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Fossils and strata of Lake Pannon, a long-lived lake from the Upper Miocene of Hungary

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Foreword

In the last few decades, Late Miocene sediments of the Pannonian basin have always been discussed within the context of the evolution of the Paratethys, a chain of epicontinental marine and semi-marine basins in central to southeastern Europe. This approach was entirely correct; however, it avoided the issue of the nature of the water body that filled the basin (was it lake or a sea?) and put a strong emphasis on the biostratigraphic duality of the sediments ("Pannonian" versus "Pontian").

In the Late Miocene waters of the Pannonian basin, various organisms, particularly mollusks, underwent a spectacular endemic radiation. Hundreds of closely related species evolved from very few ancestral forms. This kind of endemic radiation is a hallmark of long-lived lakes, both fossil and recent. The evolution of the Pannonian basin endemic biota was continuous during the Late Miocene, and is not naturally divided into an older ("Pannonian") and a younger ("Pontian") fauna.

Consequently, it is reasonable to suppose that the Late Miocene sediments of the Pannonian basin were deposited in a deep, large and long-lived lake, similar to modern Lake Tanganyika or Lake Baikal. In terms of water chemistry and actual composition of the fauna, the Caspian Lake may give an even better analogue. We call this Late Miocene long-lived lake "Lake Pannon". The term "Pannonian Lake", which is often used in the literature, may lead to some confusion, because the "Pannonian Age", as understood today, does not correspond to the entire lifetime of the lake. Therefore, we prefer the designation "Lake Pannon" because it will cause less interference with stratigraphic nomenclature.

The rapid and gradual morphological change of mollusks within Lake Pannon offers good biostratigraphic markers. The first paper in the volume gives basin-wide correlations of biozones with magnetostratigraphic and chronostratigraphic data. The four subsequent papers introduce fossiliferous outcrops of Lake Pannon sediments from western Hungary (Transdanubia). All four papers are based on M.S. theses completed at the József Attila University of Szeged, Department of Geology and Paleontology, under the advisorship of Dr. Miklós Szónoky. In these papers, we placed the biostratigraphic and paleogeographic results into a broader and more consistent framework.

All papers in this volume are products of team efforts, and were sponsored by various foundations. We acknowledge the support of the Hungarian OTKA (T 007482 and T 019679), the U.S.-Hungarian Science & Technology Joint Fund (JFNo 511), the National Science Foundation (EAR-9706230), and the Albert and Alice Weeks Fund of the Department of Geology & Geophysics, University of Wisconsin-Madison. Photos in the volume were taken by László Novoszáth (József Attila University of Szeged) and the authors, and graphic work was carried out by Mary Diman (University of Wisconsin-Madison).

Madison, Wisconsin, June 1998

Imre Magyar and Dana H. Geary

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Integrated biostratigraphic, magnetostratigraphic and chronostratigraphic correlations of the Late Miocene Lake Pannon deposits

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Due to the highly endemic nature of the Late Miocene Lake Pannon's aquatic biota, the stratigraphic correlation of the lake's sediments with the global chronostratigraphic system is dependent upon mammal biostratigraphy, magnetostratigraphy, and radio-isotopic age determinations. However, correlation of mammal localities and isotopically dated volcanic formations with the lacustrine sequence is possible only in a few instances, offering limited independent data for magnetostratigraphic interpretation. In this paper we construct a correlation chart that integrates biostratigraphy (dinoflagellates and mollusks) with the results of physical dating methods. The average time resolution of our biozones (excluding deep-water mollusk zones) is 1 Ma. There is a difference, however, in the amount of available data and consequently in the reliability of our correlations between the lower (older than approximately 9.5 Ma) and upper parts of the sedimentary sequence; the former is well-established, whereas a scarcity of radio-isotopic ages from the upper part of the sequence results in more uncertainty. In addition, a biostratigraphic subdivision of the last ca. 3 Ma interval of the lacustrine sequence is almost entirely lacking. The only opportunity for interbasinal biostratigraphic correlation within the lifetime of Lake Pannon is the Early Pontian migration event, when a great number of the Lake Pannon endemic species entered the Eastern Paratethys. However, due to the scarcity of chronological data from the Late Miocene of the respective basins, correlation of the Eastern Paratethyan Pontian Stage to the Lake Pannon deposits carries at least a 1.3 Ma uncertainty.

Key words: biostratigraphy, chronostratigraphy, extinct lakes, Lake Pannon, magnetostratigraphy, Neogene, Pannonian basin, Paratethys

Introduction

Stratigraphic correlation between the Neogene Paratethyan basins and the Mediterranean has always been a delicate issue. The prevalence of endemic evolution combined with the extinction of marine organisms due to dilution of seawater make biostratigraphic correlation extremely difficult. The same

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factors, in addition to a poor understanding of basin-filling mechanisms and consequently a confusion of facies units with stratigraphic units, have resulted in considerable uncertainties in correlations even within individual basins.

The Pannonian basin, situated within the Carpathian loop in east central Europe, became separated from the rest of the Paratethys in the late Middle Miocene (Steininger and Rögl 1985). The resulting "Lake Pannon" was huge and long-lived, comparable in size to today's Caspian Lake (Kázmér 1990). During its seven million-year lifetime, the lake's basin was filled with clastic sediments of considerable thickness. In the countries of the Pannonian basin these lacustrine sediments have great economic importance; they yield oil, gas, lignite, and water and are a dominant factor in the overall landscape in much of the area. The biostratigraphy and exact age of these formations, however, has been debated since the middle of the last century (Stevanović 1990a; Marinescu 1990a; Magyar and Hably 1994).

The sediments of Lake Pannon, overlying the restricted marine Sarmatian Stage (Papp et al. 1974), have been assigned to the Pannonian (Papp et al. 1985) and Pontian (Stevanović et al. 1990) regional stages. Since the base of the Pontian Stage in the Pannonian basin is vaguely defined and variably interpreted, we will avoid its use in this paper and will refer to the sediments of Lake Pannon without using regional stage names.

Our purpose here is to integrate the variety of new chronostratigraphic data that have been published in the last few years in order to establish a correlation that can be used for practical purposes such as hydrocarbon exploration. We use the data (but not necessarily the original interpretation thereof) from a variety of published sources. Below we briefly describe the biostratigraphic zonations and take into account the available magnetostratigraphic and radio-isotopic data. Secondly we correlate these data in a chart and explain our reasoning behind the correlations. Lastly we discuss uncertainties and contradictions resulting from our correlation as well as the implications for Central–Eastern Paratethyan regional correlation.

Biostratigraphy

The biostratigraphy of Lake Pannon deposits is based mainly on dinoflagellates, mollusks, ostracods, and mammals. Ostracod workers have elaborated high-resolution stratigraphic systems for Lake Pannon deposits (e.g. Jiříček 1985b; Krstić 1985; Sokač 1990) but because both reliable tie-points and a generally accepted zonation are lacking we did not correlate their systems with the zonations given here. With respect to mammals we will give only a short summary of their correlation to the lacustrine sequence.

The average duration of our dinoflagellate and mollusk zones (and subzones) is about 1 million years. This level of resolution is at least as good as the contemporaneous marine plankton stratigraphy and certainly disproves the notion that no appropriate biostratigraphy can be based on the endemic organisms of Lake Pannon (e.g. Elston et al. 1994 and Pogácsás et al. 1994 in Teleki et al. 1994, with a notable absence of any other references to Lake Pannon biostratigraphy in the rest of this volume).

Dinoflagellata

Study of the organic-walled microplankton of the Lake Pannon deposits began in 1979; biozones were established in 1988 (Sütő-Szentai 1988). Two of these eight zones were not real biozones; they reflect unfavorable ecological conditions for dinoflagellates or the dominance of freshwater algae rather than age. A seventh zone has been added to the six valid zones (Sütő-Szentai 1991a, 1991b); these are (from youngest to oldest):

Galeacysta etrusca Zone
Spiniferites validus Zone
Spiniferites paradoxus Zone
Pontiadinium pecsvaradensis Zone
Spiniferites bentorii oblongus Zone
Spiniferites bentorii pannonicus Zone
Mecsekia ultima Zone

The lowermost biozone is marked by the presence of *Mecsekia ultima*, the taxonomic position of which is uncertain. The rest of the zonation, however, is primarily based on the evolution of the dinoflagellate *Spiniferites bentorii* and its descendants. Morphological changes between species may be gradual, so that zone boundaries are not always sharp and minor uncertainties in their determination may occur.

Mollusca

The stratigraphy of Lake Pannon deposits was based exclusively on mollusks for almost one hundred years. The first comprehensive zonations appeared at the beginning of this century and have been modified by numerous authors since then. Views on mollusk biostratigraphy are somewhat divergent even today (Papp et al. 1985; Korpás-Hódi 1987; Stevanović et al. 1990; Müller and Magyar 1992a).

Benthic bivalves and gastropods are rarely regarded as effective tools for stratigraphy, in part because they are believed to be too conservative in evolutionary rate. In Lake Pannon, however, mollusk evolution was rapid; hundreds of new species appeared within the life span of the lake. A large variety of benthic molluskan assemblages populated the lake, whose distribution was controlled by water depth and other ecological factors. Communication between assemblages appears to have been rather restricted and each appears to have had its own evolutionary history. For stratigraphic purposes we consider only three generalized biotopes: deep water, sublittoral, and littoral.

The zonation presented here is based on elements of earlier studies by Papp (1951), Stevanović (1951), Korpás-Hódi (1987), and Müller and Magyar (1992a). It is not a formal proposal for the erection of biostratigraphic units in the Lake Pannon sediments; instead we try to establish the geochronology of evolutionary events in mollusks. The lower boundary of each biozone is defined by the evolutionary appearance of the name-giving taxon. Succeeding zones are named in the same way, regardless of the fate of the name-giving taxon or any other typical species of the preceding zone. Geologically gradual morphological changes between species can result in somewhat fuzzy zone boundaries. Instead of giving a lengthy description of the accompanying fauna of the name-giving taxon we will refer to individual outcrops or boreholes where the typical fauna of the given zone was documented.

Deep water

We distinguish two zones in the deep-water deposits: the older *Congeria banatica* and the younger "*Congeria" digitifera* Zones. The relation between these two species is not clear. It has variously been proposed that *digitifera* is a descendant of *banatica* (Stevanović 1990b), that *digitifera* comes from another ancestor designated *subdigitifera* (Stevanović 1978), that *digitifera* belongs to the genus *Dreissenomya* (hence it is not closely related to *banatica*; Stevanović 1978, 1990b), and even that the two forms belong to a single species (Széles 1971). In addition, *banatica* has been found in the youngest layers of Lake Pannon, although in a single locality only (Magyar 1991). Therefore, we cannot claim with certainty that this twofold stratigraphic subdivision (*C. banatica* Zone and *C. digitifera* Zone) is based on a phylogenetic lineage. A typical fauna of the *C. banatica* Zone has been depicted from Beočin (Stevanović and Papp 1985) and from the Szombathely-II borehole, at depths of ca. 1500 to 1700 m (Korpás-Hódi 1992; for geographic locations see Fig. 1). A representative fauna of the "*C." digitifera* Zone occurs in the Szombathely-II borehole at depths of ca. 1350 to 1400 m (Korpás-Hódi 1992; we assign her "*C. banatica*" from 1386 m to "*C." digitifera*).

Sublittoral

The *"Lymnocardium" praeponticum* Zone represents the base of the Lake Pannon sedimentary sequence in the sublittoral facies (Jámbor and Korpás-Hódi 1971). A typical assemblage for this zone was retrieved from the Lajoskomárom-1 borehole, between 663 and 670 m (Jámbor et al. 1985, 1987). For further subdivision of the sublittoral facies we use the *Congeria czjzeki–C. zagrabiensis*, the *Lymnocardium schedelianum–L. sopronіense*, and the *Congeria zsigmondyi–C. rhomboidea* (Stevanović 1978) lineages (Fig. 2). The *"L." praeponticum* Zone is overlain by the *Congeria czjzeki* Zone. The upper part of the *C. czjzeki* Zone is divided into the *L. schedelianum* and the overlying *L. soproniense* Subzones. The former is well represented in Vösendorf (Papp 1985),

Fig. 1
Localities referred to in the text

the latter in the Sopron brickyards (Vitális 1951). Both subzones contain *C. zsigmondyi*, which evolved into *C. praerhomboidea* at about the same time as *C. czjzeki* evolved into *C. zagrabiensis*. Therefore we regard *C. zsigmondyi* as indicative of the *C. czjzeki* Zone and *C. zagrabiensis* as indicative of the overlying *C. praerhomboidea* Zone. A typical fauna with *C. praerhomboidea* was described from Donja Trnava by Stevanović (1980). The *C. praerhomboidea* Zone is overlain by the *C. rhomboidea* Zone, the youngest sublittoral zone of the Lake Pannon sequence. The fauna of this zone is well known, for example, from Tirol (Marinescu 1973), Jazovnik (Stevanović 1990d), and Bátaszék (Lennert et al. this volume).

Littoral

For the lower part of the lacustrine sequence, we tentatively base our zonation on the evolutionary lineage leading from the Sarmatian *Congeria praeornithopsis* through *C. ornithopsis* to *C. hoernesi* (Papp 1951, 1953; Fig. 2). Therefore, at the base of the lacustrine sequence we distinguish the *C. ornithopsis* Zone (e.g. Leobersdorf, "zone B" by Papp 1951) and the *C. hoernesi* Zone (e.g. Leobersdorf, zone "C" by Papp, 1951). A remarkable alternative to this zonation has been put forward by Korpás-Hódi (1985).

The base of the next zone is defined by the first appearance of *Lymnocardium conjungens;* which, according to Papp (1951), is younger than the first appearance of *C. hoernesi* (Fig. 2). The fauna of the *Lymnocardium conjungens* Zone is well represented in Eisenstadt (Lueger 1980). The evolution of *L. ponticum* from *L. edlaueri* (Fig. 2) indicates the base of the *L. ponticum* Zone, the fauna of which from Dáka is described in this volume by Szilaj et al. The evolution from *Lymnocardium ponticum* towards *Prosodacnomya vodopici* (Fig. 2) was described by Müller and Magyar (1992b). The following stratigraphic units are based on this lineage: the *L. decorum* Zone (e.g. Rădmăneşti, Gillet and Marinescu 1971; Marinescu 1990b; Tihany, Müller and Szónoky 1990) with the *L. serbicum* Subzone in its uppermost part (Orešac-2; Stevanović 1990c); and the *Prosodacnomya* Zone including the following subzones: *P. carbonifera* Subzone (Kötcse-120; Strausz 1942; Müller and Magyar, 1992a), *P. dainellii* Subzone (Kurd; Lőrenthey 1894), *P. vutskitsi* Subzone (Tab; Bartha 1956), and *P. vodopici* Subzone (Grgeteg; Stevanović 1990e).

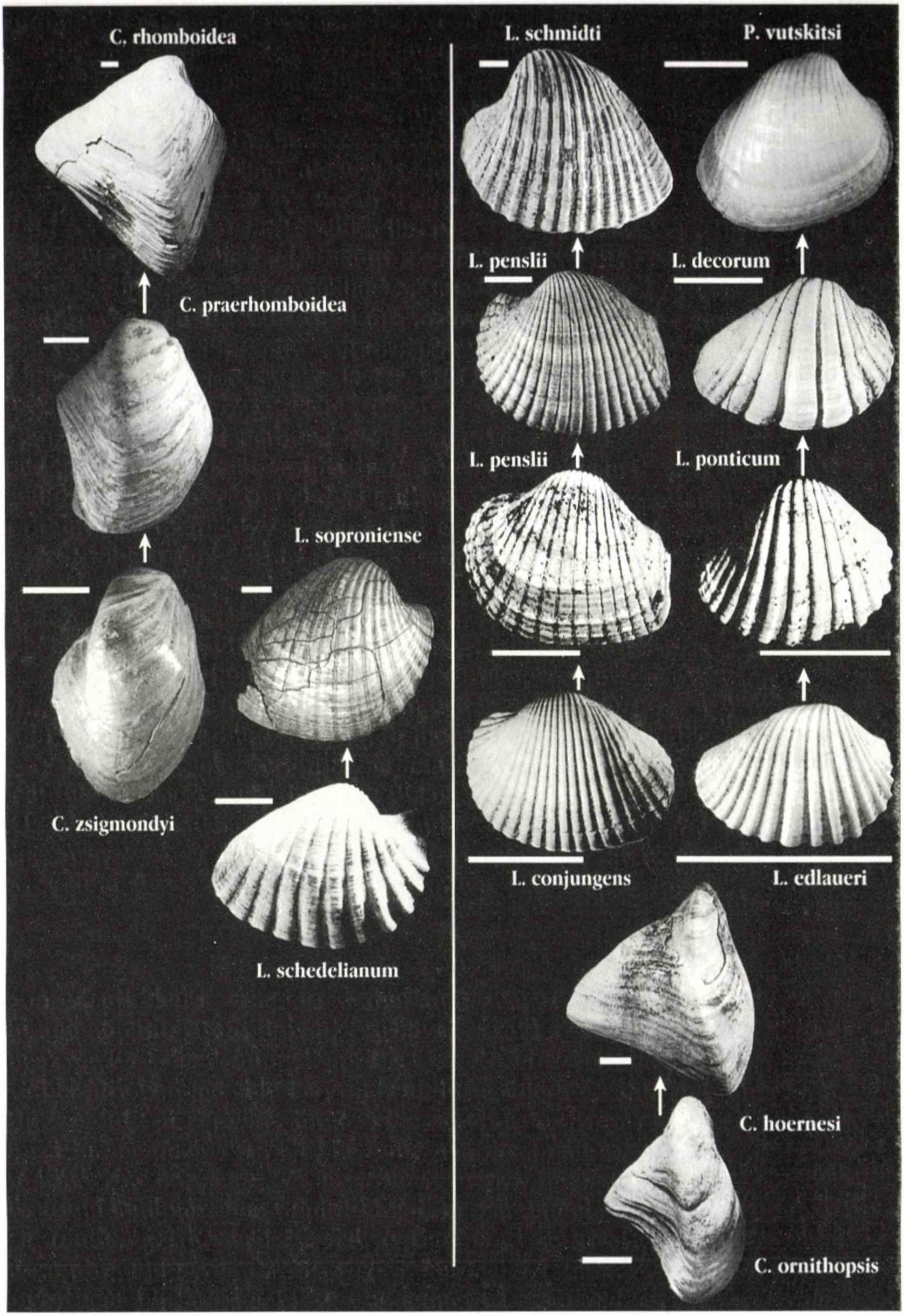
Mammalia

Mammal finds have proven very useful in correlating Lake Pannon sediments with the global time scale. In fact, Lake Pannon deposits were considered Pliocene before mammal stratigraphy established their Late Miocene age. However, they are scarcely used for intrabasinal stratigraphic correlation, because most of the mammal fossils come from fluvial or terrestrial sediments that overlie the lacustrine sequence. Therefore, usually only the minimum age of the underlying lacustrine layers can be estimated from the mammals. In some exceptional cases mammal remains have been recovered from biostratigraphically well-constrained littoral sequences: from the *C. ornithopsis* Zone at Drassburg, from between the *C. ornithopsis* and *C. hoernesi* Zones at Hovorany, from the *L. conjungens* Zone at Vösendorf, and from the *L. decorum* Zone at Tihany.

Magnetostratigraphy

The Hungarian Geological Institute drilled a series of deep, continuously cored, stratigraphic test holes to investigate the correlation and timing of deposition of Late Miocene and Pliocene strata in the Pannonian basin.

Fig. 2 →
Representation of some of the key lineages in the subdivision of Lake Pannon sediments. Sublittoral bivalves are at left, littoral ones at right. Scale bars indicate 1 cm; generally there is a significant increase in size within each lineage from older to younger species. C: *Congeria,* L: *Lymnocardium,* P: *Prosodacnomya*


C. rhomboidea
C. praerhomboidea
C. zsigmondyi
L. sopronìense
L. schedelianum
L. schmidti
P. vutskitsi
L. penslii
L. decorum
L. penslii
L. ponticum
L. conjungens
L. edlaueri
C. hoernesi
C. ornithopsis

Lithological, sedimentological, and paleontological work was supplemented by detailed magnetostratigraphic studies in ten boreholes (Fig. 1). Additionally integrated studies were also carried out on several outcrops (e.g. Budapest-Kőbánya, Tihany). Details of the paleomagnetic studies have been published elsewhere (Elston et al. 1990, 1994; Kókay et al. 1991; Lantos et al. 1992; Lantos and Elston 1995); a short summary is given here.

Unconsolidated paleomagnetic samples were placed in plastic boxes and sealed. Following the measurement of natural remanent magnetization pilot samples, representing different lithologies, depths, and inclinations were selected for progressive alternating field (AF) demagnetization. Geologic studies and subsurface correlations indicate that Lake Pannon strata accumulated rapidly and were buried promptly. Sediments display no evidence for movement of connate or ground water and the strata have remained undisturbed and unexposed since burial. Therefore the samples most likely display original magnetizations; minor secondary magnetizations disappeared at 10–30 mT (Elston et al. 1990, 1994; Lantos and Elston 1995). Several pilot samples from more consolidated rocks were demagnetized thermally. Very small differences in inclinations were observed between the thermal and AF demagnetizations (Lantos and Elston 1995). The remaining samples were demagnetized in 15–30 (40) mT.

Magnetostratigraphic studies are very scarce in the Pannonian Basin outside of Hungary. One of us (Lantos) has processed samples from the clay pit at Hennersdorf; these samples display normal polarity. One section has been published from SE Slovakia (Vass et al. 1992) but it is difficult to correlate with other sections because of the lack of tie-points, including biostratigraphically useful fossils.

Radio-isotopic ages

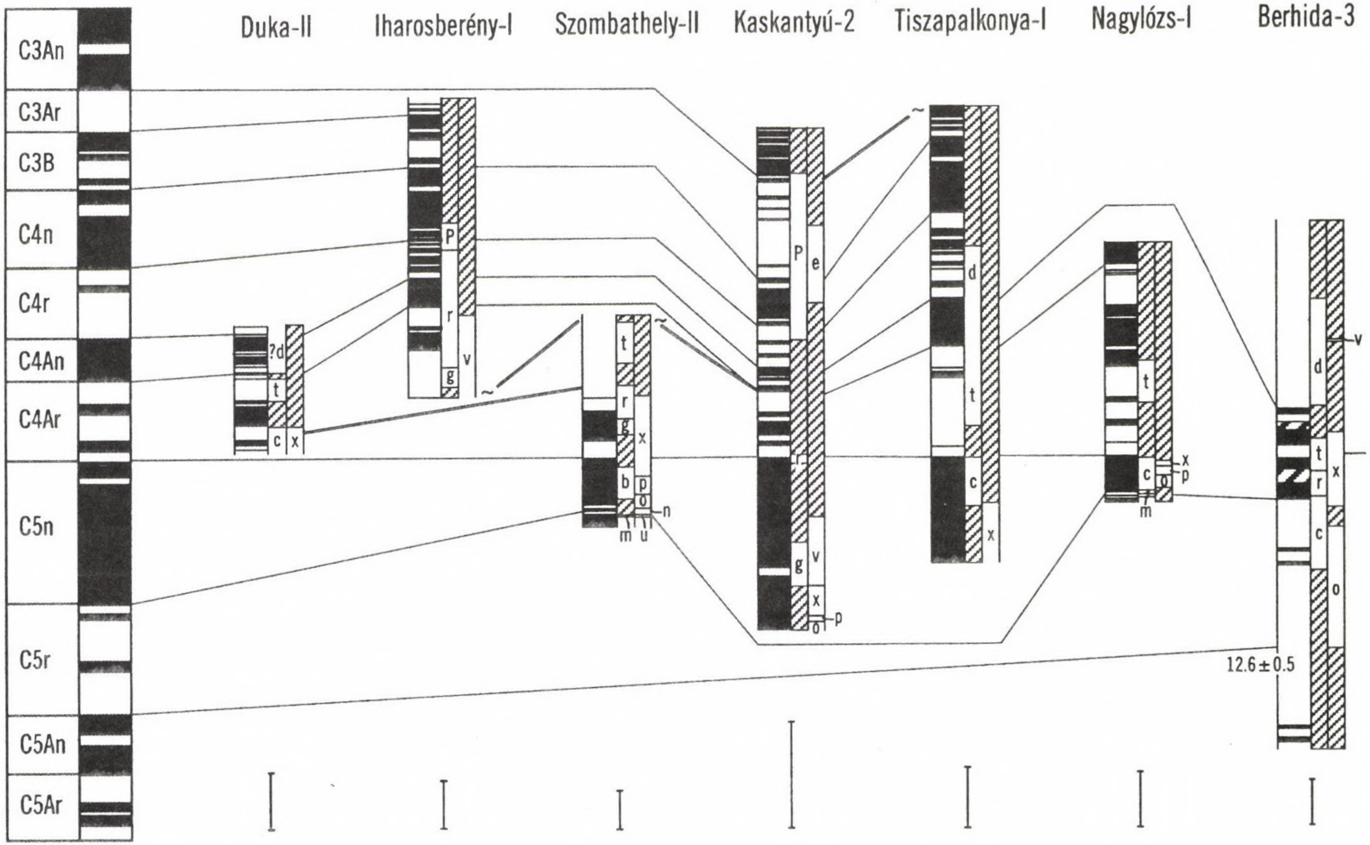
Radio-isotopic ages are crucial for age calibration of the biozones and for their correlation to the global scale. Although the Pannonian basin is rich in Neogene volcanics only a few K/Ar age determinations can be readily inserted into the sedimentary stratigraphic sequence (see Table 1). The biostratigraphic position of volcanic formations was established by direct correlation, except for the Zebrac andesite. In this case the radiometric age indicates the commencement of volcanic activity in the Călimani Mountains, and the biostratigraphic position indicates the first appearance of volcanoclastics in the sedimentary sequence of the adjacent Transylvanian basin (Marinescu 1985).

Correlation of the data

The different types of data were cross-checked in variable directions. We did not favor any particular method as calibration for the others. We present our key data here in their order of appearance in the overall correlation chart (Fig. 4).

Table 1
Correlation of radio-isotopic ages and biozones for the Lake Pannon sequence

Locality	Stratigraphic position	Material	Age ± error (Ma)	Source
Tihany	contains a clastic xenolith belonging to *P. carbonifera* Subzone	basalt volcanic bombs	7.80±1.07 (average of 21 samples)	Balogh (1995), Müller & Magyar (1992a)
Kiskunhalas-Nyugat-3 borehole, 1162–1167 m	*S. paradoxus* Zone determined from overlying marl	basalt	9.61±0.38	Balogh et al. (1986)
Bácsalmás-1 borehole, 490 m	between the *S. paradoxus* Zone and the *S. validus* Zone	biotite crystals of a rhyolite tuff	9.6±1.0	Balogh in Kovács (1992), Sütő-Szentai (1988)
Zebrac	upper part of the *C. czjzeki* Zone, upper part of the *L. conjungens* Zone	subvolcanic andesite-diorite	10.64±0.55, 10.10±0.61 (average: 10.37)	Peltz et al. (1987), Pécskay et al. (1995), Marinescu (1985)
Cavnic	upper part of the *C. czjzeki* Zone	pyroxene andesite	10.7±0.5	Edelstein et al. (1992), Pécskay et al. (1994)
Ilba	*C. hoernesi* Zone or L. conjungens Zone	quartz andesite	11.0±0.5	Edelstein et al. (1992), Pécskay et al. (1994)
Nagykozár-2 borehole, 263 m	between the *S. bentorii pannonicus* Zone and the *S. bentorii oblongus* Zone	biotite crystals of a rhyodacite tuff	11.6±0.5	Balogh in Vass et al. (1987), Sütő-Szentai (1991a)
Berhida-3 borehole, 222.6 m	between fossiliferous Sarmatian marine sediments and Lake Pannon sediments	dacite tuff in terrestrial beds	12.6±0.5	Balogh in Kókay et al. (1991)


C3An
C3Ar
C3B
C4n
C4r
C4An
C4Ar
C5n
C5r
C5An
C5Ar
Duka-II
Iharosberény-I
Szombathely-II
Kaskantyú-2
Tiszapalkonya-I
Nagylózs-I
Berhida-3
12.6 ± 0.5

Paleomagnetic record and radiometric data

The polarity zones of the test hole sections were originally correlated with the polarity time scale of Berggren et al. (1985), employing radio-isotopic ages, and the results of stratigraphic, paleontological, sedimentological and seismostratigraphic studies (Elston et al. 1990, 1994; Lantos et al. 1992; Lantos and Elston 1995; Pogácsás et al. 1994). Magnetostratigraphic correlations are anchored to the long normal polarity interval of Chron C5n (Fig. 3). No radio-isotopic data were available in stratigraphically higher parts of the sections; the age of the younger strata is thus somewhat uncertain.

Seismic stratigraphic profiles connect magnetostratigraphic test holes as well as boreholes that contain K/Ar radio-isotopic ages. The basic assumption of seismic stratigraphy is that seismic reflections are generated by bedding planes, stratal surfaces, and unconformities and are therefore essentially synchronous. The profiles provide abundant seismic reflectors, which follow stratigraphic time lines in the Lake Pannon sequence (e.g. Pogácsás 1987). The ages determined from magnetostratigraphy and K/Ar measurements were assigned to seismic horizons and then extended across the basin (e.g. Pogácsás et al. 1992, 1994). These integrated studies lead to a stratigraphic and depositional framework for the basin, the results of which are internally consistent (Elston et al. 1990, 1994; Lantos et al. 1992; Pogácsás et al. 1990, 1994).

All magnetostratigraphic records were re-examined and correlated with the new time scale of Berggren et al. (1995). The methods of correlation and tie-points were the same as before.

Dinoflagellate zones

For correlating the magnetostratigraphic record and the dinoflagellate zones, we primarily used boreholes where both methods were available.

The *M. ultima* and the *S. b. pannonicus* Zones were identified only in the Szombathely-II borehole, at the base of the Lake Pannon sediments. The top of the *M. ultima* Zone and the lower half of the *S. b. pannonicus* Zone displayed reverse polarity. We tentatively correlate this reverse polarity interval with C5r

← Fig. 3

Correlation of magnetic polarity and biostratigraphic data in seven boreholes. Only the relevant parts of the borehole profiles are shown. Magnetostratigraphic correlations are anchored to the top of the long normal polarity interval of Chron C5n. Thick lines indicate correlation along seismic profiles. Wave-shaped sign indicates ending of seismic reflectors by onlap or by erosional truncation. Columns next to polarity profiles indicate mollusk (left) and dinoflagellate (right) biozones. Shaded area shows lack of fossils or fossils with uncertain zone affiliation. Vertical scale bars below profiles correspond to 200 m, except for Berhida-3 where it is 20 m. Biozones indicated as: mollusks: P: *Prosodacnomya*, d: *decorum*, t: *ponticum*, c: *czjzeki*, r: *praerhomboidea+rhomboidea*, g: *digitifera*, b: *banatica*, m: *praeponticum*; dinoflagellates: e: *etrusca*, v: *validus*, x: *paradoxus*, p: *pecsvaradensis*, o: *oblongus*, n: *pannonicus*, u: *ultima*. See Figure 1 for location of profiles because it underlies C5n, but we cannot exclude the possibility of an even older age for this interval (see discussion below).

