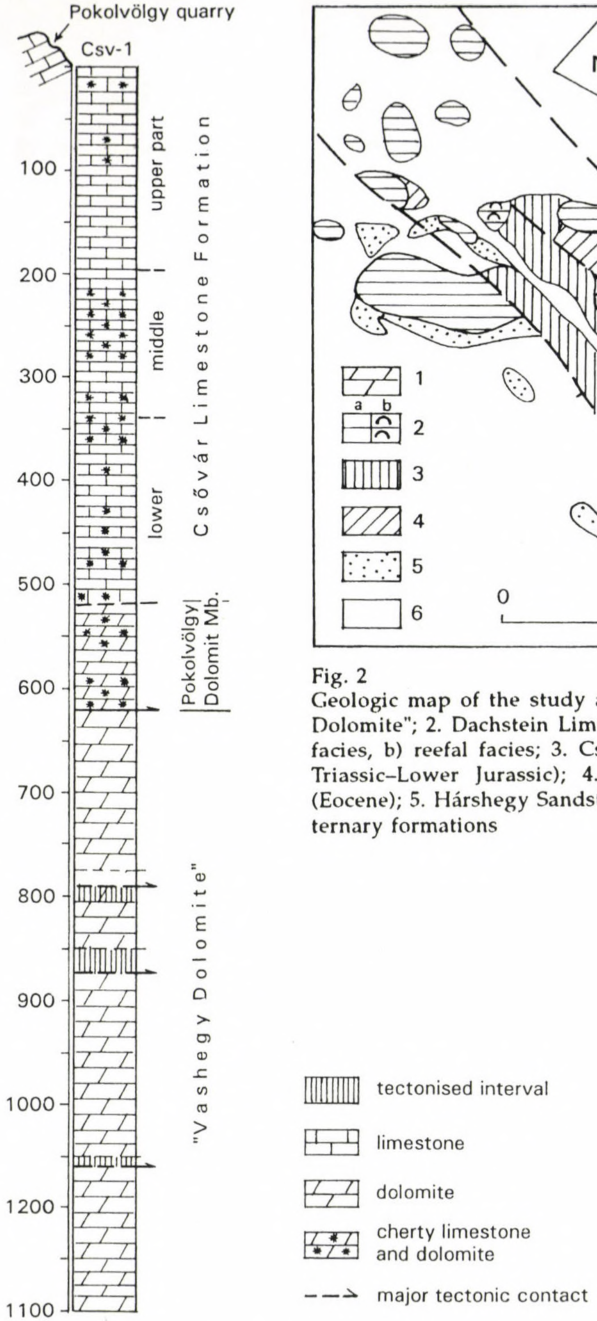


Fig. 3
Lithology and lithostratigraphical units of the core Csővár-1 complemented by the section exposed in the Pokolvölgy quarry

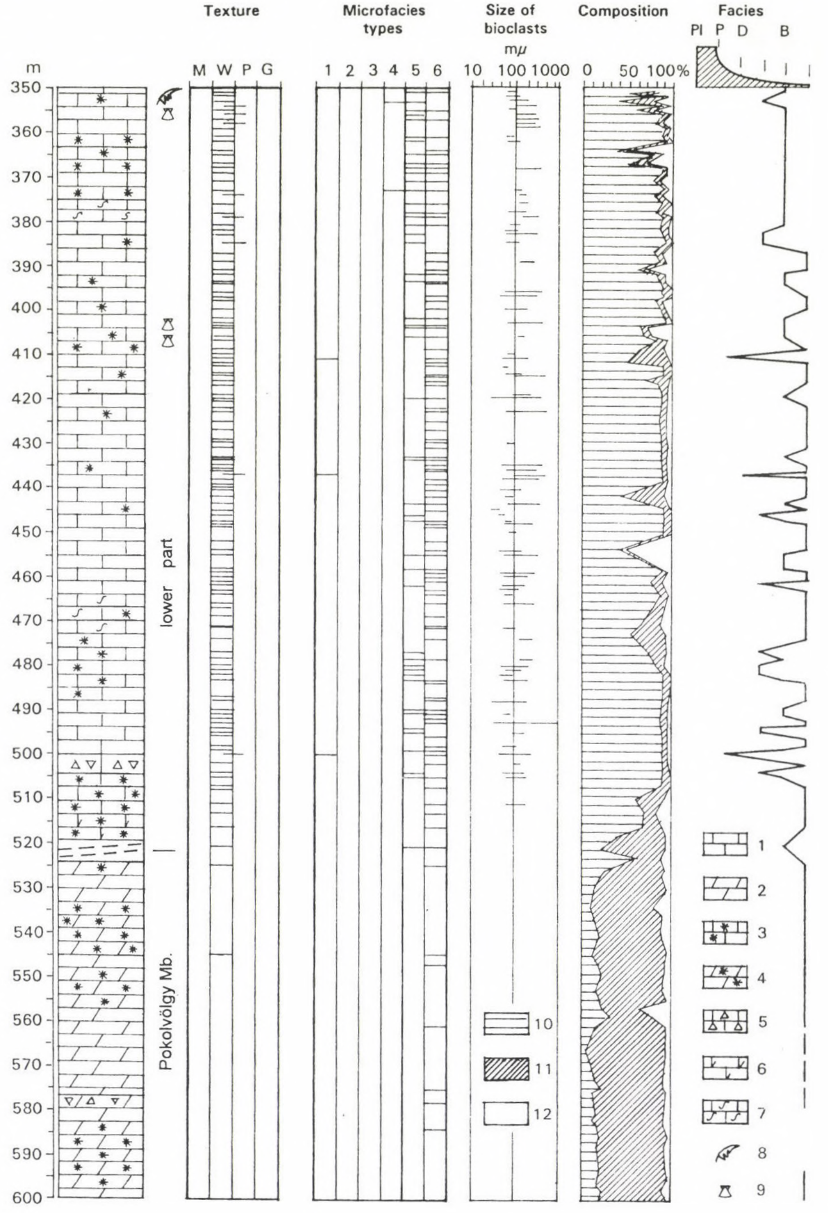


Fig. 4

Lithology, texture, microfacies-types, rock-composition and facies interpretation of the Csv-1 core, between 600–350 m. 1. limestone; 2. dolomite; 3. cherty limestone; 4. cherty dolomite; 5. lithoclastic limestone or dolomite; 6. dolomitic limestone; 7 bioturbated limestone; 8. conodonts; 9. bivalves; 10. calcite; 11. dolomite; 12 insoluble residue; *Microfacies types:* 1. packstone or wackestone with redeposited bioclasts and intraclasts of platform origin; 2. crinoidal grainstone; 3. graded laminite (alternation of graded calcarenite and calcisilt laminae); 4. fine laminite (alternation of calcisilt and micrite laminae); 5. filament wackestone (rarely packstone); 6. radiolarian wackestone (rarely packstone); Pl – platform; P – proximal toe-of-slope; D – distal to-of-slope, B – basin

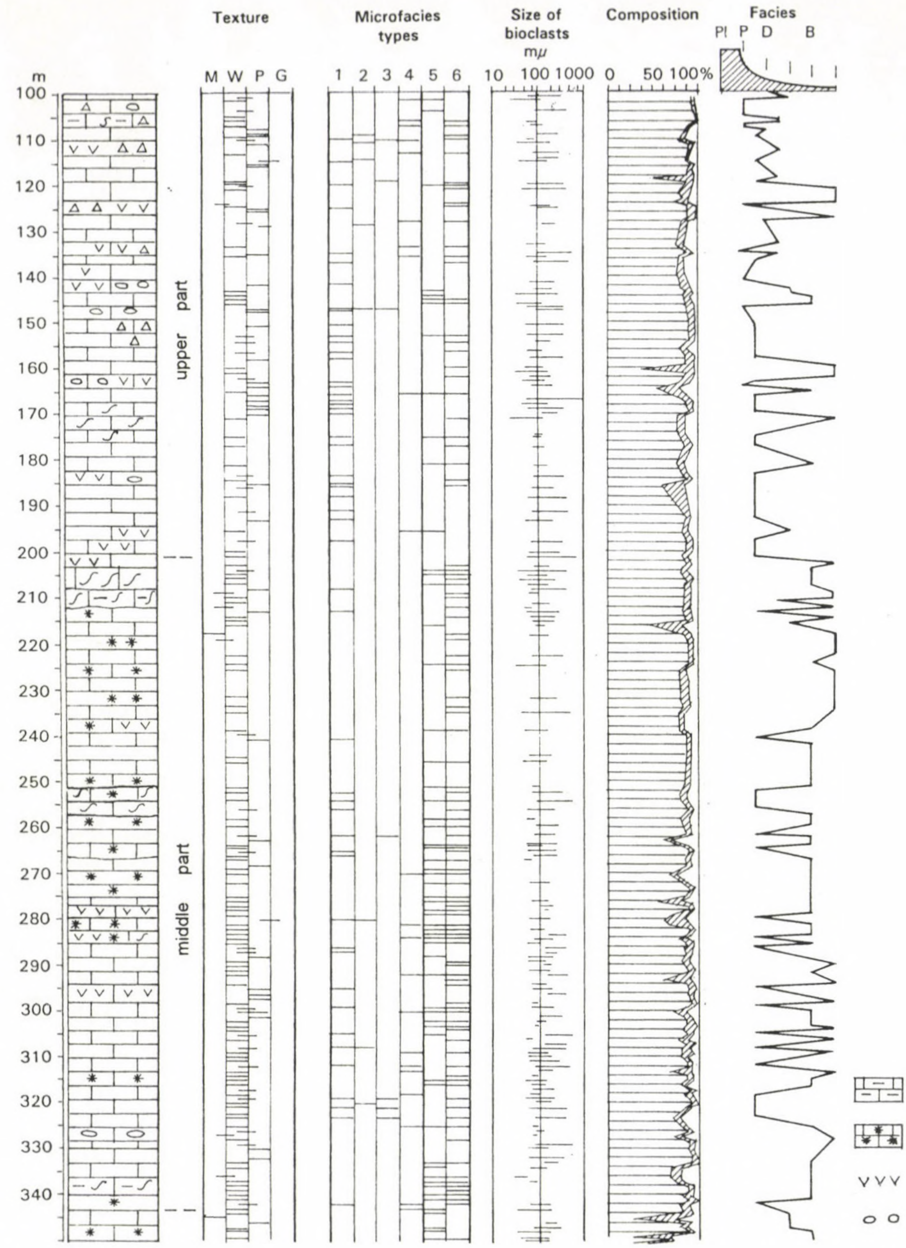


Fig. 5

Lithology, texture, microfacies-types, rock-composition and facies interpretation of the Csv-1 core between 350–100 m. For legend see Fig. 4. Additional symbols: 1. argillaceous limestone; 2. cherty limestone; 3. crinoidal; 4. plasticlasts

The middle subunit (203–345 m) is made up of dark grey or brown, drab limestones, and cherty limestones (Fig. 5). Argillaceous, bioturbated intervals with flaser bedding and bedding planes are characteristic. Crinoidal lenses as well as crinoid-coquina interbeds also appear. In some samples (e.g. 300.0–301.5 m) typical features of the allodapic limestones are visible: dark brown coarse-medium calcarenite packstone at the base of the layer was followed by a laminitic, then a convolute microlayer; another laminitic microlayer occurs at the top of the package (Plate II). Radiolaria (Plate IIIB), filament (Plate IA) or radiolaria-filament wackestones are still predominant, but layers containing bioclasts of platform origin become more frequent. Typical platform elements are porostromate algae, *Cayeuxia alpina* Flügel, *Tubiphytes* sp., *Baccinella* sp., and *Galeanella* sp. (Plates IV, V).

The middle subunit may have been deposited predominantly in the distal part of the toe-of-slope facies of a carbonate platform. A significant part of the sequence is made up of calcarenites and carbonate muds, transported by turbidity currents from the platform to the basin. However, periods of intense gravity flow activity alternated with calm intervals, when deposition of pelagic oozes prevailed.

The upper subunit (0–203 m) consists of brown, brownish-grey, and dark-grey limestones, argillaceous limestones, calcareous marls, and subordinately dolomitic limestones (Figs 5, 6). Cherty layers are scarce and limited to the upper 100 m of the section. Graded, allodapic packstone or grainstone layers alternate with laminated mudstones in many cases (Plate IIA). Intraclastic interlayers (debris-flow deposits) also occur in some horizons (Fig. 6).

As far as the microfacies is concerned, bioclastic and intraclastic packstones, frequently graded, become predominant; however, radiolarian, filament sponge spicule wackestones (Plate VI), alternating with the former microfacies, are still common. Bioclasts and intraclasts in the packstones are of platform origin as a rule. Typical fossil elements are *Cystothalamia* cf. *minima* and *Poriferitubus buseri* Senowbari-Darian (Plate VII).

This part of the section is a typical toe-of-slope facies. Debris flow breccias may have been deposited in the proximal zone at the toe of the slope. Graded allodapic carbonates were deposited from turbidity currents on the low-angle distal toe-of-slope. Microscopic observations suggest that laminitic mudstones are actually very distal turbidites (see Fig. 7), whereas the homogeneous as well as bioturbated mudstones are pelagic basinal facies which may have deposited in the "calm" periods or in the inner part of the basin, beyond the influence of the turbidity currents.

The section of the Pokolvölgy quarry

The Pokolvölgy quarry (its location is shown in Fig. 2) exposes a section which directly overlies the sequence exposed by the Csv-1 core. Although some parts of the rock succession are covered by scree, most of the approximately

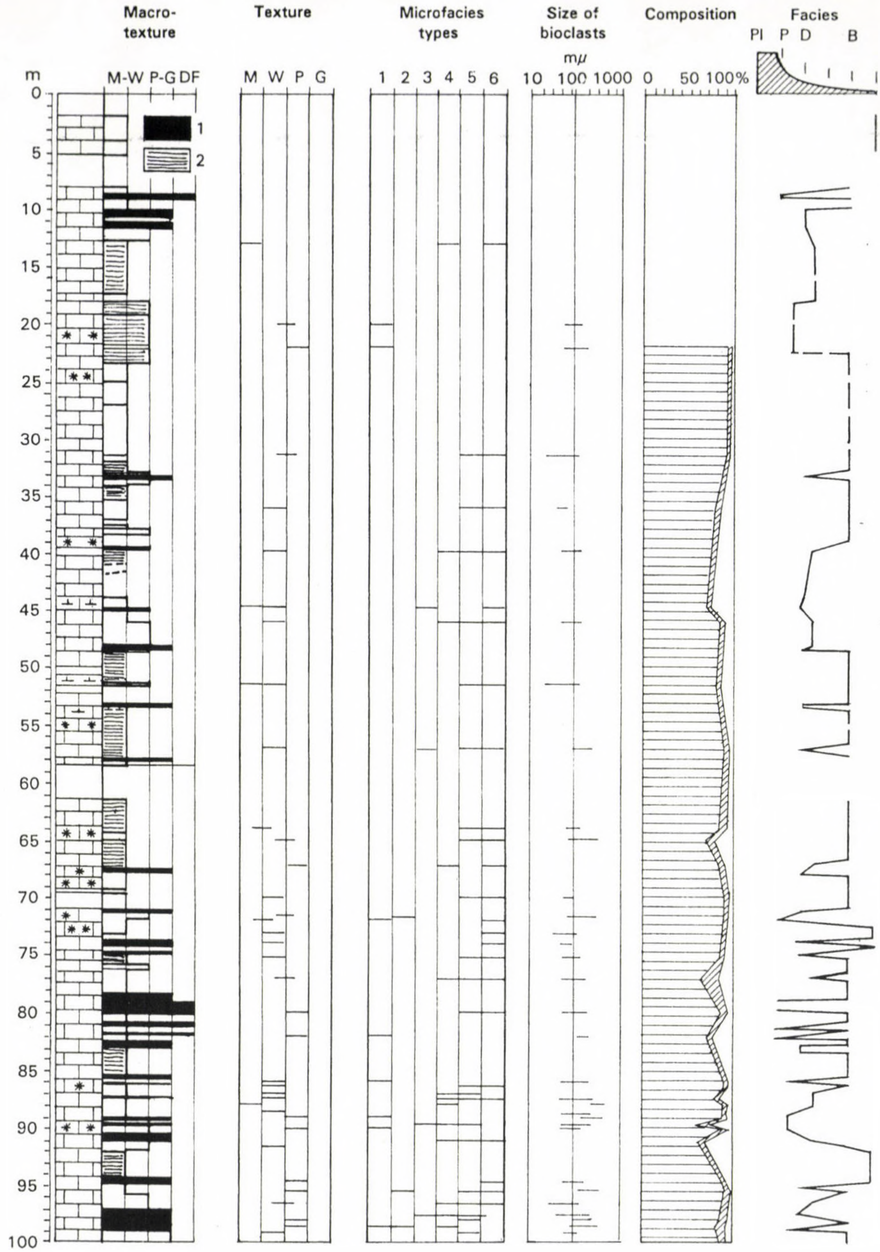


Fig. 6

Lithology, macrotexture and microtexture, microfacies types, rock composition, and facies interpretation of the Csv-1 core between 100–0 m. M – mudstone; W – wackestone; P – packstone; G – grainstone; DF – debrite. For legend see Figs 4, 5. Additional symbols: 1. turbidites and debrites; 2. laminites

Fig. 7
Location of the measured section in the Pokolvölgy quarry

30 m-thick section is well visible, offering a favourable opportunity for detailed sedimentological studies and a possibility to complete the core data.

Three sections were measured in the quarry. The location of these sections are displayed in Fig. 7. A composite section is presented in Fig. 8, together with results of the microfacies investigations.

The basal part of the lower section is covered at present. However, according to J. Oravecz and F. Góczán (pers. comm.) dark-grey calcareous marl, rich in organic material, were formally exposed beneath the limestone layers, when the quarry was operating in the early 60s. It is worth mentioning that a 2–3 cm thick, light-grey silty marl layer at the very base of the sequence yielded the richest sporomorph assemblage of the formation, and first specimens of the new algal cyst genus *Oraveczia* were also found in this layer (Góczán, present volume).

Over the scree-covered marl layers, dark-grey, brownish-grey, thin-bedded limestones are visible in a small outcrop just above the base-level of the quarry (Fig 9 Nos 1, 2). These layers are made up of alternations of calcisiltic wackestone and mudstone laminae.

Higher up a 4 m-high cliff section superbly demonstrates the most characteristic sedimentological features of the lower part of the exposed succession (Fig. 12). This section begins with a dark-grey packstone–wackestone bed (No. 3) with 0.5–3 mm sized lithoclasts. This bed is overlain

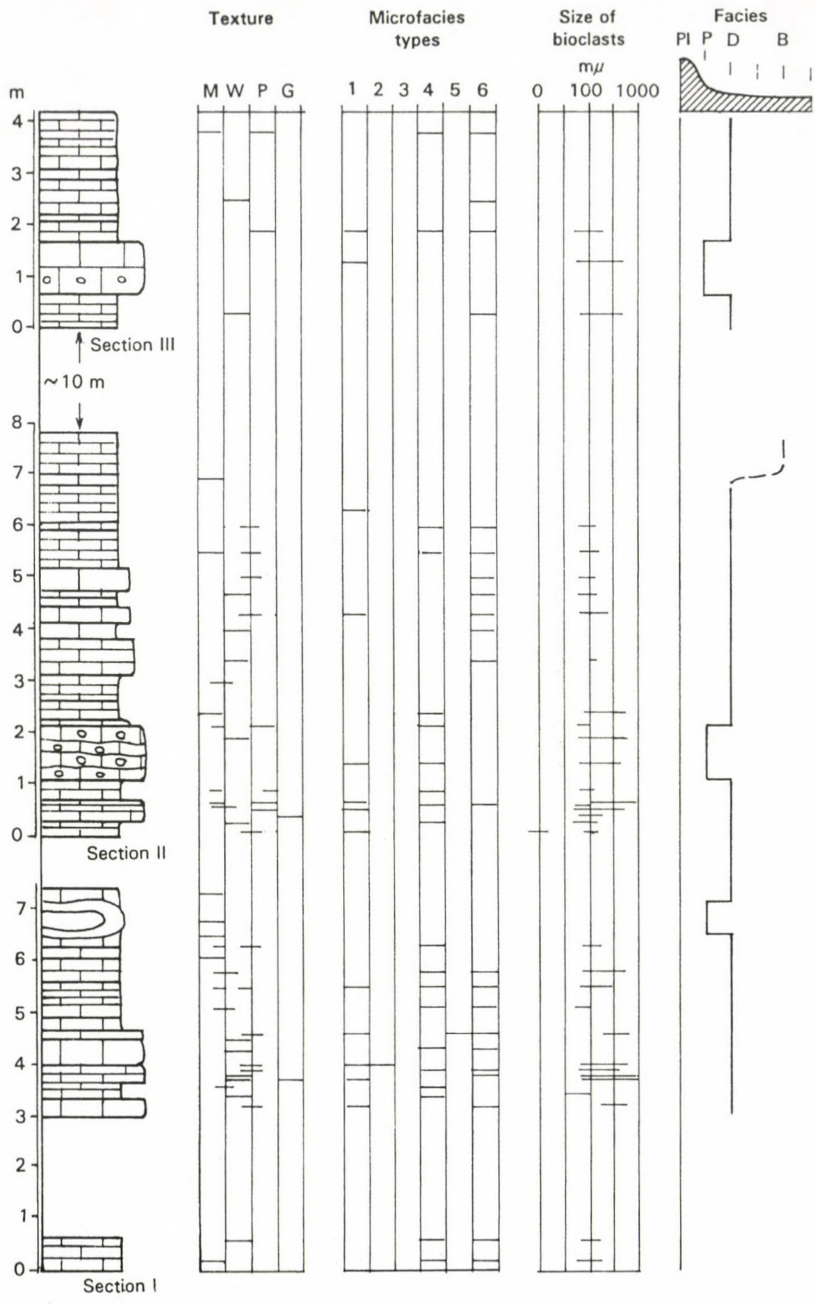


Fig. 8
Composite section of the Pokolvölgy quarry showing lithology, texture, microfacies types, and facies interpretation of the measured sections. For legend see Figs 4, 5

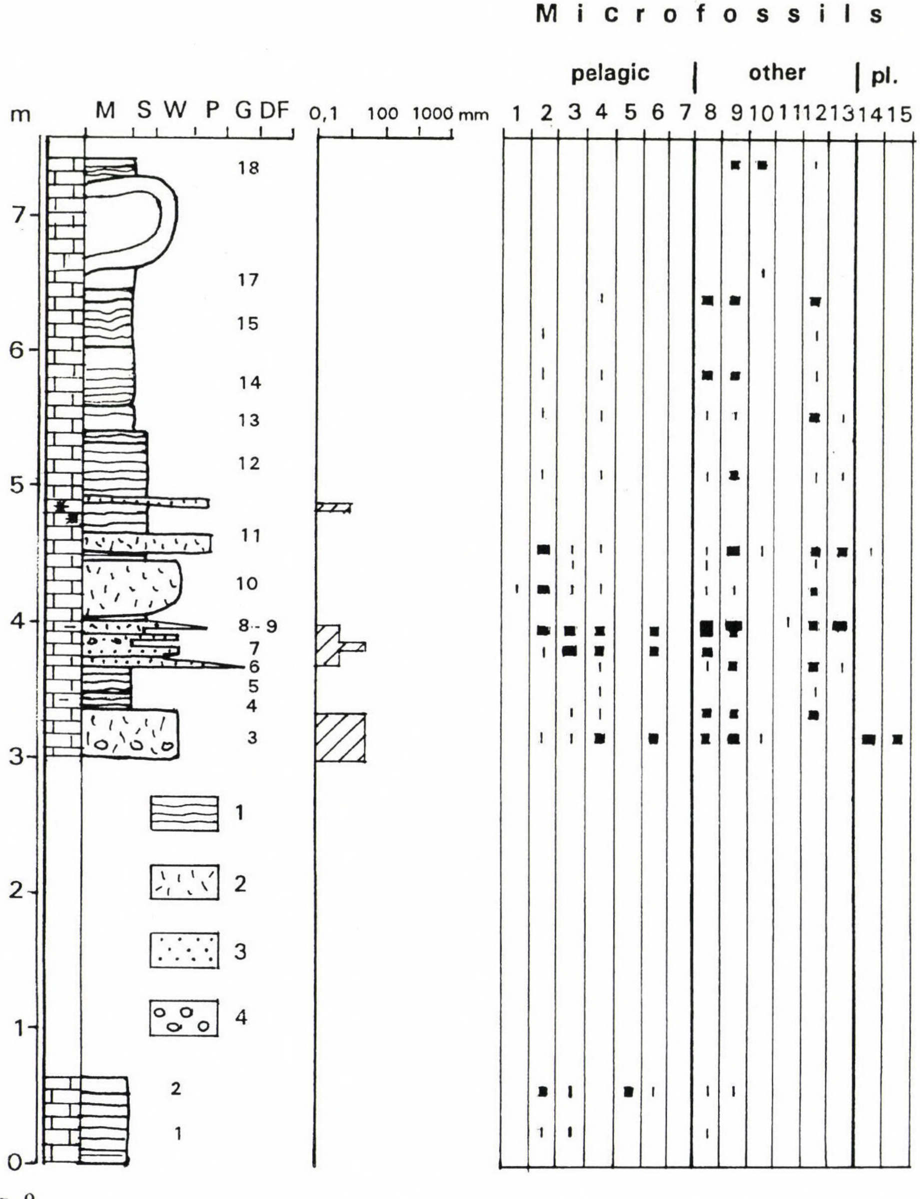


Fig. 9
Lithologic column, clasticity index and biofacies data of the lower section in the Pokolvölgy quarry. For legend of the lithologic column see Figs 4, 5. Additional symbols: 1. laminite; 2. skeletal wackestone; 3. calcarenite packstone-grainstone; 4. lithoclasts

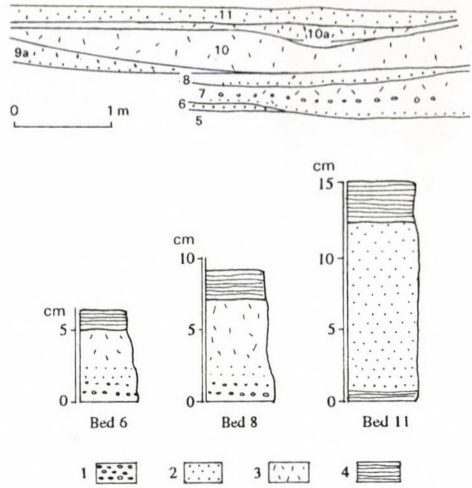


Fig. 10
Channels and channel-fill calciturbidites in the middle part of the lower section in the Pokolvölgy quarry. 1. coarse calcarenite; 2. medium to fine calcarenite packstone; 3. skeletal wackestone; 4. laminite

by brownish-grey laminitic wackestones–mudstones (Nos 4, 5). An uneven erosion surface is visible on the top of the upper laminitic layer (No. 5). Above the erosion surface, classic turbidite layers showing features of Bouma cycles were found as channel fill (see Figs 9, 10; beds No. 6–11, Plate VIIIA). The turbidite layers begin with a few cm-thick coarse calcarenites grading upward into bioclastic wackestones, and end with laminitic mudstones (Plate IX).

Upward, laminitic layers consisting of alternations of packstone–wackestone and mudstone laminae become predominant and only thin calcarenite interbeds are visible. Above these sedimentary folds and large slump structures can be observed (Fig. 9; Plate VIIIB).

The second part of the section is exposed by an 8 m-high cliff in the middle part of the quarry (Fig. 11). This part of the section begins with allodapic calcarenite layers alternating with laminitic layers, made up of alternating fine calcarenitic packstone–wackestone and calcisiltic wackestone laminae. They are followed by a 1 m-thick bed, containing dm-sized lithoclasts (Fig. 13). This matrix-supported, unsorted layer is a typical debris-flow deposit. The matrix is mudstone–wackestone, akin to the texture of the intraclasts.

Higher up in the succession, thin-bedded, laminated mudstone layers alternate with thicker bioclastic wackestone beds, grading upward into thicker layers of homogenous radiolarian mudstone. At the base of some layers uneven erosional surfaces are visible. There is an about 10 m-thick covered interval between the second and the third measured sections, which probably consists mainly of laminitic, argillaceous limestones.

The upper section (Fig. 14) begins with brownish-grey cherty wackestone. It is overlain by a 1 m-thick bed containing a large amount of coarse lithoclasts and bioclasts. It shows features of debris-flow deposits. Molluscs, fragments of silicified corals and sponges, as well as foraminifera and algae of platform origin were found in this bed (Fig. 15). A significant amount of terrigenous clastics is a peculiar feature of this layer. Monocrystalline and polycrystalline quartz, chlorite, magnetite and zircon grains were recognised (B. Árgyelán, pers. comm.).

Fig. 11
The lower measured section in the Pokolvölgy quarry. Channels filled by calciturbidites are visible in the middle part of the cliff

In the next layers, fine calcarenite and mudstone microlayers alternate, whereas laminitic radiolarian wackestones with clayey parting surfaces and silty mudstones are visible in the topmost part of the quarry.

Organic microfacies

Results of the organic microfacies studies and the interpretation of the depositional environments based on them are shown in Fig. 16.

The following microfacies types were distinguished:

1) The relative quantity of terrigenous coal detritus of wood origin is abundant or very abundant. In the coarse fraction unrounded, or only slightly rounded grains are predominant, in addition to coarse grains of autochthonous filamentous algae.

The amount and size of the terrigenous coal grains suggest the proximity of a continental area. An open marine connection to the depositional basin (lagoon) is indicated by the occurrence of organic tests of foraminifera and micro-plankton with organic skeletons. Carbonisation of the organic material indicates reductive conditions, and the angular shape of the grains indicates a low-energy

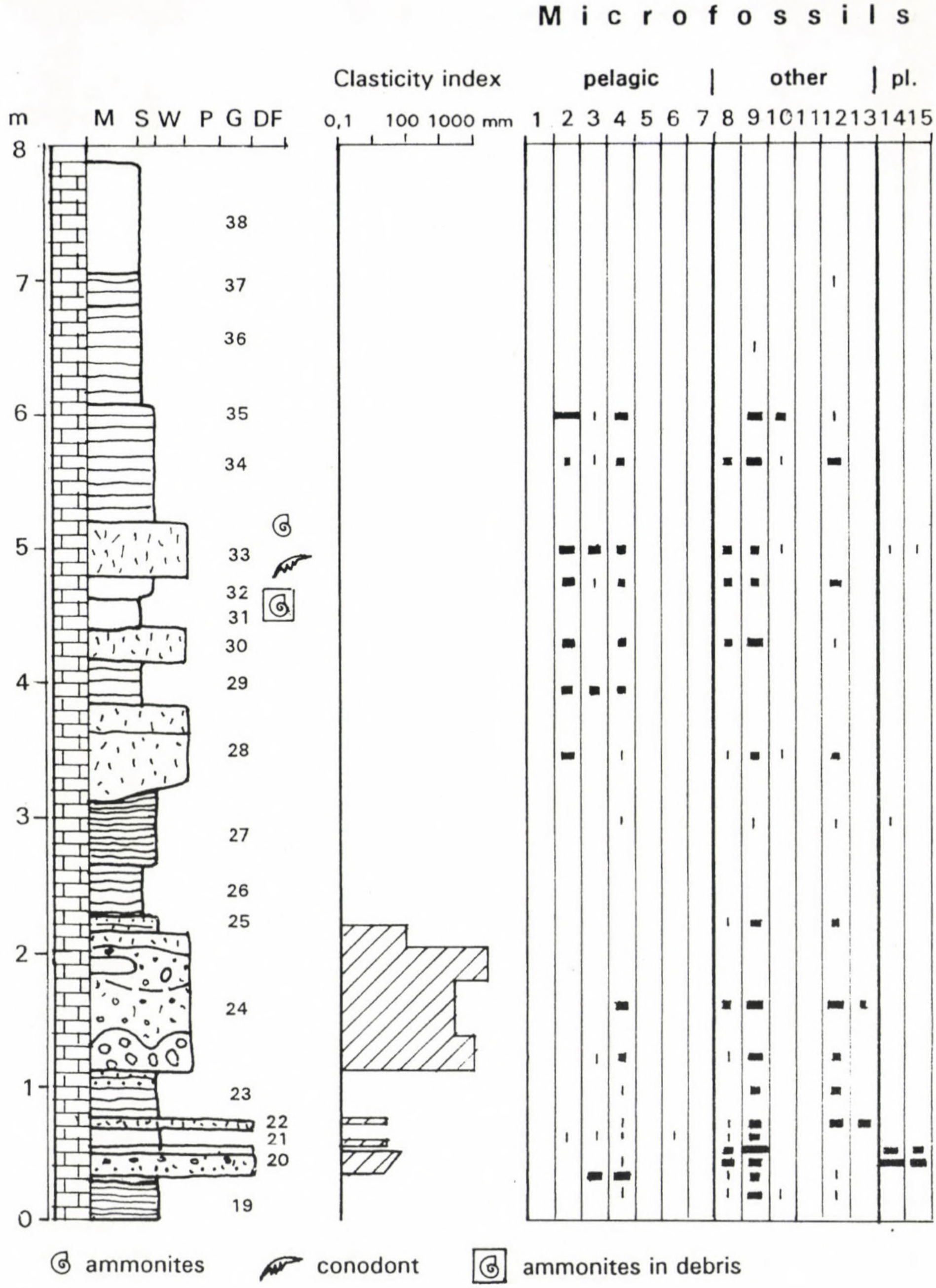


Fig. 12
Lithologic column, clasticity index, and biofacies data of the middle section in the Pokolvölgy quarry. For legend and list of microfossils see Fig. 9

environment. This microfacies type characterises the samples taken from the base of the quarry in 1961 (Plate XI).

2) Coal detritus of wood origin are abundant. They are definitely rounded, but undecayed. The size of the grains is medium or small, as is that of the detritus of filamentous algae. The amount and size of the terrigenous organic material indicate a relatively short distance to a subaerially exposed area. This microfacies type was found in the Csv-1 core at 404.0 m and 400.0 m, as well as in the samples of the basal layers (Plate XIIA) and middle part of the succession of the quarry (Plate XIIB).

3) The amount of detrital organic material is very abundant as a rule. The coarse fraction of the coal grains is only subordinate. The medium and small-sized grains are rounded, the latter being frequently decayed; the epiderm detritus and pollen grains are commonly oxidised due to the mode of their transportation at the top of water. In many cases, the amount of air-transported pollen grains is greater than that of the coal grains, but they may also be totally absent. The predominance of a fine fraction among the terrigenous organic grains indicates a relatively great distance to a continental source area. Organic microplankton tests are always present suggesting a connection with an open marine basin. Redox conditions on

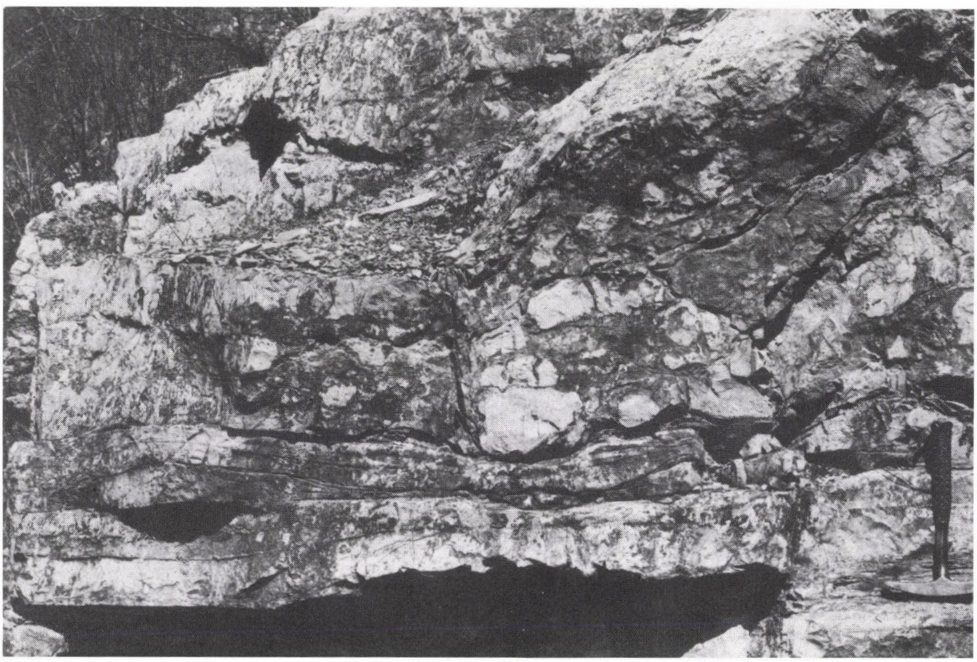


Fig. 13
Debris flow deposits in the lower part of the middle section (layer No. 24)

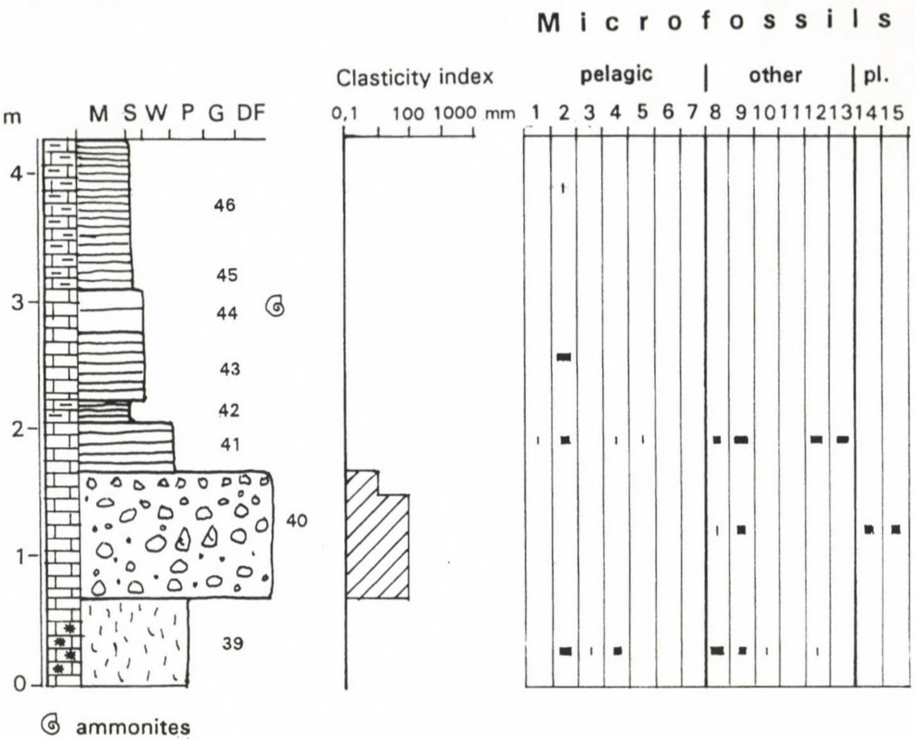


Fig. 14
Lithologic column, clasticity index, and biofacies data of the upper section in the Pokolvölgy quarry. For legend and list of microfossils see Fig. 9

the bottom may have been variable depending on the character of the various microenvironments.

Samples taken from the Csv-1 core at 520.8 m, 461.0 m (Plate XIIC), 474.0 m, 291.4 m and from the upper segment of the quarry (Plate XIID) can be classified in this group.

4) The relative quantity of the organic grains is very abundant as a rule. Terrigenous coal grains of wood origin are unsorted, generally toothed but undecayed. Among the sporomorphs the appearance of the folded, deformed exines is remarkable. Colloid material may also occur. The preservation state of the terrigenous organic material indicates rapid redeposition, suggesting a toe-of-slope depositional environment.

Samples of the Csv-1 core at 490.7 m 471.7 m and 281.6 m (Plate XIIIA) can be emplaced in this group.

5) The relative quantity of the organic material is varied, but in many cases very abundant. Every fraction of the wood grains is highly rounded, strongly or

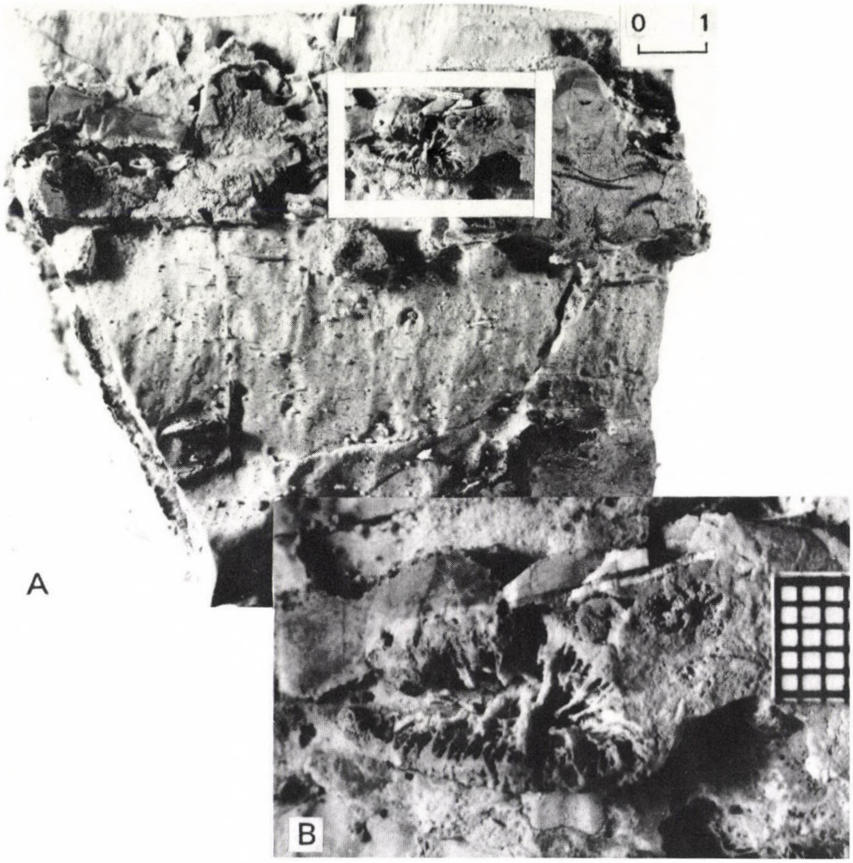


Fig. 15
A) Redeposited silicified bioclasts (corals, sponges, molluscs) and lithoclasts in debris flow deposit in the upper section of the Pokolvölgy quarry (layer No. 40). B) A detail of the upper photo showing a coral fragment

weakly decayed. The amount of colloidal material is very abundant. A significant part of the colloidal material, however, was not formed by decaying but rather by fracturation of the organic grains. The relative quantity of sporomorphs may attain medium grade, but they are always decayed, oxidised. This microfacies type may have been deposited in a distal toe-of-slope environment.

Samples 515.5 m, 507.5 m, 487.0 m, 477.2 m, 459.0–420.5 m, 410.0 m 382.6 m of the Csv-1 core may be classified into this group.

6) The relative quantity of organic material may rarely reach the abundant stage. The amount of coal, however, is only poor or medium. The coarse fraction is entirely missing and only a few decayed and rounded samples of medium-size

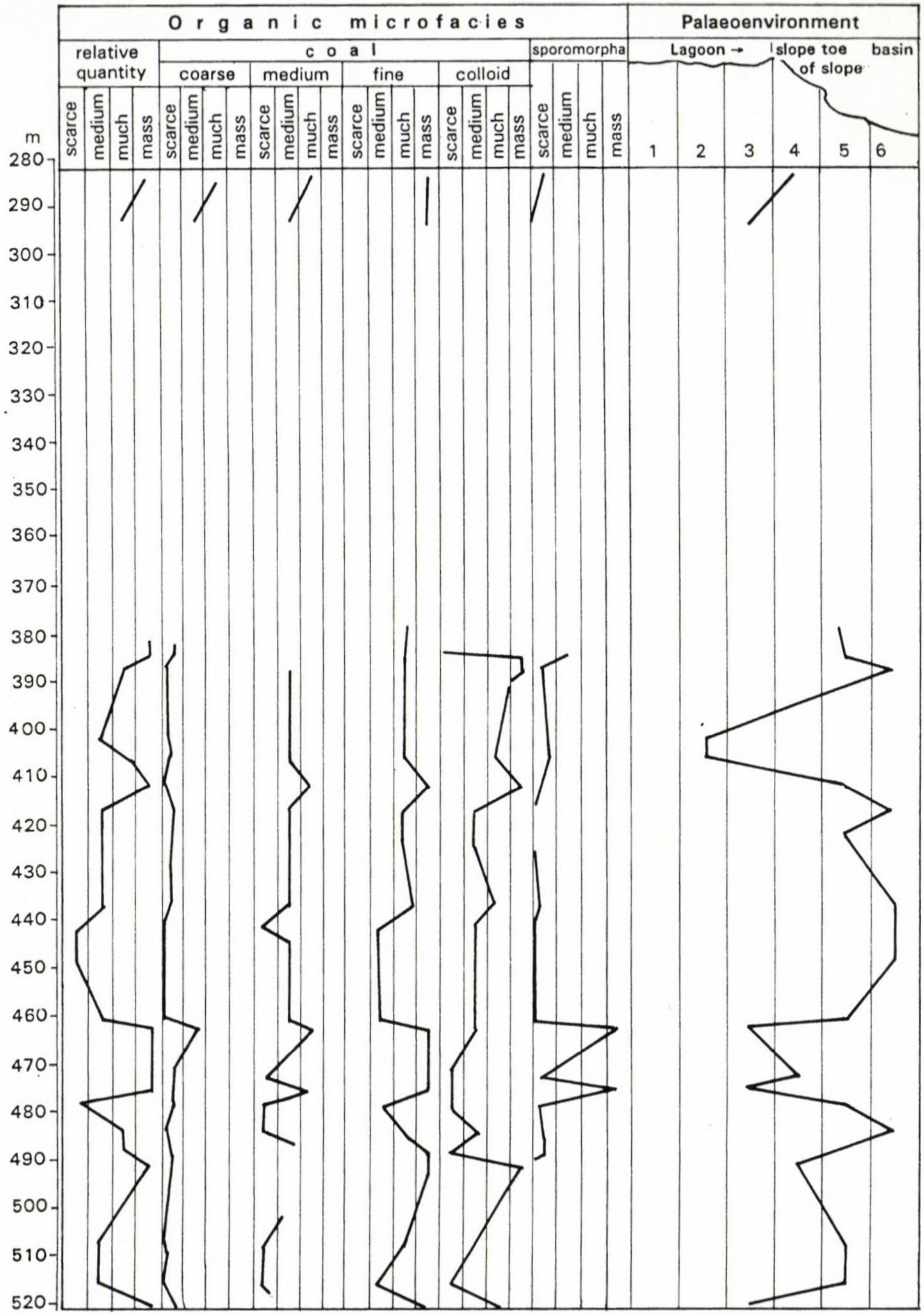


Fig. 16
Organic microfacies data and their paleo-environmental interpretation in the Csv-1 core

grains are present. Even in the fine fraction only very small (almost colloidal) grains occur. A characteristic component of the microfacies is the dispersed, grey or dark yellow colloid of organic decay origin. Decayed, oxidised fragments of sporomorphs are scarce. This microfacies-type was formed in a deeper basin.

Samples 483.2 m, 447.5 m 443.4 m, 440.3 m 435.8 m, 415.0 m and 385.0 m (Plate XIIIB) may be included in this category.

Lithostratigraphy

Based on investigations of the Csv-1 core and the measured sections in the Pokolvölgy quarry, we propose the following lithostratigraphic subdivision of the studied sequence (see Fig. 3).

The dolomitized complex with a complicated inner structure exposed in the lower part of the Csv-1 core should be distinguished from the overlying sequence. It is assumed that this complex also crops out on Vas Hill; however, this idea requires further proof. If this guess is correct, the name "Vashegy Dolomite" first proposed by Balogh (1981) for dolomites of Vas Hill should be used for the entire, strongly dolomitized complex of heterogeneous origin.

The sequence exposed in the upper part of the Csv-1 core (above 620 m) as well as in the Pokolvölgy quarry is made up of genetically related carbonate rocks, representing a facies series from the platform foreslope to the related restricted basin. It is suggested to apply the name Csővár Limestone Formation, proposed by Balogh in 1981, in an extended sense for the whole succession, also including the dolomitized rock types. It is proposed, however, to distinguish the cherty dolomite interval exposed in the middle segment of the Csv-1 core (522.0–622.0 m) as a member of the Csővár Formation under the name of Pokolvölgy Dolomite Member.

Biostratigraphy

The biostratigraphic evaluation is based on the investigations of sporomorphs and foraminifera in the Csővár 1 core and also on conodonts and ammonites from the Pokolvölgy quarry.

In the last couple of years, a relatively rich megafossil assemblage was collected from the Pokolvölgy quarry. Most of these fossils were found in debris in the quarry. A significant portion of the megafossils was subjected to silicification and consequently they are well visible on the weathered rock surfaces.

The detailed study of the quarry sections revealed that the majority of the megafossil-bearing rock-fragments may have most probably originated from layer No. 40 (see Fig. 14).

In certain beds (e.g. layer No. 40) and on the bedding planes of some beds a great number of varied megafossils are visible: sponges, corals, bivalves, gastropods, ammonites, crinoids, and echinoids. Remnants of vertebrates were also found.

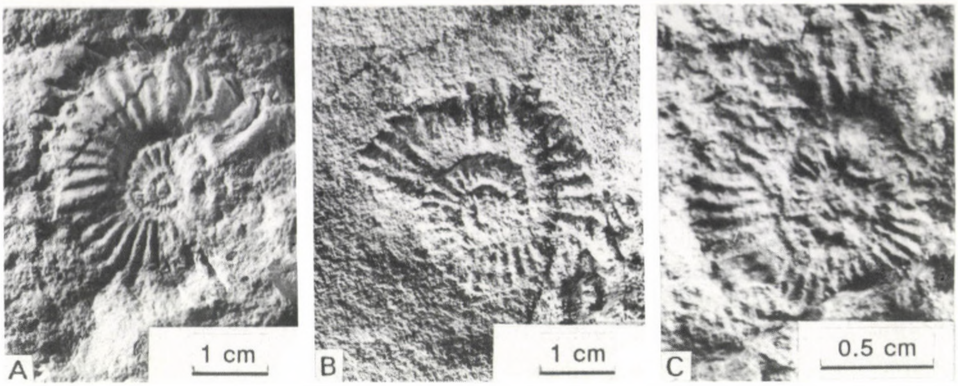


Fig. 17
Ammonites from the Pokolvölgy quarry. A) *Choristoceras* cf. *marshi* Hauer; B) *Choristoceras* cf. *marshi* Hauer; C) *Vandaites stuerzenbaumi* (Mojsisovics)

From a stratigraphic point of view the ammonites are of critical importance (Fig. 17). Representatives of the genus *Choristoceras* were found in the largest number. The majority of the specimens was collected from rock fragments, but they were also encountered in the laminated limestones in the topmost part of the quarry (see Fig. 14). Most of them were determined as *Choristoceras nobile* Mojsisovic. Recently, a specimen of *Choristoceras* cf. *marshi* Hauer was found, also in rock debris. According to its host-rock it probably originated from the layer No. 40 two other specimens were received from J. Oravecz which had been collected by him and E. Végh-Neubrandt in the same quarry in 1961. Also in rock debris, originally from the middle section of the quarry, an ammonite determined by L. Krystyn as *Vandaites stuerzenbaumi* Mojs. (pers. comm. 1995) was found. Ammonites belonging to other genera also occur, but only their cross-sections are visible as a rule.

Choristoceras nobile Mojs. occurs both in the Upper Norian and Rhaetian, so it cannot be used for detailed age determination. The *Vandaites stuerzenbaumi* Mojs. found in the middle segment of the quarry is the zonal index species of the Lower Rhaetian according to Krystyn's biozonation. The occurrence of *Choristoceras* cf. *marshi* Hauer in the topmost part of the quarry suggests the existence of the upper biozone of the Rhaetian.

In addition to the ammonites representatives of other fossil groups are also worth mentioning. Remnants of sponges (5–30 mm in size) are common, mainly in bed No. 40. Due to silicification, details of their original structures cannot be recognised. In the insoluble residue of the samples, sponge spicules frequently occur. Occurring at the base of some graded turbidite beds, corals are also silicified as a rule. Most of them are solitary; hermatypic corals are rare. Thin-walled bivalves are frequently visible on the bedding planes. In the sponge and coral-bearing allodapic limestone beds thick-walled bivalves are characteristic. Gastropods are visible on the weathered rock-surfaces, but also in the insoluble residue. Their size ranges from a few mm to 2 cm, and they can be classified into 4–5 morphotypes. Remnants of crinoids and echinoids are also frequently silicified. In the insoluble residue fragments of crinoid stems consisting of 2–3 ossicles and echinoid fragments and spines occur. Among the vertebrates several fish teeth (3 morphotypes could be distinguished) and two vertebra fragments were found in the insoluble residue.

In the insoluble residue of a sample from the middle section of the quarry a conodont specimen was found. It was identified by S. Kovács as *Gondolella steinbergensis* Mosher. This species indicates the lower part of the Rhaetian, just like the *Vandaites stuerzenbaumi* (Mojs.) which was found in the same segment of the quarry.

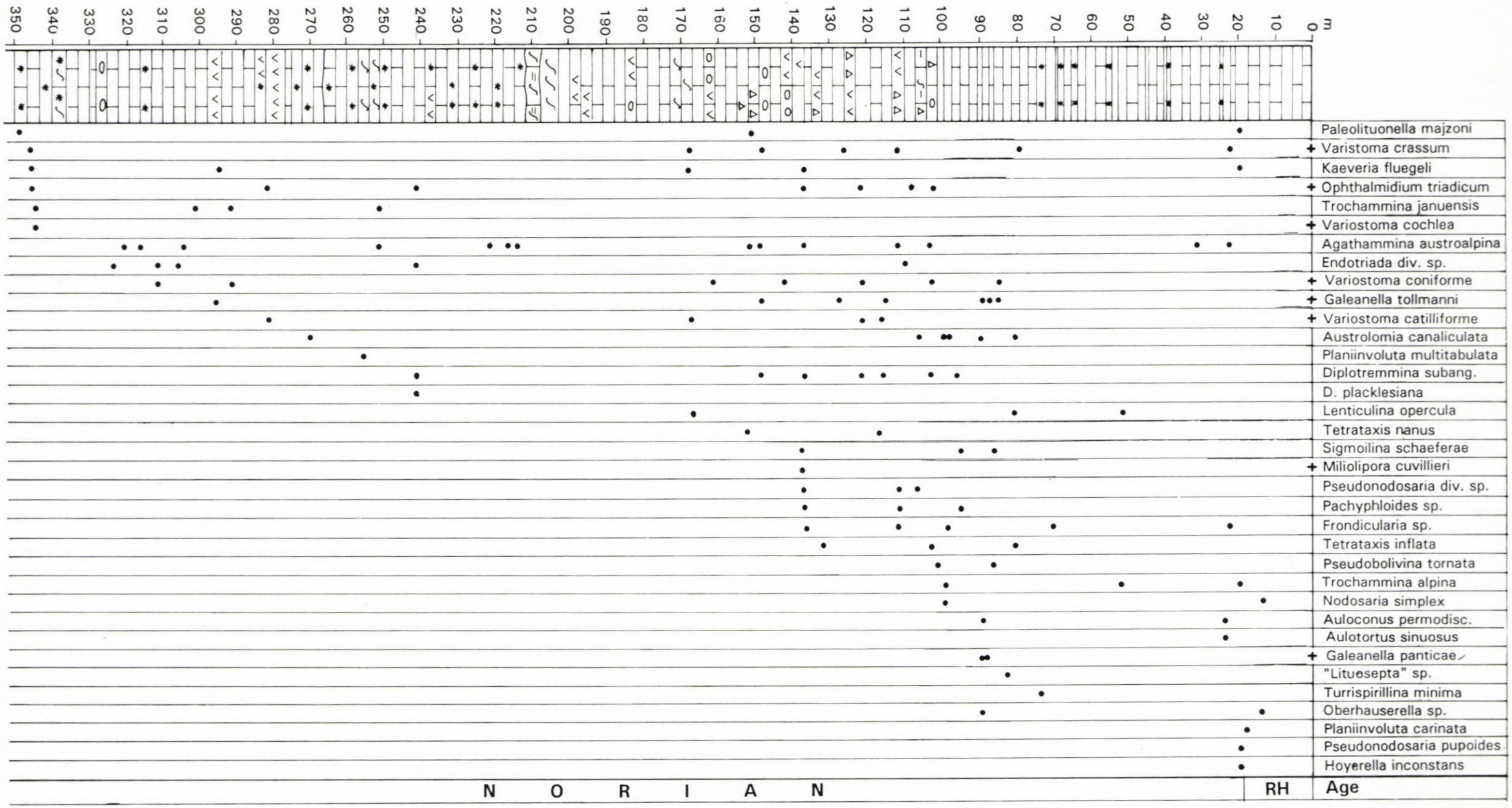
Bioturbation is common in many layers of the succession exposed by the quarry. It is worth mentioning, however, that Chondrites-type microfossils were observed on bedding planes of laminated limestones in the upper section.

Foraminifera

A rich foraminifera fauna was found in thin sections of samples taken from the upper portion of the Csv-1 core and the Pokolvölgy quarry. Some parts of the foraminifera assemblage, found in the Csv-1 core were presented previously (Oravecz-Scheffer 1987). Now our former studies were supplemented by investigations of the samples, taken in the course of resampling of the Csv-1 core, and newly collected samples from the Pokolvölgy quarry.

In the Csv-1 core, the Csővár Formation was exposed between 0.0 and 522.0 m. Within this interval the first foraminifera appeared at 411.3 m and above 354.0 m foraminifera of significant stratigraphical and sedimentological value were found (Plates XIV–XX). In Fig. 17 the determined taxa are shown, according to the order of their appearance in the samples. This succession, however, cannot be interpreted as a series of first appearances in a biostratigraphic sense, due to the peculiar paleoenvironmental setting of the sequence. The Csővár Formation represents toe-of-slope facies. Consequently it is not surprising that a significant part of the foraminifera fauna is allochthonous, containing elements originating from reefal or lagoonal environments. Since the autochthonous basinal assemblage is scarce, the allochthonous elements determine the basic character of the fauna. The basinal assemblage consists of fragile nodosariids with a long stratigraphic range (*Lecticulina opercula* (Crick et Sherborn), *Nodosaria simplex* (Terquem), *Pseudonodosaria* div. sp., *Frondicularia* div. sp., etc.) and small Oberhauserellidae, which are also referred to as "Triassic Globigerinids" (Fuchs, 1967, 1969, 1975). Instead of the foraminifera, in the microbiofacies of the basinal layers radiolarians, filaments (fragments of thin-shelled bivalves), sponge spicules, and small remnants of various echinoderms ("microechinoderms") are predominant.

Fig. 18 →
Foraminifera fauna of the Csv-1 core between 0 and 350 m. Elements of platform origin are indicated by a cross before the name of the taxon



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Paleolituonella majzoni
+ Varistoma crassum
Kaeveria fluegeli
+ Ophthalmidium triadicum
Trochammina januensis
+ Variostoma cochlea
Agathammina austroalpina
Endotriada div. sp.
+ Variostoma coniforme
+ Galeanella tollmanni
+ Variostoma catilliforme
Austrolomia canaliculata
Planiinvoluta multitabulata
Diplotremmina subang.
D. placklesiana
Lenticulina opercula
Tetrataxis nanus
Sigmoilina schaeferae
+ Miliolipora cuvillieri
Pseudonodosaria div. sp.
Pachyphloides sp.
Frondicularia sp.
Tetrataxis inflata
Pseudobolivina tornata
Trochammina alpina
Nodosaria simplex
Auloconus permodisc.
Aulotortus sinuosus
+ Galeanella panticae
"Lituesepta" sp.
Turrispirillina minima
Oberhauserella sp.
Planiinvoluta carinata
Pseudonodosaria pupoides
Hoyerella inconstans
N O R I A N
RH
Age

Among the allochthonous foraminifera fauna, a majority of the platform elements may have been inhabitants of near-reefal environments or intrareef cavities: "*Paleolituonella*" *majzoni* Bérczi-Makk, *Kaeveria fluegeli* (Zaninetti et al.), *Galeanella tollmanni* (Kristan), *Galeanella panticae* Zan. et Brönn., *Sigmoilina schaferae* Ducret, *Miliolipora cuvillieri* Brönn. et Zan., *Ophthalmidium triadicum* Kristan, and "*Lituosepta*" sp.

According to our biostratigraphic analysis, however, the foraminifera fauna of the Csővár Formation in the Csv-1 core, although it originated from various biotopes, does not suggest significant heterochronity. The assemblage of the 354.0–20.0 m interval constrains Norian age. This age determination is primarily supported by *Variostoma crassum* Kristan and *Variostoma catilliforme* Kristan (the latter is "norische Leitforaminifera" according to Tollmann, 1976) which could be found in the sequence between 354.0–22.5 m. Among them, *Turrispirillina minima* Pantic (Pantic 1967) is known also only from Norian formations.

The majority of the foraminifera assemblage is within the Norian–Rhaetian range: *Variostoma coniforme*, *Diplotremmina subangulata* Kristan, *Galeanella panticae* Zan et. Brönn., *Galeanella tollmanni* Kristan, *Auloconus permodiscoides* (Oberhauser), *Tetrataxis inflata* Kristan, etc., and consists of species appearing in the Carnian and characteristic of the entire Upper Triassic: *Aulotortus sinuosus* Weyn., *Agathammina austroalpina* Kristan, *Miliolipora cuvillieri* Brönn. et Zan., *Ophthalmidium triadicum* Kristan, *Trochammina alpina* Kristan, etc.

However, no species characterising only the Carnian (e.g. *Gsollbergella spiroloculiformis* Oravecz Scheffer, *Glomospirella capellini* Ciarapica et al., *Nodosaria ordinata* Trifonova, *Meandrospirella* div. sp.) or appearing only in the Rhaetian (they are listed later) were found. A subdivision of the Norian could not be carried out, either on the basis of the ranges or the quantity of the foraminifera taxa.

In the uppermost part of the Csv-1 core (above 20.0 m) *Hoyenella inconstans* (Michalik, Jendrejakova et Borza) (Plate XX, Fig. 18) and some *Oberhauserella* specimens, already characterising the Rhaetian, appear. These taxa are also present in the layers of the Pokolvölgy quarry. That is why it is suggested to emplace the Norian–Rhaetian boundary at 20.0 m in the Csv-1 core.

It is necessary to emphasis the biostratigraphic value of the species *Hoyenella inconstans* (Michalik, Jendrejakova et Borza), not previously referred to in the Hungarian literature. This taxon was described in 1979 from the *Rhaetavicula contorta* (Portl.)-bearing Fatra Formation ("Carpathian Koessen") of the Western Carpathians. Since then it has been found in many latest Triassic sections (Salaj et al. 1983; Ciarapica and Zaninetti 1984; Zaninetti et al. 1986; Ciarapica et al. 1987; Peybernés et al. 1988). Subsequently, as a result of wall-structure studies the taxonomic position of the species has been changed. In 1987 Ciarapica et al. placed it into the genus *Agathammina* (with a question mark!) and by including other forms, very similar morphologically, as synonyms they significantly enlarged the taxonomic range of the species. In 1994, Rettori classified the species in the Cornuspiracea superfamily and within it in the new Hoyenellidae family, and the genus *Hoyenella*. Recently, the species was also reported from Rhaetian dolomites in Tunisia, and from the area of the Sahara platform as well (Kamoun et al. 1994). They proved the extension of the Rhaetian transgression among other things by the occurrence of this species, claiming that in this time the northern and southern margins of the western basin of Tethys, which are located at present in Europe and North Africa, may have been the parts of the same huge carbonate platform.

The topmost layers of the Csővár Formation are exposed in the Pokolvölgy quarry. Thin section studies of samples taken bed by bed yielded well-preserved high-diversity foraminifera fauna (Fig. 19). Poor foraminifera assemblages were only found in the two basal layers, and in the topmost layers (samples Nos 44–46).

Although the sedimentological features of the quarry section are similar to those in the upper part of the Csv-1 core indicating toe-of-slope depositional environment, the foraminifera fauna of the quarry section is much more consistent ecologically. The number of the allochthonous elements of platform origin is lower; only some Variostomas, Milioliporas and species of *Ophthalmidium* may fall into this group. Among the definite reef-dwellers only a few *Galeanella* were found. On the other hand there are more encrusting, sessile foraminifera (*Planiinvoluta carinata* Leischner, *P. deflexa* Leischner, *P. multitabulata* Kristan-Tollmann).

In contrast to the fauna of the core section, in the entire succession of the Pokolvölgy quarry species characteristic of basinal environments are predominant and determine the general feature of the fauna. Nodosariids are generally present showing a relatively high diversity. Primitive agglutinated forms and genera of the Oberhauserellinidae family are also common.

Among the nodosariids, the fragile *Nodosaria* species are common: *Pseudonodosaria sphaerocephala* (Kristan-Tollmann), a *P. ellipsocephala* (Kristan-Tollman) a *P. vulgata multicamerata* (Kristan-Tollmann), *Lenticulina gryphea* (Kübler et Zwingli), *Vaginulina flaccida* (Schwager), *Frondicularia borealis* (Tappan), and *F. woodwardi* Howchin, etc.).

The agglutinated vagile benthic species of deeper basinal facies constitute the most characteristic group of the foraminifera fauna, especially in the middle part of the section of the quarry. They generally belong to the genera of the Ammodiscidae, Reophacidae and Lituolidae families: *Ammovertella polygyra* Kristan-Tollman, *Ammolagena clarata* (Jones-Parker), *Glomospira perplexa* Franke, *Ammobaculites suprajurassicus* (Schwagwer), *Ammobaculites latogranifer* Kristan-Tollmann, *Ammobacularia triloba* Kristan-Tollmann, *Reophax horridus* (Schwager), *Reophax rudis* Kristan-Tollmann, *Endoteba obturata* (Brönn. et Zan.), and *Verneuilina georgiae* Terquem.

The third characteristic group is constituted by the oberhauserellids (they are considered as direct ancestors of the planktonic foraminifera). Their identification is difficult in thin section, but *Oberhauserella rhaetica* (Kristan-

Tollmann), *O. quadrilobata* Fuchs, *O. norica* Fuchs and some specimens of the genera *Praegubkinella* and *Schlagerina* were recognised.

The character of the entire foraminifera assemblage of the Pokolvölgy quarry shows a striking similarity with that of the Zlambach Marl at the Fischerwiese (Kristan-Tollmann 1964). The rate of the common taxa is about 70%, and considering only the autochthonouos elements the similarity is even more pronounced. This fact is important from the point of view of the stratigraphic evaluation of the section and naturally also indicates the similarity of the depositional environments.

Although taxa of platform origin are known both from the Norian and the Rhaetian, the autochthonous foraminifera unanimously constrain the Rhaetian. The above-mentioned nodosariids and agglutinated species which were described by Kristan-Tollmann from the Zlambach Marl, *Oberhauserella rhaetica* (Kristan-Tollmann), and the *Schlagerina* and *Praegublinella* species are known only from the Rhaetian. The Rhaetian age determination is also supported by *Hoyenella inconstans* (Michalik, Jendrejakova et Borza), recognised in the topmost 20 m of the Csv-1 core as well, and *Gandinella falsofriedli* (Salaj et Samuel) also occurs only in the Rhaetian.

In the layers of the Csővár Formation exposed in the Pokolvölgy quarry we did not find taxa either of exclusively Norian range or ones appearing in the Liassic.

Sporomorphs

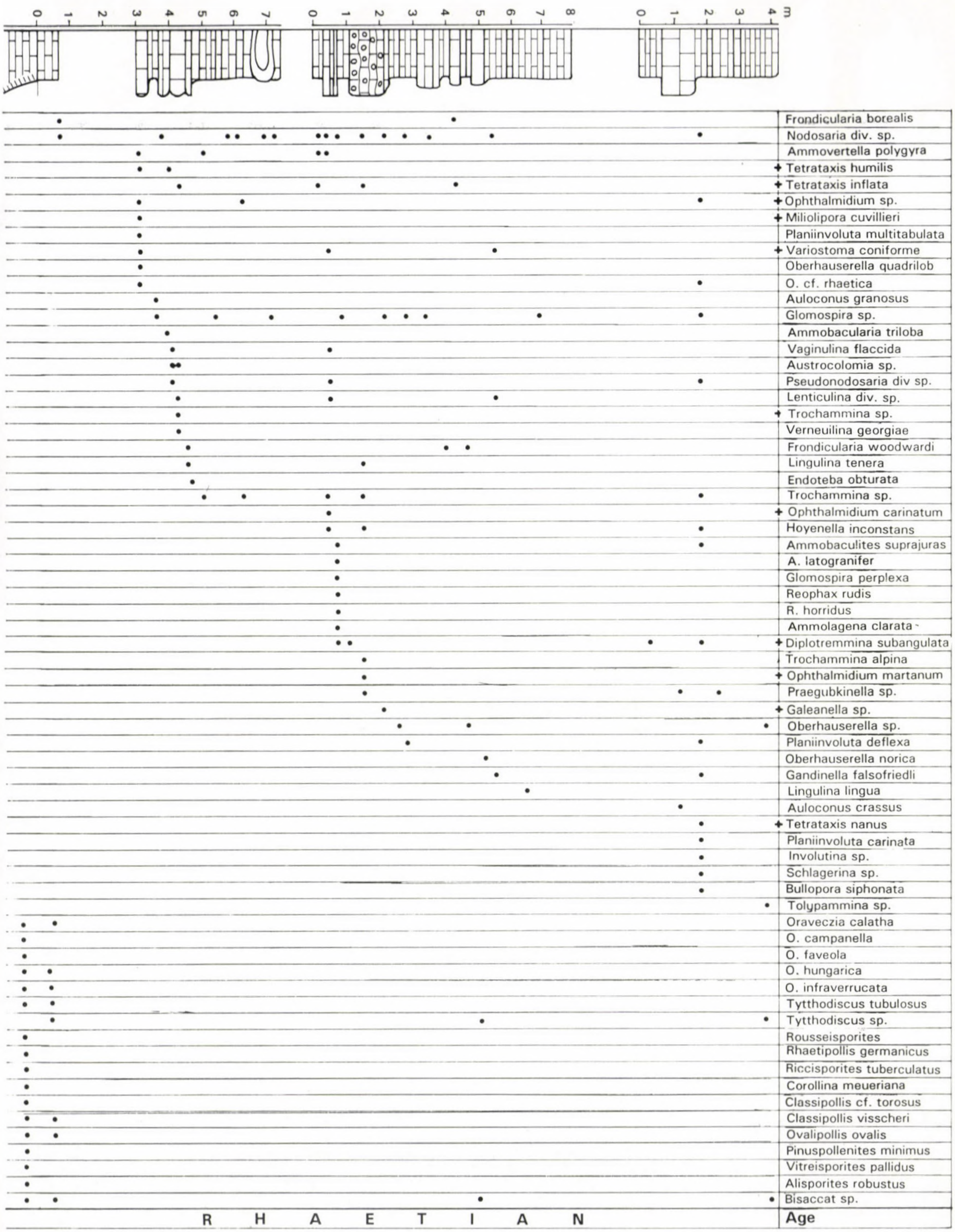
From the 281.6–520.8 m interval of the Csv-1 core, 23 samples were investigated for sporomorphs. Although the majority of the samples contained exines of sporomorphs, they were strongly oxidised and deteriorated; therefore the reliability of the determinations is variable. Sporomorphs of palynostratigraphic value were found only in 8 samples, indicated in Fig. 20.

The assemblage of taxa determined in the samples taken from 474.0 and 461.0 m indicates the upper part of the Tuvalian, on the basis of considerations listed below:

– among sporomorphs found in the Tuvalian formations of the Balaton Highland and the Zsámbék basin the species *Pseudenzonalasporites summus, Pityosporites devolvens* Leschik 1956, *Pinuspollenites minimus* (Couper 1958) Kemp 1970 and *Hevizipollenites samaroides* (Góczán 1996) occur in the samples,

– the genus *Patinasporites*, appearing in the Alpine region in the Julian, is common. Among the strongly oxidised Circumpolles, the recognition of representatives of this genus is the most reliable.

Fig. 19 →
Foraminifera and sporomorphs in the measured sections of the Pokolvölgy quarry. Foraminifera of platform origin are indicated by a cross before the name of the taxon


m
Frondicularia borealis
Nodosaria div. sp.
Ammovertella polygyra
Tetrataxis humilis
Tetrataxis inflata
Ophthalmidium sp.
Miliolipora cuvillieri
Planiinvoluta multitabulata
Variostoma coniforme
Oberhauserella quadrilob
O. cf. rhaetica
Auloconus granosus
Glomospira sp.
Ammobacularia triloba
Vaginulina flaccida
Austrocolomia sp.
Pseudonodosaria div sp.
Lenticulina div. sp.
Trochammina sp.
Verneuilina georgiae
Frondicularia woodwardi
Lingulina tenera
Endoteba obturata
Trochammina sp.
Ophthalmidium carinatum
Hoyenella inconstans
Ammobaculites suprajuras
A. latogranifer
Glomospira perplexa
Reophax rudis
R. horridus
Ammolagena clarata
Diplotremmina subangulata
Trochammina alpina
Ophthalmidium martanum
Praegubkinella sp.
Galeanella sp.
Oberhauserella sp.
Planiinvoluta deflexa
Oberhauserella norica
Gandinella falsofriedli
Lingulina lingua
Auloconus crassus
Tetrataxis nanus
Planiinvoluta carinata
Involutina sp.
Schlagerina sp.
Bullopora siphonata
Tolypammina sp.
Oraveczia calatha
O. campanella
O. faveola
O. hungarica
O. infraverrucata
Tytthodiscus tubulosus
Tytthodiscus sp.
Rousseisporites
Rhaetipollis germanicus
Riccisporites tuberculatus
Corollina meueriana
Classipollis cf. torosus
Classipollis visscheri
Ovalipollis ovalis
Pinuspollenites minimus
Vitreisporites pallidus
Alisporites robustus
Bisaccat sp.
R H A E T I A N
Age

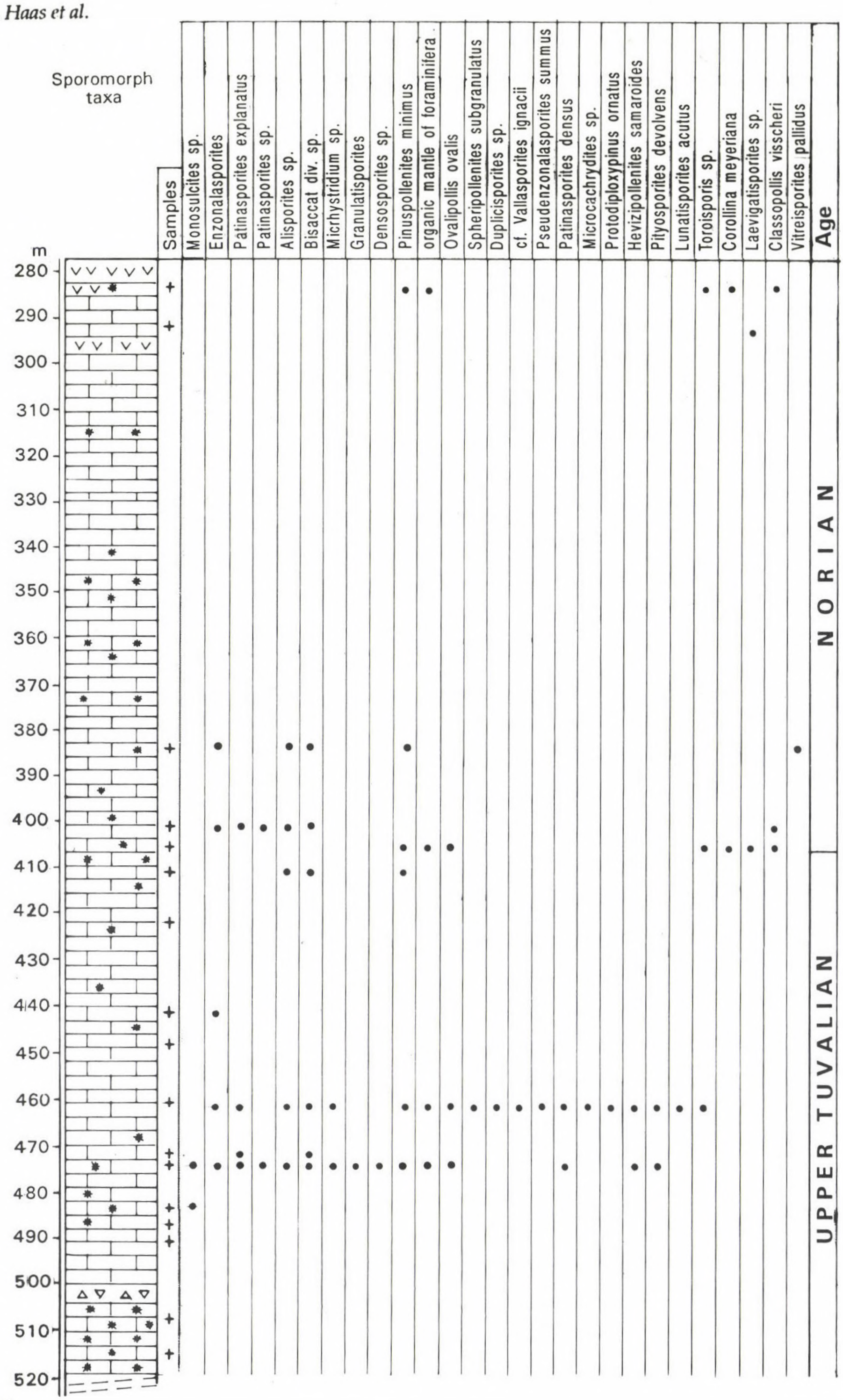


Fig. 20
Sporomorphs in the Csv-1 core between 280 and 520 m

– characteristic of the Julian, and passing upward into the lower Tuvalian, the morpho-species *Staurosaccites quadrifidus* Dolby 1976, *Ovalipollis brutus* Scheuring 1970, and *Sulcatisporites kraeuseli* Mädler 1964 are completely absent.

Based on species appearing in the samples 404.0 and 400.0 m (*Corollina meyeriana* (Kl. 1960, Venk. et Gócz. 1964), *Classopollis visscheri* n. sp.) and the lack of Carnian elements which were present in the previous sample (461.0 m) the higher part of the Csővár Formation (above 404.0 m) is emplaced in the Norian. The main arguments for this classification are as follows:

– the first examples of *Corollina meyeriana* (Kl. 1960) Venk et Gócz 1964, of a Norian–Rhaetian range, appear at this level, together with *Classopollis vissheri* n.sp. (= Forma "B" by Brugman, 1987 – it is described by Góczán in another paper in this volume), which appears in the Norian and disappears at the Norian–Rhaetian boundary.

– a large part of the typically Tuvalian elements are already missing (*Pseudenzonalasporites summus* Scheuring 1970, *Hevizipollenites samaroides* Góczán 1996, *Pityosporites devolvens* Leschnik 1956, and *Patinasporites explanatus* Leschnik 1956; Góczán 1996).

Thus, based on palynostratigraphic data in the Csv-1 core the Carnian/ Norian boundary can be defined at 404.0 m. Above this level the few determinable taxa in the samples of 382.6 m, 308.0 m, 291.4 m, and 281.6 m support the Norian age of this interval.

Today, the basal part of the Pokolvölgy quarry is covered by scree. Previously, however, when the quarry was under operation, these layers were well visible and apt for sampling. In 1961, Oravecz collected samples from the basal layers (see Fig. 10), and Góczán carried out palynological investigations on these samples. It turned out that these argillaceous limestone and marl layers contain large amounts and well-preserved sporomorphs of high paleontological and stratigraphical value.

Oravecz observed 200–300 µm-size alga-cysts on the bedding planes of the 3–4 cm thick grey, organic-rich limestones, locally in great quantity. He took the first photomicrograph of these fossils (the original photomicrograph is shown on Plate V/F, G). The re-investigation and palynological evaluation of these samples are discussed below. The new sporomorph taxa found in this interval are described by Góczán in another paper of the present volume.

The determined taxa of the basal beds are listed in Fig. 20. Typical elements of the sporomorph assemblage are shown on Plates XXI–XXIII. This sporomorph assemblage can be considered as the oldest Rhaetian association, indicating the base of the Rhaetian, close to the Norian–Rhaetian boundary.

The lack of the characteristic and widely-extended Lower Rhaetian sporomorph *Granulopeculati pollis rudis* might be explained by the peculiar relationship between the vegetation area and the sedimentary basin.

The algal cysts, typical elements of the assemblage, are described in this volume (Góczán 1997). In the thesis work of Detre (1970, p. 174) *Oraveczia* nov. gen. and *Tubulodiscus* Góczán et Venk. were still referred to as nomen nudum. The grain which was described by Morbey (1975) as *Tytthodiscus? faveolus* nov.

sp. from the Rhaetian of the Kendelbachgraben (Northern Calcareous Alps), is also classed by us into the genus *Oraveczia* Góczán 1997. Morbey described this form from the "Salzburg Facies", located between the "Kössen Facies" and the "Pre-Planorbis Beds", and corresponding to the upper part of the Rhaetian, although occurrences of the taxon (44 specimens) were reported from the entire succession, from the "Swabian Facies" to the "Pre-Planorbis Beds".

Refering to the fact that some members of the sporomorph assemblage of the Pokolvölgy quarry, namely *Rhaetipollis germanicus* Schulz 1967, *Ricciisporites tuberculatus* Lundblad 1954 and *Classopollis visscheri* nov. sp. already occur in the Norian, we would emplace the basal beds of the quarry in the Upper Norian. However, more than 50% of the members of the operculati group (*Classopollis, Corollina*), even with the absence of *Operculatipollis rudis* in the sporomorph association, strongly supports their lowermost Rhaetian age, i.e. in Rhaetian associations the share of the *Corollina–Classopolis–Operculatipollis* genera may reach 40–96% (Venkatachala and Góczán 1964).

Chronostratigraphy

The chronostratigraphic interpretation of the investigated parts of the Csővár Formation can be summarized as follows:

– according to the sporomorphs the lower part of the studied section exposed by the Csv-1 core between 410.0 and 474.0 m can be assigned to the Upper Tuvalian;

– on the basis of sporomorph appearances and foraminifera taxa (above 354.0 m), the 20.0–404.0 m interval of the Csv-1 core is classified as Norian;

– the uppermost part of the Csv-1 core (above 20.0 m) can be assigned to the Lower Rhaetian, due to the appearance of the typical Rhaetian foraminifera *Hoyenella inconstans*.

– the upper part of the formation, exposed by the Pokolvölgy quarry, is Rhaetian and the zonal index ammonites *Vandaites stuerzenbaumi* and *Choristoceras marshi* prove the presence of both biozones in the succession of the quarry.

Facies model and trend of facies changes

The sedimentological features and the ecologically mixed fossil assemblage of the Csővár Formation indicate a toe-of-slope depositional environment as well as a slope to basin transitional zone. This paleo-environmental setting is indicated by gravity-flow deposits (debris-flow deposits, graded, allodapic limestones, etc.) and frequent occurrences of remnants of platform biota, together with typical pelagic fossil elements.

These characteristics are fairly well comparable with the perfectly exposed and comprehensively studied classic foreslope deposits of the Upper Permian Capitan Reef Complex. Based on the Capitan Reef model (Hurley 1989; Bebout and Kerans 1993) as well as our observations on the Csv-1 core and the section

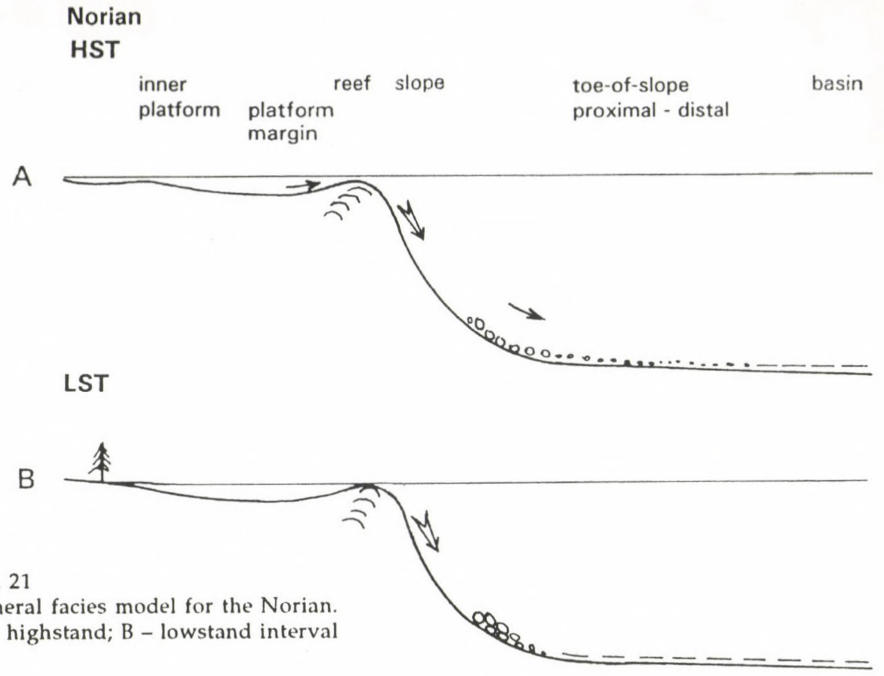


Fig. 21
General facies model for the Norian.
A – highstand; B – lowstand interval

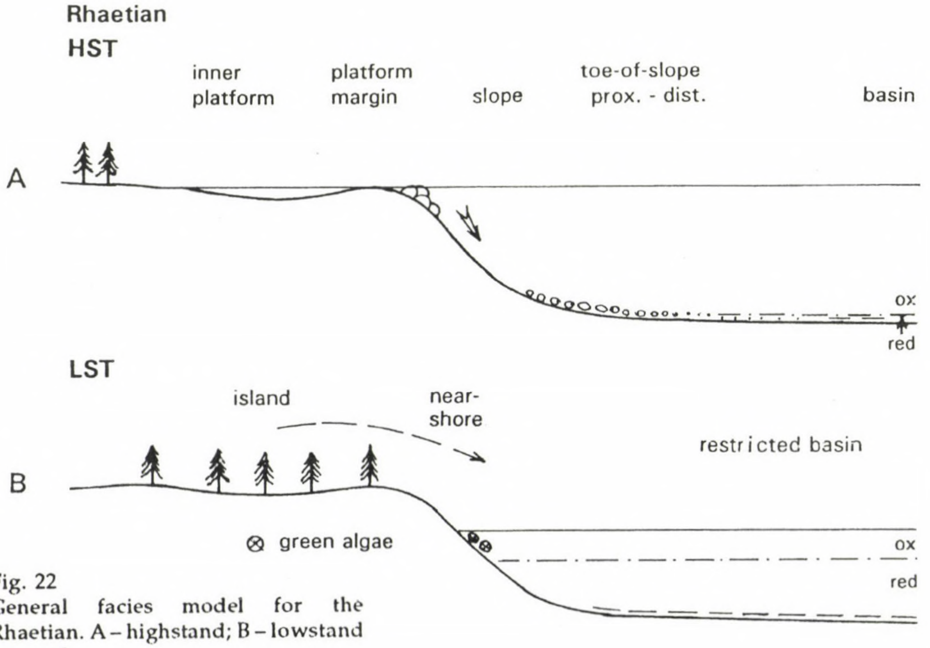


Fig. 22
General facies model for the Rhaetian. A – highstand; B – lowstand interval

of the Pokolvölgy quarry, we propose the general facies models shown in Figs 21 and 22. The main facies units are briefly characterised below.

Carbonate platform – Platform carbonates, coeval with the Csővár Formation (Upper Carnian–Norian) are known in the vicinity of the study area (see Fig. 2). The nearest occurrences of coeval platform carbonates, which can be assigned to the Dachstein Limestone Formation, are located at a distance of 2 km north-westward from the Pokolvölgy quarry. They represent patch-reef and oncoidal facies, indicating the marginal zone of a carbonate platform.

Slope – Lithoclasts in the toe-of-slope facies indicate an early consolidation of the platform-margin deposits, indicating a high-angle slope. If this interpretation is correct, the accumulation of a significant amount of sediments on the slope (upper slope) is not probable (erosional to by-pass slope).

Proximal toe-of-slope – At the toe of the slope, the angle of the slope abruptly decreases. Debris-flow deposits (smaller or larger lithoclasts and rudite–calcarenite-sized bioclast in mudstone–wackestone matrix) characterise this depositional zone.

Distal toe-of-slope – In this zone, the angle of the slope becomes gentle. The predominance of turbiditic deposition characterises this depositional environment. Above erosional surfaces (locally erosional channels), allodapic limestones, showing features of the classic Bouma sequence are visible, alternating with laminitic limestones (fine-grained turbidites) deposited from low-density turbidity currents.

Pelagic basin – In the inner basin, relatively far from the slope, the bottom is practically horizontal. Pelagic oozes ("filamentum" or radiolarian oozes) are typical. However, fine lamination (alternation of calcisilt and mudstone laminae) is common. It can be interpreted as very distal, low-density turbidites. Calcarenitic turbidites are rare and very thin (mm-thick).

The facies interpretations of the studied sequences are shown in Figs 4, 5, 6, 8. The Norian succession exposed by the Csv-1 core shows quite a clear trend. In the lower part of the succession, the predominance of the basinal facies is characteristic, whereas upwards the distal toe-of-slope facies becomes prevailing. This trend suggests progradation of the foreslope. Within this long-term trend, short-term facies changes can also be seen. These may reflect sea-level changes (Fig. 21). The increasing amount of microfossils of platform-interior origin probably indicates the highstand intervals (Reijmer and Everaars 1991; Reijmer et al. 1992; Schlager et al. 1994).

In the early part of the Rhaetian represented by the basal layers of the Pokolvölgy quarry, a remarkable facies change can be assumed. Large amount of green algae and sporomorphs of continental plants suggests a significant sea-level drop, when large parts of the former platforms may have been subaerially exposed and the restriction of intraplatform basin increased (Fig. 22).

The proximal toe-of-slope facies in the higher part of the quarry can be bound to sea-level rise, when reefs or bioherm were formed on the upper slope, whereas large part of the neighbouring platform may have remained in an emerged position providing a relatively large amount of plant remnants for the basin. This model can also explain why the inner-platform foraminifera are missing in the Rhaetian part of the Csővár Formation.

Facies relationships

The Csővár Formation shows very close similarity with the Pötschen Limestone (Schlager 1967), a characteristic Norian formation of the Hallstatt facies unit of the Northern Calcareous Alps and the Inner West Carpathians. As far as the Rhaetian is concerned, the predominantly carbonate lithology of the Csővár Formation significantly differs from that of the contemporaneous Zlambach Marl of the Hallstatt facies unit, although thin marl interlayers and argillaceous limestone beds occur. However, it is worth mentioning that toe-of-slope facies containing detritus of the Dachstein Reef Limestone was also reported from the Zlambach Marl in the Northern Calcareous Alps (Janoschek and Matura 1980). Detailed sedimentological investigations of the Pötschen Limestone section in the Gosau Valley was carried out by Reijmer (1991). His studies revealed that the succession was made up of calciturbidites containing mainly pelagic material, i.e. planktonic or pseudo-planktonic bioclasts in fine carbonate mud. The lithofacies and biofacies in the Csv-1 core show features very similar to those described by Reijmer but due to the lack of continuous core detailed studies on the facies changes and cyclicity could not have been carried out. Calciturbidites in the Rhaethian part of the Csővár Formation show practically the same characteristics as were observed by Reijmer in the Pötschen Limestone, with the exception of debrite interbeds which were not reported in the studied section of the Pötschen Limestone.

The Pötschen Limestone is also known in the Silice Nappe, in a slightly metamorphosed condition in the Torna Nappe in North Hungary (Aggtelek–Rudabánya Mts), and also in the territory of Slovakia. In the Silice Nappe, it consists of grey, thin-bedded cherty limestones with *Halobia* coquina interbeds. In the lower part of the formation intraconglomeratic and allodapic crinoidal intercalations are common (Balogh and Kovács 1981). The most frequent radiolarian and radiolarian-filament microfacies represent basin facies whereas crinoidal coquinas and intraconglomerates indicate toe-of-slope depositional environments. Based mostly on conodonts, the age of the formation is Tuvalian to Early-Middle Norian (Kovács 1986). In the Silice Nappe the Upper Norian–Rhaetian is represented by the Zlambach Marl consisting of brownish-grey marl with grey limestone interlayers. Due to its significantly higher terrigenous content this part of the sequence differs considerably from the Rhaetian part of the Csővár Formation. The Zlambach Marl is overlain by the Liassic "fleckenmergel" facies in the territory of Slovakia. In the Southern Karawanken Range (Austria), a thick Upper Triassic intraplatform basin succession was investigated by Krystyn et al. (1994) and the authors emphasised the similarity of this series with the time-equivalent formations in the NE part of the

Transdanubian Range. The intraplatform basin in the Southern Karawanken region began to form at the Carnian/Norian boundary interval. The Lower and Middle Norian are represented by 200 m - thick cherty dolomites with slump structures, sedimentary breccias and turbidites. It is followed by a 300 m - thick pelagic, platy limestone formation of Late Norian–Rhaetian age. It consists of crinoidal and radiolarian turbidites and bioturbated wackestones but no coarse clastics occur in the upper part of the succession. Just as in the Csővár block there is a continuous transition between the Rhaetian and Lower Jurassic which is represented here by grey, bedded limestones (mudstones).

Conclusions

1. The re-investigation of the 1200 m deep Csővár Csv-1 well revealed that the approximately 600 m thick lower part of the cored sequence should not be included in the normal stratigraphic succession. This segment of the core is made up of tectonically brecciated dolomites interrupted by thin slices of varied lithology and various age from the Middle Triassic to the Upper Cretaceous.

2. In the approximately 600 m thick upper part of the Csv-1 core and the Pokolvölgy quarry, cherty dolomites and, above them cherty limestones, were exposed which can be assigned to the Csővár Formation. Gravity flow deposits and redeposited bioclasts of platform origin indicate proximal as well as distal toe-of-slope and pelagic basin depositional environments.

3. According to the sporomorphs, the lower part of the Csővár Formation (in the Csv-1 core below 410 m) is of Carnian age (Upper Tuvalian between 410–174 m). Based on foraminifera, the middle part of the formation can be classified as Norian (in the Csv-1 core between 404–20 m) and the upper part as Rhaetian, which is also proved by ammonites found in the Pokolvölgy quarry.

4. In the lower (Upper Carnian–Norian) part of the Csővár Formation basinal facies are predominant. Progressing upward the toe-of-slope facies become prevailing, suggesting a long-term trend of platform progradation. In the Rhaetian part of the Csővár Formation the toe-of-slope facies is also characteristic. The large amount of sporomorphs of continental plants and the lack of platform-interior foraminifera suggest subaerial exposure of large parts of the surrounding carbonate platform (Dachstein platform) in this interval.

Acknowledgements

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Plate I

A) Filament microfacies (fragments of thin-shelled bivalves), Core Csv-1 282.0–283.9 m
B) Filament-sponge spicule microfacies, Core Csv-1 477.6 m

Plate II

A) Calciturbidite: Brownish-grey, graded calcarenite (packstone) at the base. It is followed by grey bioturbated mudstone and black argillaceous laminite. Core Csv-1 100.0 m
B) Calciturbidite: Brownish-grey calcarenite (packstone) at the base. It is followed by dark-grey laminitic, light-grey convolute, and again by dark-grey laminitic layers. Core Csv-1 301.5 m

Plate III

A) Radiolarian wackestone microfacies. Core Csv-1 65.3–68.0 m
B) Radiolarian wackestone microfacies. Core Csv-1 300.0 m

Plate IV

A) Bioclasts of platform origin: porostromata algae, *Cayeuxia alpina* Flügel. Core Csv-1 165.2–167.2 m
B) Bioclasts of platform origin: *Tubiphytes* sp., *Baccinella* sp. ("microproblematicum" algae). Core Csv-1 240.5 m

Plate V

A) Bioclastic grainstone. Core Csv-1 185.0 m
B) Bioclastic grainstone with typical foraminifera of the shallow platforms: *Galeanella* sp. Core Csv-1 283.7–284.2 m

Plate VI

A) Sponge-spicule wackestone microfacies. Core Csv-1 57.0 m
B) Sponge-spicule wackestone microfacies. Core Csv-1 73.3–76.7 m

Plate VII

A) Bioclasts of reefal origin. Sphinctozoa: *Cystothalamia* cf. *minima*. Core Csv-1 114.8–117.0 m
B) Bioclasts of reef origin: *Poriferitubus buseri* Senowbari-Darian. Core Csv-1 128.3–130.3 m

Plate VIII

A) Channelized calciturbidites. Note the outpinching layers at the hammer. Pokolvölgy quarry, lower section. Core Csv-1
B) Slump structures in the lower section of the Pokolvölgy quarry. Core Csv-1

Plate IX

A) Base of a calciturbidite layer. Note the truncated top of the underlying layer. Pokolvölgy quarry, bed No. 16.
B) Distal calciturbidites. Bioclastic packstone calcarenite and mudstone-wackestone microlayers alternate. Pokolvölgy quarry, bed No. 12

Plate X

A) Distal calcitrubidites. Bioclastic packstone calcarenite and mudstone-wackestone microlayers alternate. Pokolvölgy quarry, bed No. 34
B) Very distal calciturbidites. Bioclastic-peloidal calcisilt. Wackestone and mudstone microlayers alternate. Pokolvölgy quarry, bed No. 46

Plate XI

Organic microfacies

A) Microfacies-type 2. – Coal grains are abundant; every fraction of the coal grains is angular or subangular; they are of wood or filamentous alga origin. Pollen is also abundant; they are strongly oxidised with decayed exine. Organic test of foraminifera are also common. Pokolvölgy quarry, basal layers
B) Microfacies-type 1. – Subangular shape of the coarse fraction and large filamentous alga remnants indicate the photic zone and a low-energy environment. Pokolvölgy quarry, basal layers
C) Microfacies-type 1. – Well-preserved algal cyst (*Oraveczia infraverrucata*) with medium and small-sized coal grains and several oxidised pollen exines. Pokolvölgy quarry, basal layers
D) Microfacies-type 1. – Coal grains are abundant, angular or subangular. The organic test of the rotaloid foraminifera is well-preserved. Algal cysts (*Oraveczia*) show strong oxidisation and fracturation, suggesting reworking. Pokolvölgy quarry, basal layers

Plate XII

Organic microfacies

A) Microfacies-type 1. – Coarse fraction of coal grains of wood origin is angular or subangular with remnants of autochthonous filamentous algae and well-preserved organic tests of foraminifera. Pokolvölgy quarry, basal layers
B) Microfacies-type 2. – Coal grains of wood origin are common, strongly abraded, but undecayed with small and medium-sized fragments of filamentous algae. Pokolvölgy quarry, middle segment
C) Microfacies-type 3. – Pollen with decayed exine are abundant, a small number of remnants of filamentous algae are common, coarse angular coal grains and great number of strongly abraded medium-sized and decayed small-sized coal grains are characteristic. Core Csv-1 461.0 m
D) Microfacies-type 3. – Fair amount of medium and small undecayed grains of wood origin and a few fragments of filamentous algae; organic tests of microplankton also occur. Pollen and colloid material are absent. Pokolvölgy quarry, upper segment

Plate XIII

Organic microfacies

A) Microfacies-type 4. – Coal grains are abundant and unsorted. fragments of filamentous algae are poor. The few sporomorphs are deformed. Organic tests of foraminifera also occur. Core Csv-1 281.6 m
B) Microfacies-type 6. – Dominance of colloidal material formed by fracturation and decay of plants characterises this microfacies type. Small coal grains are rare; sporomorphs and plankton are completely missing. Core Csv-1 385.0 m

Plate XIV

Foraminifera

A) *Pseudonodosaria pupoides* (Born.), Pokolvölgy quarry, bed No. 24/a, 160x
B) *Frondicularia borealis* (Tappan), Pokolvölgy quarry, bed No. 2, 160x
C) *Lingulina tenera concosta* Kristan-Tollman, Pokolvölgy quarry, bed No. 10, 64x
D) *Hoyenella inconstans* (Michalík, Jendrejakova et Borza), Pokolvölgy quarry, bed No. 20, 64x
E) *Oberhauserella rhaetica* (Kristan-Tollmann), Pokolvölgy quarry, bed No. 3, 64x
F) *Oberhauserella rhaetica* (Kristan-Tollmann), Pokolvölgy quarry, bed No. 41, 64x
G) *Praegubkinella* sp., Pokolvölgy quarry, bed No. 24/c, 160x
H) *Hoyenella inconstans* (Michalík, Jendrejakova et Borza), Pokolvölgy quarry, bed No. 41, 160x
I) *Preaegubkinella* sp., Pokolvölgy quarry, bed No. 12, 64x
J) *Pseudonodosaria lahusani* (Uhlig), Pokolvölgy quarry, bed No. 41, 64x
K) *Oberhauserella quadrilobata* Fuchs and *Auloconus granosus* (Frentzen), Pokolvölgy quarry, bed No. 5, 64x

Plate XV

Foraminifera

A) *Reophax rudis* Kristan-Tollmann, Pokolvölgy quarry, bed No. 22, 64x
B) *Reophax* cf. *horridus* (Schwager), Pokolvölgy quarry, bed No. 22, 64x
C) *Ammobacularia* cf. *triloba* Kristan-Tollmann, Pokolvölgy quarry, bed No. 20, 64x
D) *Ammobaculites latogranifer* Kristan-Tollmann, Pokolvölgy quarry, bed No. 22, 64x
E) *Endoteba* ex. gr. *obturata* (Brönn. et Zan.), Pokolvölgy quarry, bed No. 11, 64x
F) *Planiinvoluta*? sp., Pokolvölgy quarry, bed No. 27, 160x

Plate XVI

Foraminifera

A) *Glomospira* cf. *perplexa* Franke, Pokolvölgy quarry, bed No. 22, 64x
B) *Reophax rudis* Kristan-Tollmann, Pokolvölgy quarry, bed No. 22, 64x
C) Sessile miliolids, Pokolvölgy quarry, bed No. 8, 64x
D) *Verneuilina georgiae* Terquem, Pokolvölgy quarry, bed No. 9/a, 64x
E) Sessile miliolids, Pokolvölgy quarry, bed No. 22, 64x
F) *Ammolagena*? sp., Pokolvölgy quarry, bed No. 16, 64x
G) *Miliolipora cuvillieri* Brönn. et Zaninetti, Pokolvölgy quarry, bed No. 3, 64x
H) *Bullopora siphonata* Kristan-Tollmann, Pokolvölgy quarry, bed No. 41, 64x
I) *Tetrataxis inflata* Kristan, Pokolvölgy quarry, bed No. 41, 64x
J) *Ammolagena clavata* (Jones-Parker), Pokolvölgy quarry, bed No. 22, 64x

Plate XVII

Foraminifera

A) *Auloconus crassus* (Kristan), Pokolvölgy quarry, bed No. 40, 64x
B) *Variostoma coniforme* Kristan-Tollmann, Pokolvölgy quarry, bed No. 3, 64x
C) *Gaudryina* cf. *triadica* Kristan-Tollmann, Pokolvölgy quarry, bed No. 6, 64x
E) *Diplotremmina subangulata* Kristan-Tollmann, Pokolvölgy quarry, bed No. 22, 64x
F) *Variostoma coniforme* Kristan-Tollmann, Pokolvölgy quarry, bed No. 11, 64x
G) *Trochammina januensis* Page, Pokolvölgy quarry, bed No. 24/c, 64x
H) *Variostoma coniforme* Kristan-Tollmann, Pokolvölgy quarry, bed No. 20, 64x

Plate XVIII

Foraminifera

A) *Variostoma crassum* Kristan-Tollmann, Core Csv-1 79.1–81.5 m, 50x
B) *Variostoma crassum* Kristan-Tollmann, Core Csv-1 22.0 m, 25x
C) *Diplotremmina placklesiana* Kristan-Tollmann, Core Csv-1 240.5 m, 50x
D) *Variostoma crassum* Kristan-Tollmann, Core Csv-1 147.5 m, 50x
E) *Diplotremmina subangulata* Kristan-Tollmann, Core Csv-1 114.8–117.0 m, 50x
F) *Variostoma crassum* Kristan-Tollmann, Core Csv-1 79.1–81.5 m, 50x
G) *Diplotremmina subangulata* Kristan-Tollmann, Core Csv-1 134.0–136.0 m, 50x

Plate XIX

Foraminifera

A) *Planiinvoluta multitabulata* Kristan-Tollmann, Core Csv-1 254.0 m, 125x
B) *Endotriada izjumiana* (Dain), Core Csv-1 240.5 m, 50x
C) *Kaeveria fluegeli* (Zaninetti et al.), Core Csv-1 295.0 m, 50x
D) *Tetrataxis inflata* Kristan, Core Csv-1 79.1–81.5 m, 50x
E) *Miliolina cuvillieri* Bronn. et Zan., Core Csv-1 134.9–136.7 m, 50x
F) *Kaeveria fluegeli* (Zan. et al.), Core Csv-1 134.9–136.7 m, 50x
G) *Tetrataxis nanus* Kristan, Core Csv-1 114.2–115.7 m, 50x
H) *Kaevera fluegeli* (Zaninetti et al.), Core Csv-1 167.0 m, 50x
I) *Lenticulina* cf. *opercula* (Crick et Shernborn), Core Csv-1 165.2–167.0 m, 50x

Plate XX

Foraminifera

A) *Hoyenella incontans* (Michalik, Jendrejakova et Borza), Core Csv-1, 20.0 m, 125x
B) *Galeanella* cf. *variabilis* Zaninetti et al., Core Csv-1, 104.0–141.0 m, 50x
C) *Galeanella* cf. *panticae* Zaninetti et Brönnimann, Core Csv-1, 88.9–89.1 m, 50x
D) *Ophthalmidium leischneri* (Kristan-Tollmann), Core Csv-1, 134.9–136.7 m, 50x
E) *Galeanella tollmanni* (Kristan-Tollmann), Core Csv-1, 88.9–89.1 m, 50x
F) *Sigmoilina schafera* Ducret, Core Csv-1, 134.9–136.7 m, 50x
G) *Ophthalmidium triadicum* (Kristan), Core Csv-1, 114.8–117.3 m, 50x

Plate XXI

Sporomorphs (M = 1000x)

A–B) *Rousseisporites* sp., Pokolvölgy quarry, basal layer (1961)
C–E) *Rhaetipollis germanicus* Schulz 1967, Pokolvölgy quarry, basal layer (1961)
F) *Ricciisporites tuberculatus* Lundblad 1954, Pokolvölgy quarry, basal layer
G–H) *Classopollis visscheri* nov. sp., Pokolvölgy quarry, basal layer (1982)
I) *Corollina meyeriana* (Kl. 1960) Venk. et Gócz. 1962, Pokolvölgy quarry, basal layer
J) *Classopollis visscheri* nov. sp., Pokolvölgy quarry, basal layer
K) *Classopollis* cf. *torosus* (Reiss. 1950) Venkatachala et Góczán 1962, Pokolvölgy quarry, basal layer 1961

Plate XXII

Sporomorphs (M = 1000x)

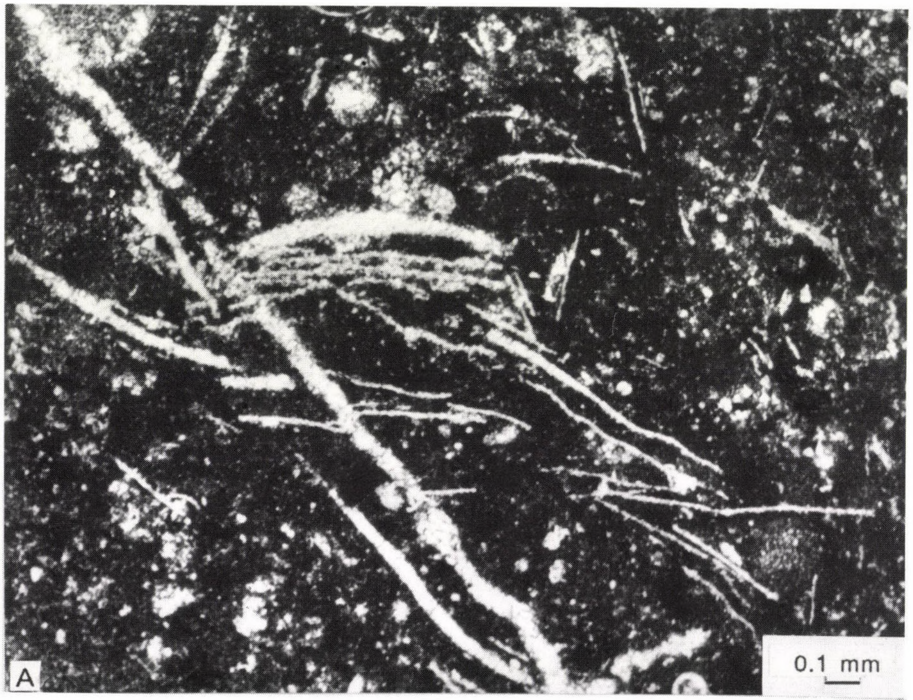
A) *Pseudenzonalasporites* cf. *summus* Scheuring 1970. Core Csv-1, 461.0 m
B) *Patinasporites explanatus* (Leschik 1956) Gócz. 1996. Core Csv-1, 474.0 m
C) *Patinasporites densus* Leschik 1956. Core Csv-1, 461.0 m
D) *Patinasporites* cf. *densus* Leschik 1956. Core Csv-1, 4 61.0 m
E) Alisporites toralis (Leschik 1956) Clarke 1965. Core Csv-1, 461.0 m
F) *Hevizipollenites* cf. *samaroides* Gócz. 1996. Core Csv-1, 461.0 m
G–H) *Ovalipollis ovalis* Krutzsch 1955. Core Csv-1, 461.0 m
I) *Microcachryidites* sp., Core Csv-1, 461.0 m

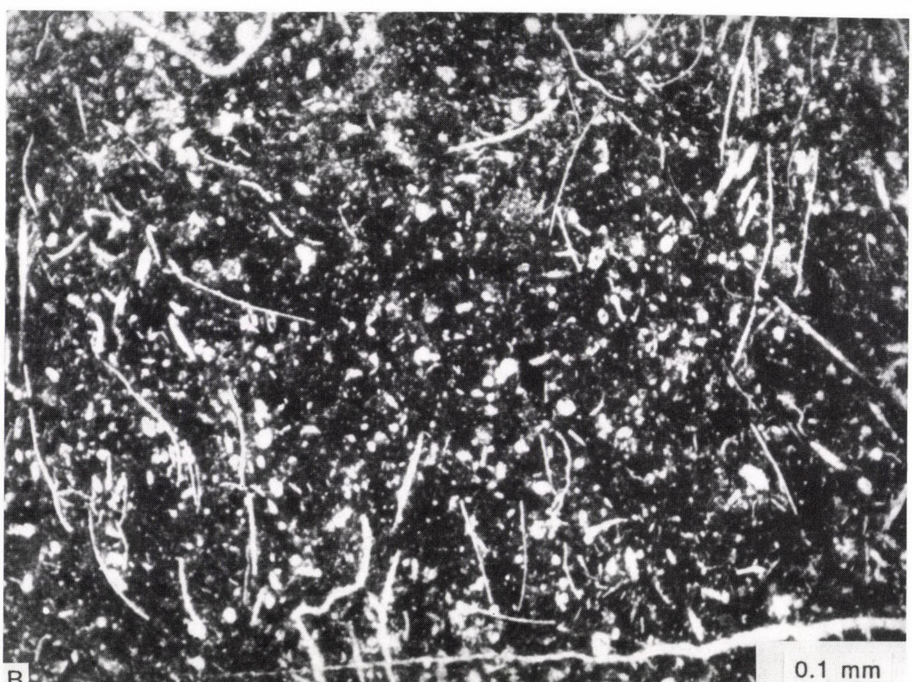
Plate XXIII

Sporomorphs (M = 1000x)