For the lower boundary of the *S. b. oblongus* Zone we have three sources of data and three different ages. In the Szombathely-II borehole this boundary is within a mixed polarity zone correlated to C5n; in the Berhida-3 borehole, where it directly overlies a volcanic layer dated at 12.6±0.5 Ma, it has been originally correlated to C5Ar (Kókay et al. 1991); on the other hand in the Nagykozár-2 borehole the radiometric age of this boundary is 11.6±0.5 Ma (Sütő-Szentai, 1991a). In our chart we accept the Nagykozár radiometric age as an overall average because the relevant mixed-polarity part of the Szombathely borehole might also be interpreted as belonging to C5r, and because the lower part of the *S. b. oblongus* Zone in the Berhida-3 borehole was identified from scarce material in a terrestrial sequence, where a hiatus may occur between this biozone and the underlying volcanic layer. For the Berhida borehole we offer an alternative paleomagnetic interpretation to the one published by Kókay et al. (1991) (see discussion below). The *S. b. oblongus* Zone was correlated to C5n in boreholes Kaskantyú-2 and Nagylózs-1.

The lower boundary of the *P. pecsvaradensis* Zone is within C5n in the Kaskantyú-2, Szombathely-II, and Nagylózs-1 boreholes. This zone is generally thin and its exact location within C5n is not known.

The lower boundary of the *S. paradoxus* Zone is within C5n in several boreholes: Kaskantyú-2 (Elston et al. 1994), Szombathely-II, Nagylózs-1, and Zsira-1. The Tiszapalkonya-I borehole simply indicates that part of the *S. paradoxus* Zone (without its lower and upper boundaries) is within C5n (Elston et al. 1994). The *S. paradoxus* Zone was identified in a core sample from directly above the 9.61± 0.38 Ma old basalt in the Kiskunhalas-Nyugat-3 borehole.

The lower boundary of the *S. validus* Zone is identified only in the Kaskantyú-2 borehole, where it is well within C5n (Elston et al. 1994). However, data from the Szombathely-II, Duka-II, and Torony-1 boreholes indicate that the top of the underlying *S. paradoxus* Zone is in C4Ar. This discrepancy is further treated in the discussion below. The *S. validus* Zone is correlated to C4Ar in the Iharosberény-I borehole, and to ?C4r in the Berhida-3 borehole. The isotopic age for the boundary of the *S. paradoxus* and *S. validus* Zones is 9.6±1.0 Ma from the Bácsalmás borehole (Kovács 1992). Correlation with the littoral mollusk zones (see below) suggests that this zone boundary is in the upper part of C4Ar.

The lower boundary of the *G. etrusca* Zone is uncertain. In the Kaskantyú-2 borehole this zone was correlated to C3B (Elston et al. 1994; Sütő-Szentai 1991b), but the lower and upper boundaries of the zone are not known in this borehole.

Deep-water mollusk zones

We have no data on the chronostratigraphic position of the lower boundary of the *C. banatica* Zone. In the Szombathely-II borehole the *C. banatica* Zone is

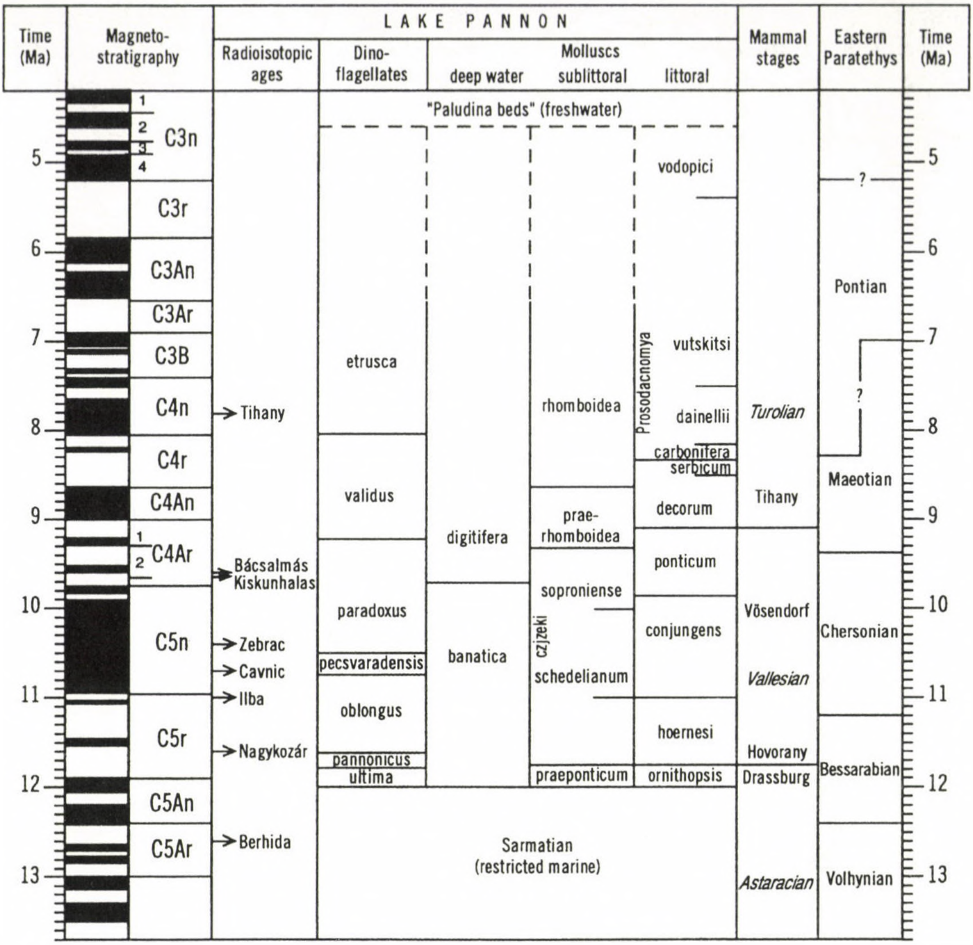


Fig. 4
Correlation chart for the Lake Pannon deposits

within C5n, and the first appearance of *Congeria banatica* in that borehole correlates with the boundary of the *S. b. oblongus* and *P. pecsvaradensis* Zones. In other boreholes the *C. banatica* Zone correlates with various dinoflagellate zones: the *S. b. pannonicus* and *S. b. oblongus* Zones in Lajoskomárom-1 (Jámbor et al. 1985), the *S. b. oblongus* Zone in Tököl-1 and in Beočin surface outcrops, and the *S. paradoxus* Zone and C4Ar3 in Duka-II.

The *C. digitifera* Zone correlates with the *S. paradoxus* Zone and magnetic chron C4Ar in the Szombathely-II borehole (Korpás-Hódi 1992), the *S. paradoxus* Zone in the Tét-5 borehole, the *S. validus* Zone and C4Ar in the Iharosberény-I borehole, the *S. validus* Zone and C5n in the Kaskantyú-2 borehole (cf. Elston et al. 1990 and Korpás-Hódi et al. 1992; see discussion below), and the *S. validus* and *G. etrusca* Zones in the Bácsalmás-1 borehole (Kovács 1992). Seismic correlations indicate that in boreholes from the Battonya area the *C. digitifera* Zone is even younger than C3Ar (Pogácsás et al. 1992; Magyar 1991; Mattick et al. 1994; see discussion below).

We emphasize the significance of these correlations of the deep water *C. digitifera* Zone with relatively young deposits. There has been a tendency to regard the deep water *C. digitifera* Zone as "Lower Pontian" as opposed to a sublittoral or littoral "Upper Pontian" "substage" (Stevanović 1990a). This distinction, however, is based on facies only and is not justified by chronostratigraphy (Magyar 1991).

Sublittoral mollusk zones

The *"L". praeponticum* Zone is the lowermost unit of the Lake Pannon sequence and correlates with the *M. ultima* Zone in boreholes Lajoskomárom-1 (Jámbor et al. 1985), Szombathely-II (Korpás-Hódi 1992), Tengelic-2 (Korpás-Hódi 1982; Sütő-Szentai 1982), and Szirák-2 (Hámor 1992), and with C5r in Nagylózs-1.

The lower boundary of the *C. czjzeki* Zone is uncertain. In the Berhida-2 and 3 boreholes the *C. czjzeki* Zone overlaps the *S. b. oblongus* Zone (Kókay et al. 1991); therefore, we tentatively put its boundary within the *S. b. pannonicus* Zone. The *C. czjzeki* Zone correlates with the *S. paradoxus* Zone in a number of boreholes from the NW foreland of the Transdanubian Central Range in Hungary (cf. data from Korpás-Hódi 1983 and Sütő-Szentai 1991b). A radio-isotopic age from the upper part of the *C. czjzeki* Zone from Cavnic is 10.7±0.5 Ma (Pécskay et al. 1994). The *L. schedelianum* Subzone is correlated to the *S. paradoxus* Zone in Umka and exhibits normal polarity in Hennersdorf. Korpás-Hódi (pers. comm.) has indicated that *L. schedelianum* occurred in the *S. b. pannonicus* Zone in the borehole Budajenő-2, which suggests that either the base of the *S. b. pannonicus* Zone is younger, or the base of the *L. schedelianum* Subzone is older, than indicated in Fig. 4. The *L. sopronienseˇ* Subzone is within the *S. paradoxus* Zone in Tata, and also shows normal polarity in Sopron.

The lower boundary of the *C. praerhomboidea* Zone, marked by the first appearance of *C. praerhomboidea* and *C. zagrabiensis*, is within the upper part of the *S. paradoxus* Zone, as proved by boreholes Berhida-2 and -3 (Kókay et al. 1991), Szombathely-II (Korpás-Hódi 1992), Tengelic-2 (Korpás-Hódi 1982; Sütő-Szentai 1982), Paks-2, or coincides with the lower boundary of the *S. validus* Zone (in Lajoskomárom-1; Jámbor et al. 1985). The magnetostratigraphic correlations are not unanimous as to whether this boundary is within C4Ar (Szombathely-II), or even in the uppermost part of C5n (Kaskantyú-2), but this may be due to uncertain determination of the transitional forms between *Congeria czjzeki* and *C. zagrabiensis*.

The lower boundary of the *C. rhomboidea* Zone, marked by a distinct evolutionary step from *Congeria praerhomboidea* to *C. rhomboidea* (Stevanović, 1978), is probably within the *S. validus* Zone. The *C. rhomboidea* Zone currently represents the youngest biostratigraphic unit in the sublittoral deposits and is coeval with the *G. etrusca* Zone in boreholes Kaskantyú-2 (cf. Elston et al. 1994 and Korpás-Hódi et al. 1992), Szentlőrinc-XII (Sütő-Szentai 1991a), and in surface outcrops in Bátaszék (Lennert et al., this volume) and Grgeteg. Seismic correlations from the Battonya area indicate that its upper boundary is younger than C3Ar (Magyar 1991; Pogácsás et al. 1992; Mattick et al. 1994; see discussion below).

Littoral mollusk zones

The littoral mollusk zones can rarely be correlated directly to dinoflagellate zones because the characteristic markers of the latter are usually absent in littoral sediments. Therefore, we used the rare co-occurrences of littoral and sublittoral mollusks, as well as magnetostratigraphic and radiometric ages as tie-points.

Lacking any microplankton or magnetic data correlation of the lowermost littoral zone (*C. ornithopsis* Zone) is based merely on its stratigraphic position. Correlation of the *C. hoernesi* Zone with the *S. b. pannonicus* Zone is based on *Melanopsis fossilis*, which according to Papp (1951) has its first appearance in the *C. hoernesi* Zone and is also known from the *S. b. pannonicus* Zone in the Berhida-2 borehole. Papp (1951) also reports the first appearance of the sublittoral *Congeria czjzeki* from the *C. hoernesi* Zone.

According to Papp (1951), the first appearance of *L. conjungens* is in the same stratigraphic horizon as that of *L. schedelianum*. The *L. conjungens* Zone is characterized by normal polarity in Ukrainian Transcarpathia (Semenenko, pers. comm., 1994), and at least in its upper part coincides with the *S. paradoxus* Zone in borehole Aderklaa-1 (cf. Papp 1951 and Fuchs and Sütő-Szentai 1991) and in Umka. The *L. conjungens* Zone correlates with the *L. schedelianum* Subzone in the vicinity of Vienna (Papp 1951) and Eisenstadt (Lueger 1980), as well as in the eastern (Boghiş) and southern (Kreka; Stevanović 1985) margins of the Pannonian basin, and even with the lower part of the *L. soproniense* Subzone in Sopron. Radiometric ages for the *L. conjungens* Zone range from 10.10±0.61 (Zebrac) to 11.0±0.5 (Ilba; Peltz et al. 1987; Pécskay et al. 1994).

A temporal overlap between the littoral *L. ponticum* Zone and the sublittoral *L. soproniense* Subzone in NW Hungary can be inferred from data of Korpás-Hódi (1983). Therefore, the lower part of the *L. ponticum* Zone can be correlated with the *L. soproniense* Subzone, and its upper part to the *C. praerhomboidea* Zone in the Berhida-3 borehole (Kókay et al. 1991). The *L. ponticum* Zone was identified in C4Ar in the boreholes Nagylózs-1, Szombathely-II (Korpás-Hódi 1992), Duka-II, Berhida-3, Tiszapalkonya-I, and within the *S. paradoxus* Zone in Berhida-3 (Kókay et al. 1991).

The *L. decorum* Zone extends from C4An to C4r in the Duka-II, Tiszapalkonya-I, and Berhida-3 boreholes. Seismic correlation with the

Hajdúszoboszló area showed that the *L. decorum* Zone there also correlates with C4r (Müller and Magyar 1992a). The *L. decorum* Zone yielded dinoflagellates belonging to the *S. validus* Zone in Tihany-Fehérpart and in the borehole Berhida-3. *Lymnocardium decorum* and *Congeria rhomboidea* were reported to occur together in Bucovăţ (Marinescu et al. 1977). The *L. serbicum* Subzone correlates to the *C. rhomboidea* Zone in the surface outcrop of Orešac-2 (Stevanović 1951).

The lower boundary of the *Prosodacnomya* Zone, and even the lower boundary of the *P. dainellii* Subzone, is within C4r in the Iharosberény-I and Kaskantyú-2 boreholes. The lower boundary of the *Prosodacnomya* Zone is constrained by radiometric ages from Tihany, where a block of silt belonging to the *P. carbonifera* Subzone was found as xenolith in a basalt agglomerate. The onset of activity of the Tihany volcano is dated at 7.8 Ma (Balogh 1995). The age of a specimen of *Prosodacnomya* cf. *dainellii* in a Hajdúszovát borehole was defined by seismic correlation as C4n. Using the same method the age of the *P. vutskitsi* Subzone in a borehole from Szarvas was C3Ar (Müller and Magyar 1992a). The *Prosodacnomya* Zone can be correlated with the *C. rhomboidea* Zone and *G. etrusca* Zone in many boreholes, including Battonya (Magyar 1991), Kaskantyú-2 (cf. Elston et al. 1994 and Korpás-Hódi et al. 1992), Tengelic-2 (Korpás-Hódi 1982; Sütő-Szentai 1982), Jánoshalma-1 (Franyó 1988), and in surface outcrops at Nyugotszenterzsébet (Bujtor 1992) and Bátaszék (Lennert et al., this volume).

Mammal zones

Mammal finds indicate that the stratigraphic range of the Lake Pannon sediments is from Late Astaracian to Late Turolian, or Mein-zone 8 to 13 (Kordos 1987; Rögl et al. 1993; Bernor et al. 1993; Rögl and Daxner-Höck 1996). The Drassburg mammal fauna from the *C. ornithopsis* Zone (Zapfe 1951) is "a typical Astaracian (MN 7/8) large mammal fauna" (Rögl and Daxner-Höck 1996). The "Hipparion"-containing (Early Vallesian) fauna of Hovorany was collected from lagoonal layers between the *C. ornithopsis* and *C. hoernesi* Zones (Čtyroký 1975; Čtyroký in Steininger et al. 1987). Thus, the Astaracian/Vallesian boundary approximately correlates with the boundary of the *C. ornithopsis/C. hoernesi* Zones. Latest results from the type area of the Vallesian (Krijgsman et al. 1996) suggest that the lower boundary of the Vallesian correlates with C5r1n or C5r2n. Our chart shows a slightly older age than C5r2n for this boundary (see discussion below).

Because the *L. conjungens* Zone in Vösendorf yields an Early Vallesian mammal fauna, and other Vallesian faunas were described from Austrian localities overlying the *L. conjungens* Zone (Götzendorf, Kohfidisch; Rögl et al. 1993; Rögl and Daxner-Höck 1996), and because the *L. decorum* Zone already yields Turolian faunas in Tihany and Sümegprága (Kordos 1987), we estimate that the Vallesian/Turolian boundary correlates with the upper part of the *L. ponticum* Zone, or with the lower part of the *L. decorum* Zone. This is more or less in agreement with the magnetostratigraphic age of the Vallesian/Turolian boundary in the type area of the Turolian (C4An; Garcés et al. 1996).

Discussion

Boundaries of the lacustrine sequence

Lower boundary

The boundary between the restricted marine Sarmatian and the Lake Pannon deposits is marked by a nearly complete faunal turnover: extinction of the marine organisms and appearance of the endemic lacustrine fauna. The hiatus between the two units appears significant in most sections. No reliable paleomagnetic record exists from this interval (a possible interpretation of the Berhida-3 magnetic record is discussed below); we must therefore rely on radio-isotopic ages of both the lacustrine and the underlying marine sediments in order to establish the numerical age of the lower boundary of the lacustrine sequence.

The oldest radio-isotopic age from a biostratigraphically well-constrained lacustrine sequence is from the Nagykozár borehole: 11.6 Ma at the boundary of the *S. b. pannonicus* and *S. b. oblongus* Zones (Sütő-Szentai, 1991a). The stratigraphic position of the 12.6±0.5 Ma old dacite tuff layer in the Berhida-3 borehole (Kókay et al. 1991) is debatable; it either underlies the Lake Pannon sequence, or is interlayered between the *S. b. pannonicus* and *S. b. oblongus* Zones. In the first case it gives a maximum age estimate for the base of the Lake Pannon deposits. In the second case, however, it conflicts with the Nagykozár radio-isotopic datum. Biostratigraphically-controlled Sarmatian radio-isotopic ages in the Pannonian basin range from 13.7 to 12.1 Ma (Vass et al. 1987; Pécskay et al. 1994). Younger radiometric ages originally assigned to the Sarmatian of the Pannonian basin lack biostratigraphic control. On this basis we mark the lower boundary of the Lake Pannon deposits at 12.0 Ma, but note that a few hundred thousand-year deviation from this figure is possible.

A more precise age determination of the base of the Lake Pannon sequence may be facilitated by a better understanding of the "*Hipparion*" datum. The concept that the sudden appearance of the American horse "*Hipparion*" in the Old World can be used as a stratigraphic tool has been discussed recently by Bernor et al. (1996) and Lindsay (1997). Efforts to determine the FAD of "*Hipparion*", marking the base of the Vallesian mammal stage, include high-resolution magnetostratigraphic studies and radio-isotopic measurements in different localities and basins around the Mediterranean. The results seem to converge at about 11 Ma. In the Lake Pannon sedimentary sequence the oldest "*Hipparion*" fossils were reported from the boundary between the *C. ornithopsis* and *C. hoernesi* Zones (Hovorany; Jiříček 1985a; Čtyroký 1987). The

FAD of "*Hipparion*" in the Bessarabian of the Eastern Paratethys correlates well with this datum. In our correlation chart (Fig. 4), however, this datum is considerably older than 11 Ma. If the synchronicity of the FAD of "*Hipparion*" were demonstrated, it would give us an excellent tool for the calibration of our basal Lake Pannon sequence. With the exception of the otherwise contradictory Berhida-3 record, the magnetostratigraphy of Lake Pannon deposits would allow such a re-interpretation. The radio-isotopic ages, however, would not; both in the Central and in the Eastern Paratethys they consistently indicate an age older than 11 Ma for the *C. hoernesi* Zone and for the Bessarabian, respectively. Correlations based on the assumption that the first appearance of "*Hipparion*" was synchronous in the Mediterranean and Paratethyan regions at about 11 Ma (Sen 1997) may be correct, but neglect the radio-isotopic data from the Paratethys.

Upper boundary

The upper boundary of the lacustrine section can be defined by the extinction of the brackish fauna and proliferation of a freshwater-fluvial fauna (Stevanovic, 1990a). Dating of this boundary would be possible in Serbia and Croatia only, because the Hungarian part of the basin dried up earlier. However, even the age of the youngest Lake Pannon beds in Hungary is still open to discussion, because no magnetic record covers the interval from C3An to C3n3 (6.6 Ma to 4.8 Ma), and even identification of C3An is uncertain. Seismic correlation of the 4.8 and presumed 6.6 Ma surfaces to southeastern Hungary, however, showed that the fossil-rich lacustrine Battonya sequence, belonging to the *Prosodacnomya*, *C. rhomboidea*, and *C. digitifera* Zones, falls into this gap. In fact, the 4.8 Ma time-line is less than 200 m above the last occurrence of brackish fossils (*Paradacna*, *Lymnocardium*) and there is no obvious sign of any significant hiatus between them. If this correlation is correct it suggests that the upper boundary of the lacustrine sequence is already within the Pliocene.

Remaining contradictions

The most outstanding controversy concerns the Kaskantyú-2 borehole. A large part of the *S. validus* Zone is still within C5n here, whereas the top of the underlying *S. paradoxus* Zone is younger than the top of C5n in other boreholes. Similarly, the lower boundary of the *C. praerhomboidea* Zone, as indicated by the appearance of *C. zagrabiensis*, and the lower boundary of the *C. digitifera* Zone are within C5n in this borehole, but above C5n in others (Fig. 3). A seismic horizon near the top of the long interval of normal polarity interpreted as C5n in Kaskantyú can be followed to the Kiskunhalas Ny-3 drill hole about 150 m above the basalt dated at 9.61 Ma (Horváth and Pogácsás 1988). These discrepancies suggest that either the biostratigraphic zones or the magneto-stratigraphic interpretation need reconsideration, but no reliable independent data are now known that would resolve these contradictions.

The correlation of the lower part of the Lake Pannon sequence in the Berhida-3 borehole also raises problems. The lower boundary of the *S. paradoxus* Zone and the upper boundary of the *S. b. oblongus* Zone lie in a reverse magnetic zone (?C5r), which contradicts data from other boreholes. According to the original interpretation (Kókay et al, 1991) the *S. b. oblongus* Zone, together with the *C. czjzeki* Zone, was identified throughout C5Ar, C5An, and C5r, right above a tuff dated 12.6±0.5 Ma. This five-cm dacite tuff layer overlies fossiliferous marine (Sarmatian) sediments and underlies fossiliferous Lake Pannon sediments. Both formations contain alternations of terrestrial and aquatic facies. As an alternative to the above correlation, the *S. b. oblongus* Zone (and the base of the *C. czjzeki* Zone) can be correlated only with C5r if one infers erosion or non-deposition in the terrestrial sequence after the tuff horizon was formed. Although lacking any sedimentological evidence, this inference can be justified by the lack of the *M. ultima* and *S. b. pannonicus* Zones in this sequence.

Implications for interbasinal correlation

Due to the endemic nature of Lake Pannon's fauna its direct correlation to the Mediterranean is difficult or even impossible. With the Eastern Paratethys, however, some correlation based on mollusks is possible, at least in the Bessarabian Substage and Pontian Stage. Mammal stratigraphy offers a rough correlation.

Bessarabian

The last significant compressional event in the evolution of the outer Carpathians resulted in thrusting of older units over the Volhynian (ca. 13.7 to 12.4 Ma; Chumakov et al. 1992), whereas the Chersonian (ca. 11.2 to 9.4 Ma; Chumakov et al. 1992) and even the upper part of the Bessarabian (ca. 12.4 to 11.2 Ma; Chumakov et al. 1992) overlie the nappes (Sandulescu 1988). This compressional event could have been the cause of the isolation of Lake Pannon from the Sarmatian sea; in this case the base of the Lake Pannon sedimentary sequence must be younger than the Volhynian and older than the Upper Bessarabian.

Based on biostratigraphic evidence the upper part of the Sarmatian in the Pannonian basin is considered coeval with the Lower Bessarabian in the Eastern Paratethys (Iljina et al. 1976; Bohn-Havas 1983). Radio-isotopic data suggesting a time span of 12.4 to 11.2 Ma for the Bessarabian strongly support this correlation (Andreescu et al. 1987; Vass et al. 1987; Chumakov et al. 1992). In addition, this correlation is in agreement with the observation that the Lower Bessarabian shows normal magnetic polarity, while the "*Hipparion*"-bearing (=Lower Vallesian) Upper Bessarabian shows reverse polarity (Pevzner and Vangengeim 1993).

Chersonian

Correlation of the Chersonian to the Central Paratethys by means of mollusks or other aquatic organisms does not seem possible. According to Pevzner and Vangengeim (1993), the mammal fauna of the Chersonian belongs to the Upper Vallesian. The paleomagnetic pattern of the Chersonian (Andreescu et al. 1987; Pevzner and Vangengeim 1993) together with its radiometric ages (Vass et al. 1987; Chumakov et al. 1992) suggests that this substage covers chronozone C5n, and some reverse magnetic intervals below and above it (11.2 to 9.4 Ma).

Maeotian

Evidence for the position of the Maeotian Stage is contradictory. The Lower Maeotian is mostly in a reverse polarity zone, although in several sections normal polarity intervals have been recorded as well, whereas the Upper Maeotian is of normal polarity (Andreescu et al. 1987; Trubikhin 1990; Pevzner and Vangengeim 1993; Krakhmalnaya et al. 1993). The radiometric ages assigned to this stage, however, range from 10.6 to 7.1 Ma (Andreescu et al. 1987; Vass et al. 1987; Chumakov et al. 1992). If we accept the correlation of the underlying Chersonian Stage as presented above, then the base of the Maeotian is about 9.4 Ma, and the upper boundary of this stage depends on how we define the lower boundary of the overlying Pontian Stage.

Pontian

The Pontian Stage of the Eastern Paratethys is characterized by a great influx of Lake Pannon mollusk species. This migration appears to have been unidirectional; no Eastern Paratethyan forms have been found in Lake Pannon deposits. Therefore, an overflow of the lake could be responsible for this migration (Müller and Magyar 1992a) and paleogeographic reconstructions of a huge, uniform Pontian basin including the entire Central Paratethys (Nevesskaja et al. 1987; Nevesskaja and Stevanović 1985) may not be correct.

In any case, correlation of the base of the Pontian with the Pannonian basin should be based on the youngest common mollusks (Müller and Magyar 1995). For instance, the appearance of *Prosodacnomya* in the Lower Pontian of the Dacian basin cannot be older than its first appearance in the Pannonian basin. The endemic evolution of this genus from pre-existing forms within the Pannonian basin has been documented, so we reject the possibility that it was an immigrant from the Dacian Basin. We correlate the Pontian of the Dacian basin with the *Prosodacnomya* Zone of the Pannonian basin because species level evolution in *Prosodacnomya* (involving at least three species) was parallel in the two basins during this time (Müller and Magyar 1992a).

This correlation is troublesome, however. According to the Iharosberény-I and Kaskantyú-2 magnetostratigraphic records and to the Tihany radiometric age, the base of the *Prosodacnomya* Zone is older than 8 Ma, whereas the base of the Pontian in the Eastern Paratethys is regarded as about 7 Ma (Chumakov et al. 1992). The Late Turolian age of the Pontian mammal faunas (e.g. Gabunia, 1990) also supports the latter date. We can think of three sources of error, each of which, or all together, can be responsible for this difference.

1. The Tihany radio-isotopic age is an average of measurements on 21 samples, and supposedly represents only the first stage of volcanic activity (Balogh 1995). Therefore, formation of the *Prosodacnomya*-bearing xenolith, and its incorporation into the volcanic rocks, may have taken place later. Similarly, magnetostratigraphic interpretations hardly have any reliable tie-points at this interval, thus these interpretations cannot be used as reference without reservation.

2. Late Maeotian (younger than 7.8 Ma) radiometric age determinations come only from Azerbaijan, from sequences lacking characteristic aquatic Maeotian fauna; their assignment to the Maeotian is based mostly on lithological correlation (Chumakov et al. 1992). This method of correlation may be a source of error.

3. Although the Pontian Stage is characterized by predominantly reverse polarity both in the Dacian (Andreescu et al. 1987; Krstić et al. 1995) and in the Euxinian basins (Pevzner 1987; Trubikhin 1990), the synchronous nature of the lower boundary of the Pontian between the Dacian, Euxinian, and Caspian basins may not be warranted (Vass et al. 1987).

Clearly, the correlation of the Pontian Stage into the Pannonian basin is still far from being solved. Further data will be required from all of the respective basins to settle this problem. The upper boundary of the Pontian Stage is even more vague than the lower boundary. The only radiometric date is from Azerbaijan (5.19±0.89 Ma; Chumakov et al. 1992), from a basin that was separated from the rest of the Paratethys by that time.

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Stratigraphy, paleoecology, and paleogeography of the "*Congeria ungulacaprae* beds" (*Lymnocardium ponticum* Zone) in NW Hungary: study of the Dáka outcrop

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The littoral "*Congeria ungulacaprae* beds" (=*Lymnocardium ponticum* Zone) of the Late Miocene Lake Pannon are widely exposed around the Transdanubian Central Range of NW Hungary, especially on its NW side. Study of the mollusk fauna of the Dáka outcrop, located near the classic locality of Kúp, showed that the fauna of this zone represents a transitional stage between the older *L. conjungens* Zone and the younger *L. decorum* Zone. The characteristic mollusk association of the *L. ponticum* Zone is apparently missing from the stratigraphic record of the southern part of the Pannonian basin. This difference can be partly explained by the paleogeographical situation. The northwestern part of the basin, where the Dáka fauna lived, was site of rapid delta progradation, whereas the southern shoreline lacked significant sediment input. However, the littoral, well-aerated, vegetation-rich environment represented by the Dáka layers must have had its counterpart in the southern part of the basin, where its sediments were either eroded or lie covered by younger sediments in a more basinward position. The age of the Dáka fauna and of the whole *L. ponticum* Zone is estimated between 9 and 10 Ma, on the basis of its correlation to magnetic polarity chron 4Ar in boreholes Berhida-3, Nagylózs-1, and Duka-II.

Key words: Neogene, Pannonian basin, Transdanubian Central Range, Lake Pannon, extinct lakes, biostratigraphy, paleogeography, paleoecology, Mollusca

*Introduction: the "***Congeria ungulacaprae** *beds"*

The term "*Congeria ungulacaprae* horizon" was coined by Halaváts (1903) to designate a local stratigraphic unit of the Late Miocene Lake Pannon sediments in the Lake Balaton region of Hungary. In this region the conspicuous shells of *C. ungulacaprae*, reworked and worn by Lake Balaton, had been associated with a fairytale and called "goat's hoofs" by the local people (Vitális 1910; Ager 1993). These shells were described and depicted as early as the 18th century (Géczy 1994), the first Lake Pannon fossils to be treated scientifically.

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Around the Transdanubian Central Range (TCR), especially on its NW side, *Congeria ungulacaprae* occurs very commonly in sandy, littoral facies with a surprisingly uniform accompanying mollusk fauna. Based on this fauna, Strausz (1942a) found that the "*C. ungulacaprae* horizon" was identifiable even in localities where *C. ungulacaprae* itself was not present. Dáka was one of these localities. A series of molluscan species, such as *Lymnocardium variocostatum, L. schreteri, L. priscae, L. ponticum, L. penslii* (early form), *Melanopsis kupensis, M. caryota* (early form) are found exclusively in the "*C. ungulacaprae* horizon". *C. ungulacaprae* itself, however, has a relatively wide stratigraphic range. Recognition of this distribution is reflected in later designations of this stratigraphic unit: "*Melanopsis pygmaea – Dreissena auricularis* Zone" (Korpás-Hódi 1983), "*Congeria ungulacaprae - Melanopsis pygmaea* Zone" (Müller and Magyar 1992), and "*Lymnocardium ponticum* Zone" (Magyar et al., this volume). For the sake of consistency, we will use the designation "*L. ponticum* Zone" in this paper.

The *L. ponticum* Zone has a peculiar feature: while most of the littoral communities of Lake Pannon were identical around the perimeter of the lake, the diagnostic fauna of the *L. ponticum* Zone seems to be confined to the northern shore (Fig. 1). What might have caused this restricted distribution? Are there indications that the facies, as well as the fauna, is unusual? How can we correlate this unit to other parts of the basin? The purpose of this study is, through biostratigraphic, paleogeographic, and paleoecological investigations, to answer the above questions. We base our work on investigation of the best available outcrop of the *L. ponticum* Zone at the village of Dáka, and compare our fossils with materials of the Geological Institute of Hungary, the Hungarian Natural History Museum, and the Geological Institute of Romania.

The Dáka outcrop

The most fossil-rich outcrop of the *L. ponticum* Zone is between the villages of Dáka and Kéttornyúlak, 8 km SW of Pápa. This outcrop is located only 5 km NW of the classic, but now inaccessible, Kúp locality (Fuchs 1870; Koch 1872; Fig. 1). The outcrop of a fossiliferous sand between Dáka and Kéttornyúlak was first mentioned by Horusitzky (1902). Later Strausz (1942a) collected fossils at the same site. The present-day outcrop, unfortunately partly refilled with waste, exposes five, prevailingly fine-grained, micaceous sand layers capped by a marl layer, in a total thickness of 5.5 m. The top of the sequence is truncated by a Pleistocene or Holocene erosional surface. The whole sequence belongs to the Somló Formation (Jámbor 1980; Fig. 2).

The lowermost layer (Layer 1) is moderately sorted limonitic sand with scattered, irregular accumulations of broken mollusk shells. Beside *Dreissena auricularis*, which is the dominant mollusk in each layer of the outcrop, valves of *Lymnocardium schreteri* were the most common bivalve remains. Among gastropods *"Pseudamnicola" margaritula* and *"Gyraulus" inornatus* were found

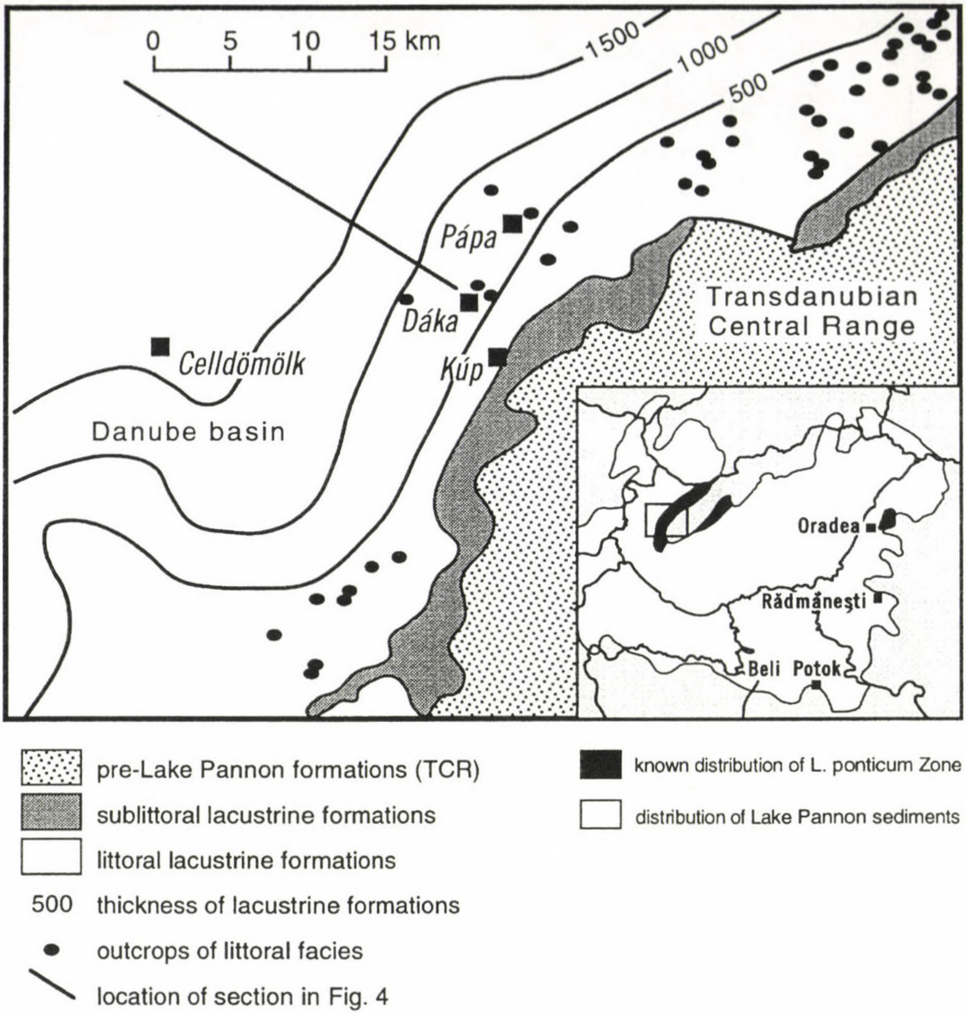

Fig. 1
Location of Dáka at the border of the Transdanubian Central Range (TCR) and the Danube basin. Geology of the region and thickness data are from Jámbor (1980); the fossiliferous localities of the littoral facies are from Strausz (1942a) and Strausz in Hegedűs (1953). The inset map shows the distribution of the *L. ponticum* Zone within the Pannonian Basin

most often. In Layer 2 the original horizontal lamination was disturbed by bioturbation. The fossil content of this layer is poor; only a few *Dreissena* specimens were found here. Layer 3, overlying Layer 2 with a sharp boundary, contains silt lenses, the upper surface of which is black from organic material, indicating the presence of vegetation in the sedimentary environment. A few cm-long root prints are also present in this layer. The most common mollusks are *Melanopsis caryota, Lymnocardium apertum,* and *L. schreteri.* The sand of Layer

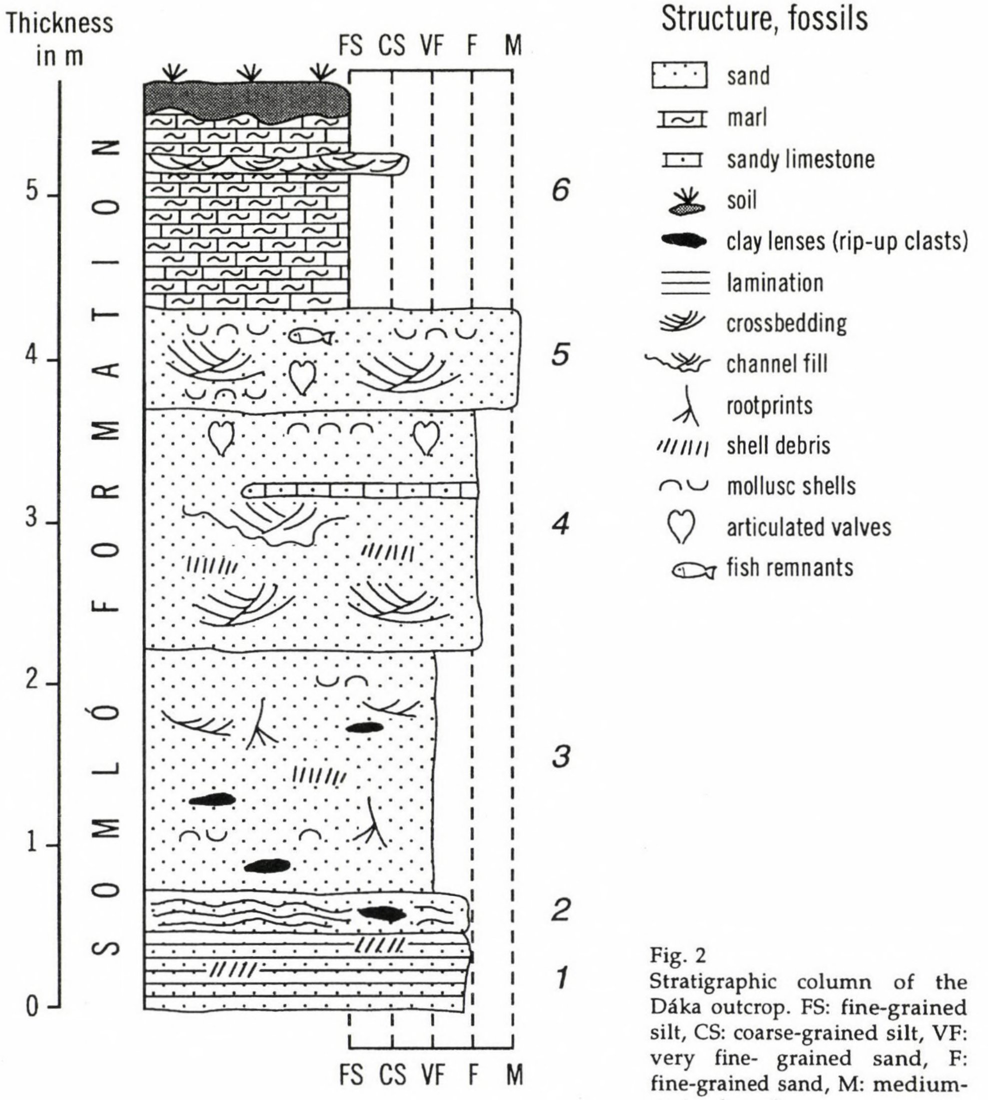

Fig. 2
Stratigraphic column of the Dáka outcrop. FS: fine-grained silt, CS: coarse-grained silt, VF: very fine- grained sand, F: fine-grained sand, M: medium-grained sand

4 is much better sorted and trough crossbedded. Mollusk shells are abundant in bedding planes and in lentiform accumulations and include broken *Lymnocardium* and relatively intact small gastropods (*"Pseudamnicola" margaritula, "Gyraulus" inornatus, Micromelania laevis, Valvata kupensis*). In the upper part of Layer 4 open but still articulated shells of *Unio mihanovici* and *Lymnocardium penslii* were often found. The disarticulated valves of *Lymnocardium,* however, are eroded and always lie convex side upwards, indicating the presence of strong currents or wave action.

Layer 5 is the most fossil-rich part of the outcrop. Mollusk shells accumulated in lenses 15 to 20 cm wide and 4 to 5 cm thick. However, many bivalves, such as *Unio mihanovici, Dreissenomya unioides,* and *"Paradacna" wurmbi,* are predominantly articulated, and even the most fragile shells fail to show evidence of significant transport or reworking. Characteristic bivalves are *Lymnocardium penslii, L. schreteri, Congeria* aff. *simulans turgida, Unio mihanovici, Dreissenomya unioides, "Paradacna" wurmbi.* Among gastropods, *Melanopsis pygmaea, "Gyraulus" inornatus,* and *Valvata kupensis* prevail. Otoliths were also found here. Layer 6 overlies the sand sequence with a sharp boundary. Its fossil content is rather poor: small (juvenile) *Dreissena* shells, ostracods, and small fragments of larger shells. The erosional surface of the Lake Pannon sequence is covered by soil.

The Dáka mollusks

612 specimens were identified in addition to the more than two thousand small (juvenile) *Dreissena* shells. Although the ratio of bivalve to gastropod species is 16:13, more than 80% of the total individuals were bivalves. The most common mollusk species are *Dreissena auricularis, Lymnocardium penslii, Melanopsis pygmaea, Melanopsis caryota, Lymnocardium schreteri, Valvata kupensis, "Pseudamnicola" margaritula,* and *"Gyraulus" inornatus.* The overall preservation of shell material was good. Many specimens were broken or eroded before burial but we also found bivalves (mostly *Unio, Dreissenomya,* and *"Paradacna"*) in life position.

Mollusk shells from the Dáka outcrop were included in a recent stable isotope study (Mátyás et al. 1996). The Dáka shells and some other samples from the *L. ponticum* and the coeval *L. soproniense* Zones displayed a significant negative anomaly of $\delta^{18}O$, as compared to the older lacustrine layers. Mátyás et al. (1996) propose that this anomaly resulted from a shift in climate towards a more humid, Atlantic type. This explanation is one of several possible; further investigations are required in order to better understand this phenomenon.

Stratigraphic significance

The Dáka outcrop yields the characteristic fauna of the *Lymnocardium ponticum* Zone. The transitional nature of this fauna between the underlying *L. conjungens* and the overlying *L. decorum* Zones is evidenced by several evolutionary lineages (Fig. 3). The early form of *Melanopsis caryota,* characteristic for the Dáka outcrop, has a morphology that is transitional between *M. impressa* and *M. cylindrica;* the former is characteristic for early deposits of the lake whereas the latter appears in the *Lymnocardium decorum* Zone (Staley 1992). The species *Lymnocardium ponticum* originates from *L. edlaueri* and is the direct ancestor of *L. decorum,* name-giving fossil of the overlying zone. The species *Lymnocardium variocostatum,* known exclusively from *the L. ponticum* Zone, is a

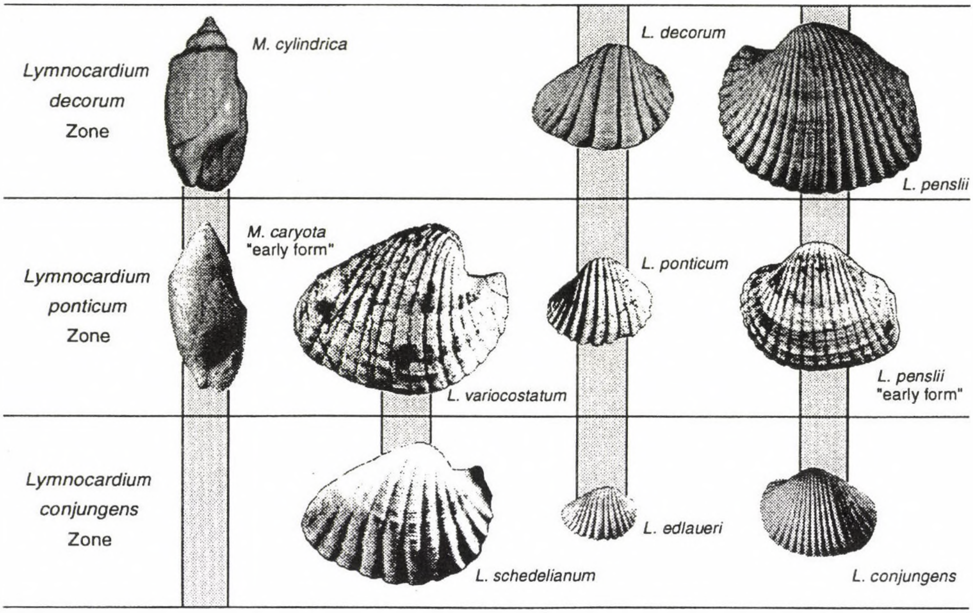

Fig. 3
Mollusks of the *L. ponticum* Zone show morphological transition between older and younger forms in several lineages

descendant of *L. schedelianum,* characteristic fossil of the *L. conjungens* Zone. The early form of *Lymnocardium penslii,* found in Dáka, originates from *L. conjungens,* name-giving fossil of the underlying zone. The late form of *L. penslii,* however, is common in the overlying *L. decorum* Zone.

Correlation of the NW Hungarian *L. ponticum* Zone with the classic locality of Rădmăneşti in the Romanian Banat (Fig. 1) seems obvious at first glance. In fact, by "*Congeria ungulacaprae* beds" Hungarian geologists have meant for decades the NW Hungarian formations whereas Yugoslav and Romanian geologists referred to formations with the Rădmăneşti fauna. Correlation of these two faunas is apparently confirmed by some rare common species described in this paper from the Dáka outcrop. The Rădmăneşti fauna is substantially richer than that of any locality in NW Hungary. Marinescu (1990) mentioned 121 taxa from this locality. In the collection of the University of Vienna we found two additional forms from the Rădmăneşti collections (*Lymnocardium dumicici* and *Melanopsis kupensis*). The unparallelled diversity of the Rădmăneşti fauna may have resulted from transport of shells from different environments. Considering the known evolutionary lineages, however, we conclude that the Rădmăneşti fauna contains many "late" forms without their ancestors; for example, *Melanopsis cylindrica* is present, but *M. caryota* is missing; *Lymnocardium decorum* is present, but *L. ponticum* is missing; only the late form of *L. penslii* is present, etc. The Radmanesti fauna thus should be correlated to the *L. decorum* Zone rather than the *L. ponticum* Zone.

No volcanic formation suitable for radioisotopic age measurement is known from the *L. ponticum* Zone. The age of this zone is assessed by magnetostratigraphic studies. According to the Duka, Nagylózs, and Berhida paleomagnetic polarity records, each obtained within a distance of 60 km from Dáka, the *L. ponticum* Zone correlates with polarity chronozone 4Ar, whose approximate age is 9 to 10 Ma (Magyar et al., this volume). (Identification of mollusks from the Nagylózs-1 and Duka-II boreholes was carried out by Korpás-Hódi; we used her determinations, but we changed her zonation).

Distribution of the **L. ponticum** *Zone within the basin*

Whereas the littoral communities of Lake Pannon were virtually identical around the perimeter of the lake, the mollusks of the *L. ponticum* Zone appear to be almost completely confined to the vicinity of the TCR. *Lymnocardium ponticum* itself was reported from the Tiszapalkonya borehole (see Magyar et al., this volume), but it occurs there without other characteristic forms of the zone. Apart from NW Hungary the only other area from which this mollusk association has been reported is the northwestern part of Munţii Apuseni, at Tataros (Brusturi) and Derna, NE of Oradea (Nagyvárad; Strausz 1942b; Fig. 1). As discussed above, the "*ungulacaprae* faunas" of Romania and Yugoslavia, including Rădmăneşti (Gillet and Marinescu, 1971) and Konopljište valley in Beli Potok (Pavlović 1928; Stevanović 1990a; Fig. 1), belong to the *L. decorum* Zone. The characteristic forms of the littoral *L. ponticum* Zone of NW Hungary seem to be totally missing in the southern and southeastern parts of the basin.

Regional geology

In the NW foreland of the TCR the deposits of Lake Pannon unconformably overly much older formations (Jámbor 1980). The older Lake Pannon biozones are missing here, and the whole lacustrine sequence belongs to the *Spiniferites paradoxus* Zone (Sütő-Szentai 1991). Further biostratigraphic subdivision of the sequence by mollusks has not been possible; only time-transgressive ecozones can be defined (Korpás-Hódi 1981, 1983).

The distribution of facies shows the following pattern: a zone of sublittoral sediments is situated directly next to the pre-Neogene TCR, and the littoral sediments are known to occur in a more basinward zone (Jámbor 1980; Fig. 1). This pattern clearly indicates that the present-day elevation of the TCR and consequent erosion of the Lake Pannon sediments from above the range is a relatively young feature. Seismic profiles show that the Danube basin was filled by progradation from the NW. Pronounced basinward dip of the originally horizontal depositional surfaces also indicates subsequent relative uplift of the TCR (Tari 1994; Fig. 4). Heavy minerals in the littoral lacustrine sand around the TCR derived mainly from metamorphic rocks, probably from the west and the north; evidence of sediment influx from the TCR is not apparent (Bartha 1963; Jámbor 1980).

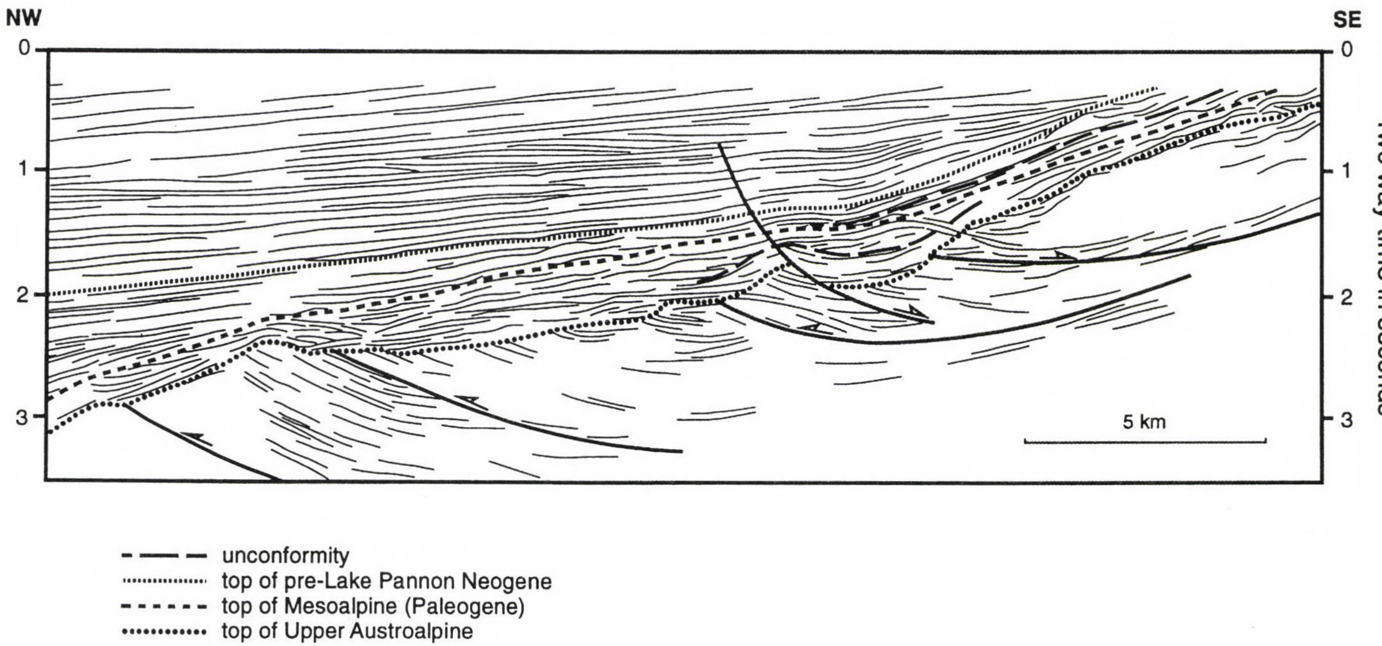

Fig. 4
Interpreted line drawing of a seismic reflection profile (Tari 1994). Location of the profile is indicated in Fig. 1. Note the subsequently tilted clinoforms of the progradational series in the lacustrine sequence. This profile, together with many others, suggests that the shoreline of Lake Pannon shifted from the NW towards the SE; when the littoral sand layers of Dáka were deposited, the offshore direction was to the SE, toward the TCR, just the opposite of what present-day morphology suggests

Discussion

Paleogeography

Based on the above stratigraphic and seismic data, we hypothesize that when the littoral sand layers of Dáka were deposited, the offshore direction was to the SE, towards the TCR, just the opposite of what present-day morphology suggests. Delta progradation from the NW was rapid in the shallow water above the TCR. The delta front shifted from at least the middle part of the Danube basin to as far as the southeastern edge of the TCR within the time interval of the *L. ponticum* Zone, and within magnetic polarity chron 4Ar (from 9 to 10 Ma). These delta deposits were later exposed and partly eroded as a consequence of the Pliocene or Pleistocene uplift of the TCR. This paleogeographic situation explains the uniformity of fauna in the NW foreland of the TCR; because the direction of progradation was more or less perpendicular to the range, the modern outcrops expose fairly synchronous deposits. Coeval deltas in the northeastern part of the Pannonian basin are probably buried under younger deposits, and they are exposed only in a few places north and east of Oradea.

Along the southern shore of the basin, sediment input and progradation were insignificant and thus the littoral deposits corresponding to the *L. ponticum* Zone probably developed only in a narrow geographical zone. Today, however, there is no trace of these formations in the southern part of the Pannonian basin. We can conceive of two possible explanations for this absence: either the *L. ponticum* Zone was deposited during lowstand, and the paleo-shoreline is buried north of the basin margin, or deposition of this zone was followed by widespread erosion, probably related to a water level drop, that destroyed the zone. Further investigations are required to validate either of these hypotheses.

Paleoecology of the Dáka mollusks

The *L. ponticum* Zone was deposited on top of the prograding deltas in NW Hungary. Sedimentology of the Dáka layers indicates a shallow lake to shoreface, vegetation-rich, well-aerated depositional environment. A strong freshwater influence is reflected by the presence of lush aquatic vegetation (as inferred from abundant herbivorous snails and root traces), as well as freshwater mollusk genera such as *Unio, Anodonta,* and diverse gastropods. A significant (-3%) $\delta^{18}O$ anomaly of the Dáka shells and other samples from the *L. ponticum* Zone (Mátyás et al. 1996) may also result from a strong fluvial influence. The shells were generally accumulated by currents or waves, but in situ burial (as evidenced by the preservation of articulated valves) is found in the deep burrower *Dreissenomya,* in some shallow burrowers like cardiids and unioniids, and also in *Dreissena.*

On the basis of the fauna of Layer 5, a distinct five-fold tiering of biotopes can be reconstructed (Fig. 5). The nominal subgenus of *Dreissenomya* comprised

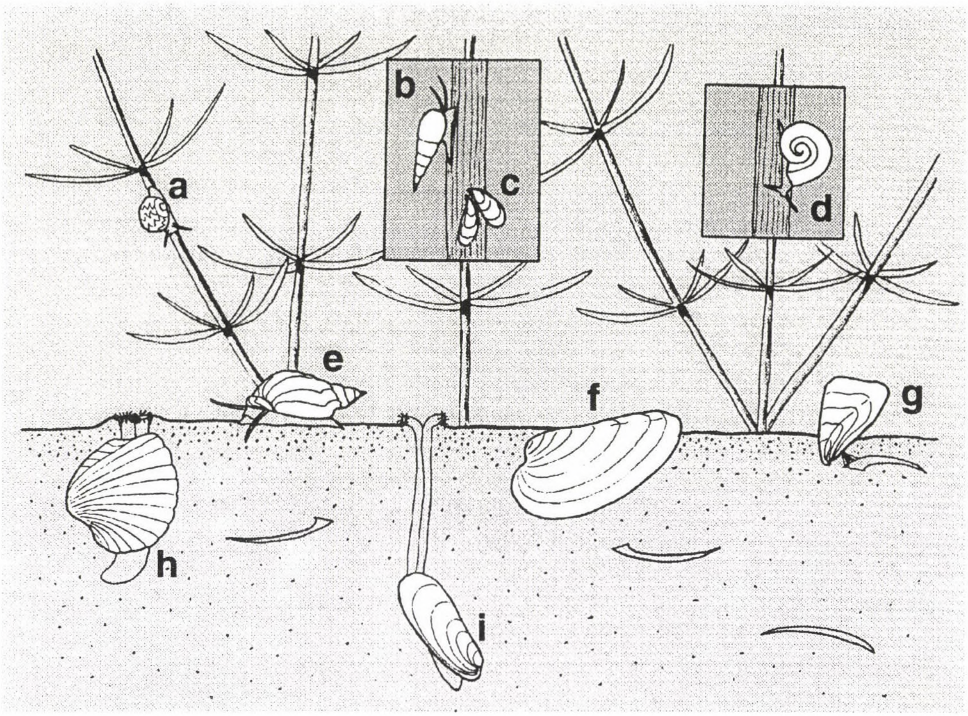

Fig. 5
Habitats of mollusks from Layer 5 of the Dáka outcrop. a: *Theodoxus*, b: *Micromelania*, c: *Dreissena*, d: *Valvata*, e: *Melanopsis*, f: *Unio*, g: *Congeria*, h: *Lymnocardium*, i: *Dreissenomya*

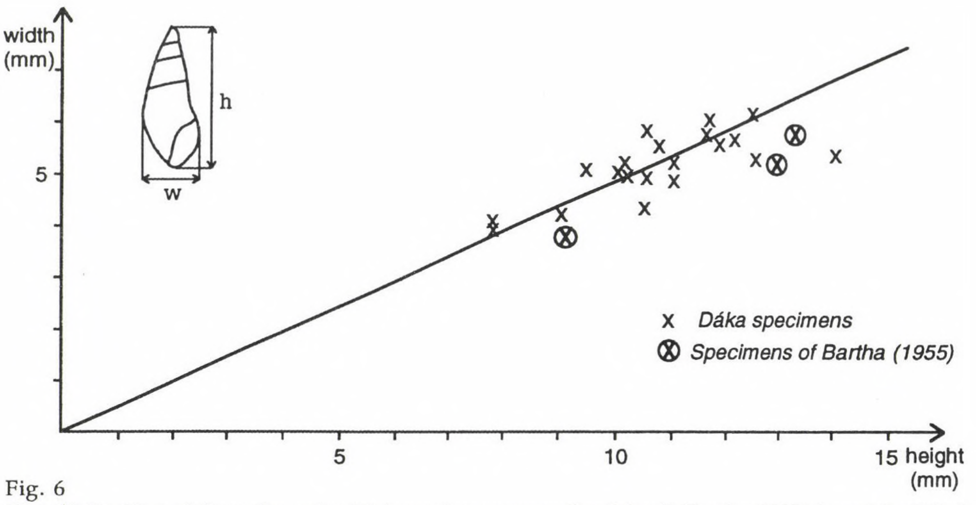

Fig. 6
The distinction of the subspecies *Melanopsis pygmaea subaudebardi* (Bartha 1955) is not justified by height and width data alone

deep burrowing forms that occupied nearly vertical life positions in the sediment, and communicated to the surface with long siphons (Marinescu, 1977). Cardiids were shallow burrowers; they usually stayed right below the sediment surface, and had short siphons. Half-sunk in the sediments lived the semi-infauna, like *Unio* and *Congeria*. *Congeria* attached to larger clasts or shells by its byssus. The epifauna included *Melanopsis*, moving on the surface of the substratum and feeding on organic particles and mud. The rooted vegetation offered food and dwelling for grazing gastropods like *Valvata*, *Theodoxus*, planorbids, and hydrobiids. The vegetation zone of lakes usually hosts an abundance of fishes; otoliths were also found in Layer 5.

The mass occurrence of juvenile *Dreissena* and the separation of juveniles from adults is characteristic of the entire Dáka outcrop. Such a pattern has also been observed in the littoral zone of modern Lake Ohrid, where adults live in the soft substratum of the sublittoral zone, whereas juveniles dwell in the littoral zone, mostly attached to the stems of the abundant *Chara* vegetation. Juveniles may be as abundant as 12,000 individuals/m^2. The main zone of the *Chara* vegetation is between 6 and 20 m water depth (Stanković 1960; Salemaa 1994). The Ohrid example serves as a model for not only the Dáka outcrop, where root traces have been found, but for many other localities of the *L. ponticum* Zone as well. Mass occurrences of young *Dreissena* have been found in other outcrops of the region (e.g. Pápa-Bárócz Hill; Strausz 1942a and collections of the Geological Institute of Hungary), and *Chara* oogonia are sometimes preserved in the layers of the *L. ponticum* Zone (Strausz 1942a).