A) *Hevizipollenites samaroides* Góczán 1996. Core Csv-1, 461.0 m
B) *Pinuspollenites minimus* (Couper 1958) Kemp 1970. Core Csv-1, 461.0 m
C–D) *Pityosporites devolvens* Leschik 1956. Core Csv-1, 461.0 m
E) *Protodiploxypinus ornatus* Pautsch 1973. Core Csv-1, 461.0 m
F–G) *Pinuspollenites minimus* (Couper 1958) Kemp 1970. Core Csv-1, F = 461.0 m; G = 474.0 m

Plate I





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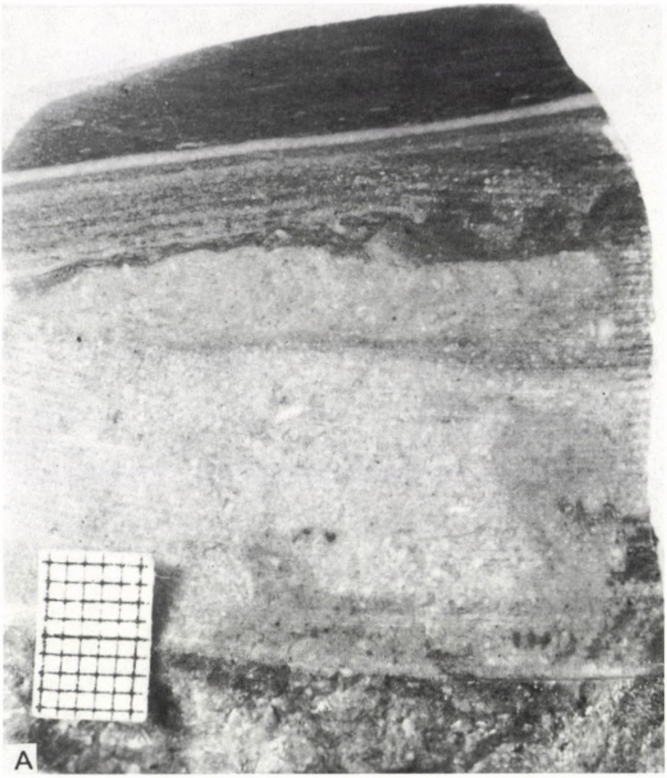

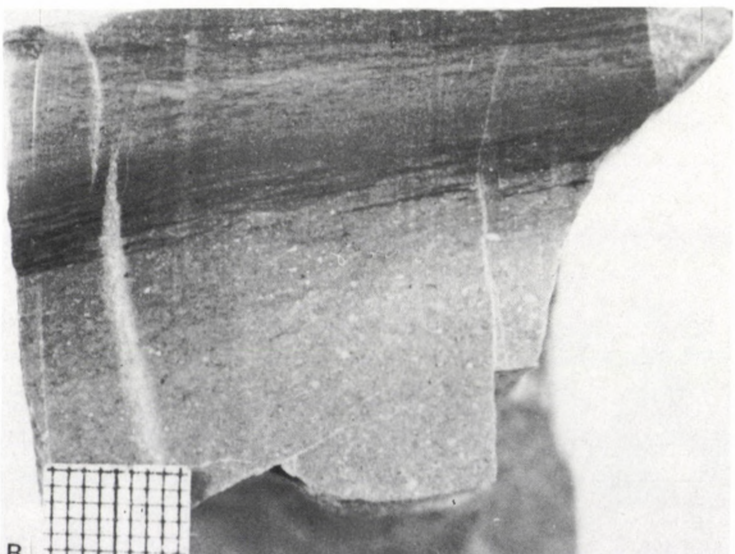

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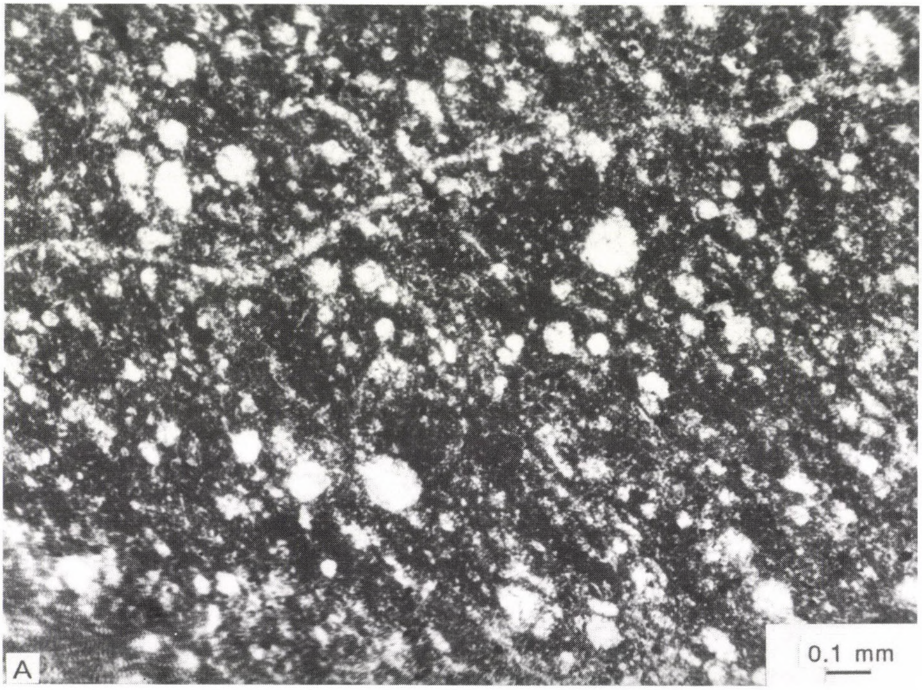

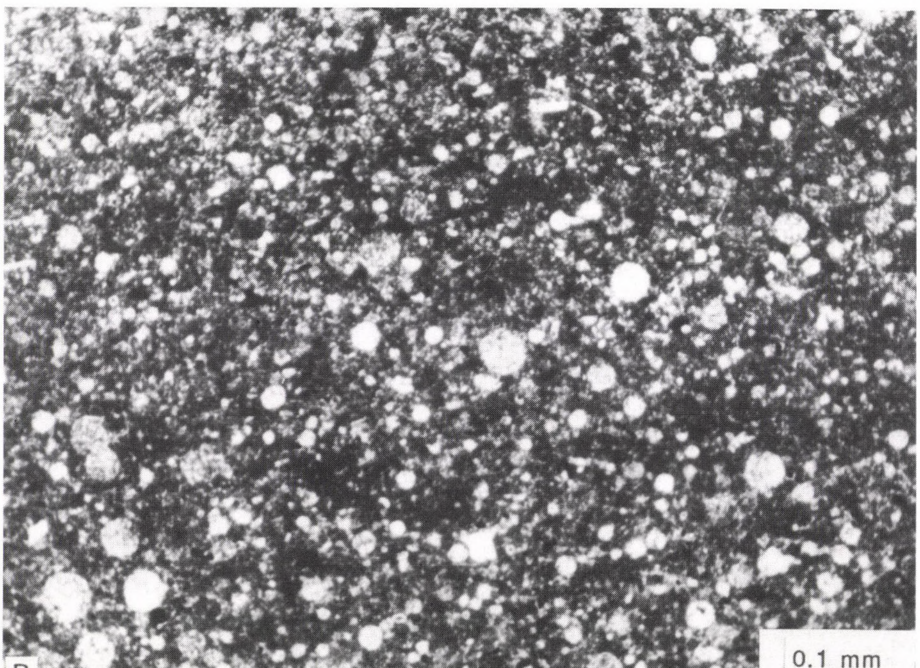

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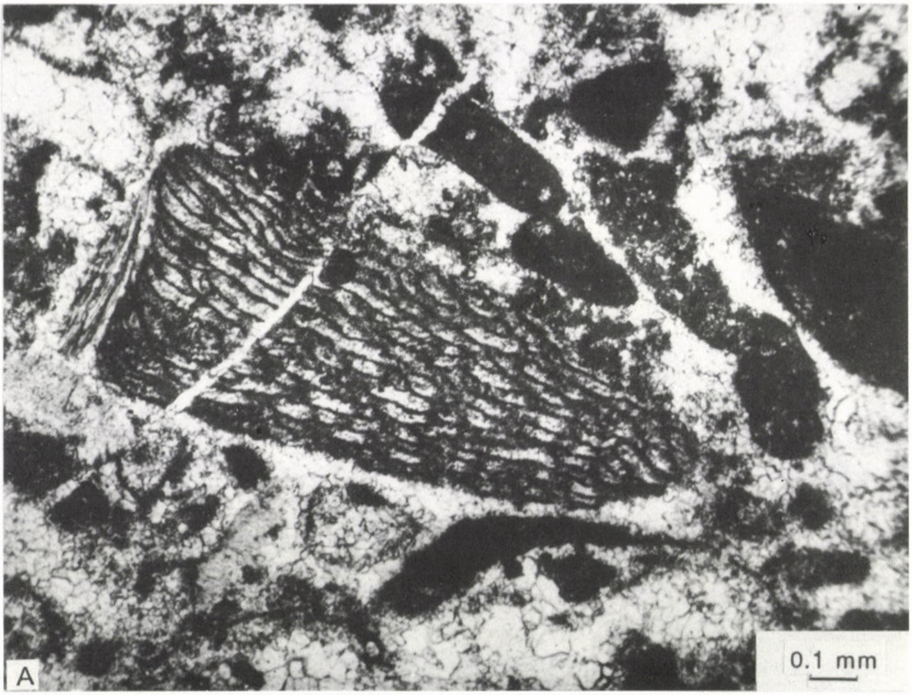

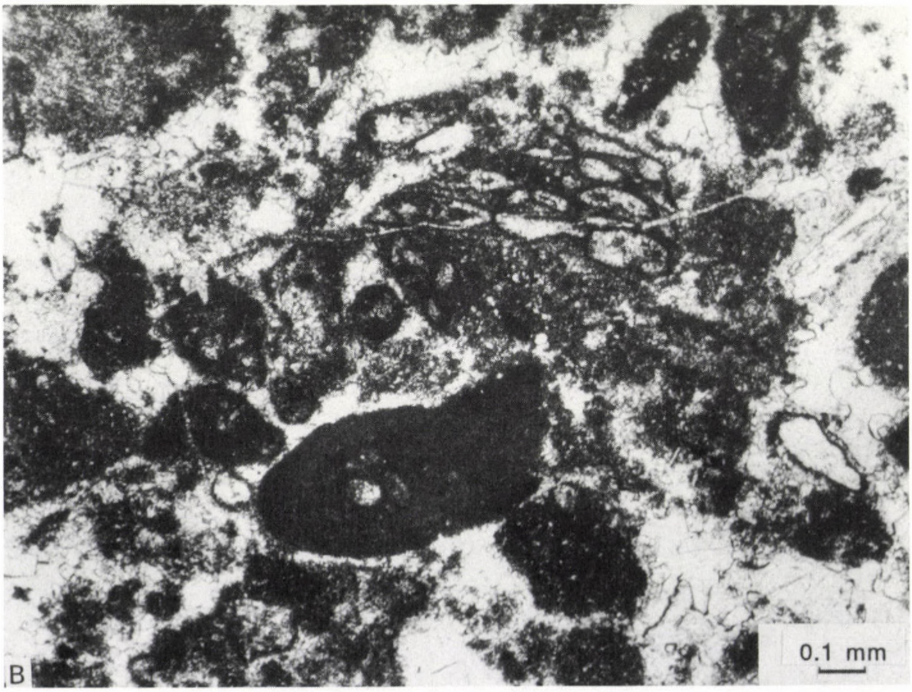

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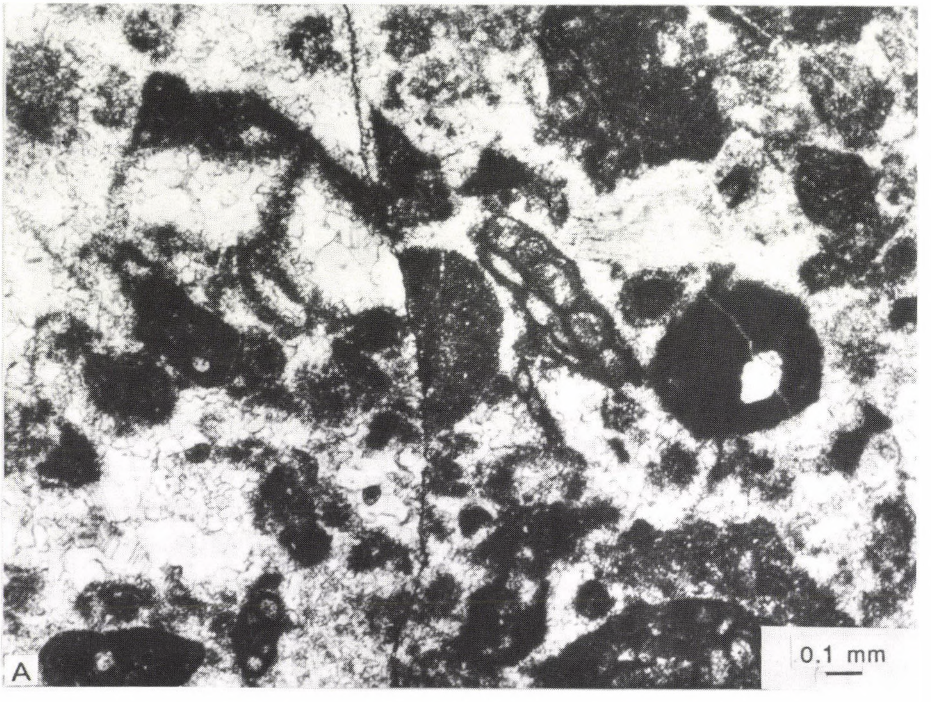

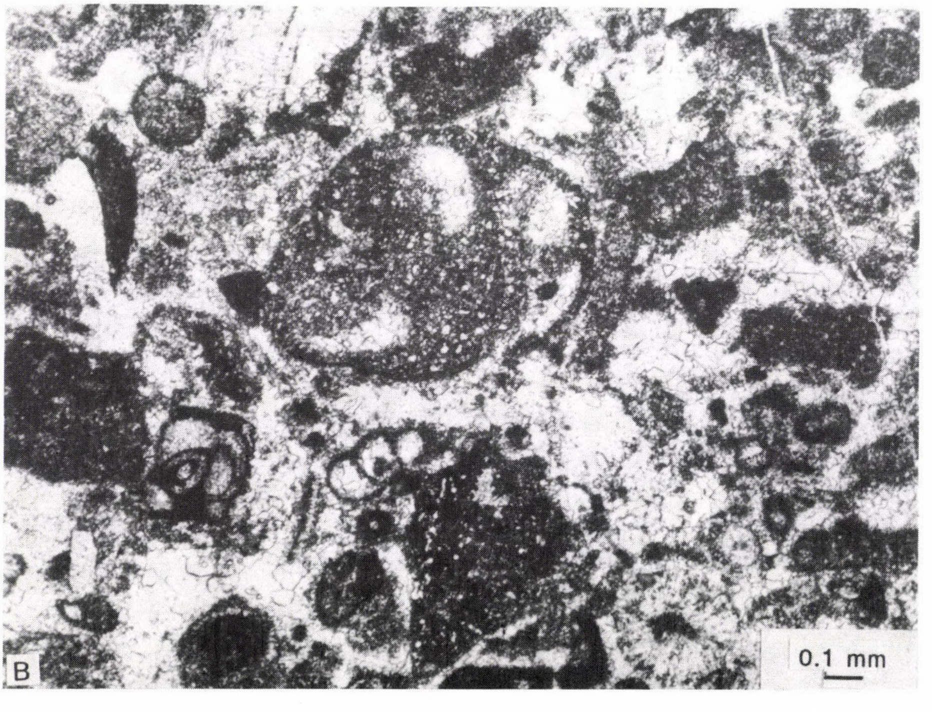

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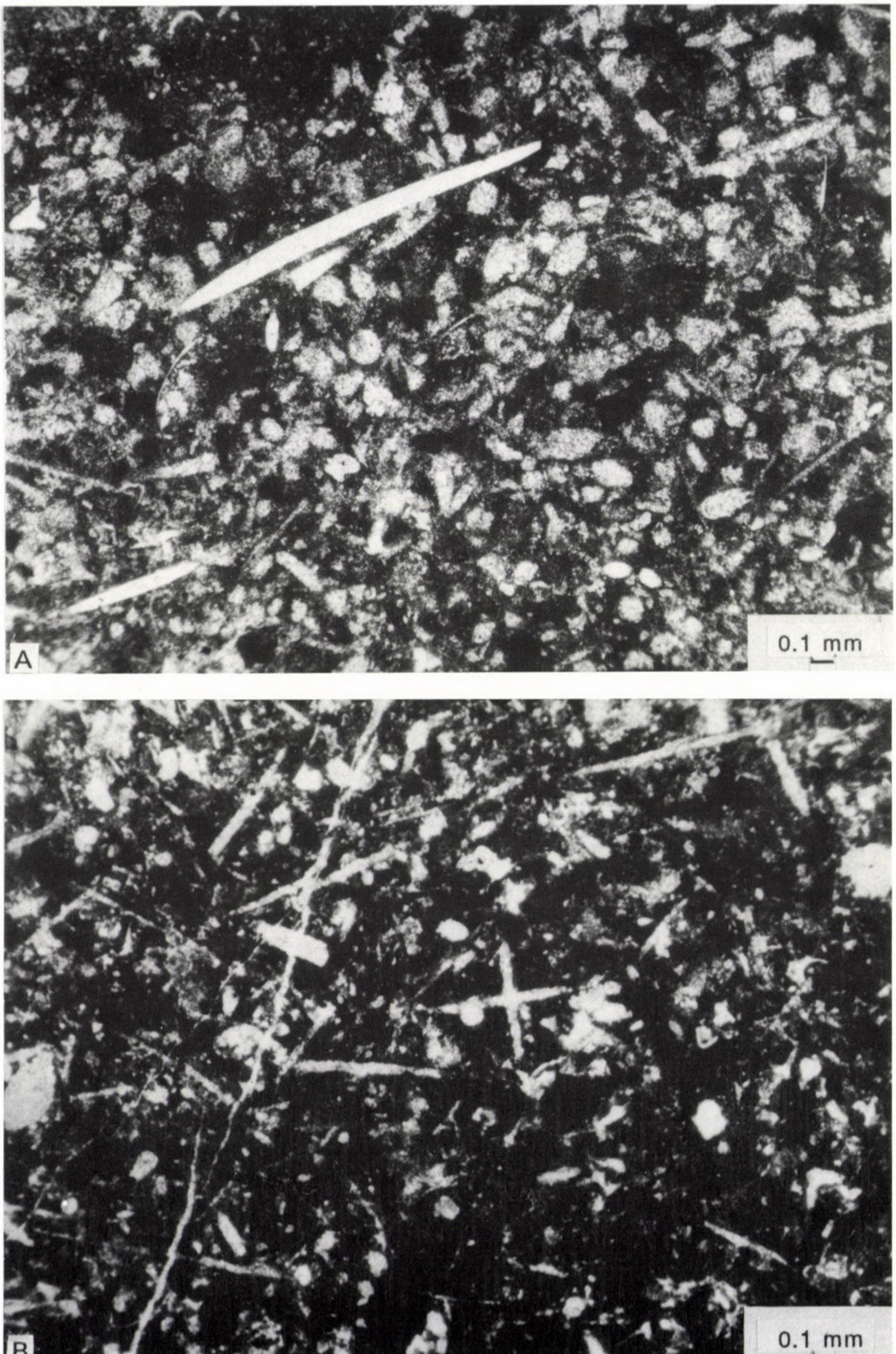

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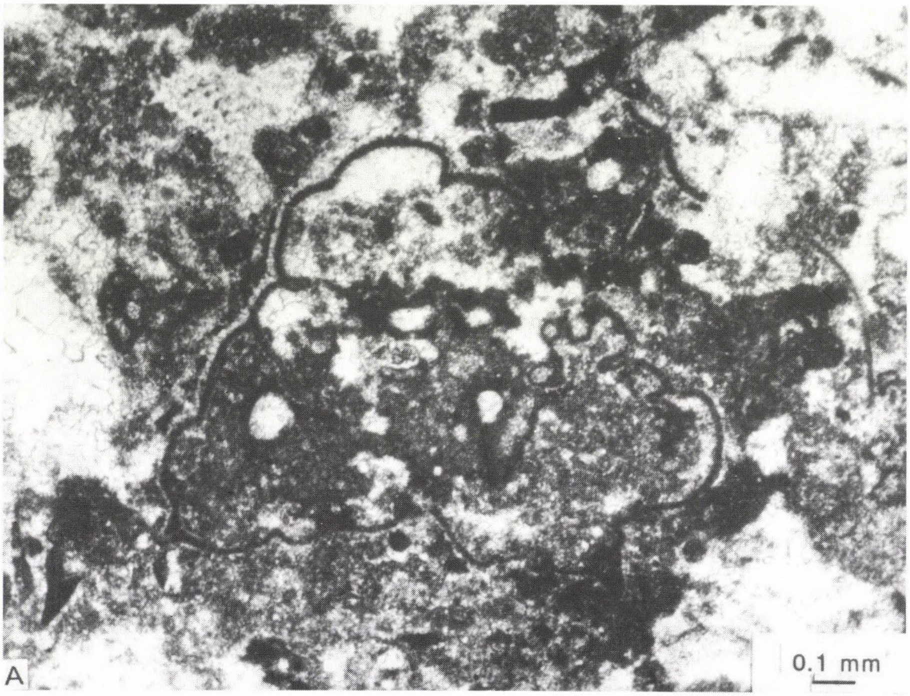


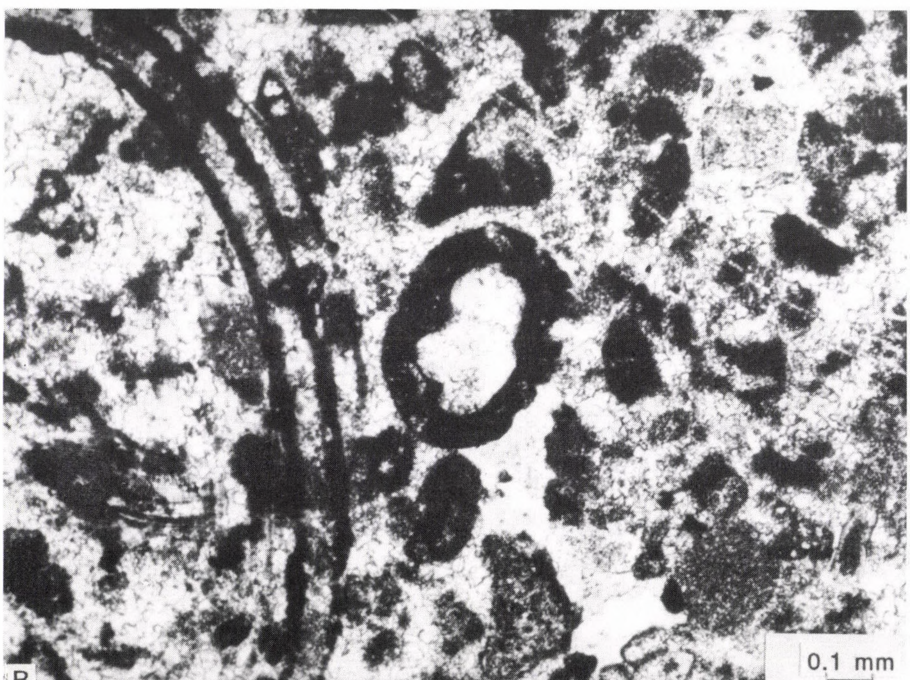
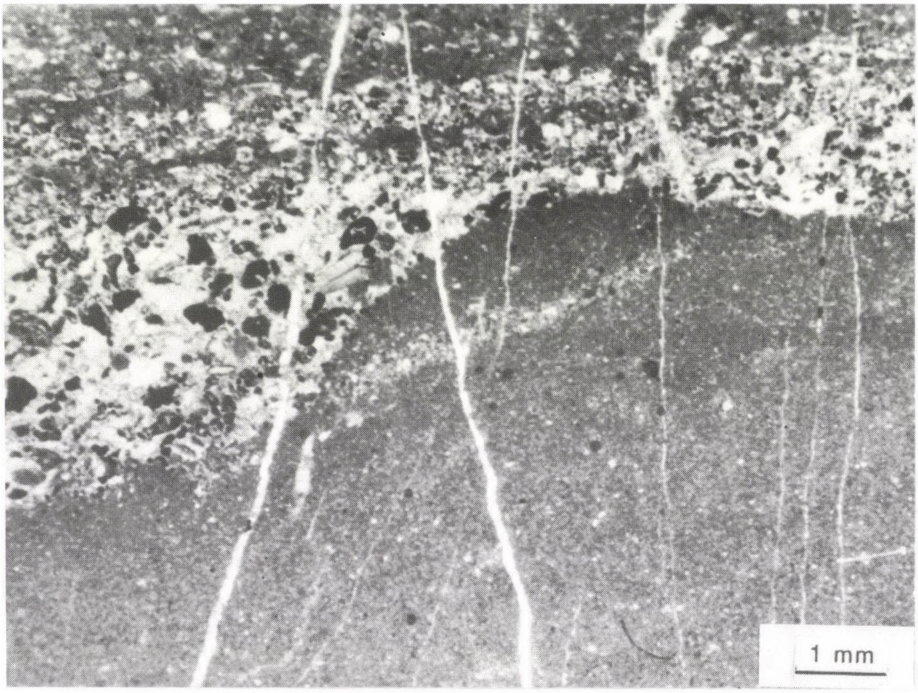


Plate VIII

Plate IX



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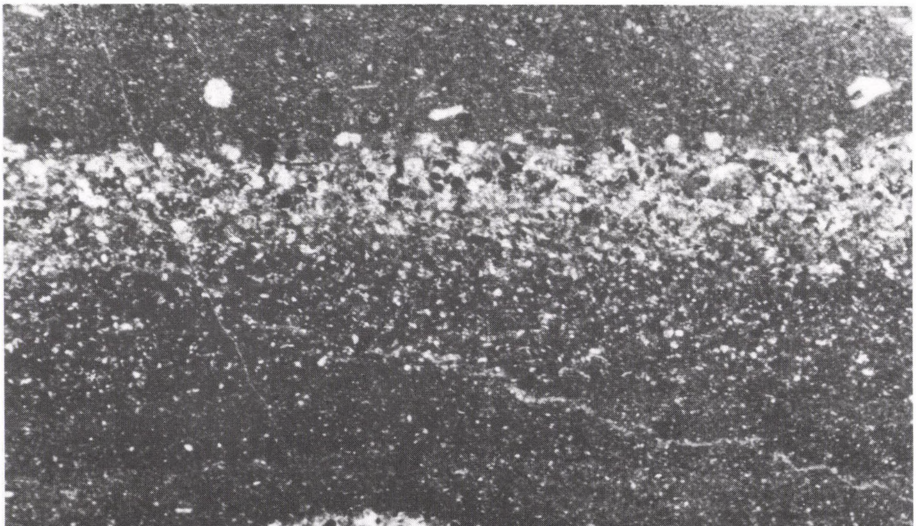
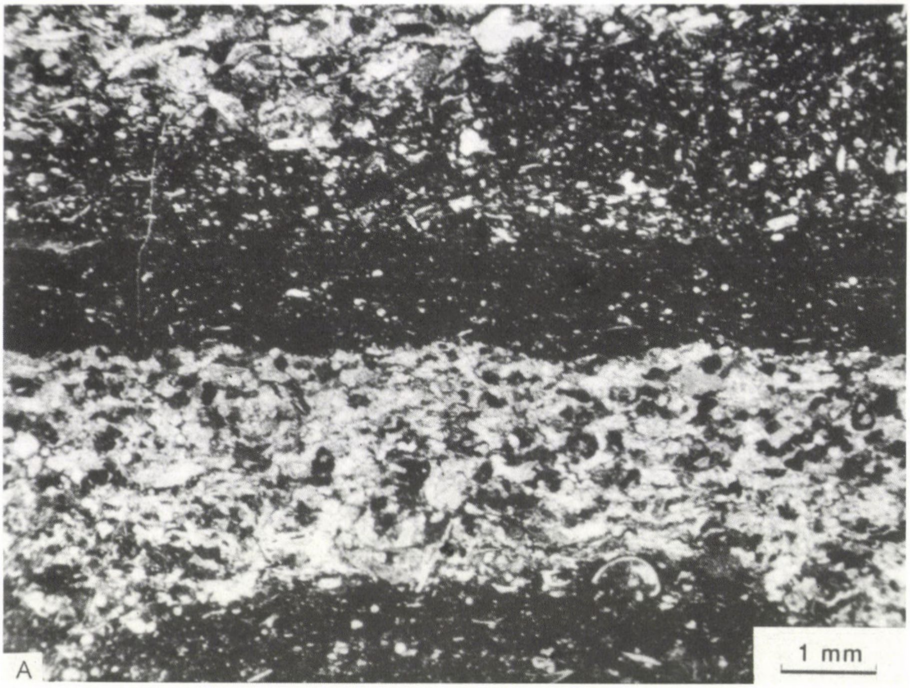
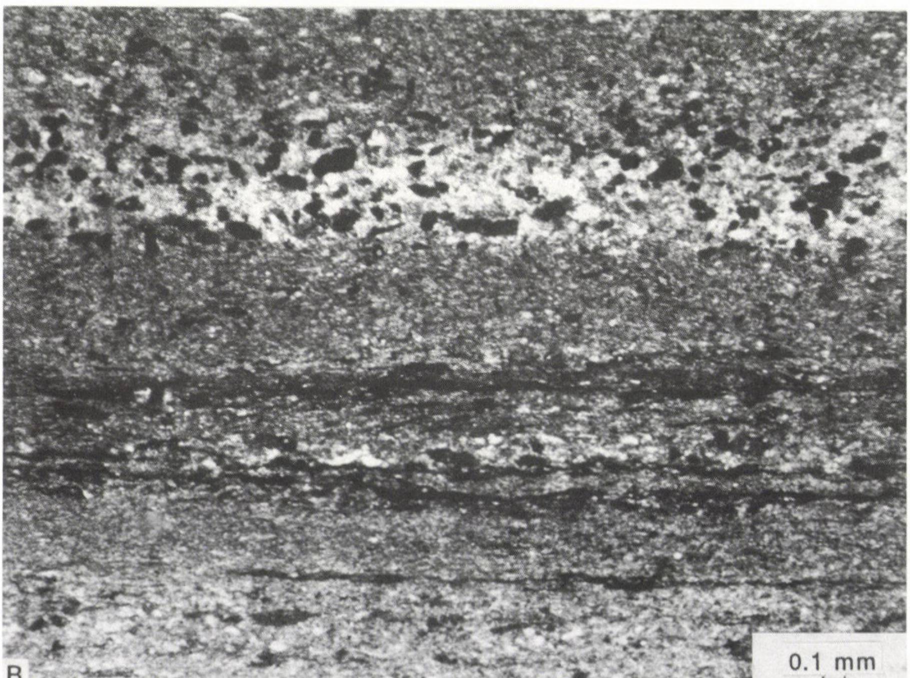


B

Plate X





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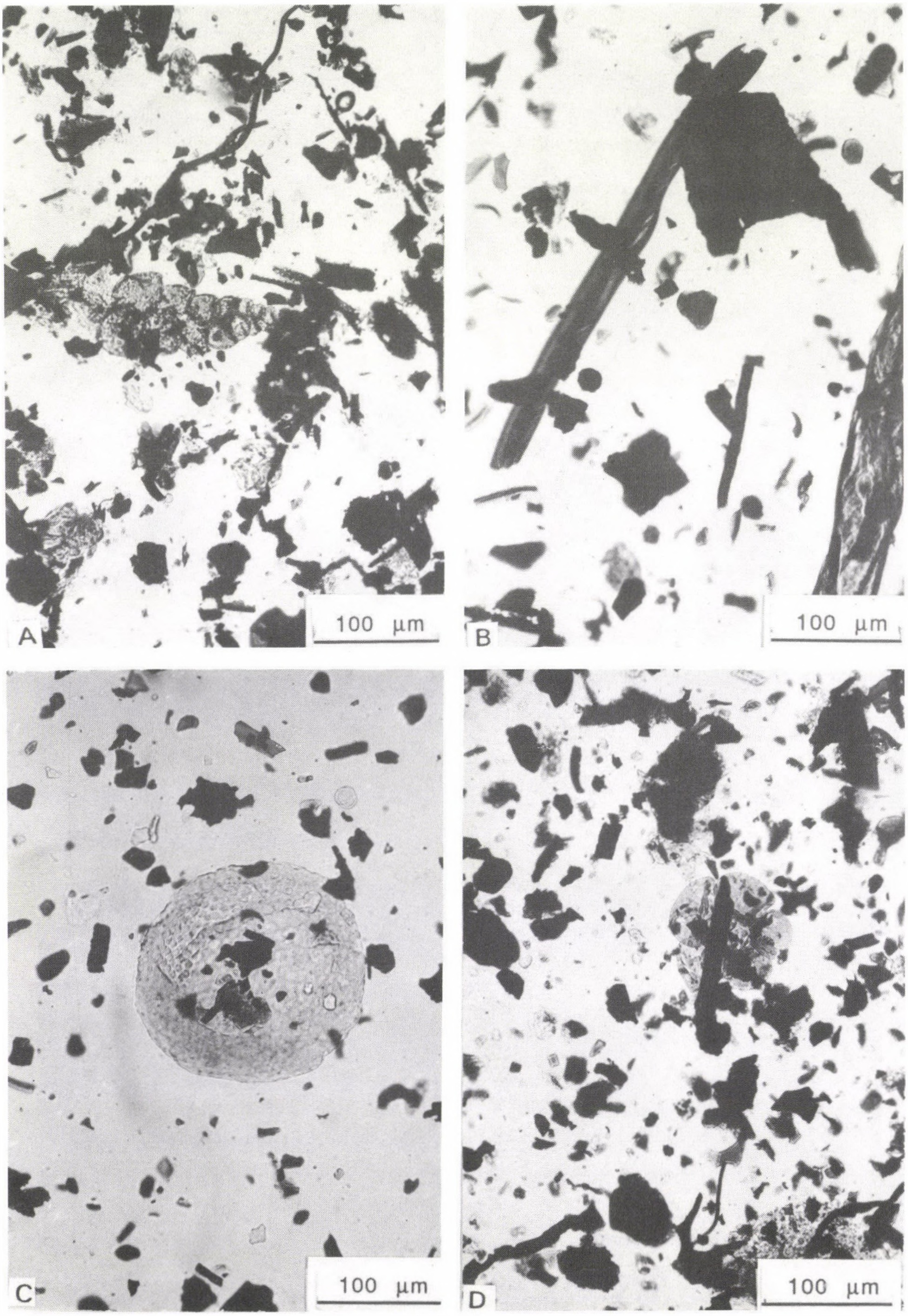

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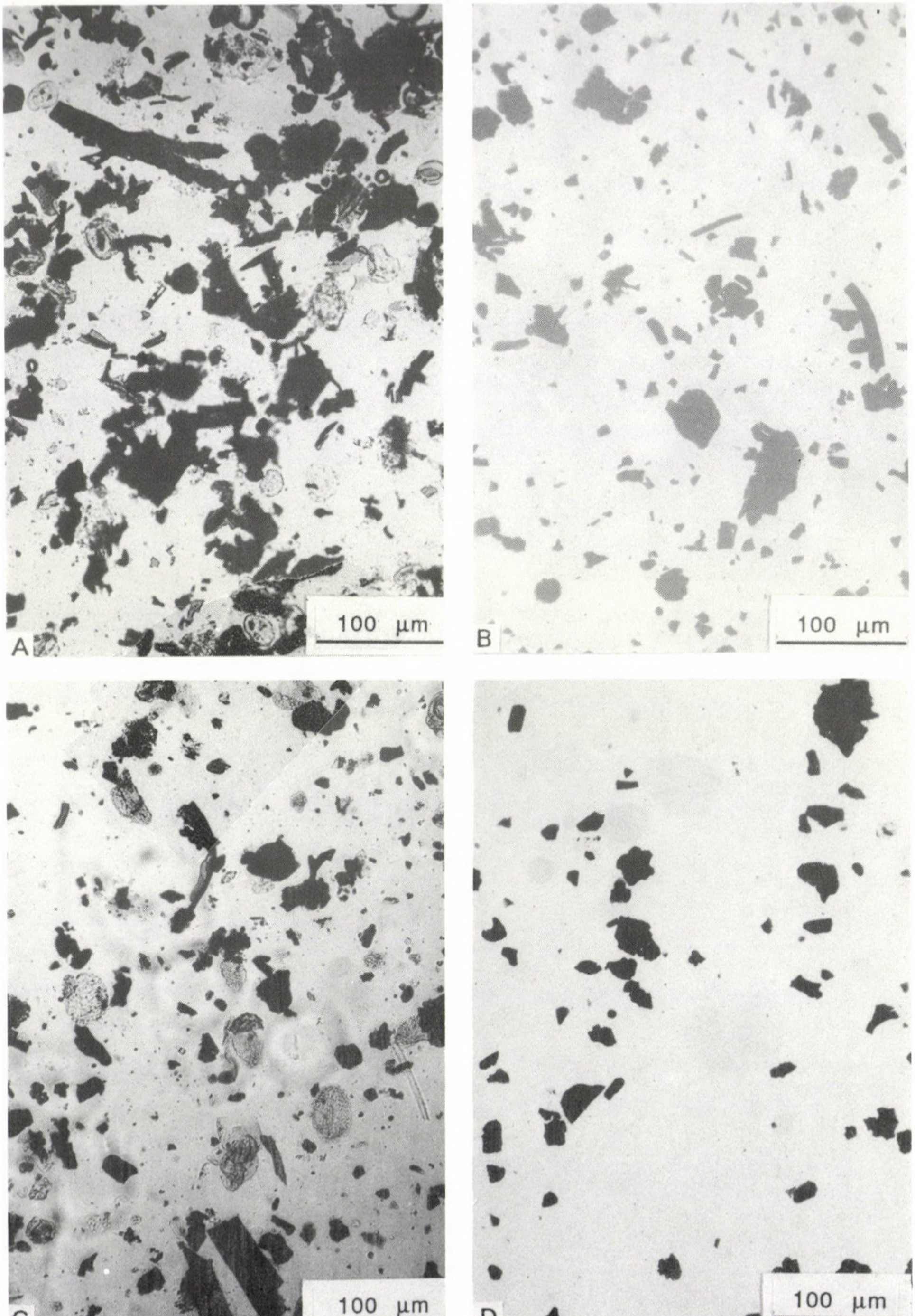

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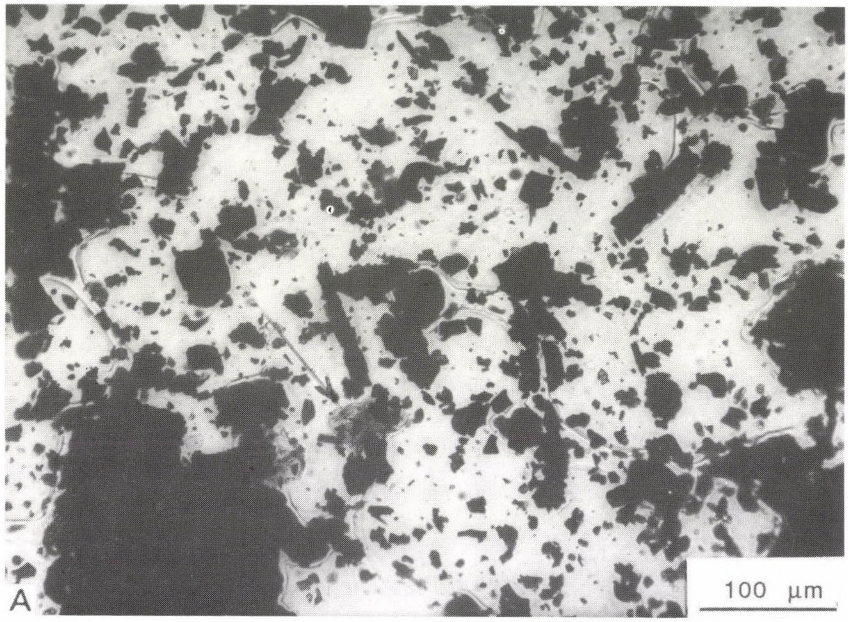

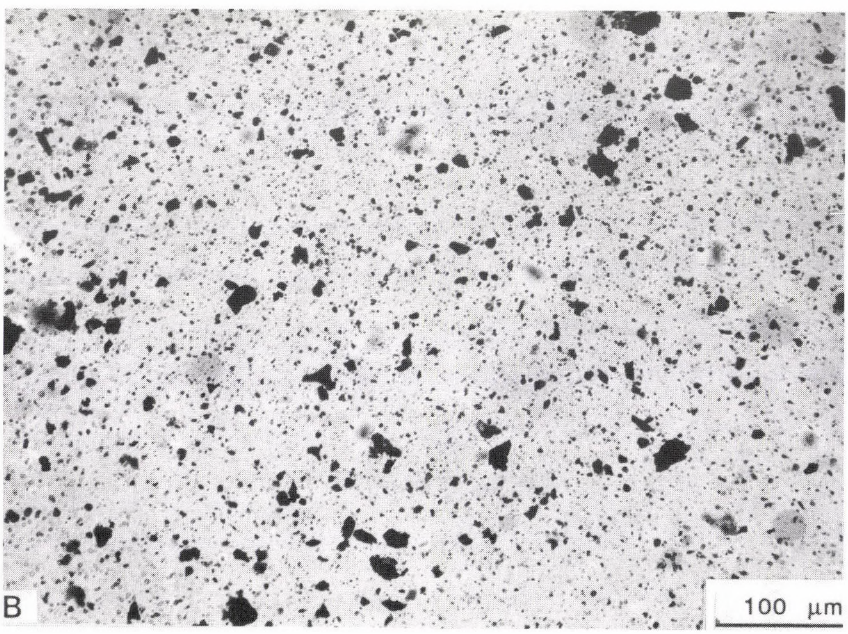

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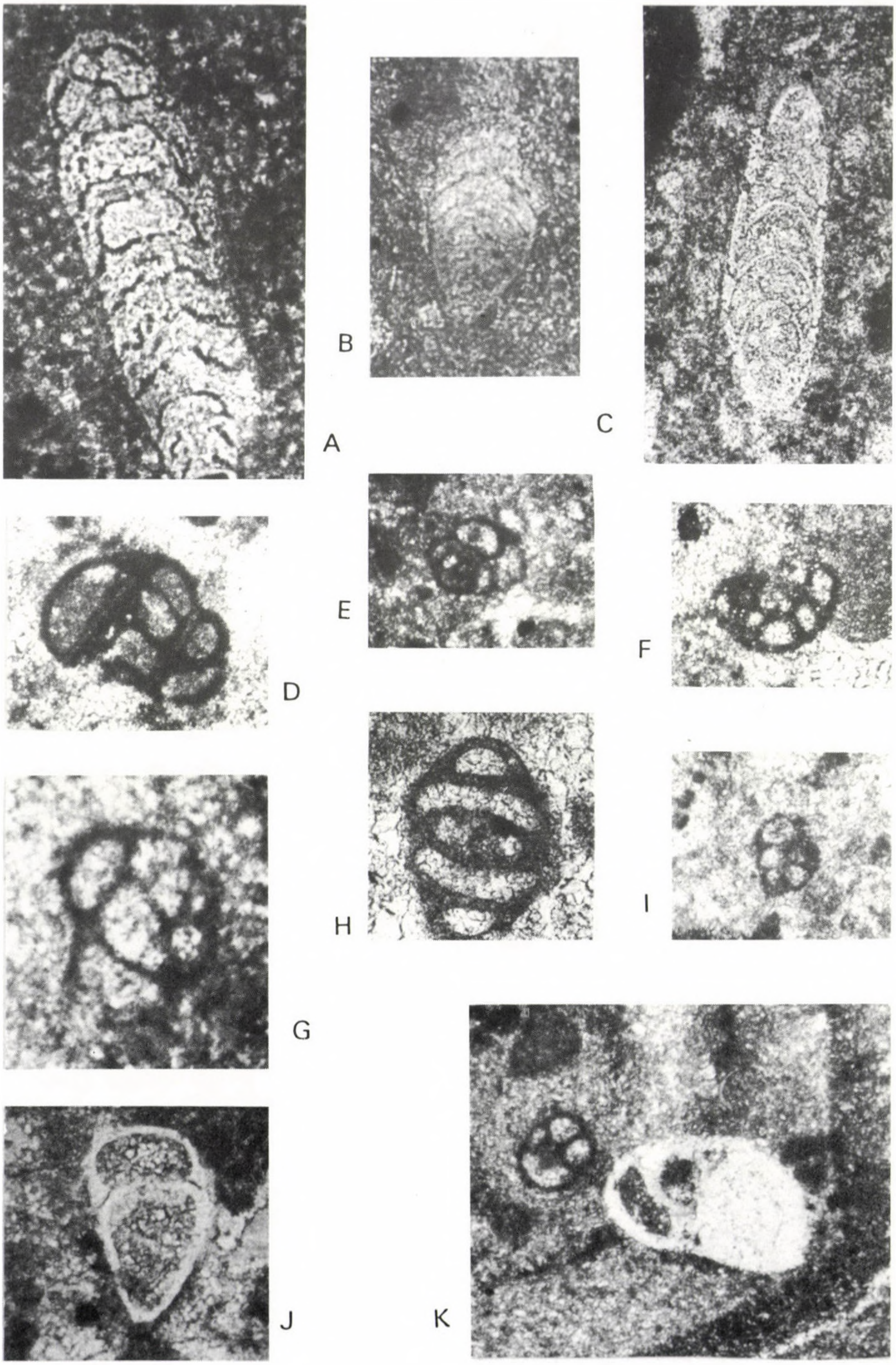

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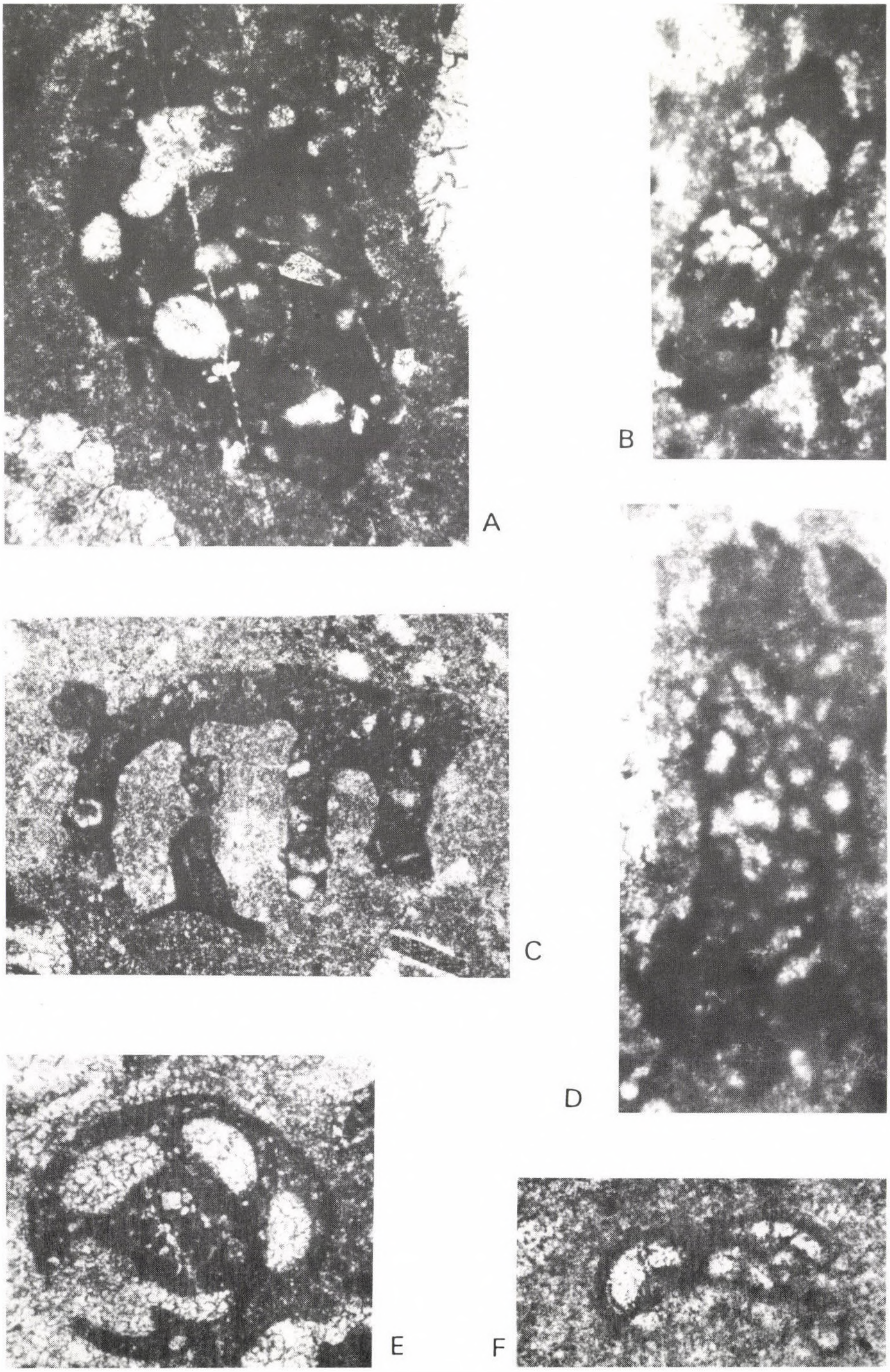

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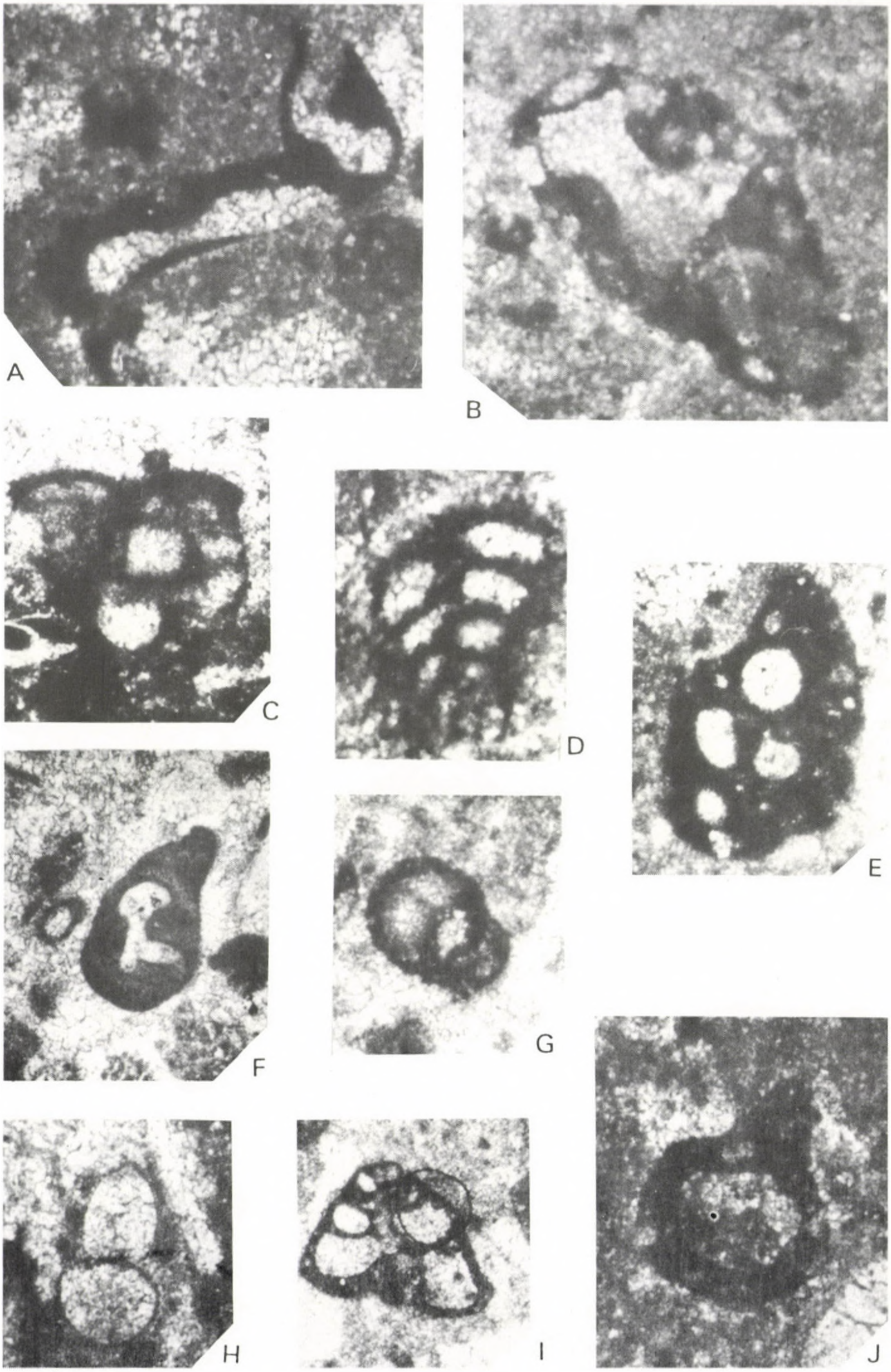
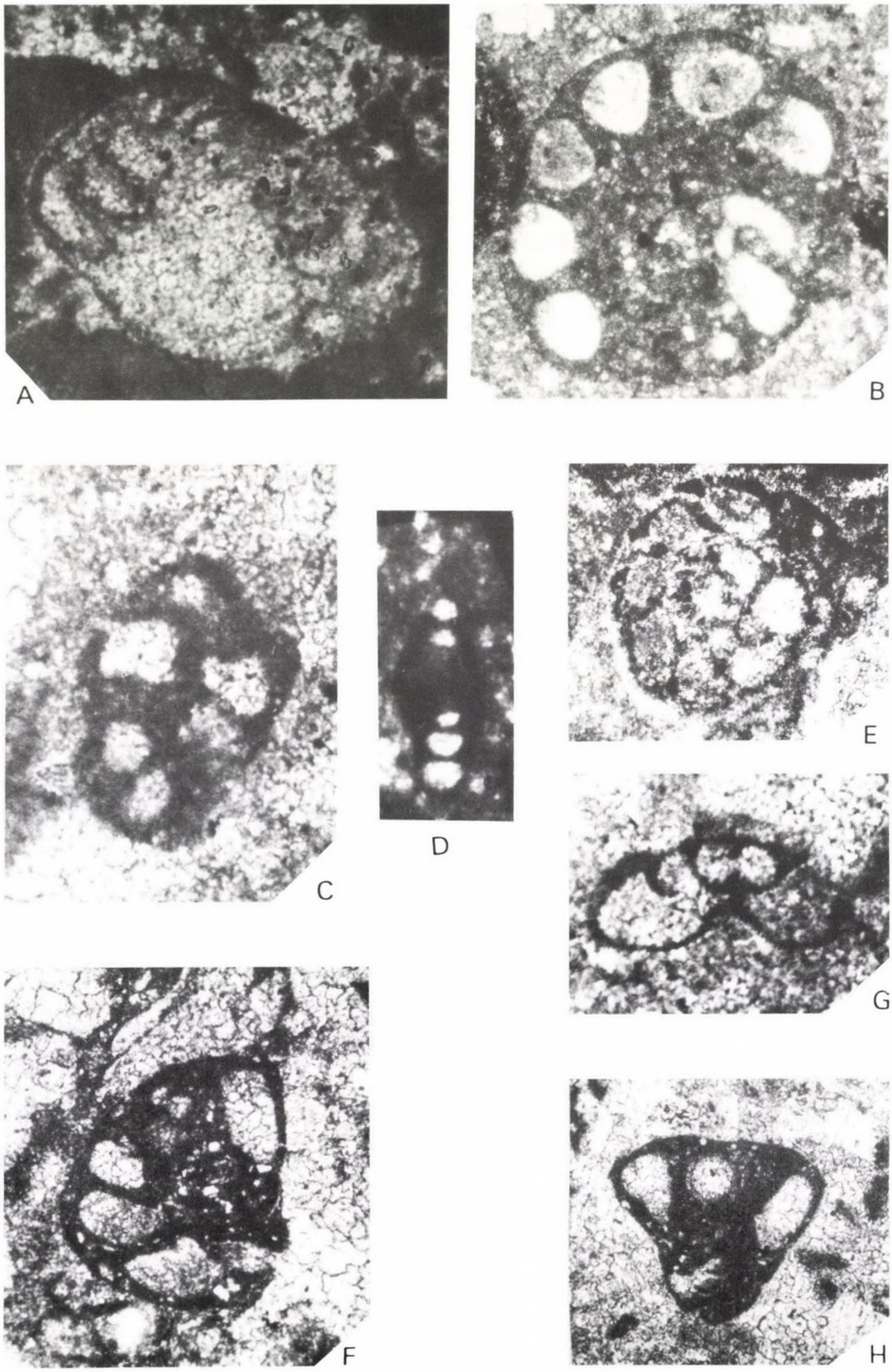


Plate XVII



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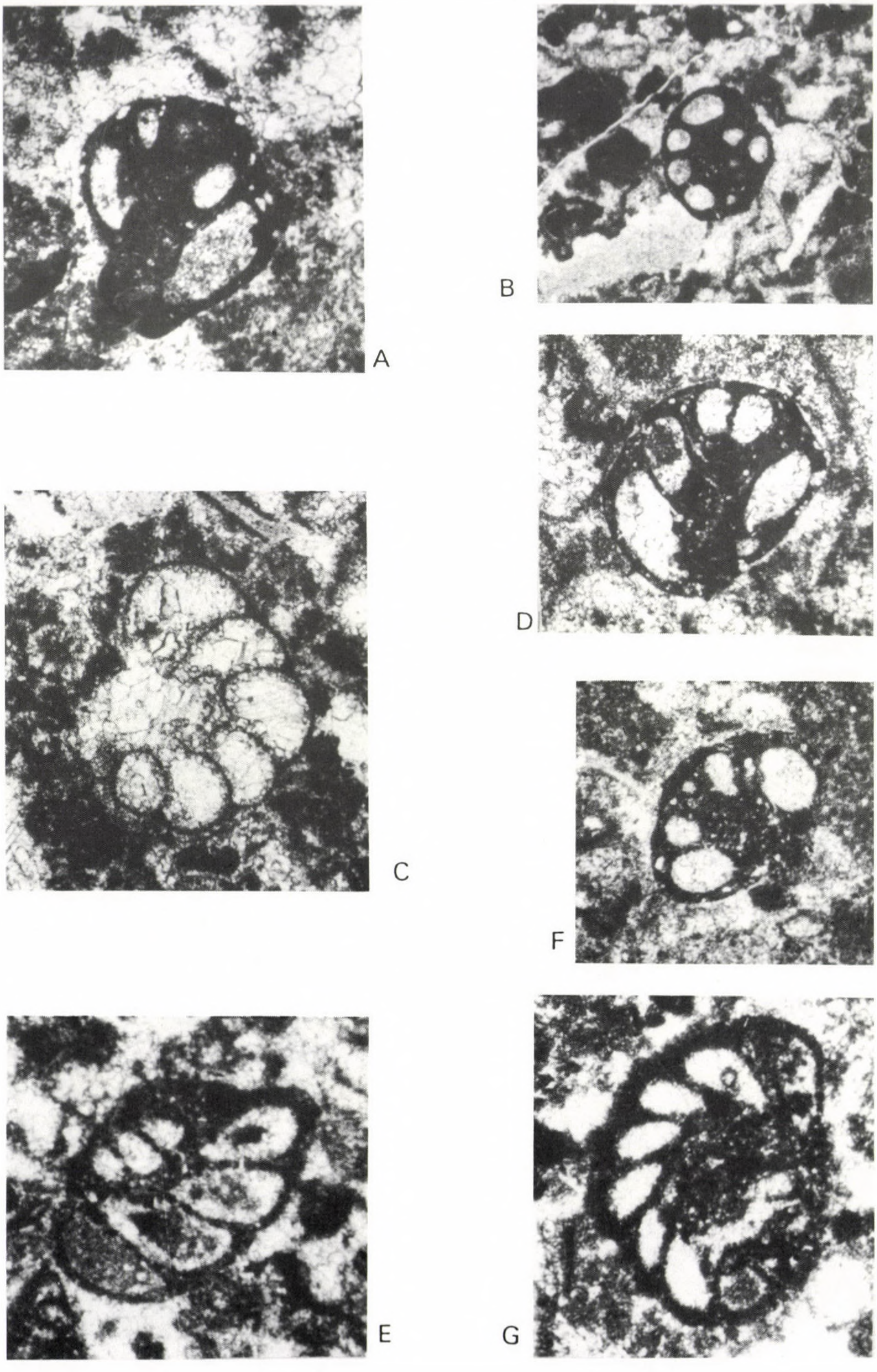

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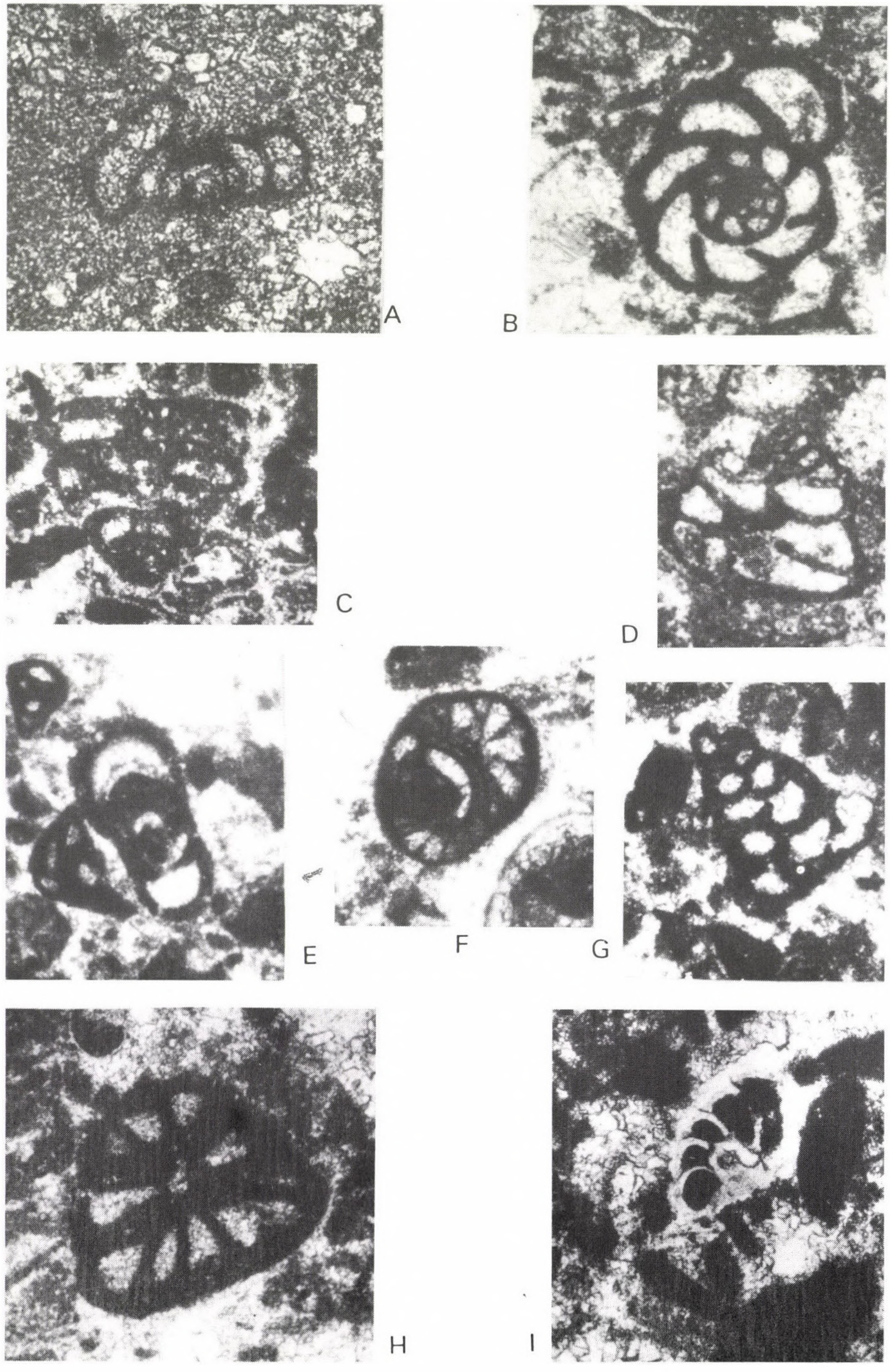

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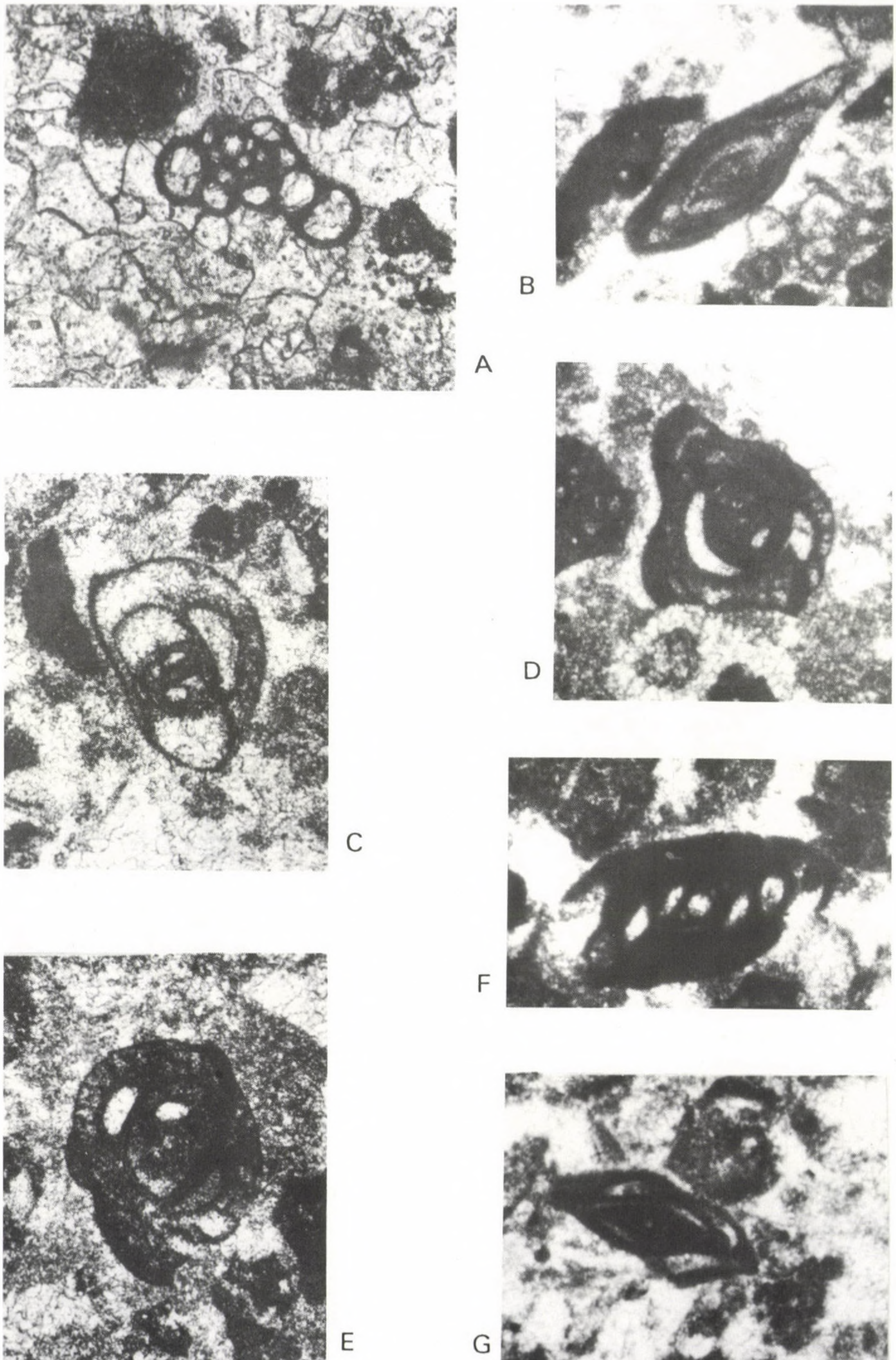
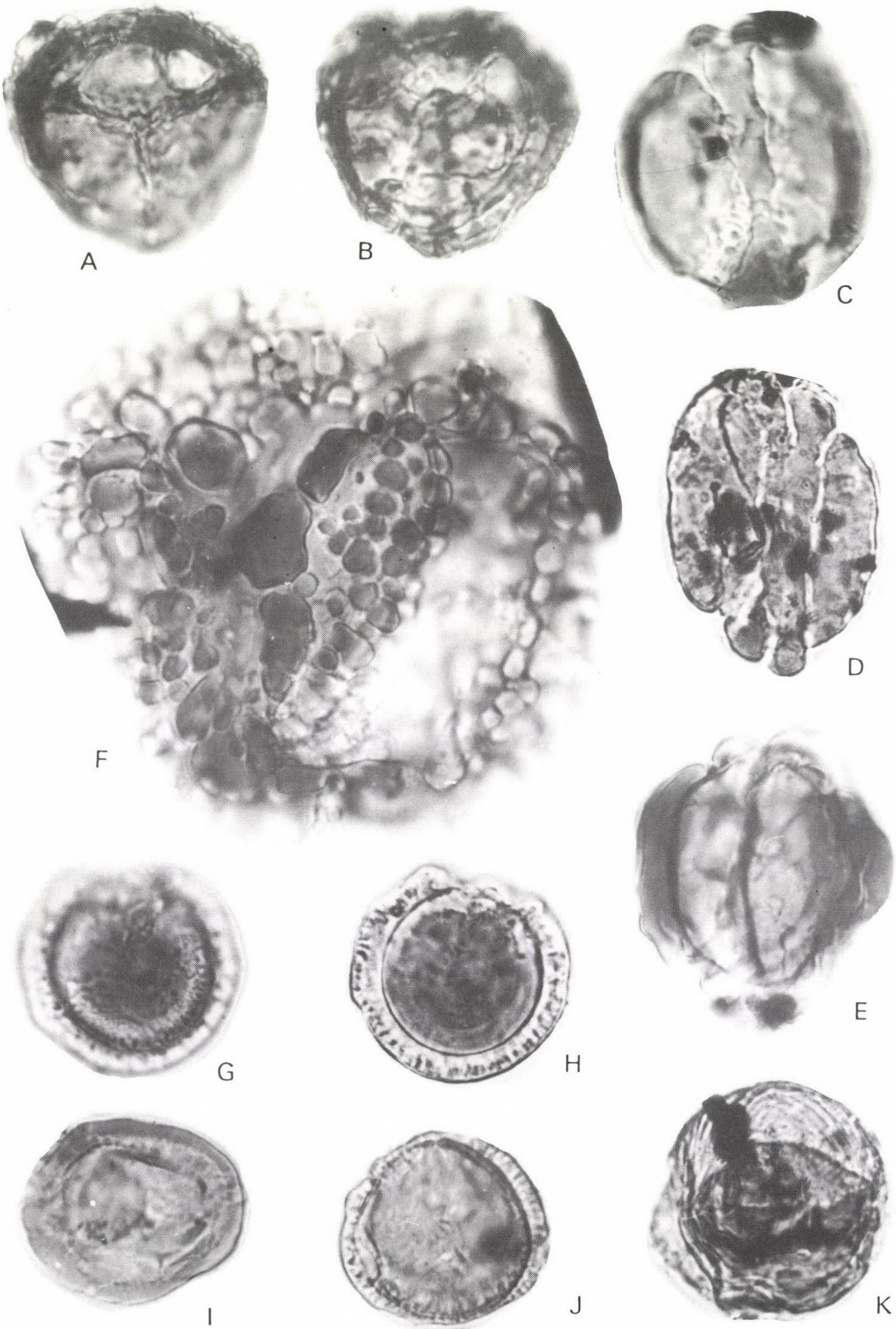


Plate XXI



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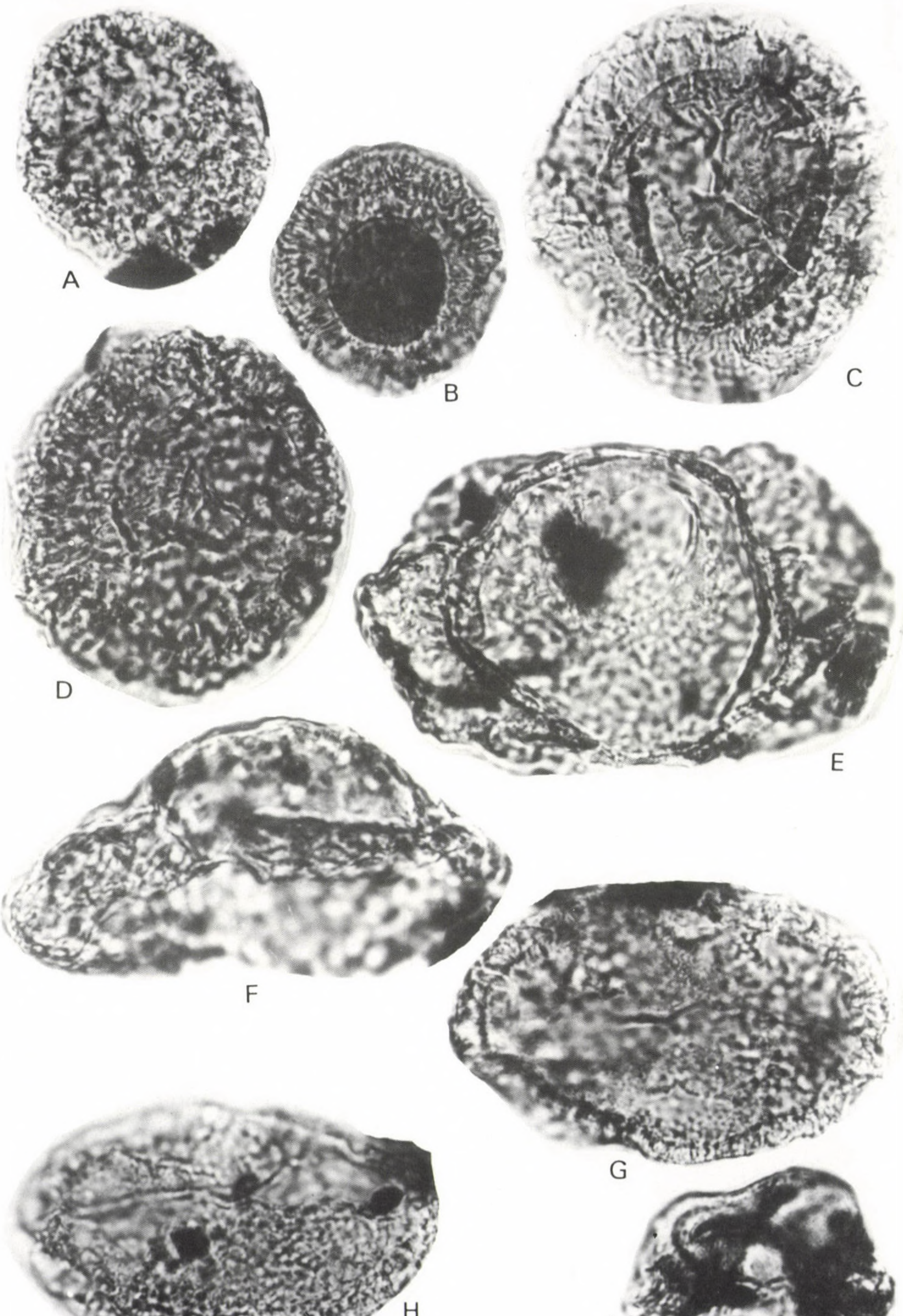

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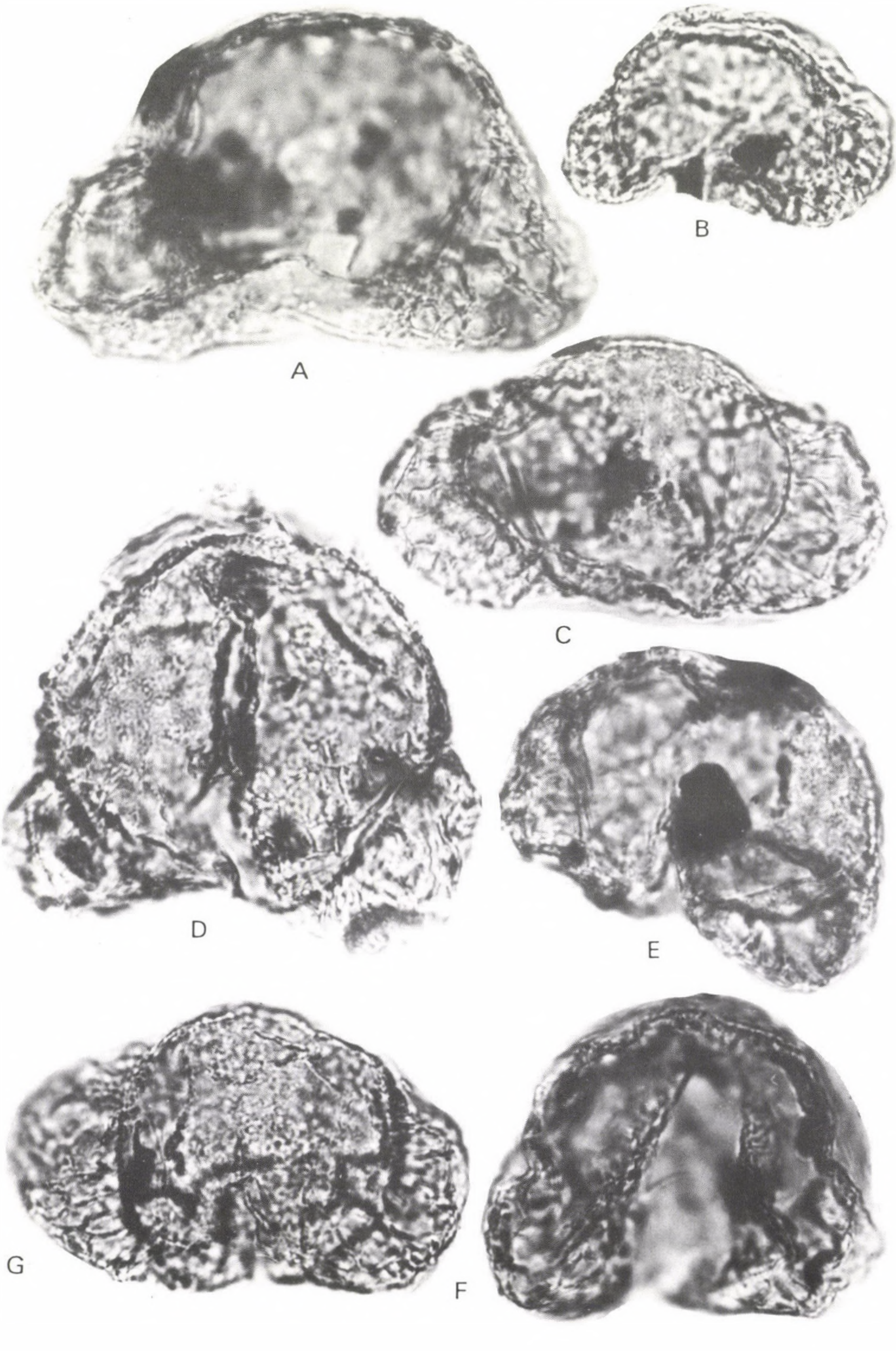

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Cretaceous insertions in Triassic (?) dolomites at Csővár, North Hungary

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Re-investigation of the cores from well Csővár Csv-1 led to the conclusion that beneath the Upper Triassic Csővár Formation, the approximately 600 m thick lower segment of the explored section should not be considered as a part of the normal stratigraphic succession. It consists predominantly of dolomites, strongly tectonized in some intervals and punctuated by thin insertions of varied lithology, Late Triassic to Late Cretaceous in age. Based on these observations, thrust sheets bounded by low angle faults and inserted slices of heterogeneous rocks belonging to a deeper tectonic unit, are assumed.

Key words: dolomite, stratigraaphy, tectonics, Cretaceous, Triassic, North Hungary

Introduction

Beneath the Upper Triassic Csővár Formation, the borehole Csővár Csv-1 encountered dolomites in a thickness of about 600 m (for location of the borehole see Fig. 2, for the general geologic column see Fig. 3 in the paper of Haas et al. in the present volume).

In 1988 Haas and Oravecz studied the preserved cores and took samples for thin-section investigations. The thin-section studies, carried out by Haas and Tardy-Filácz, led to a very surprising result. In some samples, taken from the lower, predominantly dolomitic segment of the cored sequence, rocks totally alien to the given geologic sequence, in terms of their age and facies, were found. These results were so unexpected that initially mistakes in the procedure of core conservation were suspected. As a result it was attempted to find the original drilling documentation and to obtain information from those colleagues who participated in the documentation many years ago. Finally, the original descriptions from Nagy and from the Archive of the Hungarian Geological Survey and nearly 1000 thin-sections of the first sampling by Detre were received. The investigation of this series of thin-sections and the evaluation of the original descriptions confirmed that the "odd" samples were really taken from the cored sequence, so it was necessary to find a realistic explanation for their occurrence in the lower, strongly tectonized part of the core.

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In the present brief communication, the results of the study of the dolomite complex explored in the deeper part of the core Csv-1 are summarized, with special regard to the "odd" samples found in certain intervals.

Geologic and structural setting

In the cores of Csv-1, the lower member of the Csővár Formation consists of dark or light grey, brownish-grey, cherty dolomites (Pokolvölgy Dolomite) showing microfacies characteristics akin to those in the undolomitized part of the Csővár Formation. Below it, beneath a tectonic contact, the lower part of the cores of Csv-1 (622–1200 m) is made up of darker or lighter brownish grey dolomites, strongly tectonized, brecciated, and mylonitic in some intervals. In several horizons, thin, slightly deformed greenish-grey calcareous marl insertions, locally with crinoidal limestone lenses, were also observed (Plate I).

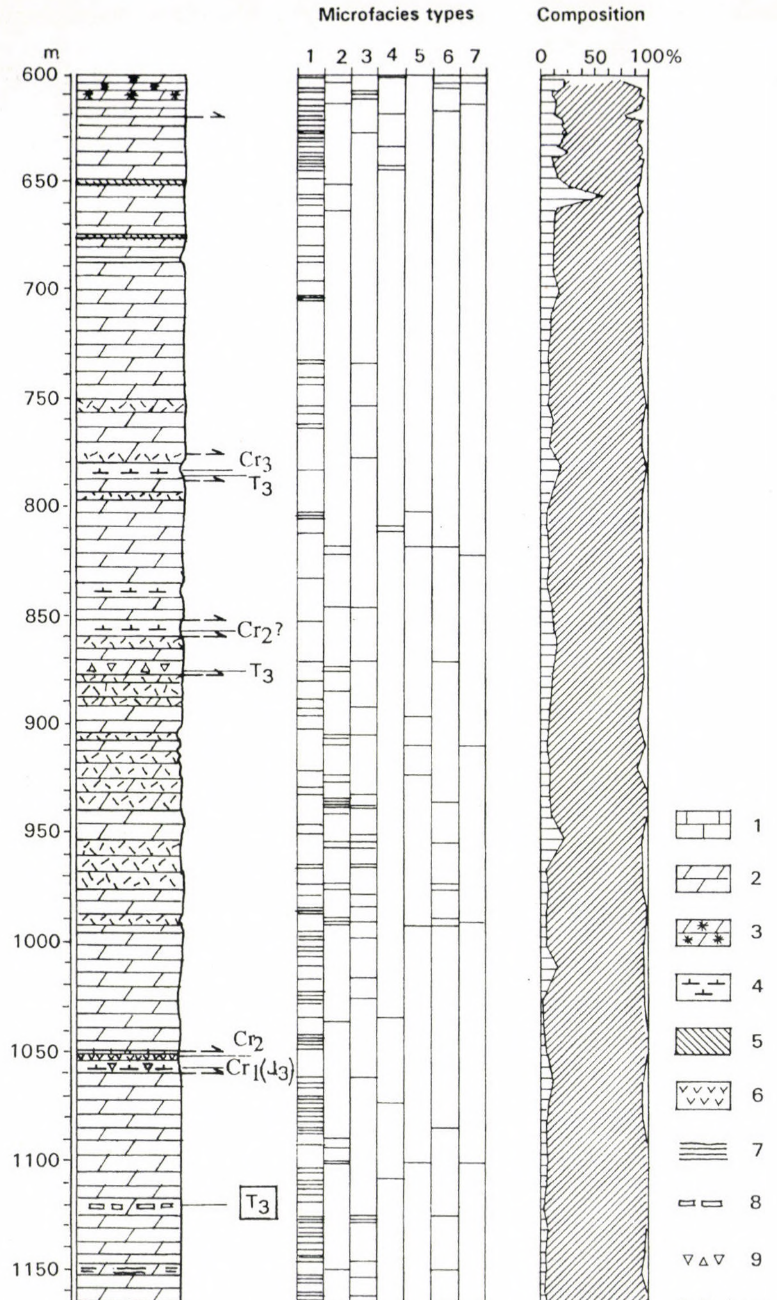
Based on the megascopic description of the cores and thin-section studies, it was concluded that the dolomitic complex exposed in the deeper part of the cores of Csv-1 is made up of thrust sheets, consisting predominantly of dolomitized rocks and bounded by low-angle faults, mylonite zones and slices of heterogeneous rocks, with extremely different ages and facies. In the opinion of the authors, this means that the dolomite complex in the deeper part of the cores of Csv-1 should not be considered as a part of the normal stratigraphic succession. Consequently, there is a major tectonic contact at the base of the Pokolvölgy Member, and the locally strongly fractured lower dolomite unit might belong to a deeper tectonic unit, or it can be considered as a tectonized boundary-interval between two nappe-like major structural units.

Lithology of the dolomites

The major lithological features and microfacies characteristics of the studied sequence are summarized in Fig. 1.

Between the thrust-zones dolomites, i.e. dolomitized rock-types, occur. The dolomites are dark grey, light grey, brownish grey, drab and finely crystalline as a rule. Locally, vague lamination is visible, however, microbial laminites are completely missing.

Fig. 1 →
Lithology, texture and rock composition of the lower segment of the cored section of Csv-1 (600–1200 m). Age data are also indicated (for details see Table I). 1. limestone; 2. dolomite; 3. cherty limestone; 4. calcareous marl; 5. calcite veins; 6. crinoidal limestone; 7. laminitic texture; 8. rip-up breccia; 9. lithoclasts; 10. tectonic breccia (mylonite); 11. tectonic contact; *Microfacies types:* 1. dolosparite; 2. dolosparite with micritic patches and relict textural elements (mainly bioclasts); 3. dolosparite with micritic intraclasts and peloids; 4. dolosparite with pyrite and other opaque mineral grains; 5. dolomicrite or dolomicrosparite with unrecognizable bioclasts; 6. dolomicrite or dolomicrosparite (partly silicified) with molds of radiolarians; 7. dolomicrite with "filaments"


Microfacies types
Composition
m
600
650
700
750
800
850
900
950
1000
1050
1100
1150
1200
Cr3
T3
Cr2?
T3
Cr2
Cr1(J3)
T3
1 2 3 4 5 6 7
0 50 100%
1
2
3
4
5
6
7
8
9
10
11

In a few horizons (1167.4–1167.8 m, 1141–1195 m), pebble-sized intraclasts of microbial mat origin were encountered (Plate IC).

Microscopic features of the dolomites are summarized in Plates II, III, and IV. The most common texture type (type 1 in Fig. 1) is finely or medium crystalline (25–70 m) dolosparite; according to Sibley and Gregg's system (1987) they can be classified into the xenotopic anhedral (A) or idiotopic subhedral (S) categories. In some instances, vague outlines of some grains are visible, but the origin of the grains cannot be traced. In a relatively large number of the studied samples, however, relicts of the original texture could be recognized (type 2 in Fig. 1). The totally recrystallized textures consist of fine to medium-sized dolomite crystals. Xenotopic–A or idiotopic–S texture-types are characteristic.

The recognized original constituents can be grouped as follows:

– Intraclasts, i.e. larger micritic grains (Plate IIIA)

– Remnants of shallow marine bioclasts (foraminifera, bivalves, gastropods, etc.) (Plate IIIB)

– Relicts of radiolarians (?)

– Relicts of thin-shelled bivalves (filaments).

In some cases, these texture elements may occur together. Microscopic investigations revealed that 1) dolomites are late diagenetic, of deep burial origin and 2) the precursor carbonates were probably heterogeneous, representing both shallow marine (platform?) and deeper basinal facies; however, it cannot be excluded that the shallow marine facies are present only as lithoclasts in basinal facies.

There is no firm evidence for the age of the dolomitized rock types. Age-diagnostic fossils were encountered in only a few dolomite samples and some of them were taken from the strongly tectonized intervals in the vicinity of the "odd" samples (see Fig. 1 and Table I) The two samples containing Carnian foraminifera of platform facies were also found in one of the these intervals.

In a sample from 987.9–991.2 m, a rich *Aulotortus* fauna indicating platform environment and Late Triassic age was encountered. There is only one single data point definitely suggesting a Norian age for some lithoclasts of platform origin in the 1118.2–1120.6 m sample (see Table I). Naturally, the matrix of the lithoclasts should be somewhat younger. However, taking into account the heterogeneity of the dolomitized rocks (relicts of both pelagic and platform fossils), it is difficult to state anything about the age of the precursor rocks. It can also be assumed that the thrust-sheets consist of dolomitized formations of significantly different age.

The microscopic investigation of dolomites also provided some information about the history of the rocks after the major phase of dolomitization. In some samples, cataclastic texture was observed (Plate IIB). Due to tectonic brecciation, pebble-sized to medium crystal-sized clasts were formed. Fractures are filled by one or two generations of dolosparites: 1) fine-crystalline xenotopic-A and/or 2) coarse, zonal (rarely saddle-type) idiotopic-S (Plate IV).

Intracrystalline pores and, occasionally, central zones of the fractures are commonly filled by pyrite.

Insertions

In Fig. 1, symbols mark those samples which could be dated biostratigraphically. As a rule, they were undolomitized samples. The exact location of these samples, their lithology and the evidence for their age are summarized in Table I.

Table I
Age, lithology and characteristic fossils the biostratigraphically dated samples

Age	Depth	Lithology	Characteristic fossils
T_3 Norian	1118.2–1120.6 m	dolomite	*Aulotortus gaschei* *Aulotortus friedli* *Aulotortus communis*
T_2	987.9–991.2 m	dolomite	*Aulotortus* sp. (abundant)
T_3	872.0 m	dolomite	Solenoporacea, bryozoans, ostracods
T_3	867.8 m	dolomite	*Aulotortus* div. sp. (abundant) *Nodosaria raibliana* *Tolypammina gregaria*
T_3 Lower Carnian	812.4–859.0 m	dolomite	*Aulotortus praegashei* *Gsollbergella spiroloculiformis* *Trochammina almtalensis*
T_3 Carnian	786.0 m	calcareous marl	Aulotortus praegashei Gsollbergella spiroloculiformis *Valvulina* sp.
Cr_1 –Cr_2 Upper Barremian– Lower Albian	1055.6–1058.2 m	calcareous marl + crinoidal limestone	*Nannoconus, Globochaete alpina* *Globigerinelloides* sp., *Cyathidites minor* *Leiotriletes rotundus,* *Gleicheniidites (Triplexisporites) triplex ,* *Gleicheniidites (Triremirsporites) minor* *Cicatricosporites hallei* *Cicatricosporites venustus* *Callialasporites dampieri* *Callialasporites trilobatus*
Cr_1–Cr_2 Upper Barremian Lower Aptian clasts: Malmian– Berriasian	1058.2–1062.6 m	calcareous marl	Crinoids, Nannoplankton, *Colomisphaera* sp., *Globigerinelloides* sp. in clasts: Saccocoma, *Calpionella* sp. *Cyathidites minor, Leotriletes rotundus,* *Gleicheniidites (Triplexisporites) triplex* *Gleicheniidites (Triremisporites) minor* *Cicatricorisporites hallei* *Cicatricorisporites venustus* *Callialasporites dampieri* *Callialasporites trilobatus*
Cr_2 Aptian(?)	1049.1–1052.4 m	limestone	Crinoids

Age	Depth	Lithology	Characteristic fossils
Cr_3 Aptian (?) clasts: Tithonian- Hauterivian	862.2 m		*Globigerinoides* in clasts: *Calpionella alpina*
Cr_3 Upper Cretaceous (?)	856.7–862.2 m	calcareous marl	*Hedbergella* sp., bryozoans, ostracods
Cr_3 Upper Campanian	783.0 m	calcareous marl	*Hedbergella* sp. *Pithonella ovalis* *Conocella ugodensis* *Hungaropollis – Krutzschipollis* assemblage

An excellently preserved sporomorph assemblage was found in grey calcareous marl in a sample taken from 783.0 m, which clearly indicates the Upper Campanian *Longanullipollis bajtayi – L. lenneri* Assemblage Zone (Góczán 1964) (Plate VII). In the thin-section from the same sample, a rich Calcisphaerulidae assemblage was found with *Pithonella ovalis* Kaufman, *Conocella ugodensis* Haas and other forms (Plate V). As in the case of the sporomorph data, the microfacies characteristics of this sample are strikingly akin to that of the Polány Marl Formation, known in the Bakony Mts (Haas 1978, 1979).

In the sample from 862.2 m, in crinoidal grainstone, *Globigerinelloides* sp. (*algerianus* ?) and *Ticinella* sp. were found (Plate VC, D), and in extraclasts *Calpionella alpina* Lorenz occurred (Plate VIA). The microfacies characteristics and microfossils of this sample are very similar to those of the Aptian Tata Limestone (Fülöp 1976; Haas et al. 1985; Lelkes 1990).

Sporomorph grains in the calcareous marl samples from 1055.6–1058.2 m and 1058.2–1062.6 m are similar to those in the Aptian–Barremian formations of the Transdanubian Range, but due to the relatively small number of taxa, an exact age determination was not possible. Aptian–Barremian age is constrained by the absence of the monosulcate angiosperm pollen grains, which are characteristic of the Albian, and of the large trilobate and striate pteridophyte spores, typical of the Upper Jurassic and deeper Lower Cretaceous formations (Plate VIIIA–D). According to the thin-section studies, the sample from 1055.6–1058.2 m contains nannoconus in rock-forming quantity (Plate VIC). In another sample, from 1058.0–1062.6 (radiolarian wackestone), a Saccocoma-bearing extraclast (most probably of Kimmeridgian-Tithonian age) was found (Plate VID).

It is extremely unusual for such a very wide stratigraphic range, from the Upper Triassic to the Upper Cretaceous, to be represented among the dated samples. Most of these rocks have shown a definite resemblance to the contemporaneous Transdanubian Range facies.

Figure 1 clearly shows that the "odd" samples occur in three horizons. According to the original description of the core, every horizon is located in a strongly tectonized interval. It is assumed that the "odd" rock-types mark the base of individual scales or thrust sheets.

It is worth mentioning, however, that according to the palynological investigations, no trace of strong tectonic effects could be observed on the sporomorph grains. Their colour, state of preservation and maturity was equal with those of the contemporaneous formations in the other parts of the Transdanubian Range.

Conclusions

Based on observations discussed above, it can be concluded that the dolomite complex exposed in the deeper part of the cores of Csv-1 consists of dolomite thrust sheets (scales) separated by thin, highly tectonized rock-slices of extremely variable lithology and age. Since rocks of these intervals are not, or only slightly, affected by dolomitization, it can be assumed that imbrication should have taken place after the major phase of dolomitization. Taking into account that, in the tectonized package, the youngest rocks are Senonian, the tectonic movements which resulted in the formation of the thrust sheets should have occurred after the Cretaceous. The definite post-deformational character of the Upper Eocene in the Csővár-block (and also in the other blocks east of the Danube) post-dates the imbrication (see Fig. 2 in the paper of Haas et al. in the present volume), which consequently may have occurred in the Paleocene–Early Eocene interval.

Although it is admitted that the proposed suggestion to solve the enigma of the 600 m thick lower part of the cored interval of Csv-1 is rather unusual, no better explanation has been found so far.

Finally, it should be noted that dolomites in the outcrops of Vas Hill (see Fig. 2 in the paper of Haas et al. in the present volume) are similar to those in the lower part of the cores of Csv-1. However, further studies are needed to prove that the two dolomite units are identical.

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Plate I

A) Grey, bioturbated calcareous marl (Barremian-Aptian) and crinoidal limestone kneading into one another. Core Csv-1, 1055.6–1058.2 m
B) Lenticular crinoidal limestone and micritic limestone clasts (Malmian–Berriasian) in greenish - grey calcareous marl (Barremian–Aptian). Core Csv-1, 1058.0–1062.6 m
C) Light grey dolomite with rip-up breccias of algal mat origin. Core Csv-1, 1167 m

Plate II

A) Idiotopic subhedral dolosparite of medium crystal-size. Core Csv-1, 1073.3–1075.8 m
B) Tectonically brecciated xenotopic euhedral dolosparite. Core Csv-1, 1027.2–1030.2 m

Plate III

A) Finely crystalline xenotipic euhedral dolosparite with micritic intraclasts. Core Csv-1, 973.5–976.7 m
B) Finely crystalline dolomicrosparite with molds of molluscs and foraminifera (?). Core Csv-1, 987.9–991.2 m

Plate IV

A) Coarse pore-filling zonal idiotopic euhedral and subhedral sparry dolomite. Core Csv-1, 743.6 m
B) Coarse pore-filling zonal idiotopic euhedral and subhedral sparry dolomite. Saddle-shaped crystals are also visible. Core Csv-1, 1007.0 m

Plate V

A) *Conocella ugodensis* Haas – Upper Campanian. Core Csv-1, 783.0 m
B) *Pithonella ovalis* Kaufman and other calcisphaerulids – Upper Campanian, Core Csv-1, 783.0 m
C) *Globigerinelloides* sp. – Aptian (?), Core Csv-1, 862.2 m
D) *Ticinella* sp. and other planktic foraminifera – Aptian (?), Core Csv-1, 862.2 m

Plate VI

A) Extraclastic crinoidal grainstone (Aptian?), *Calpionella alpina* Lorenz is visible in an extraclast grain. Core Csv-1, 862.2 m
B) Silicified crinoidal grainstone. Core Csv-1, 1052.7–1055.6 m
C) Nannoconus mudstone (Barremian–Aptian). Core Csv-1, 1055.6–1058.2 m
D) Echinoderm detritus and Saccocoma-bearing extraclast (Kimmeridgian–Tithonian) in radiolarian wackestone matrix (Barremian–Aptian). Core Csv-1, 1058.0–1062.6 m

Plate VII

Upper Campanian sporomorphs of the grey calcareous marl in the core of Csv-1, (783.0 m)

A) *Microfoveolatisporites* sp.
B) *Hungaropollis* sp.
C) *Trudopollis cretaceous* nov. sp.
D) *Trilites* sp.
E–F) *Hungaropollis* cf. *ajkanus* Gócz. 1964.
G–H) *Krutzschipollis spatiosus* Gócz. 1967
I) *Longanulipollis longianulus* (Gócz. 1964) 1967
J) *Longanulipollis* sp.
K–L) *Dinogymnium microgranulosum* Clarke et Verd. 1967

Plate VIII

Lower Cretaceous sporomorphs in the core of Csv-1 (1055.6–1058.2 and 1058.2–1062.6 m)

A) *Trilites toratus* Weyl. et Kr. 1953 Juhász 1977. Core Csv-1 1055.6–1058.2 m
B) *Callialalsporites trilobatus* (Balme 1957) Dev. 1959. Core Csv-1 1058.0–1062.6 m
C) *Cicatricosisporites baconicus* Deák 1964. Core Csv-1 1058.0–1062.6 m
D) *Cicatricosisporites hallei* Delc. et Sprum. 1955. Core Csv-1 1055.6–1058.2 m

Plate I

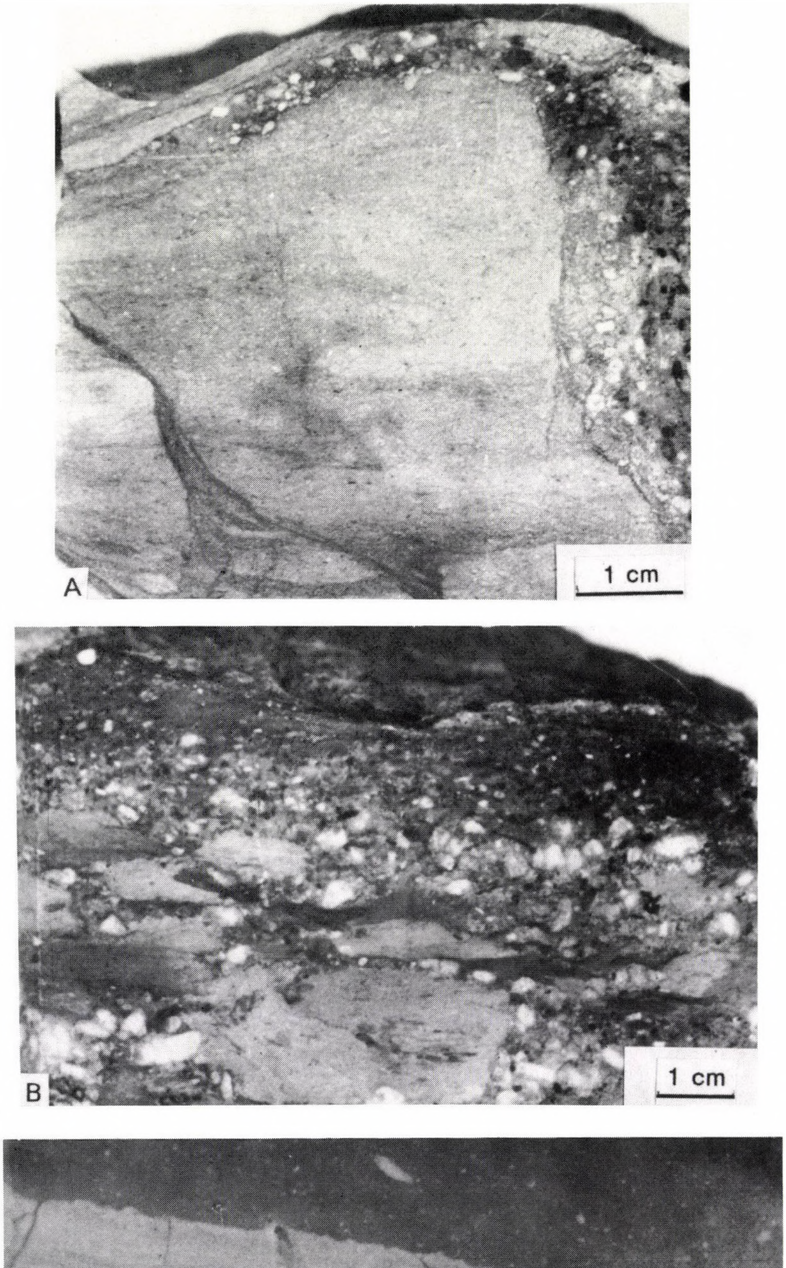
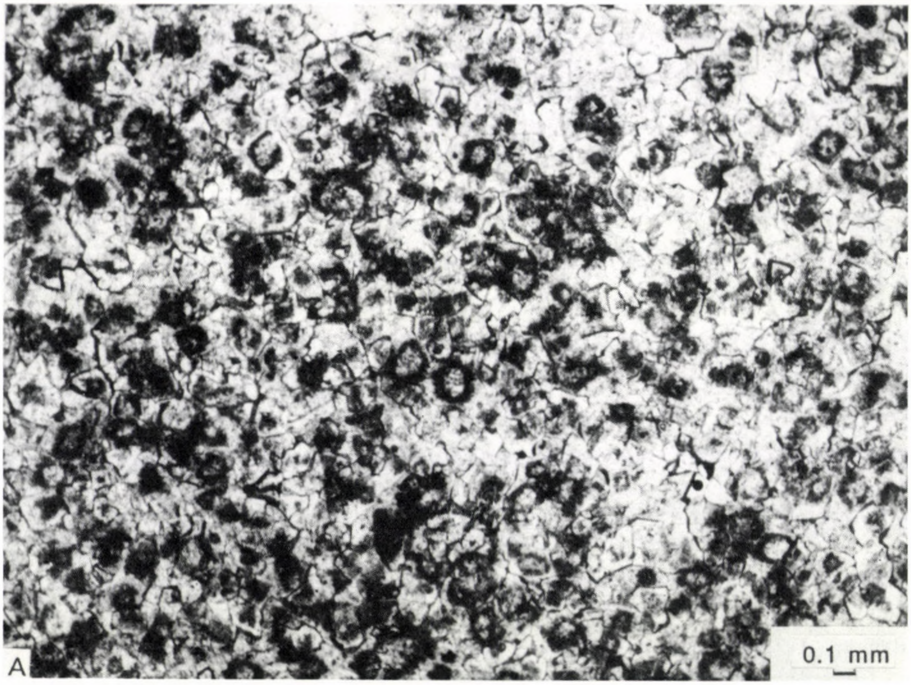


Plate II



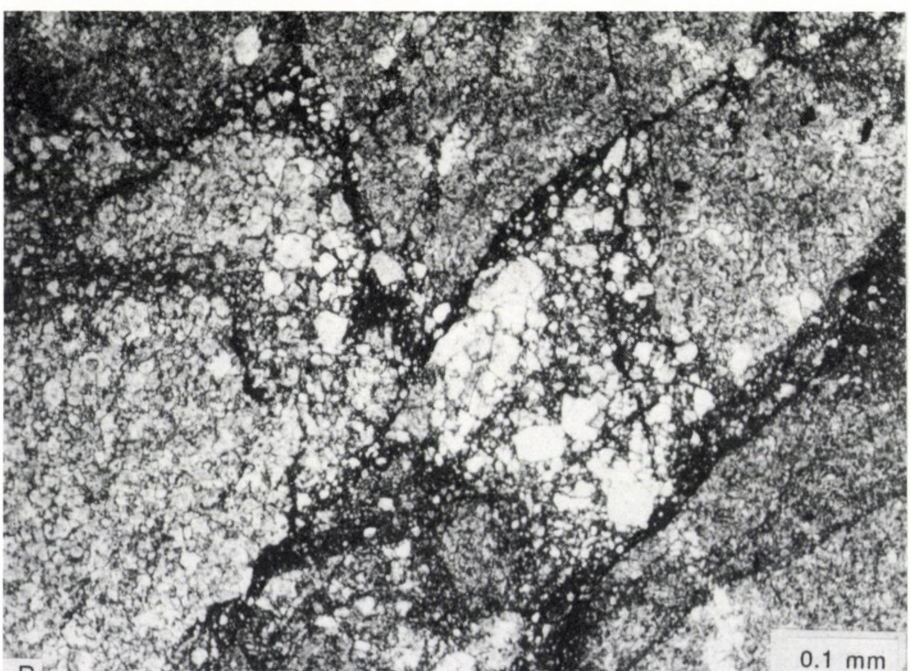


Plate III

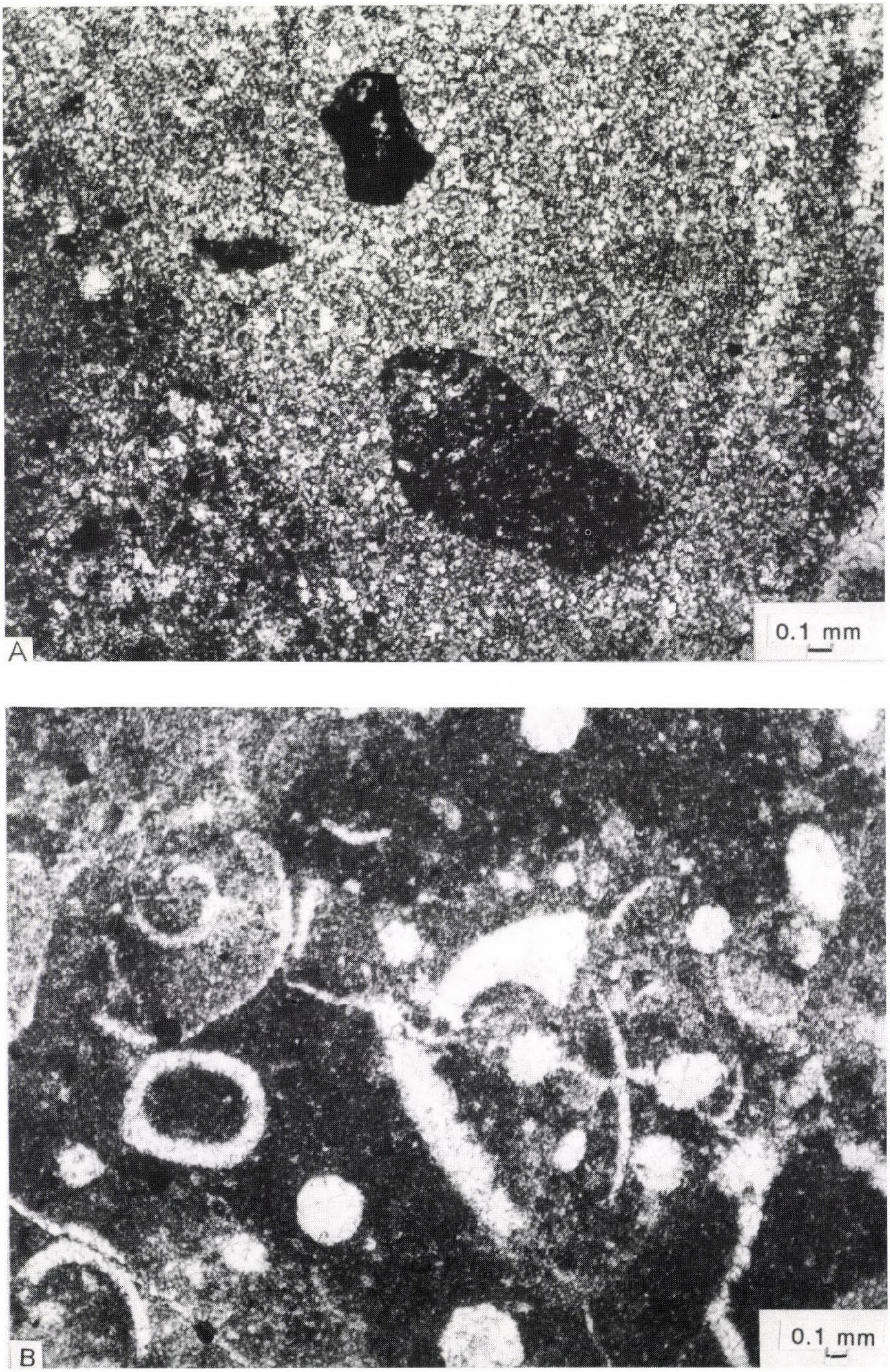


Plate IV

Plate V

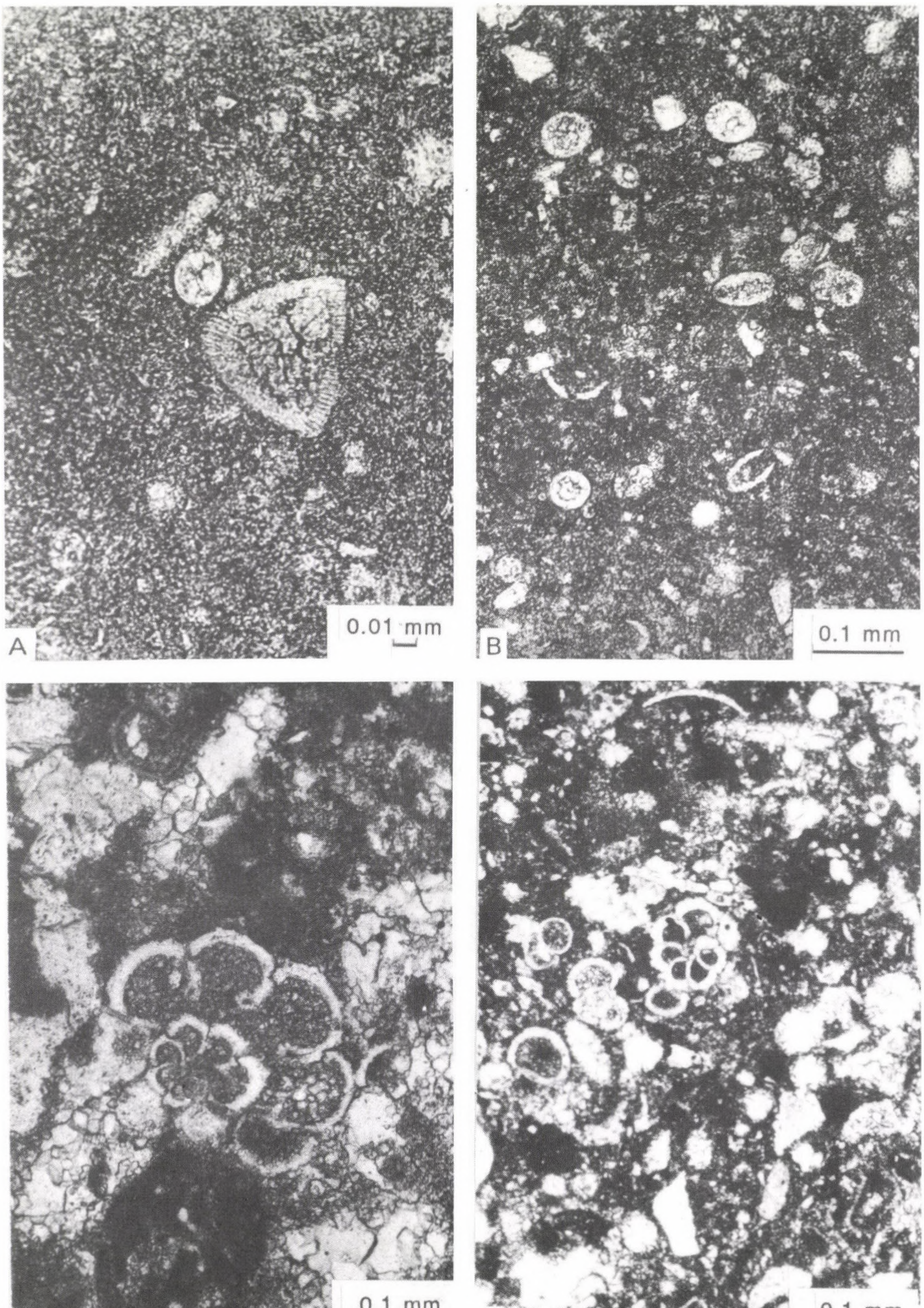


Plate VI

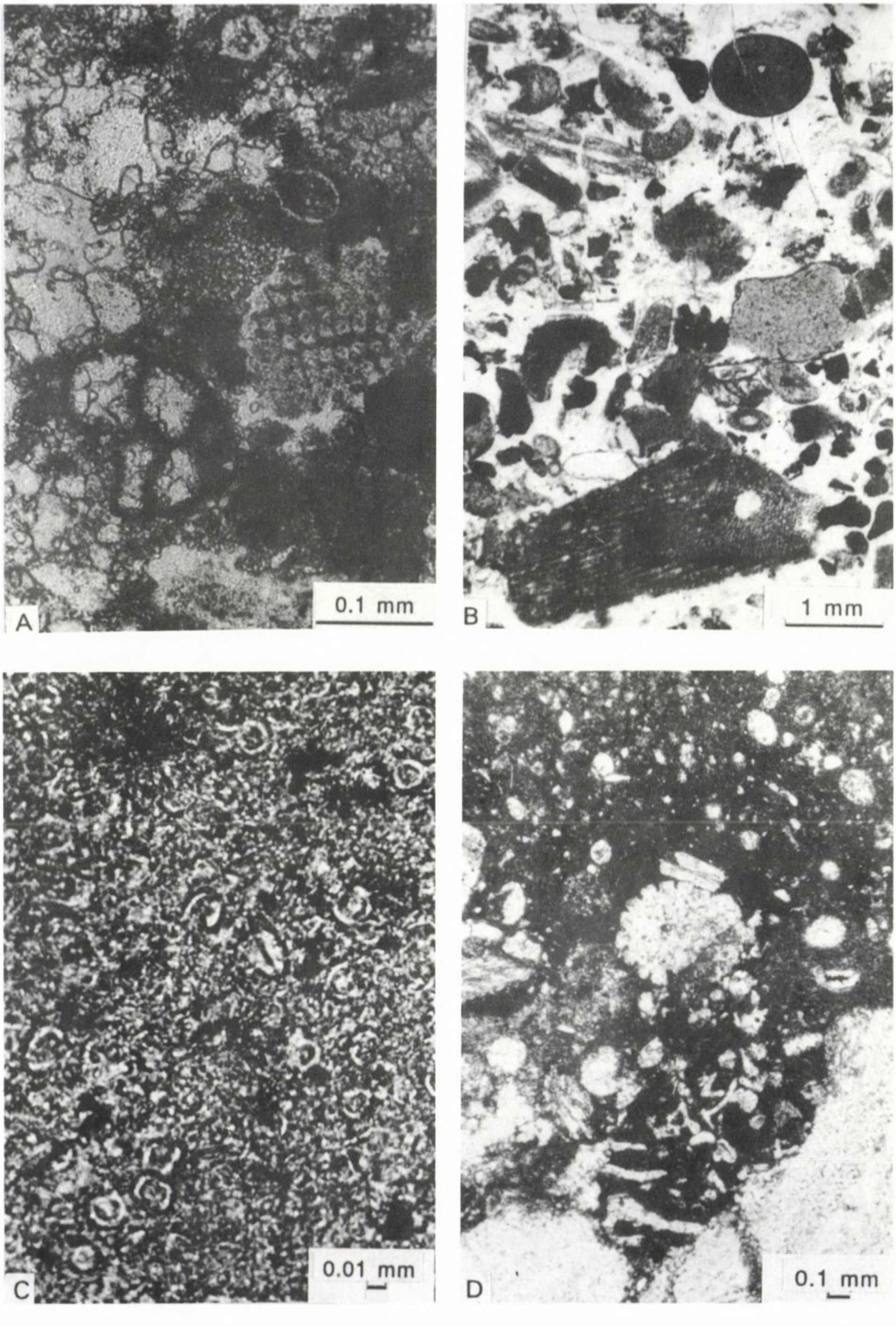


Plate VII

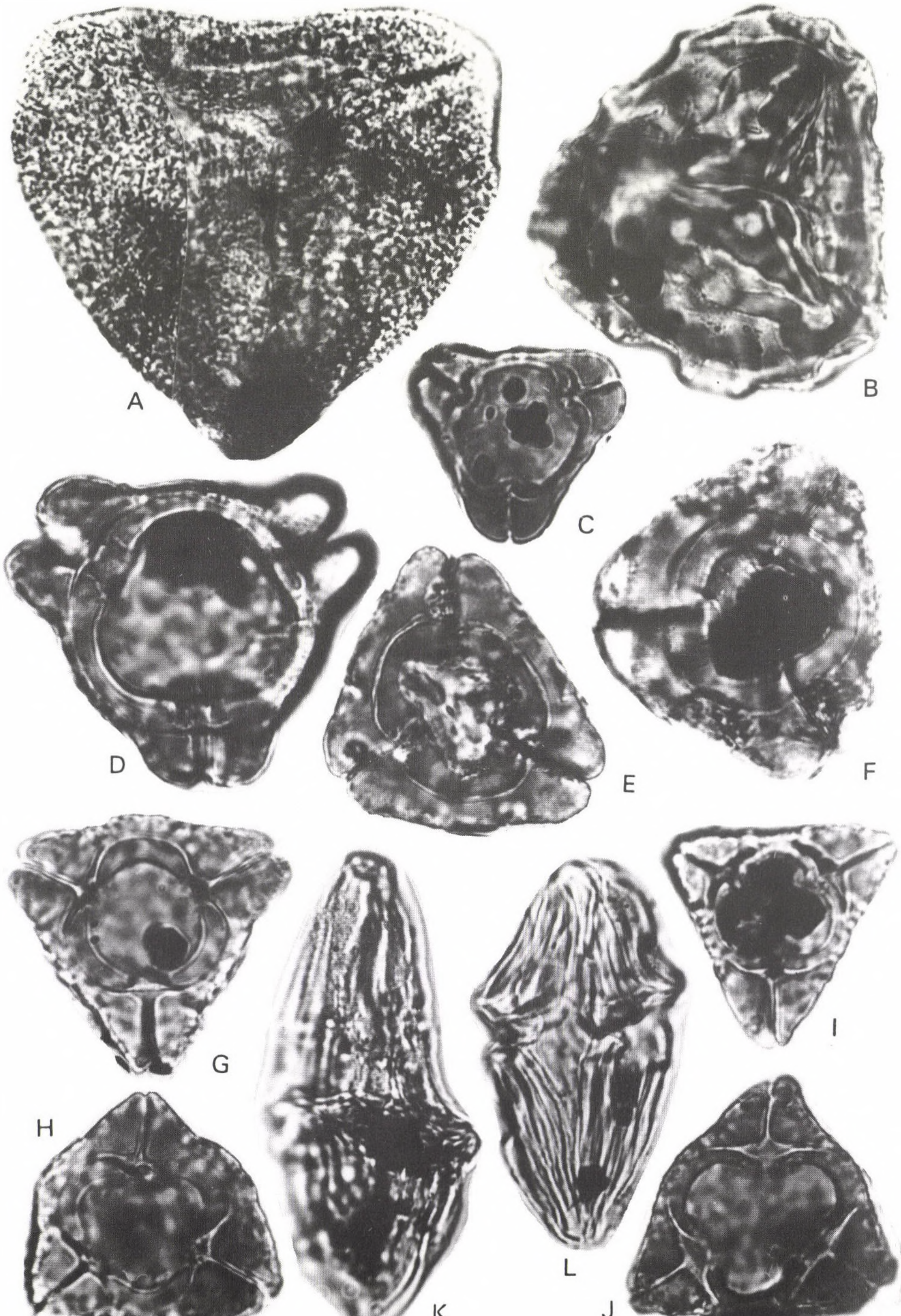
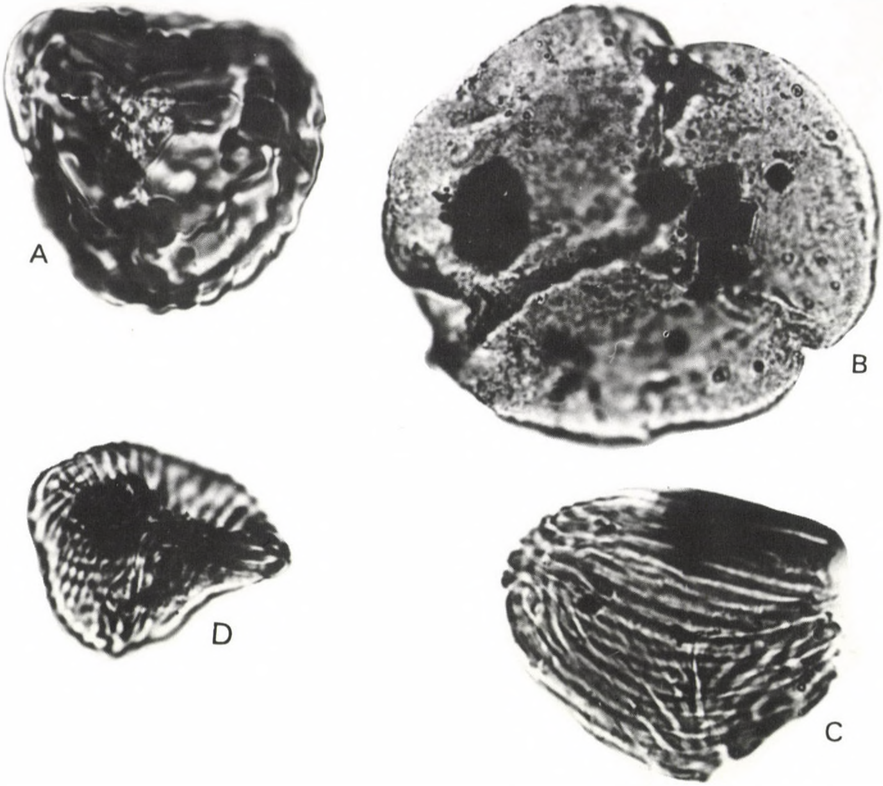


Plate VIII



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New organic framework plant microfossils in the Lower Rhaetian beds of the Csővár Limestone Formation

Ferenc Góczán

Acid treatment revealed that grains appearing like disseminated poppy-seeds on the bedding plane of 3–4 cm thick, light grey, hard, calcareous marls in the basal layers of the Upper Triassic formations (exposed in 1961) of the Big Quarry of Csővár are large, thick-walled organic framework plant microfossils. They represent 4 genera and 7 species. Among them, three genera and six species proved to be new. In introducing them, the names *Oraveczia hungarica* nov. gen. et sp., *O. doliola* nov. sp., *O. galeata* nov. sp., *Vadaszia cavernosa* nov. gen. et sp., *Sulcodiscus trisulcatus* nov. gen. et sp. and *Tytthodiscus tubulatus* nov. sp. are recommended. *Oraveczia faveola* (Morbey 1975) nov. comb. et emend. is also discussed.