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Appendix

Mollusks of the Dáka outcrop

Bivalvia:

Anodonta sp. indet.
Unio mihanovici Brusina (Plate I, Figs 1, 2)
Congeria aff. *simulans turgida* Andrusov (Plate I, Figs 3–5)
Dreissena auricularis (Fuchs) (Plate I, Figs 6, 7)
Dreissena bipartita (Andrusov)
Dreissenomya unioides (Fuchs) (Plate I, Figs 8, 9)
Lymnocardium variocostatum (Vitális) (Plate 2I, Figs 1–5)
Lymnocardium hantkeni (Fuchs) (Plate II, Fig. 6)
Lymnocardium ponticum (Halaváts) (Plate III, Fig. 4)
Lymnocardium penslii (Fuchs 1870) (Plate IV, Figs 1–7)
Lymnocardium apertum (Münster) (Plate IV, Figs 8–11)
Lymnocardium (Euxinicardium) schreteri (Strausz) (Plate III, Figs 5, 6)
"Monodacna" viennensis Papp
"Paradacna" wurmbi (Lőrenthey) (Plate III, Figs 1–3)
Parvidacna sp.
Caladacna steindachneri (Brusina) (Plate IV, Fig. 14)

Gastropoda:

Theodoxus intracarpaticus Jekelius (Plate IV, Figs 12, 13)
Valvata kupensis Fuchs (Plate V, Figs 8–13)
Valvata minima Fuchs
Micromelania laevis (Fuchs) (Plate V, Fig. 7)
Prososthenia radmanesti (Fuchs) (Plate V, Fig. 6)
"Pseudamnicola" margaritula (Fuchs) (Plate V, Figs 18, 19)
"Pyrgula" incisa Fuchs
Melanopsis caryota (Brusina) (Plate V, Figs 1, 2)
Melanopsis pygmaea Hörnes
Melanopsis sturii Fuchs (Plate V, Fig. 3)
"Gyraulus" inornatus (Brusina) (Plate V, Figs 14–16)
"Gyraulus" aff. *tenuis* (Fuchs) (Plate V, Fig. 17)
Radix sp. (Plate V, Figs 4, 5)

Species descriptions and remarks

Congeria aff. *simulans turgida* Andrusov 1897

(Plate I, Figs 3–5)

pars 1897 *Congeria turgida* Brusina	– Andrusov, p. 111, pl. 3, fig. 8 (excl. figs 1–7)
1942a *Congeria batuti* Brusina	– Strausz, p. 75.

Description. Shell outline is highly variable, triangular to rounded, up to 35 mm in length. Beak always inclines anteriorly. Anterior part of the ventral field around the byssal opening is concave, giving the whole shell an elongated, bill-like appearance. Dorsal keel is not sharp.

Remarks. In the first description of *C. turgida*, Andrusov (1897) included a specimen from Neszmély that corresponds to the Dáka form. Strausz (1942a) identified the *Congeria* specimens of the Dáka outcrop as *C. batuti*, remarking that distinction of this species from *C. turgida* is ambiguous. We compared our specimens to the type locality (Radmanesti) material of the Hungarian Natural History Museum. Accepting the subspecies-level distinction of *C. simulans batuti* and *C. simulans turgida* by Gillet & Marinescu (1971), we conclude that the Dáka form is more similar to *C. simulans turgida*. The Dáka specimens, however, are larger than the specimens of Gillet & Marinescu (1971), and never have the sharp dorsal keel common in Radmanesti specimens.

Lymnocardium (Euxinicardium) schreteri (Strausz) 1942

(Plate III, Figs 5, 6)

1942a *Limnocardium Schreteri* nov. sp.– Strausz, p. 67, pl .1, fig. 21, 26–27
1943 *Limnocardium trifkovici* Brus. – Gillet, p. 52
1971 *Limnocardium* aff. *L. trifkovici* Brusina – Gillet et Marinescu, p. 13, pl. 3, fig. 1–4
1971 *Limnocardium (Euxinicardium) subsyrmiense* Andrusov – Gillet et Marinescu, p. 17, pl. 5, figs 13–21
1971 *Limnocardium (Euxinicardium)* cf. *nobile* Sabba – Gillet et Marinescu, p. 18, pl. 5, fig. 25
1971 *Limnocardium (Euxinicardium)* aff. *L. inlongaevum* Eberzin – Gillet et Marinescu, p. 18, pl. 5, figs 26–30

Description (based on 30 broken or juvenile specimens, and many shell fragments). Shell conspicuously thin. Anterior and ventral margins rounded, posterior margin straight. Average length of valves 7 mm, average height 5 mm, maximum height 20 mm. Shell moderately inflated, slightly prosogyrous. Umbones hardly coiled. Posterior gape very narrow. 16 to 18 tightly spaced, narrow, high, rounded or somewhat sharp radial ribs. Middle ribs symmetrical, anterior ribs steeper anteriorly, posterior ribs steeper posteriorly. Intercostal space flat, narrower than ribs. Posterior part of shell has 4 to 6 sharp folds or riblets. Margin strongly serrate, shell internally deeply furrowed. Pallial line not visible. Hinge teeth:

$$\frac{\text{AI}\quad \text{AIII}\quad (3\text{a})\quad 3\text{b}\quad \text{PI}}{\text{AII}\qquad\qquad 2\quad (4\text{b})}$$

Anterior lateral tooth of left valve (AII) usually well developed, posterior (PII) missing.

Remarks. Strausz (1942a) described one left valve from Somlójenő; later he found this species in Tüskevár as well (Strausz in Hegedűs, 1953). The Dáka material closely resembles the Rădmăneşti form. We also found this species in Szák, Lázi, and Stegersbach.

Lymnocardium hantkeni (Fuchs) 1870

(Plate II, Fig. 6)

1870 *Cardium hantkeni* Fuchs – Fuchs, p. 546, pl. XXII, figs 29–31
1942a *Limnocardium hantkeni* Fuchs – Strausz, p. 70, pl. 1., figs 3, 4
1980 *Limnocardium hantkeni* (Fuchs) – Lueger, p. 113, pl. 4, fig. 6

Description (based on a right valve and a juvenile left valve). Valve is flat, almost equilateral, elongated, anterior and posterior margins rounded, ventral margin straight. Length 13 mm, height 8.5 mm. Beak central, not coiled. Posterior gape narrow. 18 flat, radial ribs, rectangular in cross-section. At the ventral margin the ribs are 1.5 to 2 times wider than the intercostal space. Posterior part of valve has 7 narrow, thread-like ribs. Shell internally furrowed, muscle scars and pallial line not visible. The juvenile specimen is 5 mm long, and has 17 ribs in the anterior field. Hinge teeth:

$$\frac{\text{AI}\quad 3\text{a}\quad 3\text{b}\quad \text{PI}}{\text{AII}\quad 2}$$

Remarks. Our adult specimen differs from the type of this species in having a less inflated umbo, in not displaying a deep sinus, and in having the ribs visible in the inner surface of the shell almost to the beak. *L. hantkeni* is known from the western foreland of the TCR at Kúp, Szák, Pápa, Jánosháza, and Tüskevár (Strausz, 1942a) from the *L. ponticum* Zone, and from Eisenstadt (Lueger 1980) in the *L. conjungens* Zone.

"Paradacna" wurmbi (Lőrenthey) 1893

(Plate III, Figs 1–3)

1893 *Limnocardium Wurmbi*, nov. form. – Lőrenthey, p. 132, pl. 3, fig.7
1895 *Limnocardium Wurmbii* Lőrenthey – Lőrenthey, p. 318, pl.. 8, figs 11, 12
1943 *Monodacna wurmbi* Lőr. – Gillet, p. 81, pl. 6, fig. 8
?1951 *Paradacna retovskii* Andrus. var. *ossoinae* n. var.– Stevanović, p. 266, 342, pl. 8, fig. 7
?1961 *Paradacna retowskii ossoinae* Stev. – Stevanović, p. 194, pl. 6, figs 4–8; pl. 8, fig. 2; pl. 9, figs 1–3
1971 *"Monodacna" wurmbi* (Lőrenthey) – Gillet & Marinescu, p. 28, pl. 9, figs 10–16
?1978 *Paradacna retowskii ossoinae* Stev. – Stevanović, pl. 9, figs 2–4
?1990 *Paradacna retowskii ossoinae* Stevanovič – Basch, p. 89, 150, pl.. 20, figs 16, 17
?1990b *Paradacna retovskii osoinae* Stevanović – Stevanović, p. 494, pl. 11, figs 1, 3–11

Description (based on 30 broken shells, most of them with two valves, and many fragments). Shell oval, very thin. Length of adults more than 30 mm. Upper part of posterior margin straight ("truncated"). Juveniles poorly, adults moderately inflated. Position of beaks median to slightly anterior to mid-length. Umbones not coiled. No posterior ridge present. Siphonal gape wide. 15 to 21 radial ribs; shell surface smooth in front of and behind ornamented part. Anterior 9 to 10 ribs narrow, rounded in cross section; intercostal spaces usually narrower than ribs. More posterior ribs angular, often wedge-shaped in cross-section, wedge pointing towards the anterior margin; intercostal spaces can be 1.5 times wider than ribs. Posterior 2 to 3 ribs flat and wide. Inner surface of shell not known. Hinge edentulous.

Remarks. The wedge-shaped ribs must have facilitated anchoring the shell within the sediment. The Dáka form completely corresponds to the Rădmăneşti material. The only difference is the "*Monodacna*-type hinge" mentioned by Gillet and Marinescu (1971): "a small, oblique cardinal tooth below the beak, and a deep socket". This pattern could not be observed in the Dáka specimens, or in the referred picture of Gillet and Marinescu (1971).

The Dáka form slightly differs from the species *"Lymnocardium" wurmbi*, described by Lőrenthey (1893) from much younger layers of Árpád, in being highly variable in many features, such as shell outline, position of the beak, and the number, width, and shape of the ribs. Lőrenthey (1893) emphasized the low variability in his material. However, we think that the Dáka-Rădmăneşti form and the younger *"Lymnocardium" wurmbi* (Lőrenthey) (?=*"Paradacna" retovskii ossoinae* Stevanovič) belong to the same species.

Melanopsis caryota (Brusina) 1902, early form

(Plate V, Figs 1, 2)

1902 *Lyrcaea caryota* Brus. – Brusina, pl. 5, figs 21–25
1942a *Melanopsis impressa* Kr. – Strausz, p. 84
1963 *Melanopsis bonelli bonelli* Manz. – Bartha, pl. 3, fig. 1
1963 *Melanopsis impressa* Krauss – Bartha, pl. 3, fig. 2
1992 *Melanopsis* aff. *impressa* Krauss – Magyar, p. 300, pl. 4, figs 9–11
1992 *Melanopsis caryota* Brusina, "early form" – Staley, p. 73, pl. 1, figs 1–2

Description (according to Staley 1992). Spire high, conical in shape along its entire length, pointed. Exterior surface smooth with rare growth lines on body whorl. Pronounced angular inflection in the outer lip at mid-height of the aperture which produces a ridge in the middle of the whorl as the shell grows, giving the body whorl a diamond shape.

Remarks. This form is usually well distinguishable from *M. impressa*, and shows a gradual morphological change towards more shouldered forms of *M. caryota* and eventually *M. cylindrica*. This early form is restricted to the *L. ponticum* Zone, where it is widely distributed and abundant.

Melanopsis pygmaea M. Hörnes 1856

1856 *Melanopsis pygmaea* – M. Hörnes, p. 599, pl. 49, fig. 13
1870 *Melanopsis pygmaea* Partsch – Fuchs, p. 545, pl. 22, figs 7–14
1942a *Melanopsis pygmaea* Partsch – Strausz, p. 88, pl. 5, figs 25–28
1983 *Melanopsis pygmaea* Partsch – Korpás-Hódi, p. 163, pl. 10, figs 2, 3
1983 *Melanopsis decollata* Stol. – Korpás-Hódi, p. 163, pl. 10, figs 4, 5

Description (based on 29 specimens). Typically 6 whorls visible; only 4 individuals have 7 whorls and 2 individuals have 5. Body whorl accounts for approximately half shell height. Height to width ratio is 1.91 to 2.78, with a mean of 2.22. Larger specimens apparently slimmer than smaller ones. Original color patterns (rectangular orange spots) preserved in seven specimens.

Remarks. Beside *Dreissena auricularis*, *M. pygmaea* is the most common form of the *L. ponticum* Zone (Strausz 1942a, with figures 25–27 from the Dáka outcrop; Korpás-Hódi, 1983). In the Dáka outcrop it was especially common in Layer 5. We found some specimens that are very similar to the subspecies *M. pygmaea subaudebardi* Soós. This subspecies is characterized by 7 to 9 whorls, and a high, very pointy spire. The height is 9.2 to 13.1 mm, the width is 3.5 to 4.9 mm, so the height to width ratio changes from 2.62 to 2.67. According to Bartha (1955), the Várpalota specimens of this subspecies were 13.5 mm high and 5.5 mm wide, having a height to width ratio of 2.45. We measured our Dáka specimens in order to determine if the slim shells within the sample form a well-distinguishable group (Fig. 6). The height to width ratio was higher than 2.4 in four specimens; however, they do not seem to form a separate group, rather represent the extremity of variation within the sample. Our conclusion is that the use of new species or subspecies names, based on the extreme forms of the population, is not justified.

Plate I

1, 2. *Unio mihanovici* Brusina
3–5. *Congeria* aff. *simulans turgida* Andrusov
6, 7. *Dreissena auricularis* (Fuchs). Juvenile specimens.
8, 9. *Dreissenomya unioides* (Fuchs)

Scale bars indicate 5 mm.

Plate II

1–5. *Lymnocardium variocostatum* (Vitális). Specimen shown in Figs 1 and 2 is from the Dáka collection of the Geological Institute of Hungary. Figure 5 shows a juvenile specimen.
6. *Lymnocardium hantkeni* (Fuchs)

Scale bars indicate 10 mm.

Plate III

1–3. *"Paradacna" wurmbi* (Lőrenthey)
4. *Lymnocardium ponticum* (Halaváts)
5, 6. *Lymnocardium (Euxinicardium) schreteri* (Strausz)

Scale bars indicate 5 mm.

Plate IV

1–7. *Lymnocardium penslii* (Fuchs)
8–11. *Lymnocardium apertum* (Münster)
12,
13. *Theodoxus intracarpaticus* Jekelius
14. *Caladacna steindachneri* (Brusina)

Scale bars indicate 5 mm.

Plate V

1, 2. *Melanopsis caryota* (Brusina), early form
3. *Melanopsis sturii* Fuchs
4, 5. *Radix* sp.
6. *Prososthenia radmanesti* (Fuchs)
7. *Micromelania laevis* (Fuchs)
8–13. *Valvata kupensis* Fuchs
14–16. *"Gyraulus" inornatus* (Brusina)
17. *"Gyraulus"* aff. *tenuis* (Fuchs)
18, 19. *"Pseudamnicola" margaritula* (Fuchs). In Fig. 18, the outer lip of the aperture is broken.

Scale bars indicate 2 mm.

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Plate I

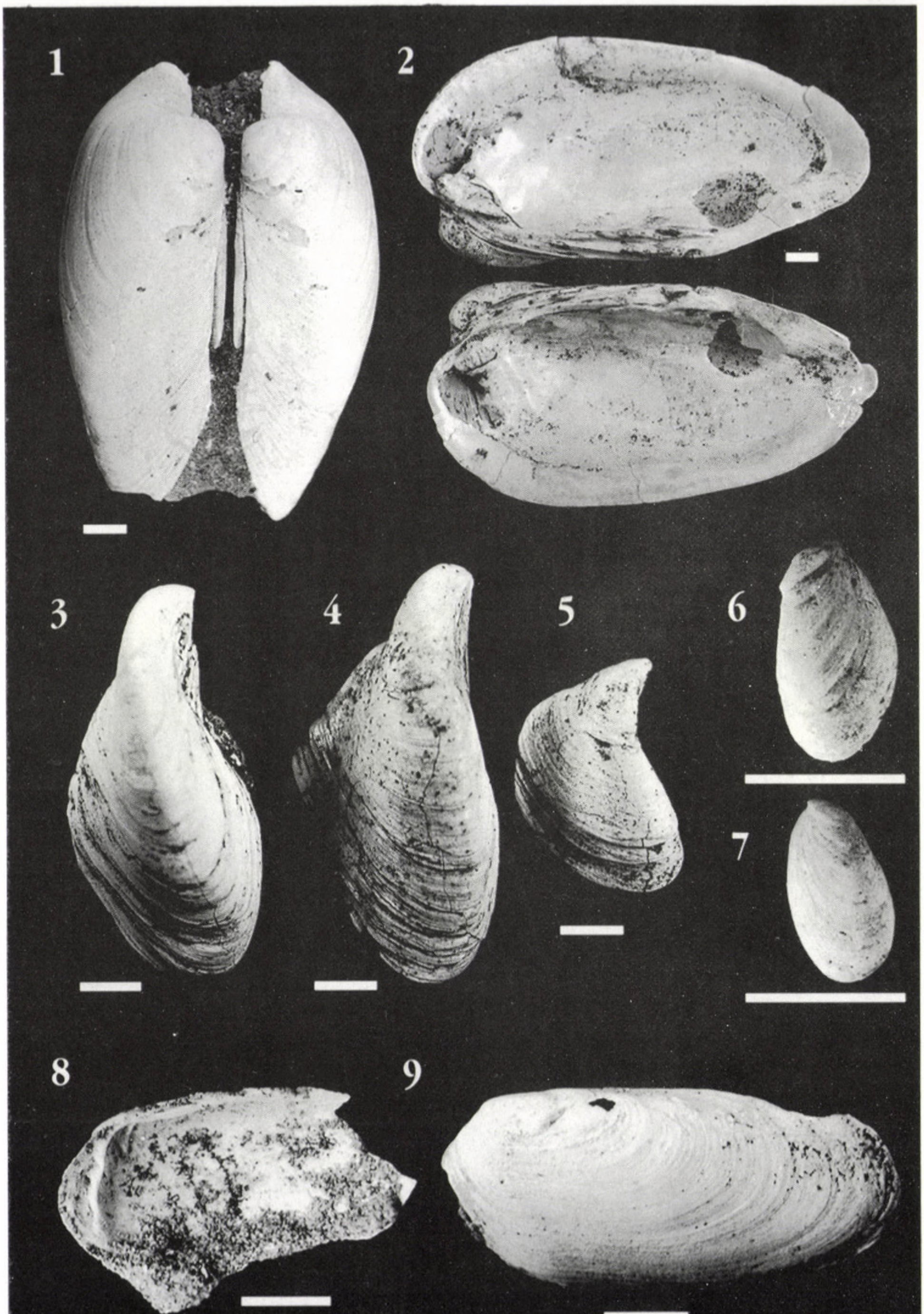


Plate II

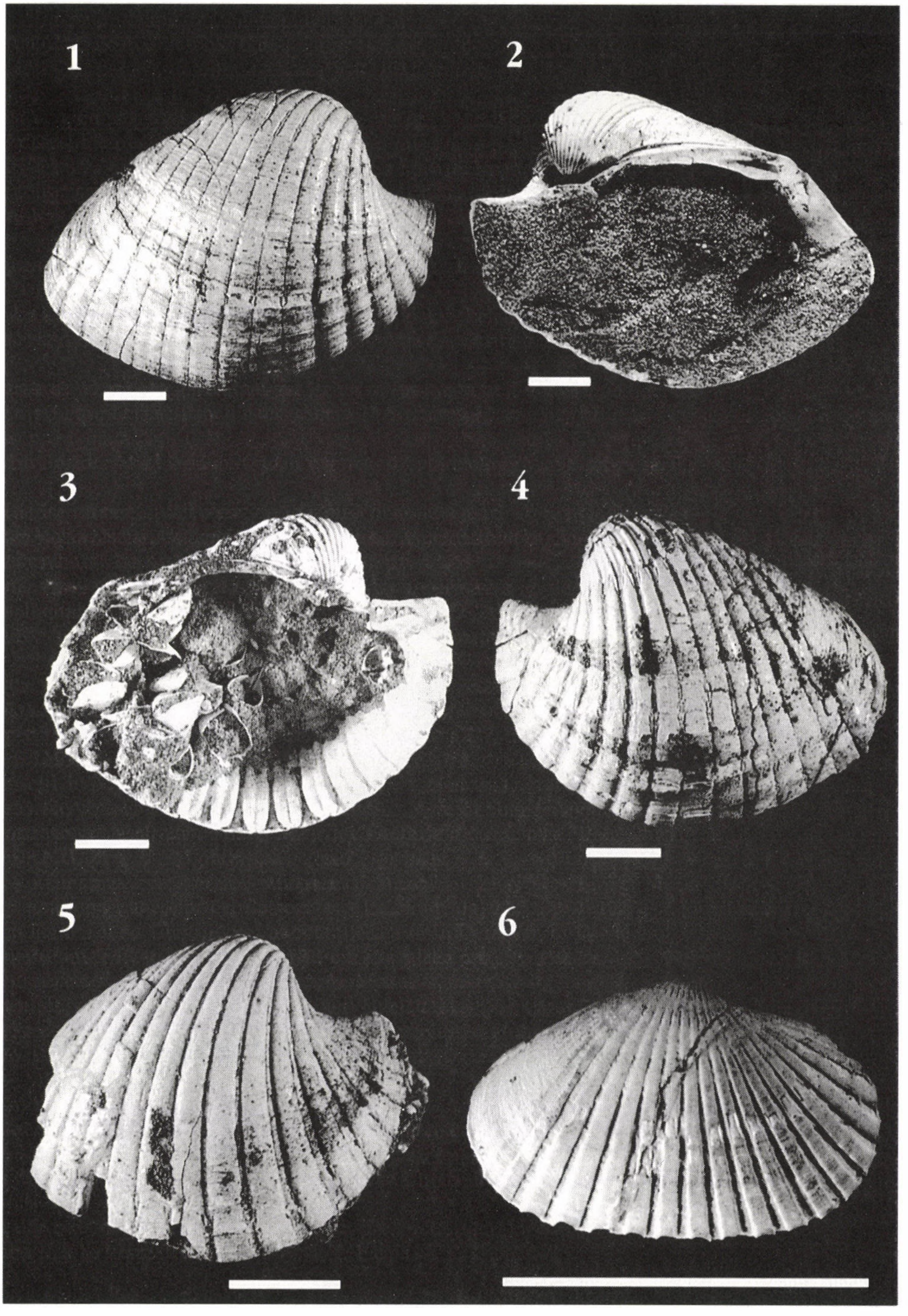


Plate III

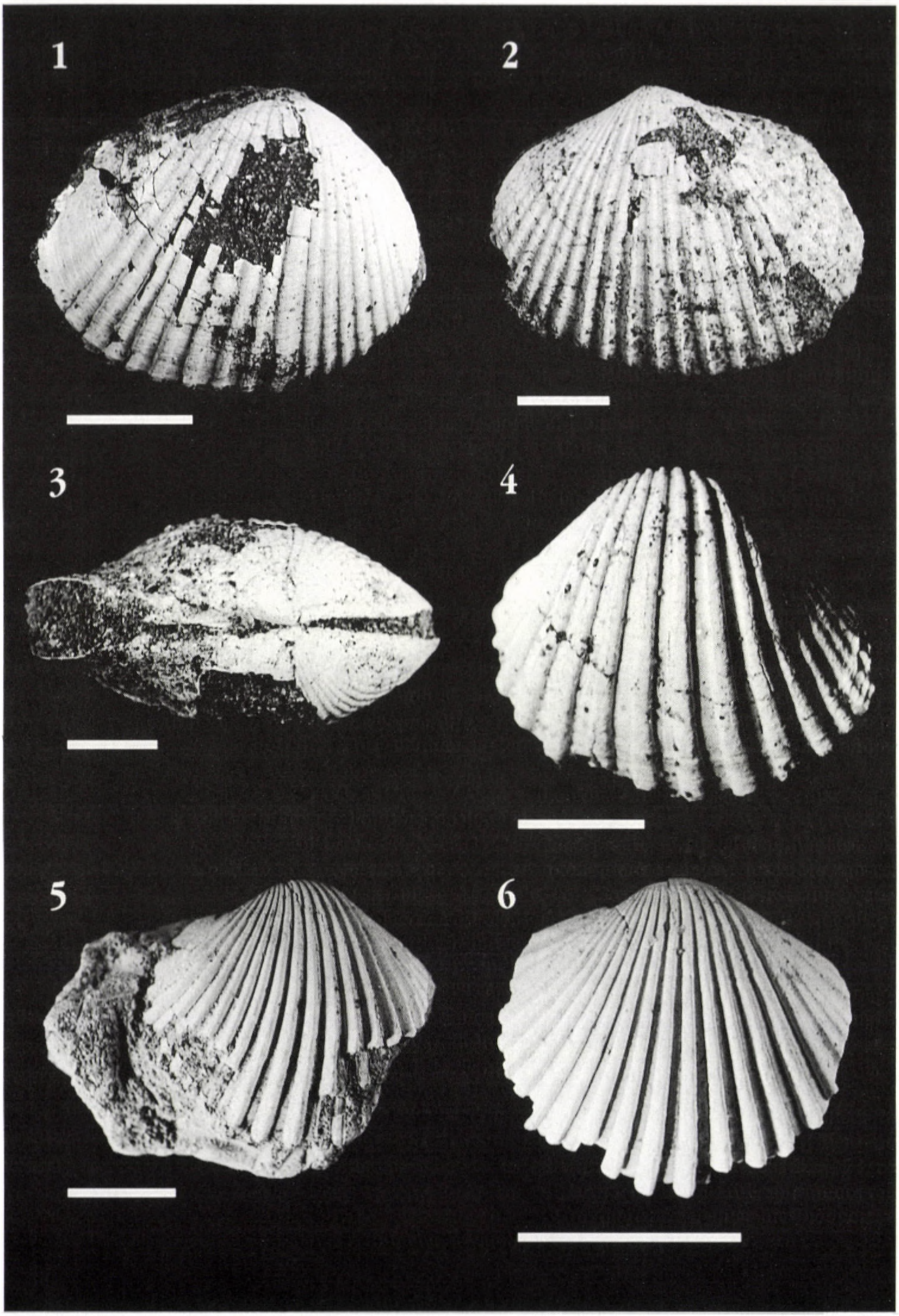


Plate IV

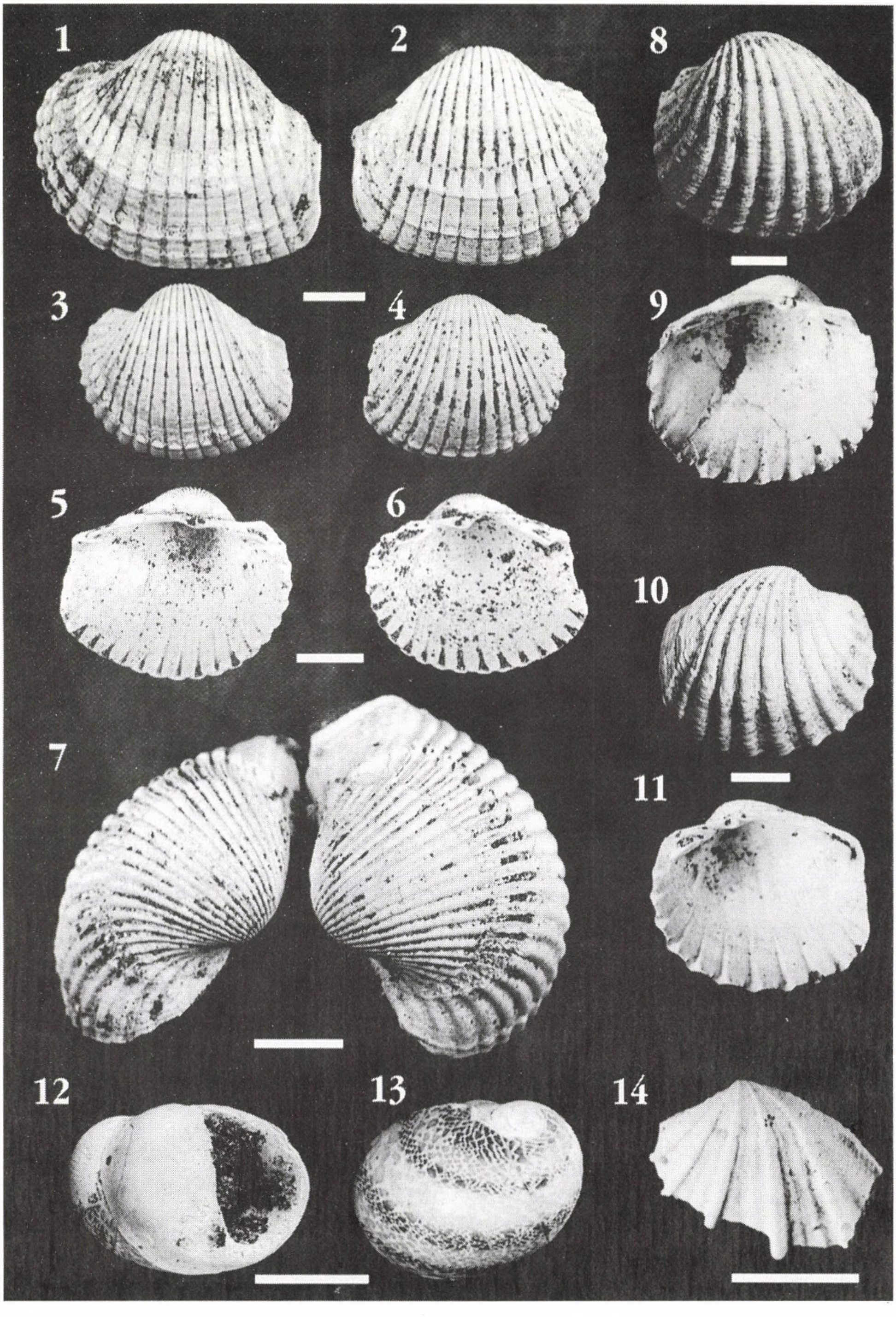
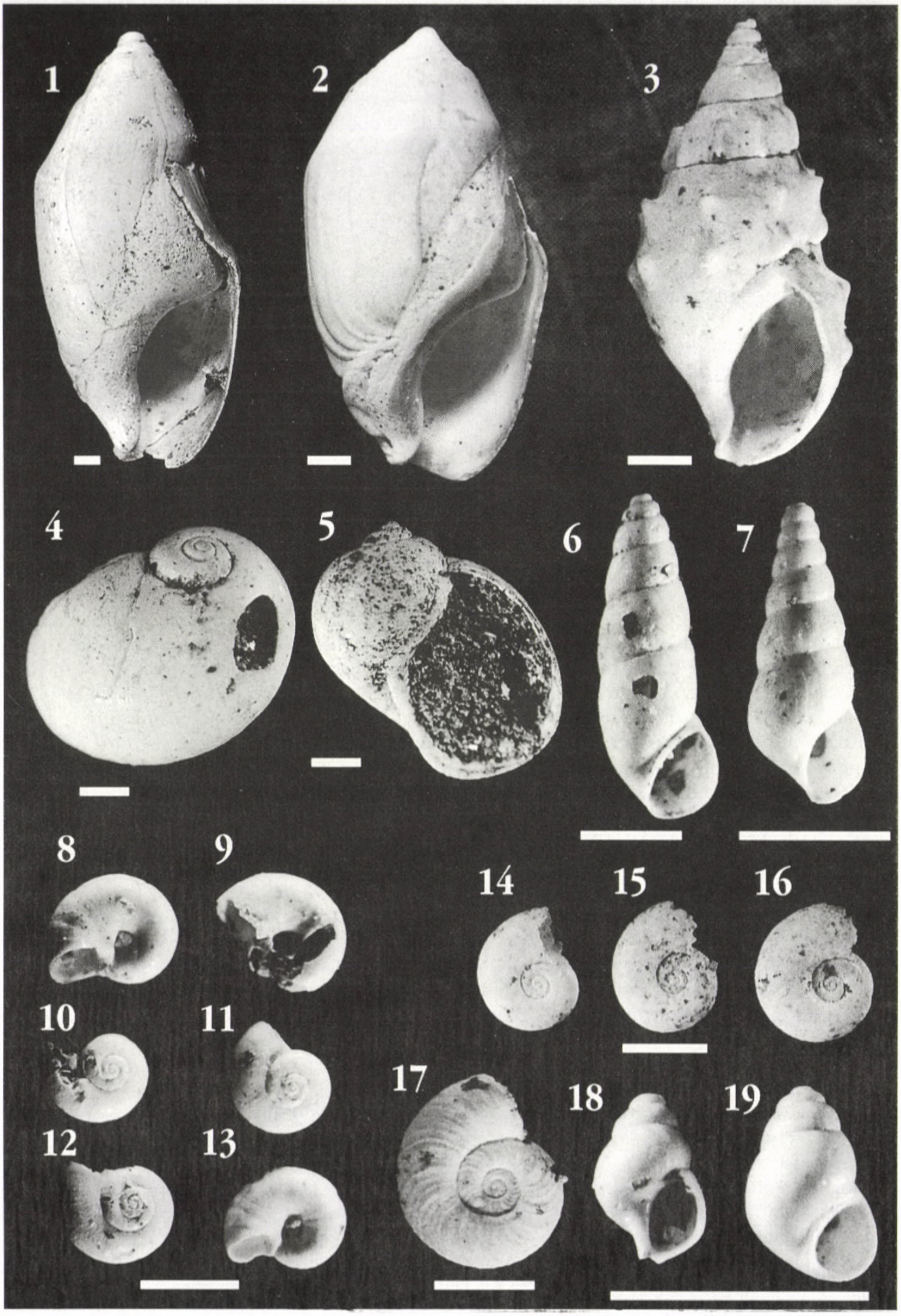


Plate V



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Mass occurrence of *Congeria balatonica* in the *C. rhomboidea* Zone of southern Hungary: mollusk fauna of the Hird sand pit (Mecsek Hills)

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Within the Late Miocene Lake Pannon mollusk fauna from the Hird sand pit, *Congeria balatonica* Partsch makes up nearly 50% of the specimens collected. The cardiids of the outcrop (especially *Lymnocardium schmidti* [Hörnes]), however, indicate that these layers are younger than the typical "*balatonica* beds" (*Lymnocardium decorum* Zone) in Tihany; they can be correlated to the *Congeria rhomboidea* Zone. The facies of the embedding rock (medium to coarse-grained sand) reflects a shallow-water, wave-dominated depositional environment. This differs from the generally silty sublittoral sediments of the *C. rhomboidea* Zone and from the coeval (but prevailingly lagoonal) deposits of the *Prosodacnomya* Zone. Mass occurrence of *Congeria balatonica* from the *C. rhomboidea* Zone has not been described in Hungary before. The sand layers also contain two rare cardiid species: *Lymnocardium kochi* (Lőrenthey), a close relative or equivalent of *L. banaticum bilogorense* Basch and *L. dumicici* (Gorjanovic-Kramberger), never before reported from Hungary.

Key words: biostratigraphy, lacustrine features, Lake Pannon, Mecsek, Mollusca, Neogene, paleoecology, Pannonian Basin, Paratethys

Introduction

Deposits of the Upper Miocene Lake Pannon and their fossils around the Mecsek hills in southern Hungary have been known since the middle of the last century. Eight new molluscan species were described by Hörnes (1859–1867) from the classic localities (Árpád, Hidas).

At the village of Hird, located 6 km east of Pécs, lacustrine layers outcrop in a sand pit situated several hundred meters south of the Budapest–Pécs highway (Fig. 1). Mollusks from this sand pit were reported by Kleb (1973; partly based on determinations of Sümeghy in Ferenczi [1937]): *Congeria triangularis, C. scharpei, Dreissena polymorpha, Lymnocardium schmidti,* and *L. apertum*. After extensive collecting from this outcrop, we have identified 15 mollusk taxa; only one species is also present in Kleb's list. In spite of the mass occurrence of *Congeria balatonica*, cardiids indicate that these layers are correlative with the *Congeria rhomboidea* Zone, which indicates that they are

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younger than the typical "*balatonica* beds" (*Lymnocardium decorum* Zone) in Tihany-Fehérpart (Müller and Szónoky 1990).

Description of the outcrop

The Hird sand pit exposes the limonitic sand zone of Lake Pannon that surrounds the Mecsek hills. The nearest outcrops of this formation are Kulcsos and Pécs-Danitzpuszta 2 and 3 km to the west, respectively, and Csokoládépuszta and Pécsvárad 5 and 8 km to the east, respectively (Fig. 1), but the sand zone can be traced well beyond these localities in both directions.

The layers in the sand pit dip slightly to the S–SE. Kleb (1973) published a geological section of the outcrop, but the appearance of the sand pit has changed considerably since then. By 1989 the lower part of Kleb's section had been filled in by debris, while its upper layers were still available for collecting, and new layers became exposed as well. The old and the new outcrops together expose a section of about 30 m.

The sedimentary sequence consists of yellowish brown, yellow, or dark brown, limonitic, well-sorted, medium or coarse-grained sand with coarser, pebbly intercalations. Limonite appears both as crust on the sand grains and as cement. The small pebbles include quartz, quartzite, Jurassic limestone, and sandstone. The sand grains are moderately rounded. According to Kleb (1973) the sandy sequence contained thin layers or lenses of lignite, but they have all been removed by now. Cross-bedding indicates a shallow-water, wave-dominated depositional environment. The lacustrine sequence is unconformably overlain by a thin layer of Pleistocene clay and gravel.

The fossils

Mollusks are common in the entire section, but they are especially abundant in limonitic concretions in four layers, the thickness of which varies between 1 and 4 m. The composition of the fauna in these layers was similar. 70% of the total of 790 specimens were internal molds, 30% were external molds. Internal and external molds preserved muscle scars and ornamentation, respectively. Cavities from dissolved shells were empty and were not affected by significant compaction. We have found only one non-limonitic layer where dissolution of shell material was not complete. Specimens from this horizon include many small and juvenile forms. Because of poor preservation and the unconsolidated nature of the embedding rock, their determination was problematic.

The fossil assemblage is dominated by large, thick-shelled specimens of *Congeria* (54%) and *Lymnocardium* (45%). The low frequency of small forms is probably due to the relatively extreme depositional environment: a high-energy, wave-dominated regime with sandy or pebbly substratum. The agitated water detached the valves of the pelecypods (only 12% of specimens are articulated,

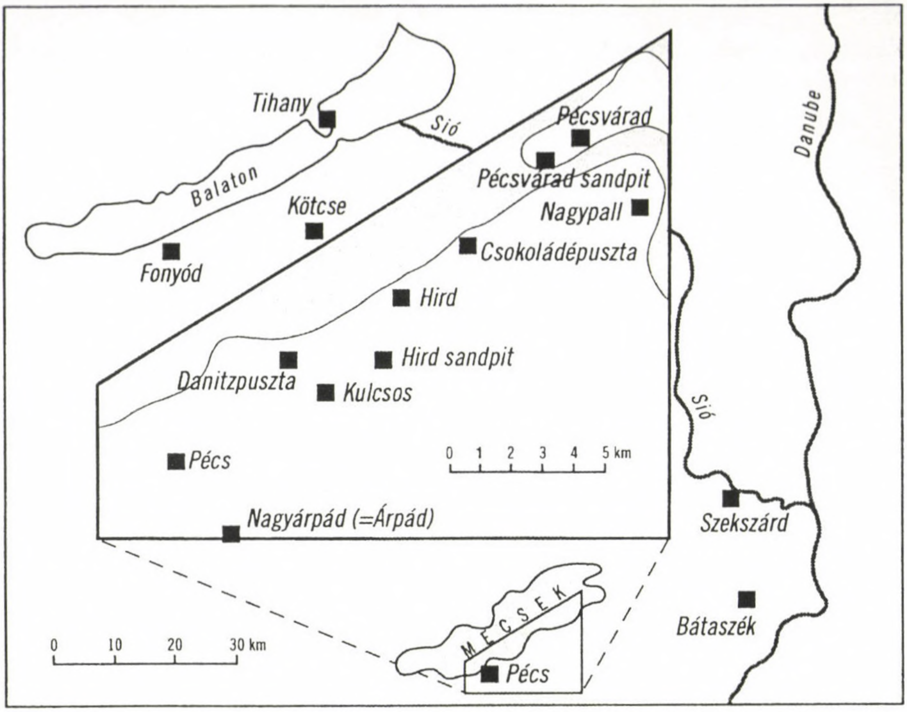

Fig. 1
Locations referred to in the text. On the inset map, white area shows the distribution of Lake Pannon sediments, whereas shaded area indicates older (prevailingly Mesozoic) formations

mostly *Congeria*), and washed together shells from different environments. The few specimens of *Viviparus*, for example, were certainly transported from a more quiet, freshwater habitat. The shells were buried by rapid sedimentation, as indicated by fining-upward gradation observed within a number of internal molds. The identified mollusk taxa are shown in Table 1.

A curiosity of the Hird fauna is the occurrence of two rare species, *Lymnocardium dumicici* (Gorjanović-Kramberger 1899) and *L. kochi* (Lőrenthey 1894). Sümeghy (1939) remarked that the species "*Limnoc. Dumici* Gorj. Kramb." (sic!) occurs in the fauna of "the Lower Pannonian layers at Pécs, Nagypall, and Mecsekszabolcs". Later Strausz (1953) published a photo of a fragment of *L. dumicici* from the vicinity of Árpád, but he identified it as a giant specimen of *L. banaticum*. Hird was the first Hungarian locality with an authentic report of this species (Mező 1989). It was more recently found in Kötcse (Müller and Magyar 1992), Fonyód, and in core samples from the vicinity of Bátaszék.

Lymnocardium kochi was described by Lőrenthey (1894) from Szekszárd, on the other side of the Mecsek hills. Unfortunately the type material has been lost but this rare species was easily identifiable on the basis of Lőrenthey's

Table 1
Mollusk taxa determined from the Hird outcrop. Because it was difficult to distinguish the internal molds of *L. schmidti* and *L. dumicici*, some of the fossils assigned to *L. schmidti* might actually belong to *L. dumicici*

Species	No. of specimen	% of total
Congeria balatonica Partsch	382	48.4
Congeria markovici Brusina	1	0.1
Congeria sp. indet	44	5.6
Dreissenomya cf. *aperta* (Deshayes)	1	0.1
Lymnocardium schmidti (M. Hörnes)	185	23.4
Lymnocardium cf. *banaticum* (Fuchs)	17	2.2
Lymnocardium dumicici (Gorjanovic-Kramberger)	15	1.9
Lymnocardium majeri (M. Hörnes)	9	1.1
Lymnocardium kochi (Lőrenthey)	1	0.1
Lymnocardium cf. *riegeli* (M. Hörnes)	1	0.1
Lymnocardium hungaricum (M. Hörnes)	4	0.5
Lymnocardium sp. indet	123	15.6
Phyllocardium complanatum (Fuchs)	1	0.1
Viviparus sp. indet	3	0.4
Gastropoda sp. indet	3	0.4
TOTAL	790	100

description and figures. A probable synonym of *L. kochi* is *L. banaticum bilogorense,* described by Basch (1990).

Apart from Lake Pannon mollusks the outcrop also yielded fish vertebrae, crushing teeth, and dolphin bones. These fossils, reworked from older, marine Miocene sediments, occur in the nearby Danitzpuszta outcrop as well, where they also include shark and whale remains (Kordos 1992).

Paleoecology and biostratigraphy

Congeria balatonica, a common species in many outcrops around Lake Balaton in Hungary, was first described by Partsch (1835) from Tihany. Agglomerations of *Congeria balatonica* shells often paved the silty or sandy substratum of Lake Pannon littoral waters (Müller and Szónoky 1990). The accompanying fauna usually indicates relatively reduced salinity (low diversity of cardiids, high number of freshwater snails). In the Hird outcrop, however, *Congeria balatonica* was found together with a variety of cardiids that are mostly known from sublittoral silts. In this case they were probably transported by waves from the sublittoral environment and mixed with littoral specimens of *Congeria*. This assumption is corroborated by the observation that the infaunal cardiid shells are almost always disarticulated (allochthonous), whereas the epifaunal

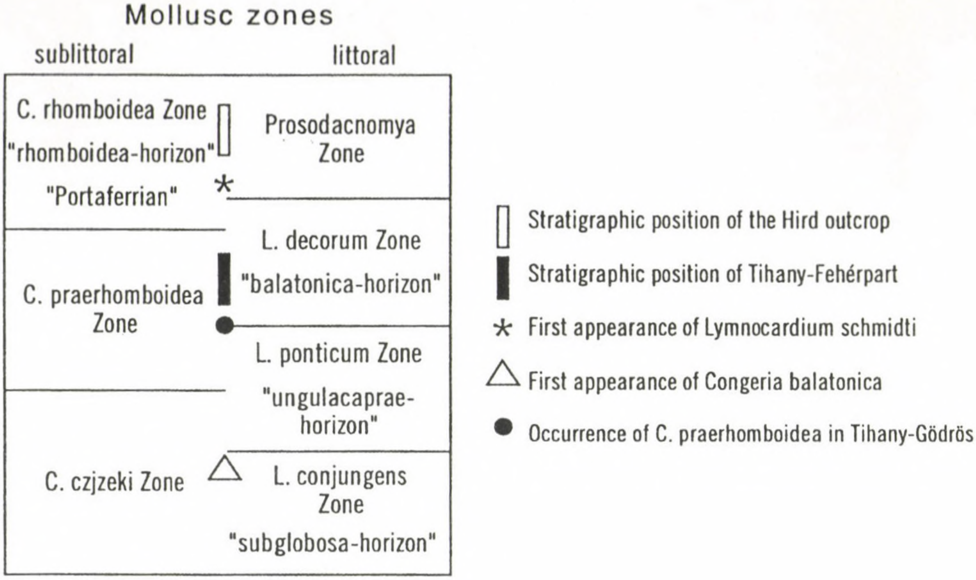

Fig. 2
Biostratigraphic positions of outcrops and events discussed in the text. Stratigraphic units ("horizons") of Halaváts (1903) and Lőrenthey (1906) are in quotation marks as well as the Portaferrian "Substage" of Stevanovic (1951). The correlation of the sublittoral and littoral zones is from Magyar et al., this volume

Congeria shells are sometimes articulated (semi-autochthonous). A noteworthy exception amongst cardiids might have been *L. dumicici.* In its Hungarian localities this species was often found together with *C. balatonica* and *Unio* sp.; we therefore conclude that it was part of the littoral *Congeria balatonica* community.

The biostratigraphic evaluation of the Hird fauna suggests that it is significantly younger than the above-mentioned Lake Balaton localities, including Tihany. In order to clarify our arguments on this issue we will begin with a short review of how the understanding of the biostratigraphic role of *C. balatonica* has developed. Due to its exceptional abundance, *C. balatonica* was one of the first "index fossils" for Lake Pannon deposits. The first Hungarian stratigraphers used large *Congeria* species for biostratigraphy; they distinguished the "*balatonica* horizon", or "*balatonica* beds", as a unit older than the "*rhomboidea* horizon" but younger than the "*ungulacaprae* horizon" and the "*subglobosa*-horizon" (Halaváts 1892, 1903; Lőrenthey, 1906; Fig. 2). Later, Papp (1951) reported the occurrence of *C. balatonica* together with *C. subglobosa* from the Vienna basin, suggesting that the first appearance of *C. balatonica* is much older than had previously been thought. In addition, scattered and uncertain reports on the occurrences of *C. balatonica* together with *C. rhomboidea* (Sümeghy 1939; Strausz 1953) and the study of the accompanying fauna of these two species led to the view that their stratigraphic ranges are the same, or that at least they largely overlap (Stevanović 1951; Strausz 1953; Korpás-Hódi 1987), or, in an extreme view, that the fauna with *C. balatonica* is younger than that with *C. rhomboidea* (Bartha 1966). The old *Congeria* index fossil system seemed to have failed.

With the understanding of morphological evolution within several bivalve lineages (Stevanović 1978; Müller and Magyar 1992), however, these stratigraphical problems can be re-addressed. Tihany and most of the outcrops around Lake Balaton belong to the *Lymnocardium decorum* Zone; the first appearance of *Lymnocardium schmidti* is already in the *Prosodacnomya* Zone (Müller and Magyar 1992). The first appearance of *C. rhomboidea* is much later than that of *C. balatonica* and correlates with the uppermost part of the *L. decorum* Zone (=*L. serbicum* Subzone) in Orešac-2. *L. penslii* showing a morphological transition towards *L. schmidti* was also reported from Orešac-2 (Stevanović 1951). In addition, Strausz (1953) reported the occurrence of *Congeria praerhomboidea* in Tihany-Gödrös (originally reported as "*C. rhomboidea* var."; revision of this form is in Stevanović 1961), immediately below the well-known Fehérpart sequence. On this basis the classic locality of the *balatonica* beds in Tihany (*Lymnocardium decorum* Zone) seems to correlate with the *Congeria praerhomboidea* Zone, whereas the Hird fauna is younger and can be correlated with the *C. rhomboidea* Zone and the *Prosodacnomya* Zone (Fig. 2).

Mass occurrence of *C. balatonica* either together with *C. rhomboidea* or in formations undoubtedly coeval with *C. rhomboidea*-bearing layers has not been noted before in Hungary, nor did Stevanovic (1951) mention any such case from Yugoslavia. The significance of the fossil assemblage of the Hird outcrop is that it demonstrates that *Congeria balatonica* pavements formed even during the deposition of the *Congeria rhomboidea* Zone.

Acknowledgements

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Plate I

1. *Congeria balatonica.* Internal (upper right) and external molds. To the left, the closed valves are in life position, showing the wide byssus opening
2. *Lymnocardium schmidti;* internal mold
3. *Viviparus* sp., internal mold
4. *Congeria balatonica.* Internal mold with the byssus opening
5. *Lymnocardium dumicici.* The rubber cast (at right) taken from an external mold (at left) shows the original (somewhat worn) surface of the shell with the characteristic narrow, flat ribs.
6. *Congeria balatonica.* The sediment is fining-upwards within this internal mold (the original position was upside-down), indicating that the shell was buried rapidly, by a single event.
7. *Lymnocardium hungaricum.* The characteristic sharp ribs are well visible in the external mold (at bottom).

Horizontal bars indicate 1 cm

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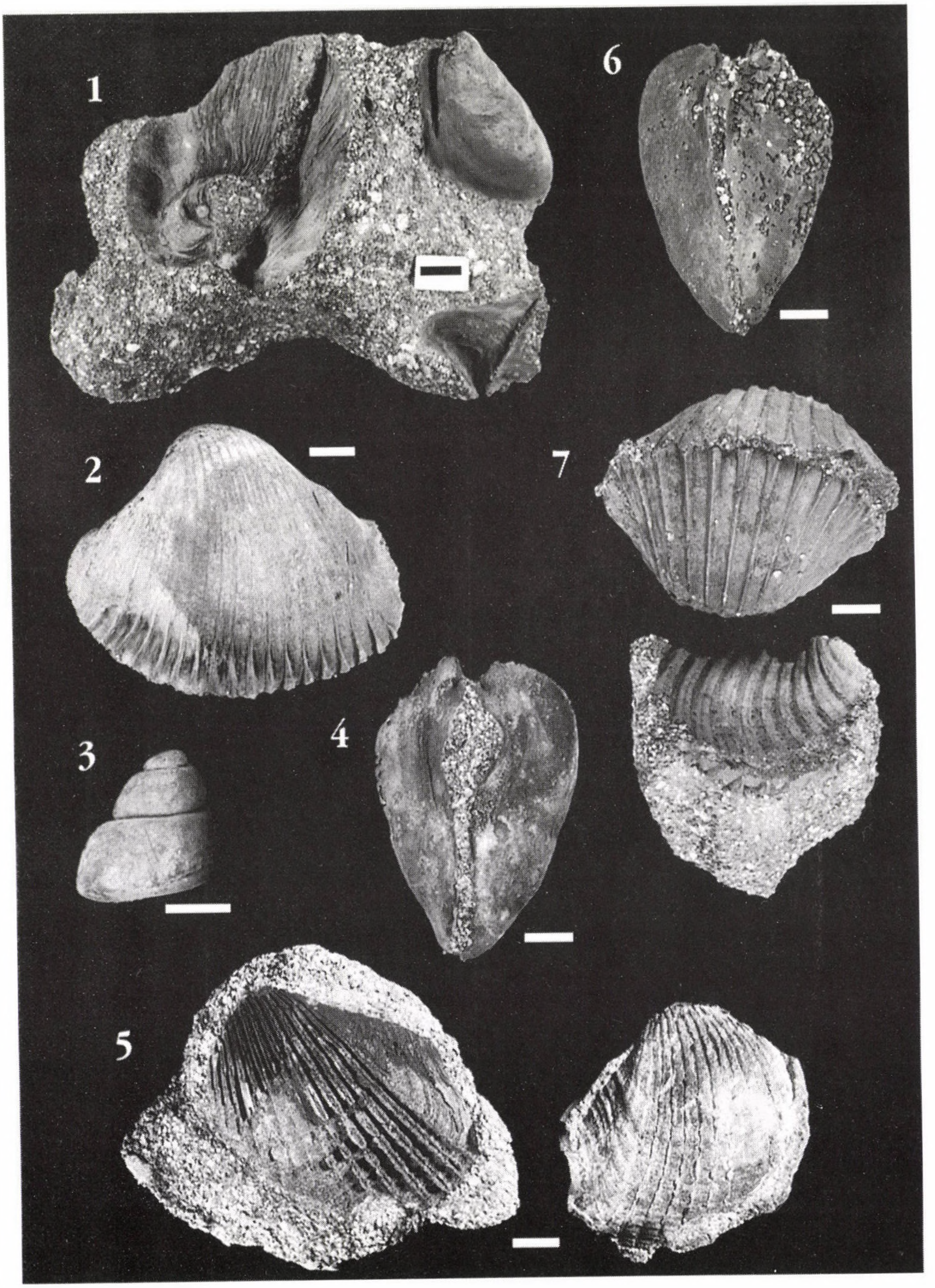
Plate I



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The Lake Pannon fossils of the Bátaszék brickyard

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The outcrop of the Bátaszék brickyard in southeastern Transdanubia of Hungary exposes clayey silt in a 20 to 30 m thickness, containing several thin, sandy, often limonitic, poorly sorted coquina layers. We interpret the silt as sublittoral deposit of the Late Miocene Lake Pannon. The sandy intercalations are probably storm deposits or sediment gravity flows from a nearby littoral environment. The mollusk fauna of the silt is largely autochthonous, and dominated by *Congeria rhomboidea* and *Lymnocardium hungaricum*. The partly reworked faunas of the sandy intercalations have a different composition. The entire fauna is very similar to the classic faunas of Okrugljak and Szekszárd, and to the fauna of Jazovnik (faciostratotype of the Portaferrian Substage). All of these deposits belong to the *Congeria rhomboidea* Zone. Other fossils from the outcrop include ostracods, fishes, algae, spores, and pollen. The dinoflagellates indicate that the Bátaszék deposits are in a slightly higher stratigraphic position than the top of the *Spiniferites validus* Zone.

Key words: biostratigraphy, lacustrine features, Lake Pannon, Mecsek, Neogene, paleoecology, Pannonian Basin, Paratethys

Introduction

The sublittoral clayey silt deposits of the Late Miocene Lake Pannon have been widely used by brick factories in western Hungary (Transdanubia). The Bátaszék brickyard in southeastern Transdanubia opened in 1973, and silt has been excavated there continuously since then. The silt and the intercalated thin, sandy layers have yielded a rich mollusk fauna consisting of 51 species,

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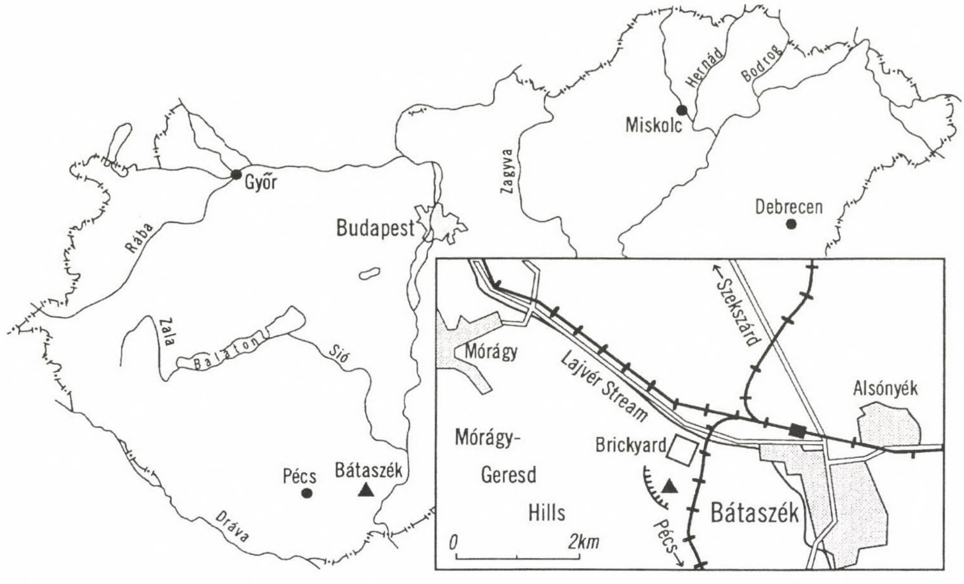

Fig. 1
Location of the outcrop of the Bátaszék brickyard (black triangle) on the eastern slope of the Mórágy–Geresd Hills, SE Transdanubia, Hungary

numerically dominated by *Congeria rhomboidea* and *Lymnocardium hungaricum*. Collection and evaluation of fossils and description of the sedimentary sequence were performed in 1983–85 and documented by Lennert as his M.S. Thesis at the József Attila University of Szeged (1986). Other fossil groups were more recently included in the paleontological investigation, and further collecting of mollusks completed the list of species known from this locality. In this paper we describe the sedimentary sequence and fossils of the outcrop, with particular emphasis on mollusks and their paleoecological patterns.

Description of the outcrop

The clay pit of the Bátaszék brickyard is situated at the eastern margin of the Mórágy–Geresd Hills in southeastern Transdanubia (Fig. 1). These hills consist of Lower Carboniferous granite. Below the transgressive Lake Pannon sequence, this granite was reached by a great number of exploration wells. The fining-upward basal series is 4 to 25 m of non-fossiliferous gravel and sand, consisting of poorly rounded or angular granite grains. Between these sands and the overlying lacustrine silt body there is a highly variable transitional layer; it is mostly gray or grayish yellow silt, but it also contains very fine-grained sand layers with much biotite, quartz, and orthoclase.

The overlying "productive series" of the brickyard is mostly well-sorted, fine-grained, clayey silt, containing very fine-grained sand intercalations

(Fig. 2). Its average thickness is 22.1 m, with a yellow, oxidized upper third and a bluish gray lower two-thirds. Due to hydrogeological considerations, the bottom of the clay pit is 1.5 to 3.0 m above the surface of the granite debris, at about 90 m above sea level. The layers dip 2 to 4 degrees to the east. Normal faults with a strike of N–S or NE–SW are common, with a throw of 0.1 to a few meters.

The bluish gray, fine-grained silt that makes up the bulk of the sequence contains a significant amount of clay minerals: mostly kaolinite, less smectite and chlorite. Mollusk shells are scattered or loosely accumulated in laterally discontinuous horizons. The only exception is Layer 21; this 20–30 cm thick, laterally continuous, gray, clayey, fine-grained silt layer is characterized by abundant, largely autochthonous mollusk shells (Plate I). The presence of bottom currents during the deposition of this layer is indicated by the accumulation of small shells and shell debris in the shadow of larger shells or bottom irregularities. Pebbles may also occur even in this fine-grained, clayey layer.

The silt contains several sandy coquina layers. These layers are thin (5–20 cm), poorly sorted (including pebbles), yellowish gray or reddish in color, and contain considerable mica. The presence of limonite in several sandy layers of the outcrop is probably a diagenetic feature. Shells within these layers are usually worn or crushed. A coalified trunk, 3 m long and oriented NNW–SSE, was found on the surface of Layer 20.

The "productive" part of the sequence is capped by a microlaminated, calcareous, very fine-grained sand. The high carbonate content of the sand (up to 43%) probably originates from dissolved mollusk shells. The carbonate cemented the sand grains. Only limonitic and manganous casts of fossils are found in this layer. The Lake Pannon sequence is unconformably overlain by Pleistocene red clay and loess.

Preservation and taphonomy of mollusks

From the Bátaszék outcrop we identified 51 mollusk species (25 bivalve and 26 gastropod), with about 4000 individuals. Shells from the gray silt were usually in good condition, sometimes even with the nacreous layer preserved (especially in *Lymnocardium majeri*). In layer 21, *Lymnocardium* shells were articulated (closed or open), whereas *Congeria* shells were either articulated or disarticulated. We regard the shells of the clayey silt as mostly autochthonous.

The preservation and taphonomy in the sandy layers showed differences between the lower and upper parts of the sequence. The fossil-rich but non-limonitic lower sandy layers, such as Layers 6 and 8, contained both weathered, worn shells that were transported from some distance and autochthonous shells. The shells of *Lymnocardium diprosopum* were always disarticulated; the shells of *L. hungaricum*, *L. rogenhoferi*, and *Phyllocardium planum* were disarticulated or articulated but open. *Congeria rhomboidea* occurred as disarticulated valves or rarely with closed valves.