The stratum typicum of the described microfossils belongs to the Csővár Limestone Formation the age of which proved to be lowermost Rhaetian on the basis of the sporomorph association occurring together with the microfossils. Characteristic sporomorph taxa are as follows: *Riccisporites tuberculatus* Lundblad 1954, *Rhaetipollis germanicus* Schulz 1957, *Classopollis torosus* (Reiss. 1950) Balme 1957, and *Corollina meyeriana* (Klaus 1960) Venk. et Gócz. 1962.

Key words: Lower Rhaetian, organic microfossils, sporomorphs.

Introduction

During the 1961 geologic investigation of the Triassic blocks on the left (west) bank of the Danube, in the left side of the quarry yard of the village of Csővár's Big Quarry, at a height of around a half metre (between the basal layers) J. Oravecz noted a 3–4 cm thick, light grey, hard, bituminous, calcareous marl layer, on the bedding plane of which he observed a mass of grains of poppy-seed size. From the acid-insoluble residue of the rock, he took microscopic photographs of spore pollen grains and microfossils of 200–300 micron size. Both the photographic material and the rock samples were handed on to me for further detailed investigation.

With a view to wall structure investigations, essential for the determination of the unknown plant microfossils, a very fierce and quick oxidation process using a mixture of nitric acid and ethyl alcohol was applied, using the organic framework fossils (separated by centrifuge from the insoluble residue of the rock material) as catalysts. As a result the majority of the originally intact but opaque grains became superoxygenated; however, the structure of the grains,

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which remained intact, could now be investigated. The greater part of the sample mounts proved to be useable even after 35 years – with minor renewal.

In the present work, the description of the determined and systematized microfossils is documented partly by original and partly by new photographs. In naming and describing the grains which proved to be new taxa, the author repays a debt he owes, partly for postponing the publication of these microfossils for 35 years, and partly as a sincere – not dutiful – homage to the former renowned researcher of the Triassic of Csővár, Professor Elemér Vadász, and to his excellent apprentice, the discoverer of these fossils, the geologist János Oravecz.

A large part of the fossils described here occurs en mass also in the Upper Norian and Lower Rhaetian limestone and calcareous marl (Csővár Limestone Formation) boundary layers of borehole Vérhalom-1, but in so strongly a biodegraded state that they are unsuitable for investigation. For the same reason, it is regrettable that the type layers of these fossils are hardly accessible any more today, being covered by 2–3 m-thick rock debris.

Palaeontology

Below, the descriptions of 7 species (among them 6 new ones) within 4 genera (among them 3 new ones) are given.

Forma genus: *Oraveczia* nov. gen.

Synonima:
Oraveczia nov. gen. Góczán, in Detre, Cs. 1970, p. 174 (nomen nudum)

Derivatio nominis: named after the discoverer of these fossils, the geologist János Oravecz

Genotype: *Oraveczia hungarica* nov. gen. et sp.

Genus diagnosis: large-sized, thick-walled organic framework plant microfossil with circular or slightly oval outline in E-view and with complex wall structure, which consists of a thin outer and a thick inner layer as well as a thin cover lamella closing the inner surface. The outer layer is smooth and unornamented. The inner layer is built up by the matrix and the thin-walled calyx-, mitre- or barrel-like structural elements embedded in it. The matrix may be compact or tubular, the structural elements are empty or tubular. The lamella, closing the inner surface, is smooth.

Differential diagnosis: by the structure of its wall, *Oraveczia* nov. gen. differs from all thick-walled plant microfossils known to date. As to the size and wall thickness, it shows the greatest similarity to *Tasmanites* Newton 1875 and *Tytthodiscus* Norem

1955. The walls of the members of both of these genera are tubular but lack the characteristic wall structural elements of *Oraveczia* nov. gen.

As to the thickness and structure of the inner wall, it resembles most *Vadaszia* nov. gen. Calyx-like, mitre-like or barrel-like structural elements of the inner wall of *Oraveczia* nov. gen. essentially distinguish it from *Vadaszia* nov. gen. with its cavernous inner wall.

There is also a significant difference in the average size of the specimens of the two genera: the size of *Oraveczia* nov. gen. grains varies between 220 and 380 microns (the mean value is 283 μm on the basis of 200 grains), while that of the specimens which can be assigned to *Vadaszia* nov. gen. is between 140 μm and 240 μm (mean value: 183 μm).

Remark: on the basis of the combination of the structural elements of *Oraveczia* nov. gen. several form species can be distinguished. Size and shape of the structural elements, however, depend strongly upon the state of preservation of the grains. During their primary or subsequent oxidation, the wall structure alters and secondary structures come into being. First, the tubules begin the resolve, then the thin walls of the calyces. In the former case the calyces are surrounded radially by small oval hollows (Plate IV, Figs 2–6), in the latter one, they are transformed into irregular, big, vesicular hollows (Plate V, Figs 2, 3; Plate XV, Figs 2–4). At an advanced stage of the resolution process, the grains are bound together only by the outer wall. Because of these misleading structures, the author feels it necessary to present the photos of as many grains as possible.

Oraveczia hungarica nov. gen. et sp.

Figs of Plates I–IV and Text-fig. 1

Derivatio nominis: named after its occurrence in Hungary

Holotype: 37; 59885; grain under coord. 17.6–105.1; Plate I, Figs 1–8

Locus typicus: Csővár, Big Quarry

Stratum typicum: light grey, hard, bituminous, calcareous marl, the basal layer of the quarry in 1961, lowermost Rhaetian

Diagnosis: large-sized, compound thick-walled plant microfossil with circular or slightly oval outline in E-view, the outer wall of which is thin and smooth, while the inner one is thick and embedded in a tubular matrix. It consists of thin-walled, small calycular elements, empty inside, reminiscent of the calyx of the tulip, as well as of a thin cover lamella closing the inner surface of the wall.

Description: the outer ends of the small calycular elements arch inward along the outer wall and in the upper third of the height of the calyces they terminate in a thin-walled ring with a 2–3 opening. On the surface of the grain the calyces

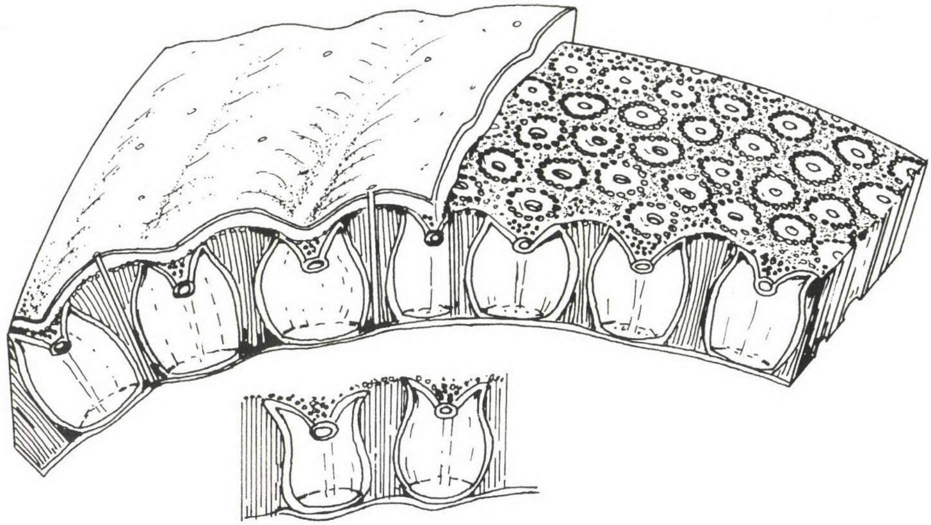


Fig. 1
Oraveczia hungarica nov. gen. et sp. – Genotype

are of regular circular shape in E-view and are arranged in regular lines along the arcs. The matrix between the calyces is densely inwrought with tubules thinner than 1 µm. In length the tubules reach from the base to the outer wall. Among them, wider small tubes (rarely of 1.0–1.2 µm diameter) can be found sporadically, which also penetrate into the outer wall in some places.

Size: the size of the grains which can be assigned here for certain, is between 250 and 290 µm. The greatest diameter of the type specimen is 250 µm. The thickness of the outer wall is 2.5–3 µm, height of the inner wall is 15–16 µm, its width is 12–14 µm. Their distance from each other within one line is 2–4 µm. Wall thickness of the calyces and the ring is 0.5–0.8 µm. Thickness of the inner cover lamella is about 1 µm.

Remark: it is a frequent species. On the type specimen, slight resolution can be observed.

Oraveczia doliola nov. gen. et sp.

Figs of Plates VI–VIII and Text-fig. 2

Derivatio nominis: named after its small barrel- (doliolum in Latin) like elements

Holotype: 71; grain under coord. 13.8–109.4; 3/1–8; Plate VI, Figs 1–7

Locus typicus: Csővár, Big Quarry

Stratum typicum: light grey, hard, bituminous, calcareous marl, the basal layer of the quarry in 1961, lowermost Rhaetian

Diagnosis: large-sized, compound thick-walled plant microfossil with circle or slightly oval outline in E-view, the outer wall of which is thin and smooth, while the inner one is thick and embedded in compact matrix. It consists of thin-walled small barrels, filled inside with tubules, as well as of a cover lamella closing the inner surface.

Description: outer ends of the small barrel-like elements are concave along the outer wall and in the upper third of the height of the barrels, in the middle, they terminate in a thin-walled ring of 1–2 opening. On the surface, the barrels are of circular shape in E-view and are arranged in regular lines along the arcs. In the individual lines, they are situated more or less at the same distance from each other.

Differential diagnosis: *Oraveczia doliola* nov. sp. shows the greatest similarity to *O. hungarica* nov. sp. Its tubular small barrels and its median wall without tubules, however, definitely separate it from *O. hungarica* nov. sp. which has a tubular median wall, empty inside and an inner wall built up by small calyces.

With its tubular structural elements, it also resembles *Oraveczia galeata* nov. sp. as well. The mitre-shaped inner wall structural element of *O. galeata*, however, is not tubular as opposed to that of *O. doliola*; some small pipes following the contour of the mitre can be found only in the wall of the mitre. The mitre-like structural element, however, is empty inside, while that of *O. doliola* is filled by tubules.

Size: the size of the type specimen is 273 µm. Its wall thickness is 19–21 µm, of which 2.0–2.5 µm falls on the outer wall. Greatest width of the small barrels is

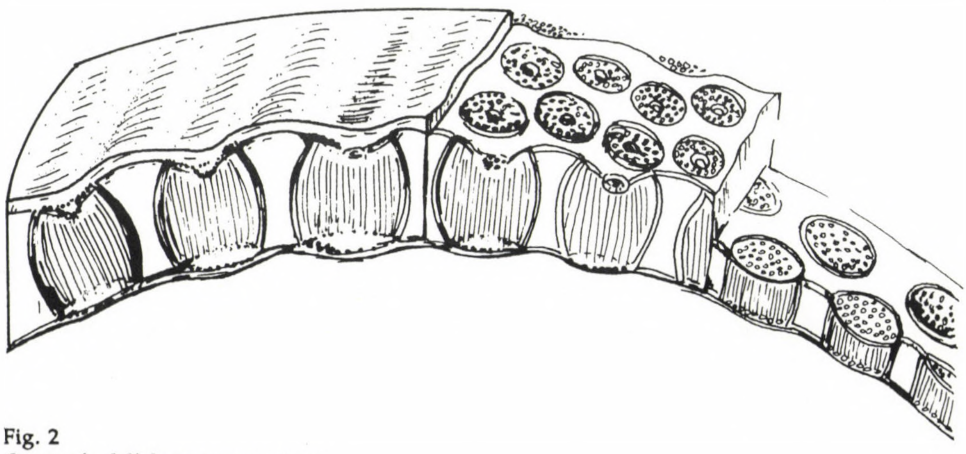


Fig. 2
Oraveczia doliola nov. gen. et sp.

12–13 µm, their distance from each other within the same line is about 2–4 µm. Thickness of the inner cover lamella is under 1 µm.

Remark: in the type layer, it is of frequent occurrence.

Oraveczia galeata nov. gen. et sp.

Figs of Plates IX–X and Text-fig. 3

Derivatio nominis: named after its helmet- (galea in Latin) like inner wall structural element

Holotype: Plate IX, Figs 1–9

Locus typicus: Csővár, Big Quarry

Stratum typicum: light-grey, hard, bituminous, calcareous marl, the basal layer of the quarry in 1961, lowermost Rhaetian

Diagnosis: large-sized, organic framework plant microfossil with circular or slightly oval outline in E-view, the compound thick wall of which is composed of a thin outer and a thick inner layer. The outer wall is smooth and unornamented. The inner wall is built up by royal mitre-like structural elements embedded in a compact matrix and by cover lamella closing the inner surface of the wall.

Description: the mitre-like structural elements are empty inside. In their walls of about 1 µm thickness 7–8 µm thin channels run in even distribution, following the outer outline of the mitre. Upper end of the mitre, along the outer wall, is slightly introsuscepted and terminates in a 2–3 µm wide, thin-walled ring with

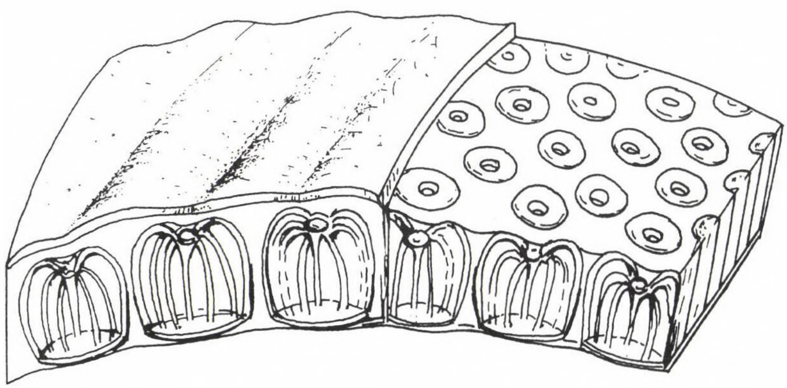


Fig. 3
Oraveczia galeata n. sp.

an approx. 1 μm opening. On the surface, the mitres are of regular circular shape in E-view, in their centre with the closing ring and its narrow opening. They are arranged in regular lines along the arcs, close to each other. The cover lamella closing the inner surface is thinner than 1 μm.

Size: greatest diameter of the holotype is 283 μm. Thickness of the wall is 20 μm, of which 2 μm belongs to the outer wall. Width of the mitres is 17–18 μm on the margin of the grain, the diameters are 13–14 μm on the surface.

Differential diagnosis: Oraveczia galeata nov. sp. shows the greatest similarity to *O. doliola* nov. sp. Its structural elements of rarely piped mitre, however, distinguish it easily from *O. doliola* nov. sp. which has wall structural elements filled with foveae and having the shape of small barrels. Likewise, it can be separated from *Oraveczia faveola* (Morbey 1975) nov. comb. and emend., having similarly tubularless median wall and inside empty structural elements reminiscent of tulip calyces on the basis of this feature.

Remark: it is a species of rare occurrence. The specimen shown on Plate X, Figs 1–5 was only assigned to *O. galeata* nov. sp. with the addition of "cf.", because although the photos taken of the surface of the grain in E-view show the marks of this taxon, on the near-margin photos the median parts and also the inside of the small mitre-like elements represent a secondary structure which makes the certain classification of the grain impossible.

Oraveczia faveola (Morbey 1975) nov. comb. and emend.

Figs of Plates XI–XV and Text-fig. 4

Synonima:

Tytthodiscus? faveolatus nov. sp. Morbey, S.J. 1975, p. 45–46. Pl. XVIII, Figs 1a–c, Pl. XIX, Figs 1a–b

Emended diagnosis: large-sized, compound thick-walled plant microfossil with circular or slightly oval outline in E-view, the wall of which is composed of a thin, smooth layer and a thick, composite, inner one. The inner wall is built up by thin-walled structural elements, empty inside, reminiscent of the calyx of the tulip, embedded in the compact matrix. Along the outer wall are outer parts of the small calyces, hollow like a funnel in the centres, and in the upper third of the calyces they terminate in a 2–3 μm wide ring. Along their basis, the calyces are closed and – covered by a 1.0–1.5 μm thick lamella – they constitute the inner surface of the wall.

Differential diagnosis: by the tubularless matrix of its inner wall and by its smooth, thin-walled, inside empty small calyces embedded in the matrix, the *Oraveczia faveola* (Morbey 1975) nov. comb. and emend. can be well separated from the *O. galeata* nov. sp., the *O. hungarica* nov. sp. as well as the *O. doliola* nov. sp., the inner walls of all of which are piped in a different way and degree.

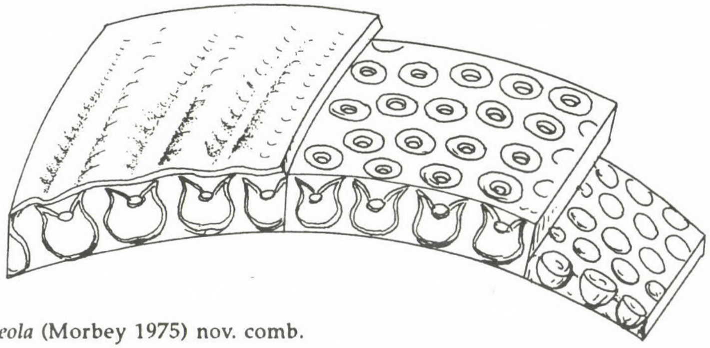


Fig. 4
Oraveczia faveola (Morbey 1975) nov. comb.

Remarks: despite the difference between the description by Morbey (1975) and my opinion on the wall structure of *O. faveola,* I believe – on the basis of the presented photos – that the type specimen of *faveola* and my specimens are the members of the same taxon.

The punctate structure, which is described by Morbey in his diagnosis (p. 45), presented in its drawn figures (p. 46) and documented by the photo of the type specimen magnified 1500 times (Pl. XVIII, Figs 1a–b), is in all certainty the result of the state of preservation of the specimen. Similar phenomena can also be observed on my specimens in the cases of different resolution stages. On the intact surface and wall sections, however, it can be unambiguously established that the tubulosity characterizing the remaining *Oraveczia* species is missing on the specimens of *O. faveola* (Morbey 1975). Certainly an original fine porosity of the compact wall cannot be excluded. *O. faveola* (Morbey 1975) is one of the most frequent taxa in the basal layers of the Csővár Quarry.

Forma genus: *Vadaszia* nov. gen.

Derivatio nominis: in memory of the former renowned investigator of the Triassic at Csővár, Prof. Elemér Vadász

Genotype: *Vadaszia cavernosa* nov. gen. et sp.

Genus diagnosis: large-sized plant microfossil with regular circular outline in E-view and with cavernous inner structure.

Description: wall of the circular or discus-shaped body consists of a thin outer and a thick inner layer. Their total thickness may attain 1/6 of the radius of the body. The thin outer wall may be smooth or foveolate. The compact matrix of the inner layer is cavernous. The foveae of the outer layer appear above the caverns of the inner wall, usually as holes of some diameter. Caverns of the inner wall appear to be hollows larger than a semicircle along the margin of the grain and regular circles in E-view on the surface. In the inner wall, tubulosity can not be observed.

Differential diagnosis: by its cavernous, thick, inner wall structure without pores or tubules, *Vadaszia* nov. gen. differs from all the large-sized, thick-walled plant microfossils known so far. As to the shape and wall thickness, it shows the greatest similarity to *Oraveczia* nov. gen. Its characteristic cavernous inner wall structure, however, distinguishes it easily from *Oraveczia* nov. gen. with an inner wall structure built up by calyx-like, mitre-like or small barrel-like elements as well as compact or tubular matrix.

There is also a significant difference between the two genera in the size of the grains: with few exceptions, specimens of *Vadaszia* nov. gen. found so far are smaller than 200 μm (mean value: 183 μm), while grains belonging to *Oraveczia* nov. gen. are of sizes between 220 and 380 μm (mean value: 283 μm).

Remark: smooth or foveolate character of the outer, thin covering layer of *Vadaszia* nov. gen. can only be observed on the totally intact specimens, which occur rarely.

Vadaszia cavernosa nov. gen. et sp.

Figs of Plates XVI–XXI

Derivatio nominis: named after its cavernous (cavernosus in Latin) inner wall

Holotype: 39; 59882; grain under coord. 15.0–111.5; Pl. XVI, Figs 3–5, and Pl. XVII–XVIII.

Locus typicus: Csővár, Big Quarry

Stratum typicum: light grey, hard, bituminous, calcareous marl, the basal layer of the quarry in 1961, lowermost Rhaetian

Diagnosis: large-sized, thick-walled organic framework plant microfossil with regular circular outline in E-view, the outer wall of which is built up by a thin, smooth or foveolate layer and the inner one by a thick, cavernous layer.

Description: the foveae of the outer wall open above the caverns of the inner wall in the form of round holes of some diameter. The caverns, embedded in the compact matrix of the inner wall, are hollows larger than a semicircle, situated at more or less regular distance from each other. On the surface, in E-view, they seem to be granulae of spherical shape, but on the margin of the grain, in optical section, it can be well seen that they are hollows and not granulae.

Size: diameter of the type specimen is 164 μm. Thickness of the wall is 10.0–11.0 μm, out of which 1.0–1.5 μm belongs to the outer wall and 9.5–10.0 μm to the inner one. Greatest diameter of the caverns is 9 μm. Size of the specimens of the paratypes is between 147 and 176 μm.

Remark: well-preserved specimens assignable to *Vadaszia cavernosa* nov. sp. show the same relative size. The only specimen larger than 200 µm found so far, displayed in Figs 1–2 of Plate XXI, may even represent a new species, characterized by the shallower caverns and thicker outer wall as compared to its size. Since it cannot be excluded that this appearance is only a consequence of more intense flattening, it is assigned to this taxon until the appearance of new, similar specimens.

Forma genus: *Sulcodiscus* nov. gen.

Derivatio nominis: named after its sulcated (sulcus in Greek) discus shape

Genotype: *Sulcodiscus trisulcatus* nov. gen. et sp.

Genus diagnosis: large-sized, thick-walled, tubular plant microfossil with sulcated surface and with discus outline in E-view.

Description: thickness of the wall may reach 1/7 of the greatest diameter. Its tubules (small pipes with a diameter approximating 1 µm) are of the same size, stand densely, penetrate through the entire thickness of the wall and open onto the surface. Their evolvement is definite, they are situated symmetrically on the surface along the longitudinal axis. Their length may reach half of the greatest diameter of the grain. They may be closed (suture-like) or open.

Differential diagnosis: by its fully developed sulci, *Sulcodiscus* nov. gen. differs fundamentally from all thick-walled, tubular plant microfossils of discus shape known so far. With its discus outline in E-view and tubular thick wall, it shows the greatest similarity to the genus *Tytthodiscus* Norem 1955. Its trisulcate structure, however, also separates it sharply from the *Tytthodiscus*.

Remark: the sulcus as a germinal aperture is a structural element of the pollens of the flowering plants. So far, trisulcate exine has not been known among the microfossils of marine plants. The perforated suture line of the operculum, occurring among the alga cysts, takes a mostly circular, elliptical, or angular shape, or a combination thereof. Nevertheless, I put the grains assignable to *Sulcodiscus* nov. gen. among those microfossils which can be regarded as alga cysts, and where also the genera *Tasmanites*, *Tytthodiscus*, *Oraveczia* and *Vadaszia* may belong, on the following considerations:

– degree and colour of carbonization are totally identical with those of the representatives of the genus *Tytthodiscus*, occurring together with it;

– tubulosity of the wall is of expressly *Tytthodiscus* character;

– degree and colour of carbonization of bisaccate and operculate pollen grains gained from the same formations by the same method of maceration differ sharply from those of it.

***Sulcodiscus trisulcatus* nov. gen. et sp.**

Plate XXII, Figs 1–5

Holotype: 37; 59885; grain under coord. 12.8–104.1; Plate XXII, Figs 1–3

Derivatio nominis: named after its three sulci

Locus typicus: Csővár, Big Quarry

Stratum typicum: light grey, hard, calcareous marl, the basal layer of the quarry in 1961, lowermost Rhaetian

Diagnosis and description: the same as those of the genus.

Size: size of the holotype is 70 x 60 μm. Its wall thickness is 5 μm. Length of the sulci is 34–35 μm. Their distribution is symmetrical on the surface of the grain. Density of the tubules within 10 mm, magnified 1000 times, is 12–14. The greatest diameter of the paratypes is between 68–70 μm.

Remark: in the basal layers, it occurs together with the following sporomorph taxa:

Riccisporites tuberculatus Lundblad 1954
Rhaetipollis germanicus Schulz 1967
Classopollis cf. *torosus* (Reiss. 1950) Balme 1957
Corollina meyeriana (Klaus 1960) Venk. et Gócz. 1962
as well as with the representatives of *Oraveczia* nov. gen., *Vadaszia* nov. gen. and *Tytthodiscus tubulatus* nov. sp.

Additional occurrence: samples 1–3 and 46 in the section of the quarry (uppermost layer).

Forma genus: *Tytthodiscus* Norem 1955

***Tytthodiscus tubulatus* nov. sp.**

Plate XXIII, Figs 1–5, Plate XXIV, Figs 1–5

Derivatio nominis: named after its piped wall

Holotype: 45; grain under coord. 11.5–118.6; D-I/13–18; Plate XXIII, Figs 1–5

Locus typicus: Csővár, Big Quarry

Stratum typicum: light grey, hard, calcareous marl, the basal layer of the quarry in 1961, lowermost Rhaetian

Diagnosis: large-sized, thick-walled, tubular organic framework microfossil of discus shape, tubulosity of which is provided by small pipes having an opening smaller than 1 μm and standing very densely. Its outer surface is closed by a thin cover lamella.

Description: wall thickness of the grain of discus outline in E-view may reach 1/10 of the greatest diameter. Tubules are open on the inner surface. They do not break through the outer cover lamella. Their arrangement along arcs on the surface gives a hexagonal outline to the matrix. The scarcer small pipes of wider opening can be focused as spines emerging from the surface. Along the margin of the grain, however, it can be established for sure that they also do not break through the cover lamella.

Size: size of the type specimen is 110 x 90 μm; thickness of the wall is 10–11 μm; the cover lamella is thinner than 1 μm. As to the density of the tubules, within 10 mm, magnified 1000 times, 15 small pipes can be counted.

Differential diagnosis: by the very fine, densely standing tubules of its wall and by the tubulosity consisting of small pipes of larger diameter situated at regular distances between them, *Tytthodiscus tubulatus* nov. sp. differs from all *Tytthodiscus* species described so far. It shows the greatest similarity to *Tytthodiscus suevicus* Eisenack 1957 (pp. 241–243, Taf. XIX, Figs 1–3) described in the Swabian Liassic and to *Tasmanites yarboensis* Pocock 1972 (p. 108, Pl. XXV, Figs 2–3) published from the Jurassic sediments of Western Canada.

From *T. suevicus* Eisenack 1957, it can be distinguished mainly by its much finer tubulosity and from *T. yarboensis* Pocock 1972 by its simpler exine structure. *T. yarboensis* has a three-layered wall, while *T. tubulatus* has only two layers even counting the outer cover lamella.

Remark: its occurrence is regular in the basal layers.

Acknowledgement

I would like to express my thanks to Dr. János Oravecz, Principal University Assistant, for letting me have his original collection and photo documentation for publication.

Plate I

Oraveczia hungarica nov. gen. et sp. Genotype, 250 μm

1. Photo of the entire genotype, 37; 59885; 17.6–105.1; 250 x.
2–6 Sections of the wall structure, 1000 x
2–5. Elements reminiscent of the calyx of the tulip, of the inner wall
3, 6. Thickness of the outer wall is 2–3
5, 6. Transition between the margin and surface with the slightly resoluted tubules
7, 8. Sections of the surface – circle-shaped outlines of the small calyces with the ends of the slightly resoluted tubules of the matrix, 1000 x

Plate II

Oraveczia hungarica nov. gen. et sp. Paratype, 280 μm

1. Photo of the entire paratype, 300 x
2. Surficial section of the calyces with the upper ends arching inwards in their middle, 250 x
3–8. Sections of the wall structure, 1000 x
3. The small calyces with the outlines of the closing ring and the inner cover lamella
4–6. Outlines of the small calyces, tubules of the matrix
7–8. The small pipes with a larger opening of about 1 μm, which can be rarely observed, penetrate clearly into the outer cover lamella
9–15. LO-analysis on the same point of the surface, from both sides of the surface – on 9, 10: the closing ring, on 12: opening of the closing ring, 1000 x

Plate III

Oraveczia hungarica nov. gen. et sp. Paratype, 283 μm

1. Photo of the entire crushed grain, 300 x, 53; 5.0–109.0.
2–5. Sections of the surface
2–3. Sections of the arrangement of the small calyces, 345 x
4–5. The small calyces with their closing rings, openings of the closing rings and the tubulosity of the matrix of wall, 1000 x

Plate IV

Oraveczia cf. *hungarica* nov. sp. Grain with exine resolved in medium degree, 296 μm

1. Photo of the entire grain, 300 x.
2–6. Sections of the surface: amalgamated small vesicles of the tubules of the matrix of wall and their radial arrangement around the small calyces, also beginning to amalgamate
2–4. 1000 x
5–6. 2000 x

Plate V

Oraveczia cf. *hungarica* nov. sp. Grain with exine resolved in high degree, 373 μm, 66; 9.6–113.2; 3/21–24

1. Photo of the entire grain, held together by the outer wall, 250 x
2. Section near the margin of the grain. Small calyces have already amalgamated into huge vesicles, wall material and fragments of the fine tubules of the structural elements are still recognizable, and in some places even the outlines of the closing rings can be distinguished. 1000 x
3. Photo of the wall along the margin of the grain. Thickness of the outer wall can be still measured, but characteristic inner wall elements reminiscent of the calyx of a tulip of *O. hungarica* have already disappeared. 1000 x

Plate VI

Oraveczia doliola nov. gen. et sp. Holotype: 71; 13.8–109.4; 3/1–8; 273 μm

1. Photo of the entire type specimen, 250 x.

2–4. Sections of the wall structure, 1000 x

2–3. Thickness of the outer wall and inner cover lamella as well as the outlines and the outer concave sides of the small barrels can be observed

4. Section near the margin of the grain – near-surface outlines of the scarce tubulus with a wider opening, penetrating into the outer wall, and of the small barrels

5–7. Sections of the surface – outlines of the small barrels in E-view with their arrangement along arched lines and their inside filled by tubules. 1000 x

Plate VII

Oraveczia doliola nov. gen. et sp. Paratype: 40; 59881; 4.5–114.5; 2/1–7; 261 μm

1. Photo of the entire ripped-up grain, 250 x
2. Section of the fractured wall. 1000 x. Outlines of the structural elements of the inner wall cannot yet be established

3–4. Section near the margin of the grain – outer, concave outlines of the small barrels can be still observed in the wall. 1000 x

5–6. Sections of the surface in E-view. Diameter and tubulosity of the calyces can be measured, tubularless character of the matrix of wall can be also established. 1000 x

Plate VIII

Oraveczia doliola nov. gen. et sp. Paratypes

1, 7. Photos of the entire paratypes, 300 x

2–6. Sections near the surface and the margin of the grain No. 1, 1000 x

2. Thickness of the outer wall and the inner cover lamella

3–6. The small barrels and their inner tubulosity

3. Small barrels of semi-lateral position with their inner tubulosity and the thickness of the walls of the barrels. 1000 x

8–9. Sections of the surface of the grain No. 7. 1000 x

Plate IX

Oraveczia galeata nov. gen. et sp. Holotype: 283 μm

1. Photo of the entire type specimen, 300 x
2–6. Sections of the wall structure. 1000 x. Wall thickness, outlines, outer infolding of the outer wall, inner cover lamella and small mitre-like elements as well as the opening of the closing ring (Figs 2, 3), and the canalicularity of the wall of the mitres, can be easily observed
7–9. Sections of the surface in E-view and in near-E-view cutting with the outlines, arrangement of the small barrels, closing rings of the small mitres, their opening and the apertures of the small wall pipes ending in rings (Fig. 9). 1000 x

Plate X

Oraveczia cf. *galeata* nov. gen. et sp. Strongly corroded specimen, 66; 13.2–115.0; 3/16–17; 331 μm

1. Photo of the entire grain, 250 x
2. Near-margin photo of the grain, 1000 x. Thickness of the outer wall and the inner cover lamella can still be measured, but outlines of the structural elements of the inner wall can only be estimated. Contours appearing to be the inside of the elements and the tubulosity of the matrix are in all certainty secondary structures which represent one of the stages of the inner resolution process and not the original wall structure
3–5. Sections of the surface in E-view in the better-preserved part of the grain. These photos of the inner wall elements show *O. galeata* structure, but the structure visible on Fig. 2 does not confirm it. 1000 x

Plate XI

Oraveczia faveola (Morbey 1975) nov. comb. and emend. Specimen of slightly open surface, 266 μm

1. Photo of the entire grain, 300 x
2–4. Sections of the wall structure – thickness of the outer wall and structural elements of the inner wall can be well observed, 1000 x
5–8. Sections of the surface in E-view, 1000 x
5. Matrix displaying the initial stage of the structural elements and the resolution
6–8. Photos, taken nearly in E-view, of the calyces without tubulosity and the matrix. In Fig. 8 the matrix can also be observed in a state of initial resolution, reminiscent of pipe ends

Plate XII

Oraveczia faveola (Morbey 1975) nov. comb. and emend. Strongly open specimen, 266 μm

1. Photo of the entire grain, 300 x
2–4. Sections of the preserved structure of the wall and the near-margin surface – thickness of the outer wall, outlines of the small calyces as well as the tubularless essential elements and the matrix, 1000 x
5–6. The essential elements in E-view. Due to the resolution, certain elements have been already transformed into bigger vesicles. On the more intact parts, however, it can be observed that neither the small calyces nor the matrix are tubular. 1000 x

Plate XIII

Oraveczia faveola (Morbey 1975) nov. comb. and emend. Specimen with corroded surface, 226 μm

1. **Photo of the entire grain, 300 x**

2–5. **Sections of the wall and the near-margin surface – thickness of the outer wall, outlines of the essential elements of the inner wall, the closing ring and its opening (Fig. 2) can be still observed in some places. 1000 x**

Plate XIV

Oraveczia faveola (Morbey 1975) nov. comb. and emend. Surficial and near-surface sections of a specimen in the stage of advanced resolution, 276 μm; 64; 11.2–109.2

1–3. Thickness of the outer wall holding together the grain and the tubularless small calyces can still be easily observed in some places, 1000 x

4–5. Due to the resolution of the matrix of the wall, the small calyces have already been transformed into bigger vesicles. 1000 x

Plate XV

Oraveczia cf. *faveola* (Morbey 1975) nov. comb. and emend. Grain in the stage prior to total resolution, 344 μm

1. Photo of the entire grain; 66; 16.2–104.2; 3/27–28; 250 x. The photo shows the final stage of the resolution of the structural elements, when a majority of the small calyces have already lost their original shape and structure and amalgamated into vesicles of different size. Essentially, the grain is held together only by the outer wall

2–3. Thickness and continuity of the outer wall can be still observed, but not the outlines and size of the elements of the inner wall any more. 1000 x

4. Wall material of the elements of the dissolved inner wall, forming a bundle. 1000 x

Plate XVI

Vadaszia cavernosa nov. gen. et sp.

1. Solution residue from the stratum typicum (photo by Oravecz, J., 1961, cca. 40 x).
2. *Vadaszia cavernosa* nov. gen. et sp., cotype, 340 x (photo by Oravecz, J., 1961).

3–5. *Vadaszia cavernosa* nov. gen. et sp., genotype, 164 μm; 39; 59882; 10.7–111.5.

3–4. LO-analysis of the type specimen, 500 x; to be continued in Pl. XVII, Figs 1–2

5. Section of the wall structure - the cavern, larger than a semicircle, embedded in the compact material of the thin outer wall and the thick inner wall can be observed. 1000 x

Plate XVII

Vadaszia cavernosa nov. gen. et sp., genotype

1–2. Continuation of the LO-analysis of the type specimen, 500 x

3. Section of the wall structure – the thick, compact inner wall and the caverns embedded in it, 1000 x.
4. Section of the near-margin surface – caverns embedded in the inner wall and – on the near-margin place of the surface, under the thin cover layer – the outlines of caverns appearing to be verrucae can be observed. 1000 x

Pl. XVIII

Vadaszia cavernosa nov. gen. et sp., genotype

1–2. Sections of the surface of the type specimen, 1000 x
1. The roundish apertures in the thin outer cover layer and beneath them the round outlines of the caverns of the inner wall can be observed
2. Through the thin outer wall, round outlines of caverns embedded in the inner wall and arranged in lines can be observed

Plate XIX

Vadaszia cavernosa nov. gen. et sp. (paratypes)

1. Ripped-up specimen, 500 x; 39; 59882; 15.0–109.2; 176 μm
2. Grain under air-bubble with the outlines, appearing to be verrucae, of caverns situated in concentric lines. 500 x; 39; 59882; 15.0–109.1; 156 μm
3. Grain of 171 μm size under co-ordinates 18.0-108.7 in mount No. 39; 59882

Plate XX

Vadaszia cavernosa nov. gen. et sp.

1. Photo of an entire ripped-up specimen, 238 μm. 500 x; 39; 59882; 9.2–104.6.

2–3. Sections of the wall structure, 1000 x. Caverns embedded in the matrix of the inner wall and the tubularless matrix can be observed.

4. Near-margin cutting. Above the caverns, ripped up outlines of foveae situated in the thin outer wall can be seen

Plate XXI

1–2. *Vadaszia* cf. *cavernosa* nov. sp., 207 μm; 37; 59885; 9.5–99.0; 1/20–24
1. Photo of the entire grain with the ripped up outer wall, 250 x
2. Section along the ripped up outer wall, 1000 x. Caverns situated in the inner wall can be easily observed

3–4. *Vadaszia cavernosa* nov. gen. et sp., paratype, 147 μm; 39; 59882; 12.0–100.3; 2/37–38
3. Photo of the entire grain with surficial air-bubble, 500 x. Arrangement of the caverns appearing to be granulae in E-view can be well observed
4. Section of the wall structure with the thin outer wall and the caverns of the compact inner wall situated beneath it, 1000 x

Plate XXII

Sulcodiscus trisulcatus nov. gen. et sp.

1–3. Photo of the entire genotype in E-view, 70 x 60 μm. LO-pictures of the surficial and wall tubulosity of the two sides as well as of the aperture of the sulcus

4. Paratype in E-view, 68 x 60 μm. The cutting shows the apertures of the tubules opening onto the surface and their arrangement. 40; 59881; 9.7–110.3; 2/8–15; 1000 x
5. Paratype in P-view from the uppermost layer of the quarry, 70 μm. 43; 59879; 17.0–116.1; 1/34, 1000 x

Plate XXIII

Tytthodiscus tubulatus nov. sp., holotype, 110 x 90 μm

1. Photo of the entire holotype in E-view, 1000 x. Tubulosity of the wall can be well observed
2. Cutting of the entire type specimen, 1000 x. Dense, fine tubules and scarce tubules with a larger opening, ending on the surface, can be easily observed
3. Section of the wall structure with the fine and coarse tubules 2000 x

4–5. Sections of the fine tubules of the wall and the outer cover lamella as well as the ends of the tubules opening onto the inner surface. 2000 x

Plate XXIV

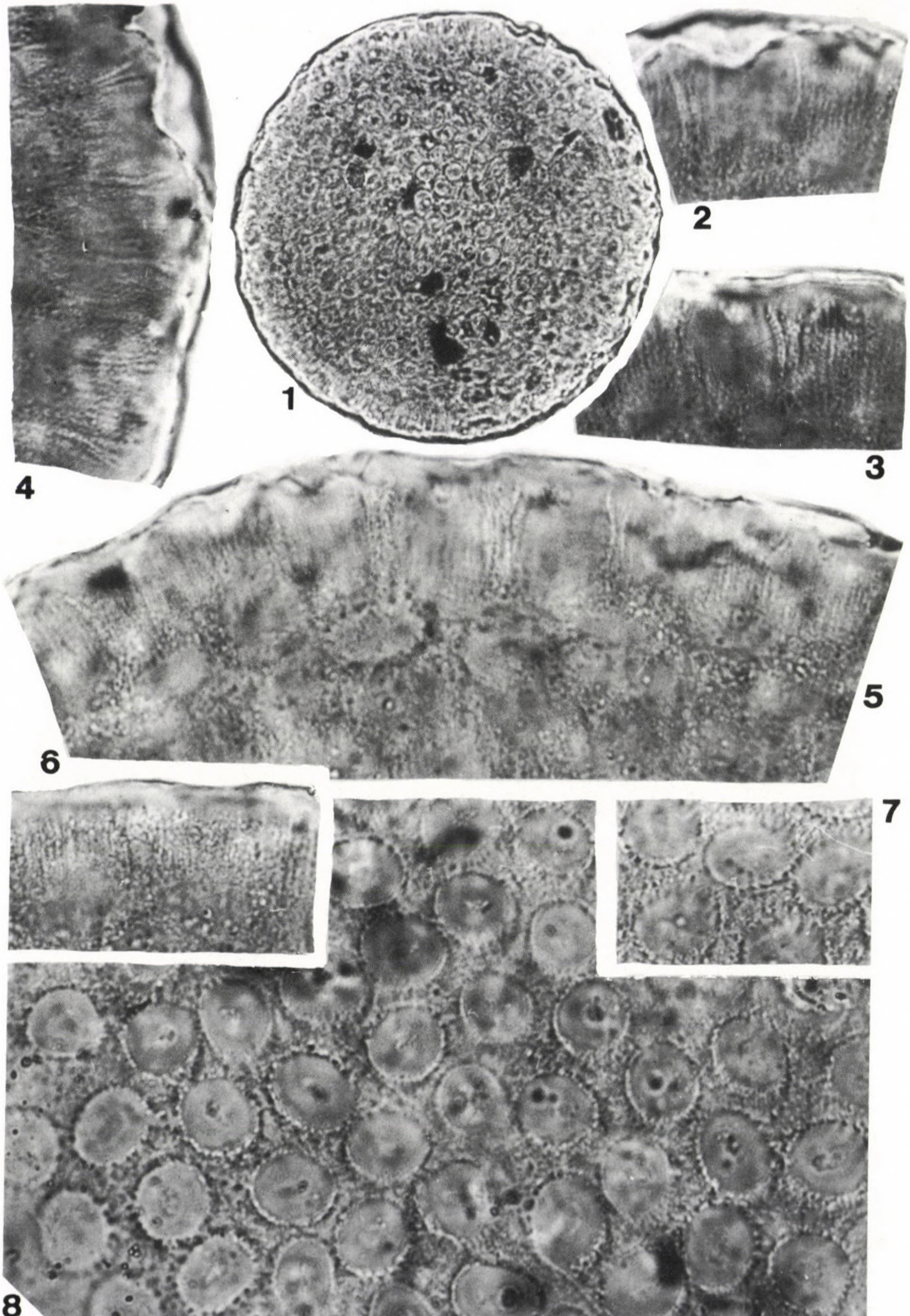
Tytthodiscus tubulatus nov. sp., paratype, 100 x 88 μm

1. Specimen of the paratype before ripping up, 1000 x. In the cutting, the wall with its densely standing fine and scarcely standing coarser tubules and their arrangement on the surface can be observed
2. The same grain after ripping up in E-view, 1000 x

3–4. Section of the tubules of the wall, their ends opening onto the inner surface and the outer cover lamella at the part of the grain marked by "x". 2000 x

5. Section of the arrangement of the fine tubules on the near-margin surface. 2000 x

Plate I



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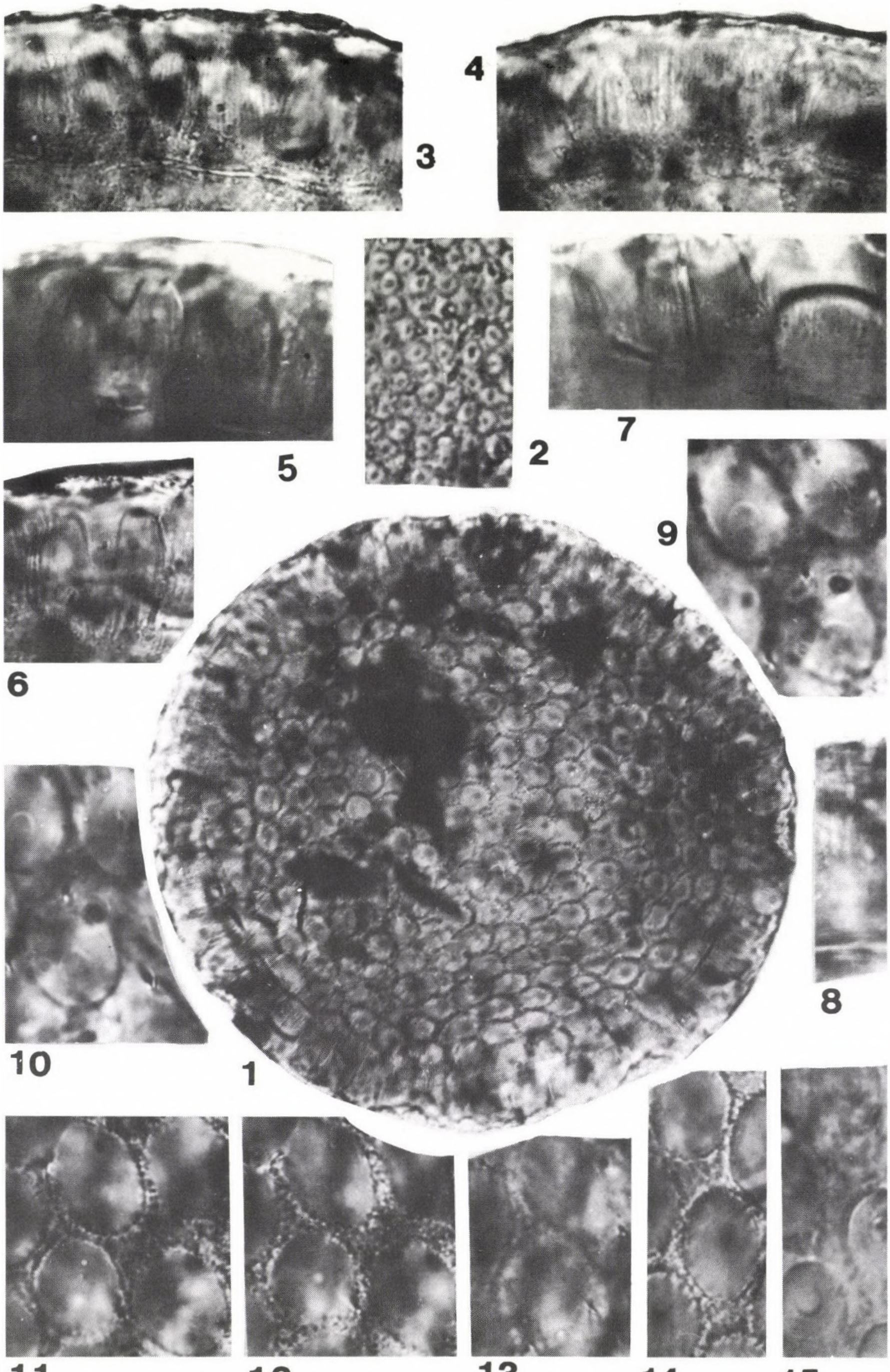

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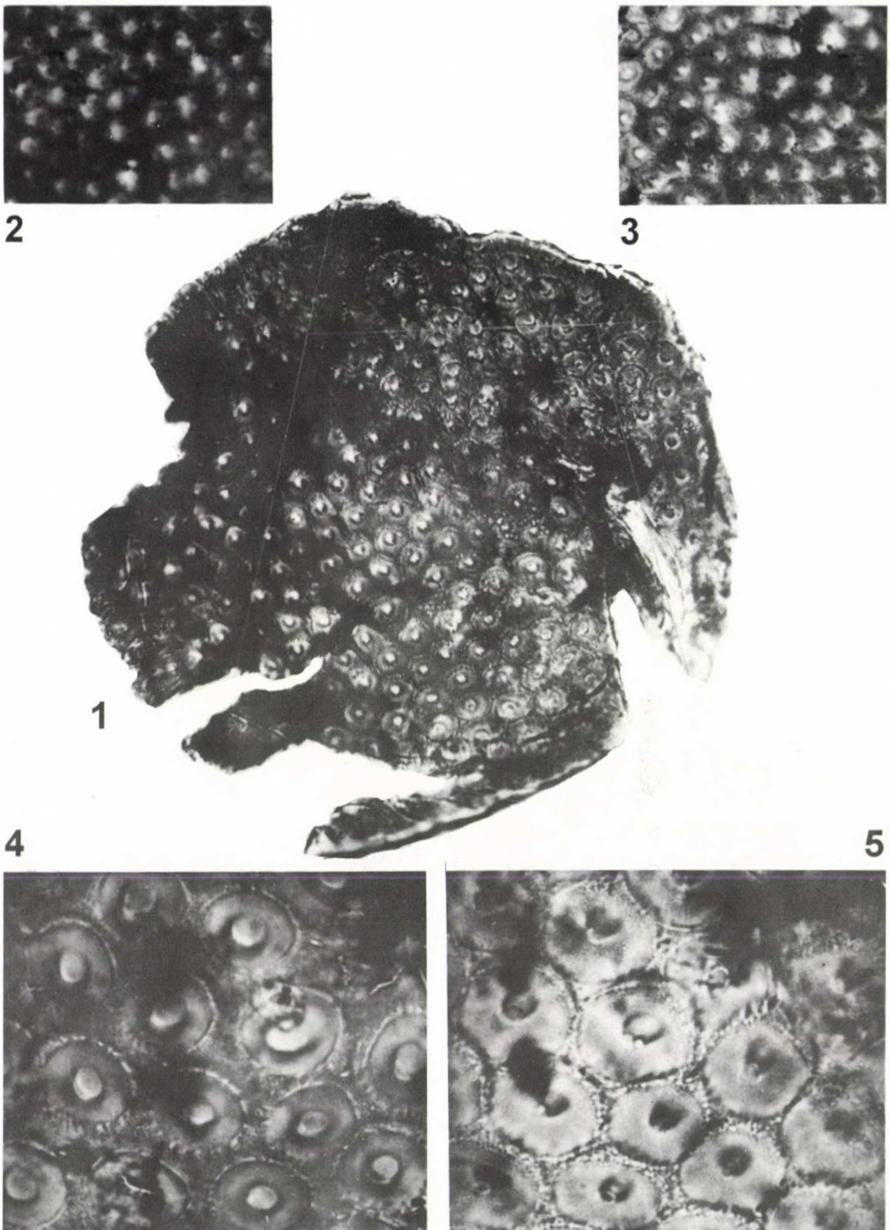

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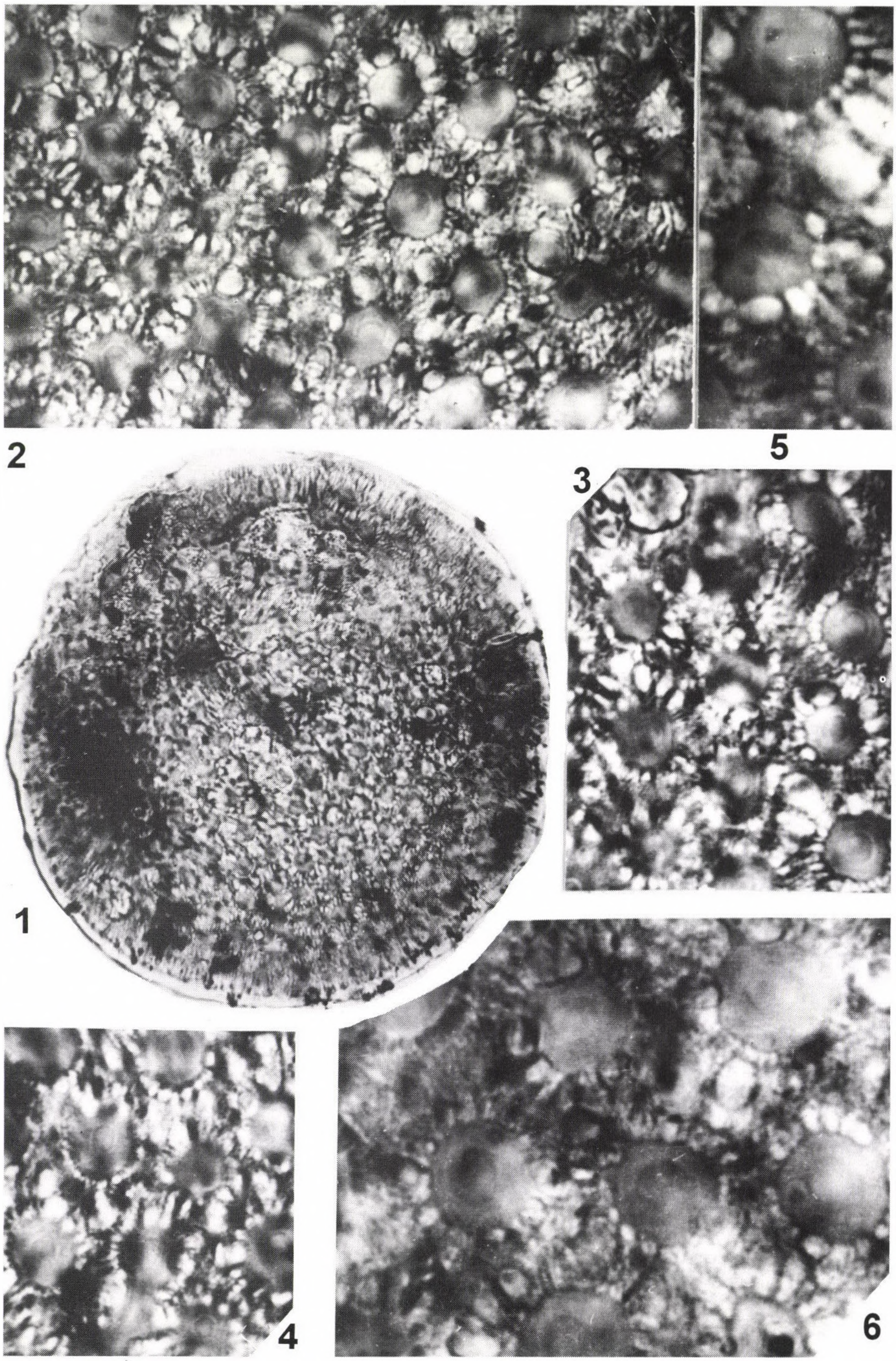

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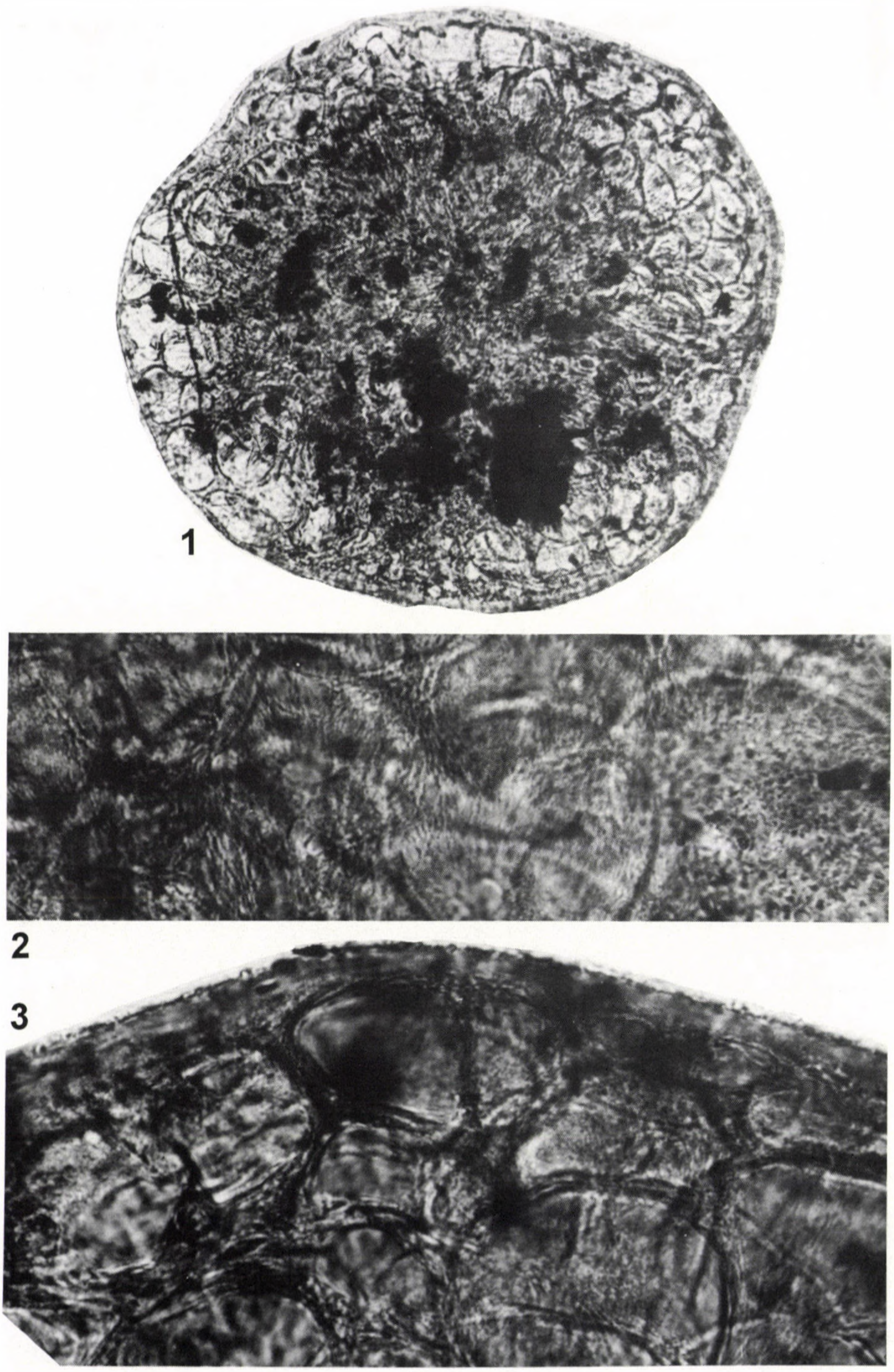

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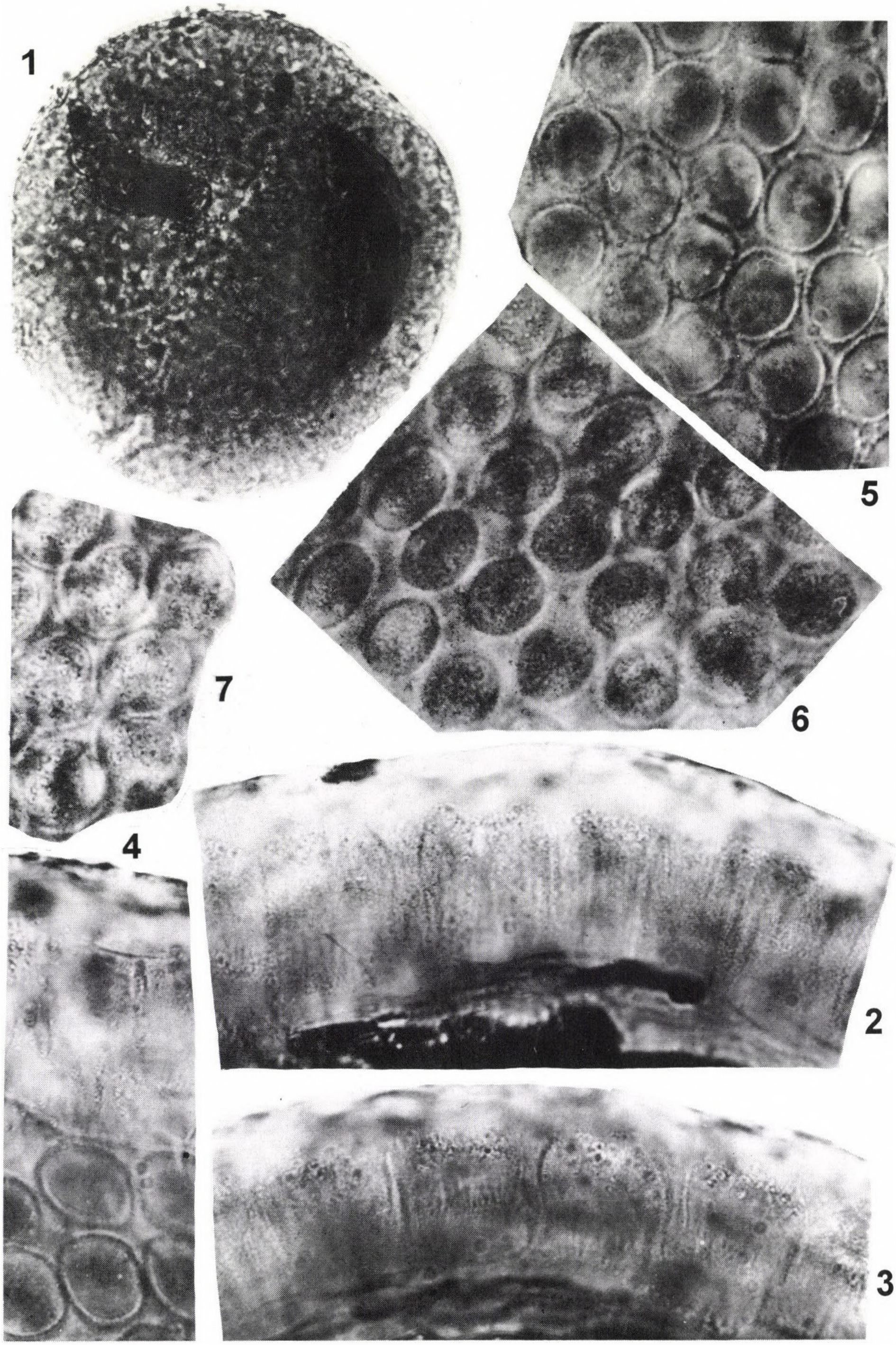

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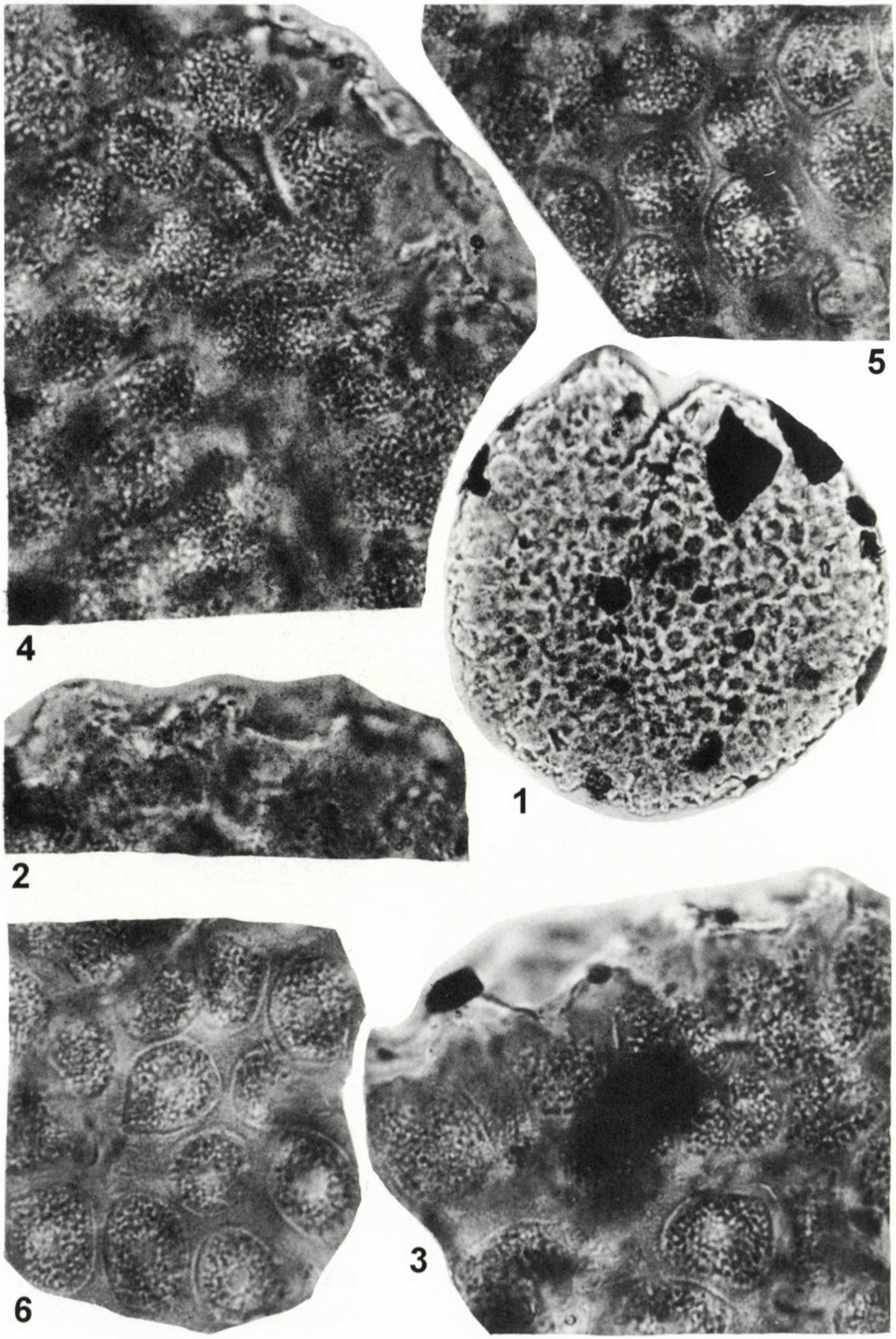

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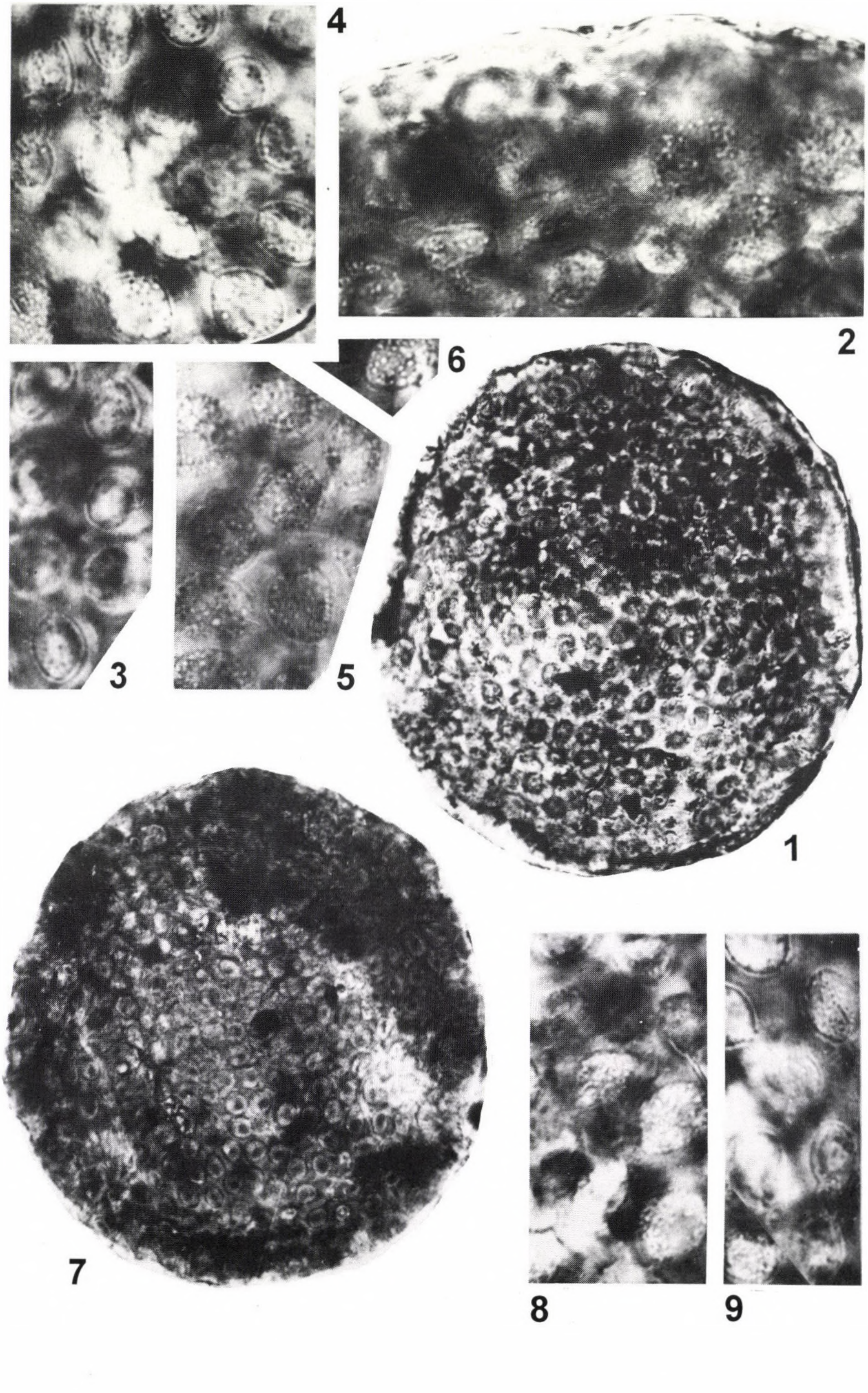

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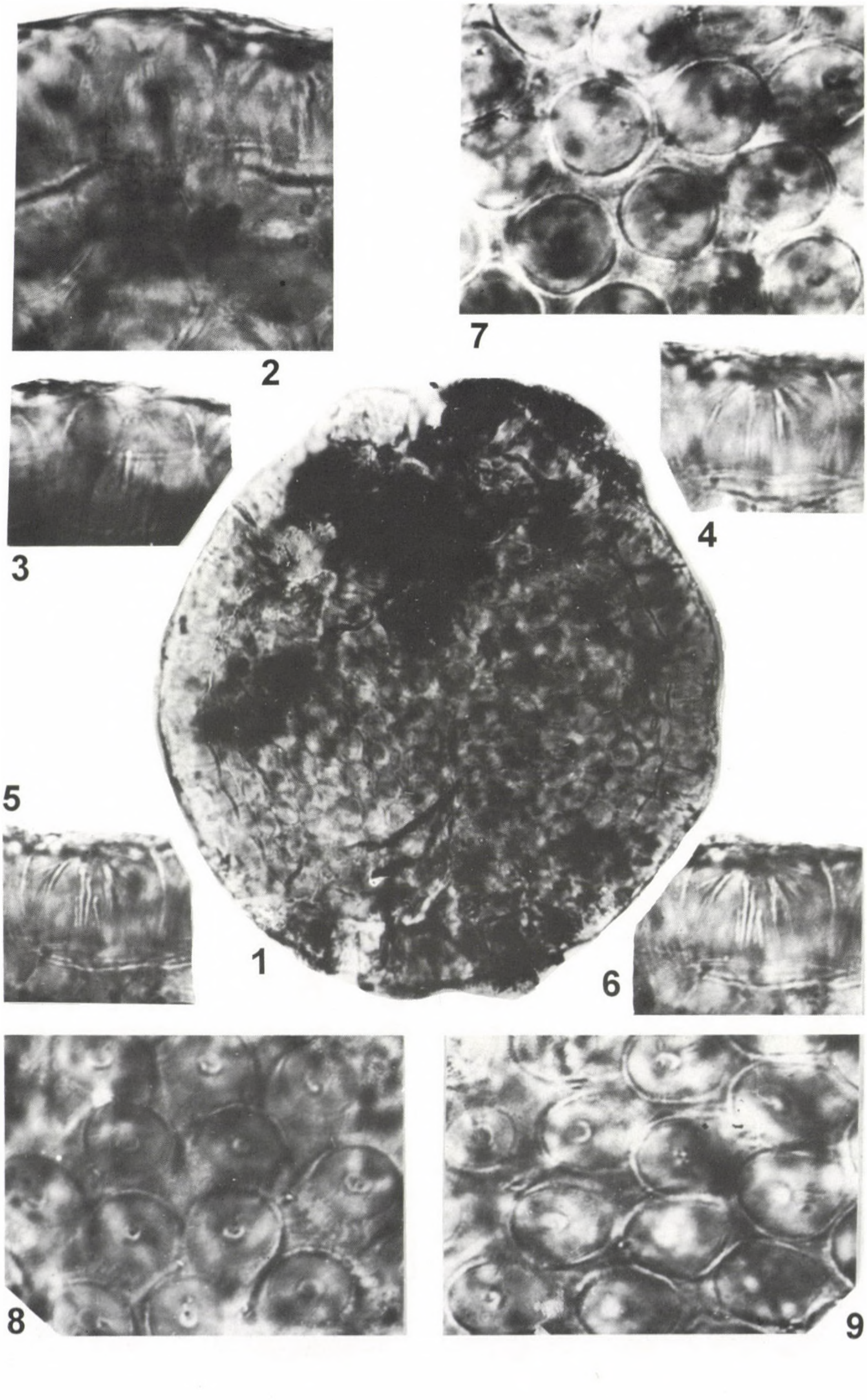

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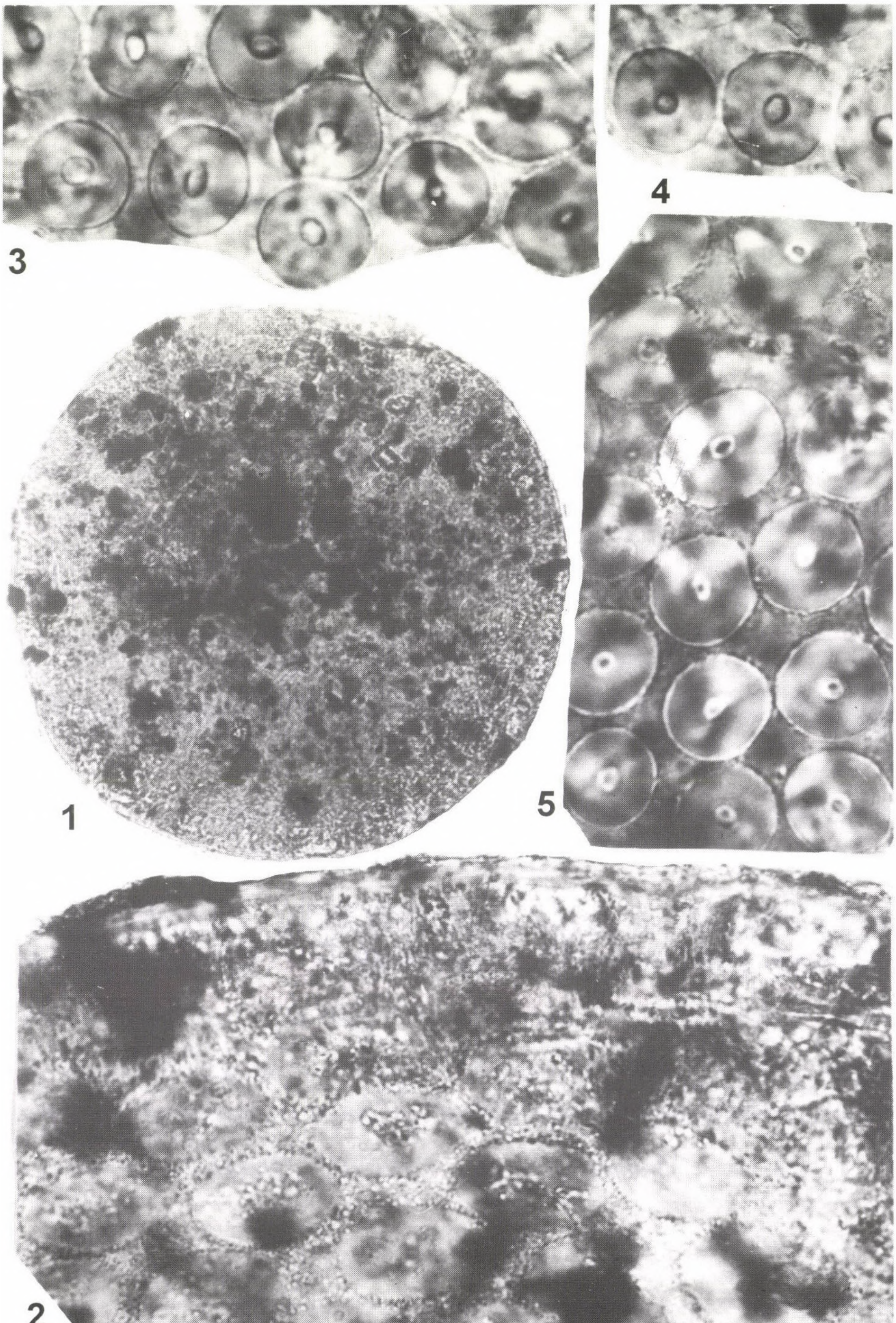

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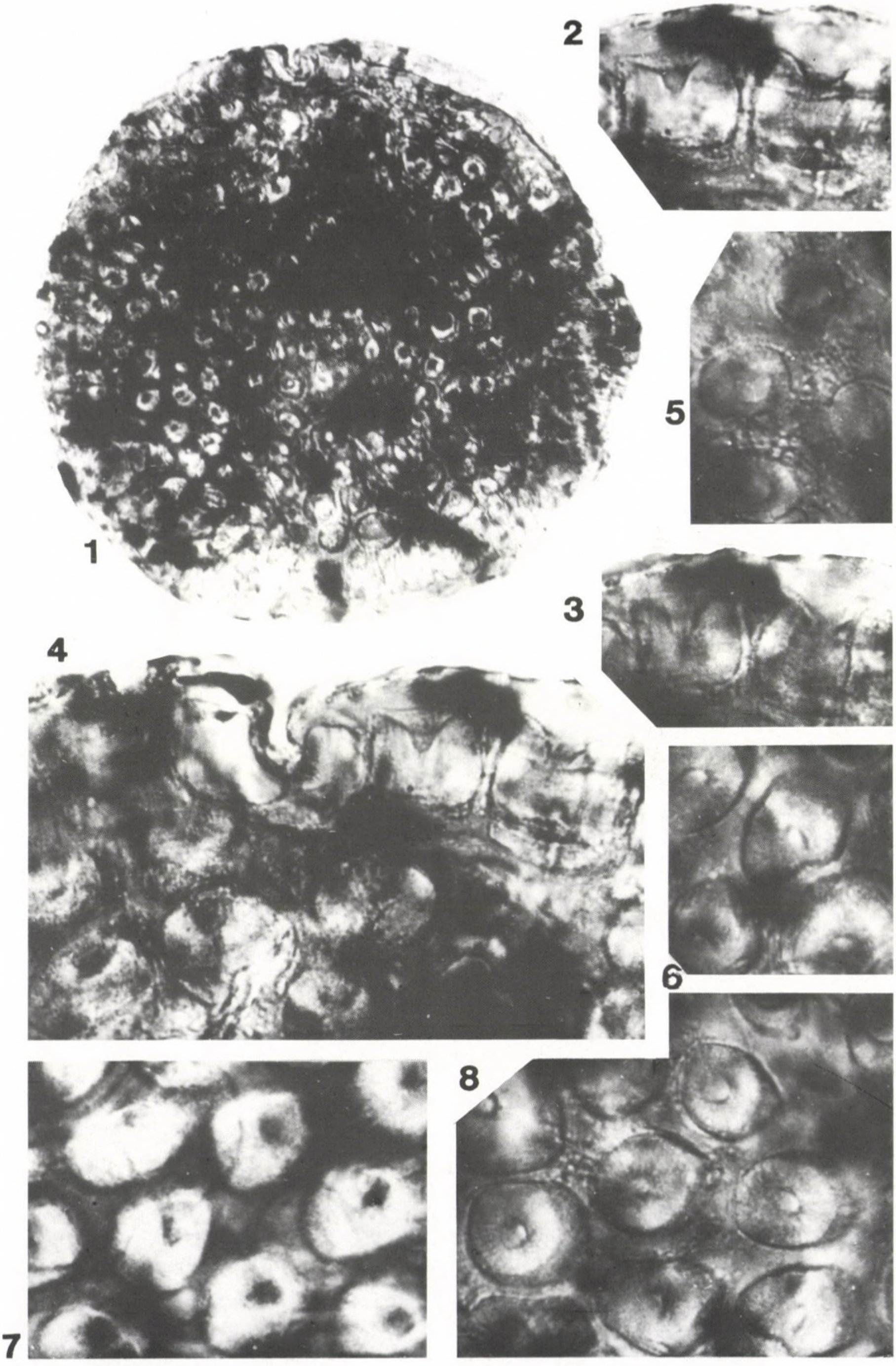

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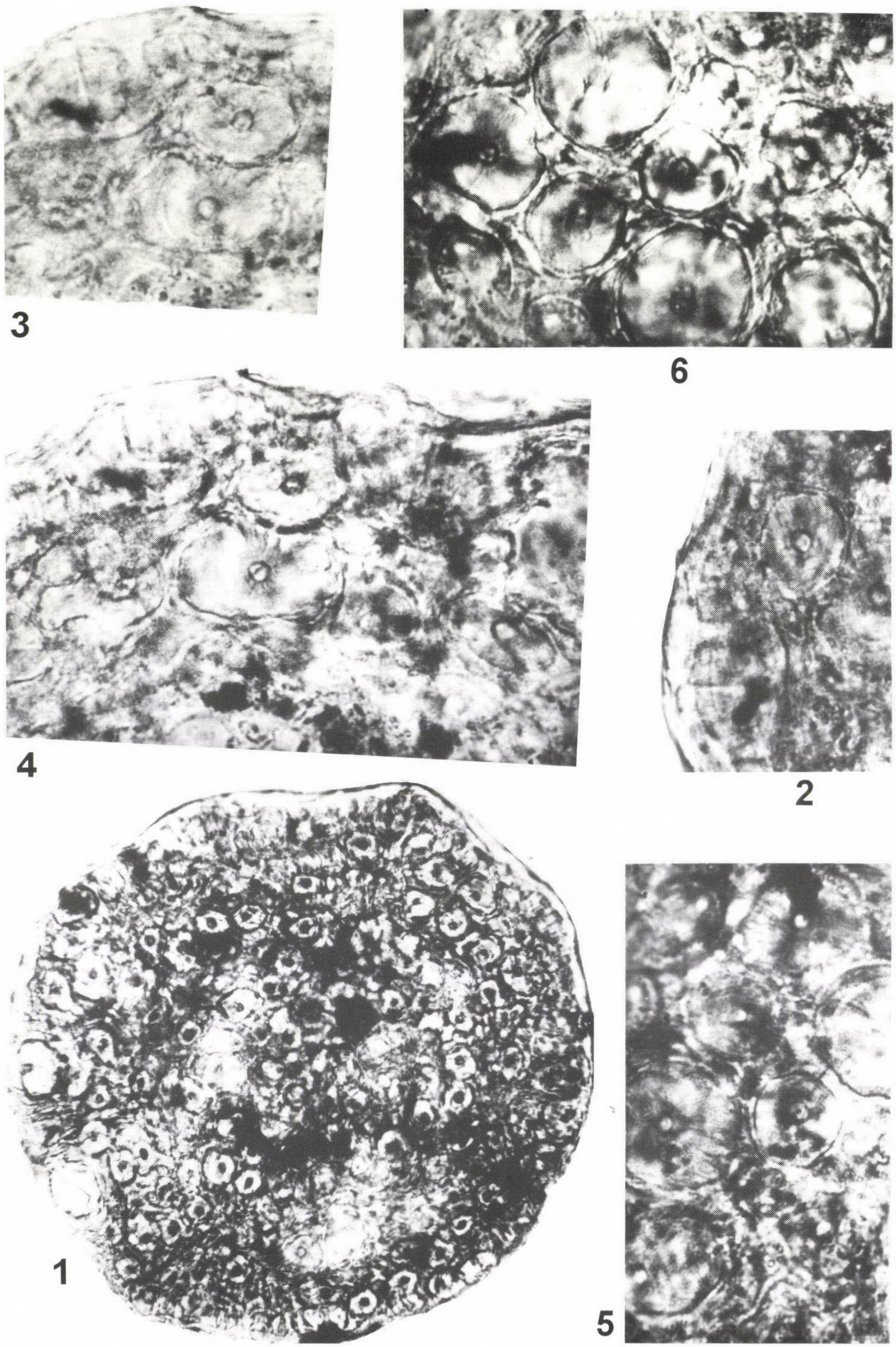

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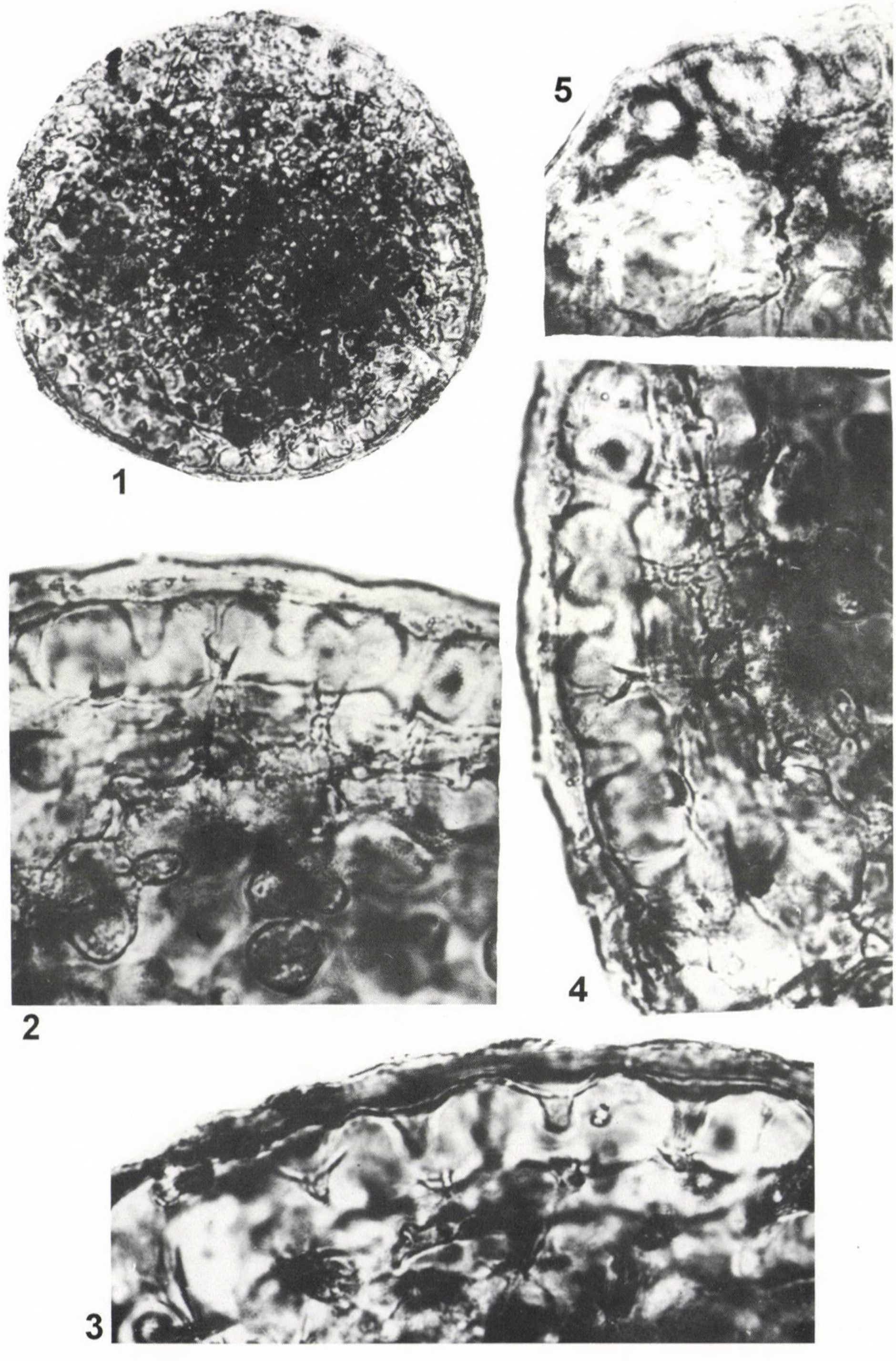

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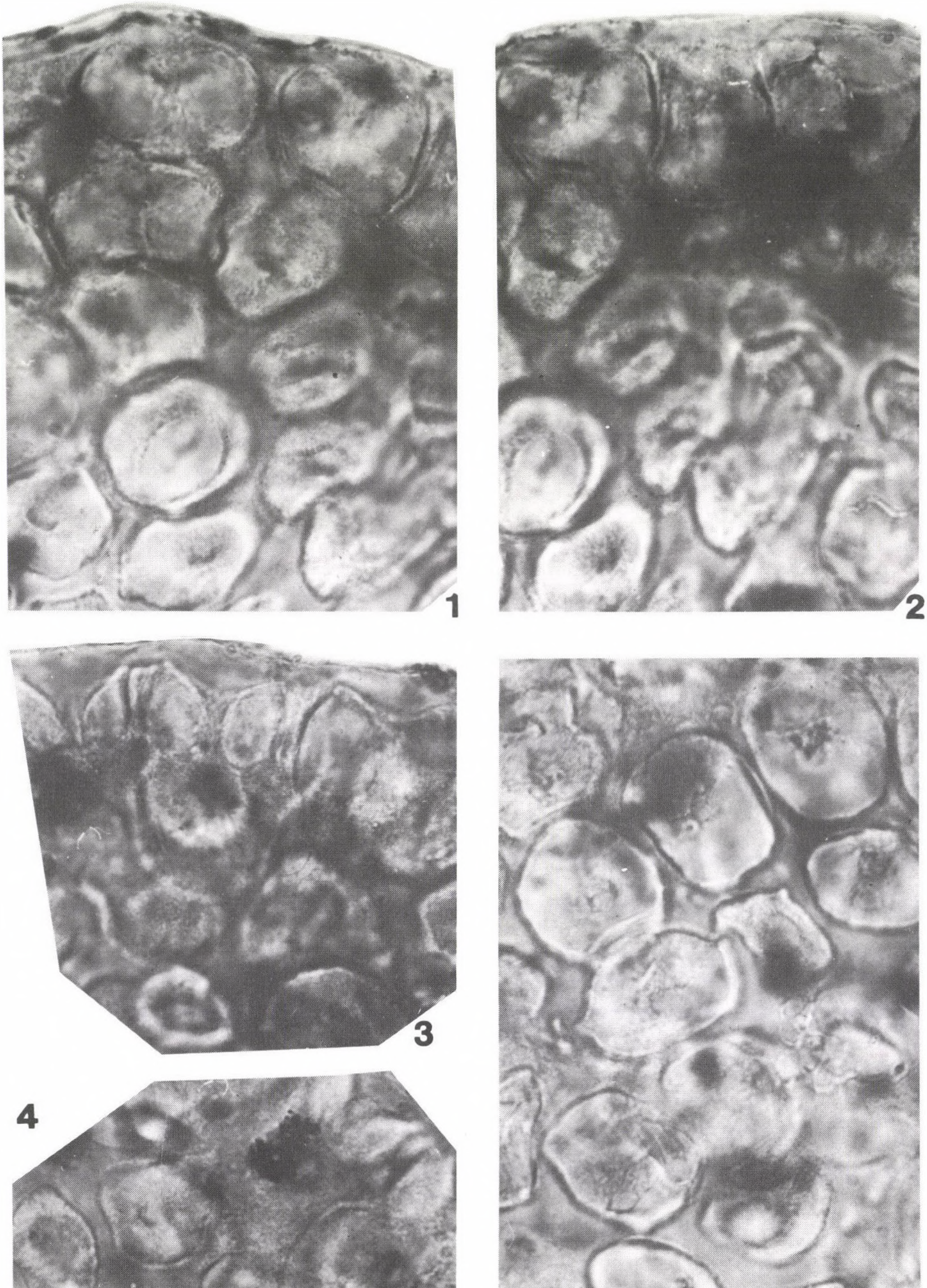

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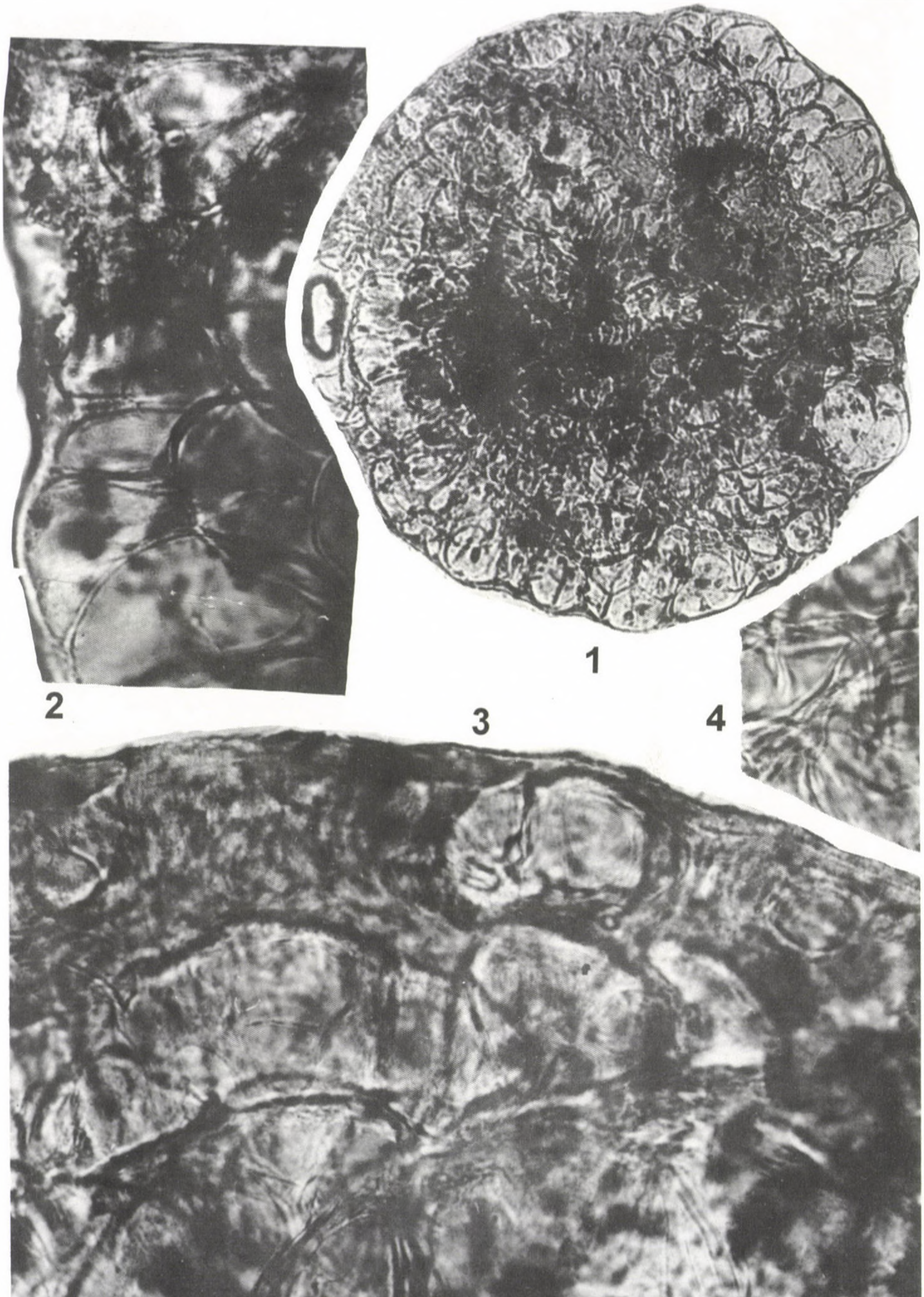

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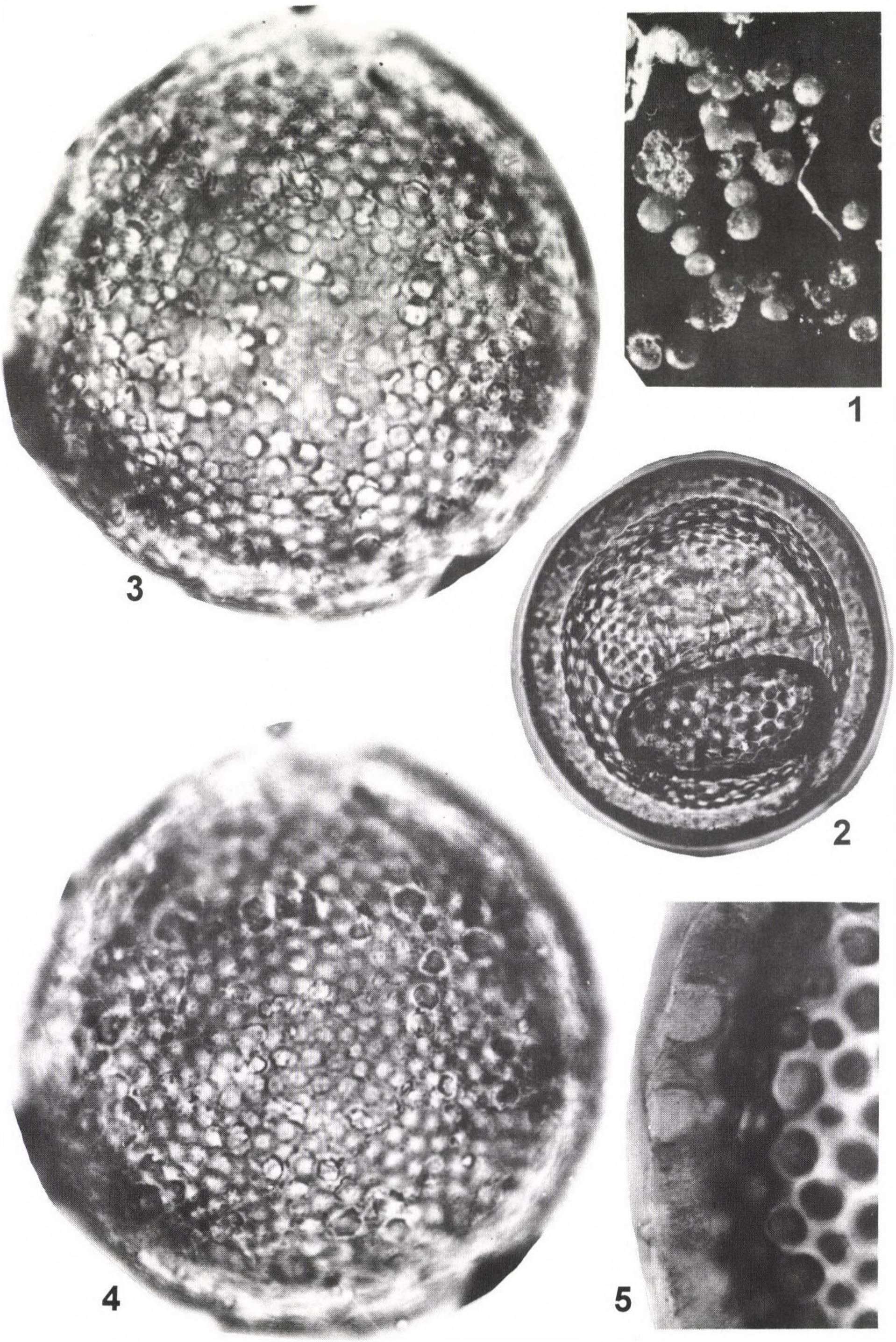