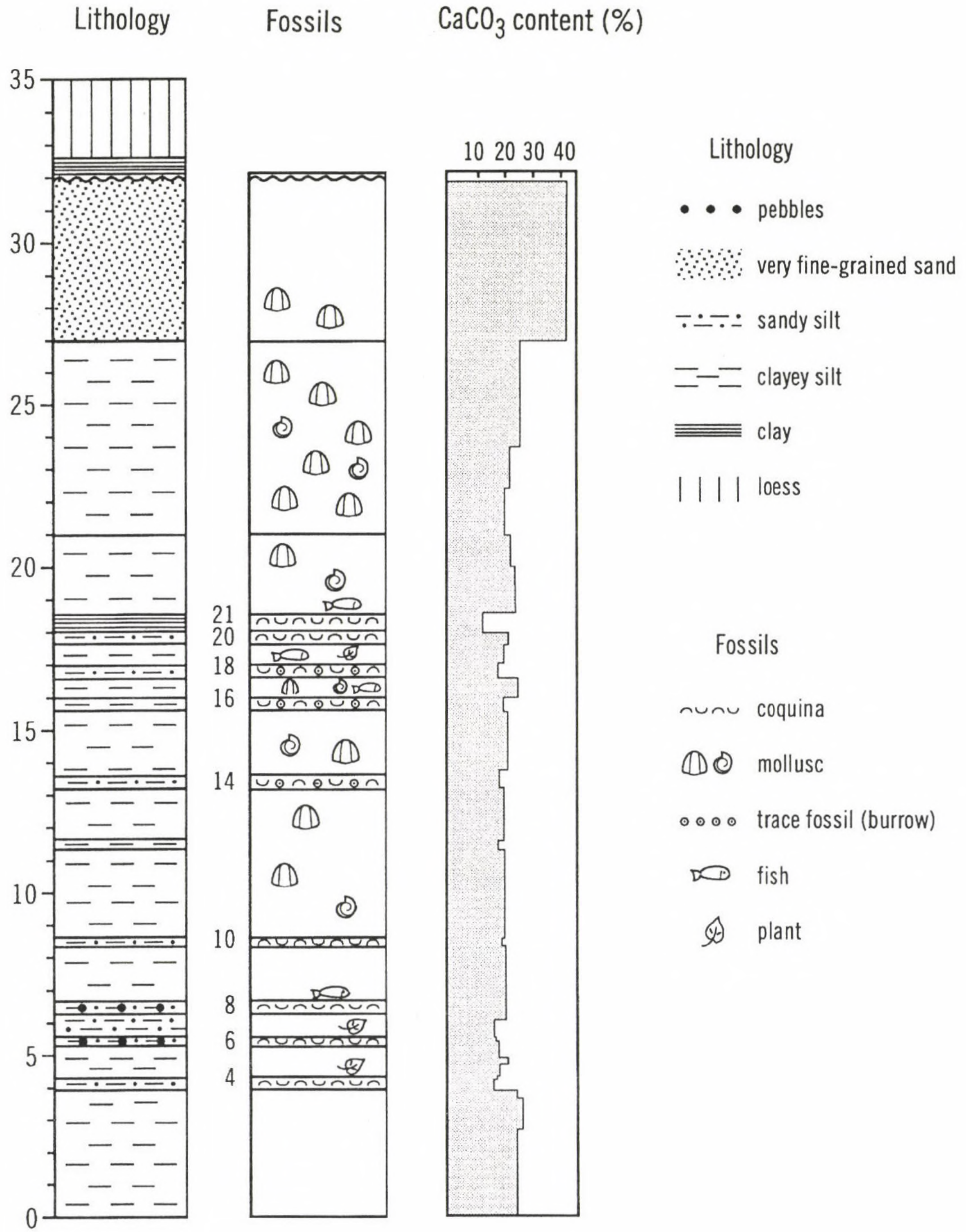

Fig. 2
Section of the Bátaszék outcrop

Most of the *Pisidium* and *Dreissenomya* shells were closed. *Dreissenomya* was buried in life position, perpendicular to the bedding planes. Shells in the upper, limonitic coquinas (e.g. Layers 18, 20), however, were eroded and broken, the bivalves being predominantly disarticulated. We regard them as allochthonous. In the yellow, oxidized upper third of the silt and overlying sand only limonitic casts were found.

Variation of shell morphology in three cardiid species

Only the fossils of three species (*Lymnocardium diprosopum, L. majeri, L. rogenhoferi*; Plate III) offered enough intact and well-preserved specimens for morphometric evaluation of their ontogeny and variability.

We observed the same pattern of ontogeny in *L. diprosopum* as described by Lőrenthey (1894) for this species from Szekszárd. We found 1–3 mm long juvenile shells in several samples. These are very inflated, with a protruding beak. The 7–15 mm long juvenile shells are flat, elliptical in outline, but with the beak still protruding. After this stage the shell becomes increasingly inflated during growth. Among adult specimens the height/length ratio never exceeds 0.90 (Fig. 3). The relatively elongate forms, similar to the figure of Brusina (1884) in the first representation of this species, are connected to the more rounded forms by a continuum of specimens. Lőrenthey (1894) in fact identified his Szekszárd form as *L. arpadense*, claiming that his material displayed continuous transition between the two species. He believed that *L. diprosopum* and *L. arpadense* merely represented different stages of ontogeny within a single species. The individual he depicted as his most *arpadense*-like specimen (Lőrenthey 1894, pl. 4, fig. 5), however, has a height/length ratio of 0.87, whereas the same ratio in adult *L. arpadense* is typically more than 1.00 (see figures in Basch 1990 and Szónoky et al., this volume). Another important difference between the two species is the robust hinge teeth of *L. arpadense*, which is already apparent in young ("*diprosopum*-sized") individuals. Therefore, we conclude that both our material and the Szekszárd material of Lőrenthey (1894) belong to *L. diprosopum* and not to the closely related *L. arpadense*.

Juvenile shells of *L. majeri* and *L. rogenhoferi* were also found in Bátaszék. Their outline is similar to adults of these species but more elongate. In all three species, both the height/length (Fig. 3) and width/length ratios increase with increasing length. During ontogeny the shell apparently grew faster along the ventral margin than at the anterior and posterior margins; the relative length of the shell decreased, and it became increasingly globular.

Paleoecology

The macrofauna at Bátaszék is dominated by bivalves (both in individuals and in species), but the microfauna washed out from the silt consists mainly of gastropods (*Micromelania, Valvata, Gyraulus*) and juvenile bivalves. Because the whole outcrop is dominated by *Congeria rhomboidea* and *Lymnocardium*

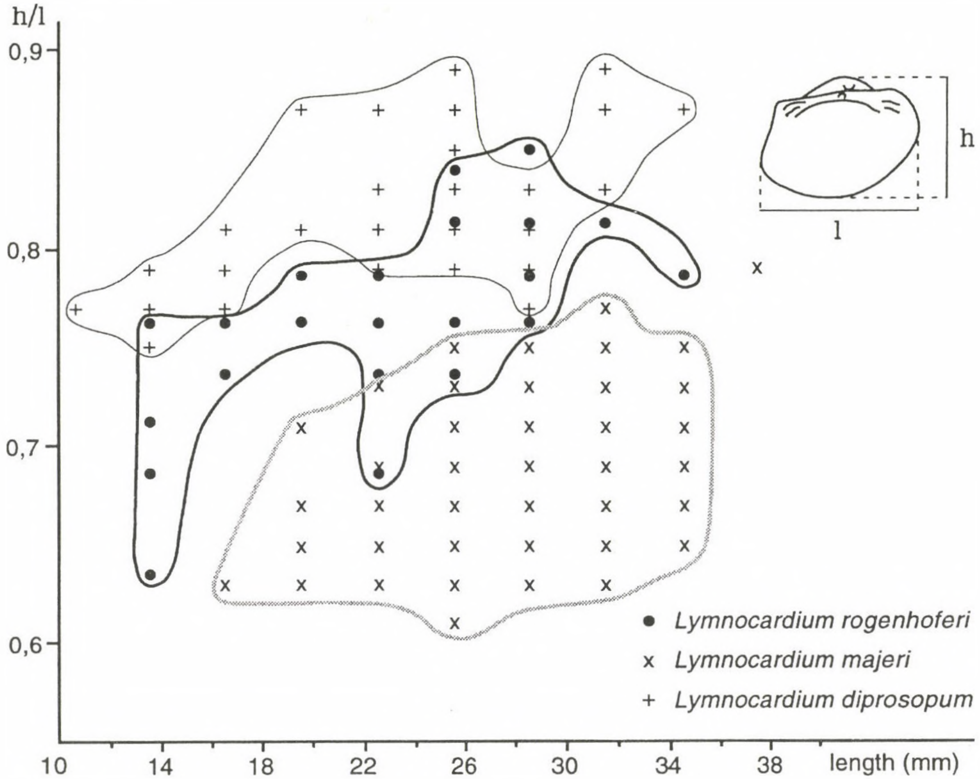

Fig. 3
The height/length ratio of three *Lymnocardium* species plotted against their length. Bigger shells tend to become higher and less elongated

hungaricum we designate this fauna the "*Congeria rhomboidea – Lymnocardium hungaricum* association". The species composition and ecology of this association was similar to that of the *Congeria zagrabiensis – Caladacna steindachneri* association of Korpás-Hódi (1983). The mollusk fauna was dominated by the suspension-filtering cardiids, belonging to the sessile infauna, and by the epibenthic dreissenids. Gastropods (*Valvata, Pyrgula, Micromelania, Gyraulus, Hydrobia*) were mostly small, deposit-feeding on organic-rich mud or grazing on small plants and algae. The family Lymnaeidae, including *Boskovicia, Zagrabica, Radix,* and *Valenciennius,* is represented by only a few individuals. The scarcity of these large grazers indicates a relatively limited source of aquatic plants in the paleoenvironment. Plant fragments were mostly found in the sandy layers, probably re-deposited by sediment gravity flows and storms.

Despite the overall dominance of *C. rhomboidea* and *L. hungaricum* there are changes in the composition of the mollusk fauna throughout the profile. The clayey silt yielded shells of *C. rhomboidea, L. hungaricum, L. schmidti, L.*

rogenhoferi, hydrobiid gastropods, and, especially in and above ayer 21, *L. majeri* and *C. zagrabiensis*. In general the infauna is more significant than the epifauna in these layers. Layer 21 contains the best-preserved and richest fossil material in the whole outcrop. The dominant bivalves are *C. rhomboida*, *C. zagrabiensis*, *Lymnocardium hungaricum*, and *L. majeri*. *Caladacna steindachneri* and *Dreissena aurcularis simplex* are also common. The ratio of infaunal (*L. hungaricum*, *L. majeri*, *C. steindachneri*, *L. rogenhoferi*, *L. riegeli*) to epifaunal species is high. Among the vagile epifauna, only large-sized lymnaeids are significant (*Boskovicia josephi*, *Radix* sp., *Velutinopsis velutina*, *Zagrabica maceki*).

The sandy intercalations, however, show a somewhat different picture. In the lower part of the section they are dominated by strongly eroded shells of *Congeria rhomboidea* and *Lymnocardium diprosopum*, and less weathered shells of *L. rogenhoferi*, *L. otiophorum*, *Pisidium* sp., *Gyraulus* cf. *striatus*, *G. constans*, *Micromelania* cf. *monilifera*, *M. cerithiopsis*, and *Zagrabica maceki*. Cardiids with thick shells and no siphonal aperture (*L. diprosopum*, *Phyllocardium planum*) are particularly common. The infauna (*Dreissenomya intermedia*, *L. hungaricum*, *L. majeri*) is subordinate. It is noteworthy that the shells of *L. majeri* are also very thick here. The abundance of gastropods indicates an organic-rich substratum. Plant and bone fragments also occur in these layers. In the upper sandy coquina layers, the most common species is *Congeria croatica*. Among cardiids, *Phyllocardium planum* and *Pteradacna pterophora* are common. Again, the epifauna is more important here than the infauna. With respect to the gastropods, small-sized species are common: *Valvata minima*, *Orygoceras fuchsi*, *Gyraulus* sp., *Micromelania cerithiopsis*, *M.* cf. *monilifera*. The ostracod fauna of the sandy intercalations also differs from that of the clayey silt. Thick-shelled (*Amplocypris*) and ornamented (*Cyprideis*, *Hemicytheria*) species are abundant in the former, whereas smooth, thin-shelled *Candona* prevail in the latter.

Pathologic alterations, such as irregular development of the nacreous layer, or even formation of half-pearls and pearls was observed in some *Congeria* and, to a lesser extent, *Lymnocardium* specimens, mostly coming from layer 21. Many of these alterations were probably caused by parasites. The bivalve tried to isolate the parasite by increased secretion of nacre. When it failed, crater-like formations (half-pearls) remained in the inner surface of the shell. We found a real pearl, 4 mm in diameter, close to the umbonal region of a *Congeria rhomboidea* shell (Plate II/3).

In summary, the *C. rhomboidea* – *L. hungaricum* association lived in the quiet sublittoral environment of Lake Pannon. We estimate the water depth to have been several tens of meters, between the fair weather wave base and storm wave base, and below the zone of abundant vegetation. The salinity of the water was probably mesohaline (5 to 9‰; Korpás-Hódi 1983). Similar environmental conditions were inferred from the ostracod fauna as well (Szuromi-Korecz 1991). The abundance of planktic dinoflagellates also indicates offshore conditions and relatively high salinity. Infaunal cardiids and epifaunal dreissenids prevailed in the mollusk association. The ostracod fauna was

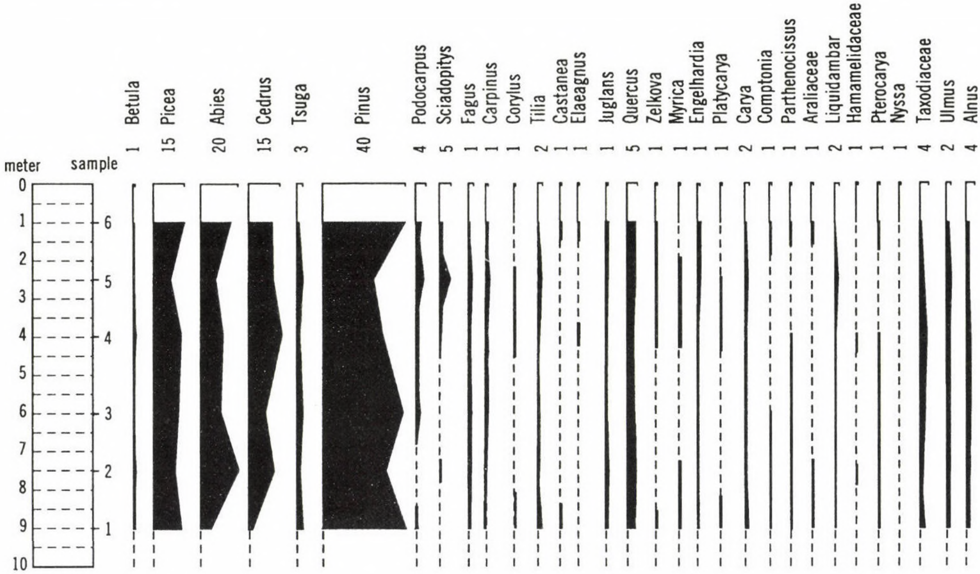

Fig. 4
The spores and pollen diagram of the Bátaszék outcrop. Six samples from a ten-meter-interval within the bluish-gray silt were analyzed

dominated by smooth, thin-shelled forms. From time to time storm waves transported sand, pebbles, eroded shells, shell debris, bone fragments, and plant material from littoral settings to this offshore environment, forming sandy coquina layers. In the predominantly allochthonous fossil assemblages of these layers thick-walled, robust ostracods prevail, and the mollusk epifauna appears more significant than the infauna. The abundance of small gastropods might have resulted from an organic-rich substratum created by storm events or gravity flows. The long trunk found on the surface of a storm deposit may have drifted offshore during the very storm that created the coquina layer.

The adjacent dry land maintained rich arboreous vegetation. As shown by palynological analysis, broad-leafed plants were very taxonomically diverse. Oak was most common, but the presence of hickory, hornbeam, and beech was also noteworthy (Fig. 4). Among conifers, the pollen of which can travel long distances, pine, cedar, fir, and spruce were most abundant whereas hemlock was subordinate.

Biostratigraphy

The Bátaszék mollusk fauna is very similar to the classic Szekszárd fauna of Lőrenthey (1894) and Okrugljak (Zagreb) fauna of Brusina (1884). The Bátaszék fauna is also very similar to the fauna of Jazovnik (Kosarlija valley), a facies stratotype of the "Lower Portaferrian" in western Serbia (Stevanović 1951, 1990).

A shared specialty of these two outcrops is the occurrence of *Orygoceras*, an aberrant gastropod that was long thought to have become extinct well before the deposition of the *Congeria rhomboidea* Zone. Its tiny shells were found exclusively in Jazovnik and Bátaszék within the upper part of the Lake Pannon sequence.

The Bátaszék outcrop belongs to the sublittoral *Congeria rhomboidea* Zone. A probably transported specimen of *Prosodacnomya* sp. establishes the correlation with the littoral *Prosodacnomya* Zone. Dinoflagellates indicate the lower part of the *Galeacysta etrusca* Zone (it corresponds to the "*Spiniferites tihanyensis* Zone" of Sütő-Szentai 1994; see Sütő-Szentai, 1995). The ostracod fauna corresponds to the "Portaferrian" fauna of Krstić (1990).

Acknowledgement

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Plate I

1. Coquina layers in the sublittoral clayey silt of the Bátaszék outcrop. The sandy, limonitic layers with allochthonous fauna (#18 and 20) are interpreted as storm deposits. The most fossil-rich layer of the outcrop, #21, however, contains mostly autochthonous shells, and the matrix is even finer-grained than in the bulk of the sequence. Scale in cm.
2. Surface of Layer 21 with shells of *Congeria zagrabiensis* Brusina, *Lymnocardium riegeli* (Hörnes), *L. majeri* (Hörnes), *Pteradacna pterophora* (Brusina), and *Valenciennius reussi* Neumayr.

Plate II

1, 2. *Congeria rhomboidea* Hörnes. Our specimens correspond to *C. alata* Brusina; however, we tentatively regard the latter as a junior synonym for *C. rhomboidea* (see also Szónoky et al., this volume)
3. A pearl in the shell of *Congeria rhomboidea*
4–6. *Congeria croatica* Brusina
7, 8. *Congeria* aff. *partschi* Czjzek (Layer 16)
9, 10. *Dreissenomya intermedia* Fuchs (Layer 8)

Scale bars indicate 1 cm.

Plate III

1–7. *Lymnocardium majeri* (Hörnes)
8–13. *Lymnocardium rogenhoferi* (Brusina)
14–19. *Lymnocardium diprosopum* (Brusina) (Layer 8)

Scale bar indicates 1 cm.

Plate IV

1, 2. *Lymnocardium hungaricum* (Hörnes) (Layer 5). Our specimens correspond to *L. zagrabiense* (Brusina); we suppose that it is a junior synonym for *L. hungaricum* (see also Szónoky et al., this volume).
3. *Pteradacna pterophora* (Brusina) (Layer 21)
4. *Lymnocardium schmidti* (Hörnes)
5, 6. *Lymnocardium majeri* (Hörnes)
7. *Lymnocardium riegeli* (Hörnes) (Layer 21)
8–10. *Phyllocardium planum* (Deshayes) (Layer 16)

Scale bars indicate 1 cm.

Plate V

1–3. *Zagrabica maceki* Brusina (Layer 8)
4. *Valenciennius reussi* Neumayr (Layer 26; scale bar indicates 1 cm)
5. *Radix (Velutinopsis) velutina* (Deshayes) (scale bar indicates 5 mm)
6. *Radix (Lytostoma) grammica* (Brusina)
7. *Gyraulus* cf. *striatus* (Brusina) (Layer 8)
8. *Boskovicia hantkeni* (Lorenthey) (Layer 19)
9. *Boskovicia josephi* Brusina
10, 11. *Orygoceras fuchsi* Kittl (Layer 16)
12. *Micromelania fuchsiana* Brusina (Layer 6)
13. *Micromelania cf. monilifera* Brusina (Layer 8)
14. *Micromelania cerithiopsis* Brusina (Layer 8)

In all photos other than 4 and 5, scale bars indicate 1 mm.

Plate VI

1–3. *Cyprinidae* sp. (determined by L. Kordos)

Scale bars indicate 1 cm.

Plate VII

1, 2. *Leptocythere (Amnicythere) multituberculata* (Livental 1929). Bátaszék, Layer 6. 1: exterior lateral view of female left valve (137 x); 2: exterior lateral view of female right valve (138 x).
3. *Leptocythere (Maeotocythere)* sp. (*radae*) Krstić (1985), pl. XIV. Fig. 10. Bátaszék, Layer 6. Exterior lateral view of male left valve (204x).
4. *Leptocythere (Amnicythere) palimpsesta* (Livental 1929). Bátaszék, Layer 6. Exterior lateral view of female right valve (152 x).
5, 6. *Leptocythere (Maeotocythere) bosqueti* (Livental 1929). Bátaszék, Layer 24. 5: exterior lateral view of male right valve (186 x); 6: exterior lateral view of male left valve (182 x).
7, 8. *Loxoconcha (Loxoconcha)* cf. *eichwaldi* Livental 1929. Bátaszék, Layer 24. 7: exterior lateral view of male right valve (133 x); 8: exterior lateral view of male left valve (133 x).

Plate VIII

1, 2. *Cyprideis (Cyprideis) triangulata* Krstić 1963. Bátaszék, Layer 6. 1: exterior lateral view of female left valve (84 x); 2: exterior lateral view of female right valve (81.5 x).
3, 4. *Candona (Thaminocypris) alta* (Zalányi 1929). Bátaszék, Layer 24. 3: exterior lateral view of female left valve (88 x); 4: exterior lateral view of male right valve (88 x).
5, 6. *Candona (Caspiolla) parabalcanica* Krstić 1971. Bátaszék, Layer 24. 5: interior lateral view of female left valve (98.5 x); 6: interior lateral view of male right valve (100 x).
7. *Candona (Lineocypris?)* sp. Bátaszék, Layer 24. Exterior lateral view of male left valve (93 x).

Plate I

1

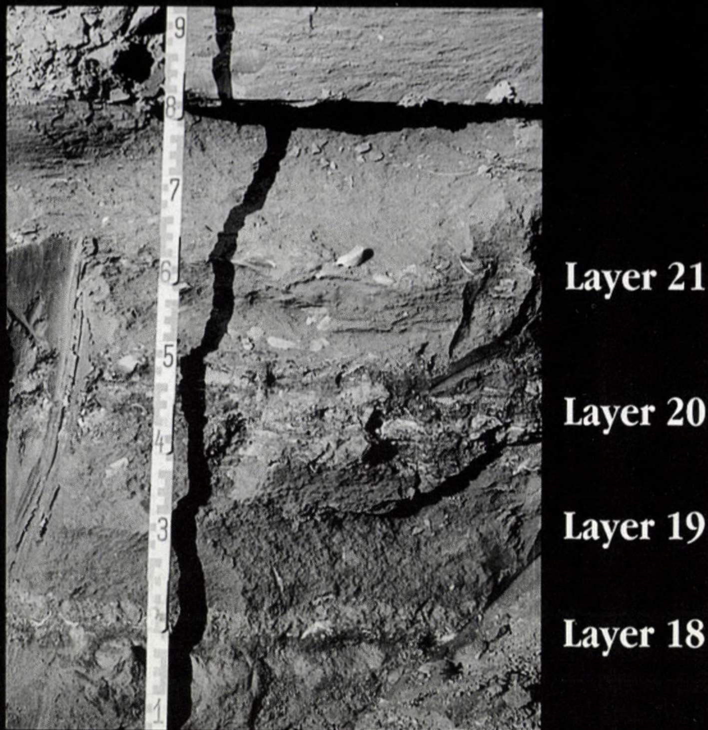

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Plate II

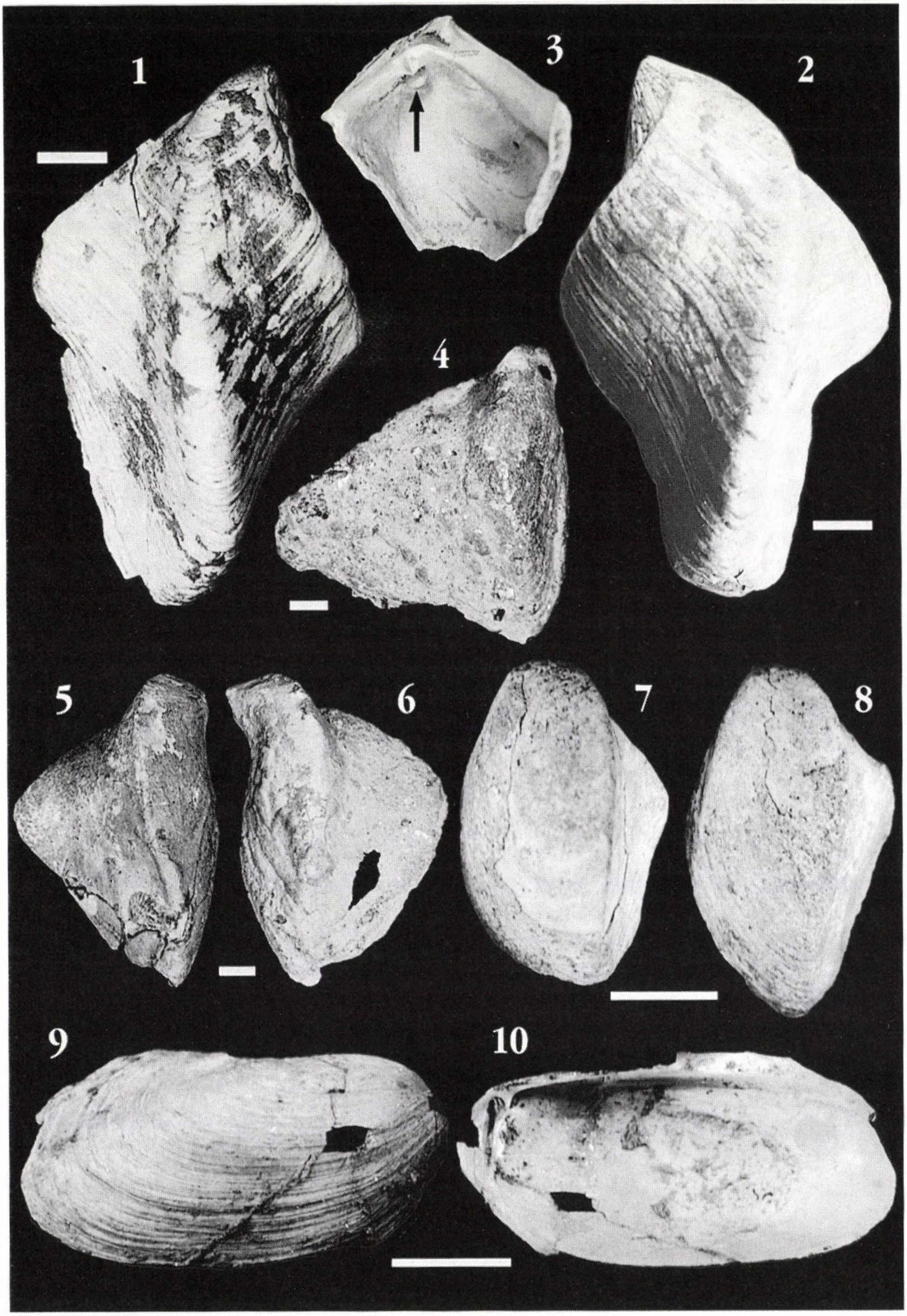


Plate III

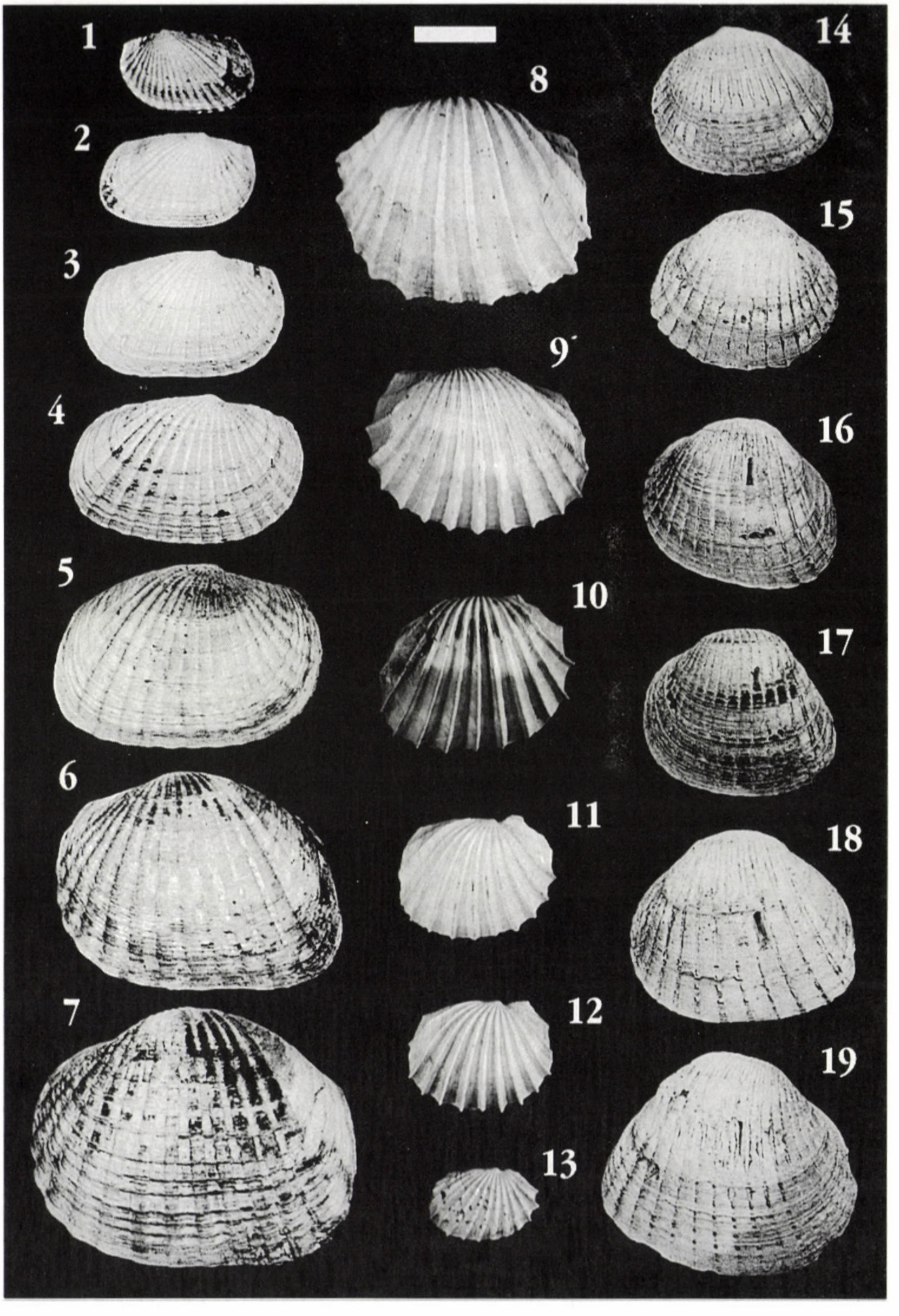


Plate IV

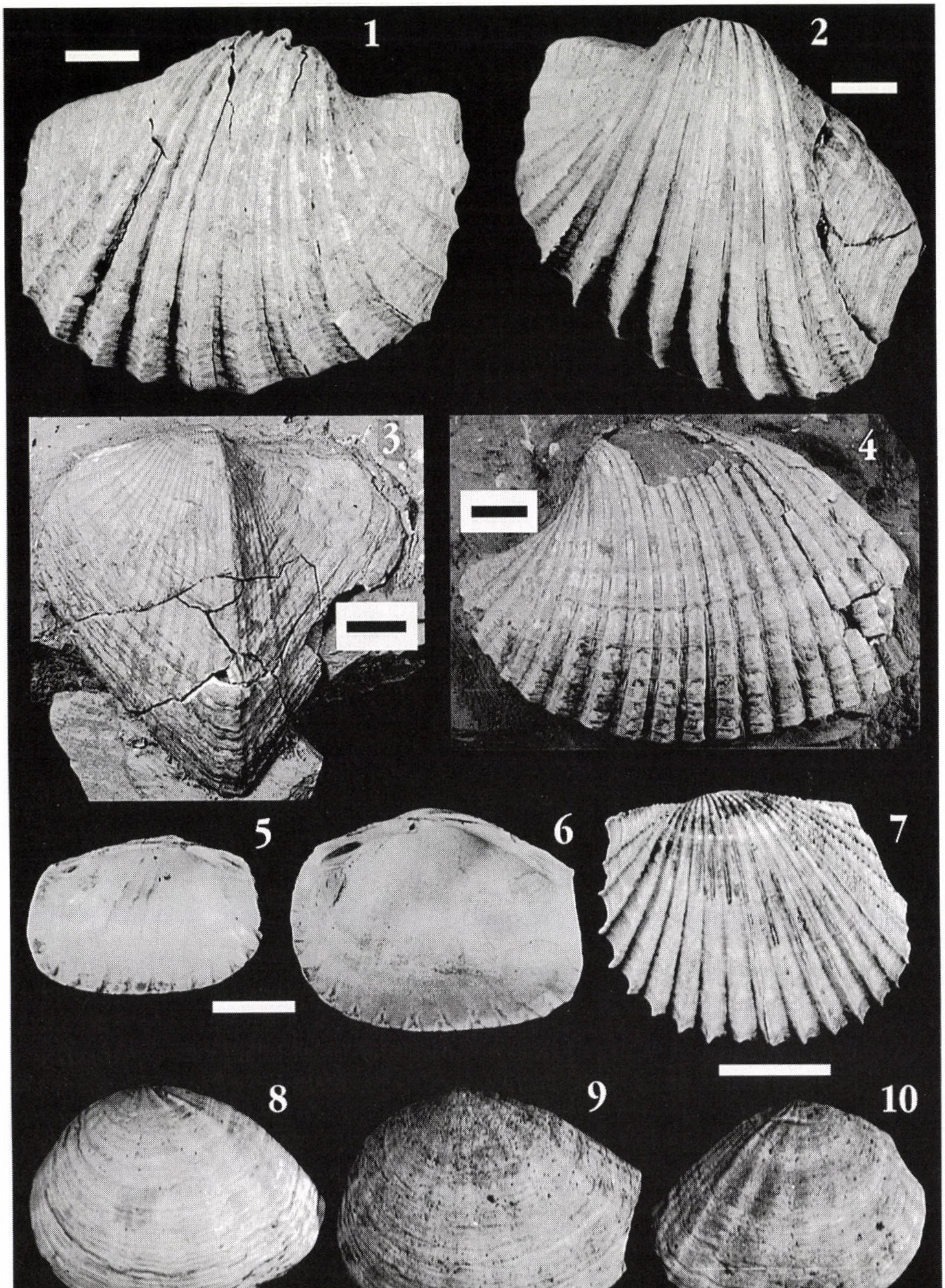


Plate V

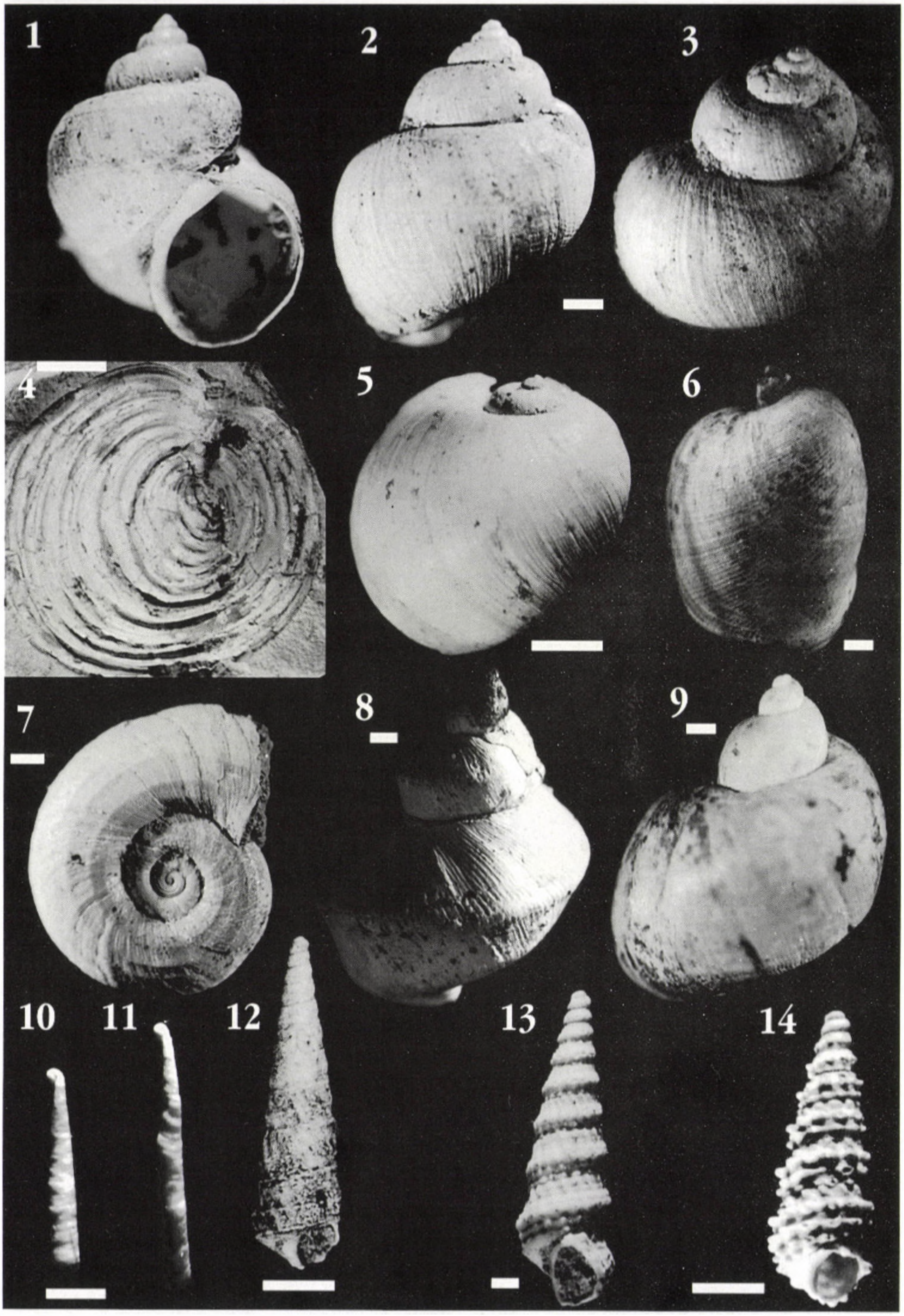


Plate VI

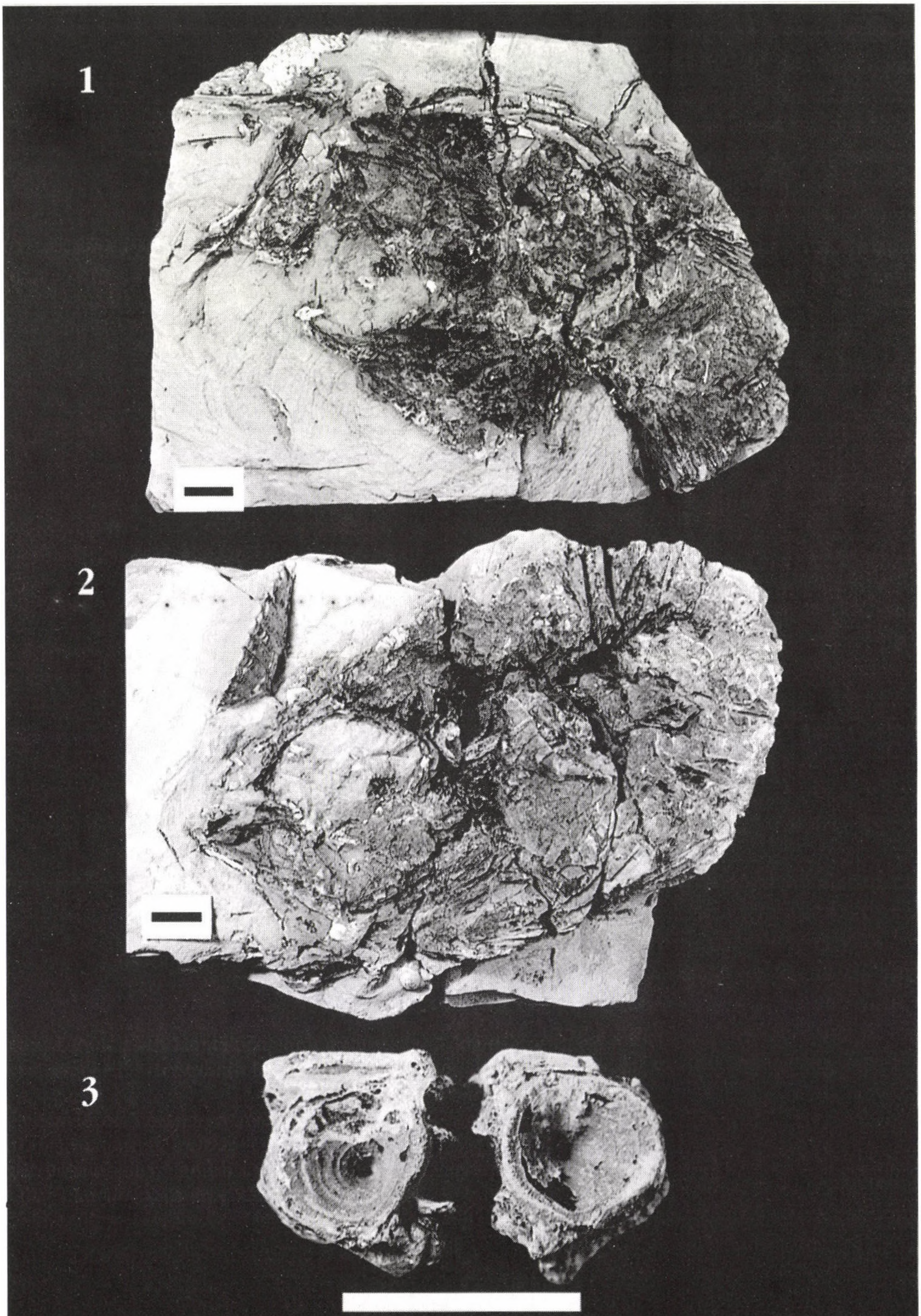


Plate VII

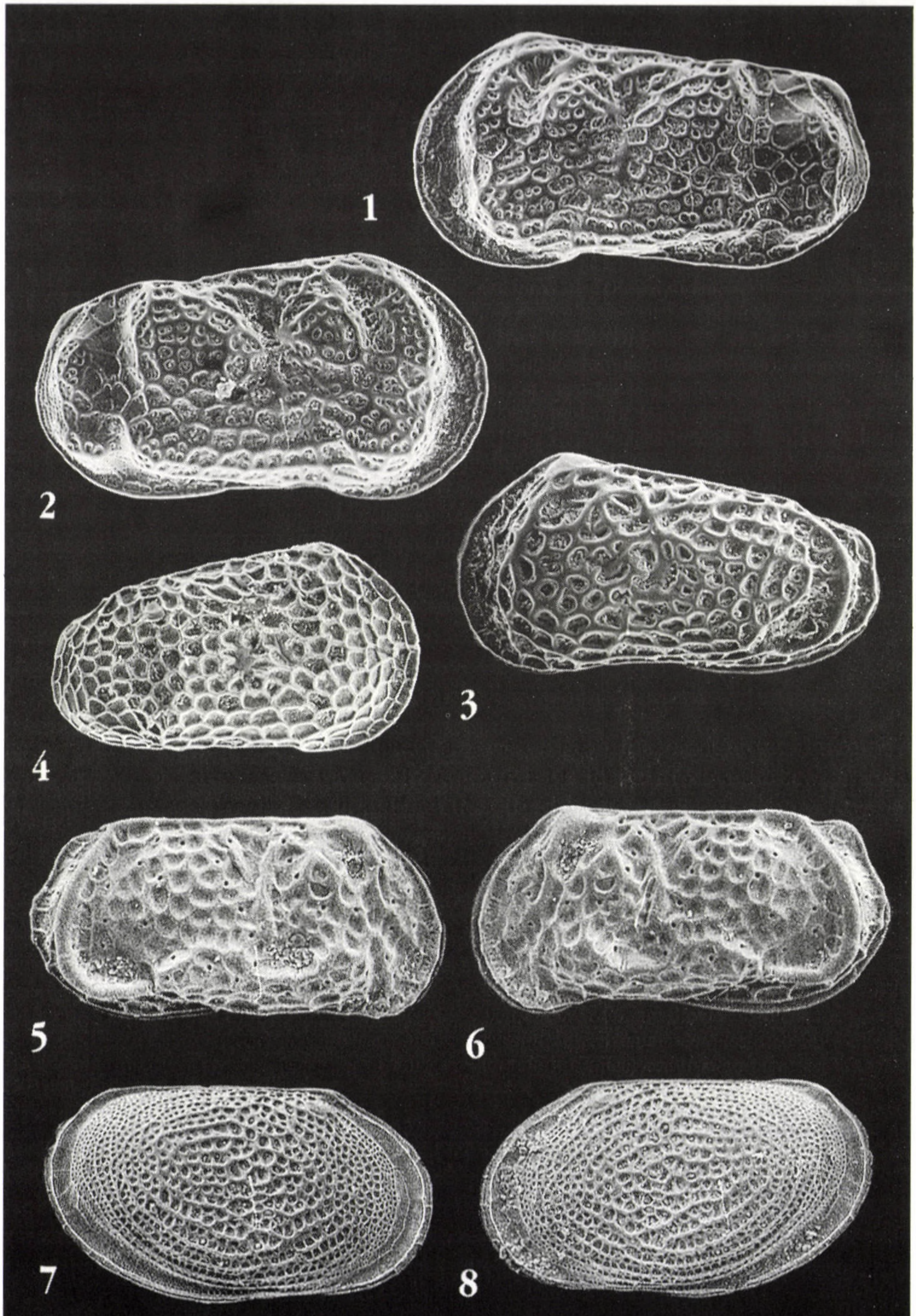
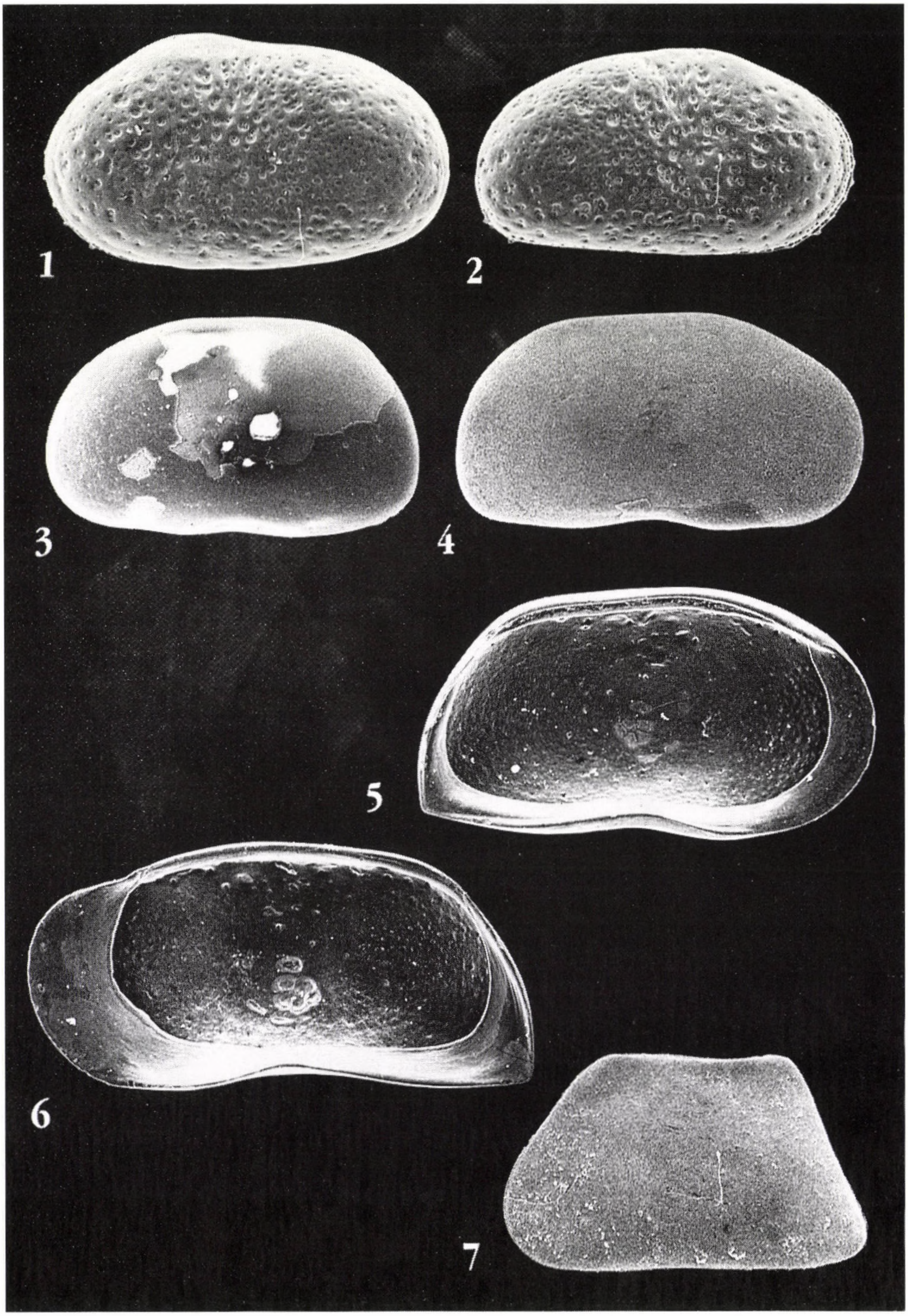


Plate VIII



Appendix

Fossils from the Bátaszék outcrop:

Bivalvia:

Pisidium sp.
Lymnocardium diprosopum (Brusina)
Lymnocardium hungaricum (Hörnes)
Lymnocardium majeri (Hörnes)
Lymnocardium otiophorum (Brusina)
Lymnocardium riegeli (Hörnes)
Lymnocardium rogenhoferi (Brusina)
Lymnocardium schmidti (Hörnes)
Lymnocardium cf. *szaboi* (Lőrenthey)
Lymnocardium pelzelni (Brusina)
Lymnocardium cf. *okrugljaki* (Basch)
Caladacna steindachneri (Brusina)
Paradacna okrugici (Brusina)
Phyllocardium planum (Deshayes)
Pteradacna pterophora (Brusina)
Prosodacnomya sp.
Dreissena aff. *auricularis simplex* (Fuchs)
Dreissena sp.
Dreissenomya intermedia Fuchs
Dreissenomya sp. indet.
Congeria croatica Brusina
Congeria aff. *partschi* Czjzek
Congeria rhomboidea Hörnes
Congeria zagrabiensis Brusina
Congeria balatonica Partsch

Gastropoda:

Valvata cf. *balatonica* Rolle
Valvata minima Fuchs
Valvata simplex Fuchs
Valvata cf. *variabilis* Fuchs
Valvata sp.
Hydrobiidae sp.
"Pseudamnicola" atropida (Brusina)
"Pseudamnicola" sp.
Pyrgula incisa Fuchs
Micromelania cerithiopsis Brusina
Micromelania cf. *coelata* Brusina
Micromelania fuchsiana Brusina
Micromelania cf. *monilifera* Brusina
Radix (Velutinopsis) velutina (Deshayes)
Radix (Lytostoma) grammica (Brusina)
Boskovicia hantkeni (Loerenthey)
Boskovicia josephi Brusina
Zagrabica maceki Brusina
Valenciennius reussi Neumayr
"Gyraulus" constans (Brusina)
"Gyraulus" cf. *radmanesti* (Fuchs)
"Gyraulus" cf. *striatus* (Brusina)
"Gyraulus" cf. *tenuis* (Fuchs)
"Gyraulus" sp.
Segmentina sp.
Orygoceras fuchsi Kittl

Ostracoda:

The entire Bátaszék section is rich in ostracods. Only some preliminary results are given here. The samples are from Layers 6 and 24. Number of specimens (in brackets) refers to valves, not individuals.

Layer 6:
Cyprideis triangulata Krstić (60 specimens)
Hemicytheria josephinae (Zalányi) (26)
Amplocypris nonreticulata Krstić (14)
Leptocythere (Amnicythere) multituberculata (Livental) (14)
Candona (Caspiolla) parabalcanica Krstić (8)
Candona (Caspiolla) ossoinae Krstić (5)
Candona (Bakunella) cf. *dorsoarcuata* (Zalányi) (5)
Candona (Sirmiella) arcuatoides Krstić (5)

Candona (Thaminocypris) alta (Zalányi) (4)
Candona (Reticulocandona) reticulata (Méhes) (4)
Candona (Serbiella) hastata Krstić (2)
Candona (Sinegubiella) sublabiata Krstić (2)
Candona (Caspiocypris?) sp. (2)
Leptocythere (Amnicythere) palimpsesta (Livental) (2)
Leptocythere (Maeotocythere?) sp. (2)
Leptocythere (Amnicythere) cf. *stanchevae* Krstić (1)
Candona (Pontoniella) paracuminata Krstić

Layer 24:

Candona (Sirmiella) arcuatoides Krstić (23 specimens)
Candona (Thaminocypris) alta (Zalányi) (14)
Candona (Caspiolla) parabalcanica Krstić (12)
Hemicytheria josephinae (Zalányi) (4)
Candona (Bakunella) cf. *dorsoarcuata* (Zalányi) (4)
Candona (Bakunella) sp. juv. (4)
Candona (Serbiella) hastata Krstić (3)
Candona (Pontoniella) paracuminata Krstić (2)
Amplocypris nonreticulata Krstić (2)
Loxoconcha (Loxoconcha) cf. *eichwaldi* Livental (2)
Leptocythere (Maeotocythere) bosqueti Livental (2)
Candona (Zalanyiella) venusta (Zalányi) (1)
Candona (Reticulocandona) reticulata (Méhes) (1)
Candona (Lineocypris?) sp. (1)
Candona (Caspiocypris?) sp. (1)
Cypria tocorjescui Hanganu (1)

Pisces (by László Kordos):

Fish teeth belonging to the family Cyprinidae were found in Layers 8, 16, and 18. In Layer 21 we found 14 cycloid scales; they are probably remnants of a single fish that died close to this site. The yellow silt yielded a whole skeleton of a *Cyprinidae* sp.

Dinoflagellata:

Chytroeisphaeridia hungarica Sütő-Szentai
Dinoflagellata form 28
Galeacysta etrusca Corradini et Biffi
Millioudodinium pannonicum (Nagy) Stover et Evitt
Impagidinium globosum Sütő-Szentai
Spongiosphaera pannonica (Baltes) comb. nova

Zygnemataceae:

Spyrogyra type 3c B. Van Geel et al.
Cooksonella circularis Nagy

Chrysophyta:

Botryococcus braunii Kützing

Chlorophyta:

Pediastrum boryanum Menegh
Pediastrum simplex Meyen

Spores and pollen:

Six samples have been processed from the lower, gray part of the Bátaszék section. The following genera were determined:

Sphagnum sp.
Selaginella sp.
Osmunda sp.
Pteris sp.
Gleichenia sp.
Polypodium sp.
Dryopteris sp.
Podocarpus sp.
Abies sp.
Cathaya sp.
Cedrus sp.
Picea sp.
Pinus sp.
Tsuga sp.
Sciadopitys sp.
Taxodiaceae gen. indet
Ephedra sp.
Myrica sp.
Comptonia sp.
Carya sp.
Engelhardtia sp.
Pterocarya sp.
Platycarya sp.
Juglans sp.
Alnus sp.
Typha sp.
Betula sp.
Carpinus sp.
Corylus sp.
Fagus sp.
Quercus sp.
Ulmus sp.
Zelkova sp.
Eucommia sp.
Caryophyllaceae gen. indet.
Chenopodiaceae gen. indet.
Magnolia sp.
Nuphar sp.
Liquidambar sp.
Hamamelidaceae gen. indet.
Rosaceae gen. indet.
Geraniceae gen indet.
Ilex sp.
Parthenocissus sp.
Tilia sp.
Elaeagnus sp.
Nyssa sp.
Cornaceae gen. indet.
Hedera sp.
Araliaceae gen. indet.
Ericaceae gen. indet.

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Árpád, a classic locality of Lake Pannon bivalves

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Ten new bivalve species, including eight cardiids, were described from the Late Miocene Lake Pannon deposits of the village of Árpád during the second half of the last century. This assemblage has played an important role in the classification of brackish cardiids since the 1860s. If we accept the current taxonomy, the 16 cardiid species known from Árpád (including those from our recent collection) represent eight genera; the genus *Lymnocardium* is represented by seven subgenera. Since most of the original species descriptions were based on donated material, the precise locality of these materials has remained unknown. Today a sandpit exposes the Lake Pannon layers south of the village. The fossil content of these layers corresponds well to the classic materials. The fining upward sedimentary sequence consists of fine sand and coarse silt layers. The lower unit is aggradational whereas the upper one reflects a minor transgression. Recurrent coquinas represent short events, probably storms, or gravitational transports from shallower settings. A one-meter thick laminite with leaves was probably deposited in a restricted part of the lake, depleted of benthic animals and without bioturbation. For most of the section, the considerable number and diversity of cardiids and large *Congeria* point to an unrestricted, nearshore, shallow lacustrine environment. The Árpád outcrop belongs to the *Congeria rhomboidea* Zone. Based on mollusks, dinoflagellates, and ostracods, we suggest that it occupies a median stratigraphic position within the *Congeria rhomboidea* Zone; it is younger than the Bátaszék fauna (described in this volume) with *Lymnocardium diprosopum*, but older than the youngest Lake Pannon fauna with large *Lymnocardium petersi* and *Prosodacnomya vodopici* in Slavonia and Syrmia. This hypothesis, however, requires further testing.

Key words: biostratigraphy, lacustrine features, Lake Pannon, Mecsek, Mollusca, Neogene, Ostracoda, paleoecology, Pannonian Basin, Paratethys

History of the Árpád fauna

In his four-part work on the fossil mollusks of the Tertiary Vienna basin, Hörnes (1859–1867) depicted ten bivalve species from the village of Árpád

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("Árpád südöstlich von Fünfkirchen")[1]. Today the name of this village is Nagyárpád, but it cannot be found on a map of Hungary because it became part of the town of Pécs ("Fünfkirchen" in German, "Pec" in Serbo-Croat). With the notable exception of *Congeria rhomboidea* all other bivalves are cardiids: *Cardium edentulum* (Deshayes), *C. planum* Deshayes, and seven new species: *Cardium arpadense, C. schmidti, C. hungaricum, C. riegeli, C. petersi, C. haueri,* and *C. majeri*. Brusina (1884) pointed out that what Hörnes depicted as a juvenile "*C. hungaricum*" was, in fact, a new species (*C. rogenhoferi*). Therefore, the figures of Hörnes represent 11 species, including 10 cardiids. The Árpád fauna thus became the first cardiid assemblage of the Late Miocene Lake Pannon known to science.

The spectacularly large, thick, and often widely gaping Árpád cardiids immediately seized the attention of paleontologists and conchologists worldwide. Conrad (1866) proposed a new genus (*Pseudocardia*) that included four of the Árpád species (*C. schmidti, C. hungaricum, C. majeri, C. haueri*), two other Lake Pannon cardiids (*C. apertum, C. conjungens*), and strangely enough, seven Mesozoic species. Two years later, Conrad (1868) changed the generic name to *Vetocardia*, and made it clear that he established this genus for Cretaceous forms (Conrad 1868). Stoliczka (1871), however, acknowledged the Árpád cardiids as representatives of a distinct genus and established *Lymnocardium*, with *L. haueri* as type species. He also introduced the family Lymnocardiidae which, aside from *Lymnocardium*, included the living brackish cardiids of the Black Sea and the Caspian Lake, united under the generic name *Didacna*.

The taxonomy of the brackish cardiids, however, has changed considerably since the time of Stoliczka. The genus *Lymnocardium* has been divided into many subgenera. *Arpadicardium*, a subgenus introduced by Ebersin (1947), includes *L. majeri* and bears the name of the village, although the type species (*A. peregrinum*) is from Kubani, Russia. Another subgenus of Ebersin's, *Tauricardium*, originally had *L. (Tauricardium) subsquamulosum* as its type species; Stevanovic (1961) subsequently established that this name is a junior synonym for the Árpád species *L. petersi*. Three Árpád species were chosen by Basch (1990) as type species of his new subgenera: *L. (Hungarocardium) hungaricum, L. (Zagrabicardium) riegeli,* and *L. (Podravinicardium) arpadense*. Because *L. schmidti* was assigned to the subgenus *Pannonicardium* by Stevanovic (1951), today all seven new cardiid species of Hörnes belong to different subgenera.

Apart from the eight new bivalve species of Hörnes, Árpád is type locality of *Dreissenomya intermedia* Fuchs (1873) and *"Paradacna" wurmbi* (Lőrenthey)

1 Note: Inclusion of the Árpád material into Hörnes' work entitled "The fossil mollusks of the Tertiary Vienna basin" has led to some confusion. For example, in the Treatise on Invertebrate Paleontology, Keen (1969) refers the Árpád species *Lymnocardium haueri* to Austria, although Árpád and the Vienna basin deposits are separated from each other by hundreds of kilometers and millions of years. In fact, none of the Árpád species occurs in the Vienna basin.

(1893). Descendants of the latter, along with *Lymnocardium petersi* and *Congeria rhomboidea*, are abundant in the Pontian Stage of the Black Sea region. *Congeria rhomboidea* is now regarded as an "index fossil", the main tool in correlating and identifying the Pontian Stage along the entire Paratethys (Gulyás 1998).

Outcrops in the Árpád area

Hörnes' descriptions were based on donated specimens, housed today at the Natural History Museum of Vienna. The donors did not provide information on the exact localities, so the precise provenance of Hörnes' material is uncertain. After Hörnes' work was published, many paleontologists tried to find fossiliferous outcrops in the area. The first locality description (at a water mill east of the village) is given by Kókán (1873, 1874). Lőrenthey (1893, 1894) described three fossiliferous outcrops; his best locality was situated between the villages of Üszög and Árpád, at the boundary of Pécs, on the slope of a clayey sandy hill. Strausz (1953) identified some species from an "Árpád sandpit", probably identical with the outcrop that exists today. This outcrop was described briefly by Stevanovic (1961). Bartha (1966) gave a sketchy section of the outcrop, mentioning and illustrating some species occurring there. Dobos-Hortobágyi and Szónoky (1991) provided a detailed sedimentological and paleontological description of the locality, tentatively assigning it to the Somló Formation (Jámbor 1980).

The present-day outcrop is a sand pit of intermittent use, situated about one and a half km away from Lőrenthey's (1893, 1894) locality described above, at the S end of the village (Fig. 1). This is currently the only fossiliferous outcrop in the vicinity. Its fossils correlate well with those of previously described outcrops. The 30 m long pit exposes a sequence approximately 14 m thick. A similar sandy formation is widespread along the southern margin of the Mecsek Mountains, found in boreholes and sand pits (Kleb 1973).

Sedimentology and fossils of the modern Árpád sand pit

Nine layers of fine-grained sand to fine-grained silt may be distinguished on the basis of grain size in the outcrop (Fig. 2); the lower three are currently covered with debris. The strata dip 8^0 SSE. Several normal faults trending NW–SE are visible in outcrop. Carbonate content increases upwards, from 5–10% near the bottom to 20–25% at the top. The upper unit of the outcrop (layers 7 to 9) shows a fining upward sequence, the lower unit (layers 1 to 6) is an alternation of silty, very fine sand and sandy, coarse silt layers. The uppermost of these (layer 6, Fig. 2) is the coarsest.