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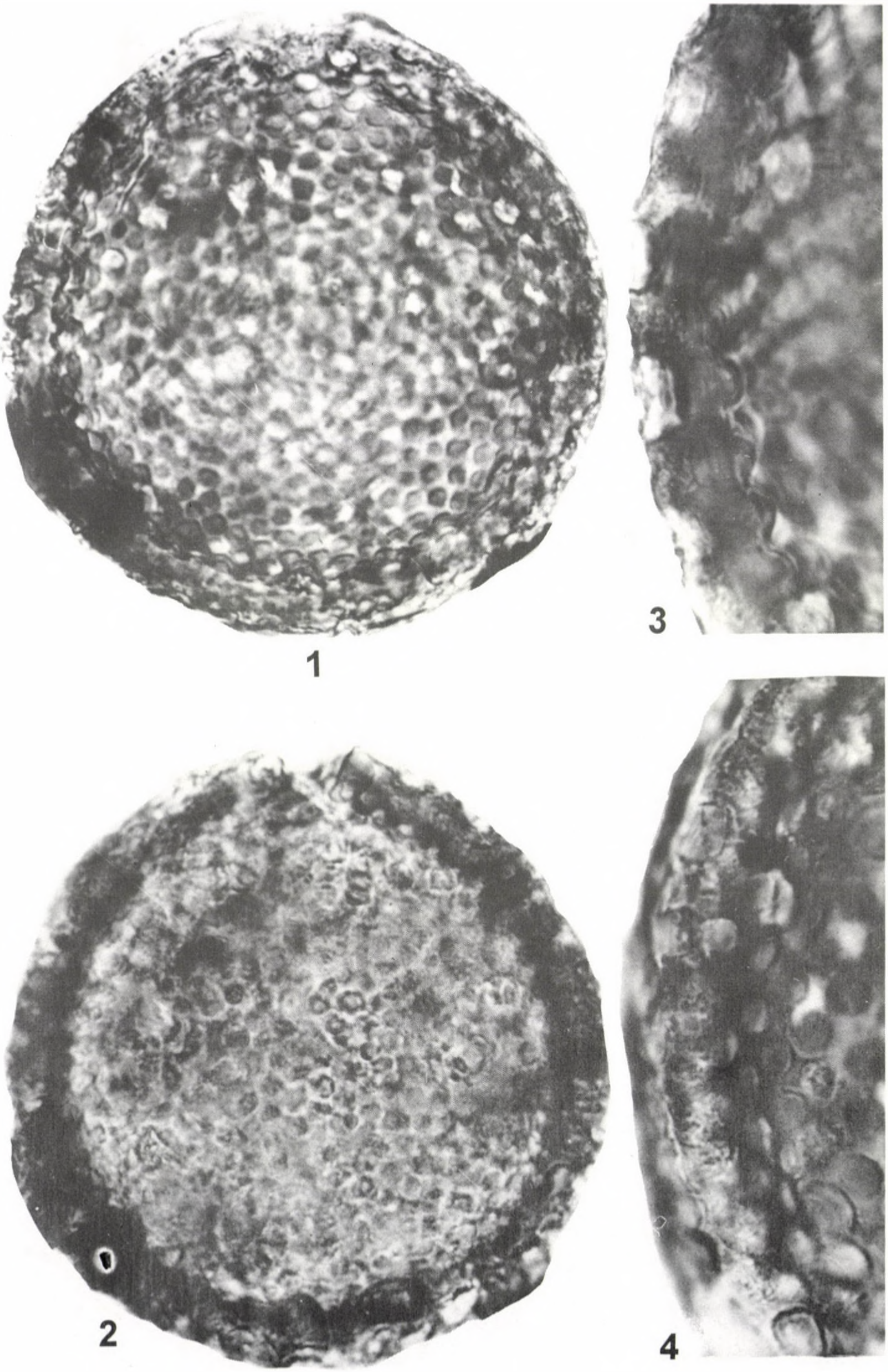

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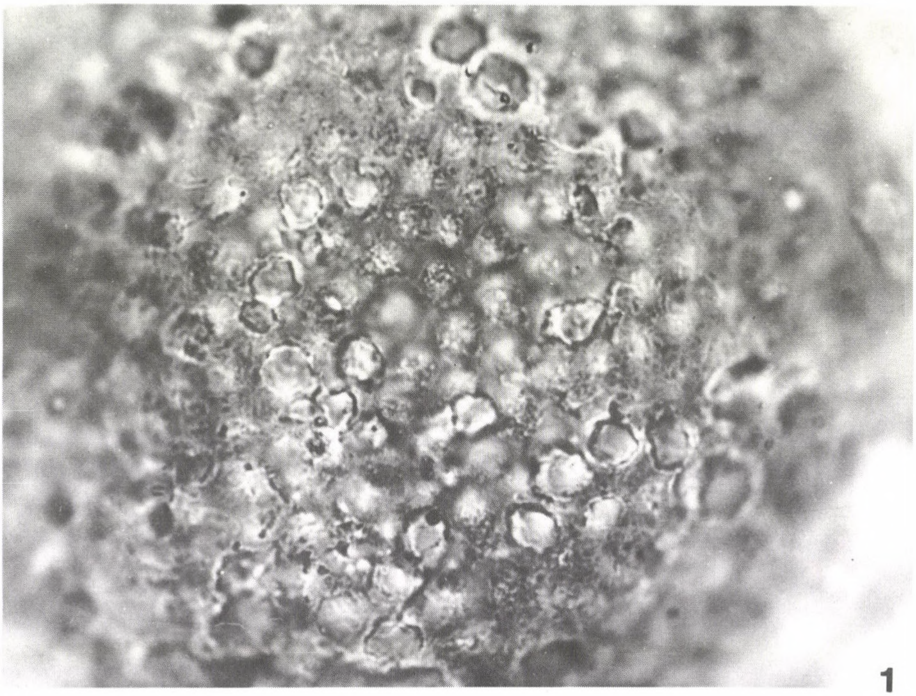

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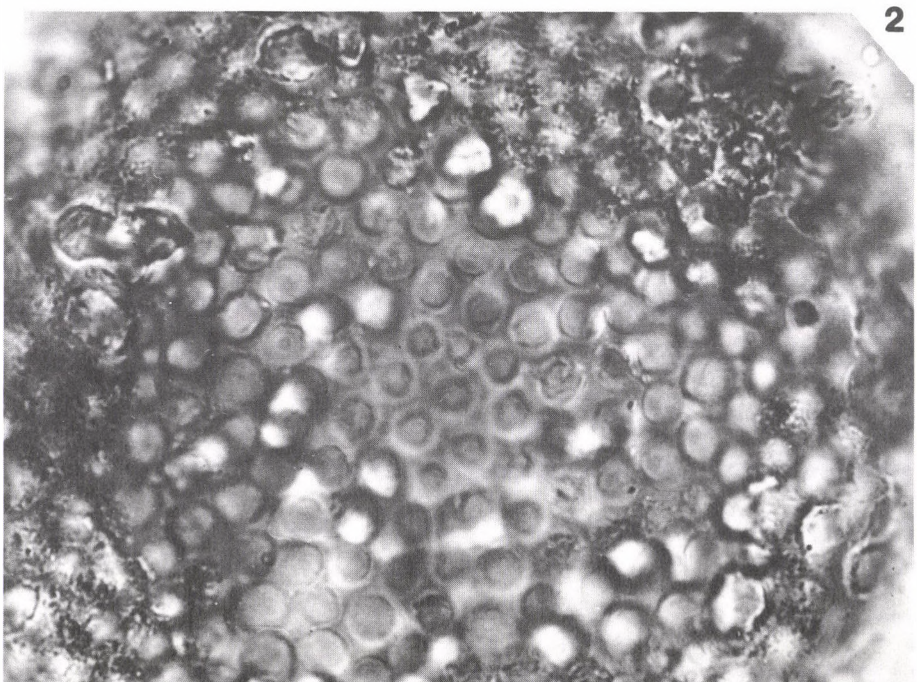

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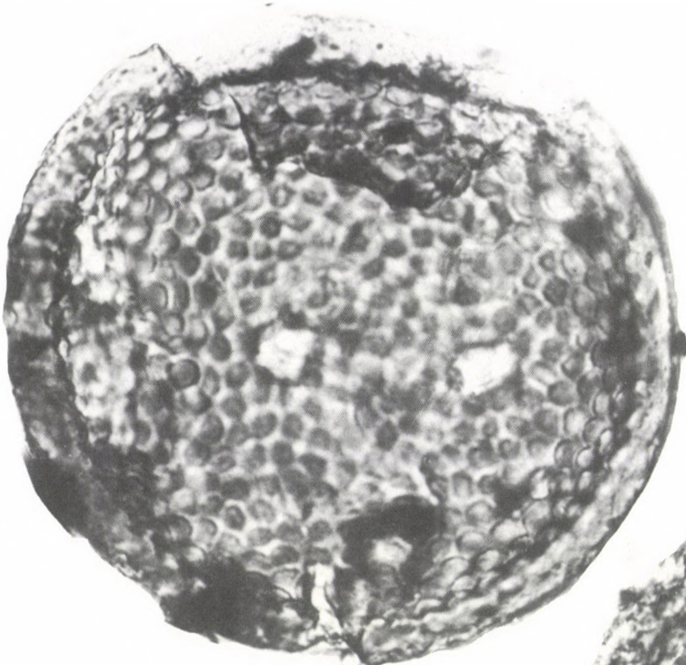

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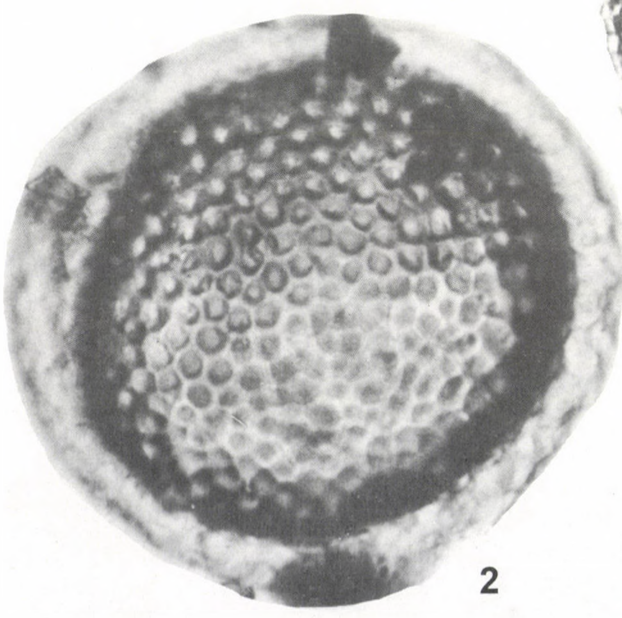

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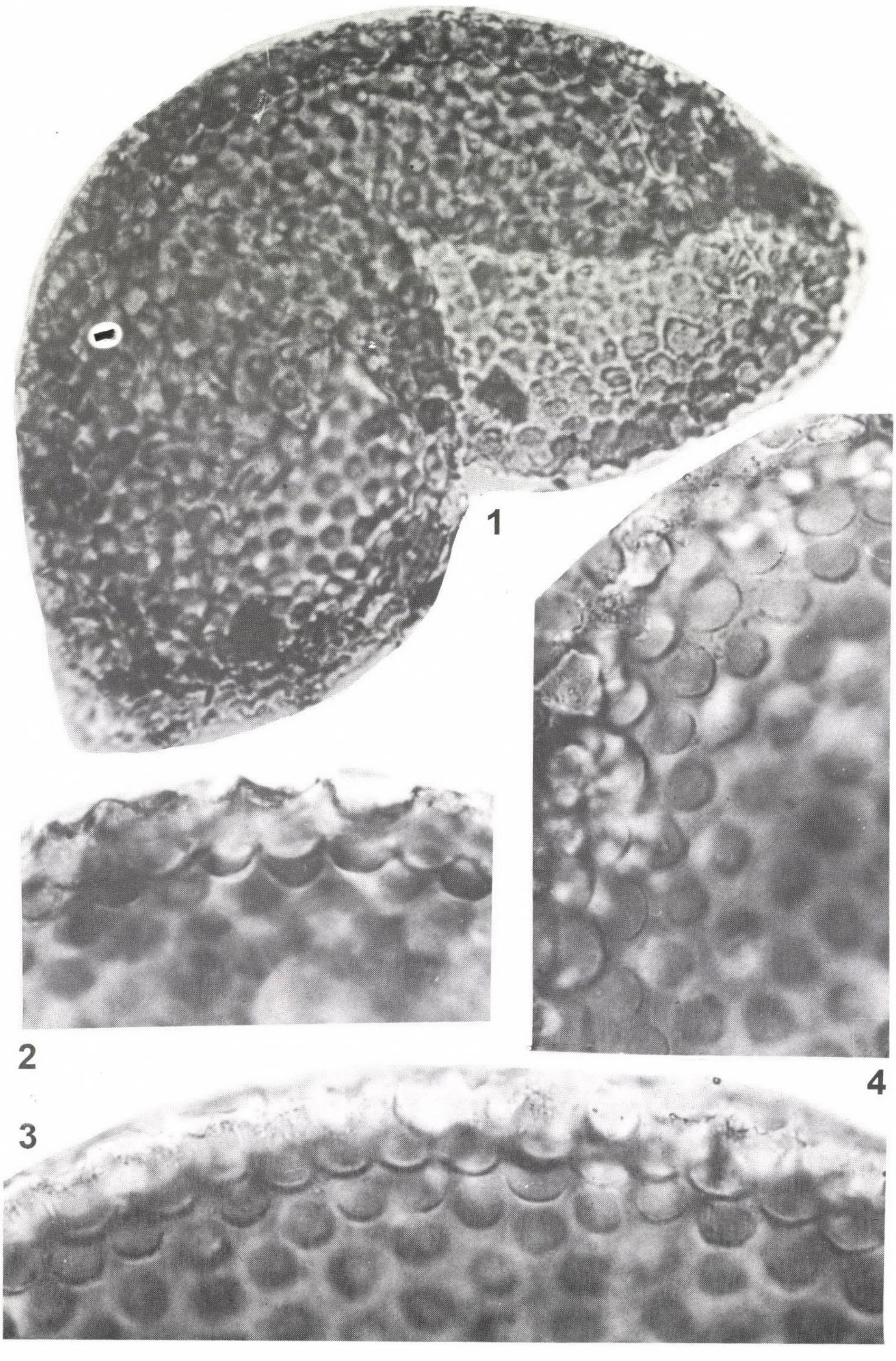

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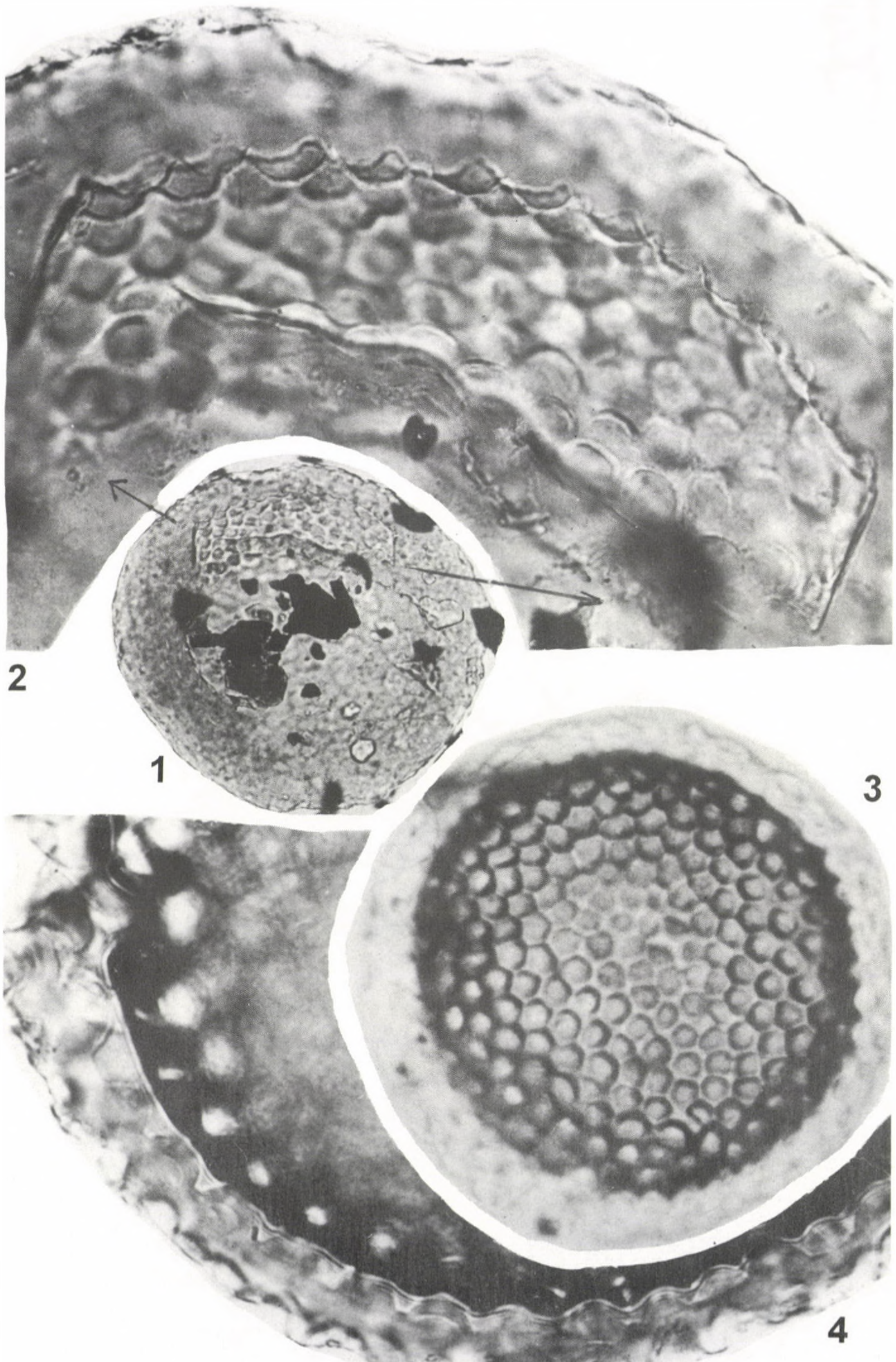

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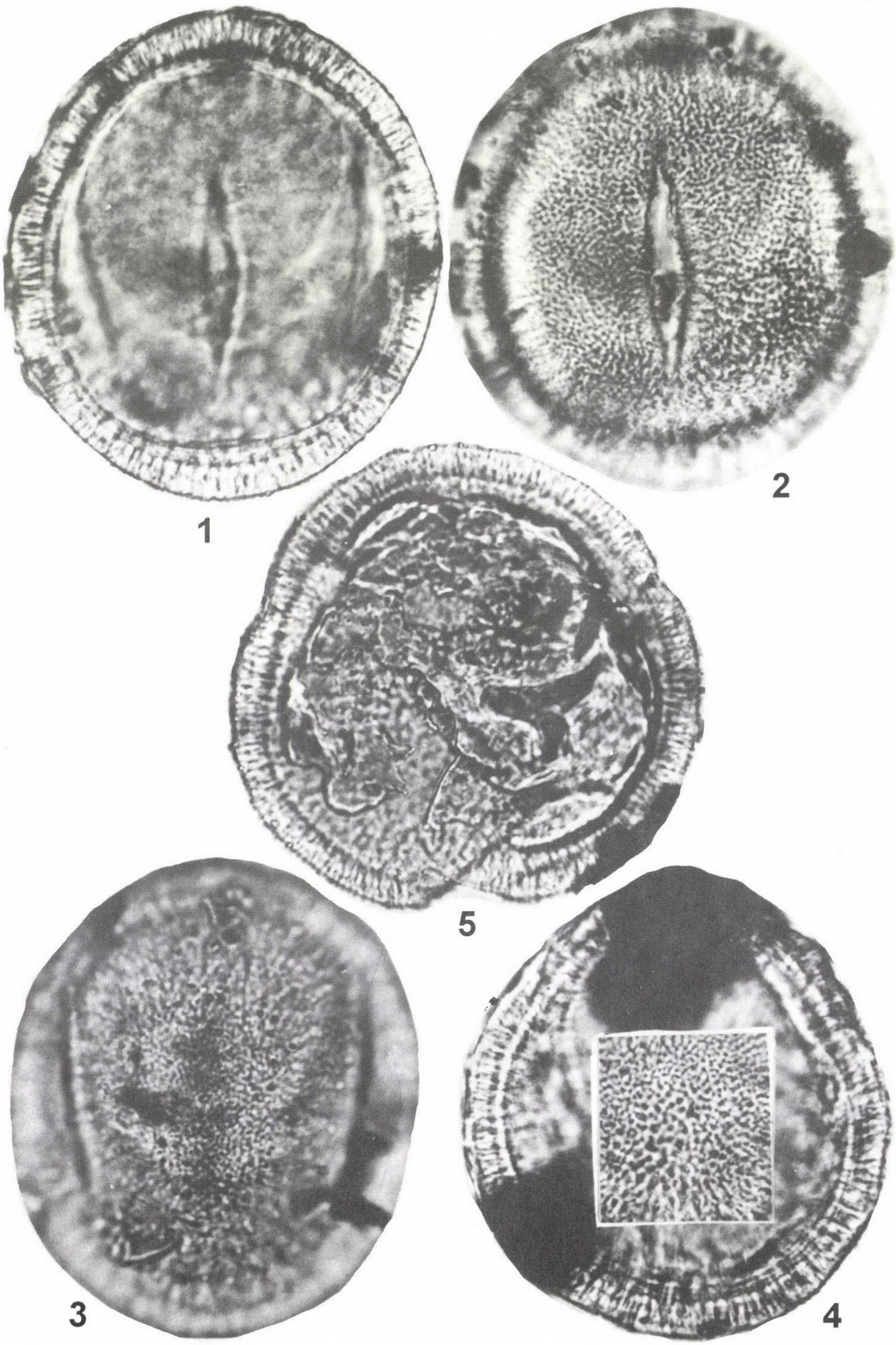

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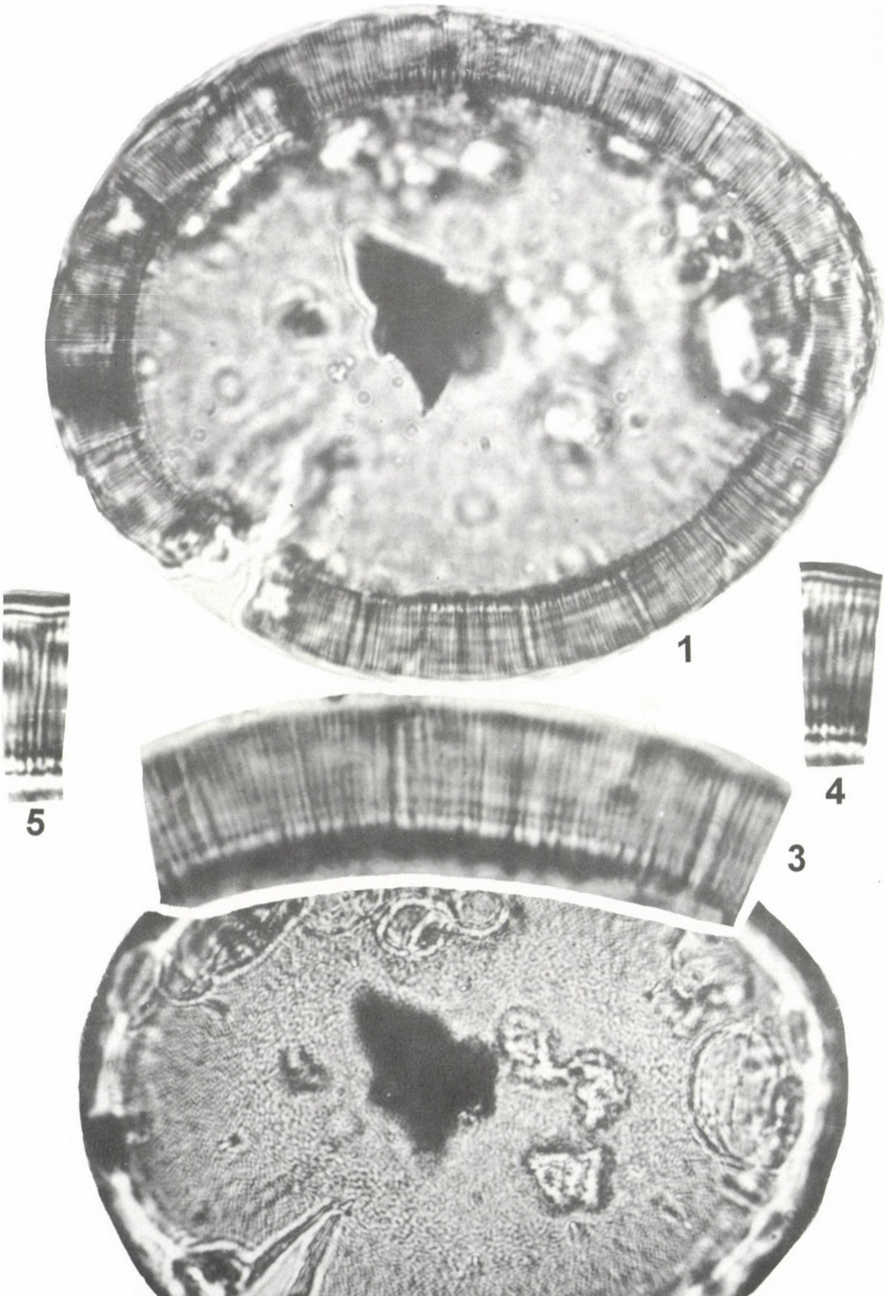

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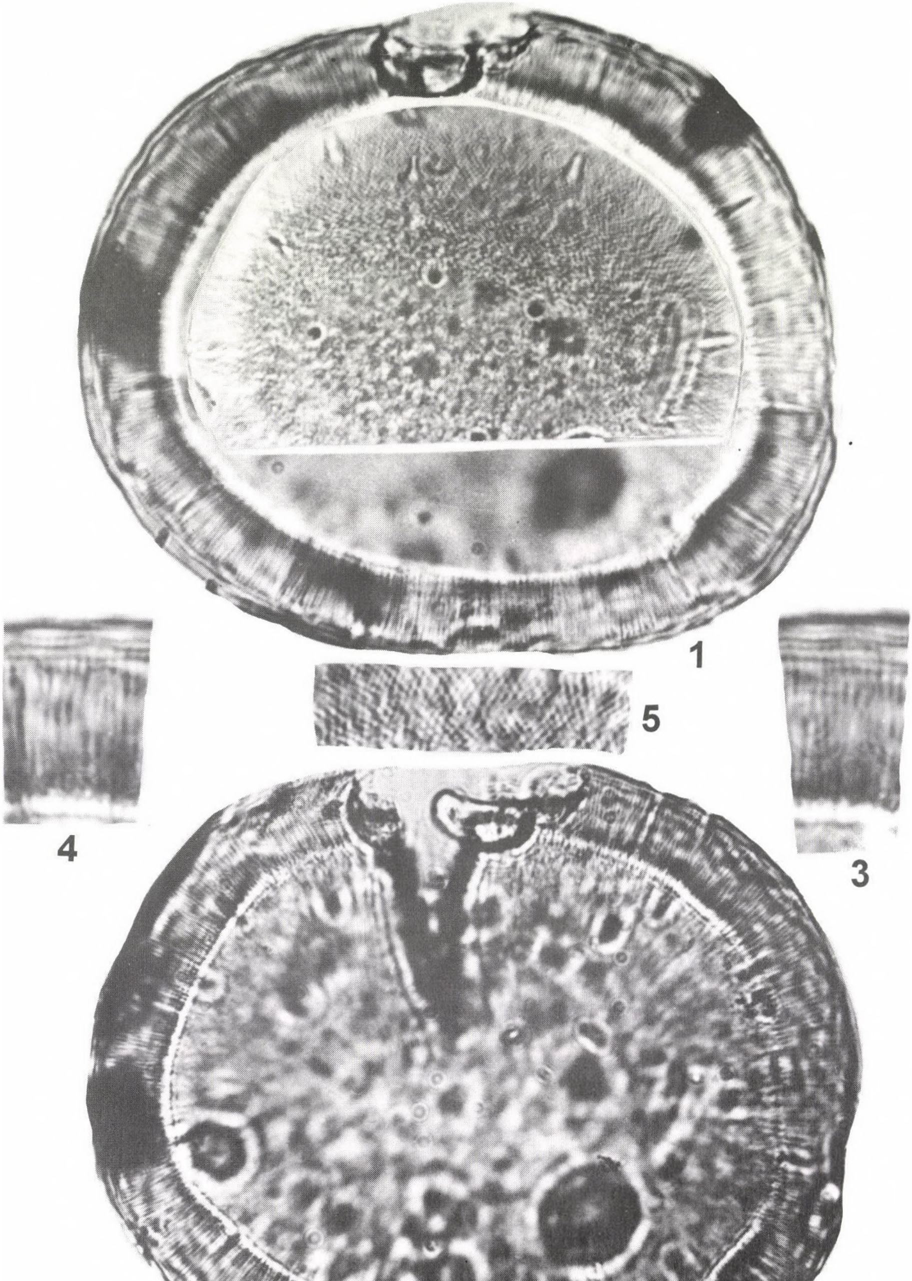

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The Miocene collapse of the Alpine–Carpathian–Pannonian junction – an overview

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The Miocene collapse of the Alpine–Carpathian–Pannonian junction is closely related to the extrusion of the Western Carpathian and Transdanubian Central Range lithospheric fragments from the East Alpine collision zone and to the Middle Miocene back-arc extension associated with the formation of the Pannonian Basin System.

The oblique collision of the Carpathians with the Bohemian Massif led to the evolution of dextral and later sinistral shear zones in the Alpine–Carpathian junction. The Early Miocene dextral shears opened the wrench fault furrow-type basins at the eastern margin of the Northern Calcareous Alps. The NE-oriented sinistral wrenching event, with a principal displacement zone along the Central Carpathian western margin, opened the Vienna Basin and the Blatné depression of the Danube Basin by a thin skinned pull–apart mechanism during the Karpatian.

The Middle Miocene subsidence both of the Vienna and Danube basins was controlled by whole lithospheric extension of the synrift stage of back-arc basin formation. The Late Miocene postrift stage of evolution, most expressive in the central part of the Danube Basin, was caused by thermal subsidence.

Key words: Miocene, Pannonian Basin, tectonics

Introduction

The Vienna and Danube basins are filled up by Neogene and Quaternary deposits in a thickness up to 6–8 km (Kilényi and Šefara 1989) (Figs 1, 2).

The Vienna Basin is superimposed upon a complex of nappe piles: the Rhenodanubian Flysch Zone, the Outer Carpathian Flysch Belt, the Northern Calcareous Alps and a part of the Central Alps and Central Carpathians (Eliáš et al. 1990; Wessely 1992; Kröll and Wessely 1993). The pre-Neogene basement of the Danube Basin is made up mainly of Central Alpine and Central Carpathian units and the buried part of the Transdanubian Range (Fusán et al.

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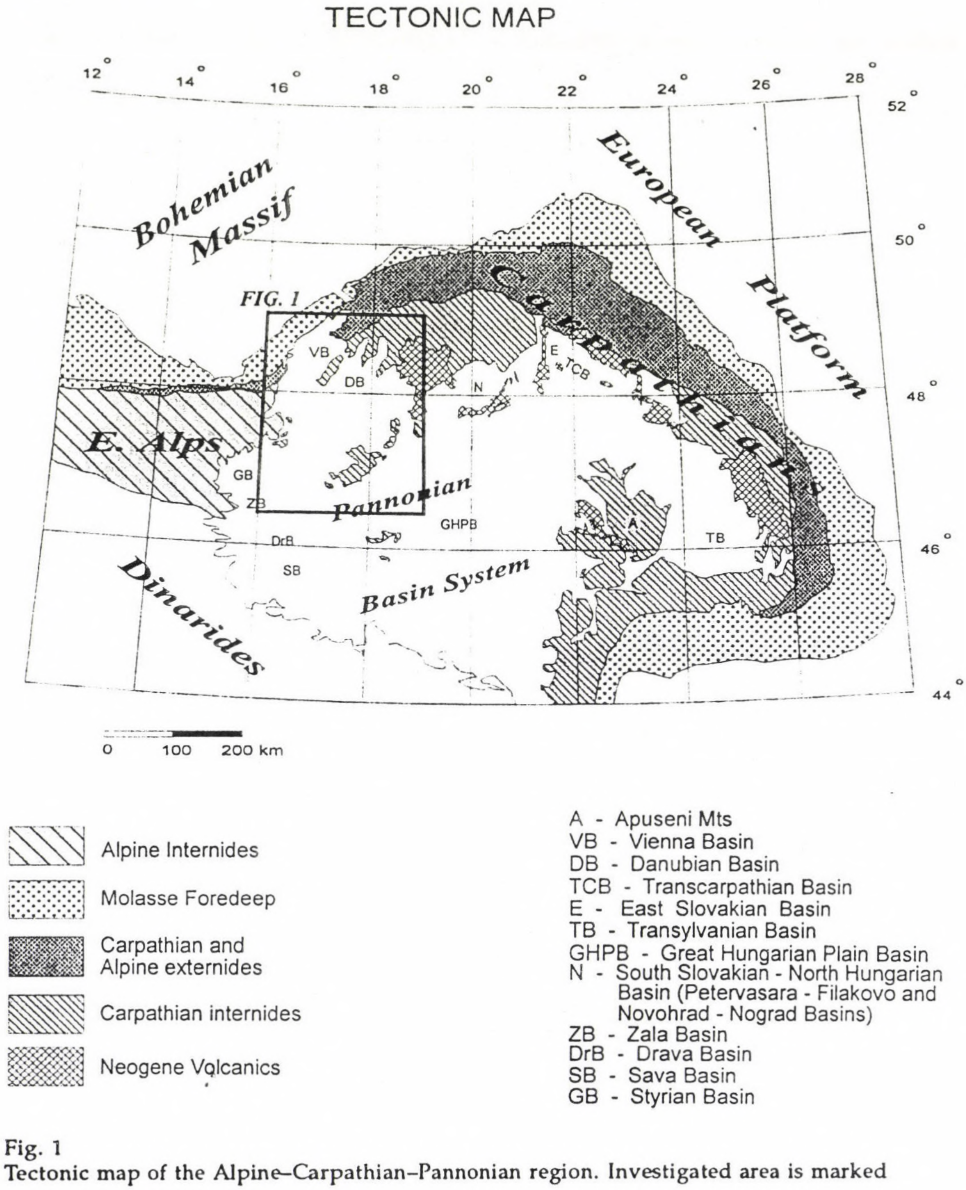

Fig. 1
Tectonic map of the Alpine–Carpathian–Pannonian region. Investigated area is marked

1987; Fülöp et al. 1987; Dank and Fülöp 1990; Balla 1995; Keith et al. 1994; Tari 1994, 1995).

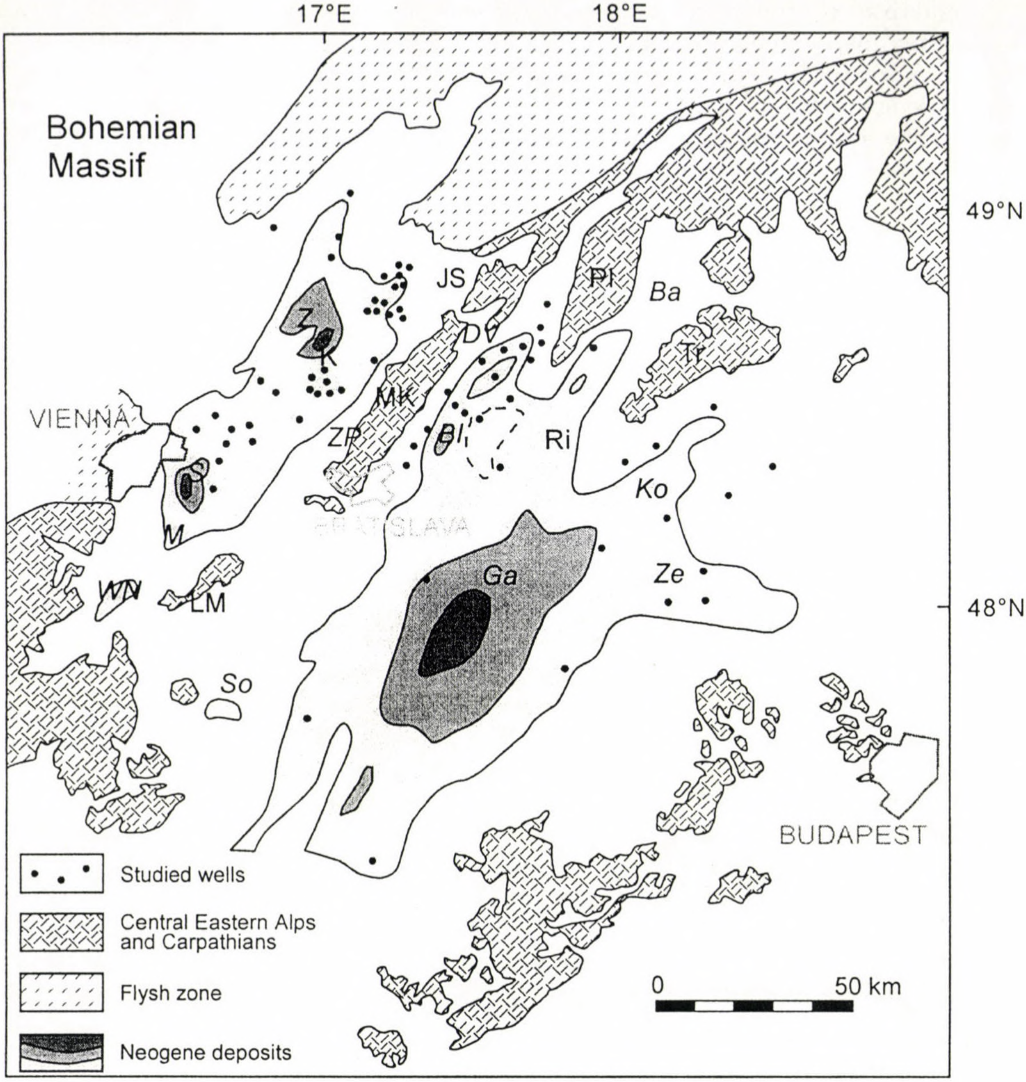

Fig. 2
Tectonic map of the Vienna and Danube basins (finger-like depressions of the northern part of the Danube basin; from west to east: Blatné (Bl), Rišnovce (Ri), Komjatice (Ko), Želiezovce (Ze) and Gabčikovo (Ga) depressions)

The Lower Miocene sedimentary fill of the Vienna, Danube and Styrian basins was deposited in close relationship to the collision between the platform and Alpine–Carpathian orogenic belt, associated with the Western Carpathian extrusion and wrench-tectonic events. The Middle Miocene mantle updoming in the Pannonian back-arc domain led to synrift and post-rift subsidence and collapse of the overthickened orogenic crust in the Alpine–Carpathian–Pannonian junction (Horváth 1993). The Neogene basin fill records the tectonic and depositional events, which led to the orogenic collapse and to widespread paleogeographic changes.

The sedimentological, biostratigraphical and paleoecological analysis of basin fill allows the reconstruction of the former shape and type of basins, as well as to take into consideration clastics sediment provenance and possible communication of water masses or seaways (Cicha et al. 1989; Kováč et al. 1989b; Oszczypko and Slaczka 1985, 1989).

The structural and paleostress analysis helped us to understand the changes in structural pattern during the Miocene and to define the type of crustal shortening and/or extension in the area of the Alpine–Carpathian–Pannonian junction (Kováč et al. 1989a; Nemcok et al. 1989, 1990; Neubauer and Genser 1990; Marko et al. 1990, 1991; Fodor et al. 1991, 1992; Fodor 1995; Csontos et al. 1991, 1992; Ratschbacher et al. 1991a, b; Kovác et al. 1993a, b, 1994, etc.).

Pre-Miocene

The Tertiary collision of the Eastern Alps with the European platform led to the flexural bend of the Bohemian Massif margin beneath the Alpine nappes during the Oligocene. This process triggered the development of the foredeep in the frontal part of the uplifting Alpine orogenic belt (Tollmann 1966). Similarly to the Alps, this collision was connected with the Carpathian foredeep development at a later time. The collision of the Central Western Carpathians with the European platform might have been oblique and shifted from the west to the east during the Miocene (Jiříček 1979; Vass et al. 1988a; Kováč et al. 1993a).

The compressional regime of the Alpine overthrust accelerated the tectonic extrusion of the eastward situated lithospheric fragments of the Central Carpathians and Transdanubian Central Range – representing a part of the ALCAPA unit (East Alpine – Central Western Carpathian – North Pannonian lithospheric segment – Ratschbacher et al. 1991a, b; Csontos et al. 1992; Horváth 1993).

The Oligo-Miocene extrusion was controlled by two main factors: north-vergent Alpine collision in the west (Decker et al. 1993) and rollback effect of the subduction of the Outer Carpathian flysch basin floor in the north and east (Csontos et al. 1992; Horváth 1993; Kováč et al. 1993a).

The Oligocene/Lower Miocene sediments of the northern margin of the foredeep are preserved on the slopes of the Bohemian Massif (Seifert 1992). The internal zone of the foredeep has been folded and thrust in the front of the Alpine flysch nappes, forming the zone of the Subalpine molasse, the Pouzdřany and the Waschberg–Ždánice units (Tollmann 1966; Krhovsky et al. 1995). The southernmost occurrence of autochthonous Egerian Alpine molasse is known from below the Northern Calcareous Alps and Rhenodanubian Flysch nappes in the borehole Berndorf–1 near the south-western margin of the present-day Vienna Basin (Wachtel and Wessely 1981).

Eggenburgian (Fig. 3)

Further compression led to thrust and wrench tectonics in the Alpine – Carpathian flysch accretionary prism, which controlled the formation of some piggy–back basins during the Early Miocene. These basins represented the embryonic Vienna Basin depocentres, situated upon the thrusted Flysch Belt. The Eggenburgian piggy–back basin deposits beyond the Central Western Carpathians are preserved in the western and northern part of the present Vienna Basin (Fig. 3). In the NE part of the basin the marginal facies is represented by the coarse Chropov conglomerates containing exclusively flysch sandstone pebble material, which was transported longitudinally from the littoral into the basinal zone of the depression. The uplifted parts of the hinterland also supplied calcareous clastic material from the south into the Winterberg conglomerates (Baráth and Kováč 1989). In the northern part of the basin the littoral Lužice sands occur. At the south-western margin of the basin variegated sandy-clayey deltaic sediments were formed. The most distal psammitic–pelitic facies is represented by Bathysiphon–Cyclammina schlier in the basin axis (Špička 1969; Jiřiček and Tomek 1981).

The oblique collision along the Outer Carpathians also led to crustal shortening at the Central Western Carpathian margin during this time. The shortening was associated with reverse fault activity and the evolution of a WSW–ENE trending dextral wrench zone (Mahel 1985; Salaj 1995; Marko et al. 1990, etc.). The dextral shears opened wrench fault furrow-type basins in the eastern margin of the Northern Calcareous Alps and in the marginal part of the Central Western Carpathians (Fig. 3) (Kováč et al. 1989a; Marko et al. 1991).

The mostly littoral Eggenburgian gravels rimming the relatively shallow sea in the NE part of the present Vienna Basin derived mainly from the Mesozoic rocks of the Northern Calcareous Alps and Central Carpathians. The orientation of island chains and intrabasinal depressions followed the principal structural boundaries of the Northern Calcareous Alpine nappes (Baráth 1993).

The open-marine basin fill is represented by pelagic marly facies in the middle part of the Váh river valley (Salaj and Zlínska 1991).

The Tatric crystalline complexes of the Alpine–Carpathian junction were still buried, as it is suggested by fission-track data, heavy mineral and pebble associations from the Eggenburgian deposits in the Central Western Carpathians (Kováč et al. 1994; Uher and Kováč 1993; Baráth and Kováč 1989).

The opening of the Turiec Basin in the central part of the orogenic belt reflects the transtensional regime of the N–S sinistral shear zone – the Central Slovakian Fault System (Kováč and Hók 1993). The Turiec graben was connected southwards with the Horná Nitra Basin, where mainly NW–SE normal faults controlled the subsidence.

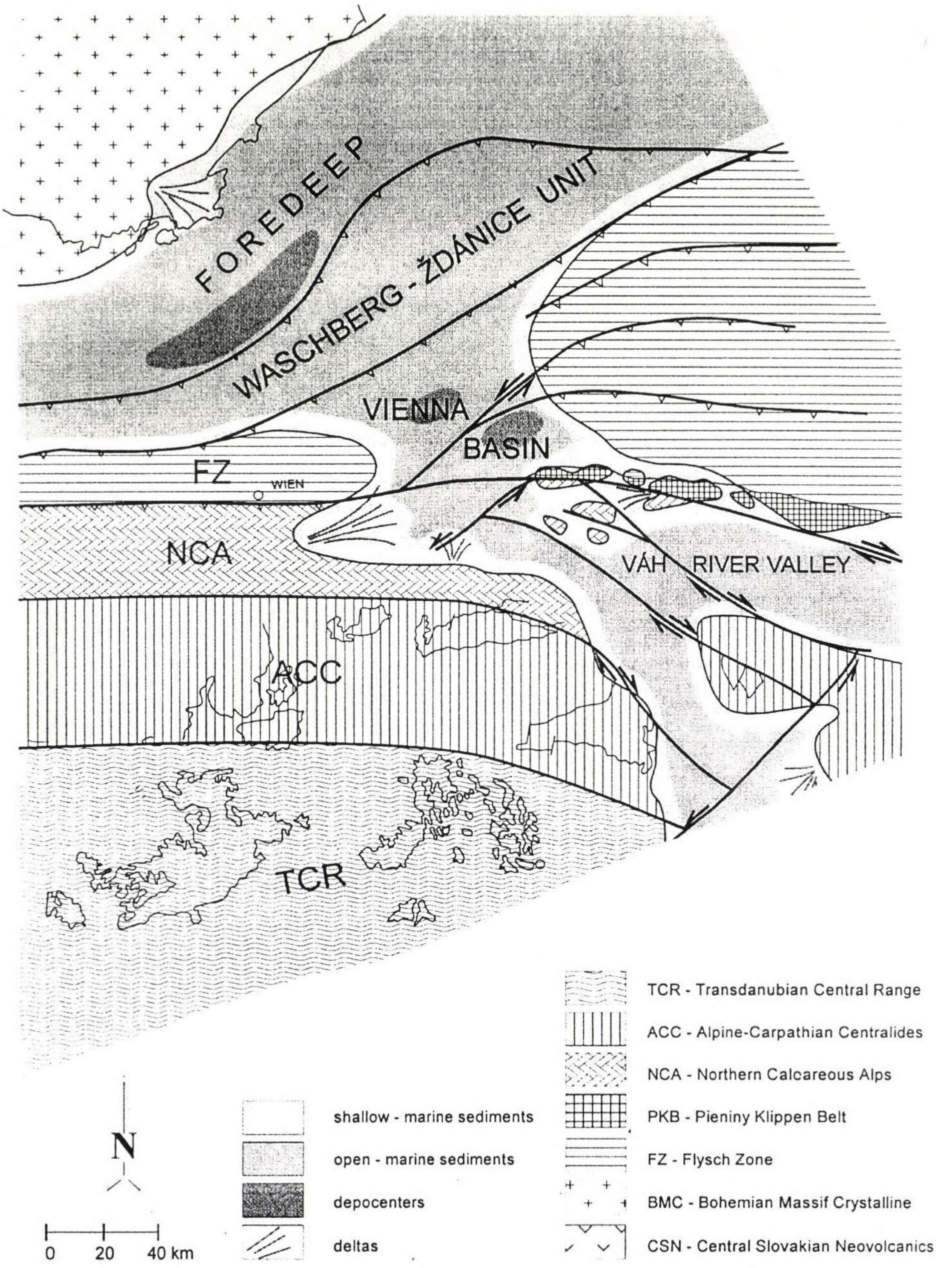

Fig. 3
The Eggenburgian–Ottnangian basin system in the Alpine–Carpathian–Pannonian junction. The positions of Vienna, Budapest, Bratislava and Graz on the geodynamic evolutionary sketch are shown in palinspastic locations

The southern boundary of the Eggenburgian sea in the central Western Carpathians is marked by marginal calcareous clastics in the Bánovce depression and in the Turiec Basin, as well as by deltaic gravels and brackish sandy sediments in the Horná Nitra Basin (Brestenská 1980; Gašparik 1969 1989; Seneš 1959).

Ottnangian (Fig. 3)

The compression in front of the Alps caused the disintegration of seaways during the Ottnangian (Rögl and Steininger 1983). The gradual uplift of the orogenic belt was followed by accelerated subsidence in the Alpine–Carpathian foredeep junction (Jiříček and Seifert 1990). The piggy-back basin upon the Rhenodanubian–Outer Carpathian Flysch Belt (the later Vienna Basin) widened. In addition to the offshore Cibicides–Elphidium schlier the littoral Lednice and Hodonín sands were also deposited (Jiříček and Tomek 1981). In the southern part of the basin the large northeastward extending deltaic–brackish complex of the Bockfliess Member developed. At the northern margin of the Malé Karpaty Mts. sedimentation continued in isolated basins, as is documented by the anoxic sandy-clayey Planinka Formation (Kováč et al. 1992).

The uplift of the Eastern Alps and the extrusion of the ALCAPA lithospheric fragment from the Alpine domain led to the opening of the Styrian Basin during this time (Fig. 1). In the Styrian Basin the east-vergent alluvial Radl, Lower Eibiswald and Lower Sinnersdorf gravel fans and further north, in the Landsee Gulf, the coal-bearing Brennberg Formation were deposited (Kollmann 1965; Ebner and Sachsenhofer 1991; Tollmann 1985; Fülöp 1983). The remnants of marine Ottnangian sediments on the eastern flank of the Transdanubian Central Range (Kókay 1971) may suggest the opening of a new seaway from the Mediterranean, following the south-eastern flanks of the Eastern Alps.

Karpatian (Fig. 4)

The change of structural pattern and paleogeography in the Karpatian was associated with extrusion of the ALCAPA lithospheric fragment from Alpine domain (Ratschbacher et al. 1991a, b; Csontos et al. 1992; Decker et al. 1993 1994; Csontos and Horváth 1995). The N–S-oriented main compression in the Central Western Carpathians (Nemčok et al. 1989) forced the displacement and tectonic subsidence along NE–SW oriented sinistral strike-slip faults, which opened the present Vienna Basin and Blatné depression (Tomek et al. 1987; Hamilton et al. 1990; Kováč et al. 1993b; Fodor 1995) (Fig. 4).

The results of forward modelling suggest that the northern part of the Vienna Basin and Blatné depression was a thin skinned pull-apart structure (Royden 1988; Lankreijer et al. 1995). In contrast to that, for the central and southern part of the Vienna Basin the results of forward modelling suggest a whole lithospheric extension (Lankreijer et al. 1995). Consequently, the pull-apart

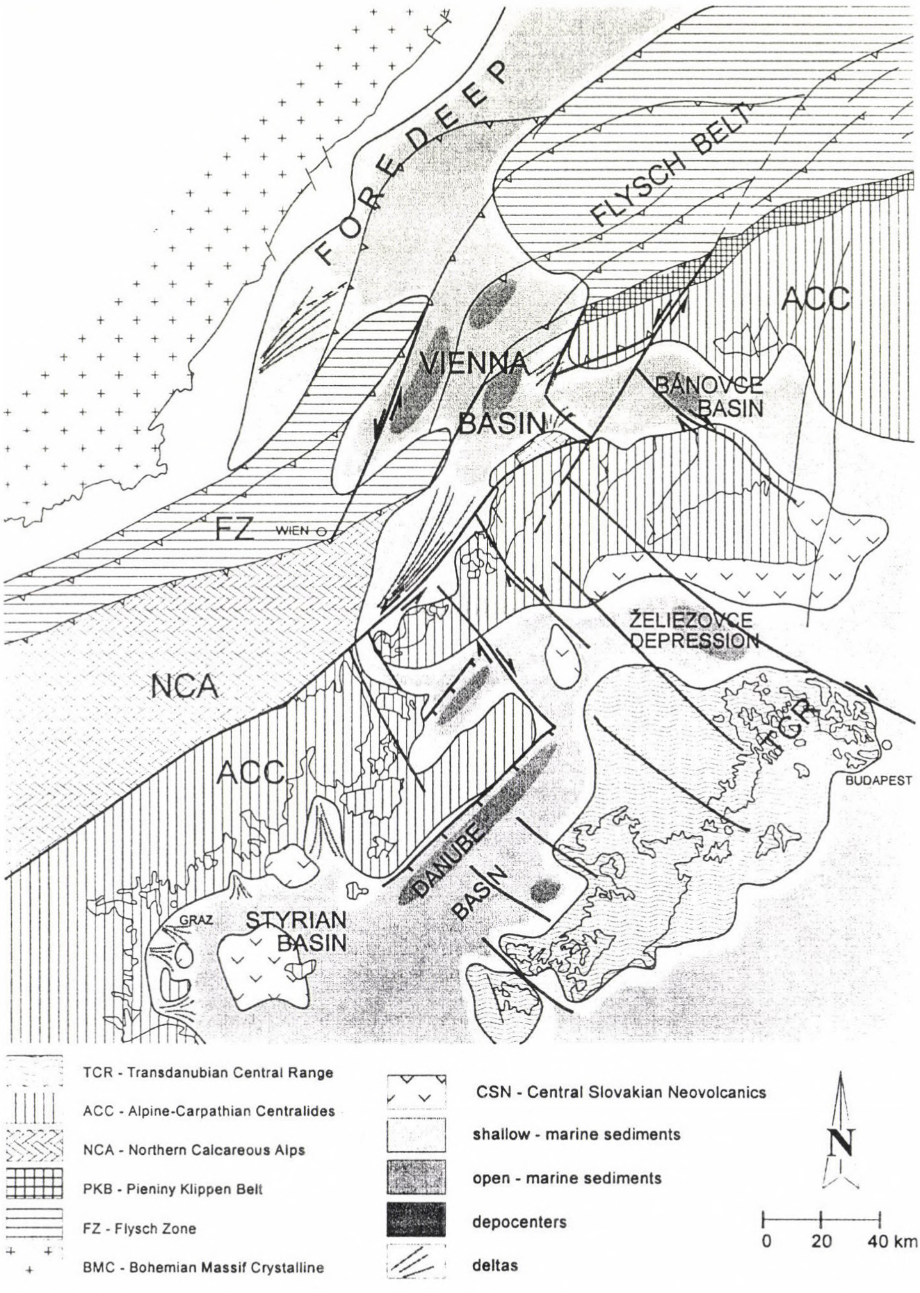

Fig. 4
The Karpatian–Lower Badenian basin system in the Alpine–Carpathian–Pannonian junction. The positions of Vienna, Budapest and Graz on the geodynamic evolutionary sketch are shown in palinspastic locations

opening was controlled here by deep-seated faults. The peri-Carpathian lineament at the eastern margin and the Schrattenberg fault on the western flanks of the Vienna Basin can be regarded as such structures (Cekan et al. 1990; Kováč et al. 1993b; Fodor 1995).

The extension of the Flysch and Upper Austroalpine nappes beneath the Vienna Basin was associated with the continuing subsidence of the marginal part of the Central Western Carpathians. The Vienna and Bánovce basins belonged to the same water circulation system, where the Karpatian is represented by shallow bathyal deposits (Brzobohaty 1987; Brestenská 1980; Kovác et al. 1993a). The statistical evaluation of the foraminiferal assemblages points to a connection toward the Pannonian back-arc basins along the transtensional Central Slovakian Fault System (Kovác et al. 1993a).

The further evolution of the Vienna Basin was strongly influenced by the changing transpressional and transtensional tectonic regimes in the Alpine–Carpathian frontal zone.

During the Karpatian, the transtensional tectonic regime widened the Vienna Basin southwards. In the northern and central part of the basin a deep marine environment prevailed with rapid sedimentation of the silty-clayey Lakšáry, Závod and Laa Formations (Špička 1969; Papp and Steininger 1979). In the south the deltaic Gänserndorf conglomerates and Aderklaa Formation pass northward into the littoral Šaštín sands (Jiriček 1990; Sauer et al. 1992). The principal source of clastic material of the Late Karpatian Aderklaa and Jablonica deltaic gravels were the Mesozoic rocks of the Northern Calcareous Alps and Central Western Carpathian nappes. At the same time, the unroofed Central Carpathian crystalline complexes also began to be eroded (Kováč 1986; Mišík 1986; Uher and Kováč 1993).

The transpression controlled the formation of the Spannberg elevation and the closure of the Lower Miocene wrench fault furrows in the northern part of the Malé Karpaty Mts. (Marko et al. 1991). Furthermore it led to the gradual uplift of Flysch nappes in front of the Central Carpathians, associated with the partial erosion of the Lower Miocene marine sequences. The overlying remnants of the Upper Karpatian–Lower Badenian braided river sediments are represented by the Čupy gravels, composed of flysch sandstone pebble material in the northern part of the Vienna Basin (Špička 1960).

As a consequence of the uplift of the Eastern Alps and initial rifting on the Danube Basin western margin coarse clastic fans appeared in the Hungarian Sopron area, in the Austrian Landsee Gulf and Styrian Basin margins. They are represented by the alluvial Ligeterdő or Auwald gravels in the Landsee–Sopron area (Vendel 1930, 1933; Tollmann 1985; Fülöp 1983; Pascher 1991), while in the Styrian Basin the deltaic and coal-bearing limnic Upper Eibiswald Member as well as the fluviatile Sinnersdorf conglomerates were deposited (Kollmann 1965; Ebner and Sachsenhofer 1991). Structural reorganization of the basin is documented by the "Styrian unconformity" at the Karpatian/Badenian boundary (Stille 1924; Friebe 1991). The crustal extension in the Styrian Basin was followed by marine transgression and acidic to intermediate volcanic activity (Ebner and Sachsenhofer 1991). The back-arc extension in the Danube Basin area was accompanied by CCW rotation of the Western Carpathians (Márton 1987; Túnyi and Kováč 1991).

Early Badenian (Fig. 4)

The Early Badenian paleogeographic situation in the Alpine–Carpathian junction reflects the Karpatian structural changes.

In the easternmost Alpine foredeep the deltaic Hollenburg–Karlstetten conglomerates were deposited, passing eastward into the sandy deposits of the Vienna Basin (Jiríšek and Seifert 1990; Jiřiček 1990). The Alpine limestone and sandstone pebbles of the conglomerates were sourced mostly from the south and south-west (Tollmann 1985). In the western part of the Carpathian foredeep the marginal Nitkovice and Holešovice conglomerates occur. Their clastic material was derived from the eroded fronts of the outer Western Carpathians flysch nappes (Benada and Kokolusová 1987). On the southwestern margin of the Vienna Basin the alluvial and littoral Gainfarn breccias and conglomerates pass into deltaic Vöslau conglomerates with mostly Alpine calcareous clastic material (Brix et al. 1988).

The steepening of the relief, connected with the uplift of the Central Alpine and Central Carpathian units along the western part of the Pericarpathian lineament (Leitha wrench zone), was associated with the development of huge talus cones and debris aprons of the Devínska Nová Ves conglomerates and breccias on the eastern flanks of the Vienna Basin. They consist mostly of crystalline clastic material (Vass et al. 1988; Kováč et al. 1991). The Leitha and Malé Karpaty Mts. became a part of uplifted horst structure in the East Alpine–West Carpathian transition zone, partly disintegrated by NW–SE dextral strike-slips and N–S normal faults (Marko and Uher 1992).

The counter-clockwise rotation of the Western Carpathians and Transdanubian Range led to extension along the south-eastern flanks of the Eastern Alps. The Lower Badenian littoral sediments of the Styrian and Eisenstadt basins belong to a former shallow sea embayment with an algal and coral reef system (Kollmann 1965; Tollmann 1955).

In the Danube Basin, the activation of low-angle listric faults (rejuvenated former compressional thrust surfaces) led to synrift subsidence (Tari et al. 1992; Tari 1994; Horváth 1993). One of the main zones of extension was the Rába lineament. Along the Rába lineament the entire Transdanubian Central Range Mesozoic nappe system moved back from above the Lower and Middle Austroalpine nappes, thus causing basin extension (Horváth 1993). The Zeliezovce depression in the eastern part of the present Danube Basin represented the deepest part of the Lower Badenian basin system (Fig. 4). Its earliest Badenian full-marine sedimentary and volcanoclastic sequence might have been deposited on a circumlittoral open-shelf plain (Seneš and

Ondrejicková 1991). The basin form as well as the rapid tectonic subsidence (Vass et al. 1993), testified by the results of forward modelling (Lankreijer et al. 1995), suggest the activity of the NW–SE-oriented dextral and normal faults.

Further south, in the Hungarian part of the Danube Basin (Little Hungarian plain) the initial point of subsidence is not well determined. While some authors suggest full marine sedimentation during the Karpatian (Hámor ed. 1988), others (Körössy 1987) described only terrestrial deposits beneath the marine Badenian. True marine foraminifera of Karpatian age were described only from the northern periphery of the Transdanubian Central Range (Metwalli 1971). Non-published micropaleontological data of Nagymarosy and Horváth, M. show that no Karpatian microfossils occur beneath the Hungarian part of the Danube Basin, and also that the oldest Badenian marine faunas and biozones are missing (lower part of NN 5, lower Lagenid zone, *Praeorbulina glomerosa* zone), i.e. the first marine beds belong to the Upper Lagenid zone, or the younger part of the Lower Badenian. This means that not only the pre-Badenian (Karpatian?) but also the earliest Badenian deposits are of non-marine origin.

The non-marine pre-Badenian can be characterized by sand beds, with coarse conglomerates indicating an anchi-metamorphic source area (Teleki et al. 1989). The occurrences are arranged into two zones: one close to the northern edge of the Transdanubian Central Range, southeast of the Rába lineament, and the other northwest of the Répce lineament.

From the late Early Badenian on, the subsidence of the southern part of the Danube Basin accelerated and the Badenian sea invaded a huge area, from the Bakony Mts. to the elevated crystalline ridge at the Sopron Mts. The Mihályi ridge formed a minor archipelago. The basin margins can be characterized either by algal limestone or by coarse-clastic sedimentation. In the intrabasinal parts siltstone and shale were deposited (Körössy 1987).

The Badenian crustal extension and subsidence in the Danube Basin was accompanied by volcanic activity (Gnojek and Heinz 1993). The belt of the alkaline and calc-alkaline intermediate volcanism extends from the Styrian Basin (Ebner and Sachsenhofer 1991) along the Rába lineament up to the Central Slovakian volcanic area. All boreholes reaching thick eruptive volcanic sequences beneath the Little Hungarian plain are situated very close to the Rába lineament system.

The extensional development of the basins, isochronous with the further northward drift of the Inner Western Carpathians, and the volcanic activity can be supposed as a consequence of the evolution of a back-arc basin.

Middle–Late Badenian (Fig. 5)

The Middle Miocene transtensional development of the Vienna and Danube basins (Fig. 5) were characterized by a paleostress field with NW–SE extension (Nemčok et al. 1989; Fodor et al. 1991; Csontos et al. 1991; Kováč et al. 1993b).

The subsidence of the central and southern parts of the Vienna Basin was controlled predominantly by the activity of NE–SW to NNE–SSW oriented normal faults (Špička 1969). The gradual basin subsidence was compensated by the deposition of sediments transported from uplifted basin margins and by great deltas of the paleo-Danube and paleo-Morava rivers which entered the Vienna Basin from the west and north during the Badenian. Some smaller deltas and alluvial fans in the southwest and northeast also brought sandy-conglomeratic material into the basin (Brix et al. 1988; Jiříček 1990; Sauer et al. 1992). On the eastern flanks, along the Malé Karpaty, Hundsheim and Leitha Mts., a transgressive sequence with gravels, sands and algal–coral reef limestones occurs (Buday et al. 1962; Fuchs 1985; Tollmann 1985; Baráth et al. 1994). The basinal facies is represented by *Spiroplectammina* and *Bulimina–Bolivina*-bearing shales (Jiříček and Cícha in Steininger et al. 1985).

The subsidence of the Danube Basin northern embayments was controlled by NE–SW oriented normal faults and WSW–ENE oriented sinistral strike-slip faults (Kováč et al. 1993b).

The Middle Badenian opening of the Blatné depression was followed by the accumulation of huge talus cones of conglomerates and breccias in the western and eastern part of the Blatné depression (Baráth 1993). The subsiding central part is filled up mainly by the pelitic Špačince Formation passing northward into the deltaic–brackish Madunice sands (Vass et al. 1990; Jiříček 1990).

The subsidence of eastward-situated grabens of the Rišnovce and Komjatice depressions was controlled by NE–SW oriented normal faults (Fig. 5). The predominantly sandy–clayey marine basin fill contains frequent tuffaceous admixture. On the northern edges of the depressions deltaic and brackish sediments are present (Vass et al. 1990). The subsidence of the central part of the basin also reflects the activity of NE–SW oriented normal fault systems (Peničková and Dvořáková 1985).

In the Little Hungarian Plain area the Middle-Late Badenian paleogeography of the Danube Basin shows the same features as in the Early Badenian. However, some trends of filling up of the basin can be observed at the end of the Badenian epoch, such as the frequent occurrence of sandstone bodies interfingering with pelagic siltstones, or the formation of two belts of algal (rarely reef) limestone patches along the margins of the basin. The volcanoclastics of trachytic alkaline volcanism in the Pásztori region also played an important role in filling up the basin (Teleki et al. 1989).

In the Styrian Basin marine sedimentation continued in a sandy–clayey facies. frequently algal, coral and biodetritic reef limestones formed on the elevations. At the western margin of the basin there is a visible reflection of salinity drop in the sandy–gravelly deposits, due to the fluviatile influence from the Alps (Tollmann 1985; Friebe 1990; Ebner and Sachsenhofer 1991).

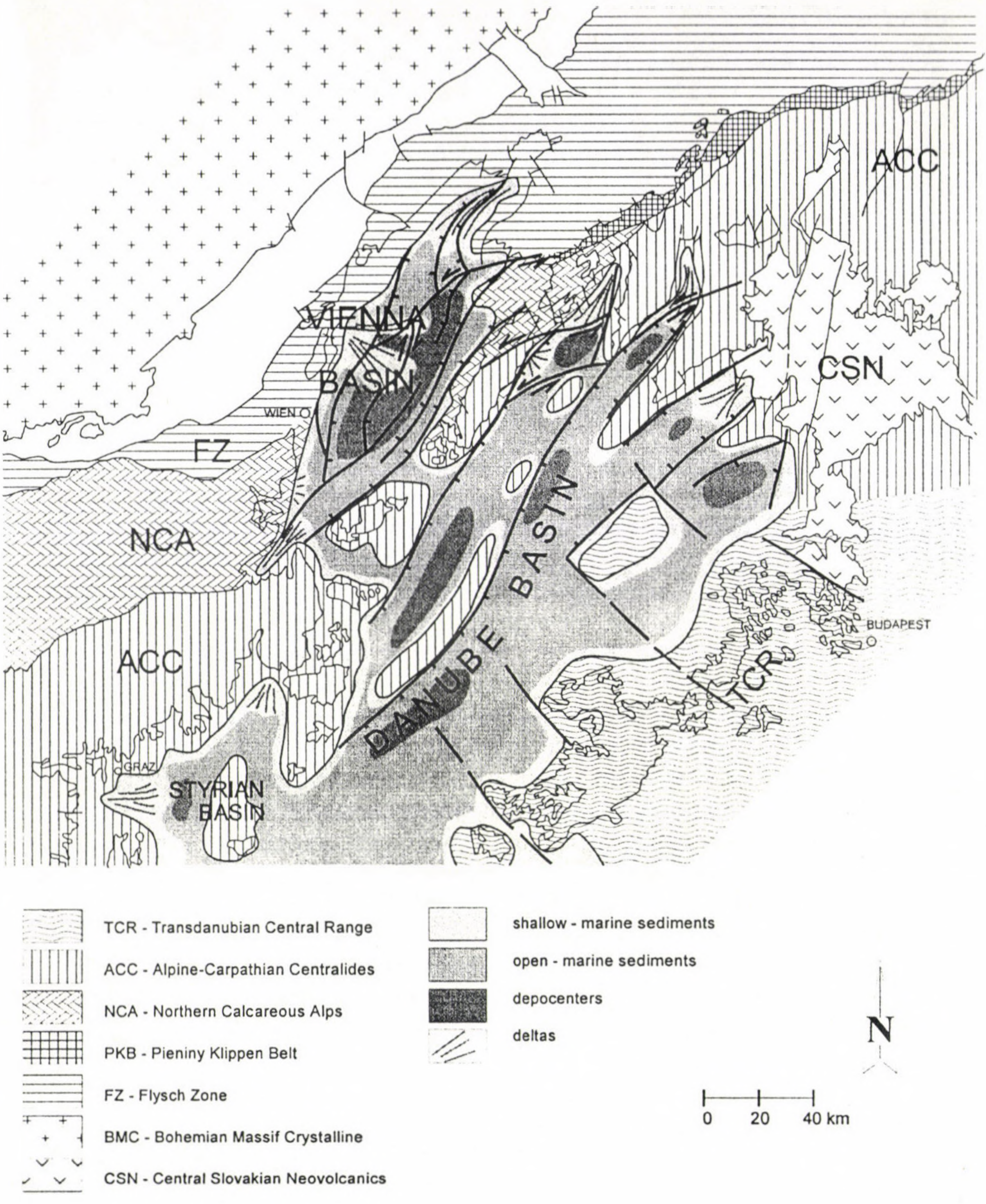

Fig. 5
The Middle Badenian–Sarmatian basin system in the Alpine–Carpathian–Pannonian junction. The positions of Vienna, Budapest and Graz on the geodynamic evolutionary sketch are shown in palinspastic locations

Sarmatian (Fig. 5)

During the Sarmatian the paleostress field with a NW–SE oriented extension activated the WSW–ENE-oriented sinistral strike slips and NE–SW to NNE–SSW normal faults, permitting the marine invasion further northward into the Koválov graben of the Vienna Basin. A similar sinistral shear zone of WSW–ENE direction has also been demonstrated at the northern margin of the Danube Basin (Rišnovce depression), extending to the Horná Nitra region (Hók et al. 1995).

In the central part of the Danube Basin, the associated N–S dextral shears controlled the sedimentation from the Late Badenian to the Pannonian (Hrušecký et al. 1993). This coincides with the N–S dextral displacement in the Klippen Belt Zone (Kováč and Hók 1993) and both can be related to the initial stage of inversion regime of the orogenic belt and back-arc region (Horváth 1995).

At the beginning of Sarmatian in the Alpine–Carpathian junction a sea level drop can be observed, accompanied by erosion of the Badenian marginal sequences (Hudáčková and Kováč 1993; Baráth 1993). Although it is not visible on seismic sections, a minor sedimentary gap must be supposed between the Badenian and Sarmatian deposits in the bulk of the area. If not postulating an uplift and subaerial erosion it is possible to suggest a slight submarine uplift and erosional effect combined with a low rate of deposition.

In the southern and partly central part of the Danube Basin (Little Hungarian Plain) the narrow belts of brackish fish-scale bearing shale and *Elphidium*- and *Nonion*-bearing schlier represent the Sarmatian sedimentation. At the basin margins sandy limestones and coarse clastics were deposited (for example a minor river delta system in Sopron Mts. (Rosta 1991).

As opposed to the central part of the Danube Basin, an accelerated synrift subsidence characterized the evolution of the subbasins situated on the northern margin of the back-arc asthenosphere updoming.

Opening of partial depocentres in the northern part of the Vienna and the Danube basins (Vass et al. 1988a, 1990) reflected the active elongation of the outer Western Carpathians due to roll-back effect of the Eastern Carpathian subduction zone (Royden 1993; Lexa et al. 1993, 1995; Csontos and Horváth 1995).

The brackish character of the Sarmatian sea was a consequence of the disintegration of the Badenian sea-ways toward the Mediterranean. The Sarmatian sea might have been shallow, reaching neritic depth as a maximum.

In the western and northern part of the Vienna Basin as well as in its central part, deltaic sequences were formed (Sauer et al. 1992). On the eastern margin of the basin, a regressive conglomerate passes upward into transgressive sandy–organodetritic limestones (Tollmann 1985; Buday et al. 1962; Nagy et al. 1993). The basinal facies is represented by sandy–marly to clayey sediments.

In the northern gulfs of the Danube Basin the development of paleodeltas continued, while in the central part of the basin sandy–clayey sediments filled up the slowly subsiding basin (Vass et al. 1990).

Sedimentation in the Horná Nitra Basin is characterized by Lower Sarmatian paralic coal accumulations and later by brackish clays (Fazekaš 1995).

In the Little Hungarian plain the Sarmatian deposits show a more restricted areal distribution than the Badenian ones. The sporadic distribution and extremely small thickness of the Sarmatian indicates beginning basin inversion and as a consequence of this, selective erosion in post-Sarmatian time (Fodor et al. 1991; Horváth 1995; Csontos 1995).

In the Styrian Basin, after a short sedimentary gap at the Badenian/Sarmatian boundary there is visible gradual transition from the Early Sarmatian marine sandy–clayey facies to the Late Sarmatian fluviatile–limnic gravelly and coal-bearing sequences (Kollmann 1965; Ebner and Sachsenhofer 1991).

Pannonian–Pontian (Fig. 6)

During the Upper Miocene rapid subsidence took place only in restricted depocenters of the Vienna and Danube basins, as is suggested by the results of forward modelling (Lankreijer et al. 1995). In the Vienna Basin a young wrenching event is assumed by the opening of the Pannonian and Pontian depocentres (Wessely 1988).

In the Gabcíkovo central depression of the Danube Basin, sedimentation was connected with the thermal post-rift subsidence of the back-arc area without any significant role of fault activity (Horváth and Royden 1981; Horváth et al. 1986; Tari et al. 1992; Horváth 1993).

In the northern part of the Vienna Basin a large south-vergent deltaic system formed with marsh lignite series during the Pannonian. The greatest delta of the paleo-Danube river entered the Vienna Basin through the Zaya graben, with the gravelly mouth in the Mistelbach area and sandy subaqueous continuation as far as into Vysoká and Zwerndorf in the central part of basin (Jiřiček 1990). In the southern part of the basin deltas of the paleo-Piesting and paleo-Triesting rivers developed coming from the Alps (Sauer et al. 1992). On the western slopes of the Leitha Mts. detritic limestone and sandy facies were deposited. The sources of clastics were mostly the eroded older Sarmatian and Badenian sediments (Tollmann 1985). The basinal facies is represented by sand and shale deposits with brackish fauna.

The sedimentation in the extensional grabens at the northern margin of the Danube Basin was influenced by a deltaic environment, forming marshes in the Blatné depression, a limnic estuary in the Rišnovce depression and a delta-influenced embayment in the Komjatice depression (Jiřiček 1990).

During the Pannonian and Pontian the lacustrine water masses invaded all previously emerged areas in the central and southern areas of the Danube Basin, including the previously uplifted Mihályi ridge as well. The occurrence of

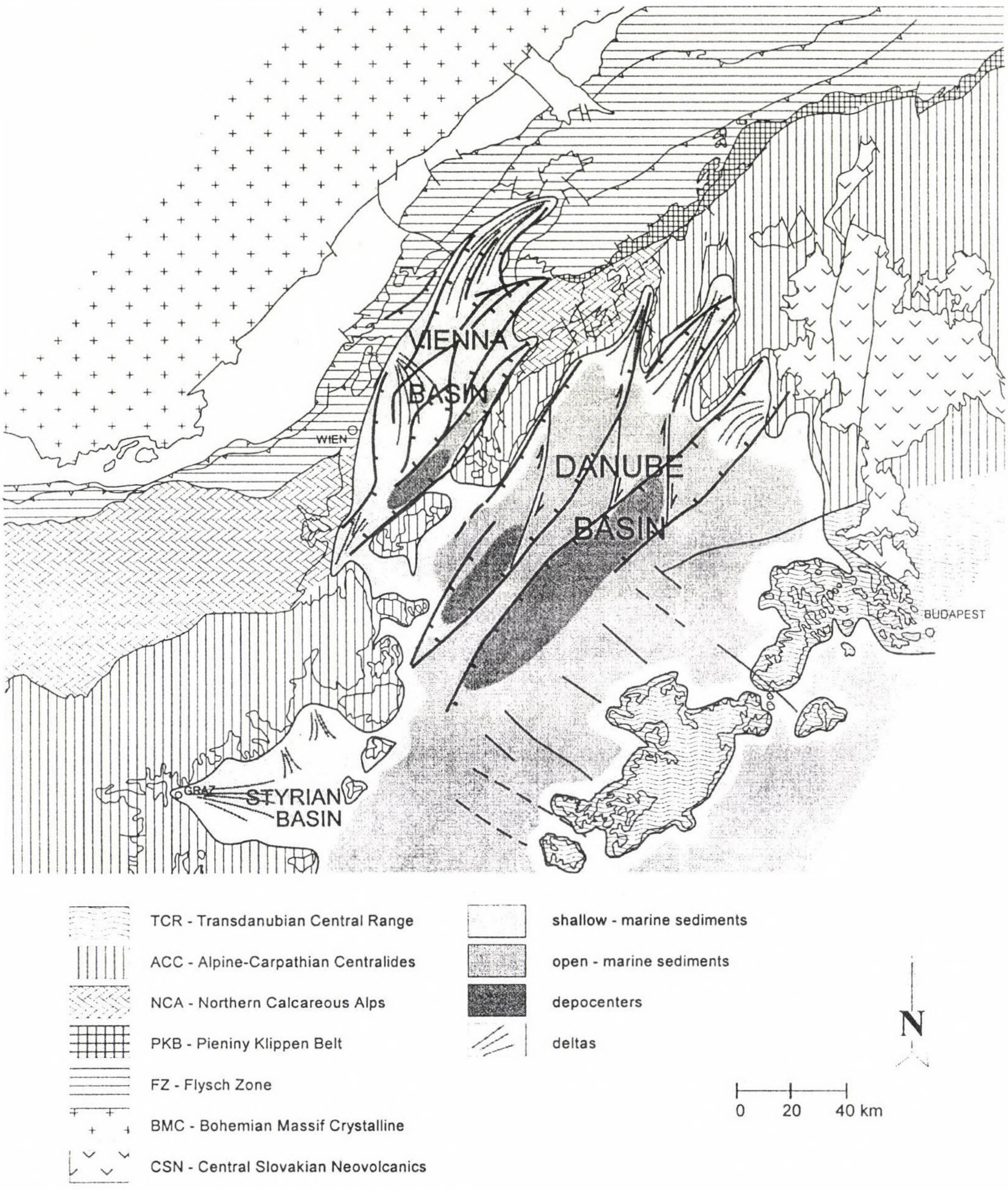

Fig. 6
The Pannonian–Pontian basin system in the Alpine–Carpathian–Pannonian junction. The positions of Vienna, Budapest and Graz on the geodynamic evolutionary sketch are shown in palinspastic locations

sediments suggest that most of the Transdanubian Central Range was probably flooded by the Pannonian lake (Jámbor 1980 1989).

In the central part of the brackish basin (Gabčíkovo depression) an alternating clayey–sandy sedimentation continued, reaching a maximum thickness of 4000–5000 metres (Buday et al. 1962; Körössy 1987). This basin was filled up by sediments transported by the rivers from the northern edge (Fig. 6).

In the brackish Styrian Basin, after a short period of marly–clayey sedimentation a limnic coal-bearing sequence was deposited during the Pannonian (Ebner and Sachsenhofer 1991).

Near the boundary between the Upper Miocene and Early Pliocene a rapid inversion took place in almost the entire Alpine–Carpathian–Pannonian (ALCAPA) area (Csontos 1995), excluding the Pliocene grabens of the Vienna Basin running in the footwall of the Leitha and Malé Karpaty Mts. (Gutdeutsch and Aric 1988; Brix, Plöchinger et al. 1988) and the Gabčíkovo–Komjatice depression in the Danube Basin (Adam and Dlabac 1969; Gaza 1984; Baráth and Kováč 1995; Horváth 1995). In this time large accumulations of fluviatile and limnic gravels originated due to differences of vertical movements in the area of the Western Carpathian orogenic belt.

Conclusions

The Oligocene–Lower Miocene Alpine collision forced the extrusion of the Western Carpathian–North Pannonian lithospheric fragments of ALCAPA domain to the E–NE.

In the frontal part of the Western Carpathians a transpressional tectonic regime controlled the marine sedimentation in the piggy-back basins situated in the Flysch Belt, and in the relic forearc and wrench fault furrow-type basins on the northern edge of the Central Western Carpathians during the Eggenburgian.

The Ottnangian compressive event led to isolation of the basins. In front of the Alpine–Carpathian orogenic belt a brackish to limnic (or anoxic) sedimentary environment prevailed. The initial back-arc extension, associated with CCW rotation of the ALCAPA domain, opened the Styrian Basin and also valleys in the Alpine–Carpathian junction area (Ebner and Sachsenhofer 1991).

During the Karpatian the oblique collision of the Western Carpathians with the North European platform was associated with the pull-apart opening of the marine Vienna Basin in a transtensional regime. The uplift of the central zone of the orogenic belt (internides) was followed by the evolution of a drainage system oriented toward the Carpathian externides. The back-arc extension accompanied by CCW rotation led to subsidence and volcanism in the Styrian Basin and initial synrift subsidence in the Danube Basin along NNE–SSW to NE–SW-oriented low-angle listric faults (e.g. the Répce and Rába lines).

The subsequent Middle Miocene events led to a gradual collapse of the orogenic belt in the Alpine–Carpathian–Pannonian junction. Astenospheric updoming in the Pannonian back-arc domain and active NE to E-trending elongation led to accelerated subsidence in its marginal zone, where an extensional tectonic regime controlled basin formation.

The Early Badenian subsidence of the Vienna Basin was accompanied by an uplift of the Malé Karpaty–Leitha Mts. horst structure. The steepening of the relief enabled the accumulation of huge alluvial and debris apron-related coarse clastics fans at the eastern margin of the basin.

The Middle and Upper Badenian basin formation was associated with the development of a WSW–ENE-trending sinistral shear zone in the Western Carpathians, thus opening depocenters in the Vienna Basin northern part and the finger-like depressions along the Danube Basin northern margin (Blatné, Rišnovce, and Komjatice depressions). The subsidence of the central and southern part of the basin was still controlled by synrift back-arc extension, which reactivated the listric low-angle faults in the Little Hungarian Plain area.

The Badenian sedimentary environment in both the Vienna and the Danube basins was marine. It is important to note the development of a new deltaic system prograding towards the back-arc basin region, due to the gradual uplift of the Outer Carpathian Flysch Zone accretionary prism.

During the Sarmatian basin extension continued only in the marginal zone of the back-arc area. The sinistral wrench zone running from the northern part of the Vienna Basin (Kúty and Kovalov depressions) eastward still controlled the sedimentation in grabens along the northern margin of the Danube Basin (Blatné, Rišnovce, and Komjatice depressions). The central and southern part of the Danube Basin were slightly uplifted and the subsidence decreased similarly as in the Styrian Basin. The marine sedimentary environment changed to a brackish one due to isolation from the Mediterranean.

The Upper Miocene brackish to freshwater sedimentation in the area of the collapsed orogenic belt was controlled by regional extension, with some compressional events. In the Vienna Basin the NE-running wrench zone activity controlled the sedimentation in pull-apart depressions and grabens along the Leitha fault zone. The sedimentation in the central and southern part of the Danube Basin was controlled by post-rift thermal subsidence combined by a widening of the depositional area documented by extended transgression during the Pannonian.

During the Pontian and Pliocene tectonic inversion occurred as is shown by the widespread depositional gap and erosion in the region considered. Important subsidence continued only in the central part of the Danube Basin in the Gabčíkovo–Komjatice depression.

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Sinistral lateral displacement in the Aggtelek–Rudabánya Mts (North Hungary) based on the facies distribution of Oligocene and Lower Miocene formations

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Renewed geologic field observations were carried out from 1980 to 1985 in the Aggtelek Rudabánya Mts (the southern extension of the Gemer Karst). Based on the stratigraphy, facies distribution, and structure of post-Mesozoic formations, a synsedimentary lateral displacement system was identified which corresponds to the Darnó zone.

Key words: allodapic limestone, Darnó-zone, Egerian, Eggenburgian, lateral displacement, oil shale, radiometric ages, schlieren

Stratigraphy

Kiscell Clay Formation, Novaj Member (Middle Oligocene, Upper Kiscellian substage)

The identification of the formation in two iron-prospecting boreholes (Rudabánya Rb-390 and Rb-414) drilled in 1953 and 1954 has been problematic for a long time (Sidó and Szőts in Pantó 1956; Majzon 1961; Radócz in Alföldi et al. 1975). The preserved cores of the Rb-390 borehole were re-sampled, while the sequence of borehole Rb-414 was correlated with the former one by means of lithological similarity.

The sequence is probably underlain by Upper Triassic sedimentary rocks. It consists of alternating glauconitic sandstone, marl and calcareous clay strata, with allodapic lithothamnium-bearing limestone lenses.

The massive, sandy limestone is a mixed sediment of intraclastic biomicrosparite and grainstone microfacies. Larger foraminifers (*Nummulites boullei* de la Harpe A, *Eulepidina* sp., *Nephrolepidina* sp., determined by Kecskeméti in Szentpétery 1988b) indicate Middle and possibly Early Oligocene age. Nannoflora (determined by Báldi-Beke in Szentpétery 1988b) from the loose sedimentary rock clearly indicates the NP 24 zone (*Sphenolithus distentus* horizon) known from several borehole profiles in Budapest and in the Great Hungarian Plain (Báldi-Beke 1977; Báldi-Beke et al. 1980; Báldi 1983).

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The lithology and origin of this sequence is similar to that of the Novaj Member of the Kiscell Clay Formation, but its stratigraphic position is different. Further studies are necessary to determine whether the member should be subjected to a twofold subdivision or not (Báldi 1983), whether this assignment is applicable, or the introduction of a new unit is needed.

Its thickness (based on an average dip of 60) is about 30 m.

Bretka Limestone Formation
(Lower Miocene, Upper Egerian)

This unit extends over a small area; erosional remnants occur in regions I and III (Fig. 1). The Bretka Limestone is a light grey to yellowish grey, coarse to medium crystalline, compact, less porous, bioclastic conglomeratic breccia with calcareous matrix containing Lithothamnium nodules and mollusc shells. Microfacies: biopelsparite, grainstone. Besides smaller foraminifers there are larger ones in rock-building quantities belonging to the group *Miogypsina gunteri* Cole *Miogypsina tani* Drooger (Varga 1977; Báldi 1983; Less 1991).

The formation is a typical abrasional conglomeratic breccia deposited at a steep cliff. Rounded grains have been formed by strong wave action, while their disintegration produced minor, angular grains.

Palaeontological data indicate normal saline, warm marine environment.

Its maximal thickness is 10 m.

Szécsény Schlieren Formation (corresponds to the Putnok Schlieren Formation)
(Lower Miocene, Upper Egerian–Eggenburgian)

Outcrops are located in region I, while borehole data indicate a common occurrence in all regions, except Nos III and V.

The dominant lithological characters indicate twofold subdivision:

1. The lower, usually 50 m-thick section starts with a basal conglomerate with rare tuffite intercalations. The glauconitic conglomerate–sandstone sequence contains rare marl beds. In region No. II in the Rb-407 borehole the Upper Egerian age of the Lithothamnium marl beds is proved by foraminifers and nannoplankton. A similar result was obtained for the Tornabarakony-1 borehole in region VI. There are frequent mollusc shell and coal strips, tuffaceous and tuffitic beds.

Sudden grain size changes in this section indicate frequent changes in depth and a mobile bottom, clearly shown by foraminifers according to Korecz-Laky (in Szentpétery 1988b). The entire fauna indicates a normal saline, shallow marine environment.

Radiometric ages measured on glauconite range from 19.4 to 24.6 $\pm$ 0.9–1 Ma (Debrecen, ATOMKI 1982 in Szentpétery 1998b). Mineralogical studies by

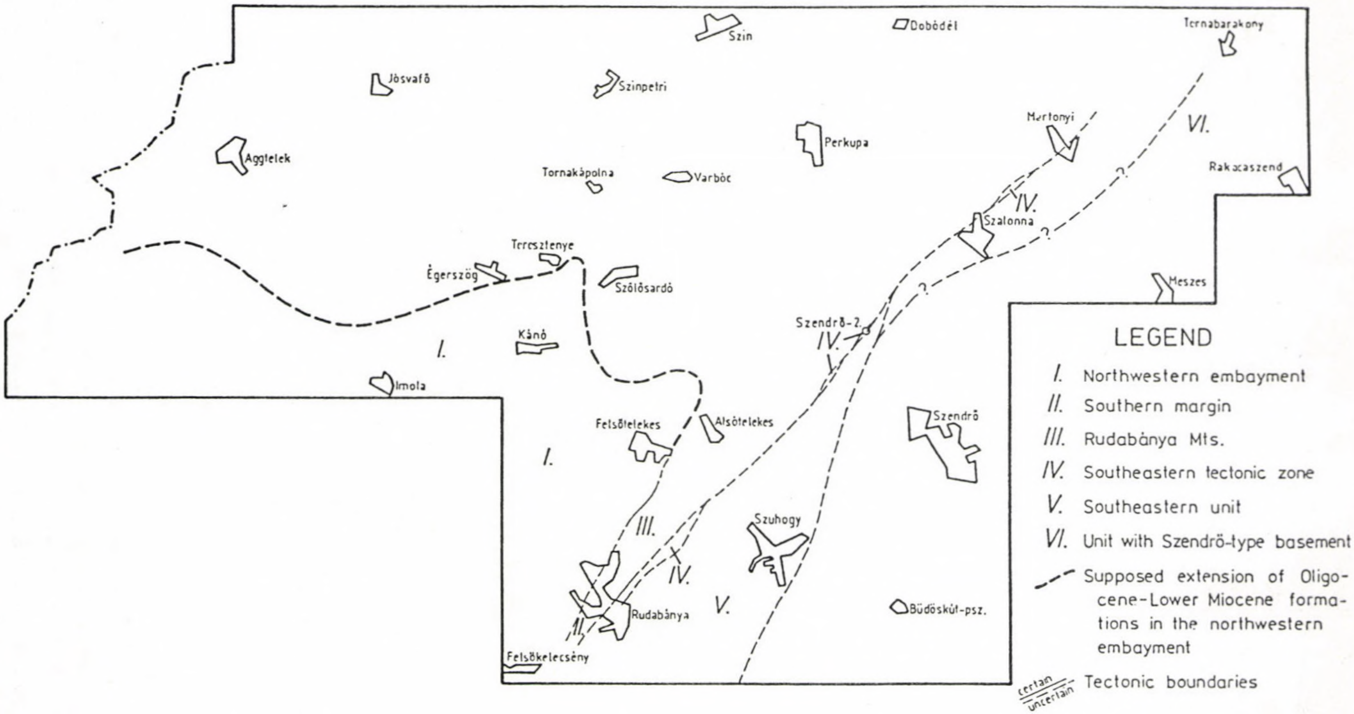

Fig. 1
Areas of occurrence of Oligocene and Lower Miocene formations of the Aggtelek–Rudabánya Mts

Ravasz-Baranyai (in Szentpétery 1988b) indicate that this age corresponds well with the expected geological age.

2. By gradual decrease of sand content the upper part of the sequence becomes silty to clayey silty with a variable carbonate content. Bedding is poor, cleavage occurs along fossils and carbonized pyritic plant fragments. There are characteristic traces of bioturbation and large muscovite flakes, indicating a quiet depositional environment.

The formation is not a typical schlieren, as shown by mollusk studies of Bohn-Havas (in Szentpétery 1988b): *Nucula nucleus* L., *N. fragilis* Chemn., *Yoldia longa* Bell., "*Amussium*" sp., *Saxolucina bellardiana* Bell., *Macoma elliptica* Bell., *Corbula gibba* Olivi, *Dentalium* sp.

This section certainly belongs to the Eggenburgian Stage. Spores, pollen and nannoplankton in the Rb-391 borehole (region II) may indicate Upper Eggenburgian to Lower Ottnangian age (Bodor and Báldi-Beke in Szentpétery 1988b).

Its maximal thickness is 200 m.

Szuhogy Conglomerate Formation
(Lower Miocene, Eggenburgian?)

This unit is a long, northward-extending band, easily followed on the surface in region V (Fig. 1). It follows the morphological boundary of Rudabánya Mts. forming a 4.5 km long, maximally 1 km-wide zone. Schréter (1952) believed it to possibly be part of the Gosau, based on the analogy of the Nekézseny Conglomerate. According to Balogh (1949), Balogh and Pantó (1952) and Pantó (1956) it is a "Mediterranean" stage.

Boulders, cobbles and pebbles of this cyclic, terrestrial fan display extremely variable size and rounding; most of them are composed of metamorphosed limestones, and to a lesser degree of shales. Most of them are similar to the Palaeozoic formations of the Szendrő Mts. There are very few non-metamorphosed grains: Lower Triassic littoral limestone, Upper Triassic basinal and platform limestone, and Jurassic–Lower Cretaceous radiolarite breccia embedded in Upper Cretaceous limestone matrix.

Among the pebbles larger than 1 cm there are no clasts from the neighbouring Rudabánya Mts. Rarely, redeposited Oligocene nannoplankton and Egerian foraminifers were found in the yellow to red sandy matrix; therefore any relationship to the Senonian Nekézseny Conglomerate Formation can be excluded.

The Szalonna-9 borehole in the southeastern tectonic zone (region IV) encountered a finer-grained variety of the formation. From the bottom up to 235.1 m there are quartz pebble beds in the red siltstone–sandstone sequence; upward, clasts from the Rudabánya Mts. Also appear, too (recognized by Less, Gy.). The palynology of the entire sequence ranged into the Eggenburgian (determined by Bodor and Bóna in Szentpétery 1988b).

Its thickness is 350 m in Szuhogy-6 borehole, but the dip is 30–60. The Szuhogy-6 borehole exposed the underlying Uppony-type Abod Limestone Formation.

Oil shale of poor quality
(Lower Miocene, Eggenburgian)

The Szendrő-2 borehole (region IV, southeastern tectonic zone) crossed a laminitic rock, made of greenish-brown and white laminae, between 10.0–74.9 m. The appearance and features of the folded, steeply dipping sequence indicate oil shale. However, the average volume weight (2.3 g/cm^3) is too high for an oil shale. The organic content (DTA-DTG-TG analysis) is 10% at most, the Soxleth bitumen is 0.117%, and the sulphur content is 0.045% (the latter values are from a sample with 10% organic matter). These values are similar to those of other localities in the Borsod Basin (Radócz 1984). Further studies are under way.

Spore-pollen studies (Bodor and Bóna in Szentpétery 1988b) indicate Early Miocene, most probably Eggenburgian age of the sequence.

Conclusions

Six structural units, separated by tectonic lines, have been recognized within the Oligocene and Lower Miocene formations (Fig. 1).

The facts and conclusions are listed as follows:

1. Flat-lying seashore sediments occur in the northern and northeastern margins of region I. These are the original margins of the Lower Miocene sea.
2. Sediments of the ENE boundary zone of the northwestern embayment (region I) indicate a mobile basement. Synsedimentary tectonic movements occurred here probably by the end of Egerian.
3. Oligocene sediments covering the Mesozoic basement of the southern (II) region are identical with those in the Novaj-2 borehole. These two localities are 50 km apart at present. They have been separated by dextral strike-slip displacement. This occurred before Egerian time, as the Szécsény Schlieren does not display tectonic disturbances.
4. There are mixed, Mesozoic blocks in tectonic position above Oligocene sediments in region II. These may have been emplaced by tectonic events prior to the termination of the Egerian.
5. The Szécsény Schlier transgressively overlies basement rocks of variable ages and lithologies in the southern margin. A new basin with a dissected bottom was formed by the end of the Egerian.
6. There are erosional relicts of the autochthonous Bretka Limestone Formation in the southeastern side of the Rudabánya Mts, where tectonic contacts with contemporaneous formations in different facies can be seen. The

material in this locality is a seashore sediment of the Lower Miocene sea, rotated and displaced several km to the northeast, together with its basement (Fig. 2).

7. Eggenburgian Szécsény Schlieren, Szuhogy Conglomerate and oil shale occur in allochthonous position in the southeastern (IV) tectonic zone. These relicts were deposited in the same basin, and emplaced in their present tectonic position by subsequent sinistral lateral displacement.

8. Rock fragments not characteristic of the Rudabánya Mts are found below the depth of 235.1 m within the fine-grained Szuhogy Conglomerate in the Szalonna-9 borehole. The formation contains these clasts above that depth. Since the entire sequence is Eggenburgian, the Rudabánya Mts (region III) and the unit containing the Szuhogy Conglomerate above Uppony-type basement (region V) were placed side by side during Eggenburgian time.

9. The clast of the antithetic Szuhogy Conglomerate contains alien pebbles. It may have been formed in the northern foreland of the Szendrő Mts, and was emplaced in its present tectonic position by Early to Middle Miocene strike-slip faulting.

10. The Tornabarakony Tbr-1 borehole contains a Szécsény Schlieren sequence over a Szendrő-type Palaeozoic basement; it can be correlated with the sedimentary sequence of the Rb-407 borehole. The two boreholes are 17 km apart from each other, indicating sinistral lateral displacement (Fig. 2).

The above list indicates a multi-stage structural evolution. The first phase occurred during the second half of the Oligocene: a single, dextral lateral displacement (Rb-390 and -414 boreholes). A probable sequence of the Early Miocene (Late Egerian–Eggenburgian) events is shown in Fig. 2. The presented sinistral lateral displacement system displays an eastward progressive displacement of segments in space and time. Separate elements of the displacement cannot be proven by one zone. The studied region is part of a large-scale tectonic zone, where each unit displays different relationships to its neighbours. The "large-scale tectonic zone" refers to the Darnó zone. It is a young lateral displacement system, as shown by Jaskó (1946), Pantó (1956), Balla (1982), Zelenka et al. (1983), Balla (1984), and Báldi and Báldi-Beke (1985). The displacement was several tens of kms in magnitude. The only help in estimating the displacement is the 17 km distance of the Rb-407 and Tbr-1 boreholes. However, it should be stressed that this is only a single point of a tectonic zone, the size of which far exceeds the size of the studied area.

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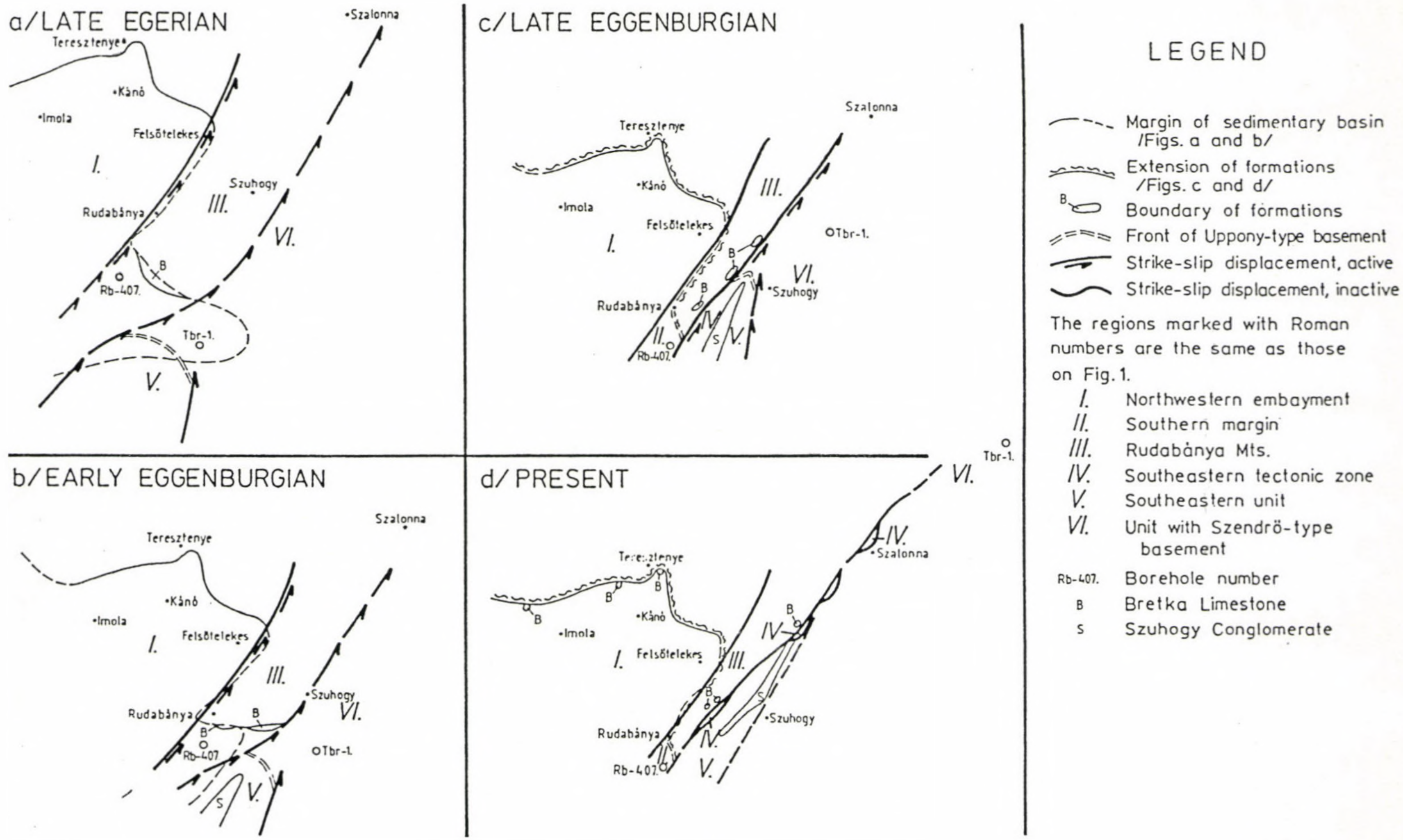

Fig. 2
Scheme of Lower Miocene terctonic events in the Aggtelek–Rudabánya Mts (not to scale)

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Cretaceous and Palaeogene sedimentary evolution in the Eastern Alps, Western Carpathians and the North Pannonian region: An overview

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On the basis of 44 sections exhibiting both the major facies and the trends in the terrigenous input, the sedimentary evolution of the Eastern Alps, Western Carpathians and the Transdanubian Range during the Cretaceous and Palaeogene is summarised. The three segments of the Alpine orogenic belt are characterised by a predominantly similar sedimentary history. It is demonstrated that two main sediment source terrains supplied the Austroalpine–Central Carpathian domain and the Transdanubian Range; one was situated along the northern active margin and the other was the intra-orogenic Tethys suture zone to the south. In spite of the fundamental similarity in the geological evolution, differences in both facies and time in the Upper Cretaceous and Palaeogene successions between the Central-Alpine domain of the Eastern Alps and the Transdanubian Range can be explained by an eastward shift of the latter unit along the Tethys suture zone.

Key words: Eastern Alps, Western Carpathians, Transdanubian Range, Cretaceous, Palaeogene, facies development

Introduction

The Eastern Alps, the Western Carpathians and the Pannonian region represent separate segments of the Alpine orogenic belt. Palaeogeographically, however, the entire sedimentary evolution belongs to the western Tethys realm and, therefore, the major events in its geodynamic history are the same in all three zones. Nevertheless, some problems arise in correlating the sedimentary sequences and tectonic units of these three regions, because Neogene and

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Quaternary sediments obscure various geological junctions. Furthermore, the circumstance that three different languages are used and that, for many decades, the political boundaries slowed down free scientific communication make the problems no easier to solve. As a result of this situation, a great variety of local formation names masks a considerable number of similarities in the facies and tectonic evolution of different sectors within the region. The purpose of this paper is, therefore, to give an overview of trends in the sedimentological character of these three neighbouring regions, which reflect their common Alpidic geodynamic evolution. The paper is especially focused on the Cretaceous and Palaeogene evolution because during these periods decisive orogenic events, such as the Eoalpine and Mesoalpine tectonic movements, occurred.

This overview is based on observations and discussions derived from eight field workshops of the Cretaceous–Palaeogene working group of the ALCAPA project, established by the Austrian, Hungarian and the Slovak Academies of Sciences. During the workshops, typical localities in the region of the eastern parts of the Eastern Alps (EA), the Western Carpathians (WCa) and the Transdanubian Range (TR), as shown in Fig. 1, were studied. Five to ten persons from each country contributed to the workshops. Details of the results will be published in separate papers. The sedimentary evolution is described by two kinds of sections: 1) stratigraphic sections exhibiting the major facies evolution and 2) sections which give information about the palaeogeographically significant terrigenous input. The numerous local formation names are listed in Table 1 and are also indicated as abbreviations on the right side of the stratigraphic columns.

For the Eastern Alps, the Cretaceous and Palaeogene sedimentary evolution is summarised as a whole or in parts by Oberhauser (1963, 1980, 1995), Tollmann (1976), Herm (1979, 1981), Flügel and Faupl (1987), Faupl and Wagreich (1992a, b, 1996), Wagreich and Faupl (1994) where further references are compiled. Overviews for the Western Carpathians (except for the Flysch zone not included in this study) have been published by Andrusov (1965a, b), Salaj and Samuel (1966), Samuel and Salaj (1968), Marschalko (1986), Birkenmajer (1977), Gross et al. (1980), Rakús et al. (1989, 1990), Gross et al. (1993) and Vašíček et al. (1994). For the Transdanubian Range summaries are given by Haas (1979, 1983, 1987), Korpás (1981), Császár (1984, 1986), Császár and Haas (1984), Báldi and Báldi-Beke (1985), Kázmér (1985b), Kázmér and Kovács (1985), Császár et al. (1987), Haas and Császár (1987), and Fodor et al. (1994).

Regional geologic setting

The Eastern Alps

The Eastern Alps (EA) are composed of three major tectonic units: the Helvetic, the Penninic and the Austroalpine nappe complexes. Northern units

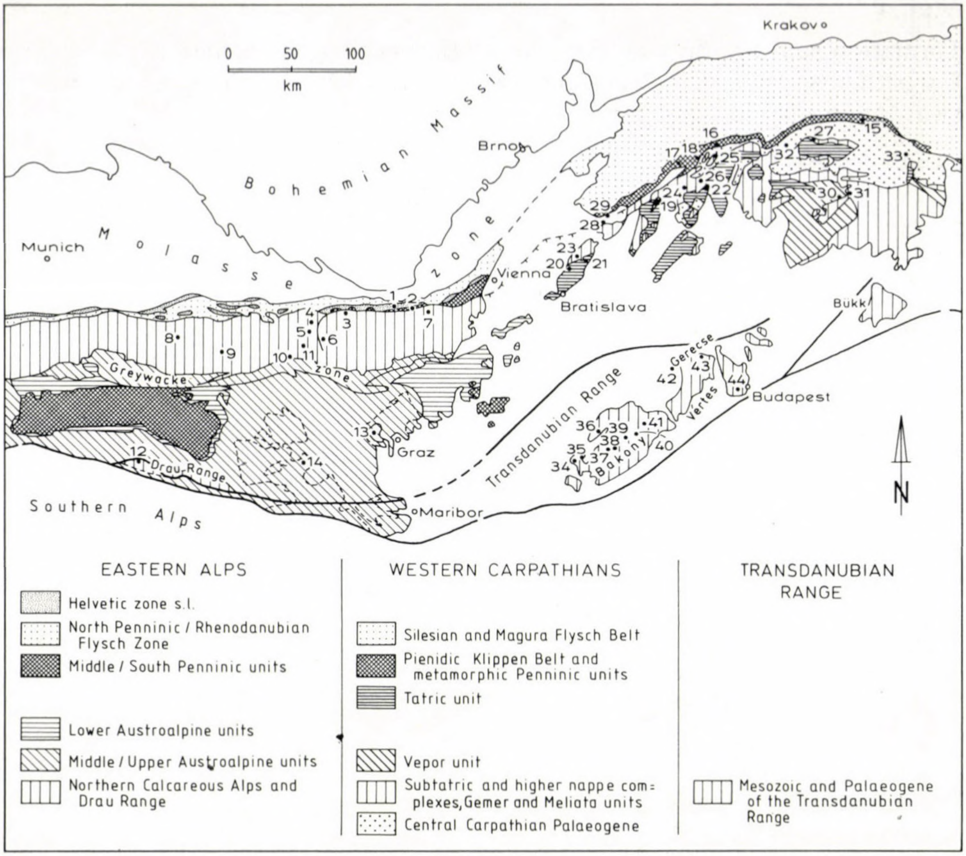


Fig. 1

Geologic map of the Eastern Alps, Western Carpathians and North Pannonian region. Numbers refer to the sections studied: *Eastern Alps:* 1. Gresten Klippen zone; 2. Ybbsitz zone; 3. Frankenfels/Ternberg nappe; 4. Reichraming nappe (Schneeberg Syncline); 5. Reichraming nappe (Ebenforst Syncline); 6. Lunz nappe (Weyerer Bögen); 7. Lunz nappe (Lilienfeld); 8. Tirolikum (Salzachtal); 9. Gosau Type Locality; 10. Gosau of Wörschach. 11. Gosau of Windischgarsten; 12. Drau Range (Lienzer Dolomiten); 13. Central-Alpine Gosau of Kainach; 14. Central-Alpine Krappfeld-Gosau. *Western Carpathians:* 15. Pieniny Klippen Belt (Czorsztyn zone); 16. Pieniny Klippen Belt (Kysuca-Pieniny zone); 17. Klape unit (Middle Váh Valley); 18. Manín unit (Manín Gorge region); 19. Belice unit (Považský Inovec Mts); 20. Tatricum (Malé Karpaty), Kuchyňa unit; 21. Tatricum (Malé Karpaty), Solírov and Orešany unit; 22. Tatricum (Malá Magura), Malá Fatra unit; 23. Krížna nappe (Vysoká unit); 24. Krížna nappe (Belá unit); 25. Krížna nappe (Zliechov unit, Polomec); 26. Krížna nappe (Zliechov unit, Strážovce); 27. Krížna nappe (Zliechov unit, Oravice); 28. Brezová Group (Bradlo facies), Brezovské Karpaty; 29. Brezová Group (Surovín facies), Brezovské Karpaty; 30. Šumiac; 31. Dobšinská Ladova jaskyňa; 32. Central-Carpathian Palaeogene (Orava area); 33. Central-Carpathian Palaeogene (Spiš region). *Transdanubian Range:* 34. Sümeg; 35. Darvastó; 36. Magyarpolány; 37. Ajka and Padrag; 38. Úrkút; 39. Háskút; 40. Olaszfalu; 41. Bakonynána; 42. Tata; 43. Lábatlan (Gerecse); 44. South Buda

Table 1
Formation names of the Cretaceous and Palaeogene of the Eastern Alps, Western Carpathians and the Transdanubian Range. Abbreviations used in the stratigraphic sections are given

Eastern Alps

Gresten Klippen Belt:

Bla Blassenstein Fm.
Bun Buntmergelserie

Ybbsitz zone:

Fas Fasselgraben Fm.
Glo Glosbach Fm.
Has Haselgraben Fm.
Kah Kahlenberg Fm.
Ybb Ybbsitz Fm.

Northern Calcareous Alps:

Bra Branderfleck Fm.
gBru Brunnbach Fm.
gHöl Höllgraben Fm.
gLSG Lower Gosau Subgroup
gNie Nierental Fm.
gRes Ressen Fm.
gSpi Spitzenbach Fm.
gWör Wörschachberg Fm.
gZwi Zwieselalm Fm.
Gra Grabenwald Mb. of the Rossfeld Fm.
Los Losenstein Fm.
Ros Rossfeld Fm.
Schr Schrambach Fm.
Tan Tannheim Fm.

Drau Range (Lienzer Dolomiten):

Lalm Lavant Fm. - lower member
Laum Lavant Fm. - upper member
Schr Schrambach Fm.

Central-Alpine Gosau occurrences:

gB Gosau-"Basisschichten"
gBc Gosau-"Basalkonglomerat"
gBit Gosau-Bitumenmergelfolge
gHau Gosau-Hauptbeckenfolge
gTur Gosau-"Turbiditfazies"
Gut Guttaring Group
gZem Gosau-Zementmergelfolge

Western Carpathians

Pieniny Klippen Belt:

BrB Brodno Beds
Chm Chmielowa Fm.
CzF Czorsztyn Fm.
Dur Dursztyn Fm.
Jar Jarmuta Fm.
Jaw Jaworki Fm.
Kap Kapusnica Fm.
Lys Lysa Fm.
PiF Pieniny Fm.
Pom Pomiedznik Fm.
Puh Púchov Fm.
Sne Snežnica Fm.
Sor Sromowce Fm.
Spi Spiš Fm.

Klape Unit:

Ihr Ihrište Fm.
Kap Kapusnica Fm.
Kra Kravárik Fm.
Nim Nimnica Fm.
Orl Orlové Fm.
Puh Púchov Fm.
Pvb Považská-Bystrica Fm.
Sfr Šafranica Fm.
Sne Snežnica Fm.
Sro Sromowce Fm.
Stp Štepnica Fm.
Tiss Tissalo Beds
Uhr Uhry Fm.
Upo Upohlav Fm.

Manín Unit:

But Butkov Fm.
Hlk Hlboké Fm.
Hrb Hrabové Fm.
Kal Kališčo Fm.
Luk Luckov Fm.
Noz Nozdrovice Breccia
Pod Podhorie Fm.
Pra Praznov Fm.
Zad Žadorec Fm.

Belice Unit:

HBF Horné Belice Fm.

Laz Lazy Fm.

Tatricum:

Luc Lučivná Fm.
Oso Osobitá Fm.
Par Párnica Fm.
Por Poruba Fm.
Schr Schrambach Fm.
SoB Somár Breccia
Sol Solírov Fm.

Krížna Nappe:

Boh Bohatá Fm.
But Butkov Fm.
Hlb Hlboč Fm.
Kos Koscieliska Fm.
Mra Mráznica Fm.
Osn Osnica Fm.
Par Párnica Fm.
PaV Padlá Voda Fm.
Pod Podhorie Fm.
Por Poruba Fm.
Schr Schrambach Fm.
Str Strážovce Fm.

"Post-tectonic" Senonian and Palaeogene; Brezová Group, "Gosau"-type deposits:

Bar Baranec Sandst. (Mb. of the Ostriez Fm.)
DeV Dedkov Vrch Fm.
DIc Dobšiná Ice Cave Conglomerate
DIm Dobšiná Ice Cave Marls
Hur Hurbanova Dolina Fm.
Jab Jablonka Fm.
Kos Košariská Fm.
Kra Kravárik Fm.
KrzB Kržlá Breccia
LuF Lubina Fm.
Men "Menilite" Fm.
Mos Mosnáčov Marl (Mb. of the Bradlo Fm.)
Ost Ostriež Fm.
Pbr Podbradlianska Fm.
Plf Podlipovec Flysch (Mb. of the Bradlo Fm.)
Pol Polianka Fm.
Pri Priepastné Fm.
PsV Pustá Ves Fm.
SBr Široké Bradlo Limestone
(Mb. f the Bradlo Fm.)
Stv Štverník Marls (Mb. of the Ostriež Fm.)
Sum Šumiac Limestone
Val Valchov Conglomerate
(Mb. of the Ostriež Fm.)

Palaeogene of the Central Western Carpathians:

BiP Biely Potok Fm.
Bor Borové Fm.
Hra Hrabník Fm.
Hut Huty Fm.
Hsm Šambron Conglomerate
(Mb. of the Huty Fm.)
Puc Pucov Conglomerate
Zub Zuberec Fm.

Transdanubian Range

Ajk Ajka Coal Fm., $^{a}K_3$
BeM Bersek Marl Fm., $^{b}K_1$
BuM Buda Marl Fm., $^{b}E_3$-Ol_1
Csa Csatka Fm., $^{c}Ol_2$-Me or $^{c}Ol_2$
Cse Csehbánya Fm., $^{c}K_3$
Cso Csolnok Fm., $^{c}E_2$
Dar Darvastó Fm. $^{dt}E_2$
Dor Dorog Fm., $^{d}E_2$
Fel Felsővadács Breccia Mb. of the Bersek Marl Fm., $^{b}{}_{f}K_1$
Gan Gánt Bauxite Fm., $^{g}E_1$
JaM Jákó Marl Fm., $^{j}K_3$
KiC Kiscell Clay Fm., $^{k}Ol_1$
Kös Köszörűkőbánya Conglom. Mb., $^{l}{}_{k}K_{1-2}$
Lab Lábatlan Sandstone Fm., $^{l}K_{1-2}$
Man Mány Fm., $^{m}Ol_2$
MoL Mogyorósdomb Limestone Fm., $^{m}J_3$-K_1
PaM Padrag Marl Fm., $^{p}E_{2-3}$
PeM Pénzeskút Marl Fm., $^{p}K_2$
PoM Polány Marl Fm., $^{p}K_3$
SüM Sümeg Marl Fm., $^{s}K_1$
Sze Szentivánhegy Limestone Fm., $^{s}J_3$-K_1
SzL Szépvölgy Limestone Fm., $^{s}E_3$
SzöL Szőc Limestone Fm., ^{s}E
TaC Tard Clay Fm., $^{t}Ol_{12}$
TaL Tata Limestone Fm., $^{ta}K_2$
Tes Tés Clay Fm., $^{t}K_2$
TöS Törökbálint Sandstone Fm., $^{t}Ol_2$
UgL Ugod Limestone Fm., $^{u}K_3$
VeS Vértessomló Siltstone Fm., $^{v}K_2$
ZiL Zirc Limestone Fm., $^{z}K_2$

of this nappe pile tectonically overlie the southern parts of the Molasse sediments which cover the southern Bohemian crystalline basement and autochthonous epicontinental Mesozoic successions. Sequences of the Helvetic zone *s. str.* are known only in the western and middle segment of the EA, up to the east of Salzburg. Further to the east, the Gresten Klippen belt has been assigned to the Ultrahelvetic zone. The Penninic realm as a metamorphic development appears in the Penninic windows, such as the Unterengadin Window, the Tauern Window and, at the easternmost border of the EA, the small windows of Rechnitz, Bernstein and Meltern as well as Kőszeg in Hungary. Beside these metamorphic series, non-metamorphic Penninic successions are confined to the Rhenodanubian Flysch zone named by Oberhauser (1968) and the Ybbsitz zone (Schnabel 1979; Decker 1990). The latter seems to be an equivalent to the Arosa zone in the boundary region of the Eastern and Central Alps. The Austroalpine domain can be subdivided into three major nappe systems; the Lower, the Middle and the Upper Austroalpine units. The distribution of the Middle and the Upper Austroalpine units, especially in the crystalline basement, is discussed in Tollmann (1987) and Frank (1987).

The sections studied in this paper are concentrated in the eastern parts of the EA and, especially, in the Upper Austroalpine units such as the Northern Calcareous Alps (NCA) and the Drau Range. External units are only documented by comparative sections of the Gresten Klippen belt and the Ybbsitz zone, the latter interpreted as having an oceanic origin within the South Penninic domain. In the Lower and Middle Austroalpine tectonic units of the eastern parts of the EA, neither Cretaceous nor Palaeogene sediments have been preserved, except for the so-called "Upper Eocene of Kirchberg/Wechsel", a small redeposited remnant of a former larger sedimentary cover.

The nappe pile of the NCA, a prominent sub-unit of the Upper Austroalpine tectonic system, comprises Upper Permian to Eocene formations. During the Cretaceous (Eoalpine) orogeny the sedimentary cover of the NCA was sheared off from its substratum and emplaced along the active northern margin of the Austroalpine microplate. This microplate was separated from the European continent by the Penninic ocean. In Late Eocene/Oligocene time, the NCA were affected by a further extensive deformational event (Mesoalpine orogeny). In a last great pulse of deformation the Alpine body was compressed and fractured during the Miocene (Ratschbacher et al. 1991; Decker et al. 1994).

The Western Carpathians

The external tectonic units of the Western Carpathians (WCa) comprise rocks of the Flysch belt (the Krosno or Silesian and Magura zones) and the Pieniny Klippen belt. Only the Ždanice unit, a Subsilesian element, and the Magura Flysch can be correlated directly with units of the EA (Waschbergzone and Rhenodanubian Flysch zone) (Eliáš et al. 1990). The Flysch belt is not included in this synthesis. To the south of the Peri-Pieniny Klippen zone lies the complex of the Central WCa, which is composed of the Tatricum and the Subtatric or Central WCa nappe complex (Krížna, Choč, Strážov units), the Veporicum and the Gemericum. This structurally complex nappe system of the Central WCa is assumed to belong to the Austroalpine realm *s. l.* The Manín unit is believed to be a tectonic element from the northern part of the Central WCa nappe system. The Czorsztyn facies of the Pieniny Klippen Belt was accumulated on a continental crust fragment situated within the mainly oceanic realm, which is believed to be the continuation of the Penninic domain. The Klape unit, represented by Cretaceous Flysch formations, contains the maximum concentration of conglomerates with exotic pebbles derived from the subducting clastic wedge of the oceanic domain (Ligurian–Penninic–Pieniny Ocean) situated between the Czorsztyn elevation and the northern margin of the Central WCa. According to Plašienka (1995), the Klape unit originated far to the south and was transported to the north as a nappe from an Ultraveporic position.

The metamorphic Upper Cretaceous successions of the Belice unit cropping out in a window in the Považské Inovec Mountains can be correlated with metamorphic Penninic units of the EA. Plašienka et al. (1994) assumed that this unit was situated palaeogeographically to the north of the Tatric units and thought it to be a unit similar to the Matrei zone from the southern border of the Tauern window in the EA. Metamorphosed Eocene sequences were only found in the boreholes beneath the Neogene sediments of the East Slovakian Basin (Soták et al. 1993). This Pozdišovce–Iňačovce unit probably belongs to the Eastern Carpathians.

Although the Tatricum of the Malé Karpaty shows similarities with the Lower Austroalpine units of the EA, the Tatricum of the Tatra and Fatra Mountains has been located at the northern edge of the Austroalpine microplate and contains specific Carpathian elements. The Krížna cover nappe system can be correlated with the tectonically lower nappes of the NCA, such as the "Cenoman-Randschuppe" and the Frankenfels nappe. The Choč and Strážov nappes can be correlated with tectonically higher units of the NCA. The Meliata unit of the southern parts of the WCa, with its Triassic oceanic facies elements (Kozur and Mock 1985), marks the Tethys suture zone and this extends westwards to the eastern segment of the EA (Kozur 1991; Kozur and Mostler 1992; Mandl and Ondrejičková 1991).

The Transdanubian Range

Within the Pannonian region, this paper is focused on the Cretaceous and Palaeogene development of the Transdanubian Range (TR). As the north-western part of the Pelso unit, the TR is situated along the southern border of the Austroalpine and Central WCa units, separated by major bounding faults (Rába Line – Scheffer and Kántás 1949; Rába and Diósjenő

Lines – Brezsnyánszky and Haas 1986). It is bounded to the south by the Balaton Line. The Central Hungarian Fault zone forms the fundamental tectonic contact between the Pelso and the Tisza units (Brezsnyánszky and Haas 1986). The Mesozoic stratigraphic successions of the TR comprise facies similar to the Northern and Southern Alps, with a closer relation to the latter (Kázmér and Kovács 1985; Császár and Dosztály 1993). Its formations are also dissimilar to the Alpine ones. Cretaceous formations of the TR crop out in the Bakony and Vértes Mountains, as well as in the Gerecse Mountains. Palaeogene deposits underlain by Middle Triassic to Upper Cretaceous formations are spread over a broad zone of the TR. Studied sections are situated in the Nyirád (Darvastó)–Padrag region (S-Bakony), the Gerecse Mountains and the Buda Hills (Kázmér 1985b; Fodor et al. 1994).

The separation of the TR began in the pre-Gosau tectonic phase. Its recent position was reached in the Early Miocene (Kázmér and Kovács, 1985). The distribution and changes in the stress field and tectonic phases were studied by Balla and Dudko (1989), Fodor et al. (1992, 1994) and Bada et al. (1996).

Facies distinguished in the sections

In the three Alpine regions, the facies development illustrated in the stratigraphic sections (Figs 2–5) is here reduced to nine major facies types. The *alluvial to fan-delta facies* comprises reddish alluvial successions, such as the basal conglomerates of the Gosau Group in Austria or the basal part of the Upper Cretaceous succession in the TR and WCa, and also including freshwater limestones and coal seams as well as transitional facies types. The occurrences of coal seams and freshwater limestones are indicated by separate symbols. In a few cases it appears necessary to infer a *lacustrine-paludal facies* as observed in the Tés Clay Fm. of the TR.

Grey shelf marls and neritic sandstone sequences are combined in the term *shallow-marine terrigenous facies*. Some sandstone successions are interpreted as storm-generated deposits. *Shallow-water limestones* comprise mainly the Urgonian-type platform carbonates of the Lower Cretaceous, the Upper Cretaceous rudistid limestones and the Eocene nummulite-bearing limestones as well. Further occurrences of rudist bioherms and nummulite beds are indicated separately.

The *deep-water limestones*, which occur frequently in the Lower Cretaceous successions, comprise pure Majolica-type limestones (Mogyorósdomb Fm. in the Bakony Mts.) as well as marly micritic limestones, typical of the upper parts of the Schrambach Fm. This latter type interfingers in several sections with thin turbiditic layers.

The *bathyal marl facies* is characterised by high proportions of planktonic foraminifera but also includes successions which exhibit transitional sediments deposited between outer shelf and slope deposits. Slump deposits occur frequently in this facies (Bersek Marl in Gerecse; Polány Marl in Bakony;

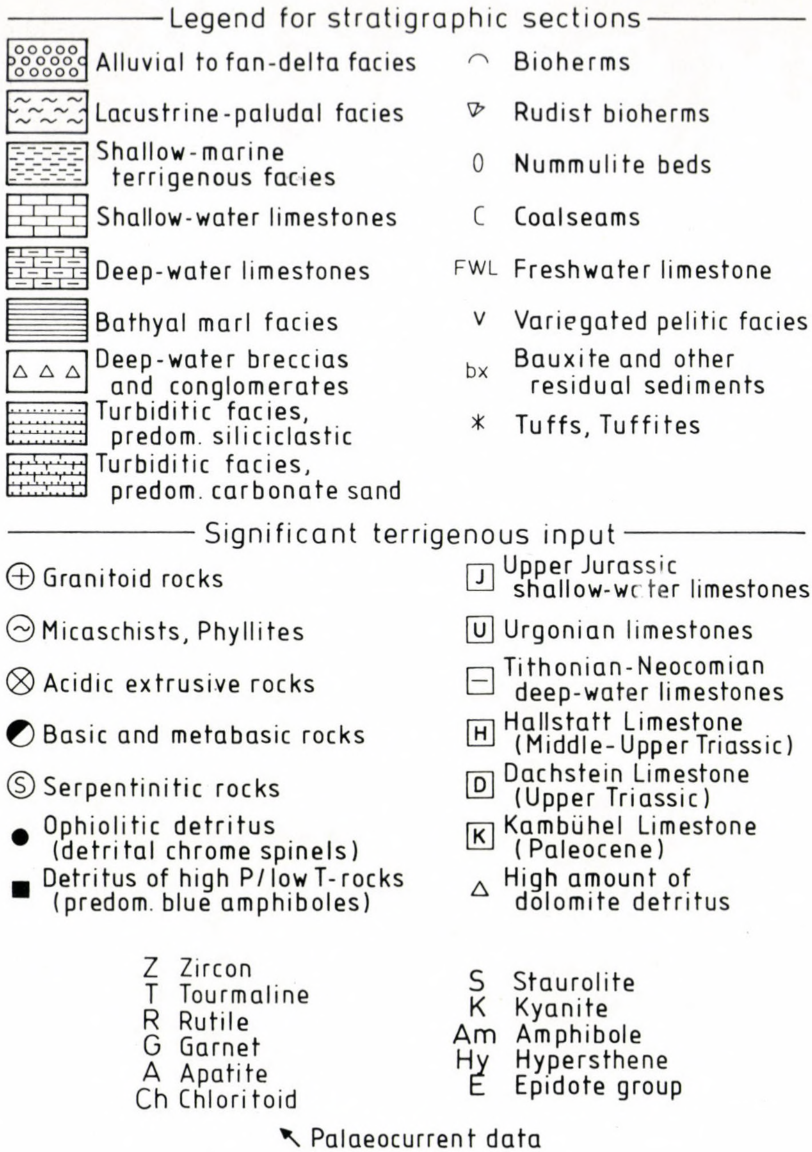


Fig. 2
Legend to the section on Figs 3–5

Nierental Fm. in the NCA). Reddish and variegated colours as well as dark grey series have been observed (Nierental Fm. in the NCA; Bersek Marl in Gerecse; Aptychenschichten in the North Karawanken; Košariská Fm. in the Brezovské Karpaty).

Fig. 3a, b
Stratigraphic sections of the Eastern Alps

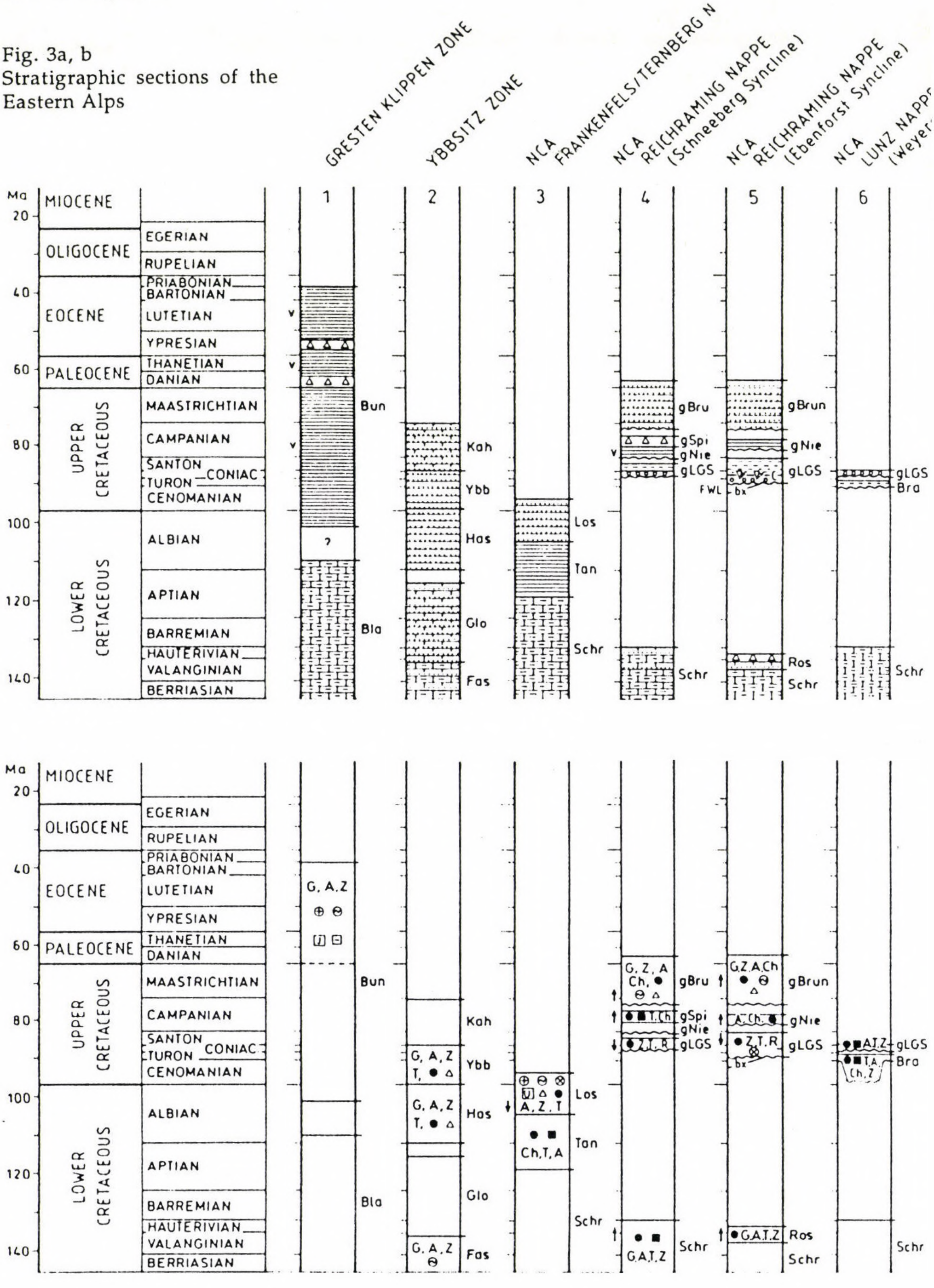
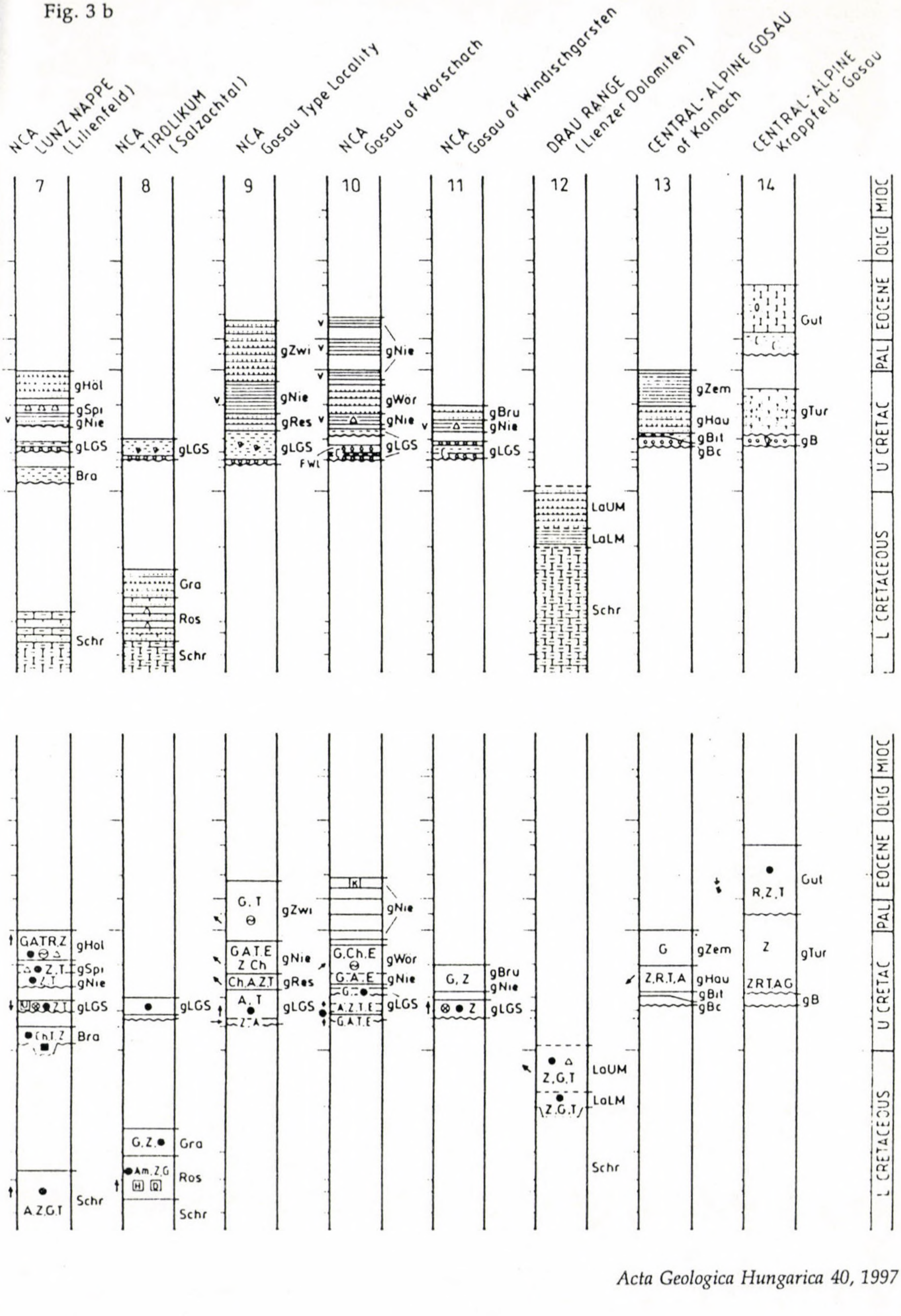


Fig. 3 b


NCA LUNZ NAPPE (Lilienfeld)
NCA TIROLIKUM (Salzachtal)
NCA Gosau Type Locality
NCA Gosau of Worschach
NCA Gosau of Windischgarsten
DRAU RANGE (Lienzer Dolomiten)
CENTRAL-ALPINE GOSAU of Kainach
CENTRAL-ALPINE Krappfeld-Gosau
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8
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11
12
13
14
MIOC
OLIG
EOCENE
PAL
U CRETAC
L CRETACEOUS
gHöl
gSpi
gNie
gLGS
Bra
Schr
Gra
Ros
gZwi
gRes
FWL
gWor
gBru
LaUM
LaLM
gZem
gHau
gBit
gBc
Gut
gTur
gB

Fig. 4a, b, c
Stratigraphic sections of the Western Carpathians

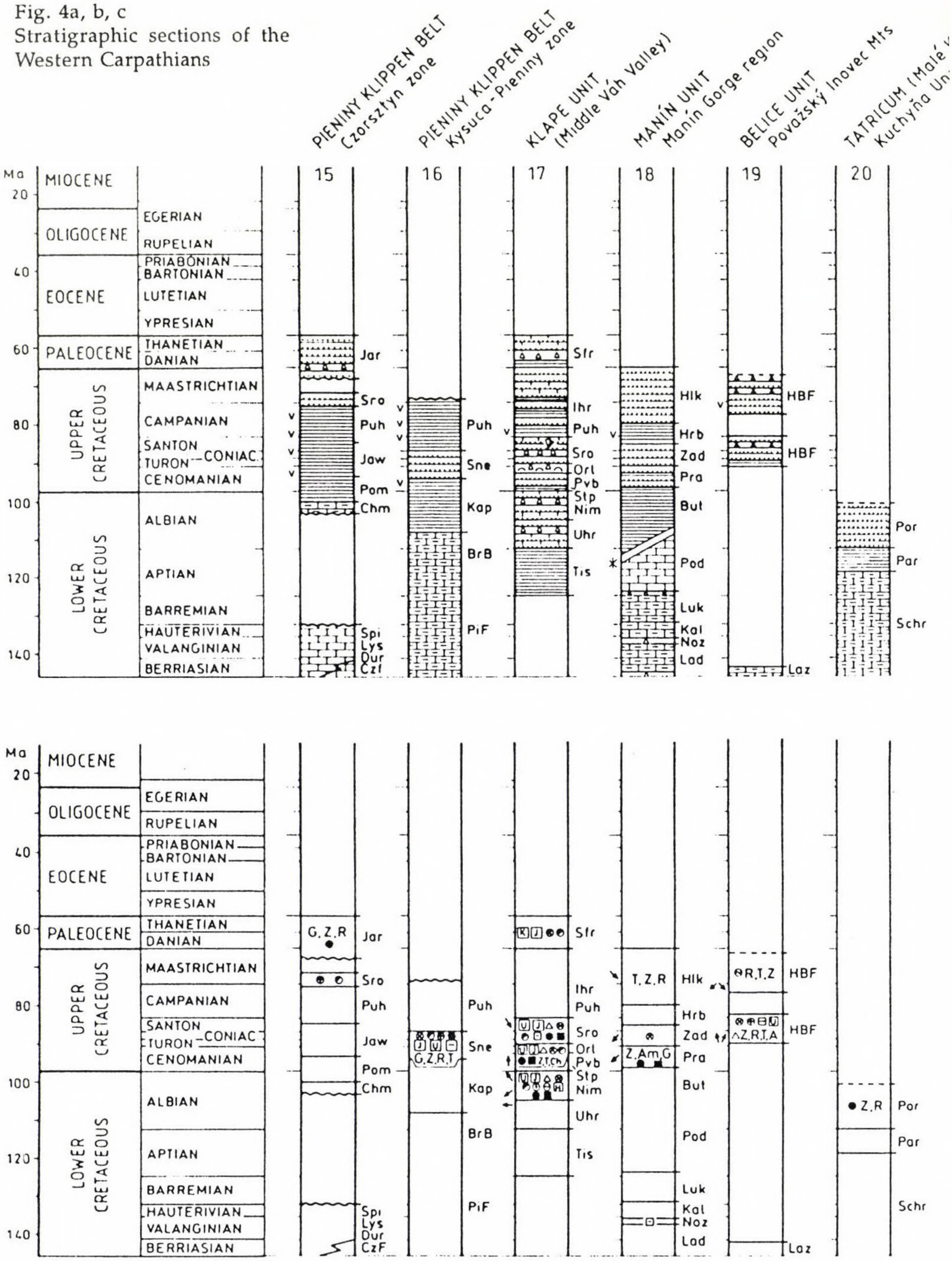


Fig. 4 b

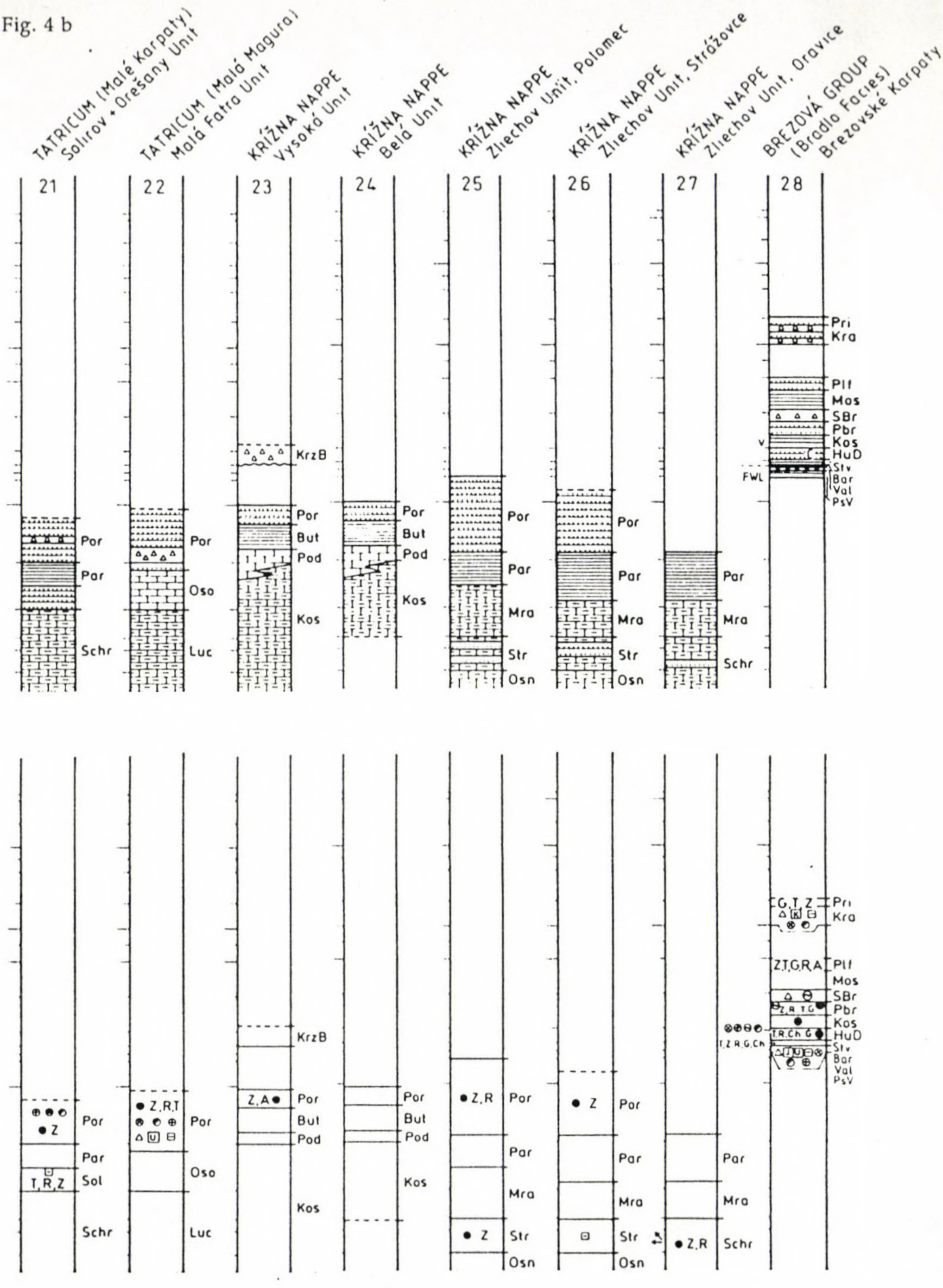

TATRICUM (Malé Karpaty) Solirov + Orešany Unit
TATRICUM (Malá Fatra Unit) Malá Magura)
KRÍŽNA NAPPE Vysoká Unit
KRÍŽNA NAPPE Belá Unit
KRÍŽNA NAPPE Zliechov Unit, Polomec
KRÍŽNA NAPPE Zliechov Unit, Strážovce
KRÍŽNA NAPPE Zliechov Unit, Oravice
BREZOVÁ GROUP (Bradlo Facies) Brezovské Karpaty
21
22
23
24
25
26
27
28
Por
Par
Schr
Sol
Oso
Luc
KrzB
But
Pod
Kos
Mra
Str
Osn
Pri
Kra
Plf
Mos
SBr
Pbr
HuD
Stv
Bar
Val
PsV
FWL
Z
T,R,Z
Z,R,T
Z,A
Z,R
G,T,Z
Z,T,G,R,A
T,R,Ch,G
T,Z,R,G,Ch

Fig. 4c

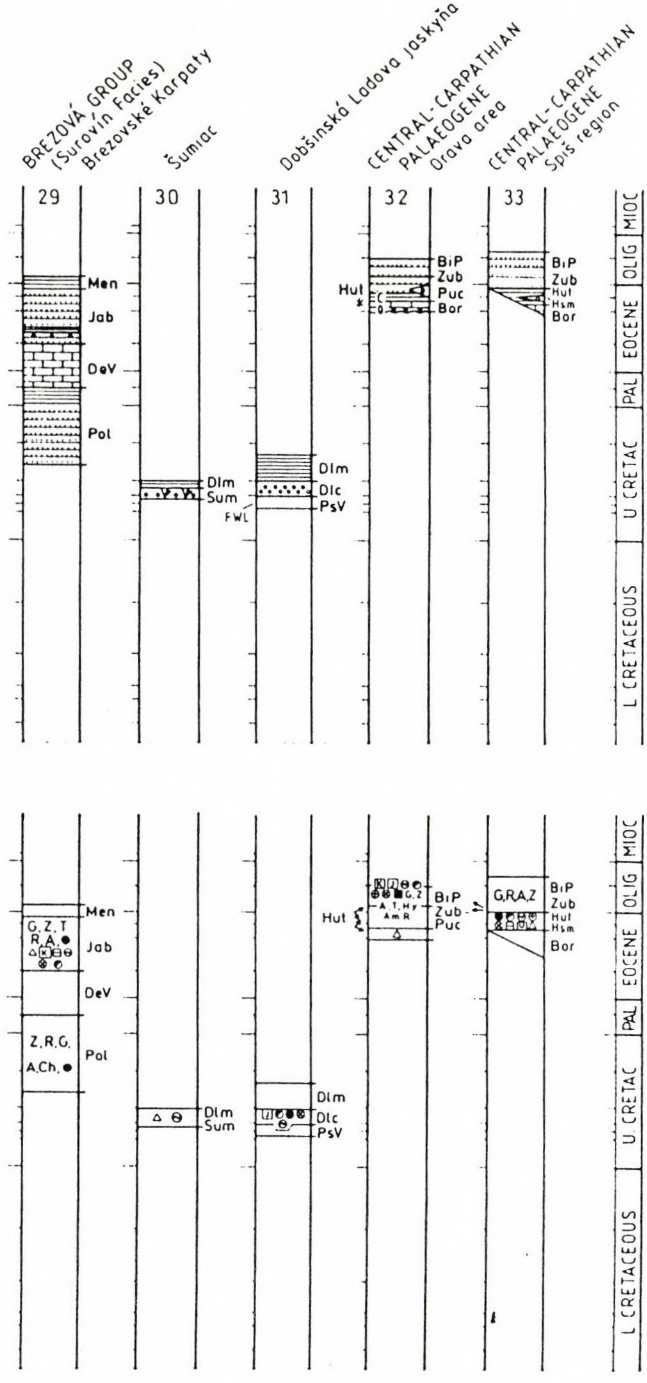

BREZOVÁ GROUP (Surovín Facies) Brezovské Karpaty
Šumiac
Dobšinská Ladova jaskyňa
CENTRAL-CARPATHIAN PALAEOGENE Orava area
CENTRAL-CARPATHIAN PALAEOGENE Spiš region
29
30
31
32
33
Men
Jab
DeV
Pol
Dlm
Sum
FWL
Dlc
PsV
Hut
BiP
Zub
Puc
Bor
Hsm
G.Z.T
R.A.
Z.R.G.
A.Ch.
G.Z
A.T.Hy
AmR
G.R.A.Z
L CRETACEOUS
U CRETAC
PAL
EOCENE
OLIG
MIOC

Fig. 5a, b
Stratigraphic sections of the North Pannonian region

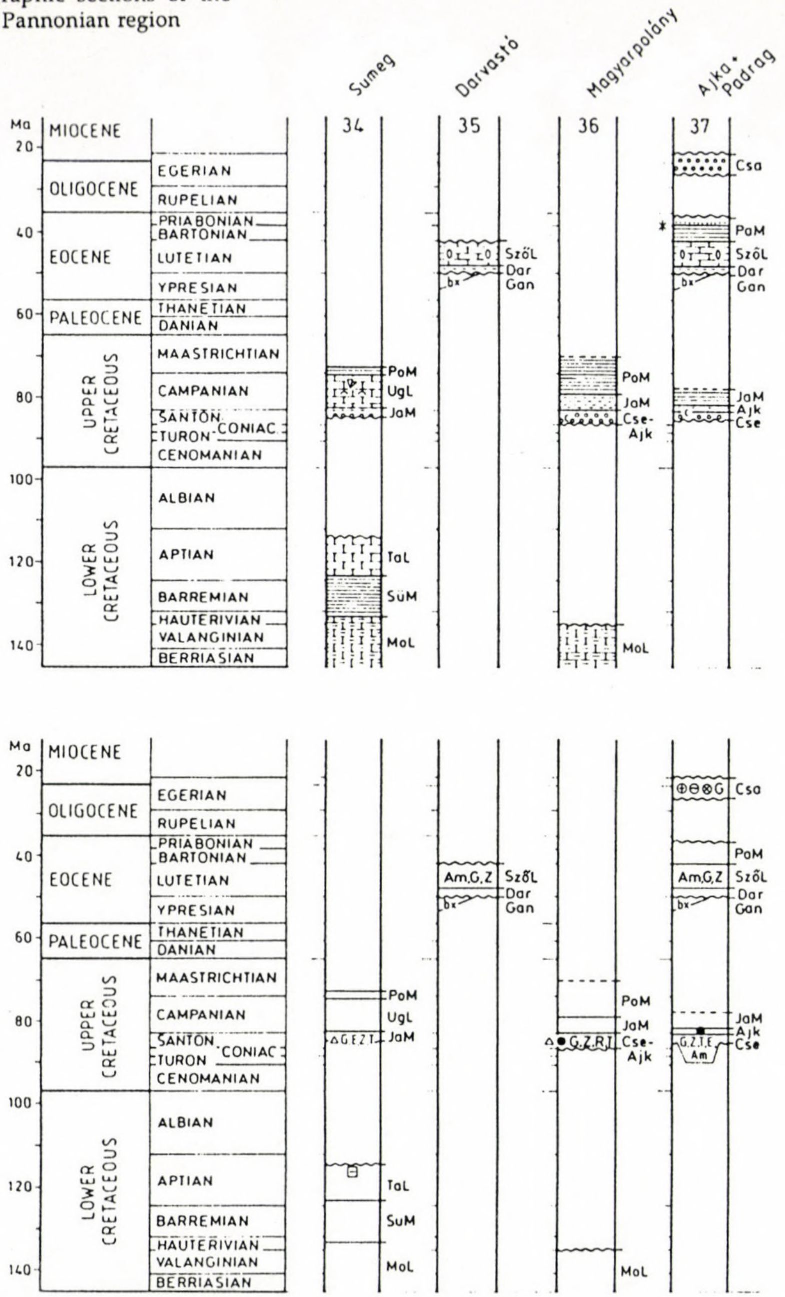
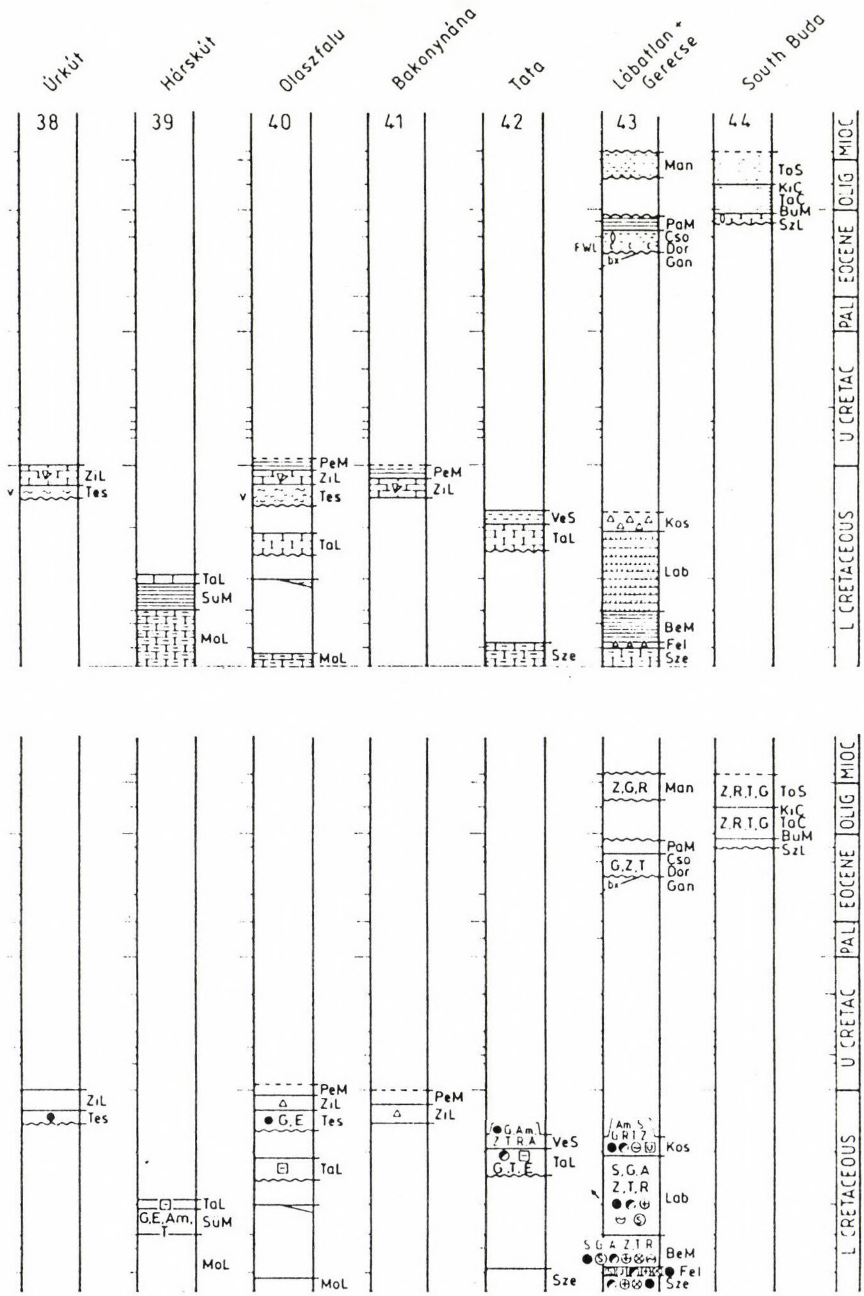


Fig. 5 b


Úrkút
Hárskút
Olaszfalu
Bakonynána
Tata
Lábatlan + Gerecse
South Buda
38
39
40
41
42
43
44
MIOC
OLIG
EOCENE
PAL
U CRETAC
L CRETACEOUS
Man
PaM
Cso
Dor
Gan
FWL
ToS
KiC
TaC
BuM
SzL
PeM
ZiL
Tes
TaL
SuM
MoL
VeS
Kos
Lab
BeM
Fel
Sze
Z.G.R
G.Z.T
Z.R.T.G
G.E
G.E.Am.
G.T.E
S.G.A
Z.T.R

The terrigenous deep-water development comprises coarse-grained facies of *deep-water breccias and conglomerates* (e.g. Spitzenbach Fm. in the NCA) and, mainly in sand grain size, the *turbiditic facies* (e.g. Rossfeld Fm., Brunnbach Fm in the NCA; Poruba Fm. in the WCa; Lábatlan Fm. in Gerecse). The latter facies is subdivided into *siliciclastic* and *carbonate-rich* turbidites on the basis of predominant clast material.

Bauxites and other residual sediments (TR and NCA) as well as *tuffs* and *tuffites* are indicated in the sections by separate symbols.

Palaeogeographically significant terrigenous material

In the highly orogenic Cretaceous and Palaeogene period, essential information can be obtained about the position and composition of palaeogeographically important regions, such as suture zones, high-pressure belts and occurrences of ophiolitic rock assemblages by studying the terrigenous detritus (Fig. 2), much of which can be classified as exotic material. Thus, several significant rock types are exhibited in the stratigraphic sections (Figs 3–5), such as granitoid rocks, metapelitic rocks (micaschists, phyllites), acidic extrusive rocks, basic/metabasic rocks and ultrabasic rocks (mainly serpentinites). In addition to these silicate rocks, some carbonate clasts seem to be highly significant palaeogeographically, such as Upper Jurassic and Lower Cretaceous shallow-water limestones, Tithonian/Neocomian calpionellid-bearing deep-water limestones, Hallstatt Limestone (Middle to Upper Triassic), Dachstein Limestone (Norian–Rhaetian), and Kambühel Limestone (Palaeocene). Exceptionally high proportions of dolomitic detritus also occur in a few sections. In most cases, these clasts are gravel-sized, although serpentinite is mainly found in sand-size fragments.

Chrome spinel and blue amphiboles in detrital heavy mineral assemblages are important indicators of the occurrence of ophiolitic complexes and high pressure/low temperature-belts in the source area and are therefore indicated by separate symbols in the sections. Both clast types mostly occur in the sand-size fraction. Among the other heavy minerals, such as those of the stable group (zircon, tourmaline, rutile) as well as garnet, apatite, chloritoid, staurolite, amphibole, hypersthene, and epidote-group minerals, only the predominant ones or those which are of particular palaeogeographic interest are documented in the sections.

To enhance the palaeogeographic significance of the sections, the available information on palaeocurrents is shown at the left side of the columns.

Lower Cretaceous

The lower parts of the Lower Cretaceous successions of the EA are characterised by a widespread deep-water limestone facies, known as the Schrambach Fm. in the Upper Austroalpine units. For the frequent occurrence of aptychus shells the term "Aptychus beds" is commonly used in the EA. This limestone facies starts normally with relatively pure, calpionellid-bearing Majolica-type limestones and passes upwards into impure marly limestones. From the Valanginian on, turbiditic layers are intercalated with these deep-water limestones to a variable extent in the NCA (Vašíček and Faupl, in prep.). In the external units, such as in the Gresten Klippen belt, these limestones are termed Blassenstein Fm., whereas in the Ybbsitz zone they are named Fasselgraben Fm.

In the WCa, as in the EA, monotonous marly limestone successions of deep-water origin are also characteristic for the bottom parts of the Lower Cretaceous (Michalík and Vašíček 1989). An exception is seen in the section from the Czorsztyn zone of the Pieniny Klippen belt, where shallow-water limestones are developed. The deep-water facies of the WCa has been subdivided into more formations than in the EA. From the sections of the Krížna nappe, intercalations with turbiditic facies are also common (Strážovce Fm.). Thick deep-water breccias have been observed in the Manín zone, and are termed Nozdrovice Breccia. In the Manín and Krížna units there are numerous small occurrences of basic volcanics.

The deep-water radiolarian, Calpionella and Nannoconus limestone (Majolica-type) facies of the TR is termed Mogyorósdomb Limestone. A similar but shallower facies without radiolarian chert (Szentivánhegy Limestone) is developed in the north-eastern localities (Tata and the Gerecse Mountains), the uppermost part of which persists into the early stage of the Early Cretaceous (Fülöp 1975).

In the EA and parts of the WCa, the deep-water limestone facies passes upwards into a bathyal marl facies, followed by turbiditic facies developments enriched with exotic material (Rossfeld Fm., Tannheim–Losenstein Fm., Lavant Fm., Párnica Fm., Poruba Fm.). This trend can also be observed in the Gerecse Mts. of the TR, where the Szentivánhegy Fm. is overlain by the bathyal slope sediment of the Bersek Marl Fm. with a breccia at its base (Felsővadács Breccia) and intercalations of graded sandstones throughout (Császár and B. Árgyelán 1994; Császár 1995). The marls grade up into the prograding turbiditic succession of the Lábatlan Sandstone and the Köszörűkbőánya Conglomerate on the higher slope.

Contrary to this trend of increasing terrigenous influx in the deep-water formation, to the SW of Gerecse (Vértes Foreland and North-Bakony) a great variety of limestones from the Upper Triassic (Dachstein Limestone) to the lowermost Cretaceous (Mogyorósdomb or Szentvánhegy Limestone) are overlain by subtidal biodetrital limestones with extraclasts (Tata Limestone), restricted lagoonal marls (Vértessomló Siltstone) and shallow-water platform carbonates, the so-called Urgonian facies (Környe and Zirc Limestone) of Aptian and Albian age. The two Urgonian sequences, both with a retrograding trend, are separated from each other by a fluvial to marine clastic formation (Tés Clay – Császár 1995). The lower formation (Környe Limestone) is interfingered with the subtidal to shallow bathyal Vértessomló Siltstone and with the Tés Clay Fm.

The development of shallow-water carbonates is also preserved in some parts of the WCa, such as in the Manín zone (Podhorie Fm.), the High Tatric unit of the Tatricum (Osobitá Fm.) and in parts of the Krížna nappe (Muráň Limestone in the High Tatras; Belá unit; Podhorie Fm.). The shallow-water biodetritus is present in fluxoturbitites and calciturbidite intercalations (Mišík 1990).

In situ Urgonian facies has not been observed in the Austroalpine units of the EA, but redeposited Urgonian clasts have been repeatedly reported from Lower and Upper Cretaceous as well as Tertiary deposits (Gaupp 1980; Hagn 1982, 1983; Weidich 1984; Schlagintweit 1987; Wagreich and Schlagintweit 1990); it is thought therefore that such carbonate platform sediments also accumulated within or in the near vicinity of the Austroalpine realm and were later totally eroded. A thick Urgonian succession (Schrattenkalk) is developed and preserved only in the western part of the Helvetic zone of the EA (Heim and Baumberger 1933; Császár et al. 1994).

In some cases, from the Late Albian on the Urgonian platforms subsided rapidly (Császár 1995) and were overlain by bathyal marls and turbiditic formations, such as the Butkov Fm. and Poruba Fm. of the WCa and the Pénzeskút Marl Fm. of the TR. It is considered that this subsidence pulse was a direct response to the mid-Cretaceous tectonic movements (Austrian phase post-collisional subsidence).

The terrigenous material of the Lower Cretaceous deposits was derived from two major source areas (Faupl 1990; Faupl and Wagreich 1992). One was situated in front of the Austroalpine realm and the Tatricum, whilst the other corresponds to the Tethys suture zone located within the Austroalpine realm.

The northern provenance area was built up by an accretionary wedge consisting of rocks of continental origin derived from the deforming Austroalpine margin and related units as well as from oceanic complexes. The oceanic series originated from obducted bodies of the Penninic–Ligurian Ocean associated with a subduction phase during the Early Cretaceous. Ophiolitic detritus is mainly represented by chrome spinel (Pober and Faupl 1988). The existence of high pressure/low temperature rocks within this accretionary wedge is proved by the occurrence of blueschist fragments in the Klape and Manín units of the WCa, in the Albian Poruba Fm. of the Humenné Mts. and also those redeposited in the Palaeogene sequence of Orava (Zoubek 1931; Mišík 1979; Mišík and Sýkora 1981; Mišík and Marschalko 1988; Ivan and Sýkora, 1993; Dal Piaz et al. 1995) and by detrital blue amphiboles and a few discoveries of lawsonite in the northernmost tectonic units of the NCA (Woletz 1970; Winkler and Bernoulli 1986; Faupl and Wagreich 1992). The Urgonian platform, which is preserved in several Tatric units, also seems to have served as a source area, but pebbles of Urgonian limestone with ophiolitic detritus are of exotic origin (Mišík 1990). From redeposited pebbles in the Losenstein

Fm. of the EA, it is known that such shallow-water carbonates occurred on the accretionary wedge (Wagreich and Schlagintweit 1990).

The existence of an intra-Austroalpine suture zone situated to the south of the NCA and acting as a source terrain during the Cretaceous and Palaeogene can also be deduced from Early Cretaceous clastic material. Within the NCA, the Rossfeld Fm. contains high amounts of detrital chrome spinel, accompanied by metamorphic heavy mineral associations derived from this suture zone (Faupl and Tollmann 1979; Decker et al. 1987). An analysis of Upper Cretaceous conglomerates in the NCA demonstrates that Urgonian carbonate successions were also deposited in parts of the NCA, but were later totally eroded. Pebbles of Urgonian-type limestones bear extremely high amounts of chrome spinel with relatively high Cr-contents, and it is believed that they were also derived from this intra-Austroalpine suture zone (Wagreich et al. 1995). Comparable chrome spinel-bearing pebbles have also been reported from the WCa by Mišík (1979), Mišík et al. (1980) and Mišík and Sýkora (1981). In the TR, the ophiolitic detritus of the Gerecse Mountains, which is also accompanied by metamorphic heavy minerals, points to the same source area. There the massive occurrence of ultrabasic and basic rock fragments (also associated with high-Cr-bearing spinel in the heavy mineral fractions – Császár and B. Árgyelán 1994) can be explained by a more proximal position of the Gerecse basin (Sztanó 1990) to this suture zone than the Rossfeld Fm of the NCA. The chemistry of detrital chrome spinels from the Rossfeld Fm. (Faupl and Pober 1991) and from the Gerecse section (Árgyelán 1996) are highly comparable. Detritus of the Urgonian-type limestones in the Gerecse Mountains was derived from the margin of the Tethys-Vardar Ocean (Császár and B.Árgyelán 1994; Császár 1995). The Tethys suture zone probably also supplied ophiolitic detritus to the Lower Cretaceous deposits of the WCa, as found in several units of the Krížna and Choč nappes (Mišík et al. 1980; Jablonský 1992), although no clear indications are available from palaeocurrent data.

Upper Cretaceous

In some parts of the tectonically lower nappe system of the NCA (Frankenfels and Ternberg nappe) the turbiditic facies of the Losenstein Fm. persisted into the Cenomanian. An analogous trend can be observed in parts of the Krížna nappe of the WCa, where the turbiditic Poruba Fm., with a comparable facies, also continued into the Cenomanian. A similar situation is known from the TR, where the carbonate platform was drowned in the Late Albian, with silty marls deposited in the Late Albian to Early Cenomanian and sandstones in the Middle Cenomanian. The Lábatlan Sandstone in the Gerecse Mountains is also supposed to extend into the Cenomanian, in the same palaeoenvironment (Császár 1995). The sequence of the Pénzeskút Marl spread over the Zirc Limestone in the Bakony Mountains in the Late Albian and the deposition of the coarsening-upward argillaceous-siliciclastic sedimentation continued at least until the Middle Cenomanian.

The sedimentary development of the Branderfleck Fm. of the NCA, resting unconformably upon external parts of the Lunz–Reichraming–Lechtal nappe system (Gaupp 1980, 1982; Weidich 1984, Faupl and Wagreich 1992) has no equivalent in the WCa or in the TR. While the Branderfleck Fm. of the western NCA represents a clastic deep-water facies, it is developed as a grey terrigenous shelf facies rich in ammonites (Summesberger 1992), inoceramids, and orbitolinid foraminifera in the middle and eastern parts.

In a particular way, the clastic mid-Cretaceous deep-water facies reflects the geodynamic activities of the Eoalpine stage. In the entire Austroalpine realm, including both the central WCa tectonic units and the TR, sedimentation was interrupted during the mid-Cretaceous as a result of Eoalpine folding and nappe stacking. Only in the "Cenoman-Randschuppe" of the western NCA did the deep-water sedimentation continue without interruption up to the Santonian/Campanian (Weidich 1984).

In the external units of the EA such as the Helvetic–Penninic domains (e.g. Gresten Klippen belt, Ybbsitz zone), and in the Manín and Klape units of the WCa as well as in the Pieniny Klippen belt, bathyal or turbiditic sedimentation passed continuously from the Early into the Late Cretaceous without major gaps. In the Klape unit an exception has been described, where oyster bioherms are embedded within a clastic deep-water sequence (Marschalko and Kysela 1980).

In the Turonian, a new sedimentary succession termed Gosau Group in the NCA (e.g. Wagreich and Faupl 1994) accumulated unconformably upon the deformed pre-Turonian successions. In the equivalent Brezová Group in the WCa sedimentation began in the Senonian (Samuel et al. 1980; Salaj and Began 1983). The Brezová Group is subdivided into the Bradlo and Surovín facies. The basal part of all these deposits consists of an alluvial to fan-delta facies (Kreuzgraben and Streiteck Fm., Valchov Conglomerate). In some places, freshwater limestones (e.g. Pustá Ves Fm.) and small bauxite deposits have been observed at the base and coal seams have been found within the transitional facies of alluvial to fan-delta environment. The alluvial to fan-delta facies passed into a shallow-marine terrigenous facies which also includes thin beds of rudist bioherms and related sediments. This shallow-marine development is overlain by a mostly reddish bathyal marl facies (e.g. Nierental Fm., Košariská Fm.) or grey marls such as the Štverník Marls, Mosnáčov Marls associated with deep-water breccias (Spitzenbach Fm., Široké Bradlo Limestone) and by a turbiditic facies (e.g. Ressen Fm., Brunnbach Fm., Höllgraben Fm., Zwieselalm Fm., Podbradlianska Fm., Podlipovec Flysch, Polianka Fm.), both deep-water developments demonstrating the rapid deepening of the basin. In many localities of the NCA, a conspicuous unconfomity developed between the Lower Gosau Subgroup, comprising the terrestrial to shallow-marine deposits, and the deep-water sediments of the

Upper Gosau Subgroup. This unconformity does not appear to occur in the Brezová Group of the WCa.

The sedimentation of the Lower Gosau Group appears to have been tectonically controlled by the oblique subduction of the Penninic–Ligurian Ocean below the Austroalpine margin. Deposition took place in a transtensional crustal regime behind the accretionary wedge, a palaeogeographic feature mentioned previously, lying in the frontal part of the Austroalpine domain since the mid-Cretaceous. The diachronous onset of the deep-water sedimentation of the Upper Gosau Group began in the north-western parts of the NCA and migrated towards the south-east. It is presumed that the new subsidence pulse of the Upper Gosau Group was triggered by a collisional event causing subcrustal tectonic erosion, during which the accretionary wedge was removed. In the sedimentary column this event is documented by deformation and erosion of parts of the Lower Gosau Group and by the discontinuous onset of deep-water sediments.

The so-called Central-Alpine Gosau deposits of the EA, resting unconformably upon Upper Austroalpine units such as the Palaeozoic of Graz (Kainach Gosau) and the Gurktal Nappe (Krappfeld Gosau, Gosau of St. Paul/Lavanttal), have an equivalent development in the western parts of the TR (Bakony Mts). In both cases, sedimentation commenced in the Santonian and comprises a terrestrial to shallow-marine lower part and an upper part of deep-water sediments. In the Central-Alpine Gosau of the Krappfeld the lower series, including extensive deposits of rudist limestones, were widely eroded and redeposited into the deep-water environment. In contrast, in the TR, fluvial-paludal (Csehbánya Fm. – Jocha-Edelényi 1988) and pit-bog facies (Ajka Coal Fm. – Haas et al. 1986; Császár et al. 1993) deposits are preserved to a much greater extent, where they developed in troughs. At the base of the Hungarian successions bauxites locally accumulated in karst depressions (Halimba Bauxite Fm.), not only in the troughs but also high on the slope. The non-marine succession is overlain by a shallow-marine marl (Jákó Marl Fm.), while the topographic highs separating the troughs are covered directly by Campanian rudistid platform carbonates (Ugod Limestone Fm.), replaced laterally and upward by shallow bathyal marls (Polány Marl Fm.). The lithoclastic (debrite) bodies (Jákóhegy Breccia Mb.) interfingering with the Polány Marl were derived from the carbonate platform and moved onto the slope a few kilometres away from the source (Haas 1979, 1983). The upper part of the Polány Marl turns into a coarsening-upward siliciclastic marl, resembling the Pénzeskút Marl Fm. Locally (Tapolcafő), the carbonate platform was raised above sea-level and karstified just prior to the drowning of the platform. There the Polány Marl shows an coarsening-upward trend starting with red clays.

A comparable development of Upper Cretaceous deposits appears to be missing in the WCa, although the successions of Šumiac and Dobšinska Ladova

Jaskyna could be interpreted as small remnants of the lower parts of the successions described from the EA and the TR.

One of the major differences between the Hungarian Senonian and the Central-Alpine Gosau of the EA is found in the rock and mineral composition of the terrigenous material. Whereas the Hungarian deposits are characterised by the occurrence of chrome spinel in the heavy mineral associations (B. Árgyelán, personal communication), this ophiolitic detritus is totally missing in Upper Cretaceous successions of the Central-Alpine Gosau. Furthermore, the facies development of the Central-Alpine Gosau and the Senonian of the TR differs considerably from those of the Gosau in the NCA and Brezová Group in the WCa (Oberhauser 1963).

The terrigenous material in the Upper Cretaceous formations was derived from the same two major source terrains as recorded in the Lower Cretaceous: the intra-orogenic Tethys suture zone to the south and the accretionary wedge in front of the Austroalpine domain to the north. The material from the accretionary wedge which supplied the Klape and the Manín units of the WCa is characterised by high proportions of ophiolitic detritus and high-pressure rock fragments, as well as by manifold suites of carbonate clasts such as Upper Jurassic and Lower Cretaceous (Urgonian) shallow-water limestones and Tithonian/Neocomian deep-water limestones. This accretionary wedge was also active in supplying chrome spinel-rich detritus and exotic pebbles to the Lower Gosau Group of the NCA (Wagreich and Faupl 1994), the Brezová Group (Wagreich and Marschalko 1995) and the unit joining them covered by Neogene sediments in the Slovakian sector of the Vienna Basin (Mišík 1994).

With the onset of subcrustal tectonic erosion which removed the accretionary wedge, intra-orogenic source areas including the Tethys suture zone began to play a more important role. As a consequence, ophiolitic detritus decreased considerably in the WCa and EA and only a local supply of chrome spinel from the Tethys suture zone can be observed. In the Gosau Group of the NCA, rock fragments, especially light micas derived from the Eoalpine metamorphosed Austroalpine crystalline basement, were deposited from Late Campanian time on (Pober 1984). The ophiolitic detritus of the Upper Cretaceous of the TR was supplied exclusively from the Tethys suture zone.

Palaeogene

In the Gosau Group of the EA, the deposition of bathyal marls and turbidites continued from the Upper Cretaceous to the Eocene. The same trend has been observed in the Gosau-type Brezová Group of the WCa. However, the shallow-water limestone facies of the Dedkov Vrch Fm. (Paleocene/Eocene) in the Surovín facies of the WCa, which forms an intercalation within the deep-water development, is an exception. This formation possibly represents redeposited shallow-water material in a proximal deep-water environment, as also suggested for the Kambühel Limestone (Lower Paleocene) at the type locality in Lower Austria. The entire succession of the Surovín facies persists up to the Rupelian. In these facies the occurrence of detrital chrome spinel is reported from the turbiditic Polianka Fm. (Maastrichtian) as well as from the deep-water deposits of the Jablonka Fm. (Eocene). Similar occurrences of ophiolitic detritus have also been reported from several turbiditic formations of the Upper Gosau Group of the NCA (e.g. Gießhübel Fm. – Sauer 1980; Brunnbach Fm. – Faupl 1983) which proves that the Tethys Suture zone was active as a source terrain during this late stage of sedimentation.

In the external tectonic units of the EA, described in this paper, a deep-water development is characteristic for Paleocene–Eocene times, such as in the Gresten Klippen belt of the EA (Buntmergelserie with deep-water conglomerates) as well as in the Pieniny Klippen belt and Klape unit of the WCa.

In the Krappfeld area of the EA, a succession of shallow-marine terrigenous facies with coal-bearing beds at its base lies unconformably on Central-Alpine Upper Cretaceous Gosau deposits (Wilkens 1989). These terrigenous formations pass up into a highly fossiliferous shallow-marine limestone facies with nummulite beds. This transgressive sequence as a whole forms a new sedimentary cycle and is here summarised under the stratigraphic term "Guttaring Group", which has been subdivided into three formations and several members by Wilkens (1991). Contrary to the comparable Hungarian deposits, which are free of detrital chrome spinel, ophiolitic detritus has been observed in sandstones of the Guttaring Group.

A very small occurrence of shallow-water limestones (Priabonian) from Kirchberg/Wechsel, resting upon the crystalline basement complex of the Lower Austroalpine unit, is not presented among the sections of the EA. It appears to be similar to the Szépvölgy Limestone in the Buda Hills of Hungary (Kázmér 1985a). The Upper Eocene–Oligocene successions of the "Unterinntal-Tertiär" (NCA) are not discussed in this paper.

In the TR, a comparable transgressive development of shallow-marine deposits commenced in the Middle Eocene, or later than in the EA. At the base, well-known large bauxite deposits (Gánt Bauxite Fm.) are widely distributed. In western parts of the TR (Darvastó at Nyirád and Padrag) the shallow-marine terrigenous facies is associated with coal seams and passes into nummulitic limestones (Szőc Limestone Fm.). The entire succession shows a deepening-upward trend with a bathyal facies at the top (Padrag Marl Fm.). In the troughs between the North Bakony and the Dorog Basin, with a varied morphology similar to the Upper Cretaceous one, sedimentation began with terrestrial sediments, including thick coal measures, followed by marls and siltstones. The topographic highs were covered directly by nummulitic and biodetrital limestone (Szőc Limestone Fm.) of Late Lutetian age. To the east, in the South Buda Hills, the transgression started later (in the Priabonian) with the Szépvölgy Limestone Fm. which also passed into a bathyal facies at the end of the Eocene. The Oligocene to the east of Buda developed continuously from the bathyal Buda Marl Fm. and is represented by fine-grained siliciclastics (Fodor et al. 1994). As far as to the west of Buda it forms a new sedimentary cycle, which began with variegated alluvial fan deposits (Korpás 1981). The Egerian transgression occurred with a shallow-marine terrigenous facies (Mány Fm.) towards the east and variegated fluvial sediments to the west (Csatka Fm.). In the South Buda Hills, the Törökbálint Sandstone Fm. directly overlies the bathyal facies of the Kiscell Clay Fm. (Rupelian) without any unconformity.

In the WCa, the transgression from the region of the Flysch Belt into the Peri-Klippen Belt area, which contains blocks of the biohermal Kambühel Limestone, had already begun in the Paleocene. In the eastern part of the Pieniny Klippen Belt, the Paleocene Proc Conglomerates include almost the same inventory of exotic rocks as the Cretaceous conglomerates of the western part of the Klippen and Peri-Klippen zones. These exotic components could not have been redeposited from the Cretaceous conglomerates because such a facies is totally absent in the eastern sector.

The transgression of the "Central-Carpathian Palaeogene" (Gross et al. 1993) commenced in the Middle Eocene, starting with the shallow-water deposits of the Borové Fm. These deposits pass rapidly via a bathyal marl facies (Huty Fm.) into the thick Central-Carpathian Flysch sequences of the Zuberec and Biely Potok Fms. The palaeocurrents show various directions, but besides local currents a general trend towards the east is characteristic for the central and eastern parts of this basin (Gross et al. 1993). From the Spiš region, ophiolitic detritus has been reported in these formations (intercalation of the "serpentinite sandstone" within the Šambron zone – Soták and Bebej 1996). In the Orava area, the occurrence of glaucophane-bearing pebbles, probably redeposited from the Cretaceous conglomerates of the Klippen Belt, is of special interest. The rapid subsidence of the Central-Carpathian Flysch basin is interpreted to have been caused by subcrustal tectonic erosion which seems to continue in the WCa for a longer period than in the EA.

Conclusions

The sedimentary history of the three segments of the Alpine orogenic belt, the Eastern Alps, the Western Carpathians and the Transdanubian Range, is predominantly similar. This similarity can be summarised as follows:

1) In the Early Cretaceous, deep-water pelagic limestones with turbiditic intercalations in them are typical for almost the entire area concerned: Blassenstein Fm. (Gresten Klippen Zone), Fasselgraben Fm. (Ybbsitz Zone), Schrambach Fm. (NCA), "Calpionellenkalk" (N Karawanken, not in the presented sections), Nozdrovice Fm. (Manín unit), Strážovce Fm. (Krížna nappe) and Szentivánhegy Limestone–Bersek Marl Fm. with the Felsővadács Breccia Mb. (TR).

2) In the Valanginian the pelagic limestones were replaced first by turbiditic marl deposits (lower part of the Rossfeld Fm. (NCA) and Bersek Marl Fm. (TR)) and then by coarser grained clastics (upper part of the Rossfeld Fm. (NCA), intercalations in the Stražovce Fm. (loc. Oravice) and Lábatlan Sandstone Fm., Köszörűkőbánya Conglomerate (TR)). The coarsening-upward tendency is conspicuous in all three regions.

3) In some zones the sedimentation of bathyal marls continued or only began in the Aptian and/or Albian: Tannheim Fm. in the NCA, lower member of the Lavant Fm. (Drau Range), Párnica Fm., Butkov Fm., Tissalo Beds and Kapusnica Fm. in the WCa, and Vértessomló Siltstone Fm. in the TR. In the EA and WCa, this bathyal marl facies marks the onset of further synorogenic clastic deep-water sedimentation (e.g. Losenstein–Branderfleck Fm., Lavant Fm., Poruba Fm.).

4) Due to the overall tectonic activity, or as a consequence of a shallowing-upward trend, in some areas platform carbonates of Urgonian facies developed: In the WCa the Podhorie Fm. (Manín unit, Krížna nappe) and the Osobitá Fm. (Tatricum, High Tatras); in the TR the Zirc and Környe Limestones (Vértes Foreland, not in the presented sections). These platforms are mainly prograding. In the EA Urgonian carbonate successions are only preserved in the western parts of the Helvetic zone *s. str.* (e.g. Schrattenkalk, not in the presented sections). However, exotic Urgonian pebbles from the EA and WCa are evidence for the existence of carbonate platforms to the north on the accretionary wedge forming the leading edge of the Austroalpine–Central Carpathian margin, as well as to the south at the active margin of the Tethys-Vardar Ocean. Urgonian-type olistoliths, found in the Köszörűkőbánya Conglomerate Mb. (Gerecse Mts.), were supplied from this southern source terrain.

5) The accretionary wedge at the northern active margin of the Austroalpine–Central Carpathian complex and the Tethys suture zone in the south seem to have been the two major sources of exotic clasts, including ophiolitic detritus from the Cretaceous up to the Palaeogene.

6) The carbonate platforms were drowned in the (Late) Albian as evidenced by the Butkov and Poruba Fm. (WCa) and the Pénzeskút Marl Fm. (TR).

7) Except in external tectonic units (Gresten Klippen zone, Ybbsitz zone, "Cenoman Randschuppe", Klape and Manín units), sedimentation was interrupted in the entire Austroalpine realm in mid-Cretaceous time, followed by an extensive erosion prior to or during the Turonian.

8) The Upper Cretaceous sedimentary cycle began with alluvial and fan-delta facies in the NCA and the WCa in the Turonian and in the central parts of the EA and the TR in the Santonian: Kreuzgraben and Streiteck Fm. (NCA), a part of the Valchov Conglomerate (WCa) and Csehbánya Fm. (TR). It was followed first by shallow-marine successions (e.g. Grabenbach and Hochmoos Fm., NCA; Baranec Sandstone, WCa; Jákó Marl Fm., Bakony Mts.) and then by bathyal facies (e.g. Nierental Fm., NCA; Štvernik Marl etc., WCa; Zementmergelfolge,

Central Alps; Polány Marl, Bakony Mts.). These marls represent predominantly deep bathyal facies with turbiditic intercalations in the NCA and WCa, whereas in the Central-Alpine Gosau and in the TR a more shallow bathyal facies with slope breccias is developed.

9) Late Cretaceous rudistid bioherms are distributed in a northern zone (NCA and WCa) while platform carbonates were formed in central parts of the EA and the TR. Its continuation is supposed to be developed in the Dobšiná area of the WCa (loc. Šumiac).

10) Continuous deep-water sedimentation up to the Palaeogene is reported from the Helvetic zone *s. l.* (Gresten Klippen zone), the Pieniny Klippen Belt and the Klape unit as well as from the Gosau and Brezová Group. A gap is documented in the Central-Alpine domain and in the TR. The terrigenous coal-bearing sedimentation in the Krappfeld area commenced in the Early Eocene and in the TR and the Central-Carpathian Palaeogene in the Middle Eocene. In all these places the succession continued either with shallow-marine nummulitic limestones (Szőc Limestone, TR; Borové Fm., WCa) or directly with pelagic marls (Padrag Marl Fm., TR). Clastic sedimentation continued with shallow bathyal facies east of Buda and as flysch in the Central-Carpathian Palaeogene.

In spite of local and temporal differences, the main events of the Cretaceous–Palaeogene history of the study area basically correlate with each other, especially within the NCA and the WCa as well as in the TR and the Central-Alpine part of the EA. However, the facies and time differences of the Upper Cretaceous and Palaeogene sequences between the Central-Alpine domain of the EA and the TR are clear evidence that the shift of the TR towards the east, away from its Permian to Early Cretaceous position, began in the mid-Cretaceous and lasted until the Late Cretaceous and Palaeogene.

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Triassic sequence stratigraphy of the Balaton Highland, Hungary

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Sequence stratigraphic analysis of Triassic formations in the Balaton Highland region revealed that in addition to sea-level-changes, climatic changes and tectonic effects also played an important role in the determination of the facies characteristics as well as the setting and features of the depositional sequences. However, the relative importance of these factors differed in the successive evolutionary stages. In the Early Triassic the moderately and uniformly subsiding shelf was very sensitive to sea-level changes. During the Early to Middle Anisian, mainly the effects of climatic changes are detectable. A drastic reduction of terrigenous input at the beginning of this stage can be attributed to a climatic change and it is primarily climatic conditions which may have determined whether syndiagenetic dolomite formation or organic rich lime mud deposition prevailed on the restricted inner ramp. From the Middle Anisian to the Late Carnian tectonic movements played the most decisive role. At the beginning of this stage segmentation of the shelf began, resulting in the differentiation of platforms and basins. Sea level changes manifested themselves in subaerial exposure and inundation of the platforms. Filling up of the basins began in the Carnian, when most probably due to a remarkable climatic change terrigenous influx increased significantly. During this period eustatic sea level changes may have played an important role in the determination of the sedimentation pattern.

Key words: sequence stratigraphy, facies analysis, paleoclimate, Triassic, Transdanubian Range

Introduction

Classic studies on stratigraphy and palaeogeography provided the fundaments for the sequence stratigraphy of the South Alpine region (Leonardi 1968; Pisa 1972, 1974; Assereto et al. 1977; Gaetani ed. 1979; Viel 1979; Pisa et al. 1980; Gaetani et al. 1986; Pasini et al. 1986; De Zanche and Farabegoli 1988, etc.). Summarizing papers were published on the relationships of the carbonate platforms and related basins, describing the basic pattern of their geometry and the rules of their evolution (Bosselini and Rossi 1974; Gaetani et al. 1981; Bosellini 1984; Brandner 1984; Wendt 1986; Doglioni et al. 1990, etc.).

The South Alpine Triassic series already played an important role in the elaboration of the global coastal onlap curves and sea-level charts (Haq et al. 1988). Since the beginning of the 90s, the significance of the sequence stratigraphic approach in stratigraphic correlation and in the interpretation of

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the evolutionary history in the Alpine region has increased remarkably, primarily in the Dolomites (Brandner 1984; Doglioni et al. 1990; Bechstädt and Schweizer 1991; Bosellini 1991; Bosellini and Neri 1991; Bosellini and Stefani 1991; Schlager et al. 1991; De Zanche et al. 1992a, 1992b, 1993, 1995; Neri 1991, etc.). Practically contemporaneously a detailed sequence stratigraphic analysis was also carried out for the Germanic Basin (Aigner and Bachmann 1992).

According to the classic sequence stratigraphic concept "stratigraphic signatures result from the interaction of tectonic, eustatic, sedimentary, and climatic processes" (Vail et al. 1991). Tectonic and eustatic processes determine the space available for sediment accumulation (accommodation space), while tectonics and climate control the quantity and quality of sediments. Due to significant differences between the mechanisms of sediment accumulation processes of siliciclastic and carbonate depositional systems their sequence stratigraphic evaluation has also been differentiated (Kendall and Schlager 1981; Sarg 1988; Vail et al. 1991; Schlager 1991, 1992).

In many cases, authors overemphasize the role of one or another factor. The orthodox sequence stratigraphers are apt to stress the predominance of the eustatic sea-level changes. In reality local or regional tectonic events may overprint the effects of global sea-level changes and climatic changes may also mask the effects of the sea-level variations. Thus every possible factor must be taken into account in the sequence analysis to determine the cause of the facies changes or to explain the nature of the sequence boundaries. On the other hand the correlation of the sequences or the sequence boundaries between the different depositional environments is not simple, either. For example, the sequence boundaries are well visible on the top of the carbonate platforms reflecting sea-level drops after the highstand platform progradations. However, due to poor biostratigraphic data, correlation of these boundaries is generally difficult. On the other hand, in the pelagic basins biostratigraphy is much more sophisticated as a rule, but recognition of the sequence-boundaries is more difficult and frequently ambiguous.

Previous studies

The present-day lithostratigraphic subdivision of the Triassic in the Balaton Highland (Haas and Császár ed. 1993) has been worked out on the basis of the latest regional geologic mapping project (1982–91) and the key-section project, running concurrently. At the same time a detailed biostratigraphic subdivision has been elaborated showing fairly good correlation with the international standard orthostratigraphic scale. In addition to the ammonite zonation it also includes zonal subdivisions of many other fossil groups, and their correlation with each other (Szabó et al. 1980; Vörös 1987; Kovács et al. 1990, 1991; Dosztály 1993; Kovács 1993a, 1993b; Góczán and Oravecz-Scheffer 1993; Vörös 1993; Kovács et al. 1994; Vörös et al. 1996). The biostratigraphic correlation between the Balaton Highland and the Southern Alps is more or less satisfying, in spite of the current debates around the Anisian/Ladinian boundary (Gaetani ed. 1993; Brack and Rieber 1993, 1994; De Zanche and Gianolla 1995; Manfrin and Mietto 1995; Mietto and Manfrin 1995; Vörös et al. 1996).

From the sedimentological point of view, the Lower Triassic formations were studied most comprehensively (Haas et al. 1988). Based on these studies and also taking into consideration the basic principles of sequence stratigraphy, a detailed facies analysis was made by Broglio Loriga et al. (1990).

Palaeogeographic and geohistoric analyses were carried out for the Early Triassic (Haas et al. 1988; Broglio Loriga et al. 1990), for the Middle Triassic (Budai and Vörös 1992, 1993; Budai et al. 1993; Vörös 1996; Vörös et al., in press), for various stages of the Upper Triassic (Haas 1988, 1993, 1994) and also for the entire Triassic (Haas et al. 1995; Haas and Budai 1995). Based on these studies a large- scale paleogeographic reconstruction of the Balaton Highland for the Late Triassic is shown in Fig. 1, and the paleogeographic evolution of the studied region is displayed in Fig. 2.

Recent investigations of the Triassic of the Balaton Highland created the fundamentals of a sequence stratigraphic analysis: detailed lithostratigraphy, a suitable biostratigraphic scale (orthostratigraphic and parastratigraphic zonation), sedimentological data and facies interpretations, and paleo-geographic and geohistoric syntheses, are available. However, because of the unfavourable exposure conditions, the sequence stratigraphic analysis can hardly be accomplished without comparing the successions of the Balaton Highland with others, in better exposed regions. Due to its close paleogeographic relationship with the Balaton Highland region and a well established sequence stratigraphy, the Dolomites seem to be convenient for comparison. However, local or regional effects (e.g. tectonic instability of both areas in the Middle Triassic) make the comparison and the sequence analysis more difficult. To overcome this problem, sequence stratigraphy established in the epicontinental Germanic Basin was also considered (Aigner and Bachmann 1992) although due to provincialism of the biota and the incompleteness of the biostratigraphic scale in the Germanic Basin, its biostratigraphic correlation with the Alpine regions is often ambiguous.

Stratigraphic chart and principles of the sequence analysis

A generalized lithologic column of the Balaton Highland is presented in Fig. 3a. The chronostratigraphic scale is given after Gradstein et al. (1994). The correlation of our successions with the standard chronostratigraphic scale is based on biozonations which have been worked out by Vörös (1987, 1993), Vörös et al. (1996, in press), Dosztály (1993), Kovács (1993a, 1993b), Kovács et al. (1990, 1991, 1994), and Góczán and Oravecz-Scheffer (1993). A facies curve and diagrams showing the vertical distribution of the terrigenous and volcanogenic components respectively, and also the major tectonic events, are given. Based on the amount of terrigenous components in the sedimentary rocks, of traces of evaporite formation, of the intensity of early dolomitization

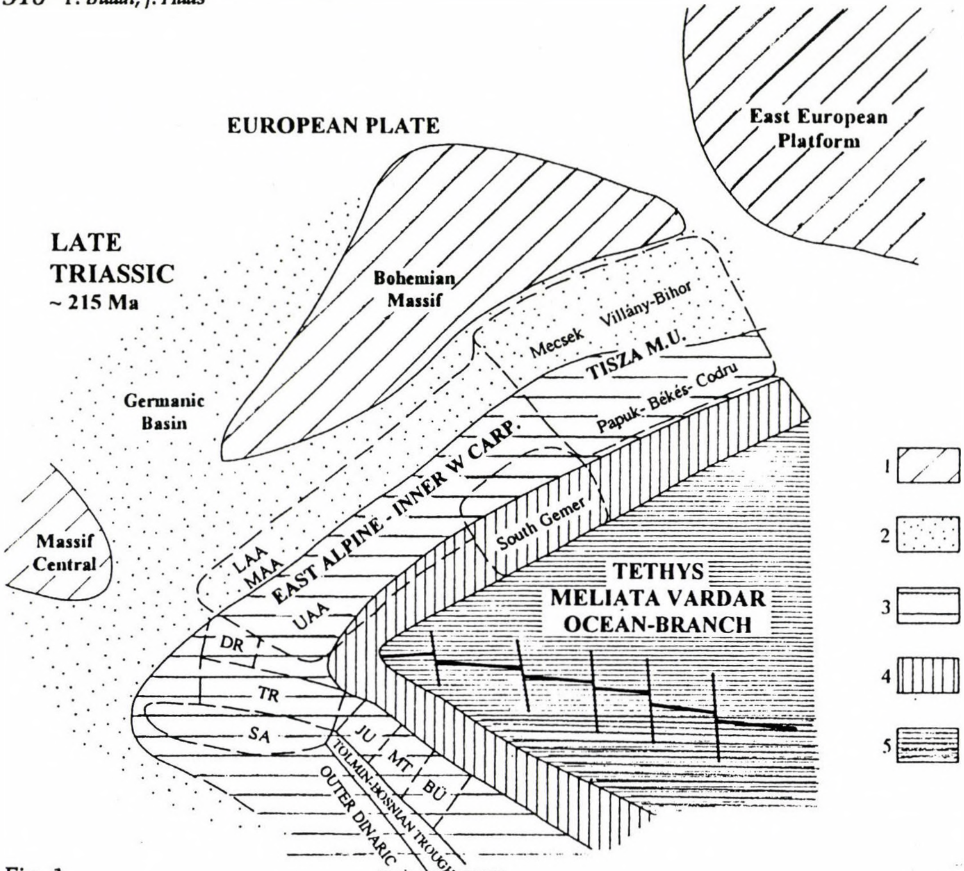


Fig. 1

Paleogeographic sketch map of the western end of the Tethys for the Norian. 1. exposed land; 2. continental (predominantly clastic) sediments; 3. platform carbonates; 4. eupelagic basin carbonates; 5. oceanic basement; BÜ – Bükkium; DR – Drauzug; JU – Julian Alps; LAA – Lower Austoalpine; MAA – Middle Austroalpine; MT – Mid-Transdanubian Unit; SA – Southern Alps; TR – Transdanubian Range; UAA – Upper Austroalpine

and of paleoecological data, a chart on changes of relatively arid and humid climatic conditions was also plotted. We also attempted to mark the boundaries of the sequences. According to considerations applied for recognition of the boundaries, three types were distinguished in Fig. 3b: 1) regional subaerial exposure surface; 2) top of the shallowing-upward cycles (without regional subaerial exposure); and 3) top of platform progradation series.

Discussion – sequence analysis and correlation

The Triassic geohistory of the Balaton Highland area can be subdivided into four major stages.

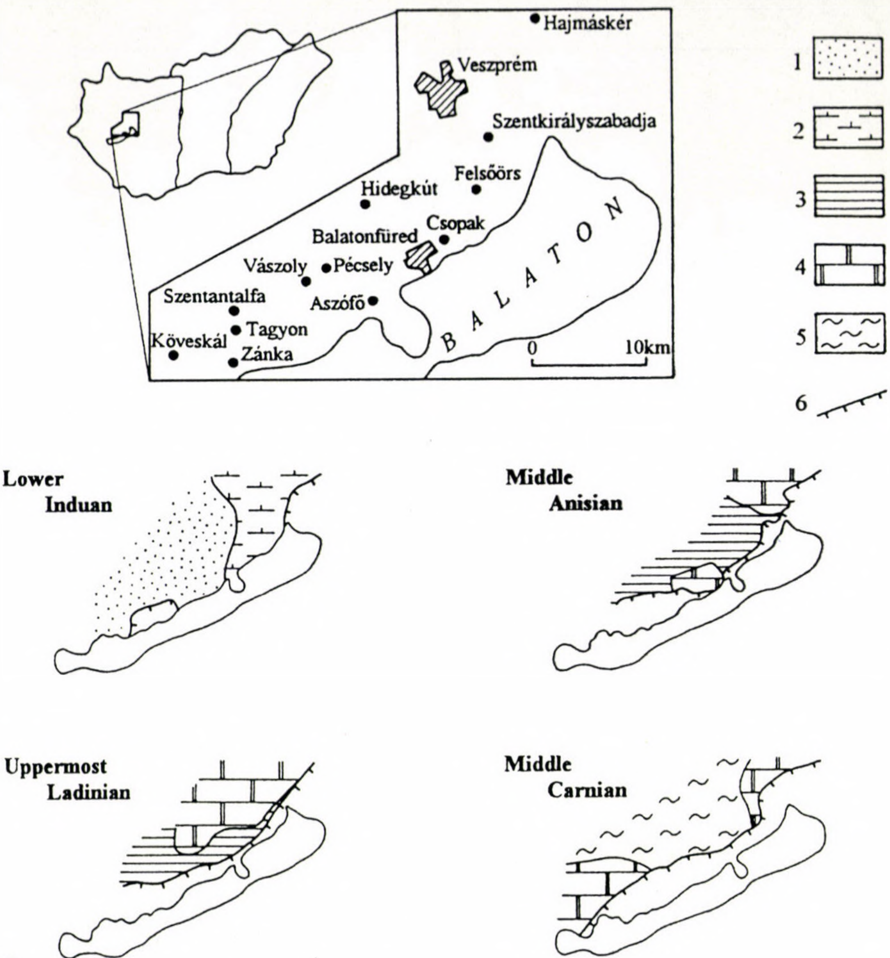


Fig. 2
Location map showing the investigated sections and facies distribution in the studied interval of the Triassic in the Balaton Highland (after Haas and Budai, 1995). 1. shallow lagoon; 2. inner lagoon; 3. pelagic basin; 4. carbonate platform; 5. intraplatform basin; 6. erosional boundary of the formations

– The first stage commenced with a significant transgression at the Permian/Triassic boundary which resulted in the inundation of the entire area. Ramp geometry and mixed siliciclastic and carbonate sedimentation characterized this evolutionary stage, which lasted to the end of the Early Triassic.

– The second stage began with a remarkable change in sedimentation pattern. Due to the cessation of terrigenous input, carbonate deposition became prevailing. This stage lasted until the Middle Anisian.

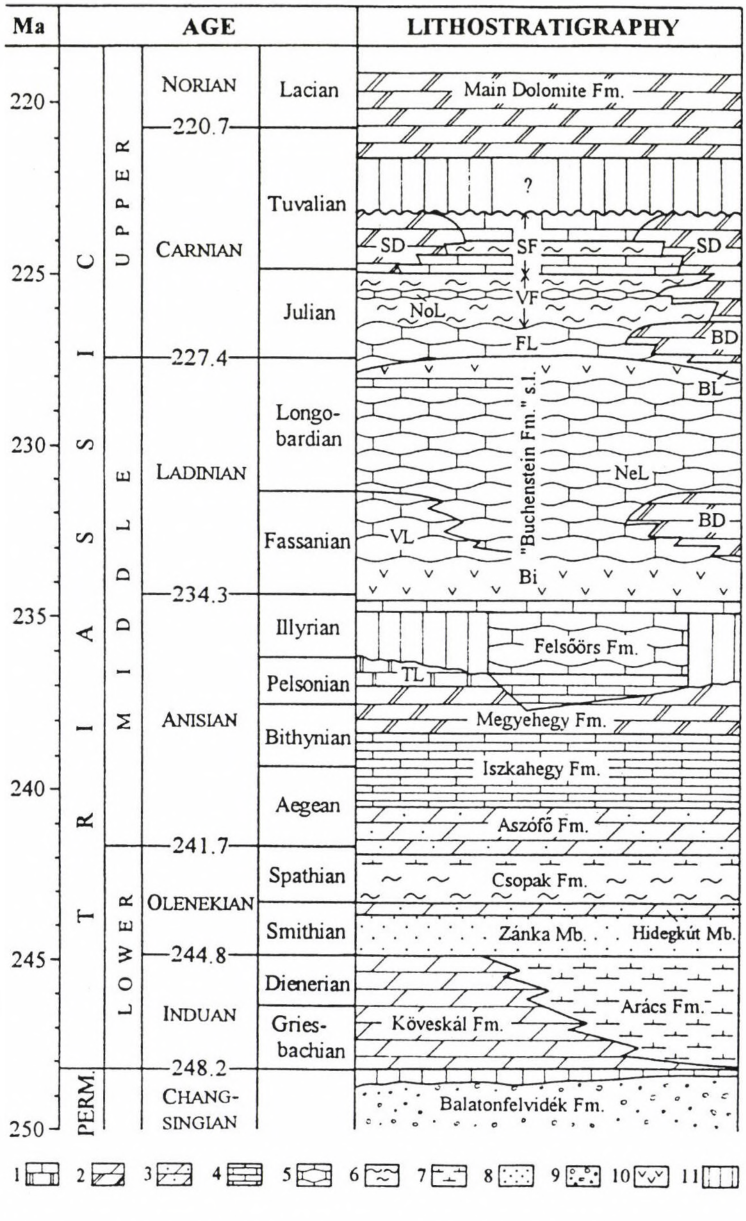


Fig. 3a
Generalized lithostratigraphic column of the Triassic of the Balaton Highland. Geochronologic scale after Gradstein et al. (1994). 1. platform limestones; 2. platform dolomites; 3. dolomites of lagoonal facies; 4. limestones of restricted lagoonal facies; 5. limestones of pelagic basinal facies; 6. neritic marls; 7. mixed (carbonate–siliciclastic) sublittoral facies; 8. sublittoral siliciclastics; 9. fluvial deposits; 10. tuffs, tuffites; 11. gap; BD – Budaörs Dolomite; Bi – "Bivera fm"; BL – Berekhegy Limestone; FL – Füred Limestone; NeL – Nemesvámos Limestone; NoL – Nosztor Limestone; SD – Sédvölgy Dolomite; SF – Sándorhegy Fm.; TL – Tagyon Limestone; VF – Veszprém Fm.; VL – Vászoly Limestone

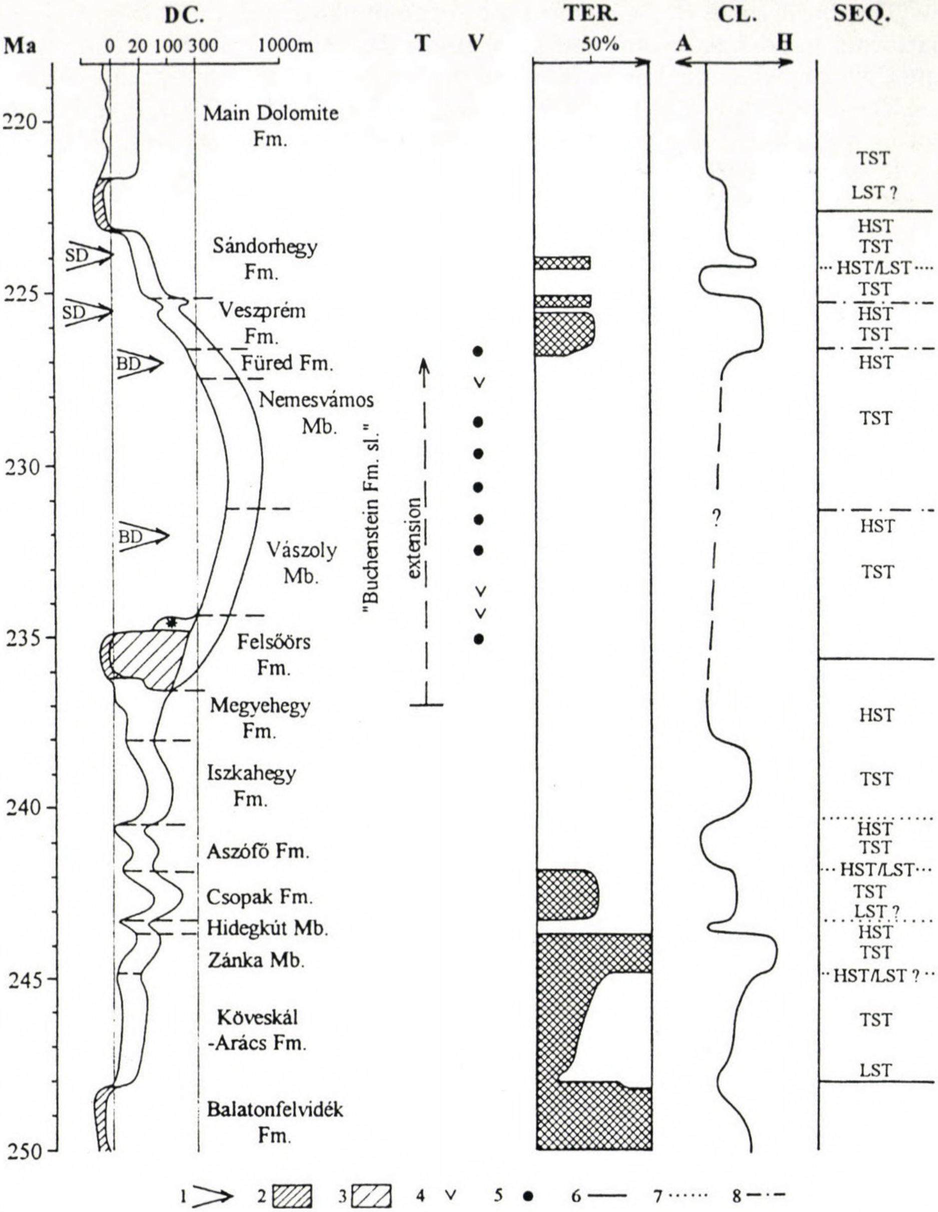


Fig. 3b
Depth curve, tectonic and volcanic events, terrigeneous influx and climatic changes during the Triassic of the Balaton Highland. 1. platform progradation; 2. subaerial exposure; 3. submarine non-deposition (slope); 4. volcanoclasts; 5. clay of volcanic origin; 6–8 types of sequence boundaries: 6. surface of subaerial exposure; 7. top of shallowing-upward cycles; 8. top of platform progradation series. * "Bivera formation" above the drowned platforms; DC. – depth curve; T – extensional tectonics; V – volcanic activity; TER. – terrigeneous influx; CL. – climate; A – arid; H – humid; SEQ. – sequence stratigraphic chart

– The third stage is characterised by differentiation of the facies pattern. Platforms and deeper basins came into being. This stage came to an end in the Late Carnian, when the basins filled up and a levelled topography was formed.

– The fourth stage represents the evolution of a huge Late Triassic carbonate platform (Dachstein Platform). However, in the Balaton Highland region, only the lowermost part of the very thick platform carbonate series has been preserved, due to the subsequent (post-Triassic) denudation. Therefore, this evolutionary stage is not discussed in the present paper.

The first stage – Early Triassic

Induan–Early Olenekian

The transgression at (or more exactly near to) the Permian/Triassic boundary which resulted in the flooding of the Late Permian coastal plains and alluvial plains was most probably triggered by a eustatic sea-level rise (Haas et al. 1988). Due to the extremely levelled topography, a significant coastal onlap already occurred in the initial phase of sea-level rise.

Following the earliest Triassic transgression, three depositional environments were developed (Haas and Budai 1995, Fig. 12):

– In the north-eastern part of the Transdanubian Range area (outside the Balaton Highland region) a shallow subtidal basin came into being (site of the deposition of the Alcsútdoboz Limestone);

– To the west, this basin was surrounded by a mud-shoal belt, where the intertidal to subtidal Arács Marl was laid down. The at most 120 m-thick Arács Marl (Fig. 4) is made up of grey and greenish-grey, intensely bioturbated marl and dolomitic marl with siltstone and limestone interlayers. The siltstones are reddish as a rule. The limestones are generally "gastropod oolites", also containing bivalve and echinoderm fragments in addition to the rock-forming microgastropods;

– Further westward, behind the mud shoals, a lagoonal environment came into existence. It was periodically affected by terrigenous siliciclastic influx. In this environment the 100 m-thick Köveskál Formation (Fig. 4) consisting of grey, porous dolomites (oolitic at the basal part of the sequences), dolomitic siltstones, and thin-bedded sandstones was formed. Evaporitic dolomites in some core sections indicate sabkha-type tidal flats located at the landward margin of the lagoon.

There are no significant vertical facies changes in the Balaton Highland region correlatable with sequence boundaries within the Induan sequences of the Dolomites and the Germanic Basin.

In the successions of the Balaton Highland there is a remarkable facies change at about the Induan/Olenekian boundary. The previously discussed formations are overlain by red or lilac, locally greyish siltstones, with a characteristic bivalve assemblage showing an upward-increasing diversity (Broglio Loriga et

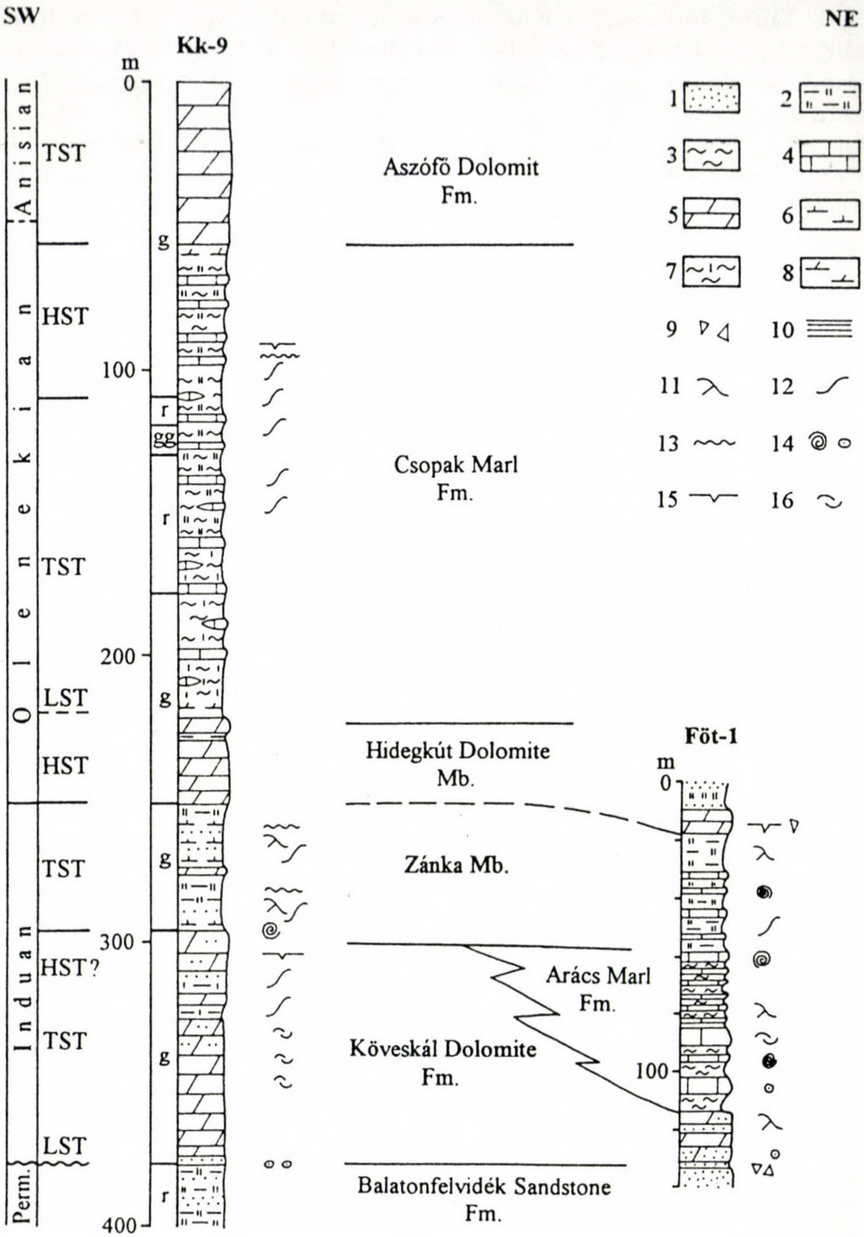


Fig. 4

Lower Triassic sequence of the Balaton Highland in the core of Köveskál Kk-9 and in Felsőörs Föt-1 boreholes (after Haas et al. 1988 and Budai 1991). 1. sand, sandstone; 2. silt, siltstone; 3. marl; 4. limestone; 5. dolomite; 6. calcareous marl; 7. dolomarl; 8. dolomite silt; 9. lithoclasts; 10. parallel lamination; 11. cross bedding; 12. bioturbation; 13. ripple marks; 14. microgastropods, oolites; 15. mud-cracks; 16. bivalves. Colour: g – grey; r – red; gg – greenish grey.

al. 1990). The 50 m-thick Zánka Member is characterized by laminated structure, i.e. alternation of red and grey laminae; however, bioturbated layers are also common. Reddish-brown limestone interlayers, generally of "gastropod oolite" facies, also occur (Fig. 4).

The siliciclastic Zánka Member is overlain by well-bedded, locally laminated, light grey, yellowish dolomites, silty dolomites or sandy limestones (Hidegkút Dolomite Member), 20–40 m in thickness. In the dolomites shrinkage cracks and bird's-eye structures are common (Fig. 4) and "rauhwacke-type" brecciated, cellular dolomites also occur in the topmost part of the member.

Lithostratigraphic analogies between the Upper Induan–Lower Olenekian units of the Balaton Highland and the Dolomites are plausible (Campil Mb. = Zánka Mb.; lower evaporitic part of the Val Badia Mb. = Hidegkút Mb.). However, our sequence stratigraphic interpretation differs in some respects from that of De Zanche et al. (1993). In our opinion the Zánka Member may represent an open ramp transgressive systems tract, whereas the Hidegkút Member (showing a regressive trend) would be a late highstand to lowstand restricted ramp deposit. In the opinion of the authors, the increase in siliciclastic influx during the Late Induan in the Balaton Highland area as well as in the Dolomites ("Campil event") can also be explained by a climatic change, i.e. increasing humidity. However, it is worth mentioning that in the Germanic Basin the most spectacular Triassic sequence boundary showing features of strong erosion ("Hardegsen Diskordanz" - Aigner and Bachmann 1992, p. 119) can be correlated with this interval.

Late Olenekian

The upper part of the Olenekian (Fig. 4) is represented by a 200 m-thick marl-dominated formation in the Balaton Highland area (Csopak Marl). Overlying the Hidegkút Dolomite its lower member consists of grey bioturbated marls, rich in molluscs. Open marine fossils such as ammonites (Tirolites) appear in the upper part of the lower member. Crinoidal and ooidic limestone interbeds (storm deposits) are common. The deepening-upward open ramp succession indicates a transgressive trend. The middle member of the Csopak Formation is made up of red silty marls and clayey siltstones with thin limestone interlayers. Open marine fossils (the ammonite genus Dinarites) occur mainly in the lower part of the middle member, representing most probably the maximum flooding. The upper member consists of grey, bioturbated silty marls and clayey siltsones, with limestone, sandstone and dolomite interbeds (HST). The diversity of the fauna shows a decreasing trend. Its transition towards the overlying dolomites (Aszófő Fm.) is gradual.

In contrast with the successions of the Balaton Highland and the Germanic Basin, three sequences can be distinguished in the uppermost part of the Lower Triassic of the Dolomites: the lower two occur within the Val Badia Member, whereas the third is actually the Cencenighe Member (De Zanche et al. 1993).

The second stage – Early to Middle Anisian

Early Anisian

At the beginning of the Anisian a significant change took place in the sedimentation of the western part of the Tethyan region: due to the radical decrease in terrigenous input the mixed marginal ramps were transformed into carbonate ramps. On the Balaton Highland the appearance of the Aszófő Dolomite marks this basic change. Similar formations are widespread in the Lower Anisian of the Dolomites (Lower Serla Dolomite), in Lombardy (Carniola di Bovegno), and also in the Northern Calcareous Alps and the Drauzug (Reichenhaller Beds). Representing the lagoonal facies, the Aszófő Dolomite is thin-bedded as a rule. Bird's-eye and tepee structures, desiccation cracks and calcite pseudomorphs after gypsum, however, indicate periodical tidal flat progradation under arid climatic conditions (Budai et al. 1993). The significant facies change close to the Olenekian/Anisian boundary is attributed mainly to a remarkable climatic change: the relatively humid climate became more arid.

In the middle part of the Lower Anisian the lagoonal–peritidal Aszófő Dolomite passes gradually upward into dark grey limestones, rich in organic material (Iszkahegy Limestone). Between them a thick transitional interval occurs, characterised by the alternation of dolomite and limestone layers, locally with intraclastic interbeds, containing limestone and dolomite clasts. These intraclastic layers may have formed in a tidal flat environment (Budai et al. 1993), suggesting maximum shoaling in the transitional interval.

In the Dolomites a definite shallowing-upward trend was observed in the Lower Serla Dolomite (Marinelli 1980). According to the sequence stratigraphic interpretation (De Zanche et al. 1993) the subtidal lagoon facies (TST) graded upward into a peritidal sabkha facies (HST). The unconformity between the Lower Serla Dolomite and the overlying Piz da Peres Conglomerate has been interpreted as a sequence boundary (De Zanche et al. 1992b, 1993).

Early–Middle Anisian

In the Balaton Highland area, above the Lower Anisian dolomites and the dolomite–limestone transitional series, the 250–300 m-thick Iszkahegy Limestone begins with dark grey organic-rich laminites, deposited on a restricted ramp under anoxic conditions. The basal laminitic unit grades upward into bedded, bioturbated limestones, formed under disaerobic conditions. This trend may indicate a decrease in the restriction of the basin which can be attributed to a sea-level rise. As to its stratigraphic position and lithology the Lombardian Lower Angolo Limestone shows striking similarity to the Iszkahegy Limestone.

The Iszkahegy Limestone passes upward into dolomite of lagoonal facies (lower part of the Megyehegy Dolomite Formation, in the sense of Budai et al.

1993). This change in the lithology can be explained by the increasing restriction of the basin (early highstand), probably under more arid climatic conditions.

It is worth mentioning that the significant sea-level drop at the end of the Bithynian which is indicated by the Voltago Conglomerate in the Dolomites (De Zanche et al. 1993) cannot be recognized in the Balaton Highland region.

The third stage – Middle Anisian to Late Carnian

Middle–Late Anisian

In the Middle Anisian (at the turn of the Bithynian/Pelsonian) disintegration of the carbonate ramp began as a consequence of extensional tectonic movements (Budai and Vörös 1992, 1993). Above the downfaulted blocks of the carbonate ramp, restricted basins were formed, characterized by bituminous laminites (Fig. 5), and also containing carbonate mud of platform origin (Felsőörs Limestone).

On the uplifted blocks isolated carbonate platforms evolved with oncoidal platform margin and dasycladacean inner platform facies (Fig. 6). They were affected by early diagenetic dolomitization in large parts of the platforms (upper part of the Megyehegy Dolomite); however, in some places the original limestone lithology survived (Tagyon Limestone).

Fig. 5
Bituminous laminated limestone in the lower part of the Felsőörs Formation (Balatonicus Zone), Aszófő

Fig. 6
Pelsonian cyclic platform carbonates at Szentkirályszabadja (from bottom to top): oncoidal subtidal facies; uneven erosion surface; and peritidal laminitic loferite

In the basins of the Balaton Highland carbonate deposition was continuous during the Pelsonian–Illyrian interval. In the Aszófő section the bituminous, laminitic limestones of the Felsőörs Formation ("Balatonites limestone") passes upward into nodular, cherty limestones before the appearance of the Paraceratites assemblage (Vörös 1987). Deposition of the laminitic limestones may have taken place in a tectonically-controlled restricted basin, receiving large amounts of bioclasts from the coeval platforms (Budai and Vörös 1992, Fig. 3), and locally redeposited slope sediments (Fig. 7). The paleoecological evaluation of the ammonoid fauna suggests a slight shallowing-upward trend (Vörös, pers. comm.) within the laminitic limestone succession.

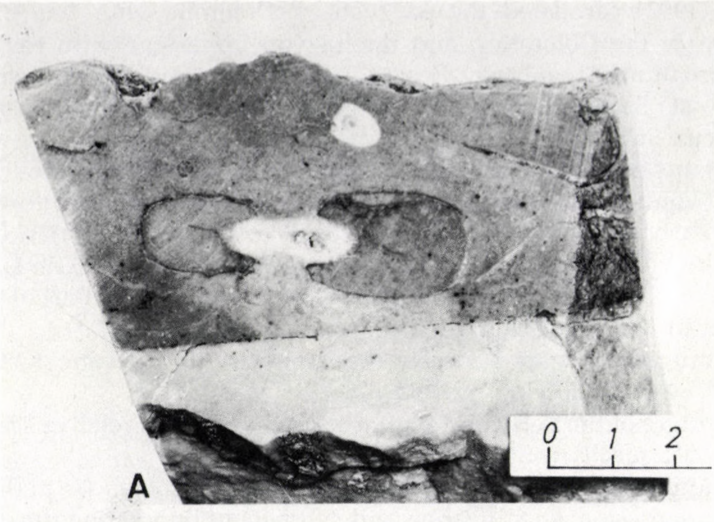
The sequence stratigraphic correlation between the Pelsonian successions of the Balaton Highland and the Dolomites is fairly clear. The maximum flooding

Fig. 7
Lithoclasts of slope origin in the lower part of the Felsőörs Limestone (Aszófő)

and even the early highstand interval appear to be within the Balatonicus Zone (De Zanche et al. 1993). The appearance of the Tethyan faunal elements in the southern part of the Germanic basin (Aigner and Bachmann 1992) may be limited to the same maximum flooding which resulted in a temporary connection between the two basins (Vörös 1992).

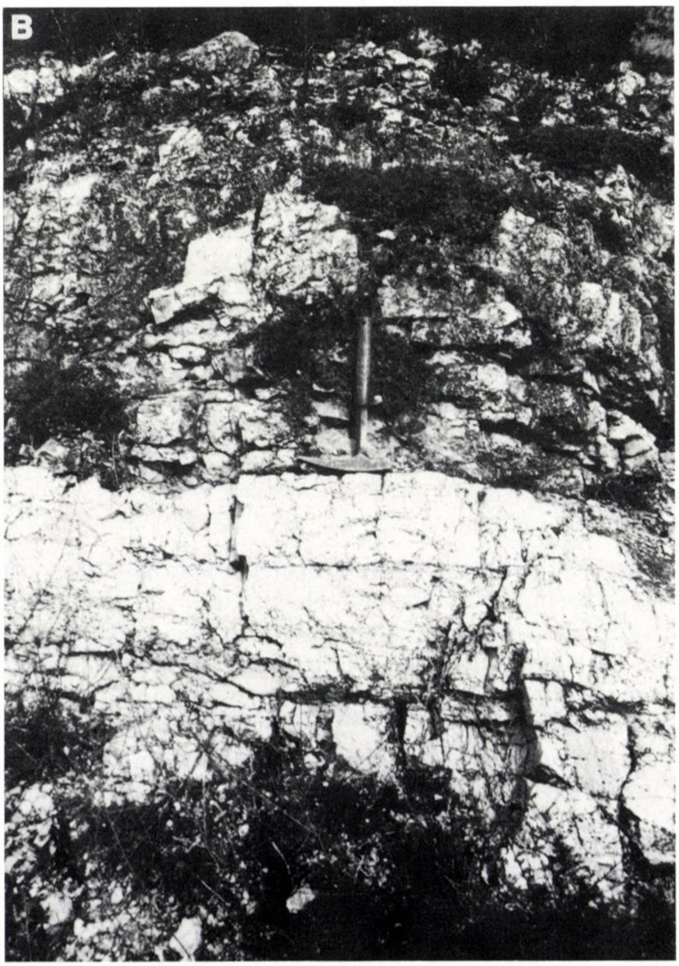
In some parts of the Anisian platforms a very spectacular abrupt vertical facies change is visible (Fig. 8); the dasycladacean platform carbonates are overlain by pelagic crinoidal (partly dolomitized) limestones with a characteristic ammonite fauna (Asseretoceras–Lardaroceras spp.). Previously the pelagic layers were assigned to the Buchenstein Formation (Budai and Vörös 1992); however, based on the latest biostratigraphic data from corresponding sections on the Balaton Highland (Szentkirályszabadja) and in the Dolomites (Mt. Rite) we suggest the correlation of these layers with the Bivera Formation (Farabegoli and Guasti 1980; Farabegoli et al. 1984; De Zanche and Gianolla 1995).

Fig. 8 →
Sharp vertical facies change above Middle Anisian platform carbonates in Szentantalfa (A) and in Szentkirályszabadja (B)



A

B



Budai (1992) correlated the Megyehegy Dolomite with the Upper Serla Formation of the Dolomites, and the Tagyon Limestone with the Dosso dei Morti Formation of Lombardy. Recently, careful analysis of a platform dolomite sequence at Szentkirályszabadja (Fig. 9) has proven the Pelsonian age (Balatonicus Subzone) of this cyclic peritidal–lagoonal series showing facies characteristics equal to those of the coeval Tagyon Limestone (Lelkes and Budai, in press ; Vörös et al., in press). The red pelagic basin sediment overlying the platform carbonates (Figs 8b and 9) proved to be Upper Illyrian (Camunum horizon) in age. According to the subdivision of Vörös (1987, 1993) and Vörös et al. (in press) the upper part of the Pelsonian and the lower part of the Illyrian is missing in this sequence (Fig. 9).

The stratigraphic gap between the platform and pelagic facies can be explained by the following models.

1) Drowning of the Pelsonian platforms due to a rapid relative sea-level rise, probably tectonically controlled;

2) Subaerial exposure of the platforms in the Pelsonian or perhaps in the Illyrian with erosion, karstification and subsequent inundation during the next significant relative sea-level rise.

In the section at Szentkirályszabadja, the erosional surface of the platform dolomite is covered by a few cm-thick red paleosol layer (Viczián, pers. comm.). That is why the latter model is preferred by the present authors. Although the Pelsonian platform carbonates show a shallowing-upward trend (Lelkes and Budai, in press) it is probable that their subaerial exposure was controlled not only by a eustatic sea level drop but also by tectonics. In the Southern Alps the significant erosion beneath the Richthofen Conglomerate marks the importance of the tectonic uplift in this interval. However, the fact that subvertical neptunian dykes were found in the platform dolomite at Szentkirályszabadja suggests that that the extensional tectonics played an important role also in the inundation. The inundation event might be correlated with the transgression at the base of the Upper Muschelkalk (Aigner and Bachmann 1992) and the appearance of the Mt. Bivera Formation above the Upper Serla Dolomite, respectively (An3/An4 boundary of De Zanche et al. 1993).

Latest Anisian–Early Ladinian

In the Balaton Highland region, the former platforms are overlain by Upper Illyrian pelagic basin carbonates which are followed by tuffaceous layers, whereas in the basins the Felsőörs Limestone is covered by relatively thick volcanic tuffs with siliceous limestone lenses or intercalations (Fig. 9).

In the area of the relatively elevated former platforms a spectacular enrichment in the cephalopod and conodont assemblage (Fig. 10) and phosphoric hardgrounds (Fig. 11) mark the interval of the maximum flooding within the lower part of the pelagic Vászoly Limestone (condensed fauna of

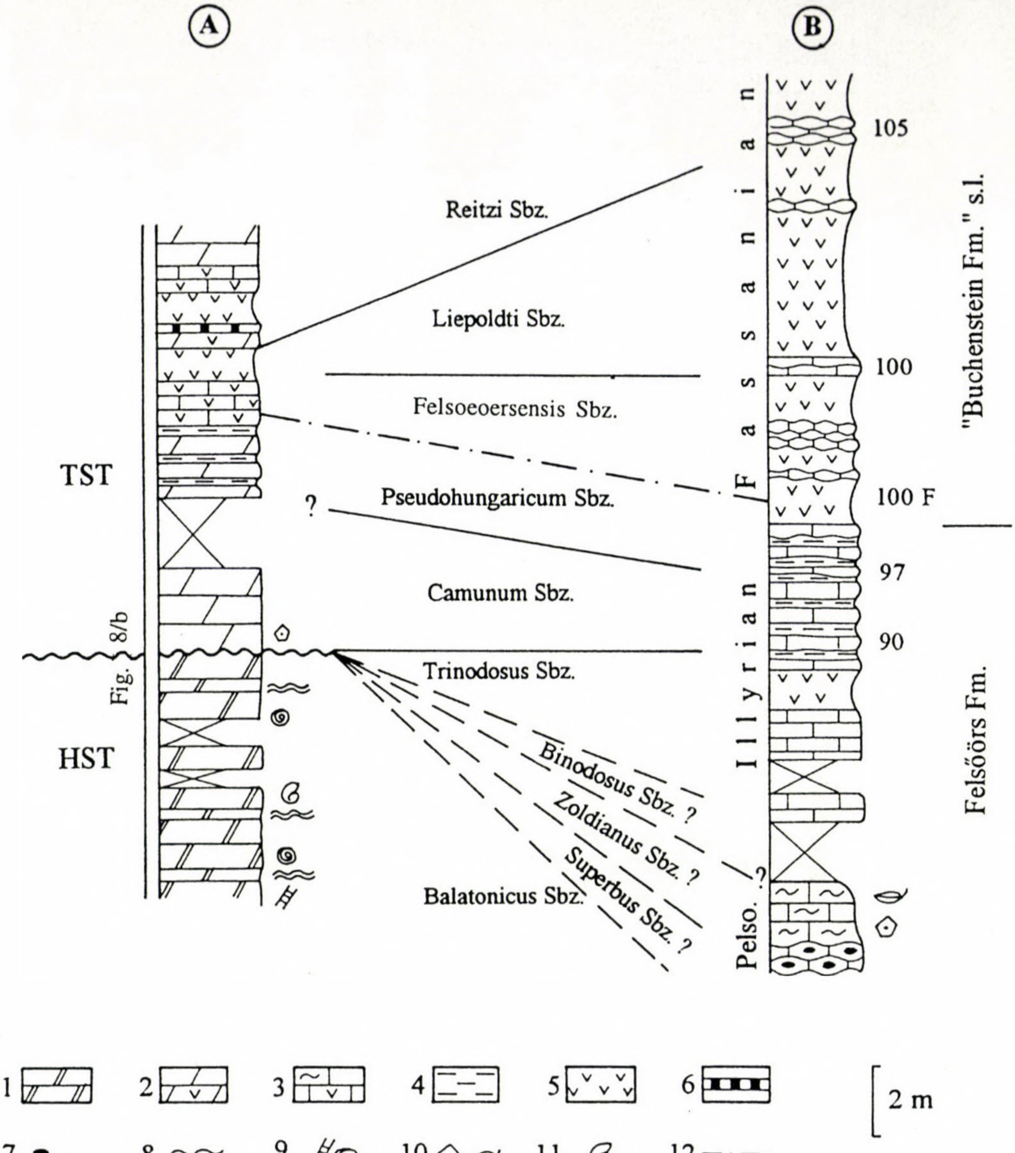


Fig. 9
Middle Triassic successions in the quarry at Szentkirályszabadja (A) and the Forrás Hill key-section at Felsőörs (B). Slightly modified after Lelkes and Budai (in press) and Vörös et al. (1996). 1. Pelsonian platform carbonates (Megyehegy Fm.); 2. dolomite of basinal facies, tuffaceous dolomite; 3. limestone, marly lmst., tuffaceous lmst.; 4. clay; 5. tuff, tuffite; 6. silicified tuffite; 7. chert; 8. algal mat; 9. Dasycladacea, oncoid; 10. crinoidea, brachiopoda; 11. *Balatonites balatonicus*; 12. Anisian–Ladinian boundary

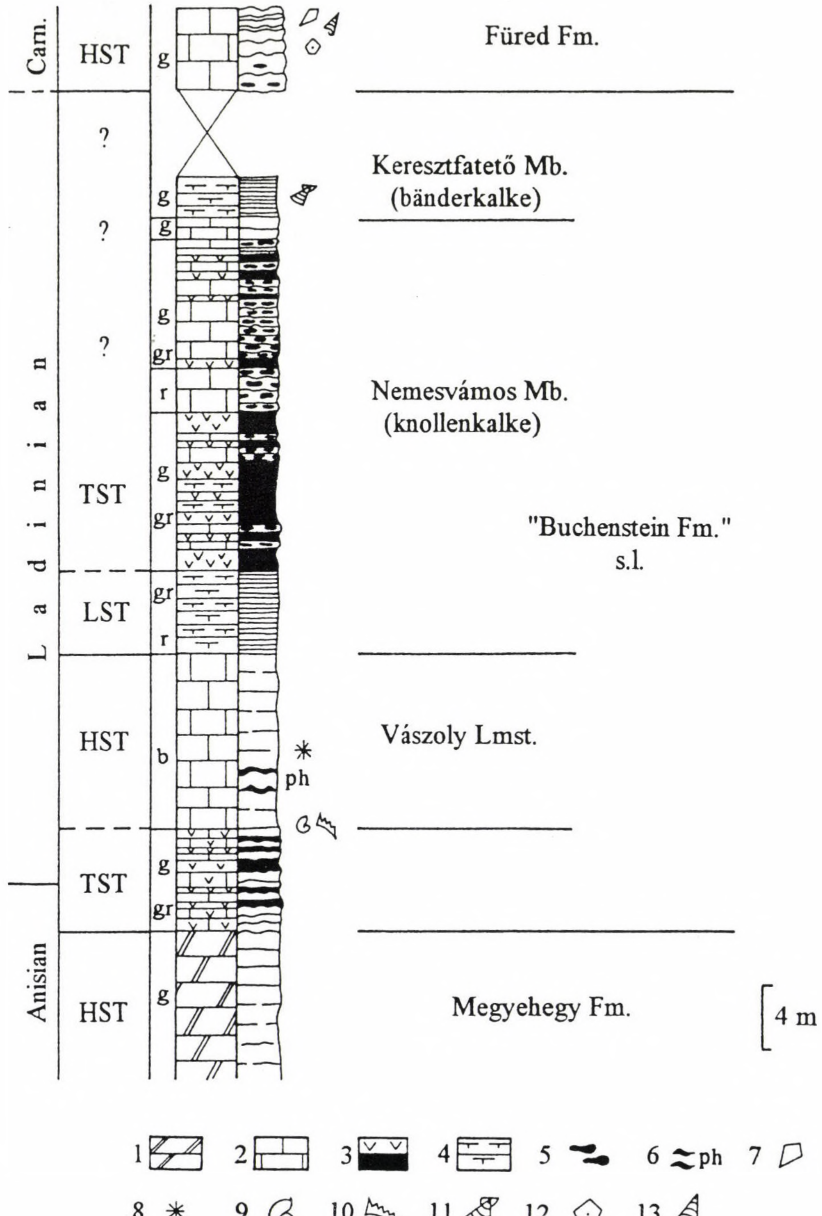


Fig. 10
Middle Anisian to Lower Carnian sequence of Öreg Hill at Vászoly (after Budai 1988 and Vörös and Pálfy 1989). For more biostratigraphic data see Vörös et al. 1996. Colours: g-grey, gr-green, b-beige, r-red, 1. dolomite, 2. limestone, 3. tuff, tuffite, 4. siliceous marl, laminated radiolarian limestone, 5. chert lenses and nodules, 6. phosphoric hard ground (ph), 7. allodapic lithoclasts, 8. *Eoprotrachyceras curionii* (*), 9–13. enrichment of fossils: 9. ammonites; 10. conodonts; 11. *Daonella* and/or *Posidonia* lumachelle; 12. crinoid fragments; 13. gastropods

Fig. 11
Phosphoric hard-ground in the lower part of the Vászoly Limestone (Curionii Zone) on Öreg Hill, Vászoly

Avisianum Subzone and the Secedensis Zone; see Vörös et al. 1996, fig. 4), while in the basins condensed pelagic red nodular limestones were deposited (Nemesvámos Limestone). The Vászoly Limestone may correspond to an incipient highstand progradation of the Budaörs Platform towards the Balaton Highland area (Curionii Zone). In the sections of the neighbouring Veszprém area, dolomite bodies within the Buchenstein Formation in the upper part of the Reitzi Zone may also be related to this process (Fig. 3a).

The upper boundary of this sequence can be well correlated to those between the Upper Muschelkalk/Lower Keuper in the Germanic Basin, and between the La1/La2 sequences in the Dolomites, respectively. The first prograding period of the Budaörs Platform can be correlated with that of the Sciliar Dolomite 1 in the Dolomites (De Zanche et al. 1993).

Late Ladinian–Early Julian

On the relatively elevated drowned platform in the central part of the Balaton Highland a new sequence appears to begin above the Vászoly Limestone (Fig. 10). The sequence begins with siliceous marls (LST), which are overlain by nodular cherty limestones with tuff and tuffite intercalations (Nemesvámos

Member). The deepest facies of the succession is probably the siliceous Posidonia–Daonella layers (Keresztfatető Mb.). The highstand period of the sequence is represented by the Füred Limestone, coeval with the second progradation of the Budaörs Platform about at the Ladinian/Carnian boundary. However, in the inner part of the Ladinian basins (Fig. 10) the platform origin of the carbonate components is less evident than in the periplatform sequences (Veszprém area), where redeposited lithoclasts and bioclasts are common in graded layers (Berekhegy Limestone). Radiolaria data suggest that the platform progradation (Fig. 12) probably began in the Late Longobardian (Dosztály, pers. comm.). In the inner part of the basin, changes in the lithofacies (Buchenstein

Fig. 12
Graded allodapic limestone layers of the Berekhegy Limestone (Uppermost Longobardian), overlain by the Budaörs Dolomite in the Veszprém area (Hajmáskér)

Fm. – Füred Fm.) which might be attributed to the platform progradation appear only at the beginning of the Carnian (Budai and Dosztály 1990).

It is worth mentioning that platform progradation at the beginning of the Carnian is also detectable in the Northern Calcareous Alps (Karwendel Platform) where graded calcarenites appear in the Partnach Beds (Brandner 1984, fig. 17).

For the time being the sequence stratigraphic interpretation of the Upper Ladinian–Lowermost Carnian interval of the Balaton Highland is a subject of debate. It is highly probable that the intense volcanic activity which may have significantly influenced the sedimentation pattern in the Dolomites during this period, affected the Balaton Highland basin only slightly or not at all, being located far from the centres of the volcanism.

Julian

Within the Lower Carnian a significant change in the lithofacies occurs; pelagic limestones are overlain by a thick marl formation (Veszprém Marl). The transitional series between the Füred Limestone and the Veszprém Marl is constituted by bituminous limestone layers a few metres thick (Budai 1992, fig. 5). They are characterized by filament-microfacies and cephalopod fauna. In the basal part of the Veszprém Formation graded bioclastic calcarenite layers are common with an upward-decreasing size of bioclasts (Csillag 1991). This trend suggests a transgressive pattern but does not explain the fundamental increase in the amount of the fine terrigenous component. A climatic change (increased humidity) may also have played an important role in the lithological change.

A 20 m-thick pelagic limestone member (Nosztor Limestone) subdivides the 800 m-thick Veszprém Marl into two parts. This is characterized by nodular "filament" limestones in the basin. However, toward the platforms it passes into a coarse bioclastic and lithoclastic facies, rich in brachiopods, echinoids, crinoids and sponges (Budai 1991; Csillag 1991). At the platform margins coarse debris of platform carbonates can be found in pelagic matrix (Buhimvölgy Breccia Mb.). It may be considered as an equivalent of the Cipit Limestone in the Dolomites (Csillag et al. 1996). The Nosztor Limestone marks the progradation of the platforms during a highstand period in the Late Julian (Fig. 13).

After the platform progradation episode a subsequent sea-level rise resulted in renewed pelagic marl deposition in the basins and backstepping of the carbonate platforms. The continuation of intense influx of fine terrigenous material led to almost total filling up of the basins by the end of the Julian.

The sudden appearance of the Veszprém Marl above the pelagic limestones can be correlated with the increase of terrigeneous siliciclastics within the Cassian Formation in the Dolomites. In the Germanic Basin channels filled by coarse clastic sediments on the surface of the Gipskeuper, indicating intense

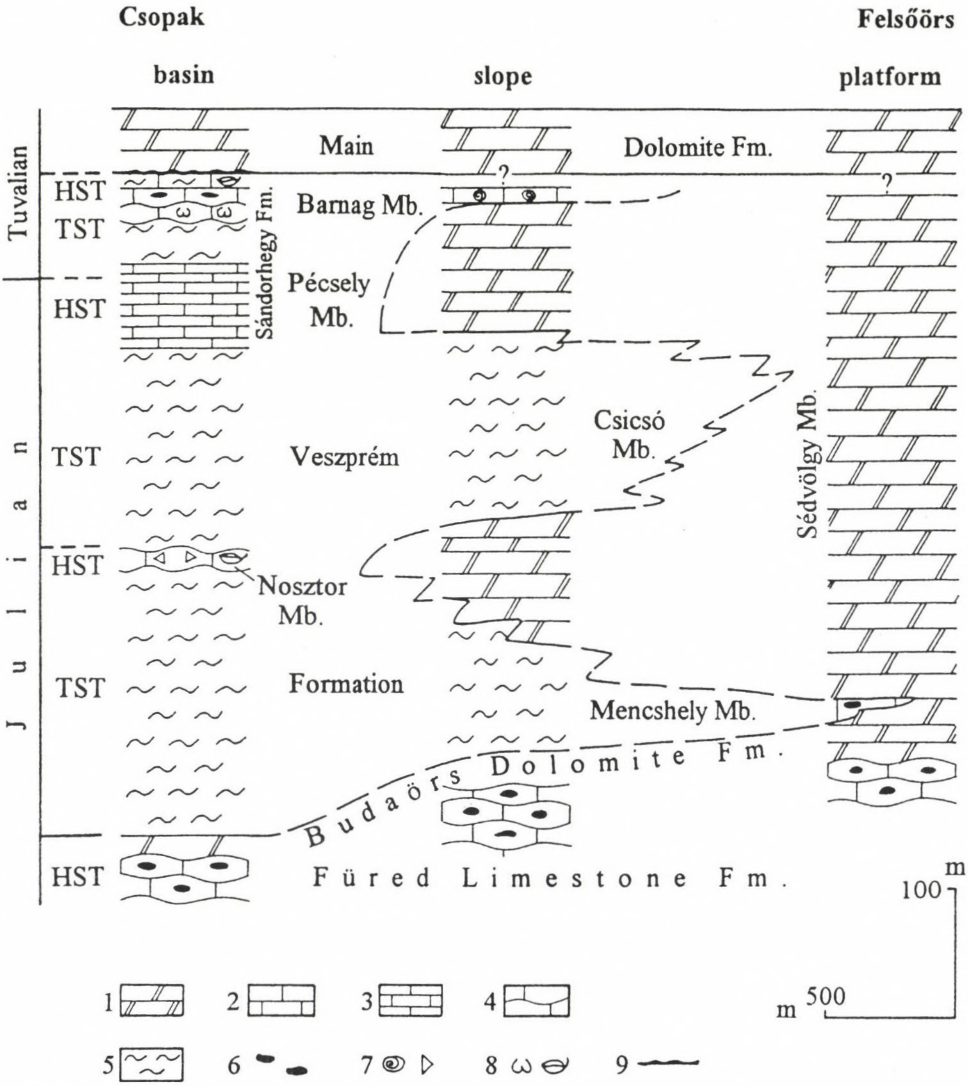


Fig. 13
Theoretical facies diagram between Nosztor valley (Csopak) and Felsőörs (after Budai 1991, Fig. 6) showing the relationships of the Upper Carnian platform and coeval basin sequences. 1. platform dolomites; 2. limestones; 3. bituminous, laminitic limestones; 4. nodular limestones; 5. marls (unexposed); 6. cherts; 7. oncoids, lithoclasts; 8. megalodontids, brachiopods; 9. calcitized lilac dolomites

subaerial erosion, were probably formed roughly coevally (Aigner and Bachmann 1992, fig. 12, 13).

Tuvalian

After the Late Julian sea-level rise resulted in the formation of the upper member of the Veszprém Marl, the hypersaline restricted lagoon facies of the Sándorhegy Formation indicates late highstand conditions (Fig. 13). Bituminous laminitic limestone lithology (Fig. 14) indicates the restriction and oxygen-depletion of the basin (Pécsely Member), whereas hypersalinity is reflected in the ostracod fauna (Monostori 1994). Appearance of megalodonts in the topmost limestone bed, bivalves of the overlying marls, however, indicate a marine environment of normal salinity, which can be explained by sea-level rise (TST). The upper member of the formation (Barnag Member) is made up of bedded, nodular limestones with megalodonts (HST). The limestone beds are overlain by a few meter-thick limestone-marl succession with large oncoids and echinoderm–bivalve–brachiopod coquinas. It may reflect a new, but probably

Fig. 14
Lower bituminous laminite (Pécsely Member) of the Sándorhegy Formation in a road-cut in the Nosztor valley, Csopak

only 4th order transgression, although the topmost part of this series is presumably truncated.

In the opinion of the authors the Sándorhegy Formation represents the final stage of the filling up of the intraplatform basins which began to form during the Anisian extension period. A large part of its carbonate component was transported from the surrounding platforms; however, in the upper part of the formation, the sea bottom entered the euphotic zone, leading to the predominance of the local bioclastic components.

Both in the Dolomites and the Germanic Basin a marked sea-level drop was found to have occurred in the middle of the Tuvalian. In the Dolomites, above an erosional surface, coarse clastics occur at the base of the Raibl Formation (De Zanche et al. 1993, fig. 23), whereas in the area of the Carnian platforms it overlies the Dürrenstein Fm. or the Cassian Dolomite with a peculiar basal sequence, consisting of variegated siliciclastics and carbonates. Porous dolomites of sabkha facies at the top of the Raibl Formation suggest a trend of shallowing (HST).

In some sections of the Balaton Highland (Fig. 13), a few meter thick lilac-red dolomites and red clay at the boundary of the Sándorhegy Limestone and the Main Dolomite probably indicate subaerial exposure (Budai 1991; Csillag 1991), which may be coeval with the aforementioned Middle Tuvalian sequence boundary (Fig. 3a–b).

The next sea-level rise led to inundation of an extremely levelled broad shelf, the site of the formation of the Main Dolomite from the Late Tuvalian to the Middle Norian. This was the beginning of the fourth stage of the Triassic history in the Transdanubian Range, but this period is poorly documented in the Balaton Highland area because of the post-Triassic denudation.

Conclusions

1. The study of Triassic formations in the Balaton Highland revealed that sea-level variation, climatic change and tectonic effects played equally important roles in determining the facies characteristics as well as the setting and features of the depositional sequences. A realistic interpretation of the evolutionary history of the region should be based only on the synoptic analysis of these factors.

2. In the first stage of the evolution, in the Early Triassic, due to its ramp geometry the moderately and uniformly subsiding shelf was very sensitive to sea-level changes. For this reason this should have been the major controlling factor; however, facies features were also influenced by the climatic conditions controlling the terrigenous input and also playing role in the dolomitization.

3. In the second stage, during the Early to Middle Anisian, it is mainly the effects of climatic changes which are detectable. The drastic reduction of terrigenous input at the beginning of the stage can be attributed to a climatic change. Climatic conditions may have determined whether dolomitization or

organic rich lime mud deposition prevailed on the restricted inner ramp. However, sea level changes may also have influenced these trends.

4. In the third stage, from the Middle Anisian to the Late Carnian, tectonic movements played the most important role. At the beginning of this stage the segmentation of the shelf began, resulting in the differentiation of platforms and basins. Sea level changes manifested themselves in subaerial exposure and the inundation of the platforms. In the basins the deposition of condensed sequences indicates maximum flooding, whereas the appearance of toe-of- slope facies suggests highstand platform progradation. Filling up of the basins began in the Carnian when, most probably due to a remarkable climatic change, terrigenous influx significantly increased. In the stage of basin filling, the significance of eustatic sea level changes may have been important.

5. Comparing Triassic sequences of the Balaton Highland with those observed in other regions, one must take into consideration the factors discussed above. In considering the Dolomites, the sequence stratigraphic correlation for the upper part of the Lower Triassic and lower part of the Middle Triassic is fairly good. In the interval between the Middle Anisian and the Carnian the sequences are primarily tectonically determined and their synchronity may reflect coeval regional tectonic effects. On the other hand, in the Ladinian–Lower Carnian pelagic successions of the Balaton Highland the interpretation of depth and climatic changes are rather ambiguous and therefore the recognition of the sequences is also uncertain. In the Carnian practically passive ("post-rift") filling up took place in the basins; consequently, the method of sequence stratigraphic correlation is successfully applicable. Correlation with the Germanic Basin is really suitable only up to the lower part of the Middle Triassic. However, the significant facies change above the Vászoly Member in the Ladinian and the appearance of the Veszprém Marl above the Füred Limestone in the Carnian correspond with sequence boundaries in the Germanic Basin, suggesting the influence of eustatic sea level changes in these cases.

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Chromite deposits of the Sagua–Baracoa range, Eastern Cuba

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The Sagua–Baracoa range includes one of the most significant chromite-producing areas of Cuba. Chromite deposits are associated with ophiolitic assemblages and show typical features of podiform chromitites: irregular shape, nodular and other characteristic textures, bimodality in Cr/Al ratio, etc. Al-rich chromites are related to the ultramafic cumulates, whereas Cr-rich ores occur in the tectonite. However, this connection is believed to be indirect and related to various compositions of successive magma batches, generated under slightly varying circumstances in the upper mantle. Chromite settled along the path of ascending magma in a spreading zone, in magma chambers and traps of various size and at various levels, as a function of the composition of the multiple magmas and equilibrium conditions. The location of this process was presumably below a marginal basin.

Key words: chromites, ophiolites, mineralogy, major oxides, microprobe analyses, origin, Eastern Cuba

Introduction

The Cuban ophiolite belt extends more than 1000 km parallel to the geographic axis of the island. The ultramafic and related rocks are exposed discontinuously forming several large massifs with a varying width of 10–50 km. In 1987–1990, in the framework of the cooperation between the Geological Institute of Hungary and the *Expedición Geológica de Santiago de Cuba*, geological mapping and ore prospecting were carried out in the Sagua–Baracoa area (Gyarmati et al. 1990) which covers roughly the easternmost massif (Fig. 1). The purpose of the work was to evaluate the mineral potential of the area which is one of the most important chromite-producing zones of Cuba.

Chromite occurrences of eastern Cuba were discovered in the 19th century, although they were described as iron ores. They have been exploited since 1916.

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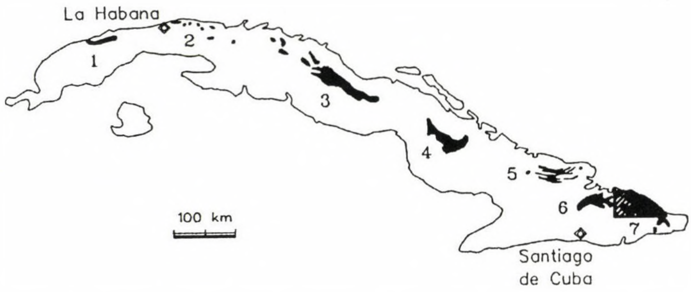

Fig. 1
Location of the studied area (shaded) and complexes of the Cuban ophiolite belt. 1. Cajálbana; 2. Matanzas; 3. Santa Clara; 4. Camagüey; 5. Holguín; 6. Mayarí; 7. Moa–Baracoa

Early comprehensive papers were published in the 1940s (Thayer 1942; Guild 1947) providing virtually the only sources of information on the chromite deposits of eastern Cuba to date. Subsequent reviews are based exclusively on these accounts (e.g., Guilbert and Parks 1986), as the results of later investigations have appeared in publications of more limited distribution (e.g., Semenov 1968; Pavlov et al. 1985). This paper presents an up-to-date account based on the recent field work and subsequent laboratory investigation.

Geological setting

The studied area (approx. 2400 km^2) is situated in Eastern Cuba, between Sagua de Tánamo and Baracoa. It is a spectacular subtropical area with variable relief ranging from sandy beaches to poorly accessible, mostly uninhabited mountainous ranges (1175 m). The region is composed largely of a Jurassic–Cretaceous ophiolite sequence (Fonseca et al. 1984) of the Moa–Baracoa massif (Fig. 2). Partly metamorphosed Cretaceous island arc volcanic rocks occur below the ophiolites as a result of Late Cretaceous obduction. The overthrusting was accompanied by formation of coarse-grained sediments (conglomerate, sandstone, olistostrome) and tectonic melange. Cenozoic formations are represented mainly by calcareous sediments and subordinate felsic tuffs. Nickeliferous laterite and refractory chromite are the most important mineral deposits of the area.

All units of the typical ophiolitic sequence are presumed to be present. The ultramafic rocks are mainly serpentinized harzburgite and dunite, but subordinate pyroxenite, wehrlite, lherzolite and plagioclase-bearing peridotite also occur. Features such as abundant kink-bands in olivines, euhedral chromite grains, and cataclastic fabrics in harzburgites and interlayered dunites suggest a large distribution of tectonite, mainly in the western part of the studied area. The other ultramafic rocks presumably belong to the cumulate assemblage. Along with more differentiated rock types (gabbroids), they are more frequent in the east. Limits of both units are obliterated by the almost complete serpentinization and deformation.

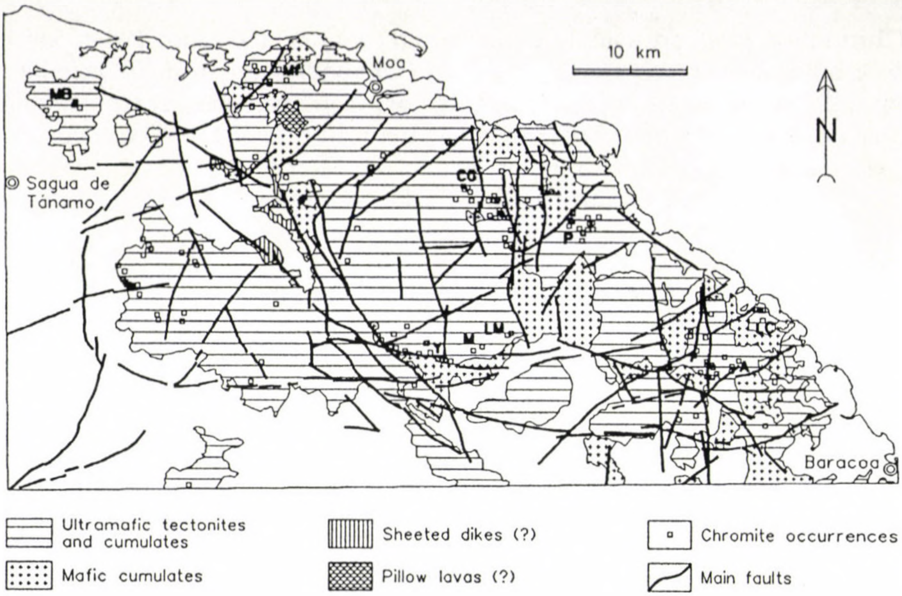

Fig. 2
Generalized map of the ophiolite suite of the Sagua–Baracoa range (after Gyarmati et al. 1990). Deposits and prospects: MB – Monte Bueno; Mf – Miraflores; CG – Cayo Guam; P – Potosí; Y – Yarey; M – Mercedita; LM – La Melba; LC – La Constancia; A – Amores

Mafic cumulates are represented largely by gabbros of both layered and massive varieties. They are mostly in tectonic contact with the ultramafic cumulates but sometimes a gradual transition through alternating layers may occur. Gabbronorite, olivine gabbro, troctolite, leucocratic gabbro, anorthosite, microgabbro and gabbro-pegmatite are also present in the mafic suite and locally have been intruded into the ultramafic rocks.

More felsic units (plagiogranites) are uncommon. The presence of sheeted dykes, pillow lavas and pelagic sediments is debatable. Abundant diabase was mapped in two areas at the border of the ultramafic rocks and the island arc volcanic rocks, but they could belong to the latter suite. The intensely weathered hyaloclastite found in the Sagua–Moa roadcut and in a nearby borehole also have ambiguous (tholeiitic and calc-alkaline) chemical character. The similarly weathered radiolarite described in the same exposure also requires further investigation.

Field relations

About 200 podiform-type (Thayer 1964) chromite occurrences are known in the studied area (Fig. 2). They are mostly small (less than 10 m in diameter) and without economic significance. Of the four largest deposits, Cayo Guam and Potosí are abandoned while Mercedita and Amores are under exploitation.

Chromite deposits are usually lenticular, but columnar (Semenov 1968), pod-like and tabular bodies were also described. Ore-bodies are usually parallel or subparallel to the layering of the host rocks, but in some cases clearly crosscut them.

Sizes of the ore-bodies vary from a few metres up to 600 m. The latter value refers to the length of an irregular lens in the Mercedita deposit. This is the largest chromite body of Cuba with a thickness varying between 1 and 16 m. Its original width is estimated as 200–250 m. Thus, the dimensions of this lens are commensurate with those of the Coto ore-body, one of the largest known podiform chromite bodies of the world (600 x 300 x 80m) at the Masinloc mine, Philippines (Dickey 1975).

The ore deposits are distributed irregularly in the studied area "as plums in pudding" (Thayer 1942). In fact, the ultramafic rocks are almost completely serpentinized and the whole ophiolite complex is extremely tectonized locally containing overturned blocks (Fonseca et al. 1984), or fragments of oceanic crust overthrusted onto each other (Andó et al. 1993). This makes prospecting for the chromite deposits as difficult as recognizing the structure of the complex.

Most ore-bodies are located within a 1 km wide zone of ultramafic rocks, below the mafic contact, as in Central Cuba (Flint et al. 1948) and many other regions of the world. The uppermost ore-bodies in the Amores mine are located in plagioclase-bearing peridotite, 50 m below the layered gabbro. As chromite deposits may be formed at the bottom of, or below, the cumulate magma chamber, a thin ultramafic cumulate suite with thickness of not more than a few hundred metres is inferred for the region.

The tectonite complex also contains numerous chromite lenses, such as in the Monte Bueno block in the western part. Their largest diameter is 20 to 30 m.

Residual ore-bodies are found in the lateritic cover. Chromite remains virtually intact and is considerably concentrated while the host serpentinites are totally lateritized and compacted. These *float ores* (Thayer 1942) may also have economic importance (e.g., Friedrich et al. 1980), and thus deserve further investigation.

Ore-bodies are often cut by dykes. Serpentinized ultramafic dykes are ubiquitous, whereas dykes of gabbro (Fig. 3), gabbronorite and troctolite are known mainly from the Cayo Guam, Potosí and Amores mines. Most gabbroic dykes are pegmatitic and crystals as large as 10 to 15 cm are frequent; Guild (1947) depicted 90 cm crystals of augite in the gabbro-pegmatites at the Cayo Guam mine. Coarse-grained gabbro also occurs, whereas medium- and fine-grained varieties are rare. The grain size, however, may vary within the same dyke. For example, in a dyke crossing a massive ore-body at Potosí, the grain size increases from the wall toward the centre, ranging from micro-gabbro through gabbro to gabbro-pegmatite (Fig. 4). The rapid cooling and crystallization near the contact implies that significant time might locally lapse between ore segregation and gabbro intrusion, in contrast to Thayer's (1964) conclusion that dykes formed simultaneously or "very soon after" the country rocks. However, such variability in grain size is rare and most dykes are equigranular.

Fig. 3
Massive chromite with gabbro dyke, Amores mine. The length of the scale is 2 cm

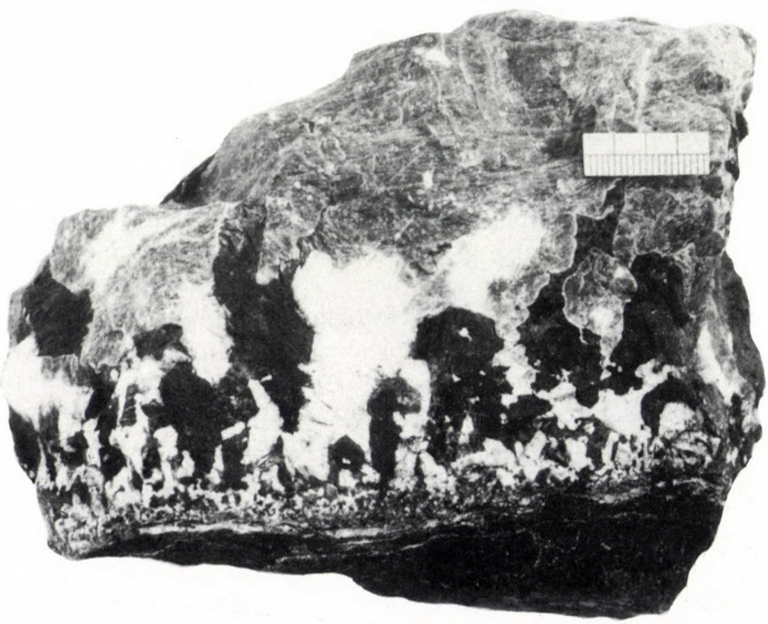

Fig. 4
Gabbroic dyke, ranging in composition from microgabbro to gabbro-pegmatite (black on the bottom: massive chromite), Potosí mine. The length of the scale is 2 cm

The dykes range in width from a few mm to 1.5 m (Thayer 1942; Semenov 1968); their length does not exceed 15 m (Semenov 1968). Autointrusion (Hughes 1982) related to tectonic movement is presumably responsible for their formation.

Ore textures

Massive, disseminated, and schlieren ores are the most common types in the area (Kovács et al. 1992). *Massive* chromites consist of medium- to coarse-grained, interlocking subhedra with a small amount of serpentine and chlorite. *Disseminated* chromites are mostly equigranular (2–5 mm), anhedral, or subhedral grains with resorbed rims. *Schlieren* type ore occurs in the Monte Bueno, Miraflores and Cayo Guam deposits.

Nodular texture has been observed in the Yarey deposit (Fig. 5). The round, slightly elongate nodules vary in size from 3 to 15 mm. The smaller ones have massive internal structure whereas the larger aggregates have a silicate-bearing core with diameter up to 8 mm, surrounded by a 2–5 mm thick chromite crust. The transition between the core and the crust is gradual. The silicate-rich nucleus contains fine, strongly corroded chromite grains as well. The nodules themselves also have corroded boundaries. Both the nuclei and the matrix of the nodules contain serpentine and subordinate chlorite with relics of olivine and pyroxene.

Most cored nodules in the Yarey deposit show pronounced concave contours (Fig. 5). These were conspicuously indented from all sides as a result of tectonic deformation of the host rocks. The silicate cores suggest that crystallization began with olivine, chromite and orthopyroxene, followed again by chromite precipitation (Greenbaum 1977).

Chromitites with *layered* fabric occur in considerable amounts in the Miraflores and Potosí deposits. They consist of alternating bands of chromitite and serpentinized dunite. The thickness of bands varies from a few mm to 2–3 cm. The chromite commonly forms fine to medium-grained aggregates, though the thicker layers locally consist of coarse grains. Transitions between the serpentinite and chromite bands are gradual through *chromite net* and *occluded silicate* microtextures. These fabrics are found both in layered intrusions and ophiolites and are regarded as evidence for cumulus origin (e.g., Greenbaum 1977).

The primary textures are often deformed as a result of subsequent intrusions and tectonic events. *Brecciated* fabrics occur in the Cayo Guam and Potosí mines where massive chromite was fractured during intrusion of coarse-grained gabbro. Cataclastic textures has affected virtually every chromite grain from all occurrences.

Pull-apart texture is more frequent in the eastern part of the Moa–Baracoa complex (Mercedita, La Melba, La Constancia, Amores deposits). Greenbaum (1977) regarded the development of such dilatant crac[illegible] in individual ore-bodies as a result of mass movements. Brown (1980) s[illegible]ed that tension

Fig. 5
Nodular chromite, Yarey deposit. Note the concave contours of most nodules. Scale in centimetres

fractures were formed by tectonic deformation and were perpendicular to the maximum direction of stress. However, tectonic pressure is more likely to result in irregular, cataclastic fracture patterns rather than subparallel cracks. The wide-spread distribution of the pull–apart texture in most podiform chromite districts (Thayer 1964) suggests extension on a regional scale, whereas subparallel orientation of fissures, as well as general lack of displacement along them implies the effect of a single couple of opposed forces.

Extension, which forms in the convex side of a flexed plate on the outer slopes of trenches and collisional foredeeps (Bradley and Kidd 1991), might give rise to such fracturing. Accordingly, pull–apart texture should be more common in ore bodies associated with the upper part of the ophiolite assemblage and less frequent downward in the sequence as the strain gradually decreases with depth (as far as a "neutral surface" at about half the thickness of the plate, below which contraction arises due to bending). In the extended side of the folded slab, opening of tension cracks is confined to the uppermost part (Bradley and Kidd 1991). This model may be applied to the studied area as the chromite deposits in the eastern part generally represent a higher stratigraphic level. The flexing probably took place before obduction.

Ore texture may change from one occurrence to another but within the same ore-body it is consistent. In some places the ore grades into the country rocks, but sharp contacts are more common, and this is not always due to faulting.

Chemical composition of chromite

Chemical composition of chromian spinels have been investigated by electron microprobe analyses. Ore chromites show a bimodal Cr distribution typical of podiform deposits (Dickey 1975; Leblanc and Violette 1983). High Cr_2O_3 contents (50.7–54.6 wt%) are characteristic of the Monte Bueno occurrences (Table 1), while the other deposits (Miraflores, Mercedita, La Melba, Amores) have low Cr_2O_3 contents (35.7–37.9%). Al_2O_3 varies between 15.0–18.2% and 27.5–31.8%, respectively.

MgO and FeO vary within a relatively narrow range (FeO: 10.6–17.3%; MgO: 11.0–17.4%). Alteration along rims and cracks of chromite grains formed high reflectance zones (width up to 0.4 mm) which are depleted in Al and Mg and enriched in Cr and Fe (Table 1).

Compositions of accessory chromites from harzburgite in Miraflores (Table 1) show higher Fe/Mg ratio than in the ores which suggests formation from a slightly differentiated magma and/or subsolidus reequilibration (Irvine 1967).

All data from the unaltered chromites project onto the field of podiform chromites in the Cr/(Cr+Al) vs. $Mg/(Fe^{2+}+Mg)$ diagram (Fig. 6a). TiO_2 does not exceed 0.44 wt% and does not show any positive correlation with major oxides, which is also characteristic of podiform chromites (Dickey 1975).

Low-chromium, high-alumina ores are known also from other ophiolite complexes, but Cr_2O_3 less then 38 wt% and Al_2O_3 more than 31% as well as $TiO_2 \geq 0.40\%$ in chromitites are rather rare. However, these values compare well with the results of previous chemical (Guild 1947; Pavlov et al. 1985; Fig. 6b) and microprobe (Ukhanov et al. 1985) data.

Samples from Amores have the lowest values of Cr/(Cr+Al) ratio (Fig. 6a, b). Consequently, a compositional series can be identified from the Cr-rich (metallurgical) ores in tectonites to the Cr-poor, Al-rich (refractory) chromites in cumulates, including the Amores deposit which appears in the highest stratigraphical level and provides the most aluminous ore (Kovács et al. 1993). Kenarev (1966) found a similar trend within the Potosí deposit.

The compositional data imply that, geologically, the Monte Bueno block belongs to the Mayarí massif rather than to the Moa–Baracoa complex (cf. Fig. 6a, b). The former, located west of the studied area (Fig. 1), is composed mostly of mantle tectonites (Fonseca et al. 1984) containing metallurgical chromite (Thayer 1942; Semenov 1968).

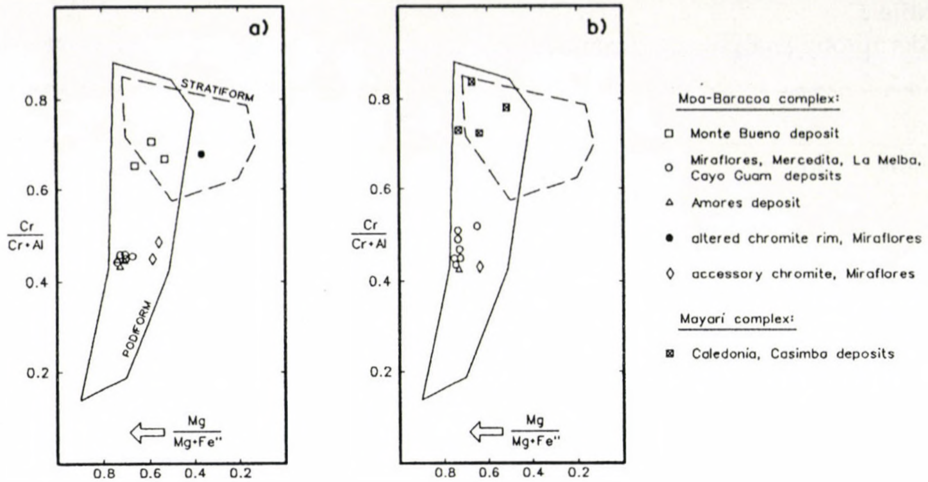

Fig. 6
Compositional diagram of chromites of the studied and adjacent areas. Atomic proportions based on: a) microprobe analyses (Table 1), and b) chemical analyses (Guild 1947; Pavlov et al. 1985). Fields of stratiform and podiform ores after Irvine (1967).

Mineral composition

Serpentine and chlorite are the most common silicates in the ore-bodies, with some relics of olivine and pyroxenes (enstatite, rarely Cr-diopside). Locally chlorite is the dominant gangue mineral (Table 2). Its formation is obviously due to the release of Al and Mg from chromite along selvedges and fractures during serpentinization (Whittaker and Watkinson 1986; Buda 1988). Amphiboles are less frequent. In the cumulates, olivines of host rocks are richer in Fe ($Fo_{90.5–91.5}$) than olivine inclusions in the ores ($Fo_{94.3–95.1}$; Table 3). However, relict olivine from ore deposits in the tectonite environment has not been found.

Microprobe analysis has revealed millerite, pyrrhotite, and a platinum-group mineral with dominant Ru and Fe and minor Os and Ir occurring as inclusions in chromite (Fig. 7). The latter has been discovered in the Monte Bueno occurrence.

Chromian garnet is a typical pathfinder mineral of chromite. Though it is essentially absent in the western part of the Sagua–Baracoa range, it occurs in numerous deposits (Mercedita, La Melba, Amores, etc.) to the east, essentially at the same places where Al-rich chromite and pull-apart textures are developed. Fine (<1 mm) euhedral garnet crystals occupy major fractures. Chemical data (Table 3) indicate compositions between grossular and uvarovite.

Table 1
Microprobe analyses of chromites

	1	2	3	4	5	6	7
Cr_2O_3	50.73	54.59	51.36	37.43	37.38	45.30	37.42
Al_2O_3	18.21	14.99	16.65	29.97	30.26	14.49	27.52
$Fe_2O_3^*$	2.71	2.10	1.78	4.20	2.57	9.17	4.74
TiO_2	0.24	0.44	0.17	0.40	0.20	0.30	0.22
FeO	12.97	15.55	17.26	11.66	13.43	22.58	17.51
MgO	14.11	12.54	11.03	16.54	15.03	7.36	12.21
MnO	0.25	0.20	0.18	0.20	0.28	0.59	0.33
NiO	0.17	-	-	0.16	-	-	0.09
Total	99.39	100.41	98.43	100.56	99.15	99.79	100.04
			Cations	per 32	oxygens		
Cr	10.040	10.936	10.506	6.904	6.960	9.455	7.184
Al	5.368	4.475	5.075	8.224	8.392	4.485	7.864
Fe^{3+}	0.512	0.398	0.346	0.736	0.456	1.835	0.864
Ti	0.040	0.082	0.031	0.064	0.034	0.060	0.032
Fe^{2+}	2.704	3.295	3.736	2.264	2.644	4.995	3.544
Mg	5.264	4.731	4.251	5.736	5.272	2.875	4.416
Mn	0.040	0.043	0.039	0.032	0.058	0.135	0.056
Ni	0.032	-	-	0.024	-	-	0.016

* Fe_2O_3 is calculated from total FeO assuming spinel stoichiometry

Sample location and lithology:

1. Monte Bueno, massive ore (average of 6 analyses)
2. Monte Bueno, massive ore (average of 4 analyses)
3. Monte Bueno, schlieren ore (average of 4 analyses)
4. Miraflores, massive ore (average of 6 analyses)
5. Miraflores, schlieren ore (average of 5 analyses)
6. Miraflores, altered rim of the former grains (average of 2 analyses)
7. Miraflores, borehole PE-2 25.9 m, harzburgite (average of 6 analyses)

Fine euhedra of secondary magnetite, presumably precipitated during serpentinization, have also been determined. Pyrite, chalcopyrite and gold minerals (electrum, native gold, calaverite) were identified in a few samples by microscopic methods.

The original wall rocks, if they are not totally obscured by serpentinization, always are dunite, as is typical of podiform chromitites (Dickey 1975). Thickness of the dunitic rim varies from 1–2 cm to several metres (Semenov 1968).

Discussion and conclusions

All chromite deposits associated both with cumulate and tectonite sequences are interpreted to be generated by cumulus processes (Thayer 1964), as textural

Table 1
Continued

	8	9	10	11	12	13	14
Cr_2O_3	37.38	37.91	37.24	37.92	36.82	35.66	37.01
Al_2O_3	30.47	30.85	31.81	29.99	30.42	31.52	31.06
$Fe_2O_3^*$	2.89	3.20	3.41	4.26	4.57	3.68	2.32
TiO_2	0.02	0.31	0.18	0.40	0.28	0.23	0.16
FeO	16.72	11.83	10.60	11.01	10.70	11.20	10.97
MgO	13.17	16.53	17.35	17.05	17.06	16.67	16.59
MnO	0.28	0.22	0.24	0.19	0.20	0.07	0.16
NiO	0.07	0.21	0.24	0.18	0.17	-	-
Total	101.00	101.06	101.07	101.00	100.22	99.03	98.27
			Cations per 32 oxygens				
Cr	6.992	6.936	6.760	6.944	6.776	6.606	6.888
Al	8.488	8.408	8.592	8.176	8.328	8.703	8.619
Fe^{3+}	0.512	0.552	0.592	0.744	0.800	0.590	0.412
Ti	0.000	0.048	0.032	0.064	0.040	0.041	0.028
Fe^{2+}	3.296	2.280	2.024	2.128	2.080	2.195	2.160
Mg	4.632	5.688	5.920	5.872	5.904	5.822	5.824
Mn	0.048	0.032	0.032	0.032	0.032	0.015	0.033
Ni	0.016	0.032	0.040	0.032	0.032	-	-

* Fe_2O_3 is calculated from total FeO assuming spinel stoichiometry

Sample location and lithology:

8. Miraflores, borehole PE-2 466.0 m, harzburgite (average of 5 analyses)
9. Mercedita mine, massive ore (average of 6 analyses)
10. Mercedita mine, massive ore (average of 6 analyses)
11. La Melba, massive ore (average of 6 analyses)
12. Amores mine, massive ore (average of 6 analyses)
13. Amores mine, massive ore (average of 3 analyses)
14. Amores mine, massive ore (average of 2 analyses)

and experimental evidence do not support a residual origin for significant amounts of spinel (Dickey and Yoder 1972). Dickey (1975) explained this feature by gravitational sinking of chromite pods as solid autoliths from the zone of segregation into the underlying tectonite peridotite. Ukhanov et al. (1985) suggested a similar mechanism for the origin of the Mercedita deposit.

However, the ubiquity of dunitic envelopes around the ore-bodies as well as the apparent relation between the ore composition and the position of deposits in the vertical rock sequence as observed in the Sagua–Baracoa range and in many other chromite districts of the world, are not compatible with this hypothesis (Brown 1980). Therefore, the process of fractional crystallization from a rising magma was suggested, which began beneath the main magma

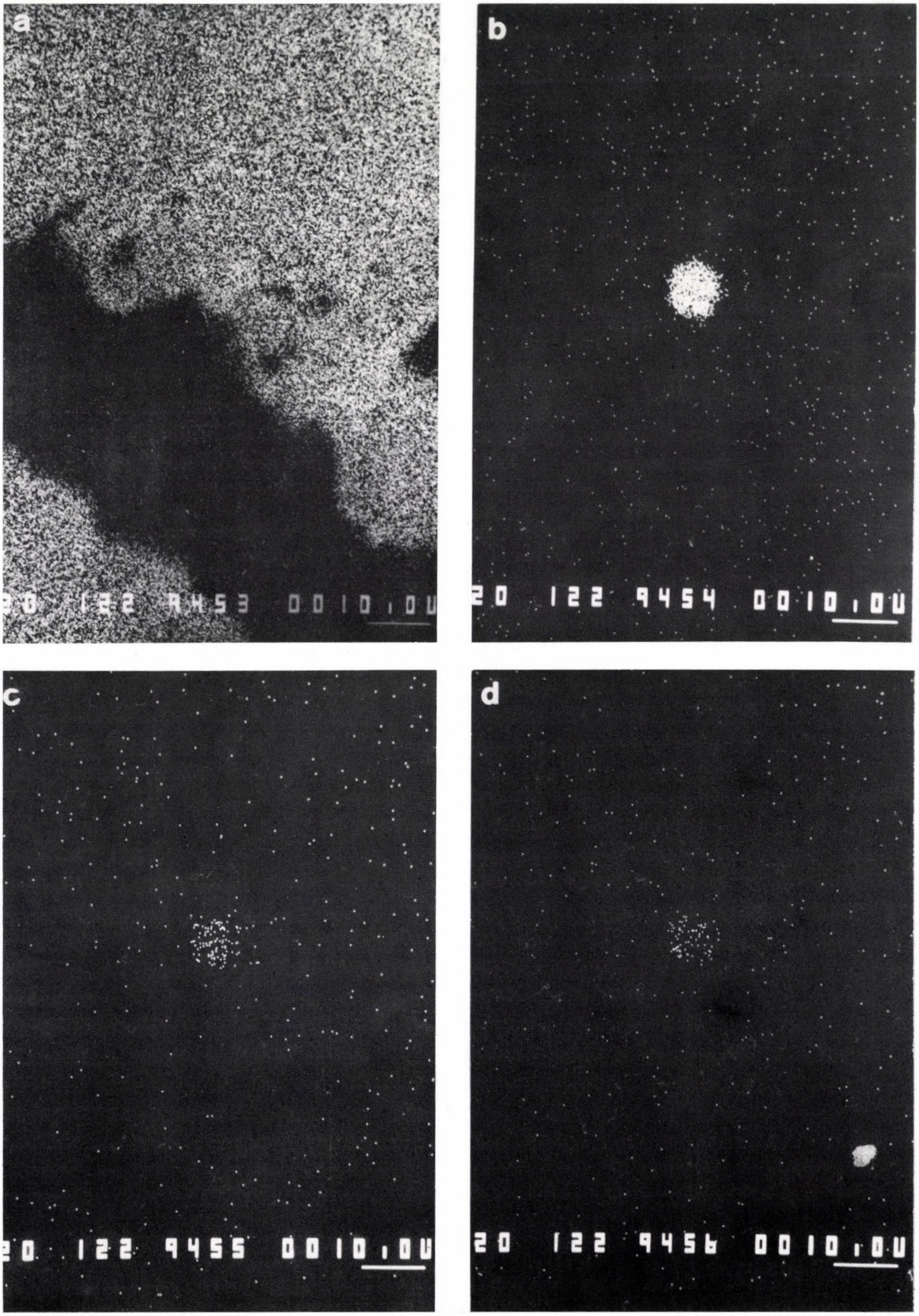
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Table 2
Microprobe analyses of orthopyroxenes (opx), clinopyroxenes and chlorites

	1	2	3	4	5
	opx	clinopyroxenes		chlorites	
SiO_2	56.55	52.34	52.61	32.14	33.08
Al_2O_3	2.02	2.68	3.05	16.49	12.95
TiO_2	0.04	0.11	0.25	n.a.	n.a.
Cr_2O_3	0.69	1.17	1.30	1.84	2.95
FeO	5.75	1.78	1.99	0.83	1.50
MnO	0.16	0.07	0.08	n.a.	n.a.
MgO	33.51	16.90	16.54	34.43	34.61
CaO	1.28	23.65	23.46	-	-
Na_2O	-	0.32	0.36	0.01	-
K_2O	n.a.	n.a.	n.a.	-	0.05
H_2O	n.a.	n.a.	n.a.	12.65	12.46
Total	100.00	99.02	99.64	98.39	97.60
	Cations per 6 oxygens			per 16 OH	
Si	1.953	1.923	1.919	6.096	6.37
Al^{IV}	0.047	0.072	0.081	1.904	1.63
Al^{VI}	0.035	0.039	0.053	1.781	1.31
Ti	0.001	0.002	0.006		
Cr	0.019	0.034	0.037	0.276	0.45
Fe	0.166	0.055	0.061	0.132	0.24
Mn	0.004	0.002	0.002		
Mg	1.724	0.925	0.899	9.732	9.93
Ca	0.047	0.931	0.917	-	-
Na	-	0.023	0.025	0.006	-
K				-	0.01
		Ca 48.7	48.8		
En	91.0	Mg 48.4	47.9		
Fs	9.0	Fe 2.9	3.3		

n.a. - not analyzed

Sample location and lithology:

1. Miraflores, borehole PE-2 25.9 m, harzburgite (average of 4 analyses)
2. Miraflores, borehole PE-2 25.9 m, harzburgite (average of 2 analyses)
3. Miraflores, borehole PE-2 41.5 m, harzburgite (average of 4 analyses)
4. Amores mine, massive ore (1 analysis)
5. Monte Bueno, schlieren ore (1 analysis)

←Fig. 7
Platinum-group mineral in chromite, Monte Bueno deposit: a) Fe, K_α X-ray image; b) Ru, L_α X-ray image; c) Ir, M_α X-ray image; d) Os, M_α X-ray image. The lengths of the bars are 10 μm

Table 3
Microprobe analyses of olivines and uvarovites (uv)

	1	2	3	4	5	6
	olivines					uv
SiO_2	40.60	40.64	41.39	42.17	42.41	37.88
FeO	8.98	9.06	8.24	5.57	4.79	n.a.
MgO	48.94	48.76	50.45	52.18	53.08	n.a.
MnO	0.13	0.11	0.11	0.08	0.07	n.a.
CaO	0.01	0.01	0.01	0.06	0.03	36.32
NiO	0.34	0.29	0.37	–	0.54	n.a.
Cr_2O_3	–	–	–	0.10	–	12.31
Al_2O_3	n.a.	n.a.	n.a.	n.a.	n.a.	12.75
Total	99.00	98.87	100.57	100.16	100.92	99.26
	Cations per 4 oxygens					12 ox*
Si	1.002	1.005	1.002	1.010	1.007	2.994
Fe	0.185	0.187	0.166	0.112	0.095	–
Mg	1.800	1.796	1.820	1.864	1.878	–
Mn	0.002	0.002	0.002	0.002	0.001	–
Ca	0.001	0.001	0.001	0.001	0.001	3.076
Ni	0.006	0.005	0.006	–	0.010	–
Cr	–	–	–	0.002	–	0.769
Al	–	–	–	–	–	1.187
Fo	90.56	90.47	91.52	94.30	95.10	
Fa	9.44	9.53	8.48	5.70	4.90	

n.a. - not analyzed

* Cations per 12 oxygens

Sample location and lithology:

1. Miraflores, borehole PE-2 25.9 m, harzburgite (average of 12 analyses)
2. Miraflores, borehole PE-2 41.5 m, harzburgite (average of 4 analyses)
3. Miraflores, borehole PE-2 466.0 m, harzburgite (average of 8 analyses)
4. Amores mine, massive ore (average of 2 analyses)
5. Mercedita mine, massive ore (average of 7 analyses)
6. La Melba, massive ore (average of 3 analyses)

chamber, within the mantle, in "mini chambers" (Brown 1980) or "cavities" (Lago et al. 1982). Thus, the persistence of dunitic halos could be explained by a fractionation sequence olivine–chromite. The Cr–Al reciprocal relationship was interpreted as a result of gradual enrichment in Al and impoverishment in Cr of the magma as olivine and chromite were precipitated (Brown 1980).

According to this assumption, the different positions of podiform ore-bodies with respect to the enclosing rocks could also be elucidated. Parallel and subparallel deposits are likely to have formed in the main chamber and in small pockets below the former, while columnar and other discordant ore-bodies have probably segregated along the path of the ascending magma.

Podiform chromite deposits tend to be confined to certain lithostratigraphic levels at several localities (e.g., Peters and Kramers 1974; Leblanc and Violette 1983) as in the Sagua–Baracoa area. This feature is generally interpreted as a consequence of abrupt changes of oxygen fugacity and/or pressure in the magma chamber (e.g., Roberts 1988). The latter process might cause indented nodules in the Yarey deposit; however, it seems likely that fluctuations in oxygen fugacity are also, if not mainly, responsible for enhanced chromite precipitation in a certain stage during the cumulus process (Peters and Kramer 1974; Watkinson and Mainwaring 1980).

The reason for the lack of correlation between the chemical composition of the ores and their stratigraphic position which may also occur (Brown 1980; Leblanc and Violette 1983) probably lies in the recurring character of the partial melting process and the fluctuation in the equilibrium conditions (Greenbaum 1977) which certainly cannot remain permanent in such a dynamic zone as a constructive plate margin. Thus, consecutive batches of magma of slightly differing bulk composition (Gass 1980) ascend in a narrow zone between the divergent lithospheric plates. As the temperature and pressure decrease upward, fractional crystallization begins at a certain depth depending on the magma composition, temperature, pressure, oxygen fugacity, etc. Primitive magma yields Cr-rich chromite as well as Mg-rich olivine (and, occasionally, platinum-group minerals) which may be precipitated below the oceanic crust, in small pockets. Therefore, somewhat differentiated melt reaches the main magma chamber, resulting in more aluminous chromite. There are numerous indications that the appearance of Al-rich chromite is roughly coeval with that of clinopyroxene and plagioclase which implies a genetic relation (Leblanc and Violette 1983). The reduced ratio of ultramafic cumulates to mafic suite, as well as high proportions of low-Cr, high-Al chromites suggest that the degree of partial melting was lower than in other ore districts.

Formerly it was thought that the site of chromite segregation was a mid-ocean ridge setting (e.g., Dickey 1975; Brown 1980). However, Miyashiro (1973, 1975), Pearce (1975) and Dewey (1976) had serious doubts about the midoceanic origin of some ophiolites which were subsequently ascribed to island arc environments (Hawkins 1980; Pearce et al. 1984). Roberts (1988) concluded that generation of the Tethyan chromites was more likely in marginal (fore-arc or back-arc) basin spreading zones. This inference may be valid also for the chromites of Cuba, as recent paleogeographic reconstructions have implied that the Cuban ophiolites represent an oceanic crust formed in a small ocean basin (Iturralde-Vinent 1988; Ross and Scotese 1988) which opened in the Middle Jurassic between Yucatan and South America. This Proto-Caribbean basin was generated by rifting in a NE–SW trending ridge, apparently independent of subduction processes. Nevertheless, the spreading centre was situated in the mantle wedge above the northeastward subducted Farallon plate, behind the

Proto-Greater Antilles island arc (Ross and Scotese 1988). This may explain the above-mentioned ambiguity of chemical compositions of the diabase dykes and pillow basalts of the study area.

Acknowledgements

We gratefully acknowledge P. Gyarmati, Zs. Peregi (Geological Institute of Hungary) and L. González (Expedición Geológica de Santiago de Cuba) for their support and assistance during field work. We also thank G. Dobosi (Laboratory for Geochemical Research, Hungarian Academy of Sciences) who performed a notable part of the microprobe analyses.

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Conference Reports

Acta Geologica Hungarica Advisory Board Meeting

On 2 and 3 June, respectively, a two days' Advisory Board Meeting of *Acta Geologica Hungarica* was organized by the Editorial Board in Budapest, Hungary. The meeting was accompanied by the "Tectonostratigraphic Terrane Maps of the Circum-Pannonian Region" Workshop, as several members of the Advisory Board are involved in the activity of the Editorial Board of the Maps, as well. On 2 June, in the morning, discussions on the Terrane Maps were carried out at the Department of Geology, Eötvös Loránd University of Sciences with the participation of Birkenmajer, K. (Poland); Hay, W. (USA - Germany); Karamata, S., Krstic, B. (Yugoslavia); Kovác, M., Michalik, J., Plasienka, D., Vozár, J., Vozárová, A. (Slovakia); Mioc, P. (Slovenia); Pamic, J., Tari-Kovacic, V. (Croatia) and rkai P., Buda Gy., Haas J., Kovács S., Lelkes-Felvári Gy., Márton E., Nagymarosy A., Szederkényi T. (Hungary). In the afternoon, a series of lectures began in the Szabó József Room, Eötvös Loránd University of Sciences, which was continued next morning in the Conference Room, Hungarian Geological Survey. The presentations included:

S. Karamata, M. Dimitrijevic, M. N. Dimitrijevic, M. Mirkovic, A. Popevic: Cryptic suture zones as possible proof of terrane boundaries. Example in the Moraca river gorge,

J. Pamic, I. Gusic, V. Jelaska: Geodynamic evolution of the central parts of the Dinarides,

P. Mioc: Relation between the Southern Alps and the Eastern Alps in Slovenia,

V. Tari-Kovacic: Late Mesozoic and Early Cenozoic Tectogenic Deposits and Geodynamics of the South Pannonian Basin and Northern Dinarides,

K. Birkenmajer: Evolution trends in basinal Jurassic -- Early Cretaceous. Examples from the Western Carpathians and Eastern Alps,

J. Michalik: Synsedimentary tectonics of the Paleoalpine cycle and the Mesozoic paleogeographic evolution of the Western Carpathians,

D. Plasienka: Cretaceous tectonochronology of the Central Western Carpathians, Slovakia,

W. Hay: Climate and erosion,

M. Kovác: Development of the Western Carpathian basins during the Neogene,

Series of presentations on the evolution of the Hungarian parts of the Pelsonia and Tisia Terranes:

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Árkai P., Lelkes-Felvári Gy., Kovács S., Szederkényi T.: Variscan and Late Variscan evolution,
Buda Gy.: Correlation of Central European Variscan granitoids,
Haas J., Kovács S., Török Á.: Triassic evolution,
Galácz A., Szente I., Vörös A.: Jurassic evolution,
Császár G., Haas J., Árkai P.: Cretaceous evolution (sedimentary and metamorphic),
Márton E.: Carboniferous–Early Miocene positions and moments,
Fodor L., Csontos L., Nagymarosy A.: Paleogene–Early Miocene dispersions and accretions.

On 3 June, afternoon, a round-table conference for the members of the Editorial Board and the International Advisory Board of *Acta Geologica Hungarica* was held on the present and future of the quarterly in the Board Room of the Hungarian Geological Survey. The meeting was attended by Birkenmajer, K. (Poland); Czurda, K. (Germany); Hay, W. (USA – Germany); Karamata, S. (Yugoslavia); Kovác, M. (Slovakia); Pamic, J. (Croatia)[Advisory Board] and Bárdossy Gy., Császár G., Dudich E., Haas J., Hámor G., Márton E., Szederkényi T. (Hungary)[Editorial Board]. Main topics of the discussion were subject-matter and professional level of *Acta Geologica Hungarica* (spectrum of papers, thematic volumes, regional problems); appearance of *Acta Geologica Hungarica* (format, quality of printing, etc.); permanent columns; application for the *Science Citation Index*.

First, János Haas, Editor-in-Chief, gave an account of the present state of the quarterly, emphasizing the stages of development. Since 1990 there has been computer-aided editing, and camera-ready materials have been sent to the Publishing House; since 1992 Henry Lieberman (USA) has been the Language Editor; since 1992 an International Advisory Board has been acting (K. Birkenmajer, Poland; M. Bleahu, Romania; P. Faupl, Austria; M. Gaetani, Italy; S. Karamata, Yugoslavia; M. Kovac, Slovakia; J. Pamic, Croatia); in 1992 the Editorial Board made wide-scale international advertising by handbills; since 1993 the inner pages have been printed on higher quality art paper; since 1994 *Acta* has been issued with a coated cover; since 1994 correspondence has been carried out by e-mail; in 1994 József Fülöp, Chairman of the Editorial Board died and György Bárdossy was appointed by the Earth Sciences Department of the Hungarian Academy of Sciences to be the new Chairman; in 1996 the Editorial Board was widened by Endre Dudich, Béla Kleb and Emï Márton; and in the same year the International Advisory Board was widened by Kurt Czurda (Germany), Jean Dercourt (France), Spela Gorican (Slovenia), Fritz Steininger (Austria) and William Hay (USA). New columns (such as Theses, Commission News, Books, Maps, Software, Letters & Discussion, Conference Reports, Anniversaires, Obituaries, Calendar – Forthcoming Events) have been introduced. In the future, thematic numbers on the Mecsek–Villány Triassic, the Pannonian Basin and Northern

Hungary, as well as a supplement on the Workshop on Magmatic Events in Rifted Basins, are planned. Distribution of the presented manuscripts regarding the subject and origin since 1991 was displayed for the participants of the meeting (see Tables I and II). Members of the Advisory Board were requested to encourage the authors in their countries to submit manuscripts fitting into the special field of Acta Geologica Hungarica.

Next, György Bárdossy, Chairman of the Editorial Board, spoke about the siginificance of getting a journal into the *Science Citation Index*, about the many factors relevant in selecting a journal for the Index and the efforts made in this respect. In 1995, the Editorial Board sent an application to the Institute for Scientific Information. After repeated applications, the Institute specified the requirements. Overall content, format, citation data and adherence to the publication schedule are among the guidelines used to evaluate a journal. By the beginning of 1997, *Acta* had fulfilled the condition of being published in time, i.e. has practically caught up with the delay in publication. Number 40/1 (1997) appeared at the beginning of this year. The quarterly still has to step up the professional standard of the papers. In January, the last number of *Acta* was sent to the ISI for study. Members of the International Advisory Board were asked to promote this effort by their personal relationships.

During the discussion which followed, the participants emphasized the importance of papers on tectonics, environmental geology, hydrogeology and engineering geology and overviews of the most progressive fields of geology in order to become one of the leading journals in the region. As the quarterly covers the geology of the entire Circum-Pannonian Region, a proposal was made for changing its title (e.g. *Acta Geologica Pannonica*) or, at least, complete it with a subtitle like "Journal of the Circum-Pannonian Region", "International Journal of Geology" or "International Journal of Earth Sciences". A change in the format of the journal (larger size) was also suggested.

Members of the Editorial Board were grateful for the contributions of the international authorities on geology. The possibility of realizing their valuable suggestions will be discussed during the next meeting of the Editorial Board.

János Haas and Gábor Schmiedl

Report on the 5th regional scientific meeting of IGCP representatives of the East-Alpine–Adriatic–Carpathian Region and neighbouring countries

Budapest, Hungary, 4–6 June, 1997

The Regional Meeting was held in the Hungarian Geologic Institute (1143 Budapest, Stefánia út 14.). Representatives from 12 countries attended the meeting (Austria, Bulgaria, Croatia, Czech Republic, Germany, Macedonia, Poland, Romania, Slovakia, Slovenia, Ukraine, Yugoslavia).

After the opening addresses of Károly Brezsnyánszky, chairman of IGCP NC of Hungary and director of the Hungarian Geologic Institute, Béla Köpeczi, chairman of the Hungarian Committee for UNESCO, György Bárdossy, chairman of the Hungarian NC for IUGS, Vladislav Babuska, IGCP Secretary (UNESCO), Paris, conveyed his thanks to Dr. Károly Brezsnyánszky and Dr. Márta Polgári for organizing this regional meeting, as well as to Prof. Mihály Rózsa, Secretary of the Hungarian National Committee for UNESCO, for the support for the IGCP activities in Hungary. He informed the participants about the present state and future perspective of IGCP, as well as about the difficult financial situation following the withdrawal of the United Kingdom funds and the reduction of the USA contribution from USD 80.000 to 35.000 in 1997.

The Programme has just undergone its third external evolution in 10 years which concluded that the IGCP should continue as a core activity of the UNESCO's Division of Earth Sciences with a focus on those geologic processes that have an impact on human living conditions, affect the global environment and which are concerned with the wise use of natural resources.

After the short personnel comments of every representative, two scientific reports were presented by the Hungarian co-leaders of IGCP projects No. 356 and 384.

Then the short summary reports on IGCP activity of the countries were presented by the representatives.

12 new project proposals and announcements were discussed, namely:

– *Correlation of Metamorphism from Central Europe to Western Asia: Alpine and Prealpine Metamorphic Evolution from the Alps to the Pontids* – proposed by V. Höckl, Austria

– *Sustainable water supply from deep aquifers for major cities in young basins* – proposed by W. Janoschek, Austria

– *Sedimentology, Correlation and Stochastic Modelling of Turbidite Sequences (Basin Analysis and Petroleum Potential Estimation Approach)* – proposed by N. Kontopoluos and A. Zelilidis, Greece; V.Vushev, Bulgaria; A. Colella, Italy

– *Organic matter in major environmental issues* – proposed by J. Pasava, Czech Republic

– *Geodynamic evolution and metallogeny on the Serbo–Macedonian–Anatolian metallogenic province* – proposed by T. Serafimovski, Macedonia

– *Magmatism, metamorphism and metallogeny in the Vardar Zone and Serbo-Macedonian Massif* – proposed by T. Serafimovski, Macedonia

– *Asthenosphere-lithosphere dynamics of Tethyan closure* – proposed b M.F. Flower (USA); V.Mocanu, Romania, Nguen Trong Yem, Viet Nam

– *Magnesites: from the origin to environmental impact* – proposed by D. Hovorka, Slovakia

– *Paleoclimate and paleoenvironmental warmth in boundary Miocene-Pliocene (and others); scenario in the future* – proposed by V. Semenenko,Ukraine

– *Sedimentation cycles in Paratethys and its mineral resources* – proposed by T. Pinchuk, Russia

Resubmissions after careful revision:

1) Geologic sources of Hg, As and Sb and their impact on the environment by ore manufacturing – proposed by S. Karamata, Yugoslavia; N. Ozerova, Russia (IGCP Pr. proposal No. is: 416)

2) Rock and Minerals at Great Depth and on Surface – proposed by F.P. Mitrafanov and D.M. Guberman, Russia; E. Dudich, Hungary (IGCP Pr. proposal No. is: 408)

Gábor Gaál, vice director of IUGS, summarized the critical comments of the new project proposals and announcements and emphasized that a good project proposal must be on a global scale or a general scientific problem, and needs the guarantee of good laboratories on an international level. Small regional project proposals have no chance of acceptance. All proposals were discussed by the representatives.

V. Babuska, Secretary of IGCP, summarized the results of the comments discussed openly and said that the tradition of the region is very good, and that after a less active period it seems that more proposals were being made again. Because of the financial restriction competition is stronger. The Regional Meeting finally supports the two resubmissions and the underlined project proposals for submission, and encourages the other proposals and ideas to be carefully revised according to the critical comments.

J. Pasava informed the representatives on the international conference of IGCP project 373 which will be held in Perslák, Czech Republic, September 21st - September 24th, 1998.

Carpathian-Balkan Geologic Association

W. Janoschek gave some brief information about the CBGA. He invited the representatives of IGCP projects to a close co-operation with the commissions of the CBGA. He informed the participants that the next (16th) Congress will take place in Vienna, August 30th - September 2nd, 1998, and presented the objectives of the Congress and the preliminary programme. He cordially extended an invitation to the Congress.

International Geologic Congress, 2004

W. Janoschek informed the IGCP representatives about the plan of the Austrian National Committee of Geology to invite the IGC to be held in Vienna August 8th - August 18th 2004. He briefly informed on the activities so far of the Austrian National Committee of Geology.

The final decision should be made at the next IGC 2000 in Brazil, but he cordially asked the representatives present to speak in favour of Vienna and Austria as host of the IGC 2004 and to support the Austrian applications.

V. Babuska, Secretary of IGCP, in his closing words found the Regional Meeting very successful, constructive and useful, and emphasized again the importance of showing the importance of IGCP to society. He again thanked the organizing committee for their work, for the active participation of the representatives of the national committees for IGCP, for the ideas which were discussed here, and for the many project proposals.

K. Brezsnyánszky, chairman of Hungarian National Committee for IGCP and director of the Hungarian Geologic Institute, closed the Regional Meeting, stressing that we were very lucky that this meeting was attended by V. Babuska, Secretary of IGCP Paris, by the previous Secretary E. Dudich, by G. Gaál, vice Chairman of IUGS, by Alexandra Partan, the Secretary of Austrian National Committee for IGCP, who organized the Regional Meeting twice, and whose experience helped a lot to prepare this Meeting. We thank her very much for her help. We are satisfied and hope that some of the proposed projects will be accepted by the Board. All the representatives are encouraged by the Geologic Institute co-operate further scientifically.

Márta Polgári

Report on the Sequence Stratigraphic Workshop for the CBGA member countries

Sümeg, Hungary, June 9–14, 1997

The workshop was organised by the Hungarian Stratigraphic and the Sedimentological Commissions and was held at the geological training base in Sümeg where bungalows served for accommodation and two stone houses for presentation of papers, for meal, informal talk and discussions.

The idea to arrange a sequence stratigraphic workshop arose in Athens at the CBGA congress in 1995, when it was noticed that the subject is underrepresented among the papers. The notion was supported by both the President of the Association (W. Janoschek) and the General Secretary of the IAS (A. Strasser). The aim of this meeting was to promote a deepening the knowledge of the discipline in the Carpathian-Balkan region with the help of outstanding keynote lecturers: A. Strasser for carbonate sequences, F. Surlyk for siliciclastic sequences and H. Leereveld for biota (their significance in sequence analysis).

The workshop was attended by 51 geoscientists (mainly youngsters) from the following countries: Austria, Croatia, Hungary, Poland, Romania, Slovakia, Slovenia and Ukraine.

The presentation of participants were preceded by a keynote lecture and a case study of the actual subject every day. In order to stimulate discussion an hour was devoted for each paper for presentation and discussion. Including keynote lectures and case studies all together 23 papers were presented. They included the stratigraphic interval from the base of the Triassic up to the top of the Pliocene. About half of the papers dealt with siliciclastic sequences and the other half with carbonate or mixed sequences. Several papers provided examples for usage of distribution of fossils in the distinction of sequences or cycles. The application of seismic methods to sequence analysis was also demonstrated by Neogene case studies.

Although the participants were active in pointing out uncertainties and clearing up situations, in accordance with the expectations in several questions the notes or comments of the keynote lecturers were fundamental. It turned out during the discussions that the selection of keynote lecturers was fortunate because their approaches were often different and their discussions offered a good opportunity to obtain an insight into the complexity of sequence stratigraphy.

After a three-day indoor meeting the workshop ended with an excursion in the Balaton Highland where the Triassic sequence stratigraphic phenomena were shown in stratigraphic order.

At the end of the workshop the participants expressed their wishes to continue this initiation which could not be realised without the financial aid of IAS and the kind contributions of the keynote lecturers.

Géza Császár and János Haas

Report on the final meeting of the DANREG Programme

Budapest, May 26–30, 1997

From the point of view of geological activity the Danube acts as a border separating countries where methods and approaches in geologic research are different therefore the results of researches are not compatible. On the other hand the Danube binds these countries tightly because it is as the main water collector of a broad region; it also collects pollutants via tributaries along its long course. The experiences of the past decades convinced some countries that a joint effort is required for the elimination, mitigation or even prevention of environmental damages of the area. This conviction manifested itself in an agreement first signed in 1990 by the Hungarian and Slovak, and later by the Hungarian, Austrian and Slovak geological authorities, aiming a co-ordinated joint work along the Danube between Vienna and Budapest.

In order to support the conclusions concerning the environment which are to be expressed in an applied geologic map, the following maps were compiled: surface geologic map, Quaternary thickness and lithogenetic map, Pannonian thickness and lithofacies map, Pontian to Pliocene thickness and lithofacies map, pre-Tertiary basement map, neotectonic map, tectonic map, gravity map of the Bouguer anomalies, map of the DT magnetic anomalies and three geologic-geophysical cross-sections. For the decision makers the engineering-geologic map, the geothermal potential map, the hydrogeologic map, the interpreted resistivity maps and the environmental hazards map help in land use planning. In addition to the explanatory notes for each map a joint study on water quality permits the drawing of conclusions.

Experts of the following institutions have contributed to the realisation of the work. From the Austrian side the Geological Survey of Austria, from Slovakia the Geological Survey of the Slovak Republic and Geocomplex, and from the Hungarian side the Geologic Institute of Hungary and the Eötvös Loránd Geophysical Institute of Hungary. The meeting was addressed by Caspar Einem, minister of Science and Transport of Austria by a letter presented by the cultural attaché.

After a seven-year work period the Hungarian Geologic Institute was the host of the final (but 1st joint) meeting of the working groups, May 26 to 27, 1997. The aim of this meeting was to give participants of the project a chance to be acquainted with other maps, call out notes and discuss unsolved questions. The maps were digitally prepared colour prints, in part in united form, in part still independent national map sheets. During the meeting devoted to the presentation and discussion of papers and posters introducing the maps and studies, it became obvious that several questions of practical and a few of principal importance require further efforts to be solved.

The meeting was also attended by experts of some other countries out of the DANREG area (Albania, Germany, Macedonia and Yugoslavia). Being conscious of the importance of this kind of joint work they suggested to broaden the activity over the total length of the Danube or geographically into other areas too.

The indoor meeting was followed by a three-day field trip in Hungary, Austria and Slovakia where the participants could view a cross-section of the geology-related significance of the area and via further discussion they get closer to a compromise in the open questions

The organisers are grateful to the National Technical and Scientific Commission for the financial support that promoted the success of the meeting.

Géza Császár

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GUIDELINES FOR AUTHORS

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

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

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

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Figures and tables should be referred to in the text. Figures are expected in the size of the final type-area of the quarterly (12.6 x 18.6) or proportionally magnified 20–25% camera ready quality. Figures should be clear line drawings or good quality black-and-white photographic prints. Colour photographs will also be accepted, but the extra cost of reproduction in colour must be borne by the authors (in 1997 US$ 260 per page). The author's name and figure number should be indicated on the back of each figure. Tables should be typed on separate sheets with a number.

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Acta Geologica Hungarica, Vol. 40/4, p. 365 (1997)

Preface to the "Hungarian Muschelkalk" special volume of Acta Geologica Hungarica

The occurrence of Germano-type Triassic sequences outside of the German basin (sensu stricto) is well known. French, Spanish, Swiss, and Bulgarian Muschelkalk sediments belong to this series; however, the presence of Muschelkalk-type sequences within the Carpathian system in the Pannonian Basin has received less publicity. This volume attempts to bridge this gap and intends to put a spotlight on Hungarian occurrences of Muschelkalk carbonates. Our knowledge is far from that of the classical German Muschelkalk, about which very detailed descriptions have been published. This collection of papers is considered to be the first step toward that stage of understanding. Hopefully, in the future our knowledge of the Hungarian Muschelkalk will be at a comparable level. A classical example to follow is the Muschelkalk volume of Schöntaler Symposium, which was initiated and realised by the untiring work of Hans Hagdorn and his collaborators Theo Simon and Joachim Szulc.

In our approach to the southern Hungarian Triassic the first milestone was the monograph of the Mecsek Triassic by E. Nagy (1968). Since then several papers were published including the monograph of the Villány Triassic (Nagy E. and Nagy I. 1976). These publications principally described the lithology and palaeontology of Triassic sequences giving further hints to the German analogy. In the past years, due to our increased knowledge and interest in palaeogeographic reconstruction, the question of German analogy became the most crucial problem of the southern Hungarian Triassic. This collection of papers therefore intends to show the similarities of Hungarian Muschelkalk sequences to Germano-type Muschelkalk on one side and tries to give a review of our present knowledge of these sequences on the other.

In this volume only the first series of the latest results are given; as a continuation of this work additional papers now under preparation will be published in the Acta Geologica Hungarica, dealing with the lithostratigraphic and structural geological revision of the Triassic, synsedimentary tectonism, the conodont fauna of the Muschelkalk, and the description of Buntsandstein clastics, Röt sequences, etc. We hope that these results will be of use to many Triassic researchers. On behalf of the authors I wish to express our thanks to the scientific and technical editors of the Acta Geologica Hungarica for providing us with a forum and we are especially grateful for their patience.

Ákos Török

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Acta Geologica Hungarica, Vol. 40/4, pp. 367–390 (1997)

Triassic ramp evolution in Southern Hungary and its similarities to the Germano-type Triassic

Ákos Török
Budapest Technical University, Department of Engineering Geology

The Muschelkalk of southern Hungary consists of Wellenkalk carbonates, Coenothyris beds and ooidal, dolomitised sequences. The depositional environment was a homoclinal carbonate ramp, which succeeded a clastics covered terrain of the Early Triassic pre-ramp phase. Major sediment reworking processes include storm activity and slumping on the gentle slopes. The lithotypes and sedimentary structures are very similar to those of the epicontinental of the German Triassic sequences, especially to the Silesian ones. The differences are probably related to the shallower water depth, more significant restriction and the different palaeolatitude. The thick Middle Muschelkalk evaporites and a part of the German Upper Muschelkalk appear to be missing in Hungary. In the Late Triassic the ramp became covered by clastics (clastics buried ramp phase).

Key words: ramp, carbonate sedimentology, Triassic, Muschelkalk, Southern Hungary

1. *Introduction*

The depositional processes of the epicontinental German Basin are well understood and based on the detailed sedimentary analysis of the Muschelkalk (Aigner 1985; Aigner et al. 1979; Hagdorn 1982, 1991; Szulc et al. 1990; Szulc 1993 and many others). Besides that of the German Basin Muschelkalk sequences are also known and described from outside the German Basin, f.i. from Spain, from France (Virgili 1958; Calvet and Tucker 1988; Calvet et al. 1990; Bodurov et al. 1993) and from Israel (Hirsch 1976). The Hungarian Muschelkalk-type carbonate sequence is less known internationally. It differs from the classical epicontinental sites in some aspects; i) its isolated position between typical alpine sequences, ii) its provenance, boundaries and Triassic palaeogeographic position are uncertain. For the description of the formations and the analogy to the German Triassic a detailed sedimentary and facies analysis were carried out. Better understanding the evolution of the Hungarian Muschelkalk carbonate ramp may help us to clarify the analogies and differences with the classical German Triassic and Germano-type ("German-like") Triassic facies types in southern Hungary. Besides the description of sediments and the interpretation of their depositional environment this work points out the Germanic features of the southern Hungarian Triassic and provides a correlation between lithostratigraphic units.

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2. *Geologic setting*

The megatectonic structure of Hungary reflects the complex tectonostratigraphic evolution of the Carpathian Basin (Kázmér and Kovács 1985; Fülöp et al. 1987; Balla 1988). Due to the formation of the Alpine–Carpathian system large-scale plate tectonics-related rearrangement of micro blocks took place during the Cenozoic resulting in the juxtaposition of tectonic units. Hundred kilometre-scale lateral displacements were documented along a dextral strike slip zone leading to the tectonic juxtaposition of two contrasting megatectonic units: the Alcapa (Pelso) and the Tisza Megaunits (Kázmér and Kovács 1985; Balla 1988; Csontos et al. 1992). The Alcapa unit and its part of the Transdanubian Central Range, located north of the Mid-Hungarian Lineament, is characterized by South Alpine-type Mesozoic sequences while the Tisza unit south of the lineament comprises Germanic Triassic and a transitional (Alpine–European) Jurassic–Cretaceous (Géczy 1973). Germanic-type Triassic sequences are found on the surface in southern Hungary in the Mecsek and Villány Mountains (Fig. 1). In the Mecsek Mts the Triassic formations are found in the central and western part forming the flanks of an

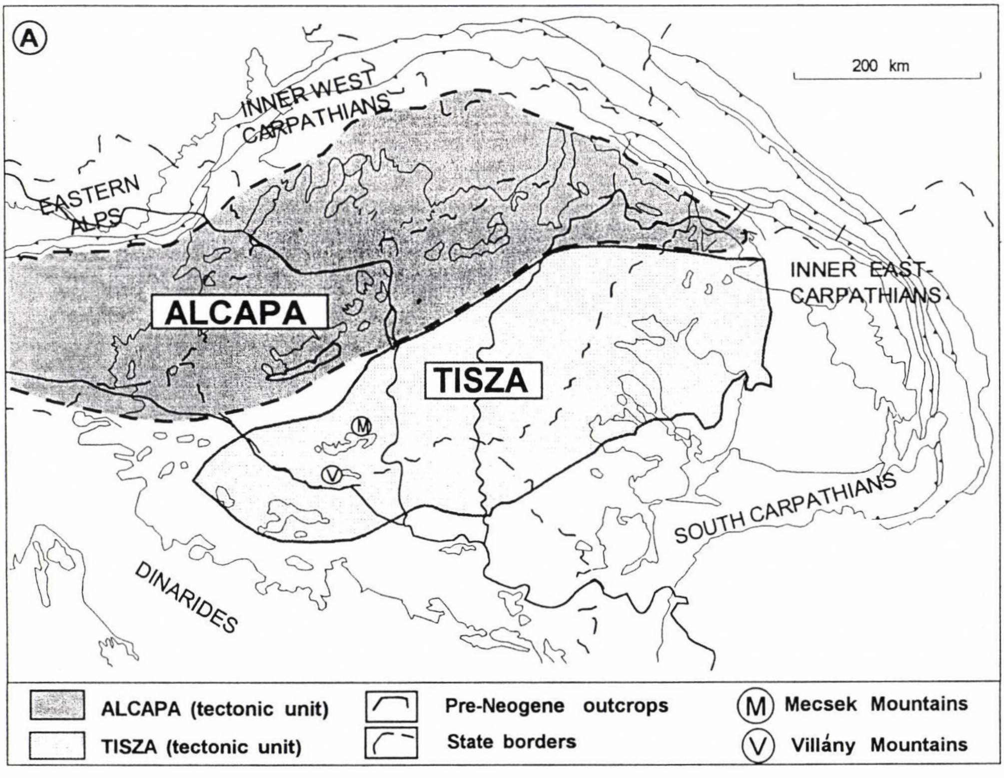

Fig. 1A
Location of the Mecsek and Villány Mountains in the Tisza Megatectonic unit south of the Alpine–Carpathian–Pannonian (ALCAPA) megaunit

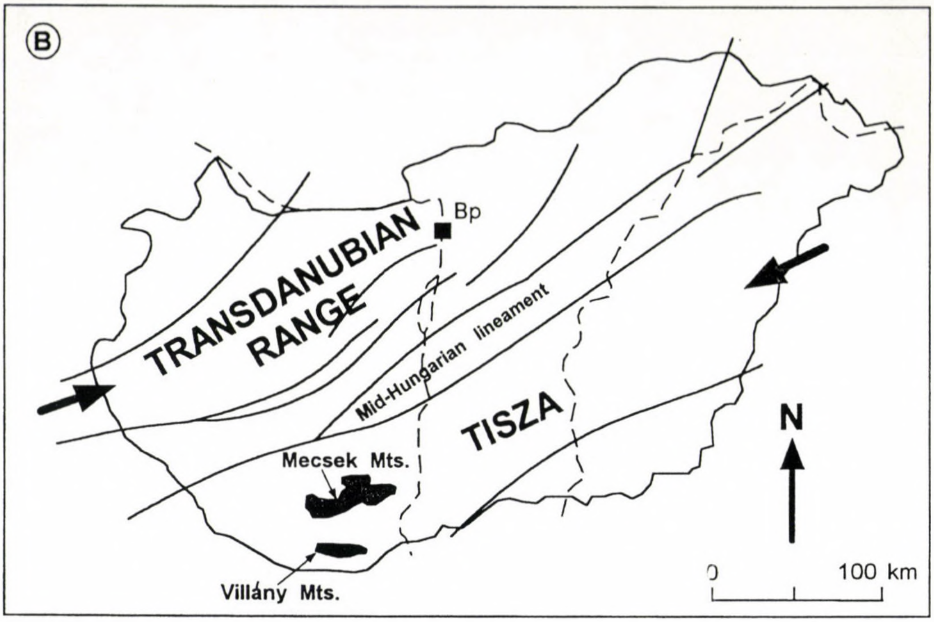

Fig. 1B
Occurrence of the Germanic Triassic in southern Hungary, in the Mecsek and Villány Mountains and its relationship to the Alpine Triassic of the Transdanubian Central range (ALCAPA-Pelso Unit) (Base map according to Császár and Haas 1984; Kázmér 1986 including the structural analysis of Balla 1988). Arrows show the relative displacement of the Tisza unit (to the SW) and the Transdanubian Range (to the NE)

anticline with an E–W axis (Fig. 2). In the Villány Mts the Mesozoic sequences with Triassic carbonates are parts of a complex thrust sheet system containing seven overthrusted NNE-verging micronappes (Rakusz and Strausz 1953; Nagy and Nagy 1976; Császár and Haas 1984) showing progressively different features eastward. Being a part of the Mesozoic basement of the Tisza unit they are covered by thick Tertiary sediments. The easternmost outcrops of this zone comprise slightly different sediments and are found on the surface in the Bihor unit of the Apuseni Mts (Bleahu et al. 1994).

Because of the uncertainties in the determination of the boundaries of the Tisza unit and the scarcity of palaeontological evidence (cf. Balogh 1981) the palaeogeographic reconstruction of subunits and the unit's relationship to the African and European shelf margin has been controversial, especially for the early Mesozoic period. This is clearly manifested in the differences of palaeogeographic reconstructions of the region (Ziegler 1982; Galácz et al. 1985; Kázmér and Kovács 1985; Haas 1987; Tollmann 1987; Vörös et al. 1990, and many others).

The Triassic of southern Hungary shows a characteristic tripartite subdivision: a Lower Triassic clastic unit (Buntsandstein); a Middle Triassic fine-grained clastics-evaporites-dolomites unit (Röt) and a carbonate unit

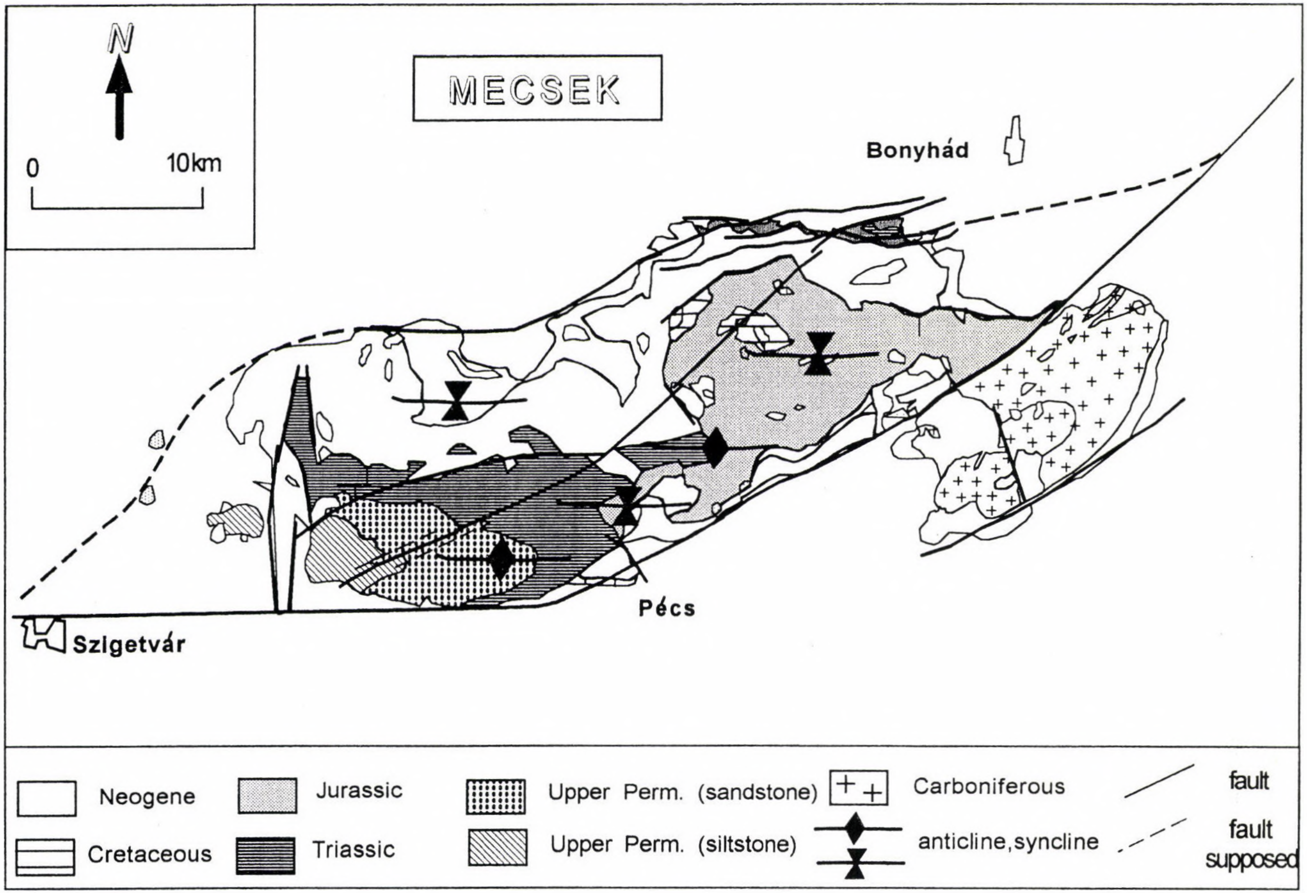

Fig. 2
Geologic map of the Central and Western Mecsek Mountains showing the occurrence of Triassic formations (Modified after Nagy 1968 with structural geological data of Benkovics)

(Muschelkalk) and an Upper Triassic clastic unit (a non-typical Keuper); (Fig. 3). The latter unit shows a different development in the Mecsek and Villány Mts, respectively. It consists mainly of thick (nearly 400 m) clastics in the Mecsek area whereas it is represented by a thin sequence of moderately cemented sandstone, clayey siltstones and dolomites in the Villány Mts. The lithostratigraphic subdivision of the Triassic is based on the outstanding works of Nagy (1968) and Balogh (1981), although later it was slightly modified (Rálisch-Felgenhauer and Török 1993) (Fig. 3). The Triassic is underlain by a Permian continental clastic sequence consisting of red-green conglomerates, sandstones and siltstones with a thickness of over 2000 m in the Mecsek zone. The Lower Triassic is also represented by a conglomerate-sandstone series with large-scale cross-bedding, channel structures, and reworked clasts reflecting a similar depositional environment (continental-fluvial sequence). This sequence gradually passes upward into the earliest Mid-Triassic evaporitic unit showing the first evidence of marine influence and representing the base of the Muschelkalk carbonates (Fig. 3).

3. *Sediments of the Hungarian Röt and Muschelkalk*

Muschelkalk carbonates in southern Hungary succeed an evaporitic sequence of sabkha facies of Röt (Fig. 4). The basal carbonate beds consist of laminated dolomites and intercalating clay and dolomitic marl horizons representing an intertidal, periodically emerged environment (Hetvehely Dolomite; Fig. 3). Calcite filled mudcracks, reworked dolomite clasts and minor contorted (primarily evaporitic) intercalations suggest this environment. Upward the sequence becomes less dolomitic and the dolomites give way to bituminous limestones (Viganvár Limestone). These limestones include several lithotypes, such as laminated limestones, bioturbated limestones (plastoclasts) and very thin and sparse bioclastic beds with small gastropods and pelecypods. Its oligospecific fauna (Nagy 1968) and the bituminous muddy carbonates indicate a restricted and periodically anaerobic high-stress environment, e.g. an intertidal lagoon. This limestone passes upward into a partially dolomitized unit. Dolomicrites with preserved fabric were formed penecontemporaneously from limestones, while coarse crystalline sucrose pervasive dolomites and dolosparitic fissure fillings are related to later processes (i.e. burial diagenesis). Dolomites are overlain by carbonates of the Wellenkalk unit representing advanced marine conditions. Wellenkalk carbonates (Lapis and Tubes Limestone; Fig. 3) consist of several lithotypes with slightly overlapping spatial and temporal distribution. Bioturbated limestones with preserved traces fossils such as *Rhizocorallium, Balanoglossites* and *Thalassinoides* are considered to form in a low-energy regime with muddy bottom conditions. Nodular limestones are composed of micritic limestone nodules of elongated and often deformed shape in a marly limestone matrix. Their plastoclast floatstone microfacies and the segregation of clayey matrix suggest syndepositional deformation which

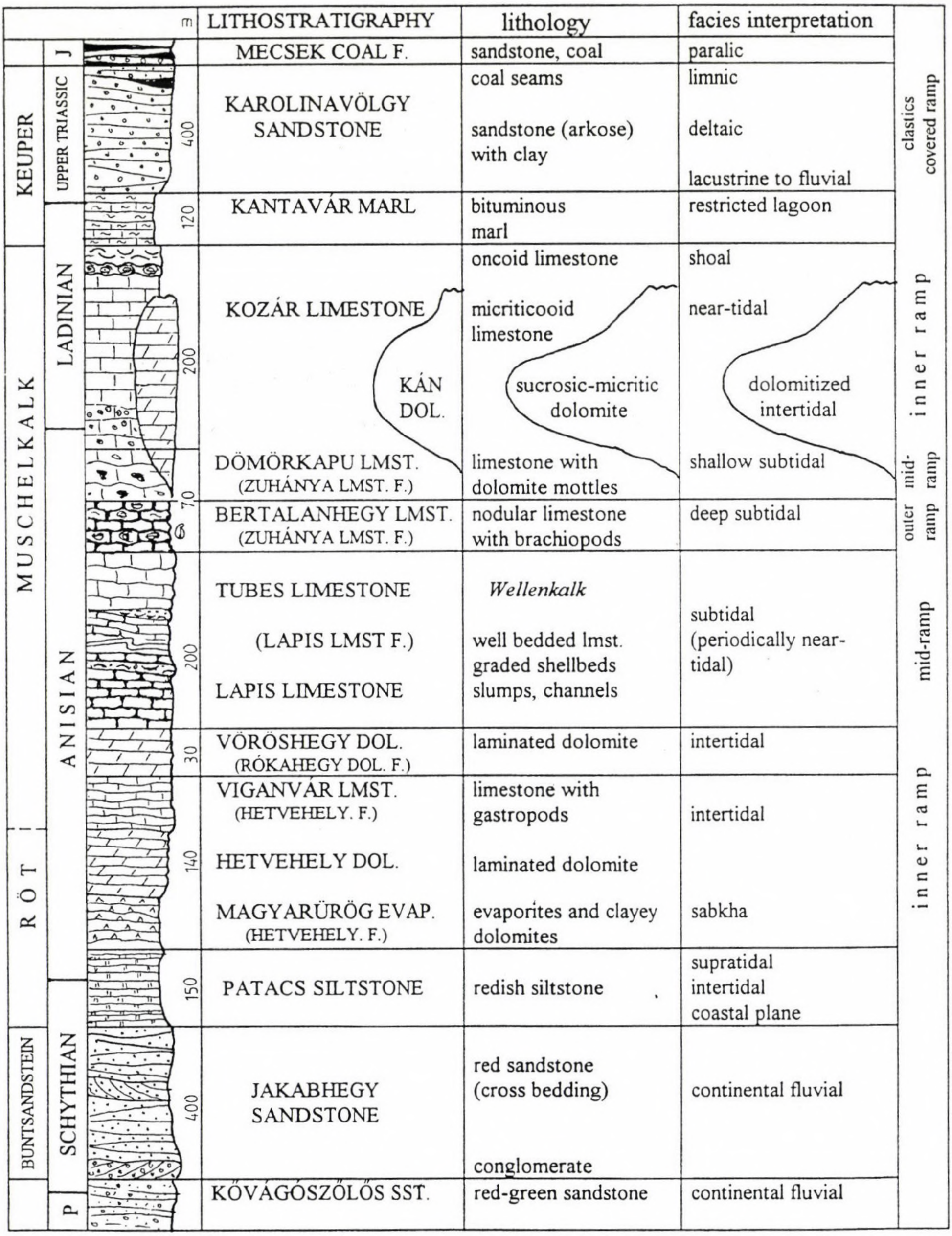

Fig. 3
Lithostratigraphic subdivision of the Mecsek Triassic with lithologies, sedimentary structures and facies interpretation (lithostratigraphy and lithology are modified and compiled from Nagy 1968; Balogh 1981; Rálisch-Felgenhauer and Török 1993 with contribution of Gy. Konrád)

Fig. 4
Ptygmatic folded evaporitic (anhydrite) beds, the Röt (Magyarürög Anhydrite, Gálosfa-1 (GF-1) borehole, core from 1602.6 m – the coin for scale is 2 cm in diameter)

was later enhanced by stylolitization. The elongated shape of some nodules probably suggest a reworked trace fossil (dwelling trace) origin and thus it is considered to be a transitional type toward a highly bioturbated limestone – the "Wurmkalk". Thin-bedded alternated limestones and marlstones are characteristic with well-expressed wavy bedding planes often show ripple marks on the tops and tool marks at the bottom of the beds. The grain size in limestone beds shows only minor variations as a rule, i.e. fining-upward microcycles are visible. The limestone beds, which have a peloid-ostracod mudstone/wackestone microfacies are generally interbedded by much thinner (few mm) marlstone layers giving the typical "Wellenkalk" appearance (Fig. 5). Crumpled horizons and slumps are common in these beds. Storm-generated skeletal beds and lenses occur both in the alternating limestone and marlstone lithofacies and in the bioturbated limestone lithofacies recording the periodic storm activity. These centimetre-thick limestones consist of small pelecypods, gastropods and crinoids in a fining-upward manner. The micro-grading, the erosive base of the bed and the sheet-like appearance suggest a storm-induced current deposition. The base of the beds is often enhanced by stylolitization, while the uneven top surface is in most cases covered by calcareous mud. Besides the sheets this lithotype may form lenticular bodies which can be

Fig. 5
Thin-bedded limestone with intercalating zones of slightly bioturbated nodular limestones. Note the thin marly horizons between limestone beds and the wedging out character (slightly deformed) in the lower middle part of the photograph (Lapis Limestone, Wellenkalk, Bükkösd-Hetvehely quarry – coin for scale in the middle is 2 cm in diameter)

interpreted as channel infills (Fig. 6). Thick to medium-bedded (2 m–0.2 m) limestones are made up of lime mudstones in the upper part of "Wellenkalk" succession. The faintly laminated mudstones contain intercalations of thin wackestone horizons, which represent a more dynamic phase of deposition, i.e. storm activity (Fig. 7). Minor slumps are also present indicating the slopeward displacement of semiconsolidated lime mud. Crinoidal limestones form small lenses rather than beds in the uppermost part of "Wellenkalk". These crinoid-intraclast packstones often show cross-stratification. These lenses are considered to be reworked crinoidal bioherms. Besides crinoids (Hagdorn et al. 1997), foraminifers are also present in this sequence indicating the Pelsonian age of these beds (*Glomospira densa* Pantic). They are overlain by brachiopod-bearing beds ("Coenothyris beds" or Bertalanhegy Limestone; Fig. 3), which are considered to be the most characteristic unit of the Hungarian Muschelkalk and were already recognised as early as the last century (Beudant 1822; Peters 1862; Haidinger 1865; Böckh 1881). Nodular limestones, limestones with calcareous marl intercalations and brachiopod shell beds of this unit represent the "deepest" sediments of the Triassic of southern Hungary. In

Fig. 6
Large-scale channel structure in the Wellenkalk sequence representing the backflow regimes of storm currents (geostrophic currents) (Lapis Limestone, Lapis road-cut – hammer for scale in the lower right)

nodular limestones articulated shells of brachiopods (*Coenothyris vulgaris* Schlotheim, *Tetractinella trigonella* Schlotheim, *Punctospirella fragilis* Schlotheim, *Menzelia menzeli* Dunker, and very rarely *Decurtella decurtata* Girard, *Aulacothyris angusta* Schlotheim and *Koeveskallina koeveskalliensis* Stur), bivalves (*Plagiostoma lineatum* Schlotheim, *P. striatum* Schlotheim, *Hoernesia socialis* Schlotheim, *Enantiostreon difforme* Schlotheim), poorly preserved cephalopods (*Paraceratites binodosus* Hauer (Detre 1973) and conodonts (Kovács and Papsová 1986) were found indicating a deeper ramp setting (for further palaeontological description see Nagy 1968; Pálfy and Török 1992; Török 1993b; Szente 1997). The limestones with calcareous marl intercalations without "diagenetic nodularity" were formed in roughly the same setting as the nodular varieties and probably reflect a deeper ramp setting similarly to Spanish Muschelkalk (Calvet and Tucker 1988).

The brachiopod shell beds are found in two distinct forms indicating the differences in their depositional style. The storm-deposited allochthonous beds contain disarticulated and often fragmented shells of brachiopods and rarely pelecypods. Their convex-up orientation, the high rate of disarticulation, the floatstone microfacies and the erosive base suggest their origin by

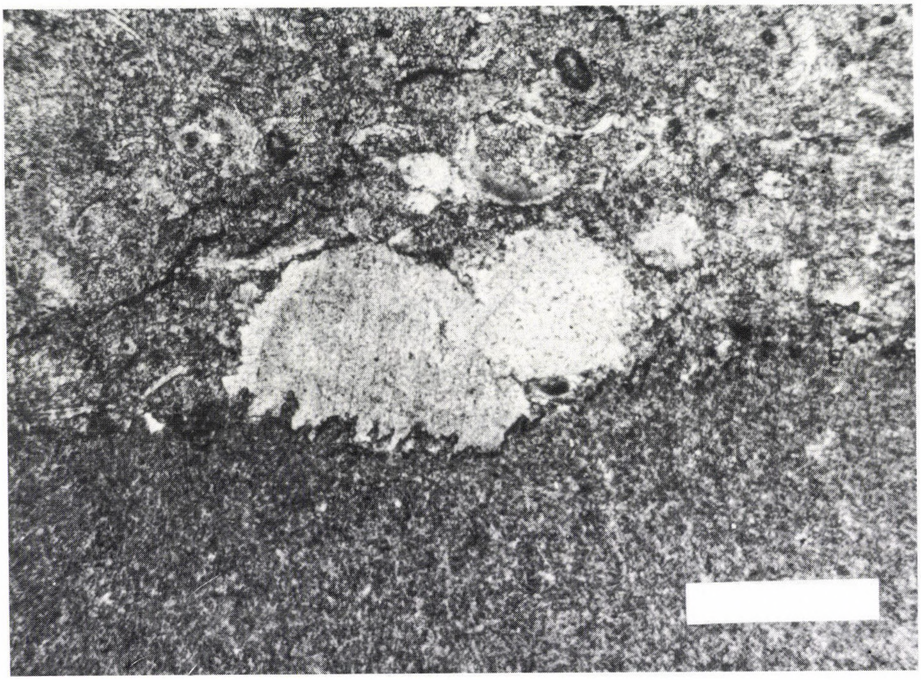

Fig. 7
Microphotograph of the lowermost part of a storm-generated bioclastic bed. The base of the bed is enhanced by stylolitization (Tubes Limestone, Bükkösd, the scale bar is 0.5 mm)

storm-induced current (Fig. 8). On the post-storm surfaces firm grounds and hardgrounds could develop with characteristic forms of *Placunopsis* encrustations and *Trypanites* borings. The composite storm events are indicated by the presence of graded crinoidal-bioclastic beds in the lower part and brachiopod-bivalve floatstone, in the upper part of the same bed. In these cases the graded bioclastic layers represent the full storm phase whereas the covering floatstones signify a rapid sediment fall-out of the storm-waning phase. The parauthochtonous brachiopod beds are also closely related to storms, but they were formed in a deeper ramp setting. They consist of densely packed shells of articulated brachiopods. The internal cavity and the space between the shells are filled with micrite. The open space was later filled by sparitic calcite (geopetal structure – Fig. 9). The remaining open space of the shell interiors suggest a very rapid burial by sediment, i.e. the formation of a sediment cover, which sealed the brachiopod shells and thus hindered the infiltration of further sediments into the shell interiors. The high rate of articulation of shells and the micritic intersediment and intrasediment indicate moderate current velocities and probably indicate a storm-entrained mud deposition (mud flow) in a deeper ramp setting. Brachiopod beds are overlain by the thick bedded

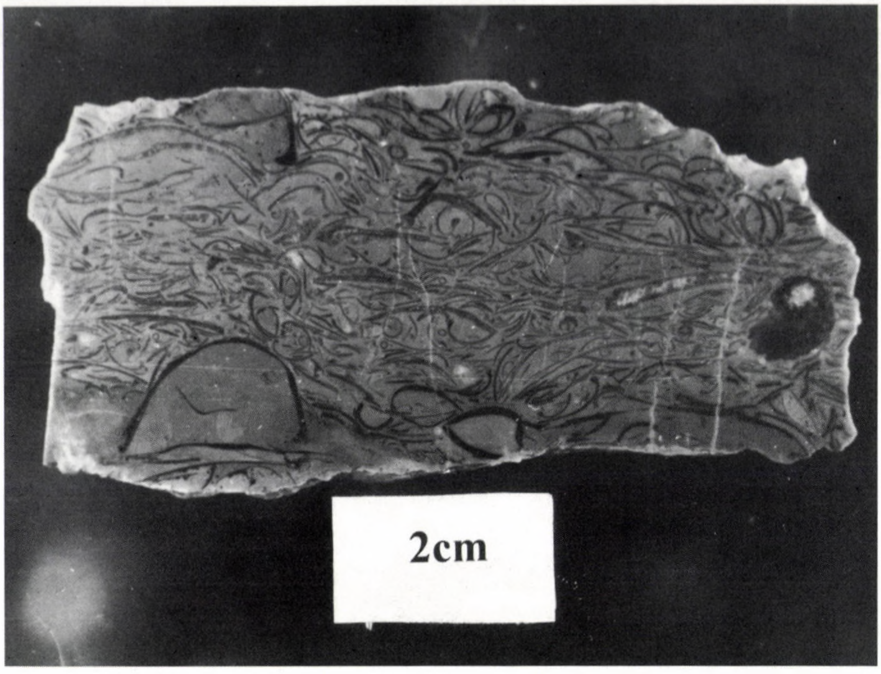

Fig. 8
Brachiopod floatstone of storm-induced origin. Note the dense packing and fragmentation of shells. (Bertalanhegy Limestone, Hetvehely road-cut)

partially dolomitized limestones of the Dömörkapu Limestone (Fig. 3) representing a shallower ramp setting. The dolomite appears in the form of irregular and usually lighter (yellowish) coloured mottles which are related to secondary dolomitisation processes. Upward in the succession it passes to a micritic limestone, which contain lenses of oolitic packstones, probably of shoal origin (Kozár Limestone; Fig. 3). In the back-shoal zone macrooncoids (3–4 cm in diameter – Fig. 10) and bioclastic shelly limestones were formed. This part of the succession is represented by light-grey micritic and pinkish sucrosic dolomites (Kán Dolomite) in the western part of the Mecsek Mts indicating a possible progressivity of dolomitization toward the west (Fig. 3). Back-shoal oncolites are overlain by finely laminated, dark-grey, ostracode-bearing bituminous calcareous marls and limestones of a restricted lagoonal origin (Kantavár Marl). This formation represents the last carbonate member of an upward-shallowing megacycle. The lagoonal marls are covered by arkosic sandstones and siltstones indicating an increased terrigeneous influx and a change of the depositional environment toward a deltaic-lacustrine-fluvial system (Karolinavölgy Sandstone, Fig. 3). In the uppermost Triassic coal seams of limnic origin already appear (Bóna 1984). The Triassic/Jurassic boundary can be drawn in the basal part of the coal formation (Bóna 1984), although it

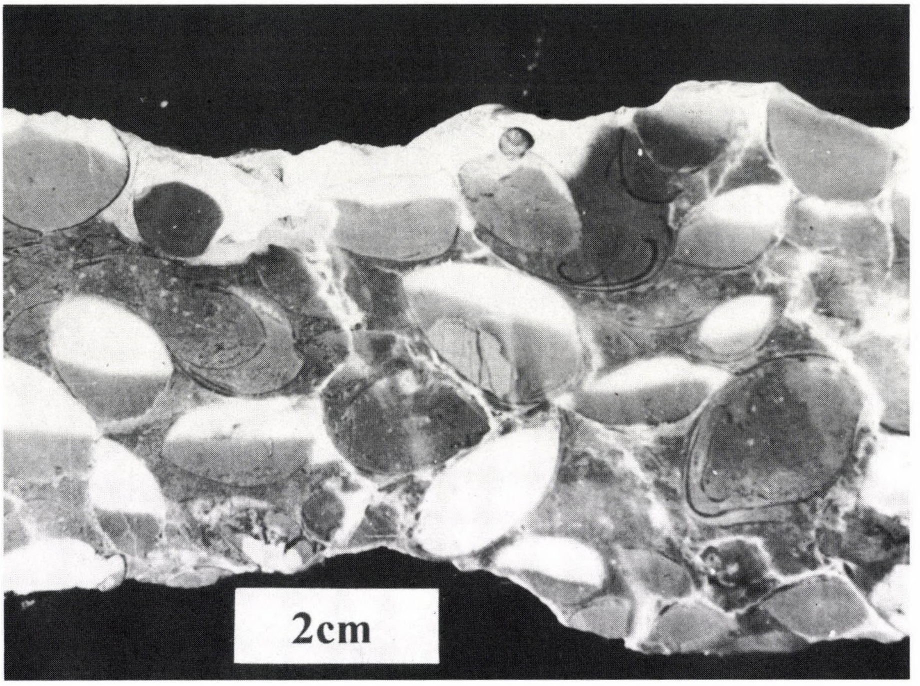

Fig. 9
Parautochtonous brachiopod bed (a storm-induced mud covered community) with articulated brachiopods showing geopetal structures. Note the late white geopetal sparitic calcite cement on the top of micritic intrasediments in the shell interiors. (Bertalanhegy Limestone, Gorica)

is very difficult to determine it in a non-marine sequence with a sparse fossil record.

4. *Facies changes, facies dynamics*

The marginal evaporites and Muschelkalk carbonates cover large areas and apart from the minor lateral facies changes they appear to be relatively uniform on a basin scale; it is therefore assumed that they were formed on a gently sloping homoclinal ramp. The ramp was characterized by a relatively high net sedimentation rate (more than 200 m of evaporites and fine-grained clastics and more than 300 m of carbonates were formed during the ca. 4 Ma period of the Anisian; cf. Fig. 3). Even this high sedimentation rate was overridden during storms when graded skeletal sheets were formed almost instantaneously (Fig. 7). The backflow regimes of storms, the geostrophic currents, the efficiency of which was highly stressed recently (Seilacher and Aigner 1991; Nummedal 1991) produced large channel structures (Fig. 6) with lenticular bioclastic infillings at mid-ramp settings. The extensive synsedimentary deformations, slumps, sigmoidal structures ("S-Faltung"), and crumpled horizons (Fig. 5) were

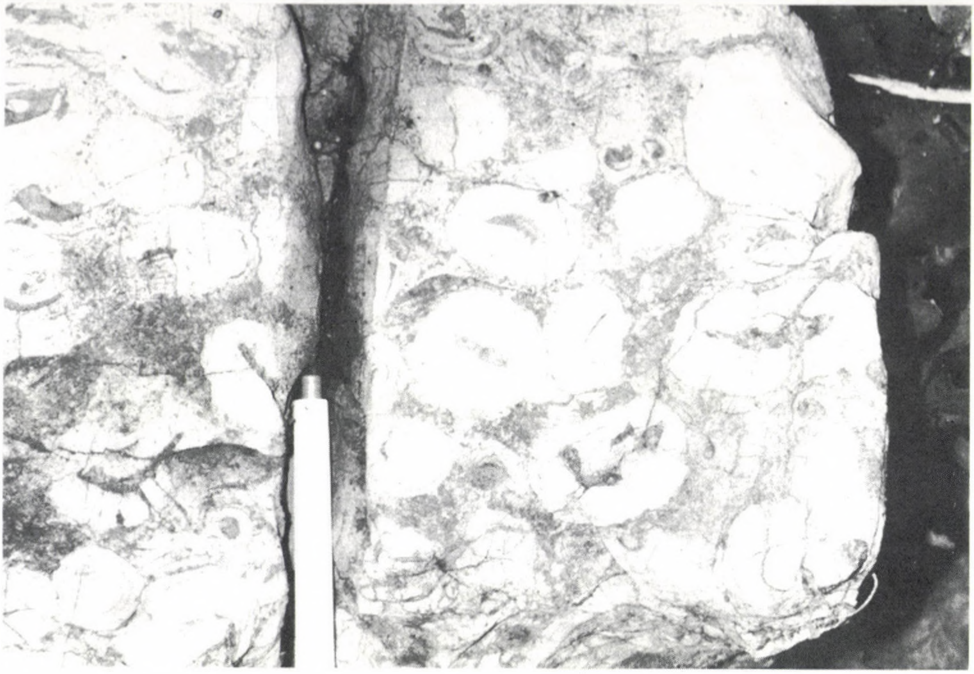

Fig. 10
Macrooncoids of back-shoal origin. Note the early plastic deformation and the later microfissures of the oncoids. Bivalve shell fragments form the nuclei of the oncoids. (Upper part of Kozár Limestone, passage close to the "U"-curve of Misina road – pen for scale)

mainly concentrated on slightly steeper slopes of the mid-ramp. The trigger mechanism could have been earthquakes as in the Polish Muschelkalk (Szulc et al. 1990) or increased bedload and shear stresses, probably of storm-induced combined flow origin. Hummocky cross-stratification of fine carbonate sand and wave ripples are signs of storm-induced waves at inner ramp zones. Nodular limestones with intercalating thin shell beds (distal tempestites) represent the deepest ramp carbonates similarly to many homoclinal ramp systems (cf. Burchette and Wright 1992). Apart from the skeletal material transport of storms mudflows were also common resulting in the formation of mud-covered parauthochtonous brachiopod beds (Fig. 9). Composite and often graded shell beds probably represent a more proximal mid-ramp deposition. The upward shoaling and coarsening is well documented by the presence of migrating ooid shoals. The youngest and the most "calm" sequence of this carbonate ramp is the Late Mid-Triassic bituminous marl. Its fine lamination, high organic content, and ostracode fauna (Monostori 1996) suggest a restricted lagoonal origin (Fig. 11).

Despite the gentle slope of the homoclinal ramp and the dominantly fine grain size of the deposited sediments (carbonate mud) the Mid-Triassic carbonates well record the dynamic processes of the depositional environment. Apart from primary sedimentary structures as slumps, channels and hummocky cross stratification the occurrence, composition and fauna of shell beds and other carbonates are evidence for storm-induced resedimentation. The

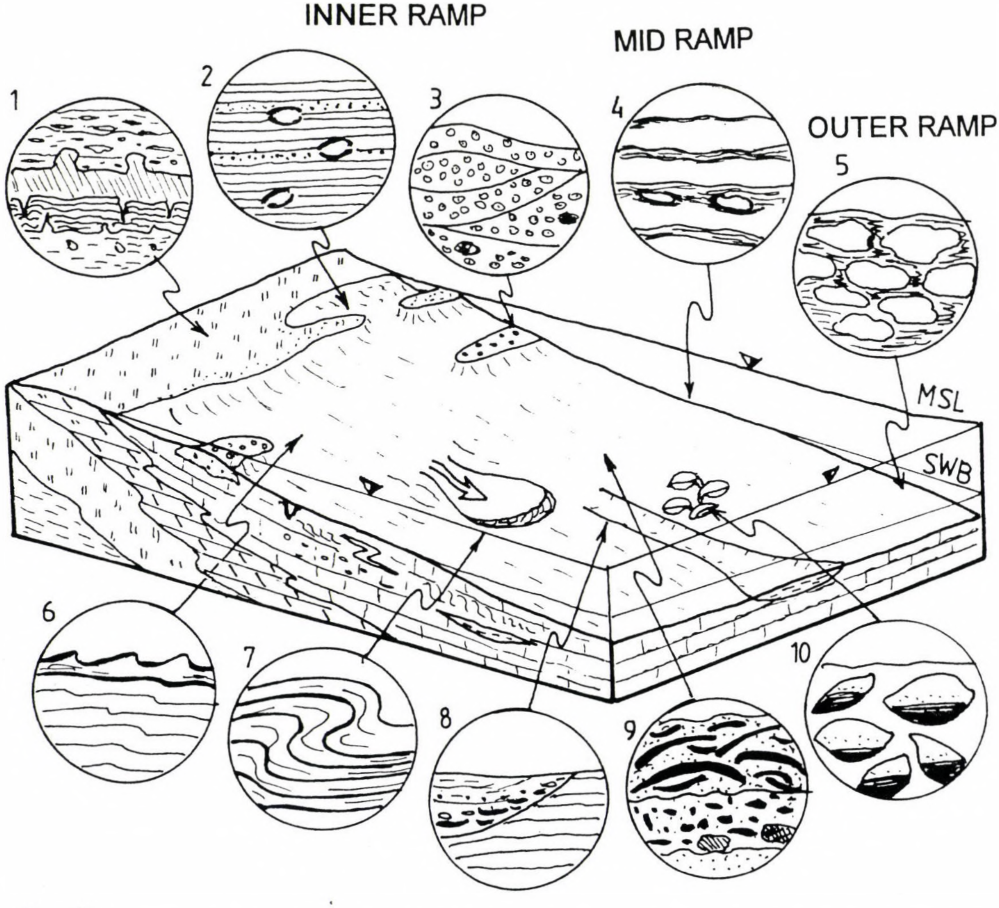

Fig. 11
Generalised depositional model of the Mecsek Muschelkalk: a carbonate mud dominated homoclinal ramp, showing the major lithofacies types and sedimentary structures. 1. Sabkha evaporites and carbonates: gypsum nodules, laminated evaporites and dolomites, entherolitic structure; 2. Lagoonal laminated bituminous marl with ostracods; 3. Cross-bedded oolitic-crinoidal shoal; 4. Alternating layers of limestones and thin marlstones with abundant trace fossils (mid-ramp); 5. Nodular limestones with brachiopods (outer ramp); 6. Ripple marks on the surface of hummocky cross-stratified beds (inner ramp); 7. Slumps (mid-ramp); 8. Channel structures of possible geostrophic current origin; 9. Storm-generated composite beds with an erosive base and a graded crinoid-bioclast packstone and a brachiopod floatstone unit; 10. Parautochtonous brachiopod beds (storm-induced mud-covered community)

importance of earthquakes as trigger mechanisms of the formation of these structures were recently emphasised in similar Muschelkalk sequences of Poland (Szulc 1993).

5. *Ramp evolution of the southern Hungarian Triassic*

The formation of the homoclinal ramp was already initiated in the Early Triassic (siliciclastic ramp stage). The transition from terrestrial clastics to siltstones and then to sabkha evaporites signifies the incipient stage of the evolution of the carbonate ramp. While clastic terrestrial input predominated in the Late Early Triassic the relative sea-level rise set up the conditions for carbonate production, as did a supposed climatic change. The "carbonate factory" was established only in the Anisian. The landward shift of the facies zones indicates the deepening of the ramp to its maximal depth during the formation of nodular outer ramp carbonates (Bertalanhegy Limestone). The mid-ramp carbonates and later the inner-ramp shoals clearly signify the reversion of ramp evolution. The lack of reefs, downslope buildups or foreslope breccias indicate the lack of organic barrier complexes and the insignificant angle of slope. The ooids may have formed isolated shoals rather than barrier banks or distinct rims on the ramp. The other evidence for this is the limited extent of back-shoal oncoid-bioclastic facies. In the Late Triassic due to a relative sea-level fall and to an increasing clastic influx carbonate sedimentation was terminated and the ramp was gradually covered by clastics (clastics-buried carbonate ramp stage). In this latter phase the formation of a half-graben structure had already begun. Thus three stages can be distinguished in the evolution of the Mecsek Triassic: 1. siliciclastic (pre-carbonate) ramp stage, 2. homoclinal ramp stage and 3. clastics buried ramp stage involving the Early, the Middle and the Late Triassic period respectively (Fig. 12). As a result of clastic cover the ramp could not evolve into an accretionary rimmed ramp or later to a rimmed shelf (cf. Read 1985).

6. *Comparison of the southern Hungarian Triassic to the Germanic Triassic*

6.1. Question of German analogy

Brachiopod-bearing beds and Triassic carbonates of southern Hungary were considered to have been Alpine analogues in the last century (Beudant 1822; Peters 1862; Böckh 1881; Haidinger 1865). Some Germanic characters were only mentioned and briefly outlined much later (Vadász 1935; Nagy 1968 and Balogh 1981). Although in the eighties a Germanic analogy was proposed (Császár and Haas 1984; Kázmér 1986) the evidence has only been published recently (Török 1986; Török and Rálisch-Felgenhauer 1990; Pálfy and Török 1992; Török 1993a). The papers listed above deal with individual aspects of the Germanic character of the southern Hungarian Triassic rather than presenting a complete

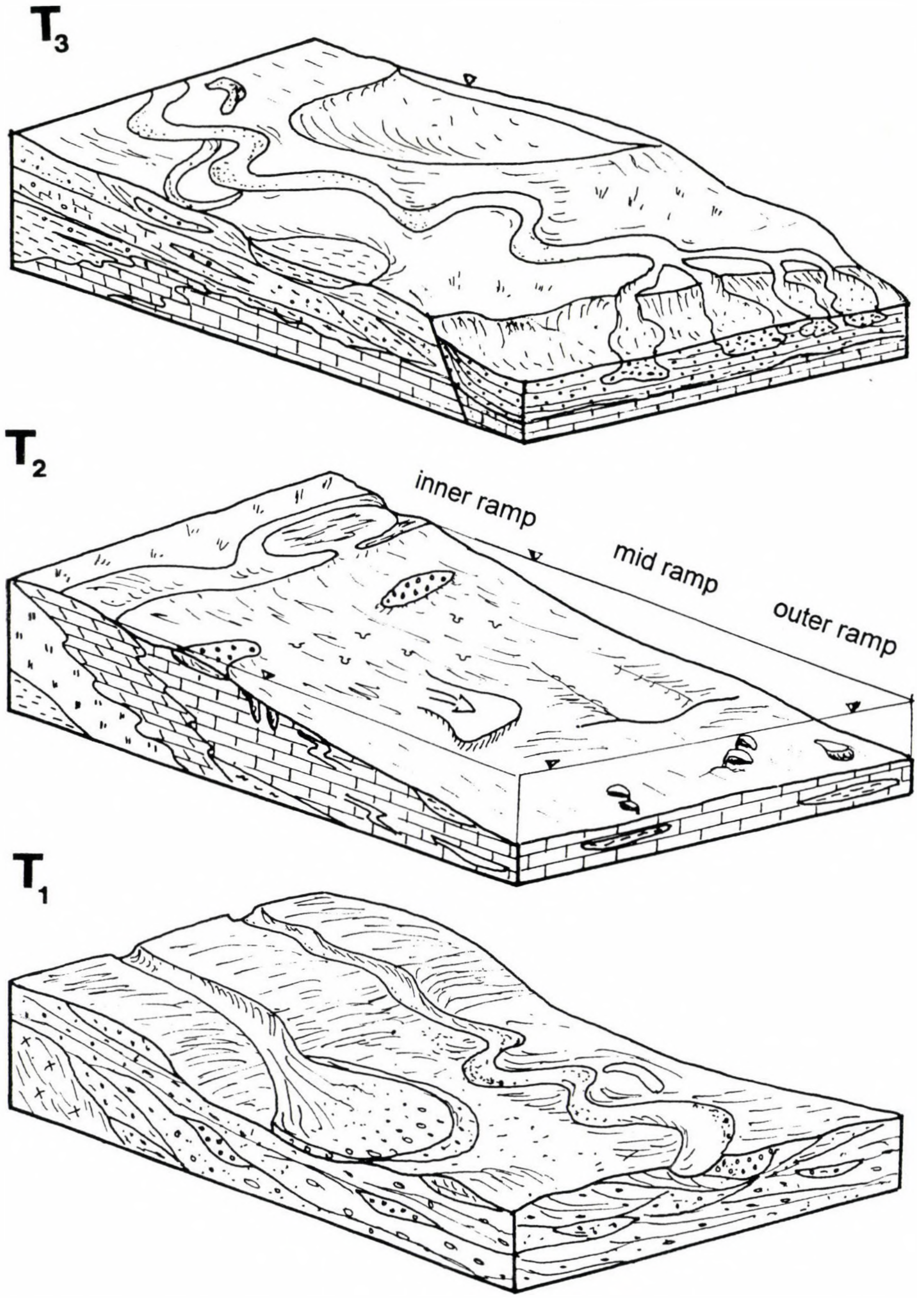

Fig. 12
Triassic platform evolution of the Mecsek Zone: Early Triassic siliciclastics covered pre-carbonate ramp stage with intense continental, mainly fluvial sedimentation (T_1). Middle Triassic carbonate ramp stage with high carbonate production on the gentle slopes of the homoclinal ramp (T_2). Late Triassic stage, temporal termination of carbonate sedimentation, a clastics buried ramp phase (T_3)

comparison including the correlation of lithotypes, similarities of depositional environments, and faunas.

6.2. Depositional environment

The tripartite subdivision of the southern Hungarian Triassic (lower clastic sequence/Buntsandstein, middle carbonate sequence/Muschelkalk and upper/ Keuper unit) is very similar to the subdivision of German epicontinental sequences (cf. Hagdorn 1991; Szulc et al. 1990 and many others); the similarity is less definite in the Upper Triassic since the Keuper in Hungary is characterized by clastics rather than evaporites. The depositional environment was a ramp, in the Muschelkalk period largely of homoclinal type very similar to that of the German Basin (cf. Aigner 1985). The processes acting on the ramp produced almost identical sediments and sedimentary structures e.g. tempestites (cf. Fig. 8, Fig. 9, Fig. 12 and Dzulynski and Kubicz 1975; Aigner et al. 1979; Aigner 1984, 1985; Hagdorn 1982; Szulc et al. 1990). The difference is expressed in the lack of large-scale barrier banks or barrier ooid complexes in the Hungarian sequence.

6.3. Correlation of lithotypes and comparison of sediments

The comparison is mainly based on Hagdorn (1991), Hagdorn et al. (1993) for Germany and Szulc et al. (1990), Szulc (1993) for Poland (Silesia) including their lithostratigraphic nomenclature; in addition the following papers were used: Kruck 1974; Schwarz 1975; Kozur 1974; Althen et al. 1980; Aigner 1985; Zwenger 1985, 1988.

The underlying clastic sequence, the Buntsandstein, is very similar in both regions. The Buntsandstein–Muschelkalk transition (Röt) is represented by coastal sabkha evaporites and marine carbonates in Poland (Röt and Gogolin Beds) whereas in Germany the Myophoria Beds and Conglomerate Bank were formed. The Hungarian sequence shows similarities to the Polish (Silesian) sequence with a transition from sabkha evaporites to intertidal dolomitized carbonates (Magyarürög Evaporites, Hetvehely Dolomites; Fig. 3). The overlying Wellenkalk carbonates are almost identical to the Mergel Beds and Wellenkalk Beds of Germany and the Upper Gogolin Beds and Goradze Beds of Poland. Both the sedimentary structures (slumps, sigmoidal beds – "S-Faltung", tempestites, and channel structures – Fig. 6) and the lithotypes (alternating thinly bedded limestones, marly limestones with wavy bedding surfaces (Fig. 5) and the bioturbated horizons ("Wurmkalk") are very similar. The Terebratula Beds of Poland and the brachiopod beds of the Lower Muschelkalk of Germany correspond to the Coenothyris-bearing limestones (Zuhánya Limestone) of southern Hungary (Figs 3, 9), representing the "deepest" sediments in the sequence (Fig. 4). Upsequence a relative sea-level fall (which is considered to have been a global one – Haq et al. 1987) is represented by bioclastic limestones and reef mounds in Poland (Karchowice

Beds) and algal oolitic shoals (Diplopora Beds – Bodzioch 1989) whereas in the Hungarian Muschelkalk the partially dolomitised limestones (Dömörkapu Limestone) are overlain by subordinately ooidal and oncoidal carbonates (Kozár Limestone – Fig. 10) in the same period. Up to the present no sign of any significant reef complexes or diplopores have been recorded. In Germany the sequence of the Middle Muschelkalk is more evaporitic (Evaporite beds) and thus less similar to the southern Hungarian Muschelkalk, although in both regions the signs of a relative sea-level fall are indicated by intense dolomitisation (Upper Dolomite Beds of SW Germany and Kán Dolomite of Hungary). A similar tendency was recognised in the Polish Muschelkalk, where a restricted lagoonal to coastal sabkha transition was documented (Tarnowice–Wilkowice–Boruszowice Beds). In the German part of the basin in the Upper Muschelkalk fossiliferous limestones and marlstones were formed indicating a more open marine environment in the Ladinian while both in Poland (Wilkowice Beds, Boruszowice Beds) and in Hungary (Kán Dolomite, Kantavár Marl) signs of restriction were observed. The Keuper is represented by different continental (lacustrine-fluvial to deltaic) sequences in Hungary, mainly consisting of sandstones and subordinately of siltstones. Polish and German examples (Aigner and Bachman 1989) indicate that the Keuper is more evaporitic and dolomitic than in Hungary. In Hungary the uppermost Triassic beds and even already the Middle Carnian ones contain minor coal seams as a precursor of the early Jurassic coal beds (Bóna 1984).

Thus the Hungarian Germanic Triassic is very similar to German epicontinental sequences and especially to Silesian (Polish) ones although it does not contain reef mounds. The difference between the German and Hungarian Muschelkalk appears in the Middle–Upper Muschelkalk, namely in the lack of significant evaporites in the Hungarian Middle Muschelkalk and the absence of the second "transgressive" carbonate cycle of Upper Muschelkalk. The differences may be related to the different palaeolatitudinal setting and to the different size of the ramps (the Hungarian one was apparently smaller and shallower, but the carbonate production rate was much higher there). Therefore a much thicker Muschelkalk succession was developed than that of the Polish or German ones. The differences of sedimentary cycles are more distinct if we compare the Hungarian Muschelkalk with the Spanish one where three major "transgressive" cycles were recognised (Calvet and Tucker 1988) unlike the one major cycle in Hungary.

6.4. *Faunistic similarities*

Apart from the lithological–sedimentological analogies between the German and southern Hungarian Triassic the fauna assemblages also show similarities. The biostratigraphic subdivision of the Hungarian Triassic is less well established than that of the Germanic Basin due to the sparse occurrence of fossils with biostratigraphic value (see Balogh 1981). The Anisian brachiopod and bivalve fauna is very similar to the German one (cf. Assmann 1937; Hagdorn 1991 with Nagy 1968; Pálfy and Török 1992; Török 1993b). The oligospecific brachiopod fauna of Mecsek (southern Hungary) is characterized by the dominance of a highly tolerant species – *Coenothyris vulgaris* (Schlotheim) – and the absence of Alpine brachiopods (Pálfy and Török 1992). Conodonts were also found in the Hungarian Muschelkalk sequences (Kovács and Papsová 1986; Bóna 1976). Cephalopods are very rare and poorly preserved in the Hungarian Muschelkalk and only found in one horizon in the Bertalanhegy Limestone, from where a *Paraceratites binodosus* (?) was described (Detre 1973). Other pelagic forms are generally absent; e.g. no Daonellas were found as have been described from the Spanish Muschelkalk (Virgili 1958).

6.5. Palaeogeography

The southern Hungarian Triassic was the part of the Triassic extensional regime of present Central Europe. Because of initial disintegration of Pangaea differential subsidence and uplifting of subunits (smaller blocks?) created a very complex system of "grabens and troughs" in a crustal extensional regime (Ziegler 1982). As a part of a passive continental margin an "epicontinental-type" and somehow restricted basin was formed which shows the characteristics of ramp evolution. Ramps often develop on the shallow dip slopes of fault blocks ("rollovers") of extensional basins (Burchette and Wright 1992). The shallowness and semi-restricted setting appeared in similar facies types as in the German epicontinental basin (cf. Aigner 1985). The presence and frequency of storm events suggest that it was located in the zone of hurricane and winter storm-dominated palaeolatitude (between 25 and 45° North) in accordance with the distribution of Triassic palaeostorm-influenced sequences (Marsaglia and Klein 1983). To set up an appropriate palaeogeographic model further data are needed; thus, several models have been proposed, not always in agreement with each other (Kovács 1982; Tollmann 1987; Haas 1987; Kázmér and Kovács 1989; Vörös et al. 1990). The uncertainties derive from the superimposing effect of the later Alpine–Carpathian deformation, i.e. the rearrangement of the tectonic units along large-scale dextral and sinistral strike-slips (Csontos et al. 1992). These post-Triassic deformations raise further difficulties in the reconstruction of the Triassic palaeogeography in those areas which are located south of the Alpine deformation front (Ziegler 1982), i.e. the southern Hungarian Triassic. Despite the uncertainties in most models it is accepted that the Tisza megaunit (southern Hungarian Triassic) was a part of the European continental margin rather than of the African plate in the Triassic (Kázmér and Kovács 1985; Vörös et al. 1990; Balla 1988).

7. Conclusions

1. Germanic-type Triassic sediments occur within the Carpathian Basin, in southern Hungary in the Tisza megaunit (on the surface in the Mecsek and Villány Mountains). This unique geographic position is related to large-scale southward displacement of this megaunit.

2. In Hungary the Germanic-type Triassic comprises the clastic sediments of the Buntsandstein, the clastic-evaporitic-carbonate Röt, the Mid-Triassic carbonates of the Muschelkalk, and the Late Triassic clastics of "non-typical" Keuper. In the Hungarian Muschelkalk there is no evidence for the formation of thick evaporites equivalent to those of the German Middle Muschelkalk.

3. The sediments of the Muschelkalk were deposited on a storm-influenced ramp. Characteristic sedimentological features, storm-generated skeletal sheets, channel infills, slump and crumpled structures of the Wellenkalk unit are very similar to their German counterparts. The brachiopod banks, the nodular lime mudstones and bipartite graded storm beds of the *Coenothyris* zone are also analogous to the German and Polish Muschelkalk, although their fauna is less diverse.

4. Compared to other Germanic Mid-Triassic sequences the difference is probably due to the different size, the shallower water depth, the increased sedimentation rate, the more significant restriction and the different palaeolatitudinal position of this ramp.

5. In the Triassic evolution of this region three major phases can be distinguished: a siliciclastic covered pre-carbonate ramp stage, a homoclinal carbonate ramp and a clastics buried ramp stage (in the Early, Middle and Late Triassic, respectively).

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Crinoids from the Muschelkalk of the Mecsek Mountains and their stratigraphical significance

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Collections of dissociated echinoderm sclerites from the Muschelkalk of the Mecsek Mts yielded different crinoid assemblages. Because they are strongly resembling crinoid assemblages from the Muschelkalk in Upper Silesia in composition and stratigraphic position, the zonal scheme of 4 crinoid biozones established for the Muschelkalk in the eastern part of the Germanic basin can be transferred to the Mecsek Muschelkalk. These biozones are: Dadocrinus zone, acutangulus zone, dubius zone, silesiacus zone (Lower Anisian to Lower Illyrian). The same community sequence or zonal scheme corresponding to a similar vertical facies sequence from soft grounds to shelly and firm grounds is observed in distant sections of the western Tethys realm during Lower Anisian through Lower Illyrian times. Significant crinoid assemblages were covering wide areas of the western Tethys during Early Middle Triassic times.

Key words: benthic crinoids (Articulata, Encrinida, Isocrinida, Millericrinida), biostratigraphy, Triassic (Anisian), Muschelkalk, palaeobiogeography, Hungary (Mecsek Mts)

1. Introduction

In the Hungarian Muschelkalk of the Mecsek and Villány Mts. fossils with biostratigraphical value are relatively rare. The biostratigraphical subdivision of Middle Triassic carbonates is based on brachiopods (Nagy 1968; Török 1993) and bivalves (Nagy 1968; Szente 1997). Additionally, ammonoids (Detre 1973) and conodonts were found in the "Coenothyris beds". Brachiopods and bivalves were also used for the correlation of the Hungarian and the German Muschelkalk sequences (Pálfy and Török 1992; Török 1993, 1997). Until now the significance of benthic crinoids for the biostratigraphical subdivision of the Hungarian Muschelkalk and its correlation with other Middle Triassic sequences has not been recognized yet.

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Vadász (1935) has already mentioned the presence of *Encrinus* sp. and *Isocrinus* sp. in the *Coenothyris*-rich "middle member" of the Anisian carbonates, but he neither figured nor described his material in more detail. Additionally to these taxa, Nagy (1968) reported the occurrence of *Pentacrinus* sp. and *Dadocrinus* cf. *gracilis* (Buch) from the brachiopod-bearing beds (Bertalanhegy Limestone), without giving detailed descriptions. However, in the stratigraphical literature the taxa *Pentacrinus* and *Isocrinus* have been used in a collective sense for any pentagonal or stellate crinoid columnals versus *Encrinus* or *Entrochus* for cylindrical columnals. In his monograph Nagy also pointed out that crinoids occurred in several lithological units of the Anisian sequence. According to the present lithostratigraphical subdivision Nagy (1968) mentioned an indetermined crinoid species from Lapis Limestone, Bertalanhegy Limestone, Dömörkapu and Kozár Limestone units (cf. Fig. 2).

The Triassic crinoids of the Balaton Highland have been described and figured in detail by Bather (1909) in his great monograph on the Triassic echinoderms of Bakony. In this work he also commented on many taxa from the Germanic Muschelkalk. However, he had no material from the Mecsek Muschelkalk. Although Bather's interest focused on systematics and morphology he was aware of the stratigraphical significance of some taxa but he did not formally use them as biostratigraphic index fossils.

The aim of the this paper is (1) to fill these gaps and to describe and illustrate crinoid remains collected in the Mecsek Muschelkalk during the last years (2) to compare the Mecsek Muschelkalk crinoid faunas and their biostratigraphical zonation with those from other sequences in the Western Tethys realm.

2. *Geological setting, lithotypes and crinoid samples*

The Anisian Muschelkalk carbonates are found in the Mecsek and Villány Mts. of Southern Hungary. For the present study, samples collected from five localities (Fig. 1) representing different positions in three major lithostratigraphic units have been analysed (Fig. 2).

(1) Bükkösd (old quarry) with a section of the lower part of the Lapis Limestone in typical Wellenkalk facies. The most significant lithofacies types are nodular marlstones, alternating layers of thinly bedded limestones and marlstones with bioturbated limestones. Dissociated crinoid ossicles mainly occur in cm-thick intercalated skeletal sheets which were generated by storms (Plate I, Figs a–c). All elements collected were determined as *Dadocrinus* sp. This determination is corroborated by a fine specimen of a slab with several articulated crowns and stems parts, which has been collected by Gy. K. from the talus at the same locality. Unfortunately this specimen was stolen from the collections of the Geological Institute of Janus Pannonius University. The photographic pictures taken by H. H. and Gy. K. still give additional evidence that the crinoid ossicles from locality (1) are *Dadocrinus*.

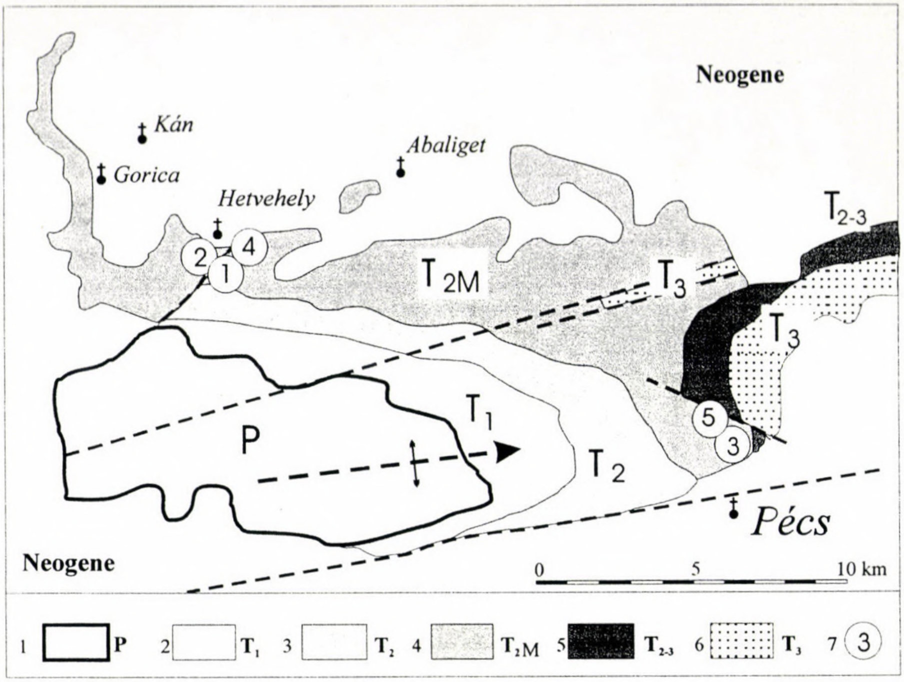

Fig. 1
Geological map of central and western Mecsek Mountains with Muschelkalk crinoid localities. 1. Upper Permian siltstone and sandstone; 2. Lower Triassic sandstone; 3. Middle Triassic siltstone, evaporites, dolomites (Röt); 4. Middle Triassic Muschelkalk carbonates; 5. Middle-Upper Triassic calcareous marl; 6. Upper Triassic sandstone; 7. crinoid localities: 1) Bükkösd, old abandoned quarry, Lapis Limestone; 2) Bükkösd valley, uppermost part of Lapis Limestone (Tubes Limestone); 3) Misina road cut, geological key section of Bertalanhegy Limestone ("Coenothyris beds"); 4) Bükkösd-Hetvehely road, abandoned old quarry near railway tunnel, Kozár Limestone, close to the top of Dömörkapu Limestone; 5) Misina road cut, U-curve, crinoidal limestone in Kozár Limestone

(2) Bükkösd valley. At the base of the section exposing the Bertalanhegy Limestone which is rich in well-preserved brachiopods, massive lenticular bodies of crinoidal limestone were found. These pillow-shaped load casts reaching a few meters in size are generally found in the wellenkalks of the uppermost part of the Lapis Limestone. Echinoderms collected from the lower surface of one of these pillows (Plate I, Fig. d) include small columnals and cirrals of *Holocrinus* cf. *acutangulus*, single columnals of *Eckicrinus radiatus* and the echinoid *Triadotiaris grandaeva* (spine, interambulacral plate).

(3) Misina road cut (geological key section) with small outcrops of the Bertalanhegy Limestone. This most fossiliferous unit of the entire Mecsek Muschelkalk contains abundant articulate brachiopods, bivalves, and extremely rare cephalopods. Single dissociated crinoid ossicles and pluricolumnals occur in nodular limestones intercalated with marls and in brachiopod beds with

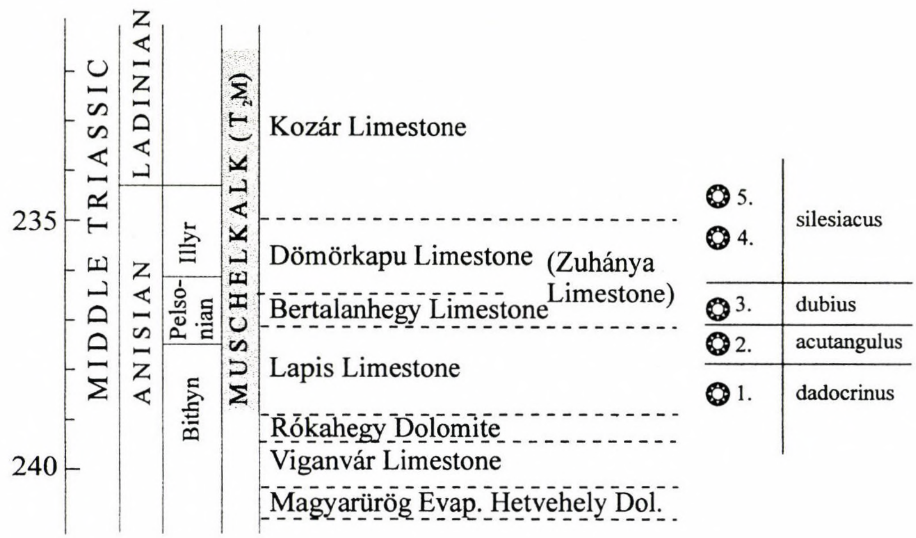

Fig. 2
Lithostratigraphical subdivision of the Mecsek Muschelkalk; position of localities 1–5 in the stratigraphical column and proposed crinoid zonation

Coenothyris vulgaris and *Tetractinella trigonella* (Plate II, Figs a–h). Composite storm beds yielded isolated and fragmented ossicles. Most of the specimens have been washed out from a marlstone sample found in between the nodular limestones. This sample contained abundant barrelshaped columnals of an undetermined crinoid, single columnals of *Holocrinus* sp. cf. *H. acutangulus* and *Eckicrinus radiatus*.

(4) Bükkösd–Hetvehely road, abandoned quarry near railway tunnel with slightly dolomitic, thickly bedded biocalcarenites and rudites of the basal Kozár Limestone. Dissociated crinoid ossicles and a few pluricolumnals were picked from the karstified surface or washed from small pockets or mottles (Plate III, Figs h–l). Due to the weathering process most of the sclerites are slightly etched The sample contains *Silesiacrinus* cf. *silesiacus*, *Holocrinus* sp. cf. *H. dubius*, *Eckicrinus radiatus* and undetermined encrinids.

(5) Misina road cut, U-curve with an outcrop in the higher part of the Kozár Limestone exposing a crinoidal limestone bed intercalated in between thickly bedded oolitic limestones. At small patches the crinoidal limestone is karstified and deeply weathered; a sample taken from this crumbly rock (Plate III, Figs a–g, m) yielded abundant crown and stem ossicles of an encrinid (*Chelocrinus* and *Encrinus*), more rarely *Holocrinus* sp. cf. *H. dubius* and one columnal of *Silesiacrinus* sp. cf. *S. silesiacus*. Additionally an interambulacral plate of the echinoid *Triadotiaris grandaeva* has been found. Unfortunately the surfaces of the ossicles are rather etched by humic acids.

3. Depositional environment of the crinoid bearing beds

The sediments containing the crinoid remains were deposited in different positions of a homoclinal ramp dominated by carbonate mud (Török 1993). During the formation of the wellenkalk carbonates (Lapis Limestone) on deeper parts of the ramp, muddy bottom conditions with soft ground benthic communities were deposited. The dadocrinids were attached by their terminal discoid hold fasts on hard ground tops, however, their small size enabled them also to dwell on soft grounds when encrusting the rear ends of mudsticking bivalves (Hagdorn 1996). Such bivalves as *Gervillella mytiloides* and *Hoernesia socialis* have been found in the Mecsek wellenkalks. It may be concluded that comparably to its occurrence in Upper Silesia and the Vicentinian Alps *Dadocrinus* in the Mecsek Muschelkalk inhabited both hard grounds and muddy bottoms as a secondary soft ground dweller.

Upsection patchily distributed firm grounds among muddy bottom may have been inhabited by holocrinids forming clusters of many individuals entangled with their cirri. Lifelong shedding of their distal stem parts caused accumulation of distal columnals and cirrals while crown and proximal stem parts are underrepesented (Hagdorn and Baumiller in press). Together with shells and other biodetritus the sclerites sunk into the muddy sea floor forming the pillow shaped load casts.

Solid shell beds and decreasing sedimentation rates allowed settlement of epibenthic low tier brachiopods and bivalves and higher tier crinoids after a slightly upward shallowing and shoaling of the ramp nodular limestones with skeletal shell beds of the Bertalanhegy Limestone.

During deposition of the thickly bedded crinoidal limestones of the Kozár Limestone on a very shallow position of the ramp conditions were most favourable for crinoids and other epibenthic filter feeders as rates of carbonate mud sedimentation were extremely low. Subsequently mud bottom communities were replaced by shell ground communities with large crinoids forming small buildups with the root calli of their terminal holdfasts.

4. Mecsek Muschelkalk crinoids

Due to the incomplete and often poor preservation of the material, definite determinations at genus or even at species level are not possible in any case. The aim of this chapter is not to give full descriptions of the recovered taxa but rather to document what has been found and how the taxa can be identified. Synonymy lists are focusing recent references.

The crinoid material of this report has been deposited in the Muschelkalkmuseum Ingelfingen (MHI).

CLASS: CRINOIDEA MILLER, 1821

Subclass: Articulata Miller, 1821

Order: Encrinida Matsumoto, 1929 [nom. transl. Hagdorn, 1987]

Family: Encrinidae Dujardin & Hupé, 1862

In the upper Kozár Limestone fauna (locality 5) encrinid ossicles represent the bulk of all crinoid remains; at locality 4 definite encrinids are scarce. They belong to 2 different genera, *Encrinus* and *Chelocrinus*. Identification of the isolated ossicles was facilitated by comparison with several articulated crowns from the Karchowice Formation of Upper Silesia. However, most encrinid ossicles are indistinct and may belong to any encrinid taxon.

Genus: *Encrinus* Lamarck, 1801

Encrinus aculeatus v. Meyer, 1847

Plate III, Figs a–b

1926	*Encrinus aculeatus* v. Meyer	– Assmann: 509–511, pl. 8, figs 1-4
1993	*Encrinus aculeatus*	– Hagdorn & Gluchowski: Figs 10, 2, 5
1996	*Encrinus aculeatus*	– Hagdorn, Gluchowski & Boczarowski: 52–53, pl. 1 a-e

Diagnosis: An *Encrinus* with flat bowl shaped cup with subhorizontal base. Radials and first primibrachials with strongly inflated dorsal sides. Interradial and basal/radial articula synostosial with distinct ligament pits. In adults, arms uniserial up to the 8th secundibrachial. Dorsal side of proximal brachials with sharp, blade shaped spines in juveniles, in adults inflated. Median and distal brachials with straight spines, which may bear knobs. Pinnules smooth.

Material: 1 first primibrachial, 1 axillary second primibrachial from the upper Kozár Limestone (locality 5); MHI 1558/1–1558/2.

Description: The 2 primibrachials show the typical dorsal inflations which have not been observed among *Chelocrinus carnalli*. The ligamentary facets are zygosynostosial with very faint aboral crenulations. Both brachials belong to a subadult individual.

Geographical and stratigraphical range: E. aculeatus is a Pelsonian to Lower Illyrian species that has been recovered in the Lower Germanic Muschelkalk of Poland (uppermost Gogolin Formation to Diplopora Dolomite) and Germany (Wellenkalk Formation) and in the Southern Alps (Giudicarie). Similarly as in the Kozár Limestone, crinoidal limestone beds in the Upper Silesian Diplopora-Dolomite contain much less sclerites of *E. aculeatus* than of *Chelocrinus*. For more detailed references see Biese (1934) and Hagdorn et al. (1996).

Genus: *Chelocrinus* v. Meyer, 1835

Chelocrinus carnalli Beyrich, 1856

Plate III, Figs a–b

1993	*Chelocrinus carnalli* (Beyrich, 1856)	– Ernst & Löffler: 224–228, Abb. 4
1996	*Chelocrinus carnalli* Beyrich, 1856	– Hagdorn, Gluchowski & Boczarowski: 53–55, pl. fig. 3, pl. 1, f–s

Diagnosis: A large *Chelocrinus* with bowl shaped to low cone shaped cup and convex base. Basals moderately to very long. Facets of cup elements synostosial. Dorsal sides of cup and arm elements smooth. Arms uniserial up to IIIBr10 or farther distal, then becoming immaturely biserial. Brachials with low wedge shaped dorsal sides. Pinnules smooth, rarely pectinate.
Material: 1 basal, 6 radials (3.8–8.5 mm width), 10 first primibrachials, 11 axillary second primibrachials (2.9–8.8 mm width), 2 first secundibrachials, 10 axillary second secundibrachials, 8 more distal beachials from the upper Kozár Limestone fauna (locality 5); MHI 1558/3–1558/52.
Description: The determination is based on cup and arm plates with smooth dorsal surface and the presence of asymmetrical second axillaries. Their smooth dorsal sides clearly differ this species from another but still unnamed Anisian *Chelocrinus* with slightly ornamented radials and proximal brachials. Among larger radials the basal/radial and interradial facets are synostosial with fairly deep ligamentary pits. The primibrachials are very low.
Geographical and stratigraphical range: Ch. carnalli is a Pelsonian (?) and Lower Illyrian species that has been recovered in the Lower Germanic Muschelkalk of Poland (uppermost Gogolin Formation to Diplopora Dolomite) and Germany (Wellenkalk Formation, Schaumkalkbank Member) and in the Southern Alps (Brachiopod Limestone, Recoaro). As in the Kozár Limestone, *Ch. carnalli* is the most abundant crinoid in crinoidal limestone beds of the Upper Silesian Diplopora-Dolomite. As most of the determinable cup and arm sclerites definitely belong to *Chelocrinus carnalli* the bulk of the undeterminable encrinid sclerites (columnals) from locality 5 may also be attributed to this species. For more detailed references see Biese (1934) and Hagdorn & al. (1996).

Encrinidae gen. et sp. indet.

Plate III, Figs c–f

Although particular characters of the encrinid stem may be diagnostic, it is impossible to generally determine their columnals at genus or even species level. This is especially true for those polyspecific faunas containing no complete articulated skeletons. Encrinid columnals have the following diagnostic characters:

– the lumen is markedly wider than the holocrinid lumen, but not as wide as the millericrinid lumen

– encrinids have short and coarse crenulae

– encrinid columnals with petaloid pattern have rounded interradii and wide rounded areolae.

Material: Some hundreds of isolated columnals from the upper Kozár Limestone fauna (locality 5); MHI 1558/53–1558/60.

Description: Among the encrinid columnals from locality 5, small to medium-sized cylindrical columnals with multiradiate crenulation and a more or less well-developed granulated perilumen are dominating. They belong to the middle and distal stem part. A few smaller cylindrical columnals which are higher than wide have a distinct epifacet. Low subcircular to subpentalobate columnals with a petaloid crenulation or short multiradiate crenulation pattern and a rounded epifacet are proximal nodals and primary internodals; nodals with cirrus scars have not been found. Low columnals devoid of an epifacet are high order internodals; immature internodals have a rounded pentalobate outline and are extremely low.

Geographical and stratigraphical range: Indistinct encrinid columnals are found in Pelsonian to Carnian carbonates and marls of the entire western Tethys realm. As cirrinodals with thick epifacet which are distinctive for Pelsonian/Lower Illyrian have not been found at localities 4 and 5, the present columnal association does not provide any biostratigraphical evidence.

A sample (MHI 1559) from Forráshegy near Felsőörs (Recoaro type brachiopod limestones, Felsőörs-Formation; Pelsonian) yielded cirrinodals with 2 to 5 cirrus sockets similar to those described from the Silesian Diplopora-Dolomite (Hagdorn et al. 1996).

Family: Dadocrinidae Lowenstam, 1948

Genus: *Dadocrinus* v. Meyer, 1847

Dadocrinus sp. indet.

Plate I, Figs a–c

Material: 7 small wellenkalk samples with crinoid columnals on the bedding planes; Bükkösd valley, old quarry (locality 1); MHI 1560/1–1560/7.

Description: The cylindrical to slightly barrel shaped columnals are small (0.8–4.4 mm). Their crenulation pattern is multiradiate with short culmina and a raised perilumen. A few columnals are slightly subpentagonal; they belong to the proximal stem part.

Disarticulated *Dadocrinus* sclerites are not diagnostic at species level. The articulated specimens on the slab from locality 1, which was stolen, were slightly corroded or weathered. Their relatively large crowns with 10 arms had conical cups with the infrabasals not exposed. The proximal stems were

subpentagonal and tapering below the cup. Being intermediate between *Dadocrinus gracilis* and *D. grundeyi* these crinoids reflect the instability of character sets among *Dadocrinus*; the *Dadocrinus* species have therefore been treated as ecophenotypes by some authors (Gluchowski 1986; Hagdorn 1996).
Geographical and stratigraphical range: Dadocrinus is a Lower Anisian crinoid which was widely distributed over the western Tethys. In Upper Silesia the genus appears at the base of the Lower Gogolin Formation and disappears above the base of the Upper Gogolin Formation. Towards the western parts of the Germanic basin, a salinity gradient prevented its westward dispersal beyond Brandenburg (Hagdorn and Gluchowski 1993). *Dadocrinus* also occurs in the Southern Alps (Formazione à gracilis, Recoaro), the Northern Calcareous Alps (Alpine Muschelkalk, Styria, Tyrolia, Vorarlberg, Brianconnais), the High Tatra (Poland). For more detailed references see Biese (1934) and Hagdorn (1996).

Encrinida fam. et gen. indet.

Plate II, Figs a–c

Material: 224 columnals washed out from marls at locality 3; MHI 1561/1-1561/224.
Description: The high barrel shaped to bead shaped columnals are small (1.2–3.5 mm). They are generally higher than wide; small barrel shaped columnals may be double as high than wide. The articulation facet is less wide than the columnal diameter. Its crenulation pattern is multiradiate with a raised perilumen. Subpentagonal columnals have not been observed. Because cups or crowns of this crinoid are still unknown its systematic position remains questionable. The crinoid may be a late dadocrinid or a small encrinid.

Similar columnals of Pelsonian age have been collected from (1) the Upper Silesian Terebratula Formation (dissociated columnals, 2 stem fragments from Strzelce Opolskie/Großstrehlitz, MHI 1564) and (2) the Lower Muschelkalk Terebratula Member of Upper Franconia, Bavaria (dissociated columnals and crown sclerites from Zeyhern, MHI 1565). This material has not been described or figured yet. The brachiopod beds in the Felsőörs Formation of Forráshegy also yielded high barrel shaped to high cylindrical columnals which may be attributed to the same crinoid genus (MHI 1559).
Geographical and stratigraphical range: If the fragmentary remains do belong to the same taxon the small crinoid with its stem resembling a chain of beads is an index of the Pelsonian substage distributed along the northeastern coast of the Bohemian Massif, in the Mecsek and Balaton Triassic of Hungary.

Order: Isocrinida Sieverts-Doreck, 1952

Family: Holocrinidae Jaekel, 1918

Genus: *Holocrinus* Wachsmuth & Springer, 1887

Among the Germanic Lower Muschelkalk holocrinids, from the earliest (Lower Anisian) *H. acutangulus* to *H. dubius* (Pelsonian, Lower Illyrian) and *H. meyeri* (Lower Illyrian), the size gradually increased and the crenulation pattern became more isocriniform. For determination, maximum size and crenulation patterns of the adult columnals in an association is diagnostic. However, preservation and specimen number of the Mecsek specimens is not fully sufficient for determination. Therefore the samples are compared with the most similar Germanic Muschelkalk taxa.

Holocrinus sp. cf. H. acutangulus (v. Meyer, 1847, 1849)

Plate I, Fig. d; Plate II, Figs d–e

1986 *Holocrinus beyrichi* (Picard, 1883) – Hagdorn: Taf. 3, Figs 1–8
1993 *Holocrinus acutangulus* – Hagdorn & Gluchowski: Fig. 6

Material: 4 small rock samples of crinoidal limestone with single brachials, columnals and cirrals at the surface; lower side of a big load cast pillow; Bükkösd valley (locality 2); MHI 1562/1–1562/4. 44 columnals and 1 IBr1 from Misina road cut (locality 3); MHI 1561/225–1561/270.
Description: The poorly preserved columnals are fairly small (1.2–3.6 mm). Their outline is subpentagonal (basaltiform) to pentagonal; the slightly higher nodals may have substellate outline. Due to the small size, the crenulation pattern is multiradiate with long culmina; only the largest columnals have a simple petaloid pattern with lanceolate petals but devoid of distinct radial bands. The nodals have deeply impressed elliptical cirrus scars with a distinct transverse ridge. The lower nodal facets are symplectical.
Geographical and stratigraphical range: In Upper Silesia the small *H. acutangulus* appears in the Late Lower Anisian Upper Gogolin Formation and dispersed westward over the entire Germanic basin (basal Wellenkalk Formation). Similar columnals have been collected in Late Lower Anisian Muschelkalk of the Northern Calcareous Alps (Reutte, Tyrolia); MHI 1566. For more details compare Biese (1934) and Hagdorn (1986).

Holocrinus sp. cf. H. dubius (Goldfuss, 1831)

Plate III, Fig. g

1986 *Isocrinus dubius* (Goldfuss, 1831) – Hagdorn: Figs 1–3, Pl. 1–2
1993 *Holocrinus dubius* – Hagdorn & Gluchowski: Fig. 8, 1–3

Material: 6 columnals (1.6–4.3 mm) from Kozár Limestone at locality 4; 1563/1–1563/6; 60 columnals (1.7–5.3 mm) from locality 5; MHI 1568/51–1568/120.
Description: Most of the poorly preserved columnals are fairly small. However, the samples contain normal sized adult columnals. The outline of the internodals is subpentagonal (basaltiform) to pentagonal or substellate; the nodals have substellate outline. The small columnals have multiradiate crenulation patterns with long culmina. Adult columnals show the typical petaloid pattern with long marginal culmina, pyriform to lanceolate petals and granulated radial bands. The nodals have 5 deeply impressed elliptical cirrus scars with a distinct transverse ridge. The lower nodal facets are symplectical.
Geographical and stratigraphical range: H. dubius is a typical Pelsonian to Lower Illyrian Holocrinus with a wide range over the entire Western Tethys realm. It is common in the Lower Muschelkalk of Germany from the Terebratelbank-Member upward, in Poland from the Terebratula Formation to the Karchowice Formation. It also occurs in the Southern Alps (Brachiopod Limestone, Recoaro) and in the Kaukasus. For more details compare Biese (1934) and Hagdorn (1986).

Genus: *Eckicrinus* Hagdorn & Gluchowski, 1993

Eckicrinus radiatus (Schauroth, 1859)

Plate II, Figs f–j, Plate III, Fig. h

1859	*Encrinus? radiatus* n. sp.	– Schauroth: 288, Taf. 1, Figs 4a-c
1909	*Balanocrinus radiatus* (Schauroth)	– Bather: 15
1979	*Laevigatocrinus radiatus* (Schauroth)	– Klikushin: 89, Fig .1
1992	*Laevigatocrinus radiatus* (Schauroth, 1859)	– Klikushin: 90
1993	*Eckicrinus radiatus* (Schauroth, 1859)	– Hagdorn & Gluchowski: 170, 174, Figs 8, 5
1996	*Eckicrinus radiatus* (Schauroth, 1859)	– Hagdorn, Gluchowski & Boczarowski: Figs 5–6, Pl. 5

Diagnosis: Columnals very low, circular to subcircular, cylindrical. Proximal columnals with marginal crenulae and granulated radial bands, petal floors pyriform. Distal columnals with long marginal crenulae, petal floors very small. Bifurcation and intercalation of extra culmina towards the periphery. Nodals not wider than internodals, only slightly higher. Up to 5 round cirrus scars, fulcral ridge indistinct or lacking; enlarged cirri with multiradiate articulation. Noditaxis very short. This taxon is known only from columnals.
Material: 20 columnals and short pluricolumnals (1.8–3.2 mm) from Bertalanhegy Limestone (locality 2); MHI 1561/271–1561/290; 2 columnals (3.4 mm) from Kozár Limestone (locality 4); MHI 1563/7–1563/8.
Description: The small columnals from locality 2 have the typical circular to subcircular outline and facets with long culmina arranged around indistinct small pyriform petal floors. Granulated radial bands are present. The proximal columnals from the Kozár Limestone are larger and their petaloid pattern with long culmina surrounding the pyriform petal floors is more characteristic (Plate III, Fig. h). The

distal columnals have extremely small and indistinct petals and very long multiradiate and bifurcating crenulae; their pattern is very similar to *Silesiacrinus* but the lumen is narrow; however, the lumen may be diagenetically filled with drusy cement (Plate III, Figs i–j).

Columnals of this species have often been confused with *Silesiacrinus* silesiacus. Distal columnals with long culmina and an atypical petal pattern may be very similar to *S. silesiacus*; however, this species has a wide lumen.
Geographical and stratigraphical range: Eckicrinus radiatus is indicative for Pelsonian and Early Illyrian. It has been recovered in the eastern part of the Germanic basin (Upper Silesia, Uppermost Gogolin Formation to Diplopora Dolomite; Holy Cross Mts, Lima striata Formation). However, it is not a Germanic but a Tethydian genus occurring in Italy (Southern Alps, Recoaro; Prags and Olang Dolomites, Cadore, Mte. Rite), Hungary (Balaton Highland, Mecsek Mountains) and from the Ladinian (?) and Pelsonian of Russia (Amur Basin, Siberia). For more details compare Biese (1934) and Hagdorn et al. (1996).

Order: Millericrinida Sieverts-Doreck, 1952

Genus: *Silesiacrinus* Hagdorn & Gluchowski, 1993

Silesiacrinus silesiacus (Beyrich, 1857)

Plate III, Figs i–m

1909	*Entrochus silesiacus*	– Bather: 11, 12, 14, 237, 256, Taf. 1, Fig. 24
1934	*Entrochus silesiacus* Quenst.	– Schnetzer: 6, 145
1967	*Entrochus silesiacus* Beyrich	– Kristan-Tollmann & Tollmann: 5, 9, 22, 26, 27, Taf. 1, Figs 8, 9, Taf. 8, 14, 15, 17, Taf. 9, Figs 7–12
1993	*Silesiacrinus silesiacus* (Beyrich)	– Hagdorn & Gluchowski: 171, 175, Figs 10, 1
1996	*Silesiacrinus silesiacus* (Beyrich)	– Hagdorn, Gluchowski & Boczarowski: 65–68, pl. 6

Diagnosis: Columnals low, circular, with straight to convex latera. Articular facets multiradiate; culmina long, often with delicate longitudinal groove. Granulated radial bands may occur. Areola very narrow or lacking. Axial canal wide, circular or weakly pentalobate, simple or complex with lensoid spatia. Holdfast irregular discoid, consisting of single elements. This taxon is known only from columnals.
Material: 26 columnals and pluricolumnals from Kozár Limestone at Bükkösd-Hetvehely road (locality 4); MHI 1563/9–1563/35; 1 poorly preserved and incomplete columnal from Misina road (locality 5); MHI 1558/121.
Description: The columnals do not reach the maximum columnal width of 13.5 mm measured for this taxon in Upper Silesian specimens; the Mecsek columnals are ranging between 2.5 and 7.7 mm. They are low, cylindrical and may have an epifacet. The articular facet has multiradiate crenulation with very long culmina becoming slightly wider towards the periphery. The number of culmina is increasing towards the periphery by bifurcation and by intercalation of new culmina. The lumen is wide and generally simple; only in a few specimens it is extended as a lensoid spatium.

Geographical and stratigraphical range: S. silesiacus is a Tethydian species which immigrated into the eastern part of the Germanic basin through the Silesian–Moravian Gate during Early Illyrian times. In the Upper Silesian Muschelkalk it is the index species of the Lower Illyrian silesiacus biozone. It has been reported from Poland (Upper Silesia, Kraków Upland: Upper part of Karchowice Formation and lower part of Diplopora Dolomite; Holy Cross. Mts, Middle Muschelkalk; Tatra Mts: Anisian), Germany (Brandenburg, Rüdersdorf (?): Schaumkalk); Austria (Northern Calcareous Alps, Karwendel, Salzburg (Saalfelden, top of the Steinalmkalk, lower Illyrian), Radstädter Tauern, Steiermark, Leithagebirge: Anisian); Italy (Grigna, Giudicaria, Anisian, Pelsonian); Hungary (Balaton Highland, "Lower Muschelkalk age"; Mecsek Mountains). For more details compare Biese (1934) and Hagdorn et al. (1996).

Discussion: The distinction of *Silesiacrinus silesiacus* from *Eckicrinus radiatus* has been demonstrated by Quenstedt (1874–76) and, more forcefully, by Bather (1909). Nevertheless, the two crinoids are still confused. Therefore many notations of *Silesiacrinus* have to be referred to *Eckicrinus*. Major reasons for this treatment may have been the attribution of *Silesiacrinus silesiacus* to the parataxonomic genus *Entrochus* and the lack of good illustrations. *Eckicrinus* appears as early as lowest Pelsonian while *Silesiacrinus* does not appear before earliest Illyrian times.

5. Muschelkalk crinoids as biostratigraphic index fossils

As benthic organisms crinoids are facies-bound. Therefore they are basically not particularly suited as index fossils. Nevertheless, Triassic crinoids are valuable for biostratigraphic zonation because of their accelerated phylogenetic history, which from the Lower Triassic produced short-lived species with rapid morphological change. On the other hand, phylogenetic lineages with gradual morphological transformation could be observed among the Genus *Holocrinus* in the Germanic Basin. Otherwise, benthic crinoids may serve at least as excellent ecostratigraphic markers, which provide valuable evidence for the geographical and chronological distribution of certain habitats or facies types.

Complete articulated crinoid skeletons from the Germanic Muschelkalk "Lagerstätten" can be used to attribute dissociated sclerites to particular taxa. Thus it was possible to define not only the palaeogeographic distribution, but also the stratigraphic ranges of Muschelkalk crinoids by means of dissociated sclerites (Hagdorn 1985). For the eastern part of the Germanic basin a para- stratigraphical biozonal scheme has been established by Hagdorn and Gluchowski (1993) which is calibrated with the index biozones (ammonoids and conodonts) as well as with lithostratigraphic and sequence stratigraphic units. Some of these five crinoid biozones (the Upper Muschelkalk liliiformis zone of Late Illyrian/Early Fassanian age included) covering an interval from Lower Anisian to earliest Ladinian can also be recognized in the Triassic of the Alps, the Tatra Mountains, the Balaton Highland Triassic and the Muschelkalk of the Mecsek Mountains. However, more research will be necessary to refine this scheme by further bed-by-bed collecting aiming more detailed knowledge in the strati- graphic range of the significant taxa and the zonal boundaries.

Due to the terrestrial facies crinoids did not reach the Germanic basin or the Mecsek Mts during Lower Triassic times. However, as early as at Late Olenekian times, a still undescribed *Holocrinus* appeared with the Cencenighe transgression in the Werfen Formation of South Tyrolia (Cencenighe Member), and in the Balaton Triassic (Csopak Marls). The earliest Triassic crinoid in the Mecsek Mts, the Vicentinian Alps (Recoaro) and in Upper Silesia is the (? Late) Lower Anisian *Dadocrinus.*

Our collections and determinations give evidence that in the Mecsek Muschelkalk the following benthic crinoid biozones as defined by Hagdorn and Gluchowski (1993) can be recognized: (1) Dadocrinus zone (Lower Anisian) (2) acutangulus zone (Lower Anisian/Pelsonian), (3) dubius zone (Pelsonian/ Lowest Illyrian) and (4) silesiacus zone (Lower Illyrian).

For the Anisian and Lower Ladinian, such a parastratigraphic crinoid biozonation is particularly useful, because crinoid remains are very common and often preserved in facies types devoid of ammonoids. Identical crinoid assemblages from the eastern part of the Muschelkalk basin (Upper Silesia, Holy Cross Mts), from the Alpine region and from Hungary (Mecsek and Balaton) demonstrate that up to the beginning of the Illyrian there has been a fairly homogeneous facies domain starting with Lower Anisian wellenkalk facies with small soft ground communities containing *Dadocrinus* and *Holocrinus* and subsequent skeletal and oolite facies with firm ground and shell ground communities dominated by large encrinids (Hagdorn 1996; Hagdorn et al. 1996). Regional facies differentiation began in Ladinian times with the break-off of the carbonate platforms in the western Tethys realm, while an impoverished and fairly endemic fauna devoid of stenohaline crinoids and echinoids managed to survive under the continuously shallow marine conditions of the Germanic and Mecsek Muschelkalk basins.

Conclusions

Due to their rapid evolutionary radiation Middle Triassic benthic crinoids may serve as biostratigraphical index fossils. Several genera or even species have diagnostic characters in both crown or stem ossicles allowing their identification not only from complete specimens but also from such crinoid lagerstätten yielding dissociated sclerites exclusively. The parastratigraphic biozonal scheme established for the eastern part of the Germanic basin can be transferred to the Mecsek Muschelkalk and – partly – to the alpinotype Middle Triassic of the Balaton Highland. However, the stratigraphical ranges of the index taxa with their first and last appearance dates in typical sections have still to be prooved. More careful bed-by-bed collections of benthic crinoid remains from any Middle Triassic sections may finally lead to a reliable crinoid biostratigraphy allowing correlations over distant areas. Moreover, the palaeobiogeographical distribution of Middle Triassic crinoid faunas can be used as a valuable tool for estimating salinity gradients in different Peritethys basins.

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Plate I

Crinoid remains from Lapis Limestone

a–b. *Dadocrinus* sp., cylindrical columnals with multiradiate facets, on surface of wellenkalk bedding planes; Bükkösd, old quarry (locality 1); MHI 1560/1, 1560/2

c. *Dadocrinus* sp. at base of small load cast; Bükkösd, old quarry (locality 1); MHI 1560/3

d. *Holocrinus* sp. cf. *H. acutangulus*, subpentagonal to subcircular columnals and cirrals, from base of big load cast, Bükkösd valley (locality 2); MHI 1562/1. mm-scales

Plate II

Crinoid remains from "Coenothyris beds" of Bertalanhegy Limestone, Misina road cut (locality 3)

a–c. *Encrinida* fam. et gen. indet., barrel shaped columnals with multiradiate facets; MHI 1561/1, 1561/2

d–e. *Holocrinus* sp. cf. *H. acutangulus*, pentagonal nodals; MHI 1561/225, 1561/226

f–h. *Eckicrinus radiatus*, circular columnals with long multiradiate culmina, indistinct granulated radial bands and pyriform petal floors; MHI 1561/271–1561/273

i–j. *Eckicrinus radiatus*, circular columnals with long multiradiate culmina, indistinct granulated radial bands and petal floors from Brachiopod beds of Felsőörs Formation, Forráshegy near Felsőörs, Balaton Highland; MHI 1559/1, 1559/2

Plate III

Crinoid remains from Kozár Limestone
a–g, m. Misina road cut (locality 5), h–l Bükkösd valley, old quarry (locality 4)

a–b. *Chelocrinus carnalli*,

a. radial from oral; MHI 1558/4

b. second, axillary secundibrachial, distal muscular facet; MHI 1558/35

c–f. *Encrinidae* gen et sp. indet., columnals,

c. proximal nodal with epifacet; MHI 1558/53

d. distal cylindrical columnal with multiradiate facet; MHI 1558/54

e. proximal internodal with short marginal crenulae; MHI 1558/55

f. proximal subpentagonal nodal with radial bands; MHI 1558/56

g. *Holocrinus* sp. cf. *H. dubius*, pentagonal columnal with petaloid pattern; MHI 1558/61

h. *Eckicrinus radiatus*, proximal columnal with pyriform petal floors surrounded by adradial crenulae; MHI 1563/7

i–m. *Silesiacrinus silesiacus*, low cylindrical columnals with long multiradiate crenulation, crenulae bifurcating and intercalating towards the periphery

i–j. distal columnals with lumen sealed by accretionary calcite growth; MHI 1563/9-1563/10

k–l. pluricolumnals; MHI 1563/11–1563/12

m. columnal; MHI 1558/121. (Note that the scale for c–f and l–m is 4.5x)

Plate I

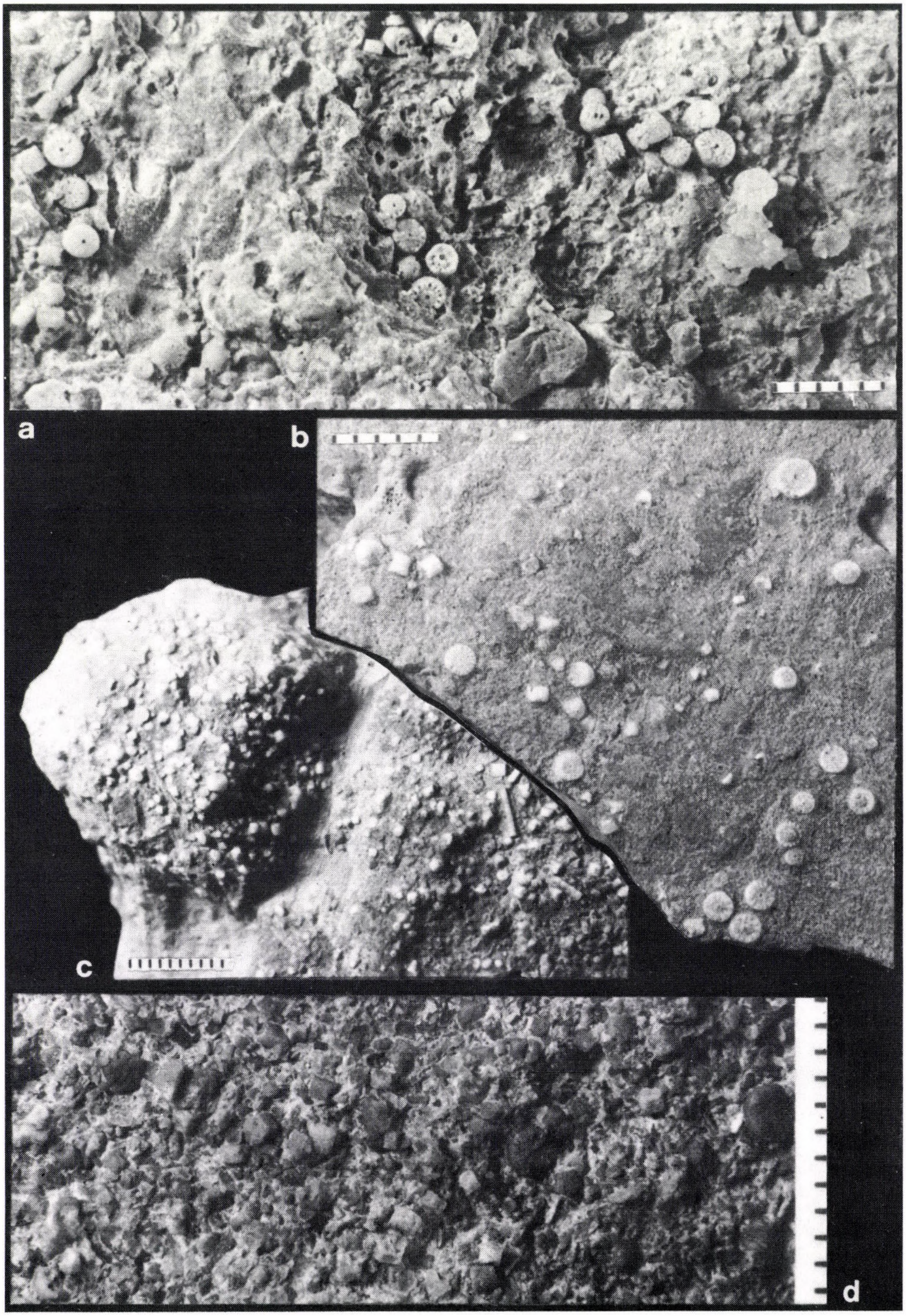


Plate II

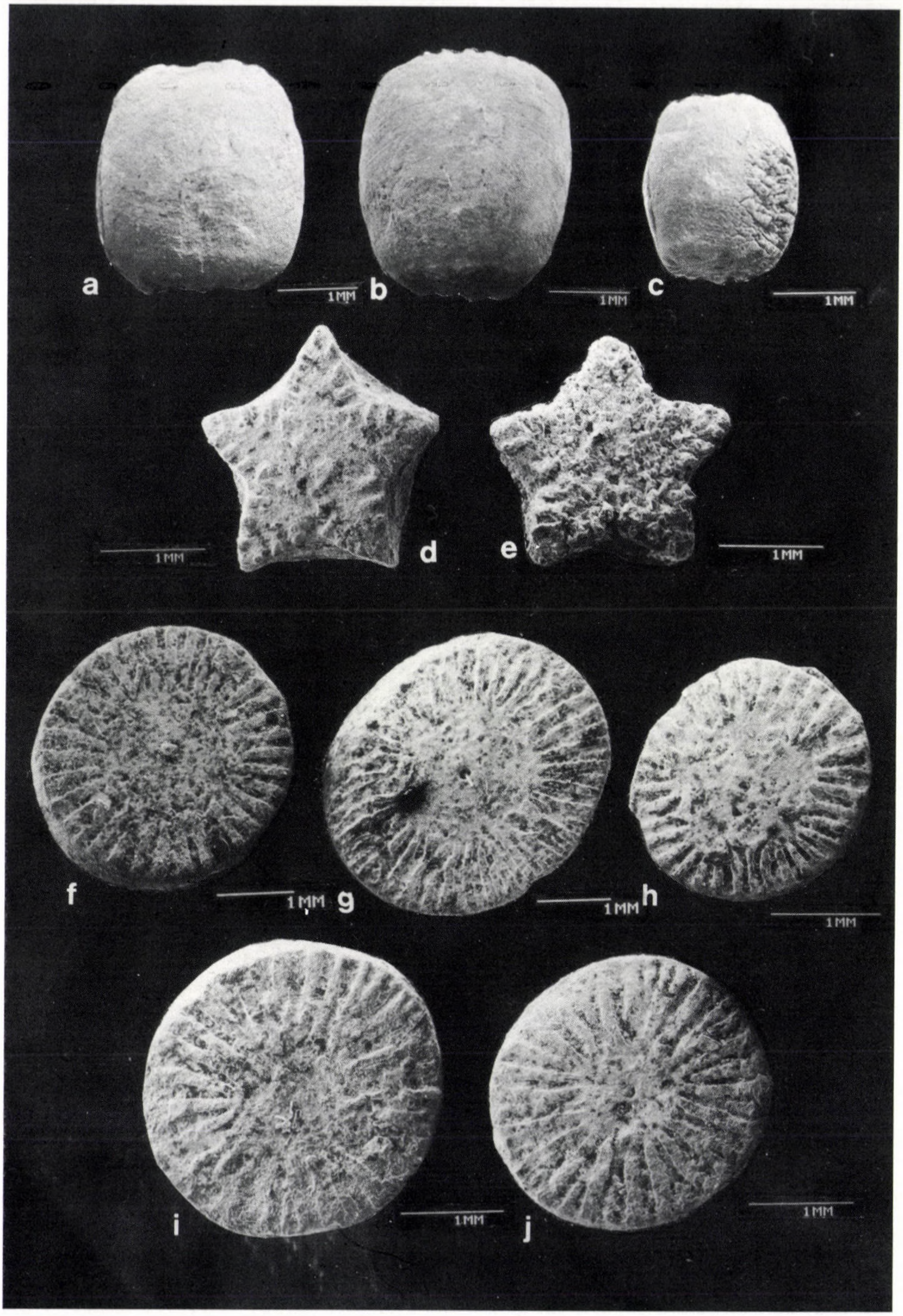
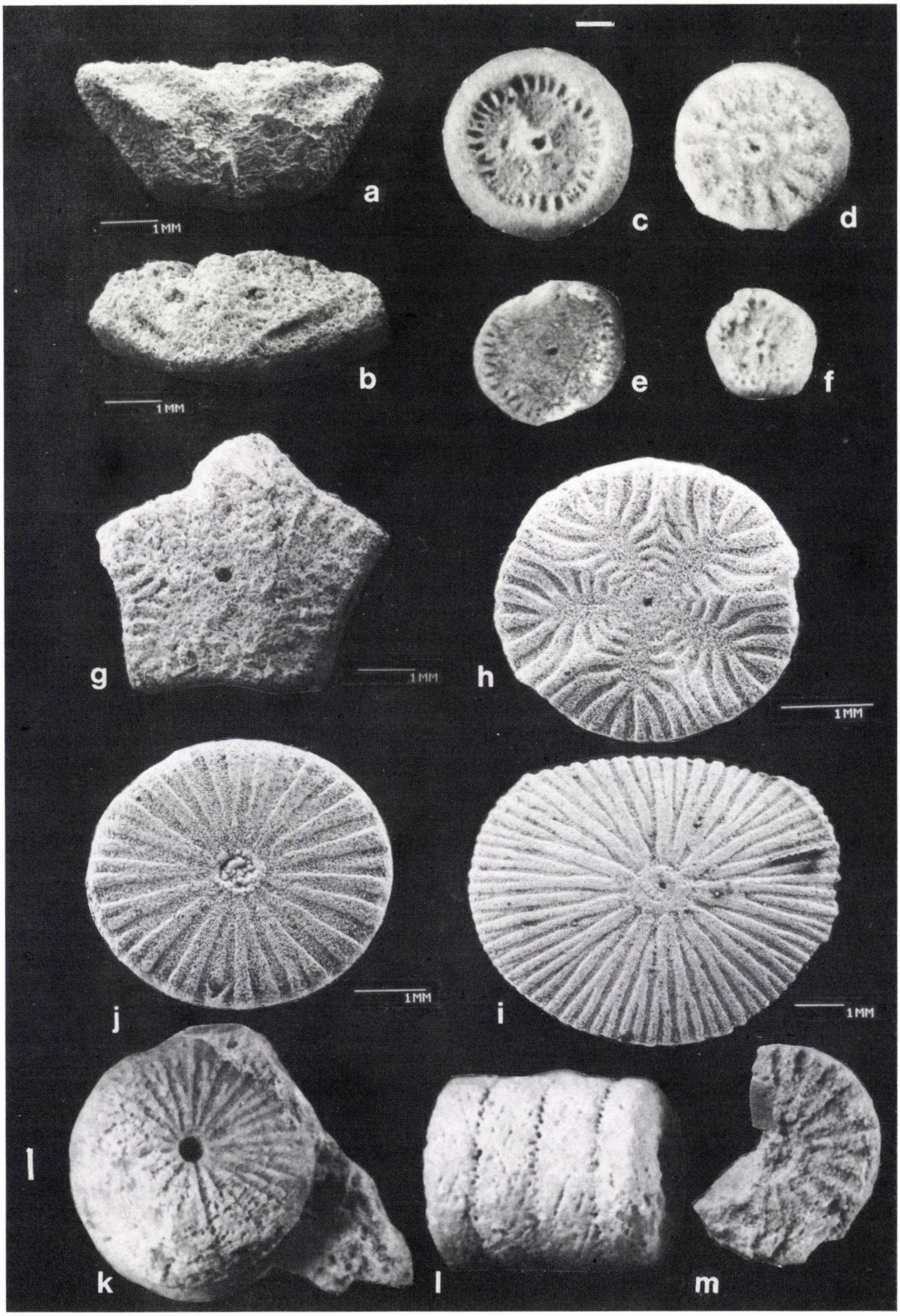


Plate III



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Bivalve assemblages from the Middle Triassic Muschelkalk of the Mecsek Mts, South Hungary: An overview

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Some units of the Middle Triassic "Muschelkalk" of the Mecsek Mts contain bivalves, often occurring as conspicuous shell-beds and pavements on bedding-planes. Corresponding to the shift of depositional environments from lagoonal through marginal marine to shallow marine settings and back, the diversity of the bivalve fauna increases upward in the sequence and reaches its highest value in the unit representing normal marine conditions. Salinity, energy level as well as consistency of substrate were presumably the main factors controlling the distribution of bivalves. Most of the species present in the Mecsek are widespread both in the Germanic and the Alpine Provinces. Some previously mis-identified taxa are revised. The most important forms and some previously inadequately illustrated ones are figured.

Key words: Hungary, Triassic, Muschelkalk, bivalvia

Introduction

Although the Muschelkalk, which is the middle part of the Triassic system of the German Facies Province was named for the mass appearance of brachiopods formerly mis-identified as bivalves (Muscheln), representatives of the latter group can also play a dominant *role* in benthic assemblages of this peculiar type of sedimentary rocks. Similarly to other occurrences of the Muschelkalk, bivalves are also frequent in the Middle Triassic of the Mecsek Mts. Although data on them have been accumulated for more than a century, this fauna is less known among palaeontologists. Nearly three decades have passed since the publication of the last comprehensive work, i.e. the valuable monograph of Nagy (1968), concerning this group of fossils. This long time, as well as the renewed attention to the Triassic of the Mecsek Mts justifies the presentation of a review. The aim of this paper is to give an overall picture, based on studies of museum material and of new collections, on the taxonomic composition, palaeoecology and palaeobiogeographical character of the bivalves of the Muschelkalk of the Mecsek Mts.

Geologic setting and stratigraphy

The Mecsek Mts are an isolated outcrop area of the Palaeo-Mesozoic succession of the Mecsek zone of the Tisza megatectonic unit, which forms the pre-Neogene

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basement of the Great Plain (see Bleahu et al. 1994). According to the most widely-accepted palaeogeographical reconstructions, the Tisza megatectonic unit belonged to the southern continental margin of Europe during at least the first half of the Mesozoic (Géczy 1973; Tollmann 1987; Vörös 1993).

Triassic rocks of the Mecsek Mts conformably overlie Permian continental sediments. Continental sandstones and conglomerates belonging to the Kővágószőlős and Jakabhegy Formations, respectively, also represent the Scythian. The coarse clastics are overlain by the silty Patacs Formation, deposited most probably in a mud-flat environment. The Anisian age of this unit was recognised only recently (Barabás-Stuhl 1993). The Patacs Formation and the overlying Hetvehely Fm. shows a marked shift of depositional environments from a pure siliciclastic littoral setting through a hypersaline lagoon (Magyarürög Evaporite Mb.) into a carbonate regime (Viganvár Limestone Mb.). This latter unit, largely representing lagoonal environments, is regarded here as the base of the up to 600 m thick Muschelkalk sequence *sensu lato*.

The marine rocks belonging to the Muschelkalk crop out in a nearly continuous belt NW of the town of Pécs, on the northern limb of the large pericline forming the western part of the Mecsek Mts. The sequence is also exposed, in a tectonically isolated position, near the village of Váralja at the eastern end of the Northern Thrust Belt (Fig. 1).

The detailed lithostratigraphy of the Middle Triassic is given in Fig. 2. The Mecsek Muschelkalk was deposited on a carbonate ramp (see Török 1993a) and the successive units show a general deepening-upward trend. The greatest depth was reached during the deposition of the Bertalanhegy Member (Rálisch-Felgenhauer et al. in Bleahu et al. 1994). The Kantavár Formation overlying the Muschelkalk contains freshwater fossils and most probably represents lacustrine environments (Monostori 1996). Fitting the Muschelkalk of the Mecsek Mountains into the standard scheme of Triassic zones or even stages is, due to the scarcity of index fossils, rather difficult. The rare ammonoids and conodonts have been yielded exclusively by the *Coenothyris* beds (Bertalanhegy Mb.) and indicate the Binodosus Zone of the Anisian (Detre 1973; Kovács and Papsová 1986). The age of the sediments underlying and following this unit, as well as the position of stage boundaries, are much less known. The lithostratigraphic subdivision is based on the work of Rálisch-Felgenhauer et al. (1993).

Previous studies

Triassic bivalves of the Mecsek Mountains were first recorded by Austrian geologists. Lipold (1858) found *"Gervilleia socialis"* in shales he believed to belong to the Kössen Formation. Peters (1862) in his pioneering work on the geology of the Mecsek Mts, mentioned some bivalves from the Triassic as well, among them *Myophoria goldfussi*.

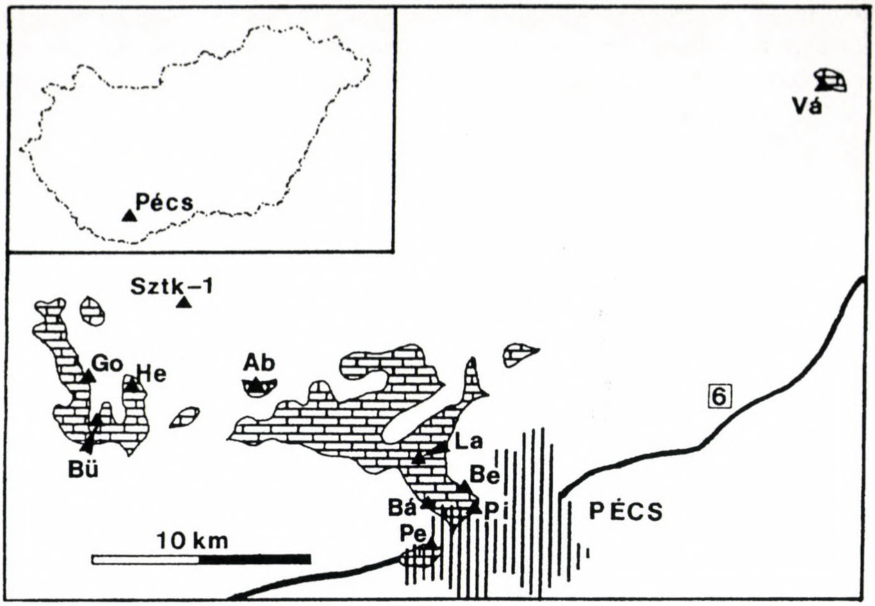

Fig. 1
Outcrop of the Muschelkalk in the Mecsek Mts with indication of localities mentioned in this paper. Sztk-1: borehole Szentkatalin-1; Go: Gorica; Bü: Bükkösd; He: Hetvehely; Ab: Abaliget; Bá: Pécs, Bárány Ravine; Pe: Pécs, Petrezselyem Spring; La: Lapis Hill; Be: Pécs, Bertalan Hill, Pi: Pécs, Piricsizma quarry; Vá: Váralja

LADINIAN	Csukma Formation	"Trigonodus Beds"	
		Kozár Mb.	Kán Dol.Mb.
ANISIAN	Zuhánya Lmst. Fm.	Dömörkapu Lmst. Mb.	
		Bertalanhegy Lmst. Mb.	
	Lapis Lmst. Fm.	Tubes Lmst. Mb.	
		Lapis Lmst. Mb.	
		Báránytető Lmst. Mb.	
	Rókahegy Dol. Fm.	Vöröshegy Dol. Mb.	
	Hetvehely Fm.	Viganvár Lmst. Mb.	
		Hetvehely Dol. Mb.	
		Magyarürög Evaporite Mb.	
	Patacs Fm.	Patacs Fm.	

Fig. 2
Lithostratigraphic subdivision of the Middle Triassic of the Mecsek Mts (after Rálisch-Felgenhauer et al. 1993). Units of the Muschelkalk are shaded

The eminent Hungarian geologist János Böckh (1881) contributed greatly to the knowledge of the Triassic faunas of the Mecsek Mts, and the extensive collections he made, housed till now at the Hungarian Geological Survey, partly form the basis of the present study as well. Böckh was invited by the community of Pécs (Fünfkirchen) to study the water supply opportunities of the area. His work resulted in a geologic map of the surroundings of the town as well as detailed descriptions of the successive lithological units, with special reference to their distinctive fossil content. Nearly all of the bivalves presently known from the Muschelkalk of the Mecsek Mts were already recognised by Böckh (l. c.).

During the eight decades after Böckh's work the Triassic of the Mecsek Mts received less attention. Only the name of Vadász can be mentioned who carried out geological mapping of the area before World War I (1914–1918). In his monograph on the geology of the Mecsek Mts Vadász (1935) provided a list of fossils from the Muschelkalk as well.

The geologic and palaeontological re-study of the Mecsek Mts began started in the early sixties. Results of biostratinomical studies of the "*Myophoria* beds" now attributed to the Viganvár Member were published by Jámbor (1967). Nagy (1968) was the first to illustrate bivalves from the Muschelkalk. Based on studies of new collections and museum material, he provided lists of Triassic fossils and illustrated the most characteristic ones. Nagy's monograph is by far the most comprehensive description of the Triassic succession and faunas of the Mecsek Mts until now.

Recently Török (1986, 1993a, b) investigated the sedimentology and fauna of the Muschelkalk of the western Mecsek Mts.

Taxonomic composition and palaeoecology of the bivalve assemblages

In the following, characteristic bivalve assemblages of the successive lithological units are briefly discussed. Bivalves do not appear evenly distributed throughout the sequence, but are concentrated in some levels, commonly occurring as dense shell-beds or pavements on bedding planes. According to Fürsich and Aberhan (1990), both autochthonous and allochthonous time-averaging can be significant in shallow-shelf environments, so that these skeletal concentrations can be considered only as mixed relics of the former communities. Some associations described from other areas can be, however, recognised in the Mecsek Mts as well.

In order to provide a sound basis for further palaeobiogeographical and palaeoecological studies, the most important species as well as some previously not or inadequately illustrated ones are presented. A full systematic description of the fauna is in progress and planned to be published in the future. Although some of the identifications published there are out of date, the general character of the fauna as well as the stratigraphic distribution of bivalves can be understood from the work of Nagy (1968).

Hetvehely Fm.

The lowest part of the Mecsek Muschelkalk, the Viganvár Member of the Hetvehely Fm., consists of thin-bedded dark, dolomitic limestones, deposited in lagoonal environments. Bivalves are preserved with shell and usually occur as monospecific or oligospecific pavements. Small, modioliform bivalves assigned to "*Gervilleia modiola* Frech" by Nagy (1968, pl. 2, figs 8, 11, 12) are dominant in the lower part of this unit. The internal features of these valves cannot, however, be studied, so their presumed bakevelliid affinity is doubtful. They represent more probably *Modiolus triquetrus* Seebach, 1861 or *M. salzstettensis* Hohenstein, 1913. The inarticulate brachiopod *Glottidia tenuissima* is also frequent in these beds.

Conspicuous and often exclusive occurrence of *Costatoria goldfussi* (Alberti in Zieten, 1830) characterises the upper part of the Viganvár Member. Formerly this part of the succession was supposed to be of early Triassic age, and for this reason the abundant *Costatoria* remains were identified as *C. costata* (Zenker, 1833) (Böckh 1881; Nagy 1968). Differences between these two widespread species were more recently discussed by Farsan (1972) who stated that the number of ribs of *C. costata* is lower than that of *C. goldfussi*, its valves are more elongated (see e.g. Gronemeier and Martini 1973, pl. 16, figs 1–3), and its posterior margin is curved. The Mecsek specimens bear 9–13 ribs and their posterior margin is markedly truncated (Plate I, Fig. 1). *Bakevellia costata* (Schlotheim, 1820), *"Gervilleia" mytiloides* Schlotheim, 1820 and *Pseudocorbula* cf. *gregaria* (Münster in Goldfuss, 1838) are the common associated faunal elements (Plate I, Fig. 2).

Márquez-Aliaga et al. (1986) interpreted *C. goldfussi* as a probably epiphytic organism. Recent trigoniids, among them species ornamented with strong radial ribs (e.g. *Neotrigonia margaritacea* (Lamarck) are, however, shallow-burrowing forms well adapted to high-energy near-shore marine environments (Stanley 1977), and probably a similar mode of life can be inferred for *C. goldfussi* as well. The low diversity of the fauna suggests extreme environmental conditions during deposition of the Viganvár Member. The lack of stenohaline organisms indicates abnormal, probably elevated salinity values.

Lapis Limestone Fm.

The most characteristic unit of the Mecsek Muschelkalk, the Lapis Limestone Fm., is a typical vermicular limestone sequence equivalent to the Germanic "Wellenkalk". It was deposited in a restricted lagoonal zone behind a wide dolomite platform covering large areas of the Tisza megatectonic unit during the Middle Anisian; at present it forms the bulk of the forest-covered hills overlooking the city of Pécs. Excepting crinoids the benthic macro-invertebrate assemblage of the Lapis Limestone consists only of bivalves and gastropods.

The Báránytető Member is a sequence of thin dolomitic limestone beds poor in fossils. Scarce remains of epibenthic and semi-inbenthic suspension feeder bivalves such as *Entolium discites* (Schlotheim 1820), *Leptochondria albertii* (Goldfuss 1838), *Modiolus triquetrus* Seebach 1861, as well as poorly preserved bakevellids have been yielded by this unit.

The Lapis Limestone Member consists of alternating beds of grey limestones with small nodules, of bioturbated limestones with the trace fossils *Rhizocorallium* and *Balanoglossites* and of crinoidal limestones. Laminated limestones and those with "S-Faltung", as well as storm-generated coquinas, are also frequent. The bivalve assemblage is moderately diverse, but usually poorly preserved. Some shell-beds are clearly dominated by endobenthic suspension feeders. Very striking bedding planes, with pavements of bivalves assigned to *Pleuromya* cf. *elongata* (Schlothei, 1822) can be seen at the type locality, near the road from Pécs to Lapis Hill. The exhumation of valves of deep burrowing forms indicates that these shell beds were generated by very severe storms (Plate I, Fig. 4). Subordinately, assemblages dominated by semi-infaunal bivalves and gastropods can also be found (Plate II, Fig. 8).

Among the bivalve species appearing first in this unit *Lyriomyophoria elegans* (Dunker 1848) (Plate II, Fig. 13) is the most characteristic. This shallow burrowing form occurs frequently in shelly, often crinoidal limestones, in association with the epibyssate *Entolium discites* (Schlotheim 1820) (Plate II, Fig. 10). The preference for coarse substrate and high-energy environments displayed by *Lyriomyphoria* was already observed by Hagdorn (1991). Pavements of *Pseudocorbula gregaria* (Münster in Goldfuss 1838) valves also occur on bedding planes.

The increased number of bivalve species as compared to the Viganvár Member indicates more favourable environmental conditions for benthic life. The very low diversity of definitely stenohaline taxa suggests that salinity was probably a main factor controlling distribution of organisms, which was, as inferred from the different guild structure of the assemblages, also influenced by the consistency and grain size of the substrate and the energy level.

Zuhánya Formation

The Bertalanhegy Member of the Zuhánya Formation is a sequence of nodular marl and limestone beds. Articulate brachiopods are confined to this unit in the Triassic of the Mecsek Mts, which is called "Coenothyris beds" after the overwhelming abundance of *Coenothyris vulgaris* (Schlotheim 1820), (see Detre et al. 1986, 1992; Török 1993b).

This unit yielded the most diverse bivalve fauna. Epifaunal and semi-infaunal taxa dominate while *Pleuromya* (Plate I, Fig. 3), *Arcomya* and other endobenthic forms are usually less frequent. Pteriomorphs are usually preserved with shell, while other taxa can be found as moulds. In addition to most of the species already occurring in the Lapis Fm., several forms appear exclusively in the *Coenothyris* beds. Among them *Hoernesia socialis* (Schlotheim, 1823) (Plate II, Fig. 1) is the most common. Limids, such as *Plagiostoma lineatum* (Schlotheim

1823) (Plate II, Fig. 7), *Plagiostoma striatum* (Schlotheim, 1823) (Plate II, Fig. 6) and *Plagiostoma cf. costatum* (Goldfuss 1838) (Plate II, Fig. 5) are also confined to this unit, as are *Enantiostreon difforme* (Schlotheim 1823). *Placunopsis ostracina* (Schlotheim 1820) encrusts *P. lineatum*, which latter often covers bedding planes ("Lima-Platten"). *Bakevellia (B.) costata* (Plate II, Fig. 2), *Entolium discites, Leptochondria albertii* (Plate II, Figs 3–4), and *Pseudocorbula gregaria,* the most persistent forms in the Middle Triassic of the Mecsek Mts, also appear in the Bertalanhegy Member.

Bivalve-dominated assemblages have been found in this unit only in two cases. At Gorica the infaunal shallow burrowing bivalve *Pseudocorbula gregaria*, identified by Nagy (1968) and Török (1986) as *Nucula* sp. constitutes about 90% of the fauna, while *Coenothyris vulgaris* (10% of the 169 specimens counted) is subordinate. A mass occurrence of *Pseudocorbula gregaria* was reported recently by Márquez-Aliaga et al. (1986) from the Triassic of SE Spain.

Near Váralja *Lyriomyophoria elegans* (Plate II, Fig. 12) have been found abundantly, in association with crinoid ossicles, rare *Hoernesia socialis* and *Tetractinella trigonella.* The myophoriid *Neoschizodus laevigatus* (Alberti 1834) (Plate II, Fig. 11) occurs sporadically.

A small, single pectinid identified as *Pecten margheritae* Hauer, 1850 was illustrated by Nagy (1968, pl. 5, fig. 4) from the Bertalanhegy Member exposed in the Piricsizma quarry at Pécs. The specimen, which is an incomplete left valve, is now identified as *Praechlamys* cf. *reticulatus* (Schlotheim 1823) (Plate II, Fig. 9).

P. margheritae described from the Ladinian? of the Southern Alps was not included in the revision of Triassic pectinids by Allasinaz (1972), and except for the original description only very little information has been available on it. Authors of the present paper (ISz) studied the syntypes housed at the Geologische Bundesanstalt in Vienna (two specimens, registration number 1851/01/35) and found that they differ both from one another and the Mecsek specimen in their dimensions and in the arrangement and density of the plicae. The Mecsek specimen, on the other hand, shows features similar to those of *P. reticulatus* illustrated in the literature. Comarginal plicae, which result in a reticulate pattern when interacting with radial ones in typical "reticulatus" fashion, are less prominent. The lack of such ornamentation can be plausibly explained with the small size of the valve, since comarginal plicae seem to occur only on large, presumably adult specimens and the juvenile region of the valves bear only radial plicae (see Assmann 1937, pl. 11, figs 17–18; Hagdorn and Simon 1993, p. 201, fig. 9). Considerable intraspecific variability of plication demonstrated in Jurassic *Preachlamys* species by Johnson (1984) and Szente (1996) can also be inferred in *P. reticulatus.*

Environmental energy, substrate conditions (consistency and grain size), and salinity were most likely the main factors controlling distribution of bivalves during the deposition of the Coenothyris beds. The abundance of byssally attached epibenthic and semi-inbenthic organisms suggests the availability of favourable hard substrate. According to Fürsich (1993), Mesozoic benthic communities dominated by *Hoernesia* and *Bakevellia* lived in brackish-marine, brachyhaline environments. These genera, as well as *Entolium* thought to be confined to the upper part of the brachyhaline regime, are always associated with articulate brachiopods in the Bertalanhegy Member and do not form distinct assemblages.

Csukma Formation

The "*Trigonodus* Beds" of 1–2 m thickness is the uppermost bivalve-bearing unit and is characterised by the mass occurrence of large, thick shells most probably belonging to *Trigonodus* Sandberger in Alberti 1865. Occasionally small, indeterminate gastropods are the only associated faunal elements. Repeated attempts were made to extract better-preserved shells from the very indurated limestone in order to study internal features of the valves, but they have remained unsuccessful. The very low diversity of this assemblage and the total faunal turnover, as well as the lack of definitely marine organisms so frequent in the underlying units, all indicate extreme environmental conditions, most probably a strongly lowered salinity. The impoverished fauna of this unit does not seem to correspond to the moderately rich assemblage of the "Trigonodusdolomit" of the Germanic Facies Province (see Herb 1957).

Palaeobiogeographical affinities of the bivalve fauna of the Muschelkalk of the Mecsek Mts

Results of sedimentological and palaeontological investigations carried out during the last decades suggest that the Triassic of the Mecsek Mts shows affinities to the Germanic Facies Province (Nagy 1968; Pálfy and Török 1992; Török 1993a). Recently, based partly on admittedly existing differences between the fauna and stages of environmental evolution of the Mecsek and those of the Germanic Basin, Kozur and Mock (1987) doubted the Germanic character of the former succession. A brief palaeobiogeographical evaluation of the bivalve fauna is, therefore, justified. It is worth mentioning, however, that bivalves usually do not serve as suitable tools in resolving palaeogeographical problems since they are almost exclusively benthic organisms whose distribution is strongly controlled by local environmental factors. The vast majority of bivalves develop through planktotrophic larval stage, giving them considerable dispersal potential.

Márquez-Aliaga et al. (1986) analysed the palaeobiogeographical pattern of Middle Triassic bivalves of Europe and, based on the presence of characteristic species, defined three provinces (Sephardic, Germanic, and Alpine, respectively). *Enantiostreon flabellum* (Schmidt 1935), *Gervillia joleaudi* (Schmidt 1935) and *Placunopsis teruelensis* Wurm 1911 were selected as characteristic of the Sephardic Province. These species were not found in the Mecsek Mountains.

On the other hand, all of the species recorded from the Mecsek Mts are known both from the Germanic and the Alpine Province, with the Germanic endemic *Praechlamys* cf. *reticulatus* as an exception. Several species have also been documented from remote areas (see Farsan 1972).

One bivalve species previously recorded from the south Transdanubian Muschelkalk would have been of palaeobiogeographical importance. Wéber (1992, pl. 2, fig. 1) illustrated a myophoriid bivalve identified as *Myophoria kefersteini* from borehole Szentkatalin-1 drilled in the northern foreland of the western Mecsek Mountains. *M. kefersteini* (Münster 1828) is a very variable species confined to the Carnian Raibl Formation of the Southern Alps (see Diener 1923; Kutassy 1931; Lieberman 1979; Fantini Sestini 1966).

The specimen, which was kindly sent us by Béla Wéber, is an approx. 6 mm long left valve bearing a marked umbonal sulcus. It is interesting that no similar bivalve has been found on the surface until now. The umbonal sulcus is, however, not a distinguishing feature of *M. kefersteini* since it is possessed by other species of *Myophoria* as well, for example by the well-known *Myophoria vulgaris* (Schlotheim 1820), or *M. transversa* (Bornemann 1856); see Müller (1985) or Baumgarte (1975). The occurrence of *M. kefersteini* in the Triassic of south Transdanubia thus cannot be confirmed.

The composition and low diversity of the Mecsek assemblages recall the Germanic rather than the Alpine Province.

Similarly to the changes in sedimentary environments, the stratigraphic distribution of bivalves as well as the occurrence of bivalve-dominated biofacies were controlled by local factors and do not correspond strictly to those in the Germanic Basin.

Conclusions

The Middle Triassic Muschelkalk of the Mecsek Mts yielded a moderately diverse bivalve fauna. Changes in the bivalve assemblages reflect the shift of sedimentary environments from lagoonal and marginal marine to shallow marine settings and back. Salinity, energy level as well as grain-size and consistency of substrate were the main factors controlling the distribution of bivalves. Although nearly all of the species found in the Mecsek Mts are known both from the Germanic and Alpine Provinces, the composition and the low diversity of the fauna resembles more the assemblages of the Germanic Muschelkalk than those of the coeval Alpine formations.

Acknowledgements

Mr. Béla Wéber (Pécs) made the bivalves he collected borehole Szentkatalin-1 available for study. Several specimens were donated by Dr. Gyula Konrád (Abaliget). Their help and kindness is gratefully acknowledged. This work was supported by the OTKA (Grant No. 2671).

Plate I

The specimens are coated with ammonium chloride. The letter T and four-figure numbers are registration numbers of specimens housed at the Hungarian Geological Museum. Other specimens are unregistered and are housed at the Department of Palaeontology, Eötvös University.

1. *Costatoria goldfussi* (Alberti in Zieten, 1830). Pécs, Bárány Ravine, Viganvár Member, T 4976, 1x.
2. *Costatoria goldfussi* (Alberti in Zieten, 1830); *Bakevellia* cf. *costata* (Schlotheim, 1822); *"Gervilleia" mytiloides* Schlotheim, 1820); *Pseudocorbula gregaria* (Münster in Goldfuss, 1838). Hetvehely, Viganvár Member, T 4821, 1x.
3. *Pleuromya* sp. Bükkösd, Bertalanhegy Member, 1x.
4. Bedding plane with *Pleuromya* cf. *elongata* (Schlotheim, 1822 Abaliget, Lapis Member, T 4884, 1x.

Plate II

1. *Hoernesia socialis* (Schlotheim, 1823). Bükkösd, Bertalanhegy Member, 1.2x.
2. *Bakevellia costata* (Schlotheim, 1823). Bükkösd, Bertalanhegy Member, 1.5x.
3. *Leptochondria albertii* (Goldfuss, 1838). Gorica, Bertalanhegy Member, T 4818, 2x.
4. *Leptochondria albertii* (Goldfuss, 1838). Gorica, Bertalanhegy Member, T 4820, 1.5x.
5. *Plagiostoma* cf. *costatum* (Goldfuss, 1838). Bertalan Hill, Bertalanhegy Member, T 4963, 1x.
6. *Plagiostoma striatum* (Schlotheim, 1823). Gorica, Bertalanhegy Member, T 4828, 1.
7. *Plagiostoma lineatum* (Schlotheim, 1823). Bükkösd, Bertalanhegy Member, 1x.
8. Characteristic bedding plane from the Lapis Member with poorly preserved *Pleuromya* sp., *"Gervilleia" mytiloides, Pseudocorbula* sp. and indeterminate gastropods. Bükkösd, Lapis Member, T 4811, 1x.
9. *Praechlamys* cf. *reticulatus* (Schlotheim, 1823). Pécs, Piricsizma quarry, Bertalanhegy Member, 2x (figured Nagy, 1968, pl. 5, fig. 4 as *Pecten margheritae* Hauer, 1850).
10. *Entolium discites* (Schlotheim, 1820). Pécs, Petrezselyem Spring, Lapis Member, T 4951, 2x.
11. *Neoschizodus laevigatus* (Alberti, 1834). Bükkösd, Bertalanhegy Member, 1x.
12. *Lyriomyophoria elegans* (Dunker, 1848) (internal moulds). Váralja, Cigánysor quarry, Bertalanhegy Member, 1x.
13. *Lyriomyophoria elegans* (Dunker, 1848). Pécs, Petrezselyem Spring, Lapis Member, T 4951, 2x.

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Plate I

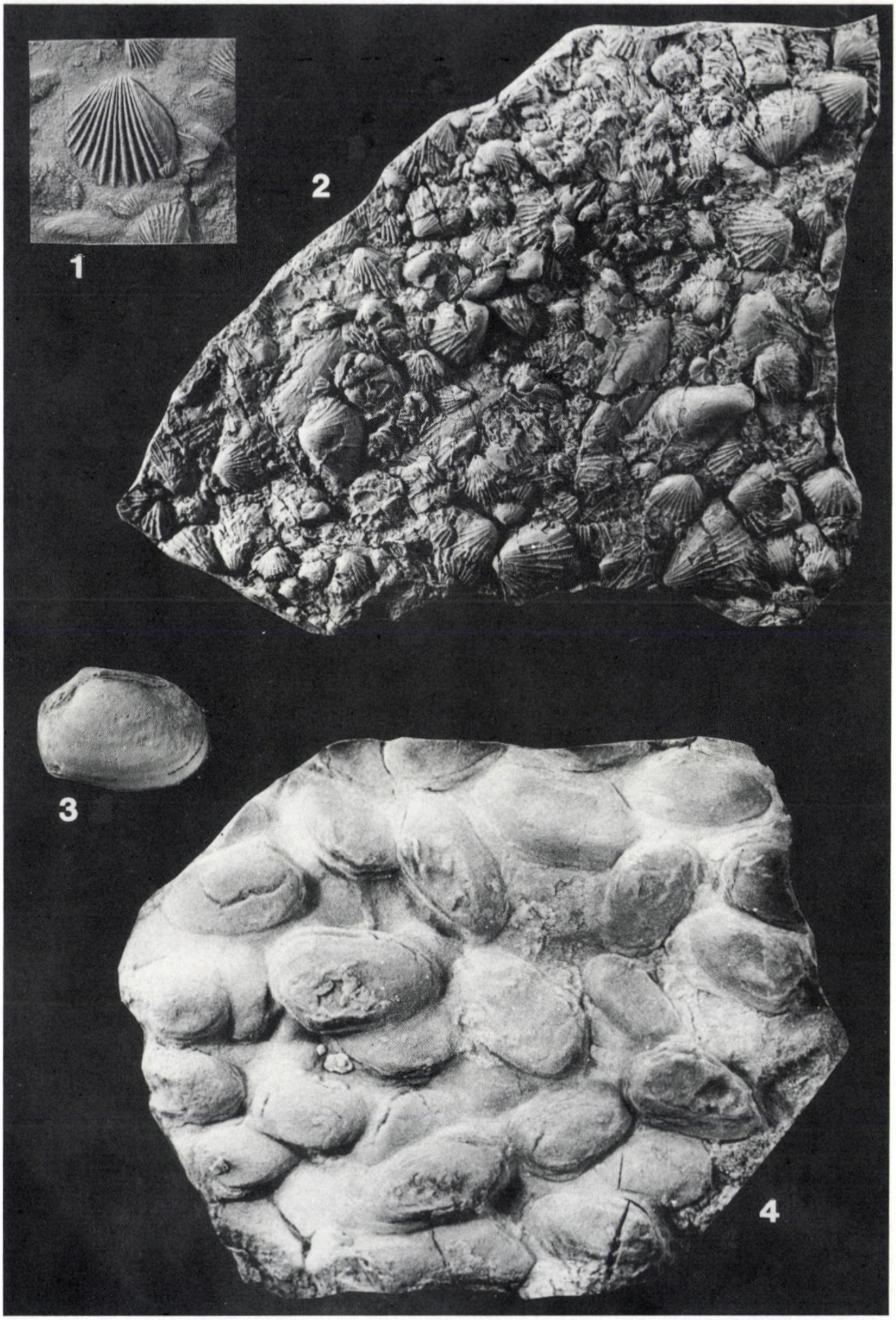
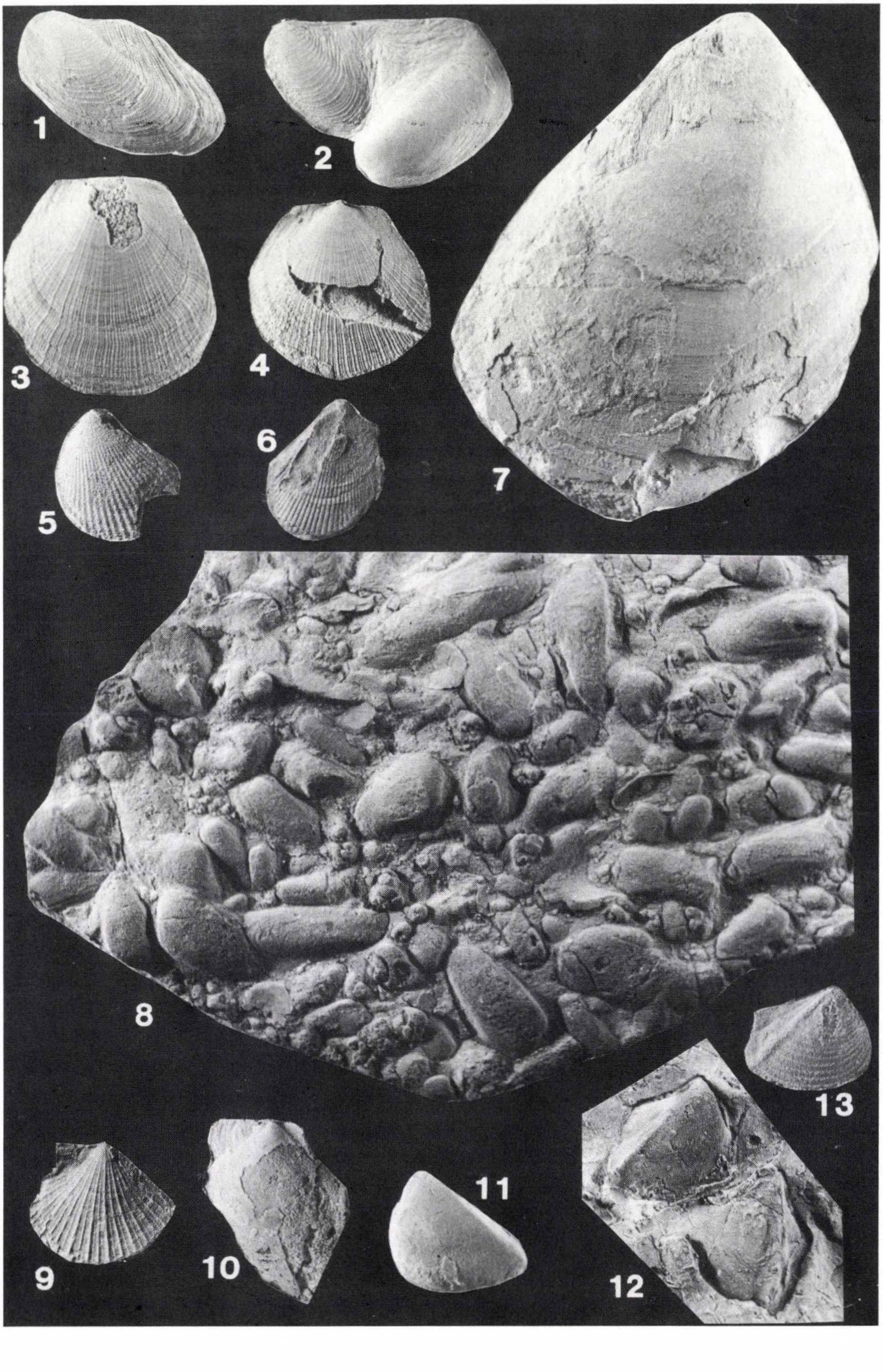


Plate II



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Folding in the Abaliget road cut (Mecsek Mts)

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Structures of a road cut exposure near Pécs, in the Mecsek Mts (SW Hungary) were revisited and analysed in detail. The section cuts through Lower Anisian folded carbonates and clays. Two different styles of folding were found: an asymmetric flexural slip folding with slightly reversed NW limbs producing major folds and a more parallel, disharmonic, box fold-type folding producing smaller folds. These latter folds are associated with decollements and ramps, most often in the long normal limbs of the major asymmetric folds. Most structures are NW-vergent and are related to Late Cretaceous tectonogenesis. Different fold styles in the same development are probably the result of ongoing shortening in the vicinity of an underlying regional detachment horizon (evaporites). Later normal faults cut earlier structures.

Key words: asymmetric folds, decollements, ramps, pressure-solution cleavage, reconstruction

1. *Introduction*

Structural studies in the Mecsek Mountains go back at least 50 years. Authors dealing with the Mesozoic (Vadász 1935; Wein 1967a, b; Némedi Varga 1983) had already stated that the Early-Middle Triassic strata were deposited during an extensional episode. Vadász (1935) postulated a Late Triassic uplift phase with denudation, which was denied by Wein (1961) based on his observations in the eastern part of the mountain. In the Jurassic a very thick sedimentary sequence was deposited, followed by flows of Early Cretaceous volcanics in an extensional phase (Harangi and Árva-Sós 1993). After the Early Cretaceous the area was uplifted. The first important shortening in a NW–SE direction occurred during the Late Cretaceous (probably the austrian phase). Wein (1967b) assumed a two-stage folding which was strong enough to produce overturned, imbricated folds. During the Miocene the Mecsek Mountains were subjected to normal faulting (Wein 1967b; Hámor 1970; Bergerat and Csontos 1988; Csontos and Bergerat 1992) without having been covered by a thick sedimentary

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overburden (Dunkl 1992). This extensional period was followed by another shortening in the Late Miocene–Pliocene.

In the present study we give the structural description of a key stratigraphic section, in order to contribute to the knowledge of the tectonic evolution of the area. The Mecsek Mountains is one of the important outcrop areas of the Tisza unit (Fig. 1; see also Kovács 1982; Balla 1984; Brezsnyánszky and Haas 1991; Csontos 1995); thus the results of the study could be extrapolated to parts of the unit covered by thick Tertiary deposits (Great Hungarian Plain). Surprisingly there has been no detailed description of deformed Triassic sections yet in the Mecsek Mountains.

The selected section is located in the road cut of the Pécs–Abaliget road, between kilometres 6.9 and 7.2 (Fig. 1). Parts of this section were published, however, in some earlier works (Brezsnyánszky 1984; Rálisch-Felgenhauer 1988). The entire section was reinterpreted for this study.

2. *Stratigraphy*

The detailed stratigraphic description of the outcrop was performed within the framework of the Key sections program of the Hungarian Geological Institute (Rálisch-Felgenhauer 1988). The formation is composed of alternating thick and thin beds of dark, micritic limestone and thin clay beds, attributed to the Lower Anisian (early Middle Triassic) Viganvár Limestone Formation. The carbonates were dolomitized in some places. A few layers contain storm deposits. Some fossil-rich layers (geopetals) permit a more detailed facies analysis (see Török, this volume). The more competent layers are often fractured; several calcite precipitations are seen along these fractures. For the stratigraphic description an overturning of the sequence was not considered. The intensive folding of the rocks was attributed to the close vicinity of a stratigraphically underlying evaporite. Our goal was to complete this description with structural information.

3. *Methodology*

In the outcrop bedding plane attitudes, fold axes, different types of cleavages and fault slip data were collected. Dips are indicated with azimuth of dip followed by the angle of dip. The relative chronology of different structural elements was also noted. Detailed drawings were made of important parts.

A photomontage of the outcrops was prepared. A line drawing of this photograph was compared with and completed by more detailed drawings of the outcrop, then once again controlled in the field. Present-day outcrop conditions were also compared to another photomontage made a decade ago by a university field camp group (Konvalinka et al. 1983). This earlier picture gave details on those parts of the section now covered.

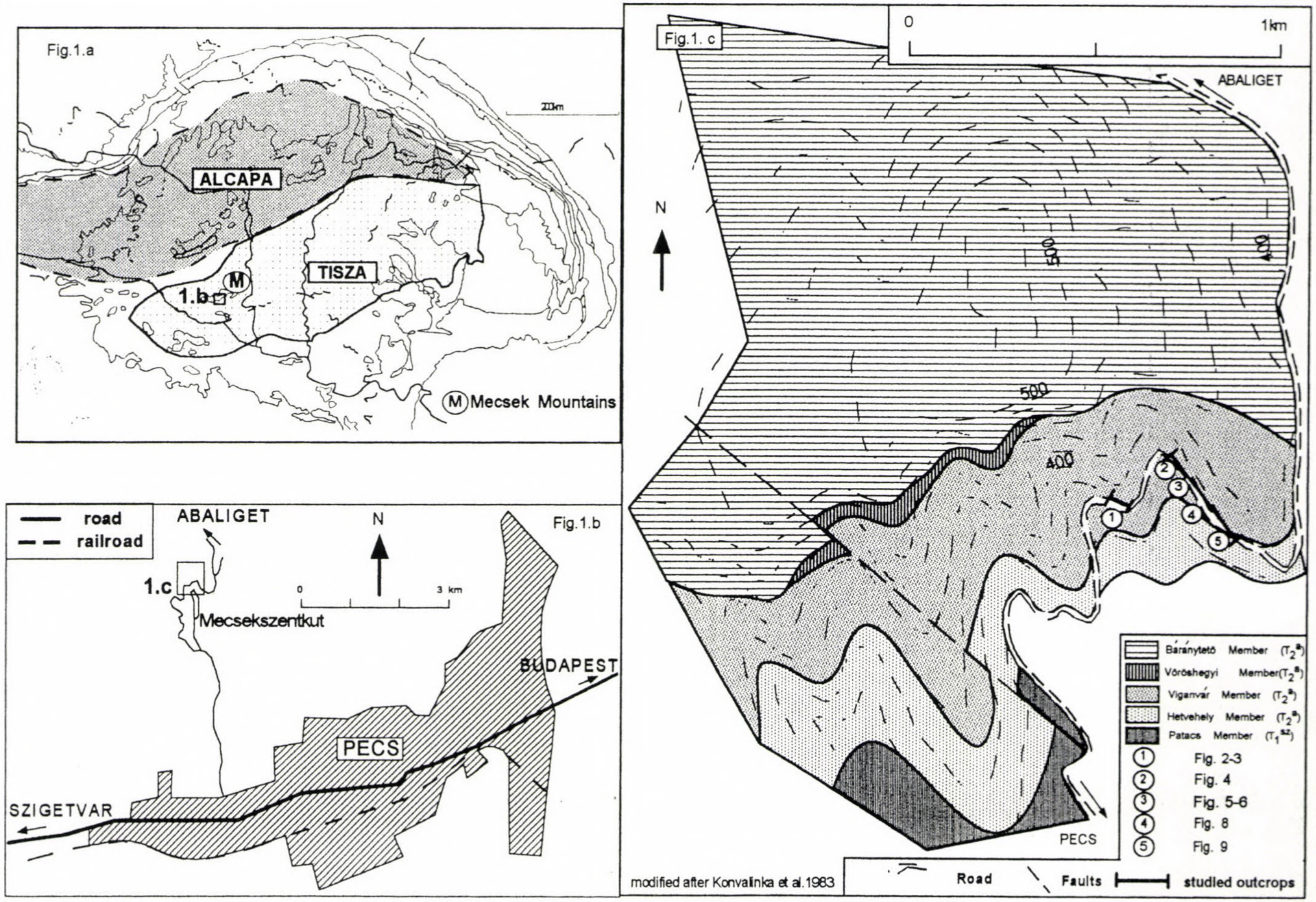

Fig. 1
Simplified geologic map of the studied section, based on the field mapping report of Konvalinka et al. (1983). Inserts: a, Location of the Mecsek Mts. in the Carpathian area (based on Csontos 1995); b, Location of the geologic map within the Mecsek Mts

4. Description

The section of the Pécs–Abaliget road above the small shrine of Mecsekszentkút was visited and described from Pécs to Abaliget. The section is composed of five more or less distinct parts of good outcrop, linked by more or less covered segments (Fig. 1). The outcrops form a series of meso-scale anticlines, with short and steep, sometimes overturned NW limbs and longer, flat-lying normal limbs. Outcrop-scale higher rank structures are found in both limbs. Details of the structure are described in five sites. A view of the entire structure and an interpretation is given in a later paragraph.

Site 1

The outcrop (Figs 1, 2) is found in the turnpike just above the shrine of Mecsekszentkút. The Anisian formation is strongly folded. A talus covers the middle part of the exposure, inhibiting the correlation of most layers. The western part of the outcrop is composed of three greater folds, some of which carry smaller-order parasitic folds (Figs 2, 3a). The parasitic folds indicate an overturned limb and drag along the bedding planes in a flexural slip fold.

In the core of the third, partly overturned fold, the axis generally measured throughout the outcrop (040/10) gradually turns upward into the axial direction 035/12 of a straight fold (Fig. 3b). This suggests a more pronounced shear in the core of the fold than on its envelope.

Site 2

The next larger outcrop area is found after a sharp turn of the road (Fig. 1). This exposes a flat, open, upright anticline and its flank dipping more and more steeply toward the southeast (Fig. 4). On the SE limb smaller parasitic folds of NW vergency are seen. The fold is cut and offset by a smaller normal fault dipping toward the northwest.

Between sites 2 and 3 smaller exposures are found along the road, but the quality of these outcrops does not enable us to interpret this part. The observations made in this part of the section are reported, however, on the integrated cross-section. Folding style as well as measured axes are similar to the previously described ones. Some steeply-dipping layers are present, however.

Site 3 (small parking place)

This is one of the better-exposed parts of the section (Fig. 1), and was illustrated previously (Brezsnyánszky 1984; Rálisch-Felgenhauer 1988). The section is divided into a lower, less deformed, and a higher, intensively folded compartment (Fig. 5). The difference in folding style implies a detachment surface between the two compartments, part of which was indeed found and measured. The flat-lying lower layers of the SE part can be followed up to the

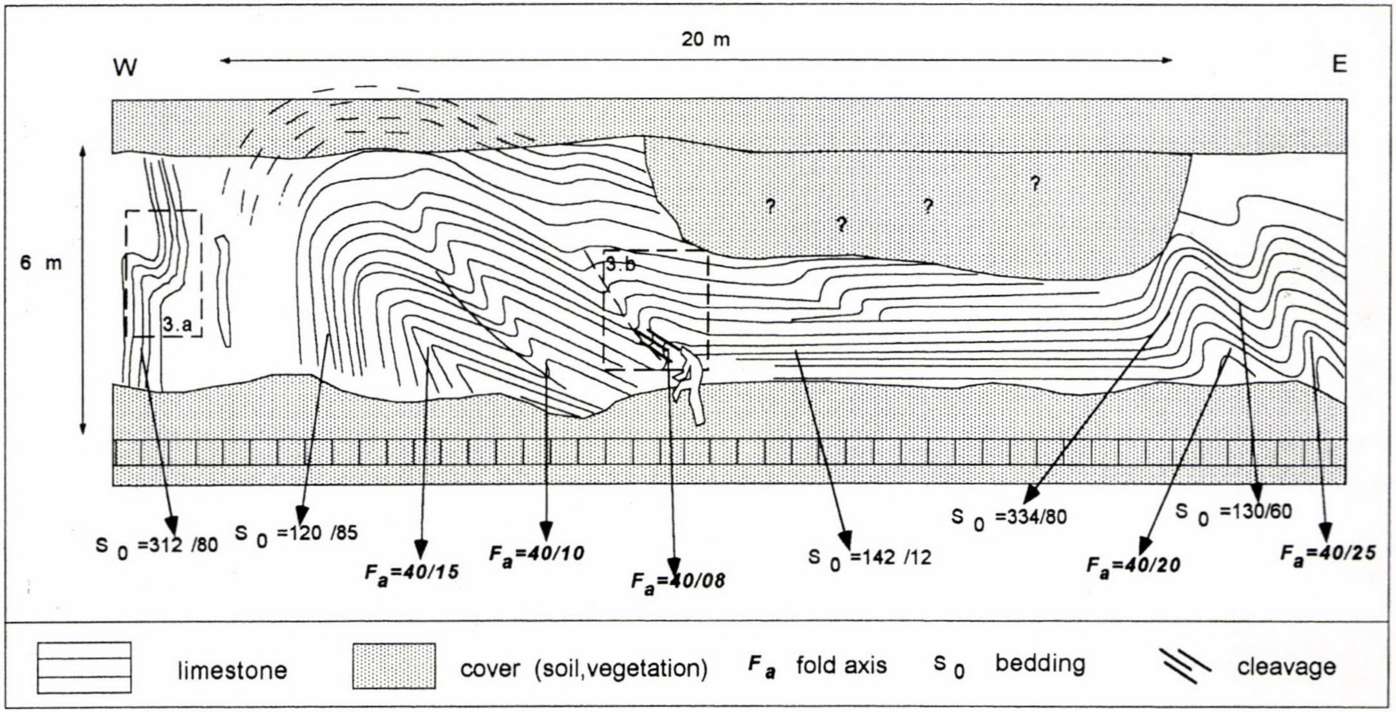

Fig. 2
Site 1. Bedding traces (S0) marked as thin black lines. Thin dashed lines parallel to the axial planes indicate nascent cleavage. Description in the text. Bedding attitudes are marked in normal, fold axes in bold characters. Inserts indicate location of Figures 3a and 3b

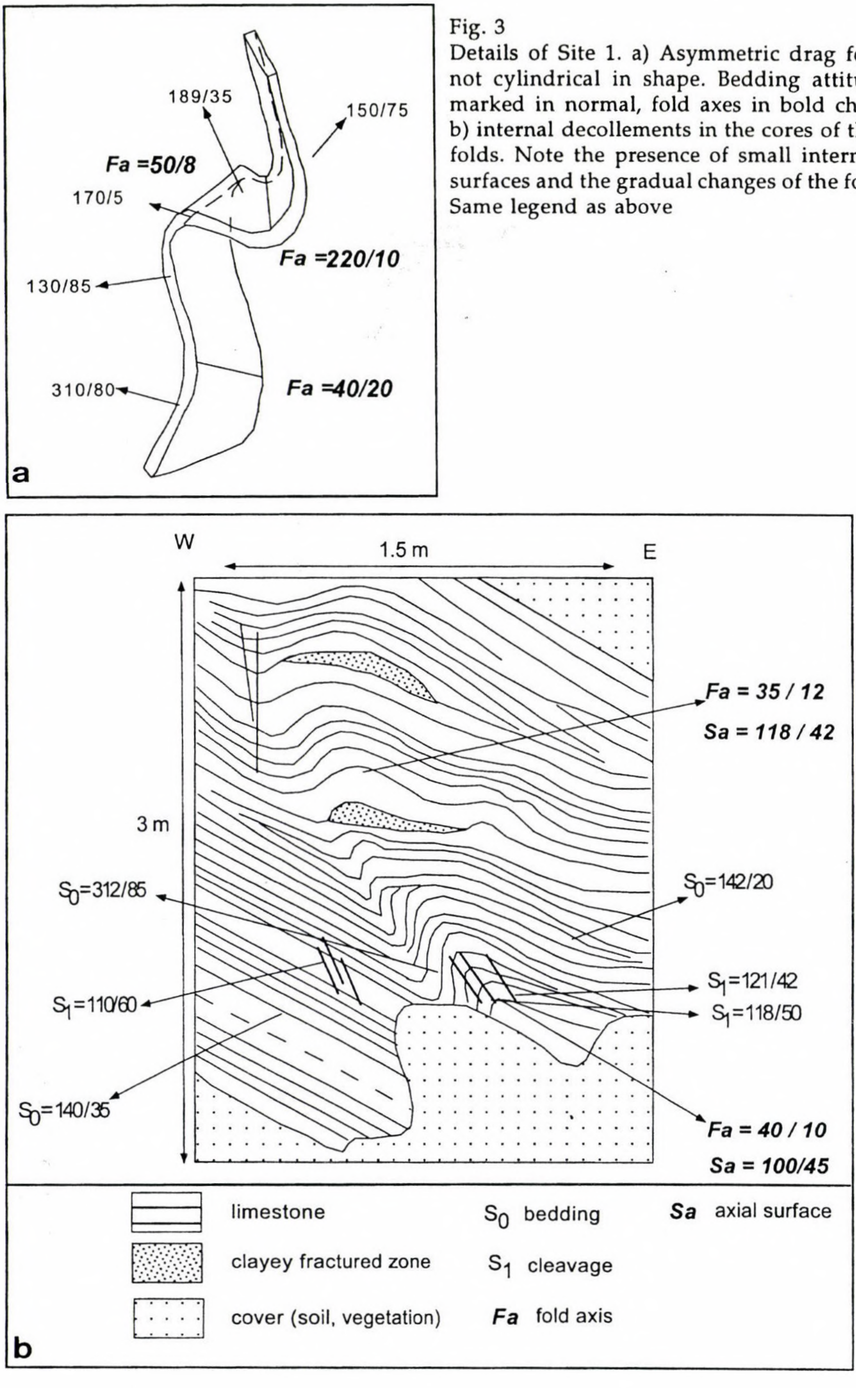

Fig. 3
Details of Site 1. a) Asymmetric drag fold. It is not cylindrical in shape. Bedding attitudes are marked in normal, fold axes in bold characters. b) internal decollements in the cores of the major folds. Note the presence of small internal shear surfaces and the gradual changes of the fold form. Same legend as above

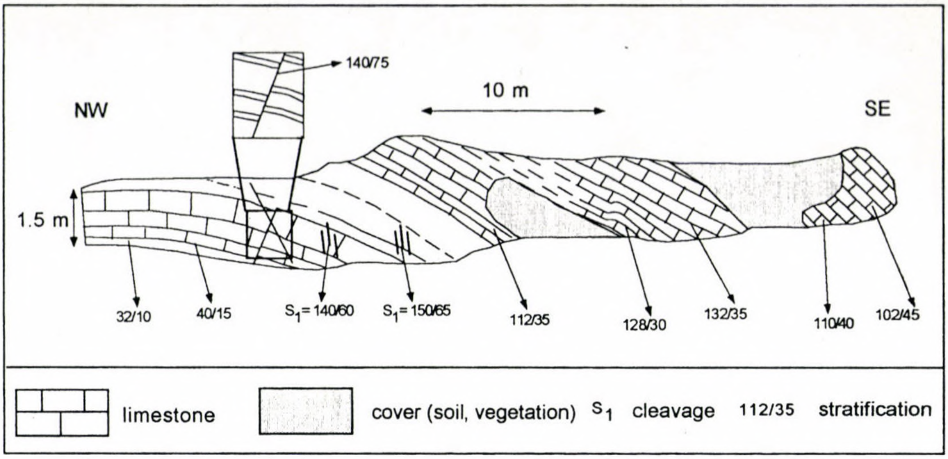

Fig. 4
Site 2. Fold limbs become steeper toward the southeast. Fold cut by late normal fault (insert)

middle part of the exposure where they become gently tilted to form a flat, open fold in the lower block. In the core of the fold several small thrust faults are observed (Fig. 7). Steeply-dipping faults produce a smaller offset, while a flat-lying SE-dipping fault cuts many layers and thrusts the upper layers on the NW limb (Figs. 5 $d_1 \dots d_n$, 6a). The open fold was amplified by this thrust. The SE limb of this gentle fold served as a ramp for the overlying block (Figs 5, 6a). Flats attached to this ramp are found in the lowest part in the SE part and at about 2 m height in the NW part of the exposure. The detachment mostly follows one single layer; very few layers are cut by it in the footwall (Fig. 6a). The hanging wall is composed of a more clay-rich, thinner-bedded formation and was thus more easily folded. Folds in this part (especially in the SE half of the outcrop) are disharmonic box-folds or parallel folds. Disharmony is relaxed at internal smaller detachments (Fig. 6b). It increases towards the core of the folds, especially above the ramp of the main detachment. This detachment seems to cut off some layers in the SE third of the outcrop, but mostly runs parallel to the layering of the hanging wall. The great structure is slightly unsymmetrical and with a NW vergency.

In the NW third of the exposure the hanging wall is backthrust along a fault rising from the upper flat of the detachment (Figs 5, 7). The backthrust branches to produce a divergent fan of thrust faults transporting material towards the SE. The uppermost slice acts as a ramp anticline above the steeper part of the thrust, which is in fact a ramp. Smaller embryonic ramps are found in the core of the hanging wall anticline. A particular layer (indicated by hatching on Fig. 7) is overthrust from both the southeast and the northwest.

In the large hanging wall block (Fig. 5) the folds are a little bit more open toward the southeast and tighter toward the ramp. The lower middle part of

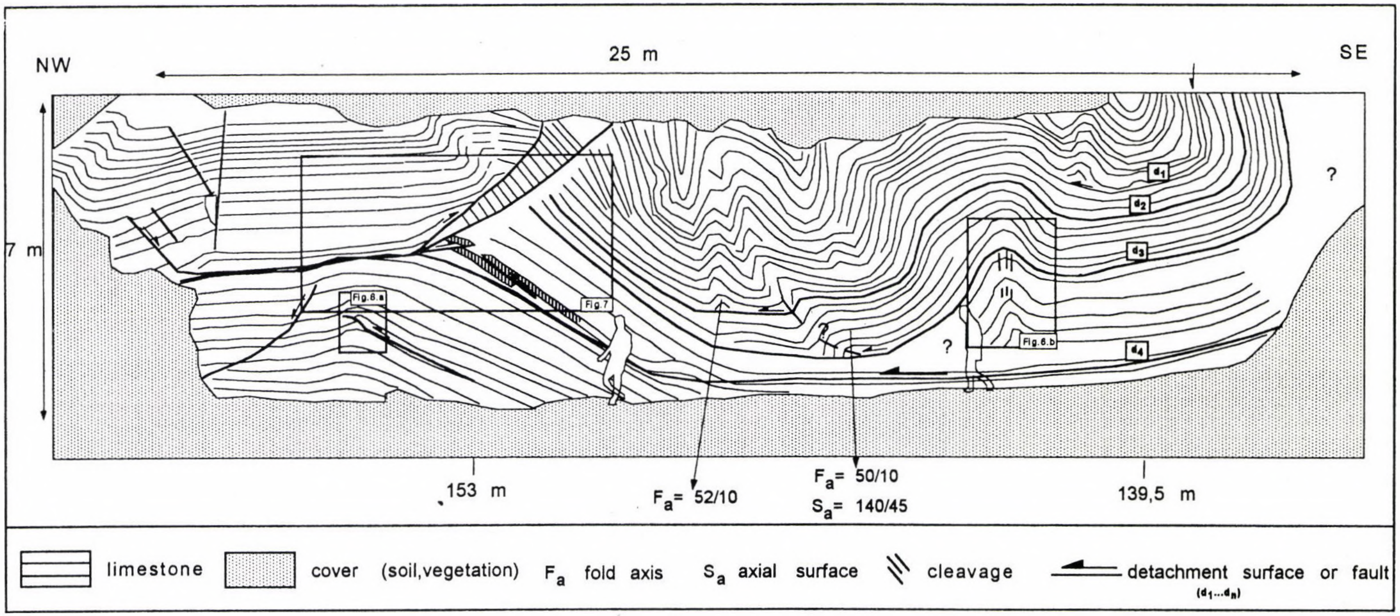

Fig. 5
Site 3 (small parking place) Bedding traces marked as thin black lines. Thin dashed lines parallel to the axial planes indicate nascent cleavage. Detachments and ramps are marked thicker, with arrow for the transport direction. Inserts indicate location of Figures 6, 7. Description in the text

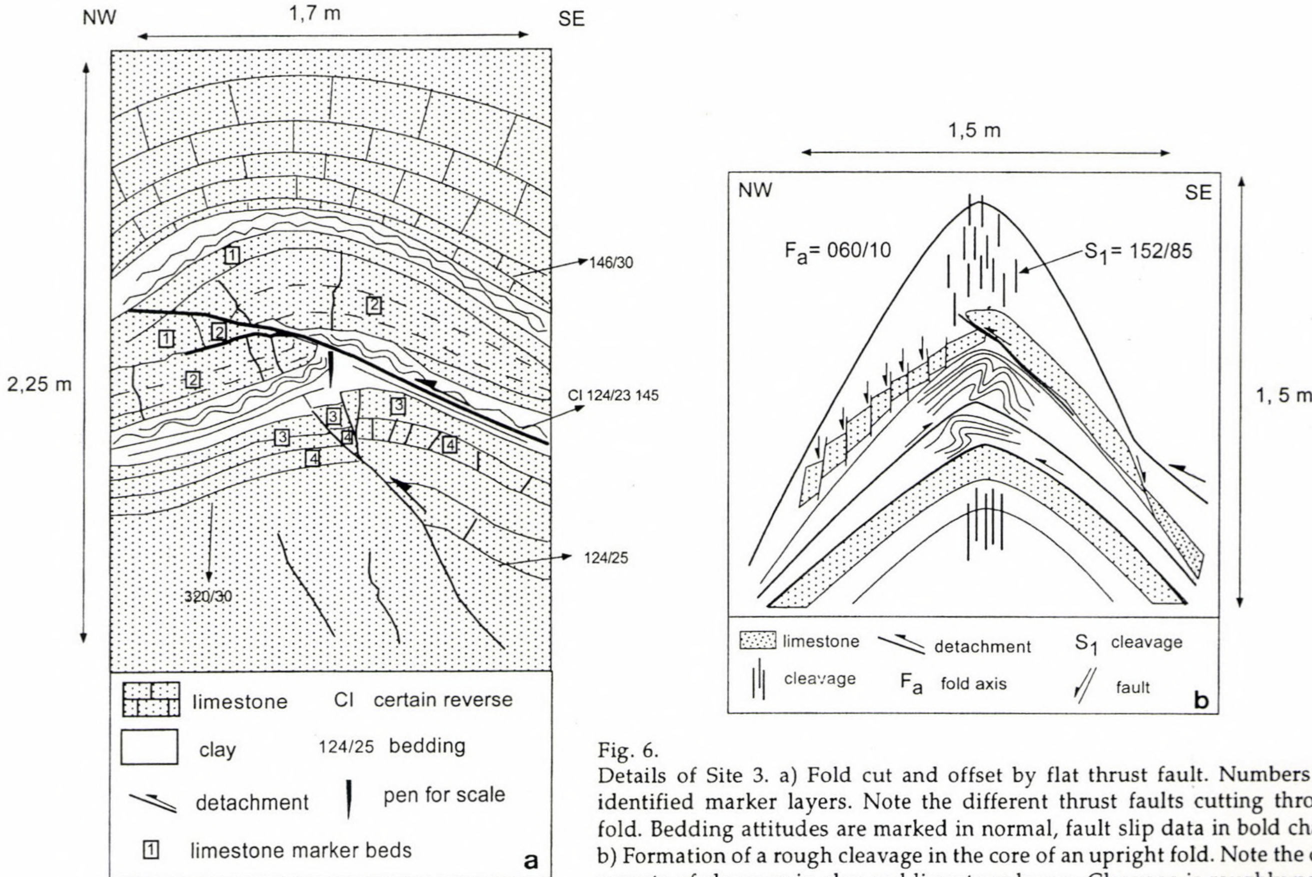

Fig. 6.
Details of Site 3. a) Fold cut and offset by flat thrust fault. Numbers refer to identified marker layers. Note the different thrust faults cutting through the fold. Bedding attitudes are marked in normal, fault slip data in bold characters. b) Formation of a rough cleavage in the core of an upright fold. Note the different aspects of cleavage in clay and limestone layers. Cleavage is roughly parallel to the axial surface of the fold

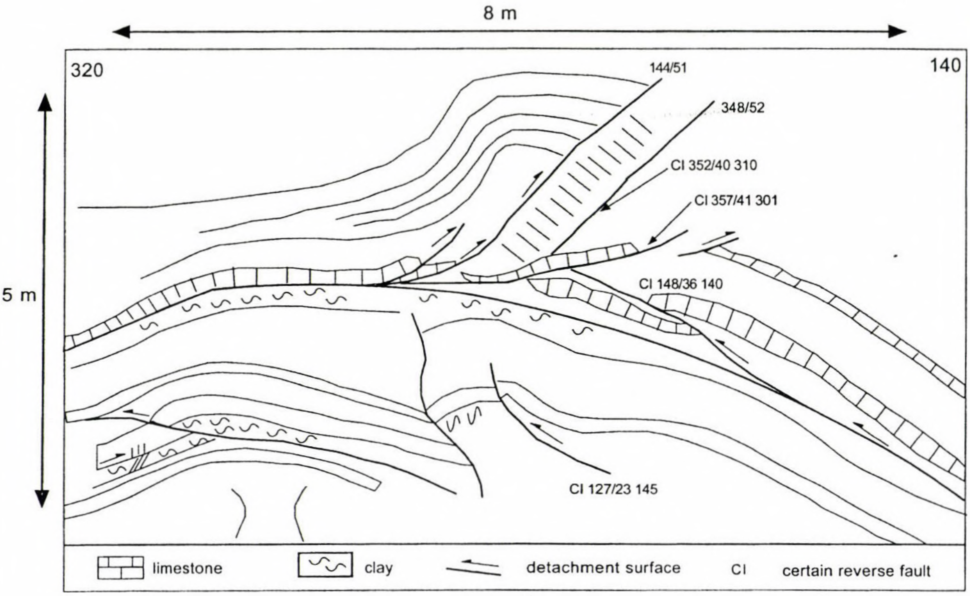

Fig. 7
Detail of Site 3. Drawing of the notebook to show a blow-up of the backthrust. Bedding attitudes are marked in normal, fault slip data in bold characters

the hanging wall is folded in very tight folds. These tight folds refold an earlier structure (Fig. 6b). Smaller ramps (detachments) are seen in this part, which have been refolded as well (see $d_1...d_n$ on Fig. 5). Later the lower detachment surface (d_4 on the Fig. 5) was refolded along an underlying detachment surface which is not exposed. Because of very frequent cutoffs and offsets of beds, the individual layers are difficult to follow in this part. A weak cleavage is observed in the core of the straight upright antiform (box 6b in Fig. 5). This is developed in the clay layers as a rough cleavage and in some thinner limestone layers as a very weak fracture cleavage (Fig. 6b). This cleavage is associated with folds with a 55–60 axial direction.

Between sites 3 and 4 the footwall block of site 3 is gently folded, then forms a steep, overturned limb.

Site 4

This is a large outcrop at km 7 of the road (Figs 1, 8). The description of the type section of the Viganvár Limestone Formation comes from this outcrop (Rálisch-Felgenhauer 1988). The outcrop exposes a major chevron fold with two plane limbs. The NW limb is subvertical and slightly overturned (Fig. 8). A thick pair of resistant limestone beds is followed throughout the outcrop.

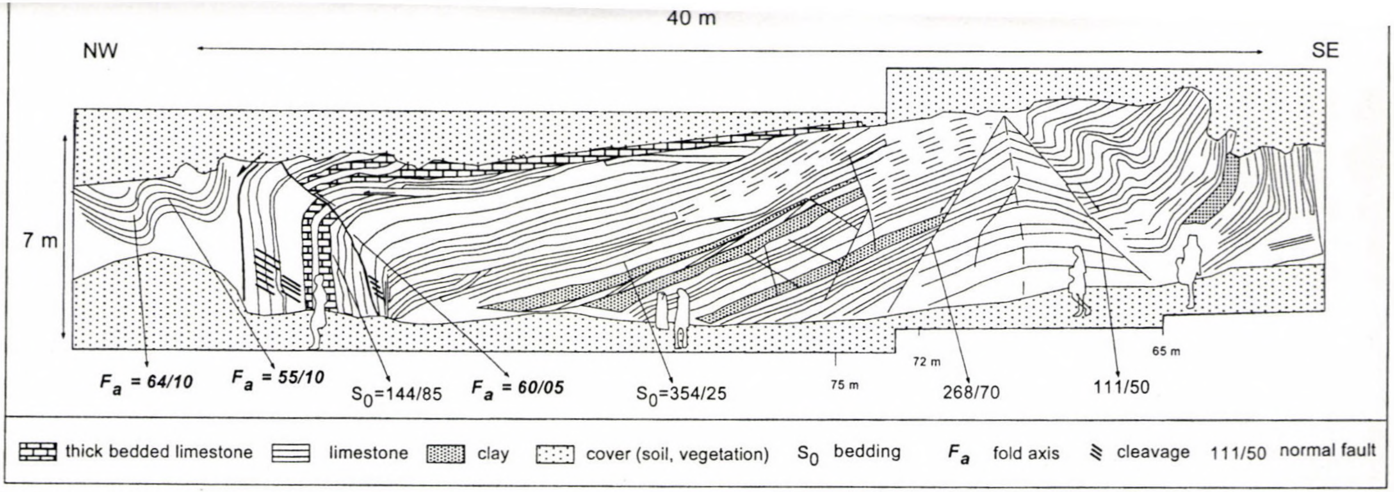

Fig. 8
Site 4. Major chevron fold with internal thrusts and late normal faults. Different rasters indicate identified beds. Bedding attitudes are marked in normal, fold axes in bold characters

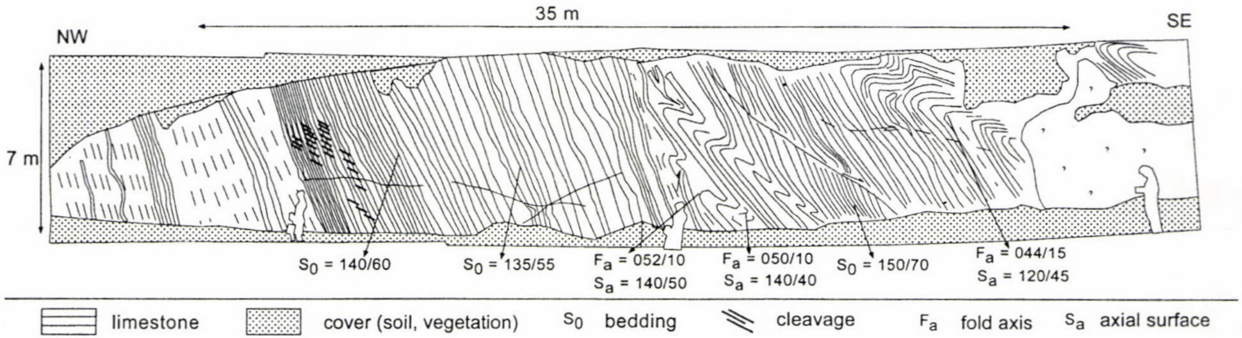

Fig. 9
Site 5. Steeply overturned limb of a fold, the hinge zone of which emerges at the SE end of the outcrop. Note asymmetric drag folds at the centre and weak, but parallel cleavage (light dashed traces) in the NW third of the outcrop. All structural features indicate inverted limb. Evaporitic layers are in subcrop at the SE end of the outcrop

These beds are offset by one metre by a thrust fault, cutting through the core of the fold. Smaller incompetent layers are dragged along the fault.

Smaller shears are also observed on the relatively flat-lying normal limb. Most of them transport material northwestward, but some are minor backthrusts. These cut through some layers rather than detach along incompetent layers. A very weak cleavage is developed on the overturned limb of the major fold. This cleavage is roughly parallel with the fold's axial plane.

On the two extremities of the outcrop smaller-wavelength, medium tight folds develop. The fold on the SE end of the outcrop shows a clear disharmony: its link with the underlying synform is most probably provided by a detachment surface.

The fold is cut by a conjugate pair of late normal faults. They produce an offset of one m toward the northwest and 2–3 m toward the southeast.

Site 5

This is the southeasternmost exposure along the road (Figs 1, 9). A very steeply (80) southeastward-dipping sequence is exposed. Two folded structures are observed: a set of asymmetric drag folds in the middle part of the exposure and the hinge of a major fold in the SE end. At this location a dolomitic-anhydritic sequence is found which might already belong to the underlying formation. The drag folds indicate shear toward the hinge, suggesting an overturned limb of a flexural slip mechanism fold.

A weak, but parallel cleavage is observed at the middle-northern part of the exposure. This cleavage cuts through clay and limestone layers. Cleavage is less steep than the bedding, which also suggests an overturned limb. This interpretation is supported by detailed sedimentological studies as well (Konrád and Török pers. comm.).

5. *Measurements*

Measurements related to folds and faults were treated separately, but in the interpretation the data from both sets was taken into account (Fig. 10).

Axes of the major folds are generally oriented 050/10. The bigger folds with overturned limbs and with cleavage (sites 4 and 5) have this attitude. Because of the presence of pressure solution cleavage (almost exclusively on more sheared overturned limbs) parallel to their axial surface these medium tight folds are thought to be the primary structures. They have been formed by flexural slip on incompetent interlayers, under the highest thermodynamic conditions; however, these did not exceed strong diagenetic conditions (between 100–200 °C – Dunkl 1992). Because of the more sheared overturned limb a fault propagation fold model (Suppe 1985) is suggested. We note, however, that faults associated with the folds were reactivated resulting in the

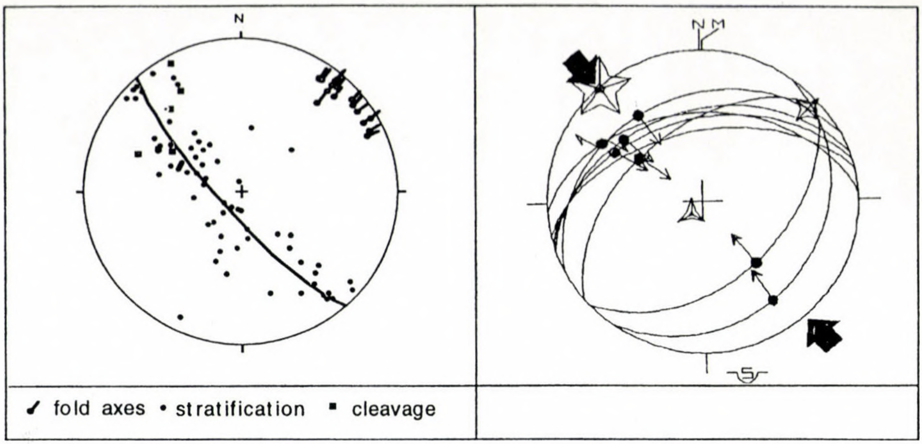

Fig. 10
Plots of measured structural data. a) data on folds: Schmidt projection, lower hemisphere. points: poles to bedding; barbed points: measured fold axes. Note the slight changes in axial directions. b) plots of the thrust fault slip data. Faults are represented by their trace on the lower hemisphere Schmidt projection. Slickenslide lineations are marked as arrows indicating movement of the hanging wall. Black arrows show the computed principal stress directions (Tector program of Angelier 1976). Note the similarity of shortening directions with respect to folding. Same legend as for a)

cutting of the earlier-formed folds. Nevertheless these two deformations were part of the same phase.

Some other folds are more of a box-fold or disharmonic fold character (sites 2 and 3). They are associated with detachments and ramps. These often cut earlier folds (Figs 5, 6, 8). Some folds are directly linked to the activity of ramps (Fig. 7). All these folds have an axial direction of 040/20. Especially based on the squeezed hanging wall of site 3 these folds and faults are linked to a progression of shortening of earlier folds, and were formed due to local space problems on flanks of the folds.

Fault slip data show that most thrust faults transport toward the northwest (Fig. 10 b). The calculated stress-field shows $\sigma 1$ directions oriented NW–SE. Although these structures have a different style than the primary ones, they are interpreted as belonging to the same, possibly Late Cretaceous shortening phase. The direction of the structures is roughly the same and the thermodynamic conditions do not differ very much from those of the primary phase.

All these structures were later cut and offset by normal faults. The age of this deformation is not known.

We attempted the construction of an integrated structural section (Fig. 11). This is characterised by wavy folding of the Viganvár formation in asymmetric medium tight folds. The core of the structure is found most probably in the

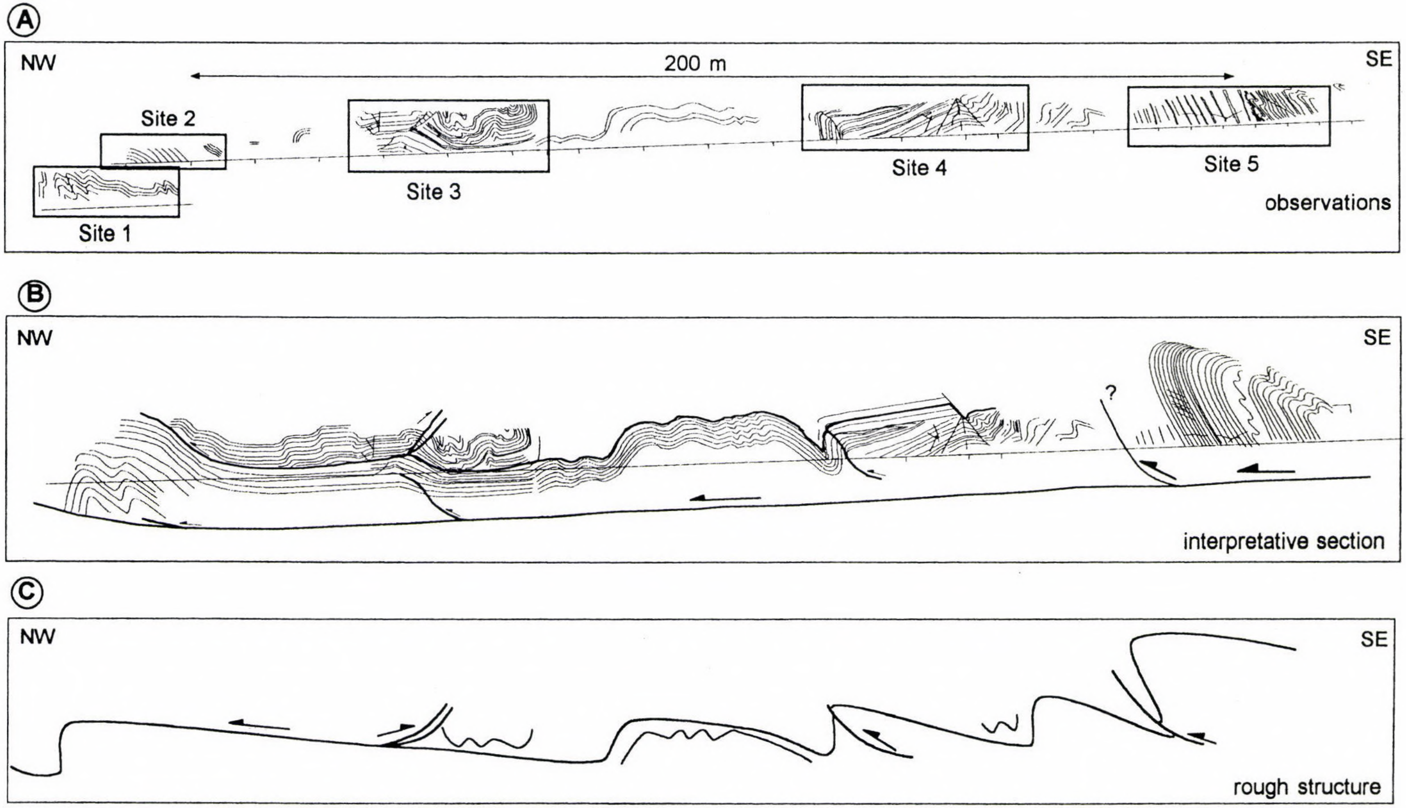

Fig. 11
Constructed section taking all data into consideration. A) raw data, lines indicate bedding; B) interpretative section. Lower detachment and ramps indicated as bold lines, arrows indicate transport direction; C) rough structure of the road cuts at Abaliget

SE end of the section, where the oldest layers are exposed. The section shows a fold train, the wavelength of which gradually increases toward the northwest. All members of this train are characterised by short, steeply inverted limbs and long, flat-lying normal limbs. Especially these flat normal limbs are cut by ramps and detachment surfaces, along which material is shortened again. Most of these ramps transport toward the northwest, but occasional backthrusts also occur. These detachments (ramps) limit blocks with more intensive local folding. Because of the presence and vicinity of an evaporitic horizon, we suppose that both folds and ramps detach along this stratigraphic horizon. Increased wavelength toward the southeast might be explained by more intensive shearing or an original propagating thrust zone out of the evaporitic horizon.

6. *Conclusions*

A detailed structural description can contribute to a better stratigraphic description and correlation of the Viganvár Formation. Folds and related faults of different styles were described. These different styles of folds with slightly different attitudes (55 and 40 respectively) were formed, however, during the same continuous deformation phase, probably during the Late Cretaceous. The different style of the folds is partly due to the decrease of shortening from southeast to northwest, which might be explained by a decrease of shear in the underlying evaporitic detachment horizon or by the northwestward propagation of a shear zone rising from this detachment. There are internal detachments within the Viganvár formation which enable local folds and ramp anticlines to develop during a later stage deformation (Late Cretaceous). All faults were consistent in that they indicated a NW–SE shortening and mostly northwestward transport. These ductile and ductile-brittle deformations were followed by later normal faulting.

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Dolomitization and karst-related dedolomitization of Muschelkalk carbonates, in South Hungary

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Middle Triassic carbonates of the Villány Mountains were deposited on a homoclinal carbonate ramp system. Within the 700 m-thick carbonate ramp sequence many of the carbonates show partial or complete dolomitization. The study describes the major petrological and geochemical constraints of the dolomitization process of the Zuhánya Limestone Fm. The dolomites occur in the limestone as small, irregular mottles with a maximum diameter of 20 cm. The characteristic textural types are planar-e dolomites, planar-s dolomites and nonplanar xenotopic dolomites.

The initial phase of dolomitization is characterised by the formation of scattered dull maroon to non-luminescent dolomite rhombs, in the later-phase xenotopic dolomites and stylolite/ fracture-related saddle dolomites were formed. These dolomites differ significantly from earlier ones both in CL luminescence and in geochemistry. The calcite cementation appears in several distinct phases. The first phase, the replacement of aragonite and high-Mg calcite by low-Mg calcite precedes the dolomitization. Later-phase karst-related calcite cementation is characterised by multiple generations of calcite fissures with generally no to dull luminescence having a highly negative stabile isotope composition.

The first phase of dolomitization is proposed to occur in a shallow subsurface setting, and was enhanced by further overburden resulting in the formation of xenotopic and saddle dolomites. Later-phase karst-related processes caused the partial dedolomitization of former dolomites.

The complexity of dolomite cementation, calcite cementation and dedolomitization is related to the complex geohistory of the region, to changes of fluid geochemistry and migration. Fissures caused by tectonic activity could have served as pathways for migrating fluids.

Key words: carbonate diagenesis, dolomitization, Middle Triassic, ramp, stable isotope, cathodoluminescence, South Hungary

Introduction

Dolomitization processes are very common in the geologic record although the origin of many ancient dolomites are still not well understood. Several existing models for dolomitization include evaporation of marine waters, Coroong-type evaporation, mixing zone, burial and sea-water related (e.g. Kohut convection); (for a summary see Tucker and Wright 1990 and Purser et al. 1994). The dolomitization of thick carbonate ramp sequences has received less attention than that of the carbonate platforms although several examples were studied in detail (Shukla 1988; Calvet et al. 1990; Burchette and Wright 1992; Kruger and Simo 1994).

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The Middle Triassic carbonates of southern Hungary in the Villány Mountains form a very thick (more than 700 m) carbonate ramp complex of which the greater part is dolomitized. The dolomitization shows significant variations both in scale and grade but in general the entire succession contains partly or entirely dolomitized rocks.

The principal aim of this paper is to explain the patchy appearance of dolomite in a limestone unit and to present a model which describes the petrology, geochemical composition and origin of these types of dolomites. In addition it is intended to describe one of the Muschelkalk dolomites of southern Hungary.

The difficulties in finding any descriptive models arise from the facts that the region is very intensely tectonized, that the partly dolomitized rock bodies are located in overthrusted zones, and that the surface occurrences are relatively small. In addition paleokarst phenomena and some sub-recent karstification, partly of thermal origin, also interfere.

Regional setting

The Villány Mts are located in South Hungary forming the western termination of a SW–NE trending tectonic unit, the Villány–Bihor zone, extending northeastward in the basement of the Great Hungarian Plain. The easternmost outcrops of the unit are in the Bihor Mts (Apuseni Mts in Romania – Bleahu et al. 1994). Similar but much less dolomitized Middle Triassic "Germano-type" ramp sequences are found in the Mecsek Mountains north of the study area (Fig. 1). Those sediments show similarities to Polish, German and Spanish Muschelkalk carbonates (Török 1993, 1997).

The Villány–Bihor unit belongs to the Tisia megaunit (terrain) which is interpreted as a part of the northern margin of the Triassic Tethys. Its present geographical location is a result of large-scale displacements of the terrains from the late Mesozoic to the early Tertiary (Kázmér and Kovács 1985; Csontos et al. 1992).

The Villány Mountains consist of 6 northward thrusted structural units and one southward-thrusted one forming imbrications, which are generally parallel to the ENE–WSW orientation of major strike slips (Fig. 2). The northward thrusting may have taken place in the Late Cretaceous (Wein 1967) or in the Neogene (Nagy E. and Nagy I. 1976; Némedi Varga 1983). At least three additional tectonic phases were documented by microtectonic analyses from the Miocene onward (Bergerat and Csontos 1988). The true nappe structures have not been proved yet.

Stratigraphic setting

In the Villány Mountains and in the surrounding subsurface zones Mesozoic rocks are underlain by thick Permian, dominantly clastic sediments (Kassai 1976). The Lower Triassic is represented by red sandstones and conglomerates of Buntsandstein type. The overlying nearly 100 m-thick fine clastics, red to

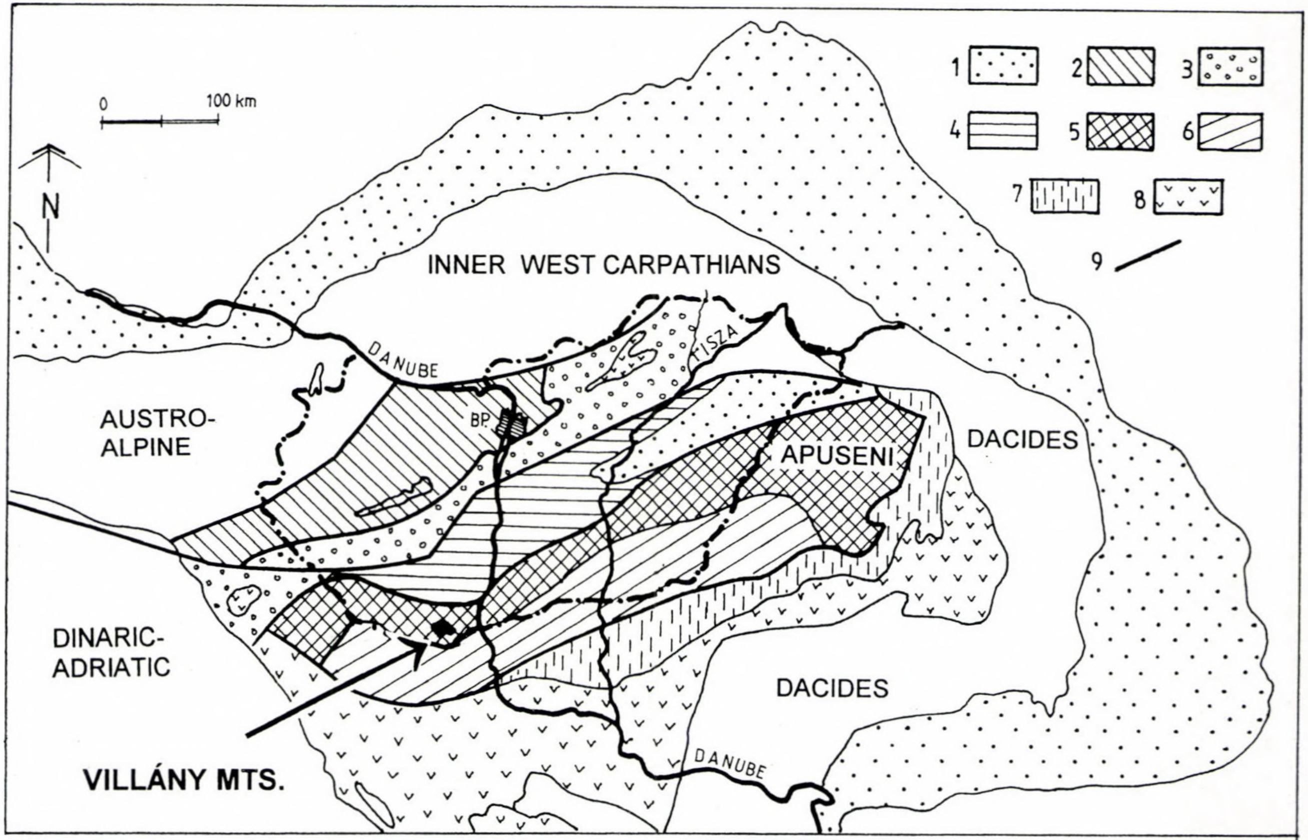


Fig. 1
The location of Villány Mountains in the Mesozoic tectonofacial units of Hungary. 1. Foredeep and flysch units; 2. Transdanubian and Drauzug units; 3. Bükk and Inner Dinaric unit; 4. Mecsek unit; 5. Villány–Bihor unit; 6. Papuk–Lower Codru unit; 7. N. Backa–Upper Codru, Persani unit; 8. Oceanic nappes (Vardar, Meliata, Mures and Olt); 9. Boundaries of tectonofacial units (redrawn after Csontos et al. 1992)

greenish siltstones, sabkha evaporites and lagoonal dolomites are also only known from the subsurface. The oldest surface occurrences, the Anisian dolomites, belong to the lower part of the nearly 700 m-thick Triassic carbonate sequence (Nagy E. and Nagy I. 1976). These intertidal to shallow subtidal dolomites are overlain by inner ramp limestones and dolomitized limestones of the Wellenkalk unit. Higher-grade dolomitization and a lower frequency of the characteristic sedimentary structures (slumps or storm generated coquinas) highlight differences with the development of the Wellenkalk of Mecsek Mts.

In the upward-deepening succession the inner to mid-ramp, partly dolomitized carbonates are overlain by thick-bedded to thin-bedded, brownish, recrystallized limestones containing characteristic dolomite mottles (Zuhánya Limestone Fm.). The present paper focuses on this unit. The mottled dolomitic limestone gradually passes upward to a porous, micritic to saccaroidal dolomite which has the highest carbonate content of the entire succession. The overlying marly dolomite represents an intertidal to supratidal depositional evironment. With a gradually increasing clay content a thin "Keuper"-type sequence with some sandstones, variegated clays and cellular dolomites forms the upper part of the Triassic (Fig. 3).

The Jurassic and Creataceous sequences are incomplete. Following a long regional hiatus the oldest Jurassic rocks rest unconformably on the Triassic surface. The lower part of the Jurassic (Pliensbachian sandstones and limestones) represent a regional transgression. The Middle Jurassic is also very condensed and represented by a thin Bathonian limestone layer (8 cm), Callovian iron oolites, LLH stromatolites and oncidal beds (40 cm). The latter one (Villány Limestone Fm.) contains a well-known famous ammonite fauna (Lóczy 1915). The Oxfordian–Kimmeridgean is characterized by proto-globigerinid limestone and by "pelagic ooids" showing an upward-shallowing trend. A regional unconformity is indicated by the presence of Early Cretaceous (Berriasian?) karst bauxite. The overlying dark grey limestones with Characeas, and further upsection the Lofer cyclical peritidal-subtidal carbonates with orbitolinids and pachyodonts (Császár 1989) indicate a deepening-upward trend and the re-establishment of marine sedimentation.

Depositional environment

The entire Middle Triassic carbonate unit was deposited on a homoclinal ramp similarly to that in the Mecsek Mountains (Török 1993, 1997), Polish (Szulc 1993), German (Aigner 1985) and Spanish Muschelkalk (Calvet and Tucker 1988; Calvet et al. 1990). Meanwhile some of the the primary sedimentary structures were partly destroyed by dolomitization.

In the general evolution of the basin a long-term cyclic trend could be observed. It begins with depening and followed by shallowing. The characteristic microfacies types are mudstones, bivalve ("filament") wacke-stones, peloidal mudstones/wackestones and brachiopod floatstones. The

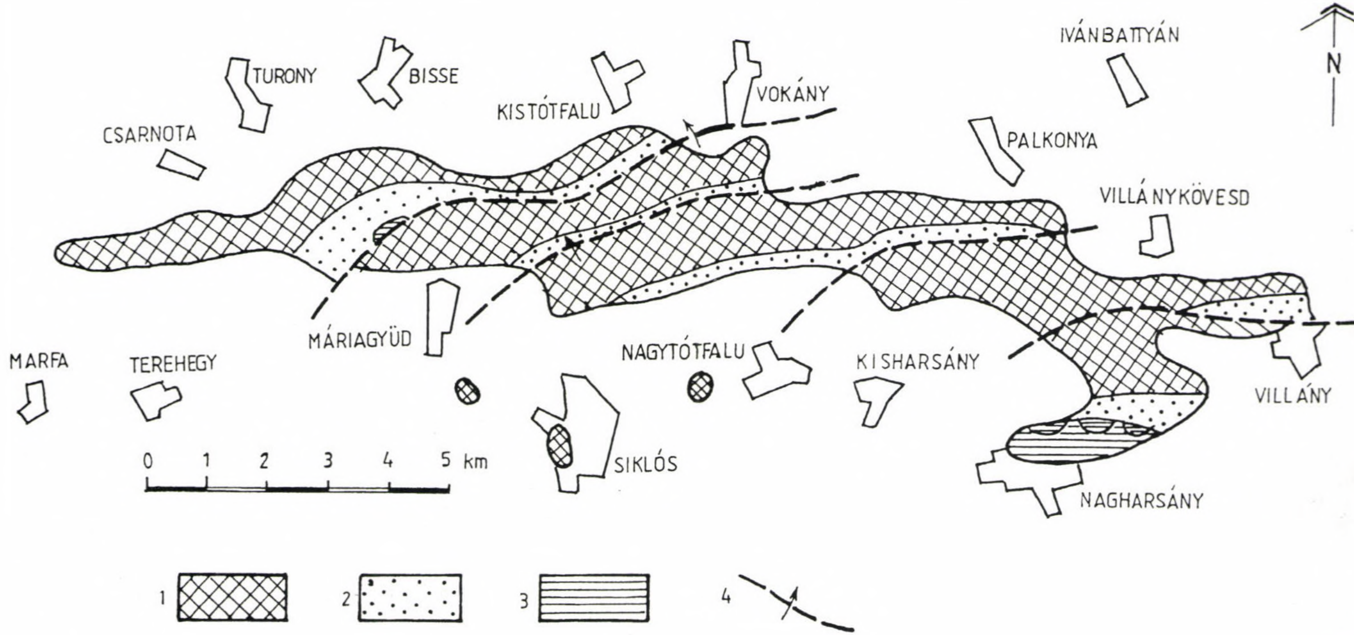
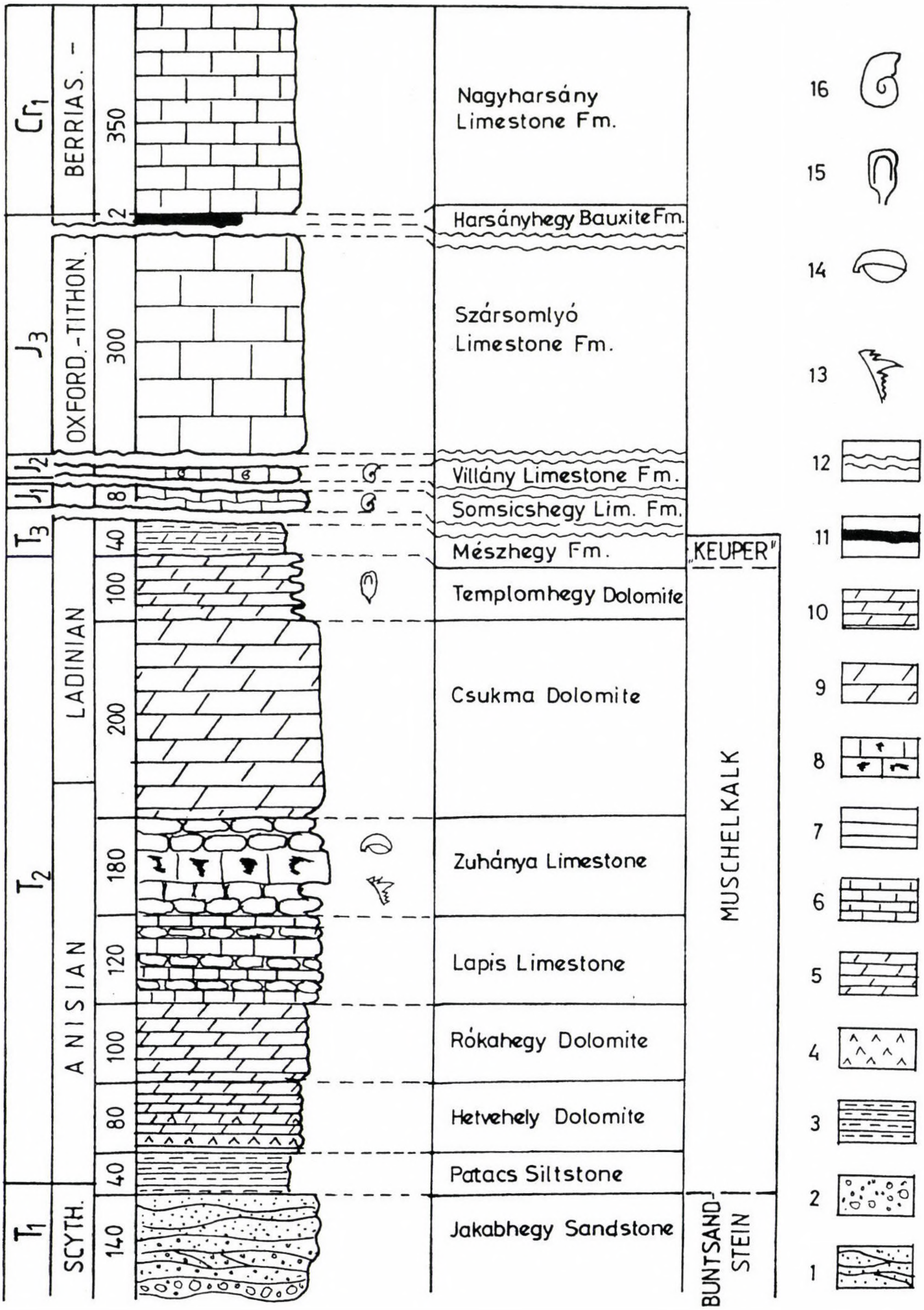


Fig. 2
The simplified geological map of Villány Mountains showing the distribution of the Mesozoic rocks. 1. occurrence of Middle Triassic carbonates (subordinatelly Upper Triassic); 2. Upper Jurassic carbonates (subordinately Middle Jurassic); 3. Lower Cretaceous bauxite and limestones; 4. imbrications, which divide the mountain into "nappes"


Cr1
BERRIAS. -
350
2
J3
OXFORD.-TITHON.
300
J2
J1
8
T3
40
LADINIAN
100
200
T2
ANISIAN
180
120
100
80
40
T1
SCYTH.
140
Nagyharsány Limestone Fm.
Harsányhegy Bauxite Fm.
Szársomlyó Limestone Fm.
Villány Limestone Fm.
Somsicshegy Lim. Fm.
Mészhegy Fm.
Templomhegy Dolomite
Csukma Dolomite
Zuhánya Limestone
Lapis Limestone
Rókahegy Dolomite
Hetvehely Dolomite
Patacs Siltstone
Jakabhegy Sandstone
„KEUPER"
MUSCHELKALK
BUNTSAND-STEIN
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intercalation of shell beds are related to storms and are considered to be tempestite coquina beds. These sediments and the underlying nodular carbonates, laminated limestones and dolomites suggests a carbonate mud dominated, periodically storm influenced homoclinal ramp setting, similarly to that in the Mecsek Mts (Török 1989). No any evidence of shelf edge rimming buildups were found. The presence of conodonts (Bóna 1976) indicates a relatively deep, outer ramp setting.

Dolomite petrography and texture, calcite cementation

Dolomite occurs in several different petrographical and textural forms in the Middle Triassic succession. The following description concentrates mainly on the mottled dolomitic limestone of the Zuhánya Limestone Formation with only minor hints of the other dolomites.

The underlying dolomitic limestones of the Lapis Limestone Fm. (Gyűd Limestone) are characterized mainly by micritic limestones and dolomicrites, less frequently by dolosparites. The dolomitization process was incomplete since most of the carbonates are either limestones or dolomitic limestones (Nagy E. and Nagy I. 1976).

The "mottled" thick-bedded dolomitic limestone is only a part of Zuhánya Limestone Fm. The formation also comprises nodular limestones and thin-bedded limestones (Rálisch-Felgenhauer and Török 1993). The dolomitic limestones show varied petrography. Macroscopically the dolomite occurs as irregular patches or mottles in the limestone. The mottles are variable in size reaching up to 20 cm in diameter. The mottled appearance is related to brecciation (Konrád 1990), i.e. there are cemented "intraclasts" of different sizes in the brownish to greenish-grey limestone. The intraclasts are locally incorporated in calcite or dolomitic calcite matrix. In this case the matrix is slightly yellowish to brownish. In addition to the dolomitic rock-types mentioned above dolomites also occur in thinner fissures and stylolites. Most of the dolomite "mottles" are apparently related to fissures and stylolites. From these fluid pathways the dolomitizing fluids "spread out". Despite these textural differences the rocks are strongly cemented and compacted, and therefore are commonly used as ornamental stones. The quarried thicker beds (up to 13 m!) are often called "green marble" because of their greenish colour.

Dolomites appear in several textural forms ranging from idiotopic mosaics (planar-e, planar-s) to xenotopic mosaics (nonplanar). The most characteristic

← Fig. 3

Lithologic and lithostratigraphic sequence of Mesozoic formations in the Villány Mountains. The studied rocks belong to the Middle Triassic, Zuhánya Limestone Formation. 1. sandstone; 2. conglomerate; 3. laminated siltstone; 4. evaporites; 5. thin bedded dolomite; 6. thin bedded limestone; 7. nodular limestone; 8. thick bedded limestone with dolomite mottles; 9. thick bedded dolomite; 10. dolomitic marl; 11. bauxite; 12. gap (unconformity); 13. conodonts; 14. brachiopods (mainly Terebratulids); 15. *Lingula*; 16. ammonites dolomite textures in order of decreasing frequency are planar-e (planar-euhedral), planar-s (planar-subhedral), nonplanar and saddle dolomites (Török 1989).

Planar-e dolomites

The texture is characterized by scattered euhedral dolomite rhombs which are floating in micrite. The rhombs are 10 to 200 micrometers in size. The most typical feature of these rhombs is the zonation. The inner part of the rhombs are often cloudy as a result of calcite inclusions. The outer part of the crystals is clear and limpid, showing a zonal pattern (Fig. 4). The CL colour also reflects this zonation. The inclusion rich core has a very dull maroon luminescence. The core is overgrown by a bright orange luminescent Ca-rich zone often composed of a thin calcite layer. The outermost, thicker, clear rim is non-luminescent black. Many of the scattered dolomite rhombs show signs of dedolomitization. The locally resorbed crystal faces (Fig. 5) and changes in luminescence pattern (occurrence of brighter zones) are indicative of this process (see below).

Planar-s dolomites

This type of dolomite is supposedly related to the planar-e type. It is believed that with the crystal growing and with the coalescence of scattered rhombs planar-s texture could develop. The product is visible as a bright orange-brown overgrowth on former dolomite rhombs (Fig. 6). Planar-s dolomites are composed of dolomite rhombs of 150–400 micrometers in size. The crystals are inclusion-rich. A very characteristic feature is that on the surface of the dolomite crystals and in some parts of the intercrystalline space a brownish, opaque, ferruginous substance is visible (Fig. 7). It is amorphous iron-hydroxyde, organic matter and iron sulfide which supposedly inhibits the growth of dolomite crystals, similarly to the Ellenburger Dolomites (Lee and Friedman 1987). The presence of this material is indicative of a higher-temperature (over 100 °C) dolomitization (Gregg and Sibley 1984). The planar-s dolomites often contain non-dolomitized fossils. Echinoderms, bivalves and brachipods are common (Fig. 8).

Planar-e and planar-s dolomites make up the bulk of the dolomitized mottles. Besides these types a minor amount of other types of dolomites are also present which play an important role in detecting diagenetic phases.

Nonplanar (xenotopic) dolomites

These dolomites occur only in the central part of dolomitized mottles or more frequently are associated with dolomitized fissures. The crystals are in the order of 100 micrometers in size and have a higher iron content.

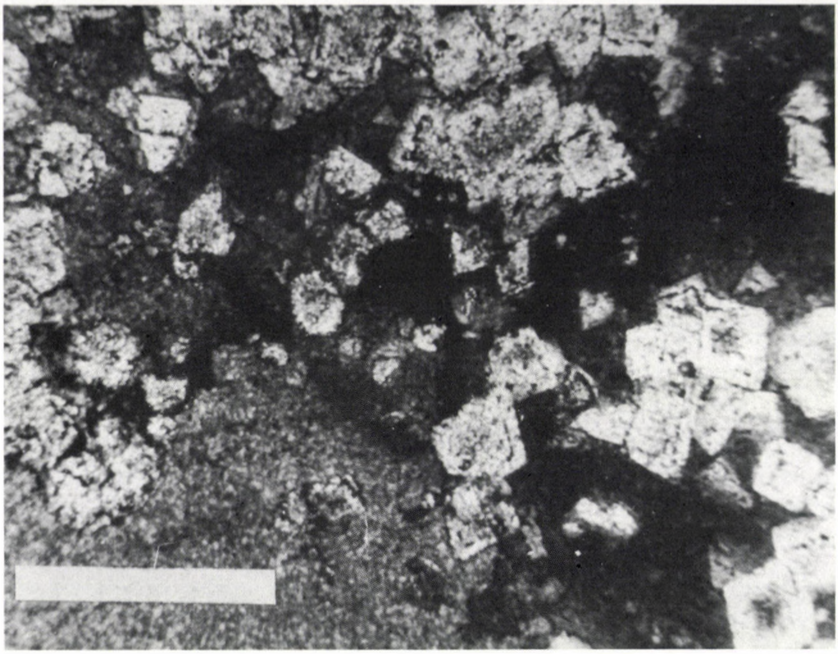


Fig. 4
Scattered dolomite rhombs in micrite (planar-e). The rhombs show a zonal pattern having cloudy, inclusion rich cores and limpid outer zones. (Stained thin section photograph, sample: Zuhánya quarry, bed no. 9, scale bar is 0.5 mm)

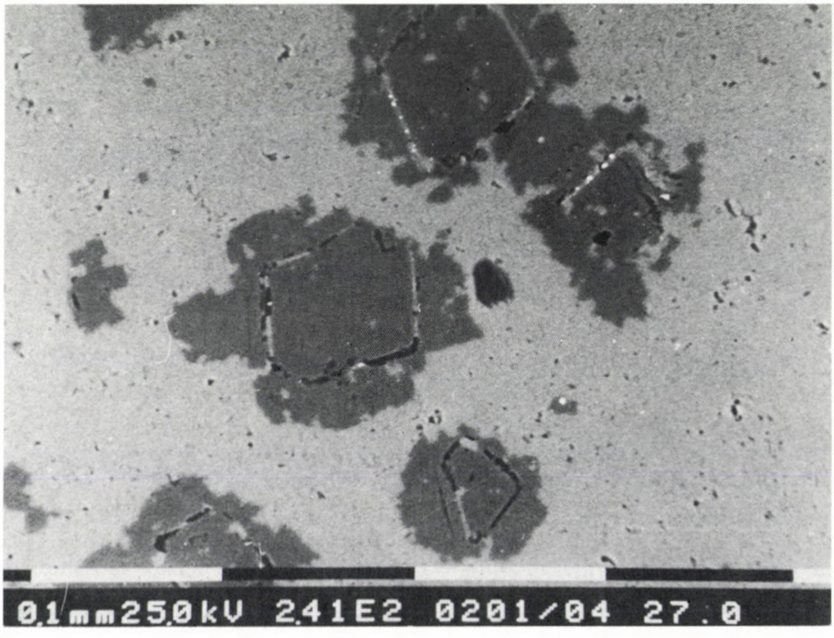


Fig. 5
Scattered dolomite rhombs (dark grey) in micrite (light grey) having resorbed, dissolved outlines owing to dedolomitization. Note the thin light grey calcite zone within the dolomite rhombs and the small light grey calcite inclusions in the core of the rhombs. (Backscattered scanning electron microscope image, sample: borehole S-VIII, 27.0 m, scale bar has 0.1 mm divisions)

Late dolomite cements, saddle dolomite

One of the late dolomite cement types is characterized by small rhombohedrons with undulating extinction of the size of 20–50 micrometers. This type is often related to stylolites showing a relatively higher iron content. The other and more characteristic type is the pore-rimming saddle or baroque dolomite with curved crystal faces and undulating extinction. This phase of cementation predates burial calcite fissure fillings since the remaining pore space is always filled by calcite (Fig. 9). These dolomites are non-luminescent.

Dedolomitization

The dedolomitization process appears in many of the above-listed dolomite texture types. At least two phases of dedolomitization were recognised. The first one, which seems to be less important, occurs in the scattered euhedral dolomite rhombs. Some of the rhombs have resorbed serrated outlines clearly indicating the dedolomitization process (see Fig. 5). Under CL microscope only a thin bright yellowish-orange discontinous zone is visible on the surface of

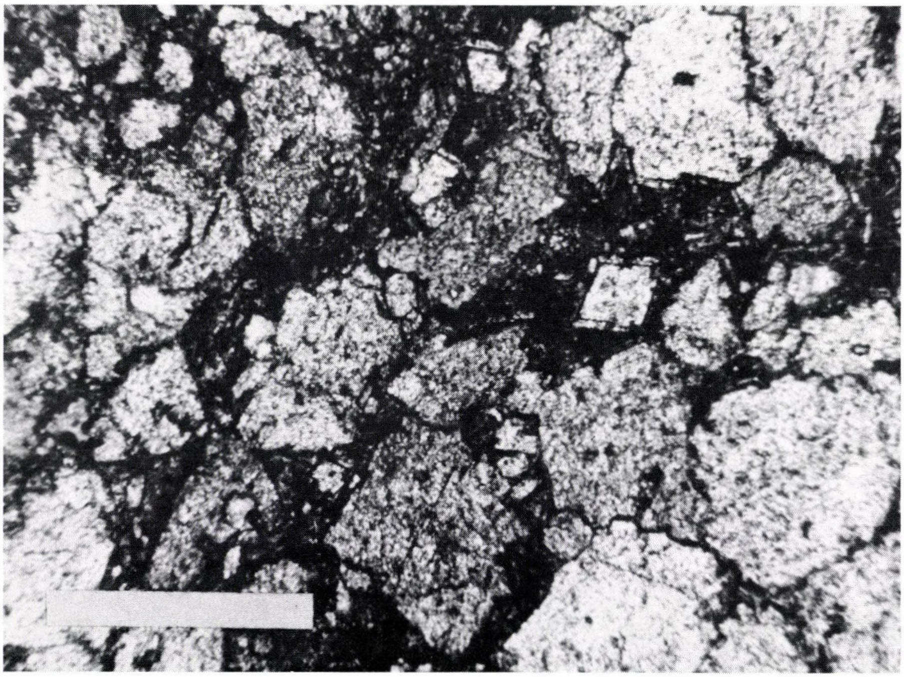


Fig. 6
Cathodoluminescence image of planar-s dolomites. The dolomite rhombs show a distinct CL zonation: core is dull "spotty" maroon, thin bright orange calcite (light zone) and non-luminescent (black) outer limpid zone. The overgrowth on dolomite rhombs have browh orange luminescence. Note the presence of bright yellow (white) irregular calcite patches in between and within the dolomite rhombs, causing dedolomitization. (CL photograph, sample: Zuhanya quarry Z1/B, scale bar is 0.2 mm)

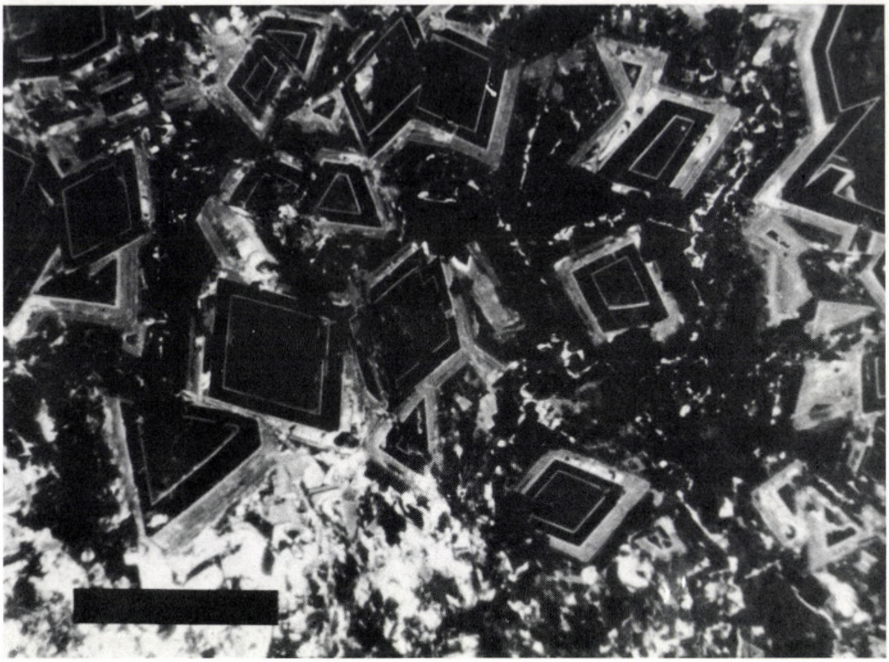


Fig. 7
Coalescence of dolomite rhombs, formation of subhedral to anhedral crystals (planar-s to nonplanar texture). Note the inclusions within dolomite crystals and the presence of dark amorphous and opaque material between the rhombs. In the lower left corner the crystals have stylolitic contacts. (Stained thin section photograph, sample: borehole S-VIII, 60.3 m, scale bar is 0.5 mm)

the corroded rhombs, or an irregular patchy zone can be observed within the dolomite crystals. This zone is not identical with the inner continuous and usually isopachous orange calcite zone of dolomite rhombs.

The second phase of dedolomitization is related to calcite cementation. This bright yellow, zoned calcite cement is responsible for the dedolomitization of fissure and stylolite-related dolomite cements (see Fig. 6). In many cases the internal zones of dolomite rhombs are replaced with calcite (Fig. 10). Besides the dedolomitization this diagenetic phase is also characterized by calcite cementation. The complex zoned pore rimming cements (crystal groups) and the calcite-dolomite alternations are mainly related to this process.

Calcite cements

Because of the predominance of the fine mud-supported texture types and the subsequent recrystallization the early marine calcite cements are not visible. The earliest recognisable diagenetic process is the replacement of the aragonite and high-Mg calcite of the bioclasts and the micrite matrix by low-Mg calcite. A clear indication of this process is the presence of small dolomite inclusions in the calcite bioclasts. The dolomitization postdates the replacement since the calcitic bioclasts remained undolomitized (see Fig. 8).

In later phases calcite fissure fillings were formed mainly in the post-lithificational diagenetic stage. The earliest fissure-filling phase is an extensional-type white one, which is non-luminescent with some dull tips. The next calcite occurs as pore-filling cement and often postdates the dolomite cements of the stylolites and fissures (see Fig. 9). Under CL it is also non-luminescent.

Significantly different calcites are the karst related fissure fillings. One type preferentially forms seams rather than regular fissures, having a zoned, bright yellow luminescence (dedolomitization). In the final phase transparent calcites were formed which occur as fissure filling and cross-cut earlier fissures. The fissure fillings show differences both in isotope and minor element composition

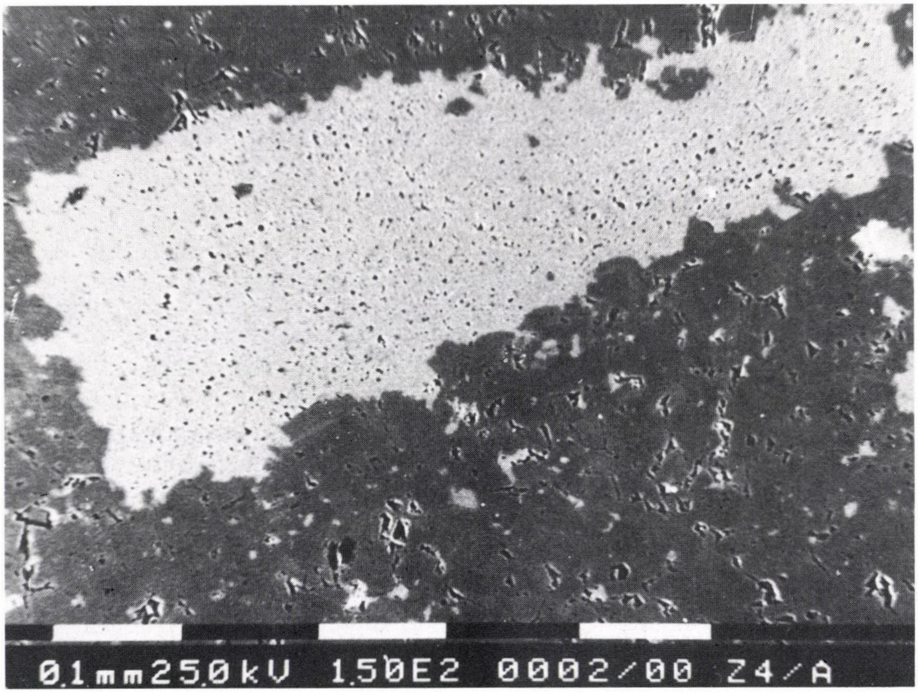


Fig. 8
Undolomitized (light grey) echinoderm fragment in dolomite (dark grey). The small dark grey spots within the echinoderm represent dolomite inclusions and are indicatives of early phase (pre-dolomitizational) high Mg calcite low Mg calcite transformation. (Backscattered scanning electron microscope image, sample: Zuhánya quarry, Z4/A, scale bar has 0.1 mm divisions)

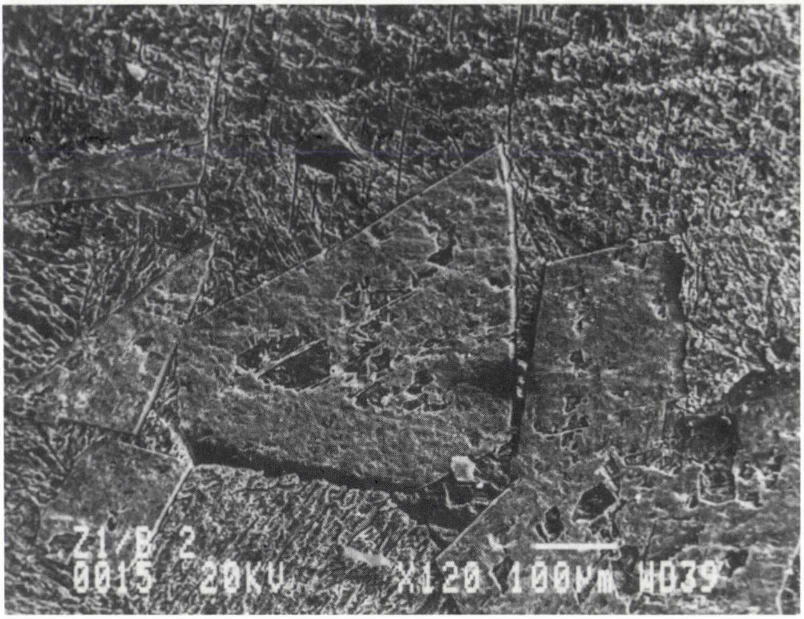


Fig. 9
Pore rimming saddle dolomite in the middle. Note the curved crystal faces and the dissolved interior part. The latter is related to dedolomitization. The remaining pore space is filled by calcite mosaic. Etched (acetitic acid) slab sample, the dolomite shows a positive relief. (SEM photograph, sample: Zuhánya quarry, Z1/B, scale bar is 0.1 mm)

(see next chapter). The paragenetic sequence of the major dolomite and calcite cement phases are shown on Fig. 11.

Dolomite and calcite geochemistry

Carbon and oxygen stable isotope (VG isotopes model MM903), XRD (Philips PW 1380 goniometer, Cu radiation and graphite monochromator), ICP (Philips 50 MHz source ICP with PV8210 air path spectrometer) and microprobe tests (JEOL Superprobe, EDAX system, ZAF software) were performed at the Postgraduate Research Institute for Sedimentology, Reading University (UK). SEM, back-scattered electron microscopy (BSEM) and cathodoluminescence analyses were carried out both at Reading University (PRIS) and at the Geological Survey of Denmark, Department of Reservoirgeology.

Table 1 shows the summary of XRD, ICP and stable isotope analyses of the mottled dolomitic limestone, the overlying dolomites, the underlying dolomitic limestones, limestones and for comparison a sub-recent speleothem.

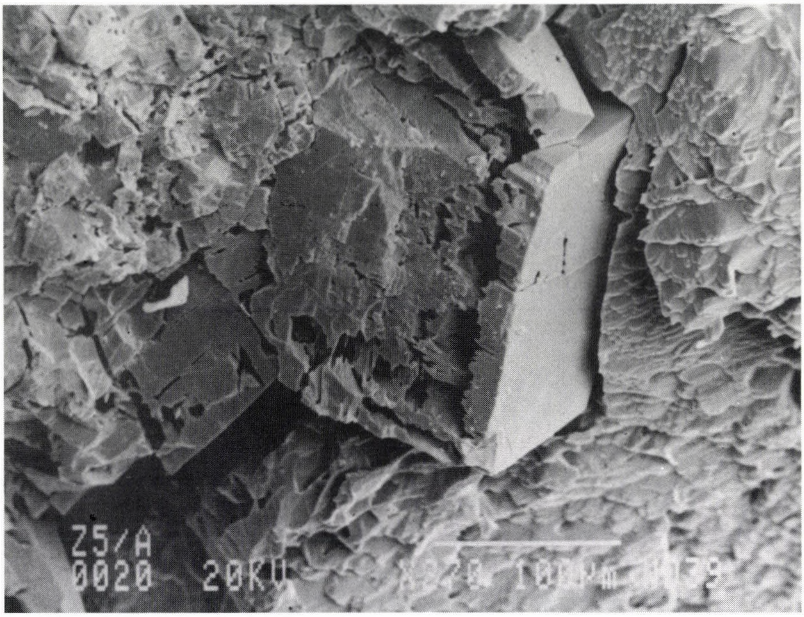
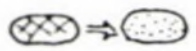
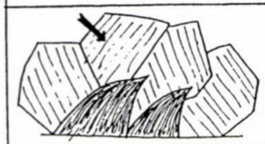


Fig. 10
Pore rimming dolomite crystal (positive relief with smooth crystal faces) with interior calcite zone, which appears in the form of cavities due to acetitic acid etching. The calcite is related to dedolomitization as it is indicated by the resorbed, serrated contact surface of the cavity. The final pore filling calcite surrounds the dolomite rhomb. (SEM photograph, sample: Zuhánya quarry, Z5/A, scale bar is 0.1 mm)

As can be seen most of the samples – except for the karst-related calcites – have a $\delta\ ^{13}C$ value of +2.6 to +1.8 indicating that no significant exchange took place between the system and outer fluid flushing. In addition the $\delta\ ^{13}C$ values of non-dolomitized micrites and bioclasts are in the same order as those of the dolomitized zones. Thus it is assumed that a calcite precursor was the major carbon source and no outer organic carbon sources were involved in the dolomitization.

The $\delta\ ^{18}O$ values are less consistent and reflect some differences in the carbonate cementation and dolomitization. The "original" micrite has values of $\delta\ ^{13}C$ + 2.6 and $\delta\ ^{18}O$ - 4.0 on average. The later cementation phases can be divided into three different groups according to $\delta\ ^{18}O$ values. The planar-e and

Fig. 11 →
The simplified paragenetic sequence of calcite cementation, dolomitization and major cement types, cathodoluminescence colour, (NL – non luminescent, DL – dull, BR – bright) diagenetic phases. In the last coloumn major changes in pore water chemistry is indicated by arrows

CEMENT TYPE AND CL COLOUR	CEMENT DESCRIPTION	EARLY DIAGENESIS SHALLOW SUBSURFACE LATE BURIAL KARST	MAJOR CHANGES IN WATER CHEMISTRY
	aragonite & high Mg calcite transform. → low Mg calcite		
 NL	calcite mosaic		
 DL maroon	cloudy core of scattered dolomite rhombs		⇦
 BR orange	middle calcitic zone of dolomite rhombs		⇦
 NL	outer limpid dolomite zone of rhombs		
 NL DL zoned	fissure filling calcite mosaic		
 BR-DL brown orange	overgrowth on dolomites and dolomite cement		⇦
 NL	saddle dolomite		
 NL	drusy calcite mosaic		⇦
 BR yellow-orange zoned	dedolomitizing calcite		⇦
 NL	calcite fissure filling		

planar-s dolomites together with the stylolite-related dolomite cements show fairly enriched composition ranging from -1.4 to -3.8 $\delta^{18}O$. This minor shift toward the positive values showing similarity to that of the overlying beds ($\delta^{18}O$ -1.5) indicates a possible downward migration of relatively cold pore fluids.

The negative (δ 18O -6.3 to -6.4) values of the dolomitized zones and burial calcite fissure fills are presumably related to the influence of ascending warmer subsurface fluids.

The karst-related cements have significantly different isotopic compositions. Both in the terms of $\delta^{13}C$ and $\delta^{18}O$ they show a depletion. The most striking difference is visible in the $\delta^{13}C$ where -5.0 to -7.2 and -10.1 values clearly indicate a different fluid system. The highly negative $\delta^{18}O$ values are probably signs of a thermal karst event rather than a cold karst system.

The Sr content of the micrites and bioclasts is one order of magnitude higher than that of the dolomites and karst-related fissures. This clearly indicates that the micrites and bioclasts were formerly aragonite or high Mg calcite. The low Sr content of dolomites is assumed to be related to two factors: 1) dolomitization took place during late diagenesis since the early replacive dolomites have higher Sr contents as a rule, and 2) the dolomites generally have a lower Sr content (nearly half) than that of the precursor calcites.

The iron and manganese accumulate in stylolite-related dolomites and dolomitized limestones. This general trend is indicative of subsurface diagenesis. Such an accumulation is correlative with the enrichment in $\delta^{18}O$. However, it is surprising that the earlier calcite fissures do not have significant iron and manganese contents. The iron and manganese were beneath the detection limit in the micrites and bioclasts as well.

The dolomite mottles are composed of non-stoichiometric dolomites having an average mole % $CaCO_3$ of 54,9 %. In contrast the overlying pervasively dolomitized beds are composed mainly of nearly stoichiometric dolomites.

Dolomitization model

A model for this kind of dolomitization must explain the following:

– the host limestone was deposited in a mid-ramp/outer ramp setting
– the overlying Mid-Triassic beds are entirely dolomitized, whereas the Jurassic and Cretaceous are not effected
– patchy appearance of dolomites in the "mottled" limestone
– part of the dolomites are associated with stylolites
– presence of baroque or saddle dolomite

Petrographic and geochemical analyses showed that there are three distinct dolomite types. The earliest scattered euhedral dolomites (planar-e types) are belived to be formed in the earliest phase of dolomitization (Török 1991). These

Table 1

Summary of XRD, stabile isotope and ICP analyses of Zuhánya Limestone (mottled dolomitic limestone unit), the underlying beds, overlying beds and for comparison a subrecent speleothem (nd. = non detectable component)

sample description (no. of samples)	calcite (%)	dolomite (%)	del 13 C (PDB)	del 18 O (PDB)	Fe (ppm)	Mn (ppm)	Sr (ppm)
neokarst , cave spelothem (1)	100	-	-10.1	-9.0	nd.	nd.	53
overlying dolomite (2)	-	100	+2.5	-1.5	nd.	nd.	95
red karstic fissure filling (2)	100	-	-7.2	-7.5	3100	52	79
karstic calcite fissure (2)	100	-	-5.0	-7.4	1200	460	81
calcite fissure (early paragenetic) (3)	100	-	+1.8	-6.4	nd.	nd.	190
entirely dolomitized limestone (2)	3	97	+1.8	-6.2	904	nd.	118
stylolite related fissure fill. dolomite II(1)	11	89	+2.8	-1.4	6932	81	315
stylolite related fissure fill. dolomite I (1)	11	89	+2.0	-3.7	7239	nd.	259
dolomite mottle (4)	13	87	+2.3	-3.6	3085	37	210
bioclasts (1)	100	-	+2.1	-5.2	nd.	nd.	1778
micrite (3)	100	-	+2.6	-4.0	67	nd.	3232
underlying limestone (2)	100	-	+3.1	-5.5	46	nd.	713
underlying dolomitized limestone (2)	59	41	+3.4	-4.8	557	nd.	450

dolomites are supposingly of shallow subsurface origin. This is indicated by the very similar isotopic composition of the dolomite mottles to that of the micrites. At the same time the significant enrichment in iron content suggests a different pore fluid system. In addition the lowered Sr content also indicates a non-penecontemporaneous origin. The scattered appearence of dolomite rhombs is indicative of a closed-system dolomitization (Sperber et al. 1984). The dolomites are non-stoichometric and have a high Ca content. This non-stoichometry and fine crystalline appearance implies a relatively early diagenetic dolomitization.

Higher temperature is assumed for the formation of planar-s mosaics and for xenotipic dolomites. One piece of evidence for this is the presence of organic and iron-rich substances and the xenotopic dolomite texture (Gregg and Sibley 1984). These cloudy crystals indicate a recrystallization process, since during the early stages of dolomitization the pore fluids were unable to dissolve the calcite completely. In addition it can be supposed that the formation of cloudy crystals is related to dilute dolomitizing fluids (Sibley 1982). These dolomites are postdated by one generation of stylolites.

Saddle dolomites are found as pore-rimming cements and often associated with stylolites. Geologic evidence showed that saddle dolmites are formed within a temperature range of 60–150 °C (Radke and Mathis 1980; Gregg 1983). Most authors agree that saddle dolomites are of burial origin (Radke and Mathis 1980; Mattes and Mountjoy 1980; Zenger and Dunham 1988; Qing and Mountjoy 1989; Amthor and Friedman 1991; Mountjoy and Halim-Dihardja 1991; Coniglo et al. 1994; Mountjoy and Amthor 1994; Zeeh et al. 1995, and many others). The high iron content and non-luminescent character (iron acts as a quencher) of the saddle dolomites also suggests a burial origin.

The initial dedolomitization occurred in relatively early diagenetic phase. It followed the formation of the scattered euhedral rhombs. Therefore it is believed to be of shallow subsurface origin.

The major dedolomitization is related to bright yellow luminescent calcites. It mainly appers in fissures and it also affected some of the saddle dolomites. The isotopic signature – the negative $\delta^{18}O$ (-7.4) and negative $\delta^{13}C$ (-5.0) values – the elevated iron and manganese content and its late appearance in the paragenetic sequence all suggests a karst related event. The responsible fluid may also have had a slightly elevated temperature.

The general isotopic trend of the karst related fractures is not consistent and indicates changes in the temperature or fluid source since the karst related cements show significant differences in $\delta^{13}C$, $\delta^{18}O$ and minor element content (see Table 1).

Discussion

The dolomitization began in a shallow subsurface setting. The fluid flow system may have been closed since only scattered rhombs were formed. The geochemistry of the dolomitizing fluid is not know, although it is quite probable, that it was of normal marine origin. The Mg source for this dolomitization thus is belived to be mainly of intraformational. The irregular distribution of dolomites is supposingly related to differences in organic/clay content of the sediments. It is not possible to explain the first dedolomitization event on the basis of available petrographic and geochemical evidences, however it clearly indicates a significant change in pore water chemistry. In the later dolomitization phase nonplanar (xenotopic) dolomites and saddle dolomites were formed indicating an elevated temperature. It is still an open question whether this elevated temperature is merely attributed to the burial diagenesis or partly related to ascending hot fluids. As the Mg source both early intraformational dolomites and under or overlying dolomites were available. The multi-stage stylolite formation partly predates and partly postdates the dolomitization.

The major (observable) dedolomitization process is of late diagenetic. According to its geochemical characaters it is related to karstification. Complex dolomite calcite cementation and dedolomitization can be related to unconformity surfaces (subaerial exposure horizons) (Holail et al. 1988), although the major Late Triassic/Jurassic unconformity is located more than 400 m higher up in the present succession.

The early dolomites were formed supposingly in the Mid-Triassic. Considering the geohistory and burial history of the region the maximal overburden and related late dolomitization may have taken place either during the Late Triassic or the Early Cretaceous. This uncertainity is related to the uncompleteness of sequences and consequently the difficulties in the tectonic interpretations. Karst related late diagenetic dedolomitization may have happened during the Early Cretaceous. However the Jurassic or post-Mesozoic age of dedolomitization can not be entirely excluded.

Conclusions

1. The depositional environment of the Middle Triassic dolomitized limestones was the outer and midramp zone of a homoclinal carbonate ramp.

2. The original sediment was aragonite and high Mg calcite (with high Sr content) which were replaced by low Mg calcite before the dolomitization as it is evidenced by the presence of calcitic bioclasts in dolomite mottles.

3. The inital phase of dolomitization took place under shallow subsurface conditions in a relative close sytem. It resulted in the formation of scattered dolomite rhombs having a distinct CL zonation of non-luminescent cloudy dolomitic cores, bright orange thin calcite and outer non luminescent limpid dolomite zones. The same origin was inferred from the stabile isotopic, minor and trace element composition.

4. The later phase dolomitization is characterised by cloudy xenotopic mosaics and later pore filling, non-luminescent saddle dolomites, which were

formed at elevated temperatures. In addition to the increasing burial depth migrating hot fluids may have also contributed to this form of dolomitization.

5. Before and after the late phase dolomitization stylolites were formed. A part of the later dolomite cements is associated to these stlylolites.

6. Calcite cementation shows several distinct phases: burial calcite mosaics, karst related calcite fissure-filling generations. Major dedolomitization is related to the zoned bright yellow luminescent isotopically light calcites of karstic origin. Some of these fissures can be the results of tectonic activity and could serve as pathways of descending cold (karst) and ascending hot fluids (thermal karst).

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Conference report

World Meeting of Hungarian earth scientists HUNGEO '96

The World Meeting of Hungarian earth science professionals – geologists, geophysicists, geographers, and cartographers – was held between 15 and 22 August 1996 in Budapest, Balatonalmádi and Vörösberény.

The principal aims of the event were:
– to provide an opportunity for earth scientists living in Hungary, in neighbouring countries and throughout the world to become directly acquainted with each other's activity;
– to promote an exchange of views between researchers, teachers and practical experts of earth sciences as well as communication, co-operation and joint publishing between certain professional fields and regions;
– to begin the elaboration of a uniform Hungarian earth science terminology as well as a modern subject matter of instruction.

The organizer of the event was the Hungarian Geological Society. It was supported by the Hungarian Academy of Sciences, the Association of Hungarian Geophysicists, the Hungarian Geographical Society, the Hungarian Society of Surveying, Maping and Remote Sensing as well as the Balaton Academy as co-organizers.

The event began with plenary sessions. The first lecturer, György Komlóssy (Budapest), analysed the problems of Hungary, and in a broader sense of the entire Central and East-Central European region, in connection with the integration into the European Union, particularly from the point of view of earth sciences. László Trunkó (Karlsruhe) presented his English language "Geology of Hungary" which is now in press. How much is known about the geophysics of Hungary? – was the question raised by László Vető (Budapest) in the title of his lecture. Perspectives of the earth sciences were outlined by Tibor Beder (Miercurea Ciuc/Csíkszereda) in his lecture "Earth sciences and strategies for the future". István Klinghammer (Budapest) presented the main features of the evolution of Hungarian cartography, which has a great past, from the scribe Lázár to the digital maps of our days.

The most important results of the last 10 years of earth science research activity in Hungary were also presented (only by lecturers from Budapest) in a plenary session. Following a historical introduction the present state of Hungarian geology was discussed. After this lecture, reviewing every aspect of the profession, the 110 professionals from 12 countries were able to become acquainted in detail with

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the results of some important fields of study (radiometric dating, geologic nature conservation, geologic mapping, exploration of natural resources) by means of lectures specifically devoted to the fields in question.

There is not enough space here to give an account of the more than 50 lectures of the section sessions; it is possible only to mention the groups of subjects and some of the more interesting topics.

In the geologic section 15 lectures were presented on mineralogy, mineralization, tectonics, hydrogeology, palaeontology and geologic nature conservation. Among them lectures on the microminerals of the meteoric shower at Mócs (Kolozs County, Transylvania) and on the Early Tertiary trace fossils in the surroundings of Eger were particularly interesting. Even the lecture on the Saint Patrick Cave on the British Isles had Hungarian aspects.

The geographical section presented a rich scientific harvest (in nearly 30 lectures). The section dealing with educational methodology and nomenclature included lectures on the subjects of maps, of teaching assistance, of the spelling of geographical names and of different teaching methods in different regions. Lectures on the topic of nature conservation and environmental protection displayed the most pressing problems of the present and the methods of dealing with them. It is remarkable that tourism was regarded as a special natural resource by one of the lecturers. Lectures belonging to the subject of political and ethnic geography dealt with serious and complicated questions; among others there were landscape and geopolitics, ethnic maps and their changes, and geography of the religions of the world. In the section on geomorphology and landscape geography lectures were held on volcano morphology, relief evolution, on the geomorphic role of sediment transport and on the questions of historical and landscape geography as well as geoecology.

In the geophysical section lectures were given on the special geophysical questions of the Pannonian Basin, on geoid undulation, and on archeological geophysics.

The closing plenary session provided a forum for reviewing as well as summarizing topics. The lectures dealt with the evolution of continents, regional hydrogeologic phenomena, the disposal of radioactive waste materials, the emerging geomuseology as well as a uniform Hungarian geoscience terminology.

The lectures were completed by visits to institutions and poster presentations (12 topics).

In the evenings debates and workshop meetings were held, common projects were elaborated, and video films were presented. Two one-day long field trips (Balaton Highland, Great Hungarian Plain) and one two-days' one (Selmecbánya [Banská Stiavnica], Northern Csallóköz [Zitny Ostrov]), the Eötvös Loránd commemoration exhibition in Tihany as well as a visit to the Hungarian Geographical Museum in Érd enriched the programme.

Perhaps the greatest benefit of the event was the possibility of having personal meetings, establishing connections, and getting acquainted with each other's work and results. The exchange of information was of inestimable value. Among other benefits the common projects, the co-ordination of research activities, the possibility to participate in field trips, the acquisition of knowledge on the spot by excursions and the possibility of integration into Hungarian scientific life should be stressed.

Because of the considerable success of the event the decision was made that the HUNGEO initiative should be continued and that the meeting should become a regularly scheduled event.

Tibor Kecskeméti

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GUIDELINES FOR AUTHORS

Acta Geologica Hungarica is an English-language quarterly publishing papers on geological topics. Besides papers on outstanding scientific achievements, on the main lines of geological research in Hungary, and on the geology of the Alpine–Carpathian–Dinaric region, reports on workshops of geological research, on major scientific meetings, and on contributions to international research projects will be accepted.
Only original papers will be published and a copy of the Publishing Agreement will be sent to the authors of papers accepted for publication. Manuscripts will be processed only after receiving the signed copy of the agreement.
Manuscripts are to be sent to the Editorial Office for refereeing, editing, in two typewritten copies, on floppy disk with two printed copies, or by E-mail. Manuscripts written by the following word processors will be accepted: MS Word, WordPerfect, or ASCII format. Acceptance depends on the opinion of two referees and the decision of the Editorial Board.

Form of manuscripts

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

- the title of the paper (with minuscule letters)
- the full name of the author
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- an abstract of not more than 200 words
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The SI (System International) should be used for all units of measurements.

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

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Figures and tables should be referred to in the text. Figures are expected in the size of the final type-area of the quarterly (12.6 x 18.6) or proportionally magnified 20–25% camera ready quality. Figures should be clear line drawings or good quality black-and-white photographic prints. Colour photographs will also be accepted, but the extra cost of reproduction in colour must be borne by the authors (in 1997 US$ 260 per page). The author's name and figure number should be indicated on the back of each figure. Tables should be typed on separate sheets with a number.

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