The lower unit (# 1–6) contains several coquina beds. Bivalve shells are usually disarticulated, in both concave upward and convex upward positions, but articulated shells are also present. The coquina in the uppermost layer of the lower unit (# 6) contains mostly concave-upward, large cardiids and

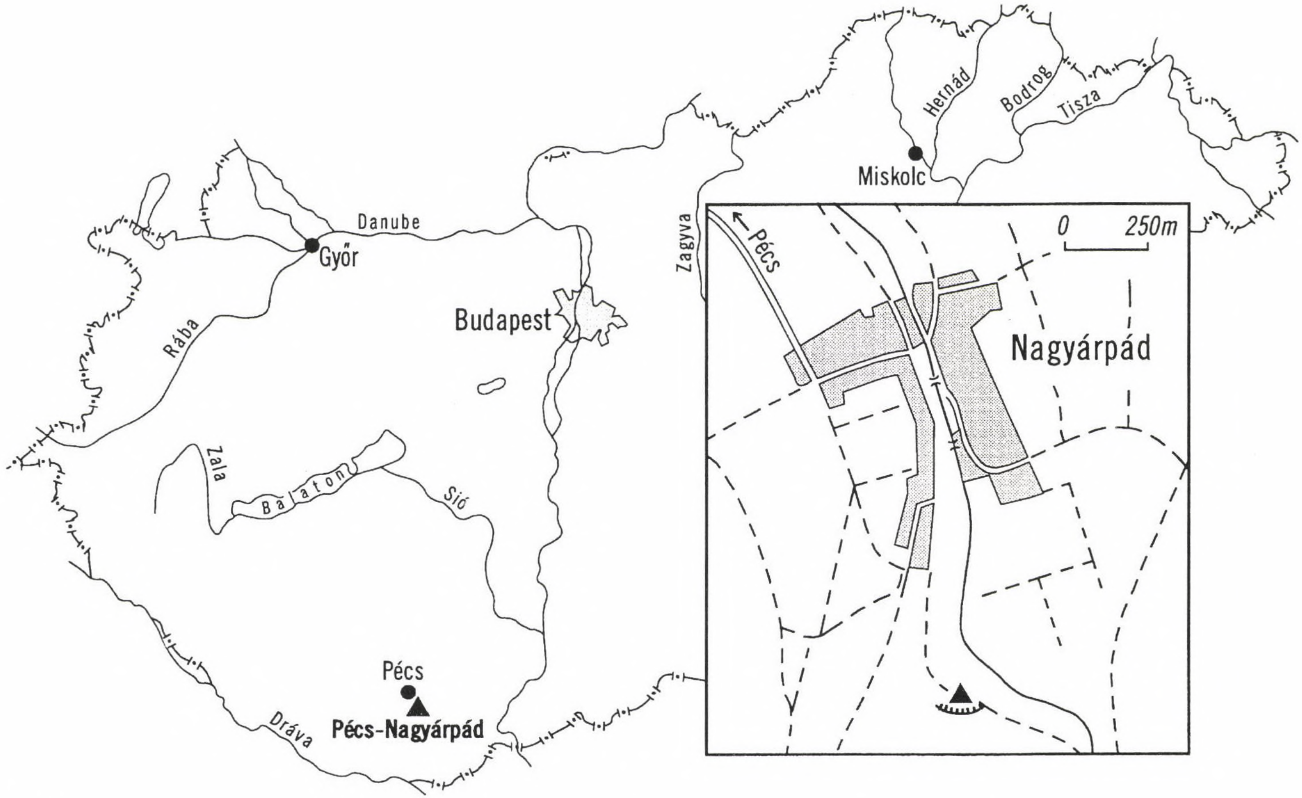

Fig. 1
Location of Nagyárpád (=Árpád) in Hungary, and location of the sand pit next to the village (black triangle)

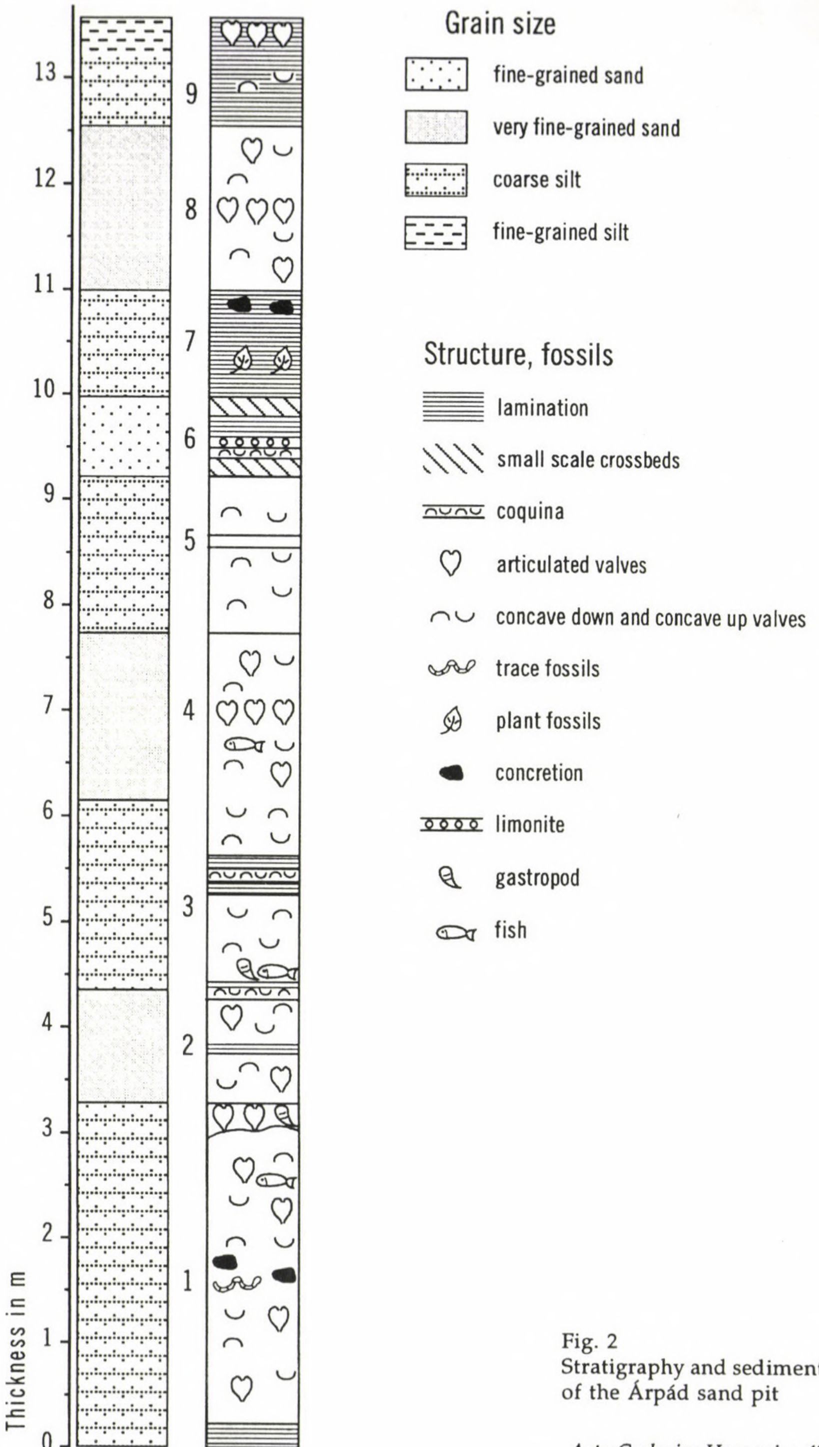

Fig. 2
Stratigraphy and sedimentology of the Árpád sand pit

Congeria rhomboidea, the cavities of which are filled with well-sorted mollusk bioclasts. Laminated sand and coarse silt occur in some levels. Layer 6 contains two small-scale crossbedded intervals. Otherwise, bedding is obsolete, often marked only by accumulations of shells along planes. Clear marks of bioturbation are present at mid-height of layer 1, as irregular burrow fills. In layer 4 articulated *Dreissenomya intermedia* were found in 5–7 cm deep burrows. The mollusk fauna of the lower unit is dominated by large cardiids and dreissenids (most abundant are *Congeria rhomboidea, Lymnocardium schmidti, L. hungaricum*, and *L. arpadense*) which occur scattered in the matrix or within some coquinas.

The upper unit (#7–9) is comprised of two coarse to medium-grained silt layers, intercalated by a fine sand layer, the material of which intrudes into the underlying silt as bag-shaped depression fillings. A laminated level in layer 7 yielded deciduous leaf remnants. The intercalated sand (layer 8) is similar to sands observed in the lower unit. Layer 9 is the finest-grained part of the entire outcrop. It is rich in scattered, mostly articulated bivalve shells, often in a butterfly position: *Lymnocardium rogenhoferi, L. otiophorum, Parvidacna chartacea, Paradacna okrugici, L. hungaricum, Caladacna steindachneri, "Paradacna" wurmbi*, and *L. majeri*. Gastropods are also present, including *Zagrabica* sp. and *Valenciennius* sp. Preservation of mollusks in this layer is excellent but only moderate in other parts of the outcrop. A rich and well-preserved ostracod fauna, dominated by the genus *Candona*, was recovered from this layer. The ostracod fauna of the uppermost layer indicates a miohaline to mesohaline (3 to 9‰), low energy but littoral (shallower than the wave base; 10–15 m) environment.

Thus, the lower unit reflects aggradation, whereas the upper one reflects transgression resulting in a slightly deeper environment. The considerable number and diversity of cardiids and large *Congeria* indicate an unrestricted nearshore shallow lacustrine environment. Recurrent coquinas represent short events, probably storms, or gravitational transports from even shallower settings. Bioturbation (probably caused mainly by cardiids) has destroyed bedding planes in most of the section. Bioturbation caused by deep-burrowing *Dreissenomya* is obvious in a level (layer 4) where shells were preserved in living position within burrows. The irregular burrow fills in layer 6 may be attributed to polychaetes. Laminites with leaves (layer 7) were probably deposited in restricted parts depleted of benthic fauna, without bioturbation.

Synonyms for the Árpád bivalves

The Árpád fauna was the first carefully described and depicted significant bivalve assemblage of the Late Miocene Lake Pannon. The large Árpád bivalves were very different from the Lake Pannon faunas of the Vienna basin, the Balaton region, or the Banat, which were described in the 1870s. It maintained its "exotic" status until the Croatian paleontologist S. Brusina began to publish his material from the southern part of the Pannonian basin. Brusina not only found many of the Árpád species in Croatian localities but also described new species which seemed to be very closely related to the Árpád ones. Such "paired" species include *Congeria rhomboidea* – *C. alata*, *Lymnocardium arpadense* – *L. diprosopum*, *L. schmidti* – *L. croaticum*, *L. hungaricum* – *L. zagrabiense*, *L. majeri* – *L. ellipticum*, and *L. petersi* – *L. baraci*. There has been much confusion about whether Brusina's names were junior synonyms of the Árpád species or valid independent taxa. For example, Brusina (1884) expressed the view that his *C. alata* is a synonym for *C. rhomboidea*, but subsequently (1897, 1902) depicted *C. alata* as a valid species. Stevanovic (1961) put an equals sign between *Lymnocardium hungaricum* (Hörnes) and *L. zagrabiense* (Brusina) but later treated them as distinct species (Stevanovic, 1990c). It is our opinion that the intraspecific variability in these bivalves has been usually underestimated; the occurrence of geographic variants does not necessarily justify introduction of new species names. In the two examples above, however, we simply do not have enough data to estimate the range of variability, so conclusions about the validity of these species would be premature (Table 1).

Biostratigraphic evaluation of the Árpád mollusk fauna

The mollusk fauna of the modern Árpád outcrop contains both littoral and sublittoral elements, the latter being especially common in layer 9. In terms of sublittoral mollusk biostratigraphy, the Árpád fauna belongs to the *Congeria rhomboidea* Zone, the youngest biostratigraphic unit of the Lake Pannon sequence (Magyar et al., this volume). The presence of some littoral cardiid species in the outcrop, however, allows us to set up a hypothesis about the more specific stratigraphic position of the Árpád outcrop within the *Congeria rhomboidea* Zone.

Lymnocardium arpadense and *Lymnocardium diprosopum* are undeniably closely related, yet distinguishable species (Basch 1990; Lennert et al., this volume). What is the nature of their relationship? We suggest that *L. arpadense* is a descendant of *L. diprosopum*, for the following reasons. The first representative of this lineage is *L. subdiprosopum* in the *L. conjungens* Zone (Fig. 3A). Through time, *L. diprosopum* became larger, and disproportionally higher, as indicated by specimens from subsequently younger localities (Fig. 3B, C). The size and overall morphology of *Lymnocardium arpadense* (Fig. 3D) suggests that it is the end-member of this lineage. The alternative to this hypothesis is that *L. diprosopum* and *L. arpadense* are in fact one and the same species, growing large under certain conditions and remaining small under others (Lőrenthey 1894). We have two reasons to doubt this suggestion. First, because both *L. arpadense* and *L. diprosopum* characteristically occur in littoral sands, there seems to be no geologic argument in favor of different habitats. Secondly, juvenile individuals of *L. arpadense* already display features that distinguish them from *L. diprosopum* of the same size; they have fewer ribs and more robust hinge

Table 1

Species with Árpád as type locality	Synonyms
Congeria rhomboidea M. Hörnes 1867 p. 364, pl. 48, fig. 4	?*C. alata* Brusina 1872 *C. oppenheimi* R. Hörnes 1901 *C. hilberi* R. Hörnes 1901 *C. rhomboidea dilatata* Stevanović 1961
Dreissenomya intermedia Fuchs 1873 p. 23, pl. 3. figs 1–6	
Lymnocardium arpadense (M. Hörnes) 1862 p. 198, pl. 29, fig. 1	
Lymnocardium haueri (M. Hörnes) 1862 p. 198, pl. 29, fig. 1	
Lymnocardium hungaricum (M. Hörnes) 1862 p. 194, pl. 28, fig. 2 (excl. fig. 3)	?*L. zagrabiensis* (Brusina) 1884
Lymnocardium majeri (M. Hörnes) 1862 p. 195, pl. 28, fig. 5	*L. ellipticum* (Brusina) 1872 *L. majeri multicosta* (Gillet) 1943 *L. majeri intermedia* (Stevanović) 1951
Lymnocardium petersi (M. Hörnes) 1862 p. 199, pl. 29, fig. 3	*L. baraci* (Brusina) 1884 *L. subsquamulosum* (Andrusov) 1903
Lymnocardium riegeli (M. Hörnes) 1862 p. 195, pl. 28, fig. 4	
Lymnocardium schmidti (M. Hörnes) 1862 p. 193, pl. 28, fig. 1	*L. croaticum* (Brusina) 1884 *L. patrulii* (Marinescu) 1973
"Paradacna" wurmbi (Lőrenthey 1893) p. 132, pl. 3, fig. 7	*"P." retovskii ossoinae* Stevanović 1951

teeth. In other words, *L. arpadense* is not simply a large *L. diprosopum;* it is a different species (Lennert et al., this volume).

If the origin of *L. arpadense* from *L. diprosopum* included branching, then we would expect to find fossils of these species together, or an occurrence of *L. diprosopum* above *L. arpadense* in outcrops and boreholes. In all of the publications or materials that we have seen, this is not the case. According to Basch (1990), among the 12 localities of *L. diprosopum* and 9 localities of *L. arpadense* in Croatia, none bears both species. From the area of Yugoslavia, Stevanovic (1990a, 1990b) reports *L. arpadense* only from two localities; in one of them, *L. arpadanse* occurs in a stratigraphically higher position then *L. diprosopum.* The other locality lacks *L. diprosopum* entirely. In Hungarian localities other than Árpád, *L. arpadense* and *L.* cf. *arpadense* have been reported from three boreholes (Korpás-Hódi 1982; Korpás-Hódi in Sütő-Szentai 1991; Korpás-Hódi et al. 1992). Here we disregard uncertain identifications ("*L.* cf. *arpadense*") because fragmentary specimens can be easily mixed up with *L. diprosopum.* In the Tengelic-2 borehole, *L. arpadense* appears above *L. diprosopum.* In the Kaskantyú-2 borehole, the latter species was not identified at all (in this borehole, the original publication (Korpás-Hódi et al. 1992) indicated *L.* cf. *arpadense;* according to Korpás-Hódi, (pers. comm.) however, some of these

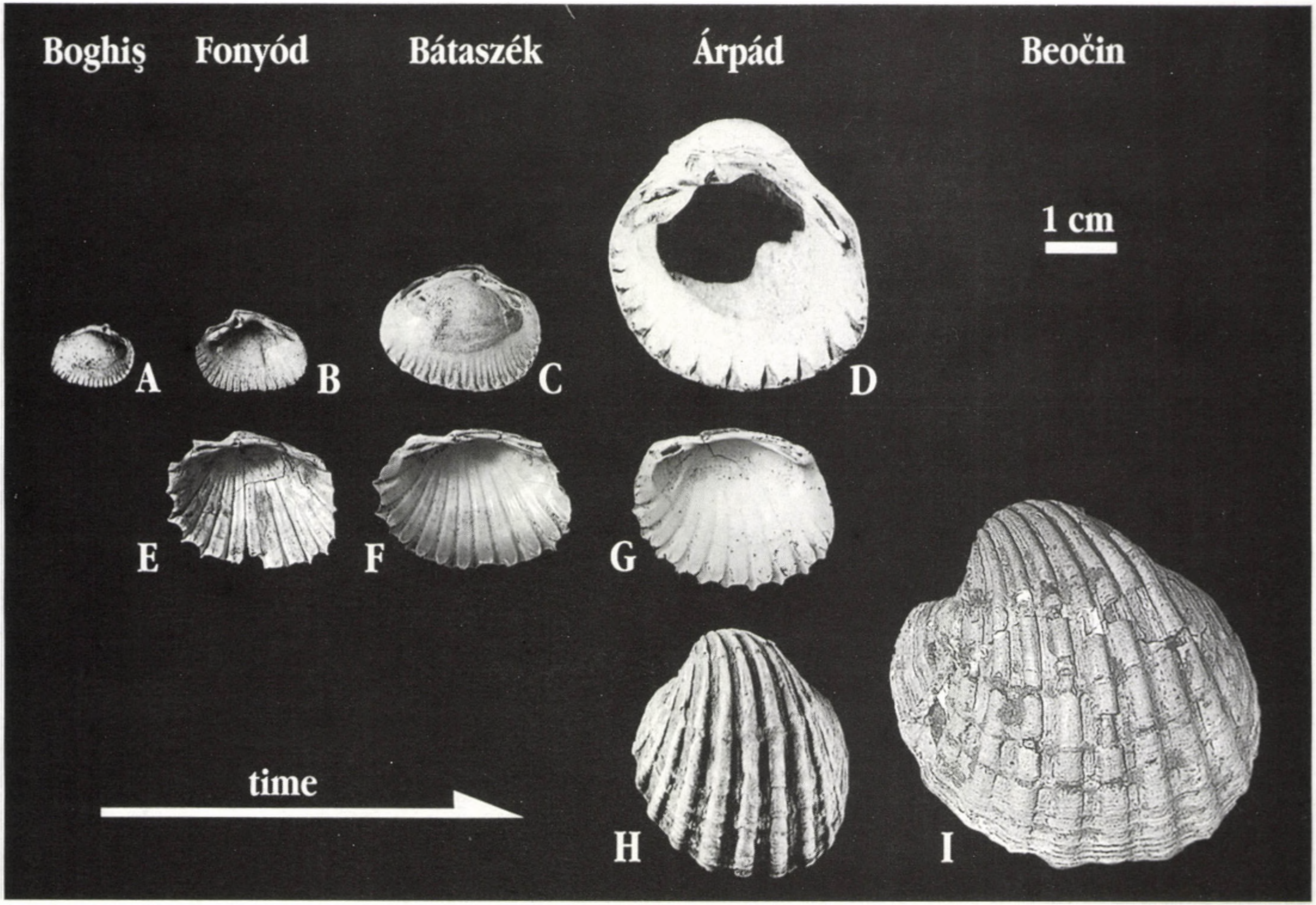

Fig. 3
Evolutionary lineages of cardiids based on the hypothesized stratigraphic position of the Árpád fauna. A – *Lymnocardium subdiprosopum* Stevanovic, B, C – *L. diprosopum* (Brusina), D – *L. arpadense* (Hörnes), E, F, G – *L. rogenhoferi* (Brusina) H, I – *L. petersi* (Hörnes)

samples almost certainly belong to *arpadense*). There is only one unpublished datum that does not fit our hypothesis. From the Paks-3 borehole, Korpás-Hódi (1987) reported *L. arpadense* from below the occurrence of *L. diprosopum*. We have not had the opportunity to check this material, however, but we expect that the occurrence of transitional forms between *L. diprosopum* and *L. arpadense* pose difficulties in identification. Therefore, we hypothesize that *L. arpadense* originated from *L. diprosopum* by anagenesis, a phenomenon apparently common in Lake Pannon mollusks (Geary 1990; Müller and Magyar 1992). It follows from this hypothesis that localities of *L. diprosopum* such as Bátaszék (Lennert et al., this volume) are older than localities with *L. arpadense*.

Biostratigraphic evidence from dinoflagellates and ostracods supports this temporal ordering of localities. Although a sample from the Árpád layer 9 did not contain dinoflagellates, in the Kaskantyú-2 borehole *L. arpadense* occurs in the uppermost part of the *Galeacysta etrusca* Zone (cf. Korpás-Hódi et al. 1992 and Sütő-Szentai 1994). In contrast, the *L. diprosopum*-bearing layers of Bátaszék belong to the lower part of the *Galecysta etrusca* Zone (=*Spiniferites tihanyensis* Zone; Lennert et al., this volume). Among ostracods, the abundance of *Candona (Lineocypris) branka* suggests that this fauna is probably younger than the Bátaszék one (Lennert et al., this volume.) This species has been found in only one other locality in Hungary: in the Nagyharsány-1 borehole, 105 to 170 m, where it correlates to the uppermost part of the *Galeacysta etrusca* Zone (cf. Sütő-Szentai 1994).

If the Árpád fauna is younger than that from Bátaszék, the gradual nature of evolution in two other lineages becomes clear. Specimens of *L. rogenhoferi* from Fonyód, Bátaszék, and Árpád consistently differ from each other (Fig. 3E–G). Fonyód is clearly the oldest of these localities; if we accept that Bátaszék is older than Árpád, the lineage exhibits gradual morphological change; the height of the umbo above the hinge gradually decreases, and a second (upper) posterior lateral tooth develops in the right valve (Fig. 3E–G).

In terms of their size, the Árpád specimens of *L. petersi* (Fig. 3H) occupy a median position between the smaller Okrugljak specimens (named *L. "baraci"*; see Basch 1990) and the significantly larger specimens from Slavonia and Syrmia (Fig. 3I). In the latter area, large *L. petersi* occurs together with *Prosodacnomya vodopici* (Stevanović 1990a), the youngest representative of the *Prosodacnomya* lineage in the Pannonian basin (Müller and Magyar 1992; Magyar et al., this volume). Thus, increasing size over time seems to characterize the *L. petersi* lineage.

Acknowledgements

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Appendix

List of the mollusk and ostracod species identified by the authors from the modern Árpád sand pit

Bivalvia

Congeria rhomboidea Hörnes
Dreissena sabbae Brusina
Dreissenomya intermedia Fuchs
Lymnocardium arpadense (Hörnes)
Lymnocardium haueri (Hörnes)
Lymnocardium hungaricum (Hörnes)
Lymnocardium majeri (Hörnes)
Lymnocardium pelzelni (Brusina)
Lymnocardium petersi (Hörnes)
Lymnocardium riegeli (Hörnes)
Lymnocardium rogenhoferi (Brusina)
Lymnocardium schmidti (Hörnes)
"Pontalmyra" otiophora (Brusina)
"Paradacna" wurmbi (Lőrenthey)
"Paradacna" aff. *wurmbi* (Lőrenthey)
Paradacna okrugici (Brusina)
Caladacna steindachneri (Brusina)
Pteradacna pterophora (Brusina)
Parvidacna chartacea (Brusina)
Phyllocardium planum (Deshayes)

Gastropoda

Valvata sp.
Micromelania sp.
Gyraulus constans (Brusina)
Radix sp.
Valenciennius sp.
Boskovicia sp.
Zagrabica sp.

Ostracoda (from layer 9)

Hemicytheria josephinae (Zalányi)
Candona (Caspiolla) parabalcanica Krstić
Candona (Caspiolla) cf. *ossoinae* Krstić
Candona (Sirmiella) arcuatoides Krstić
Candona (Zalanyiella) cf.longissima Krstić
Candona (Serbiella) hastata Krstić
Candona (Pontoniella) paracuminata Krstić
Candona (Hastacandona) sp.
Candona (Lineocypris) branka Krstić
Candona (Thaminocypris) alta (Zalányi)
Candona (Bakunella) balcanica (Zalányi)
Candona (Sinegubiella) sublabiata Krstić
Candona (Typhlocypris) sp.
Candona (Typhlocyprella) lineocypriformis Krstić
Hungarocypris sp. (juvenile specimen)
Amplocypris nonreticulata Krstić

Plates

The following abbreviations are used: NHMW: Naturhistorisches Museum Wien (original material of Hörnes); TTM: Természettudományi Múzeum, Budapest.

Where no abbreviation is used, the material is from the authors' collection. Where not indicated otherwise, scale bars represent 1 cm.

Plate I

1, 6. *Lymnocardium haueri* (Hörnes), NHMW
2. *Lymnocardium haueri* (Hörnes), NHMW
3. *Lymnocardium haueri* (Hörnes), NHMW
4. *Lymnocardium haueri* (Hörnes), Layer 6
5, 7. *Lymnocardium haueri* (Hörnes), NHMW

Plate II

1. *Lymnocardium arpadense* (Hörnes), NHMW
2. *Lymnocardium arpadense* (Hörnes), NHMW
3, 4,
5. *Lymnocardium arpadense* (Hörnes), NHMW
6, 7. *Lymnocardium arpadense* (Hörnes), from debris
8, 9. *Lymnocardium arpadense* (Hörnes), from debris
10, 11. *Lymnocardium arpadense* (Hörnes), from debris

6–11 have common scale bar

Plate III

1, 2. *Lymnocardium hungaricum* (Hörnes), NHMW
3, 5. *Lymnocardium hungaricum* (Hörnes), TTM
4. *Lymnocardium hungaricum* (Hörnes), TTM
6, 7, 8. *Lymnocardium rogenhoferi* (Brusina), NHMW
9. *Lymnocardium rogenhoferi* (Brusina), Layer 9
10. *Lymnocardium rogenhoferi* (Brusina), Layer 9
11. *Lymnocardium rogenhoferi* (Brusina), Layer 9

Plate IV

1, 2. *Lymnocardium schmidti* (Hörnes), NHMW
3, 4. *Lymnocardium schmidti* (Hörnes), Layer 6
5. *Lymnocardium petersi* (Hörnes), NHMW
6, 7, 8. *Lymnocardium petersi* (Hörnes), TTM
9, 10. *Lymnocardium petersi* (Hörnes), from debris

Plate V

1, 2. *Congeria rhomboidea* (Hörnes), NHMW
3. *Congeria rhomboidea* (Hörnes), NHMW
4. *Congeria rhomboidea* (Hörnes), NHMW
5. *Lymnocardium majeri* (Hörnes), NHMW
6. *Lymnocardium majeri* (Hörnes), Layer 9

Plate VI

1, 2. *Phyllocardium planum* (Deshayes), Layer 8
3, 4. *Pteradacna pterophora* (Brusina), NHMW
5, 6,
7, 8. *"Paradacna"* aff. *wurmbi* (Lőrenthey), with very narrow posterior gape, Layer 9
9. *Parvidacna chartacea* (Brusina), Layer 9 (scale bar indicates 5 mm)
10. *"Paradacna" wurmbi* (Lőrenthey), with wide posterior gape, Layer 8
11. *Caladacna steindachneri* (Brusina), with long spines on the primary ribs, Layer 9

Plate I

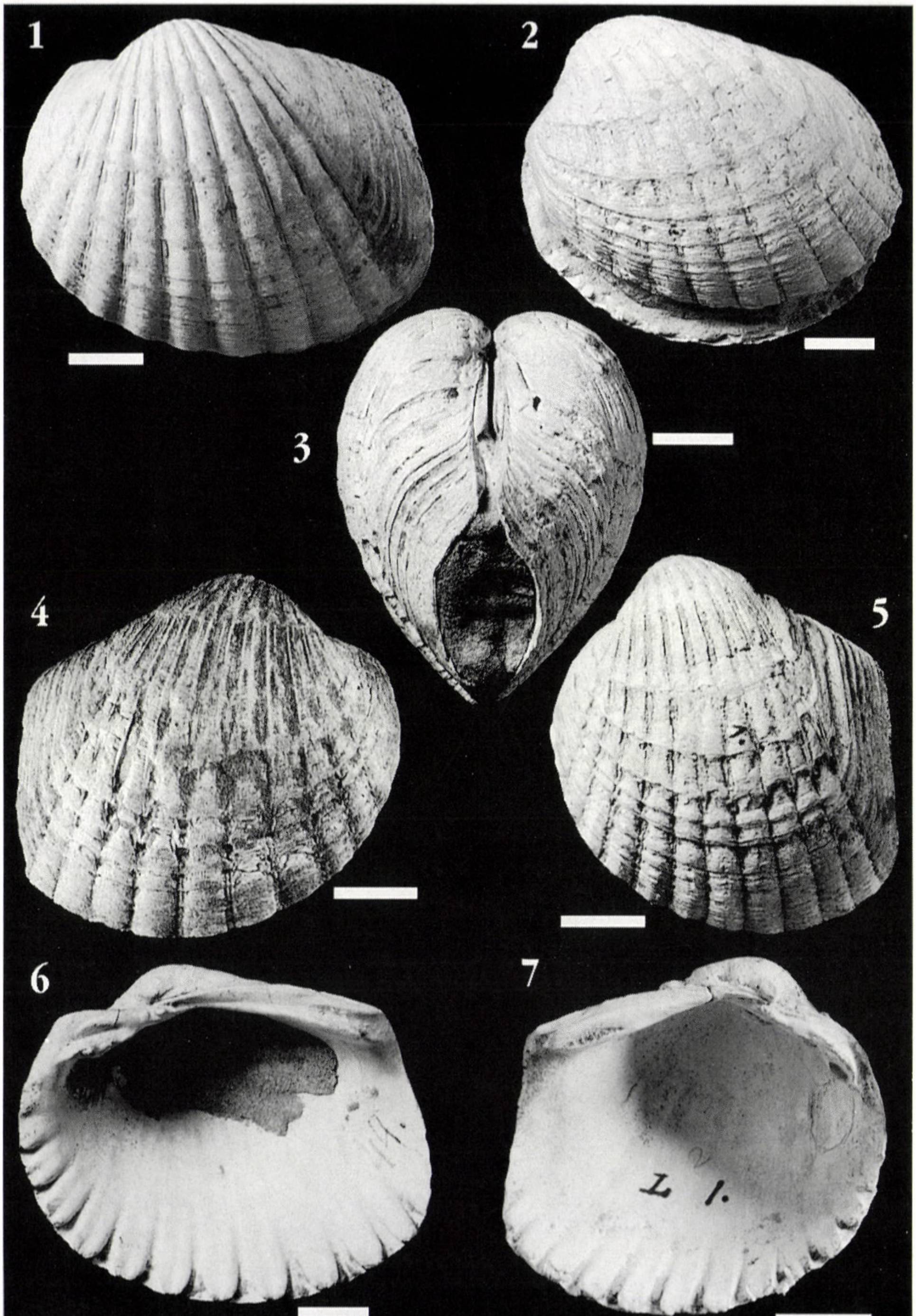
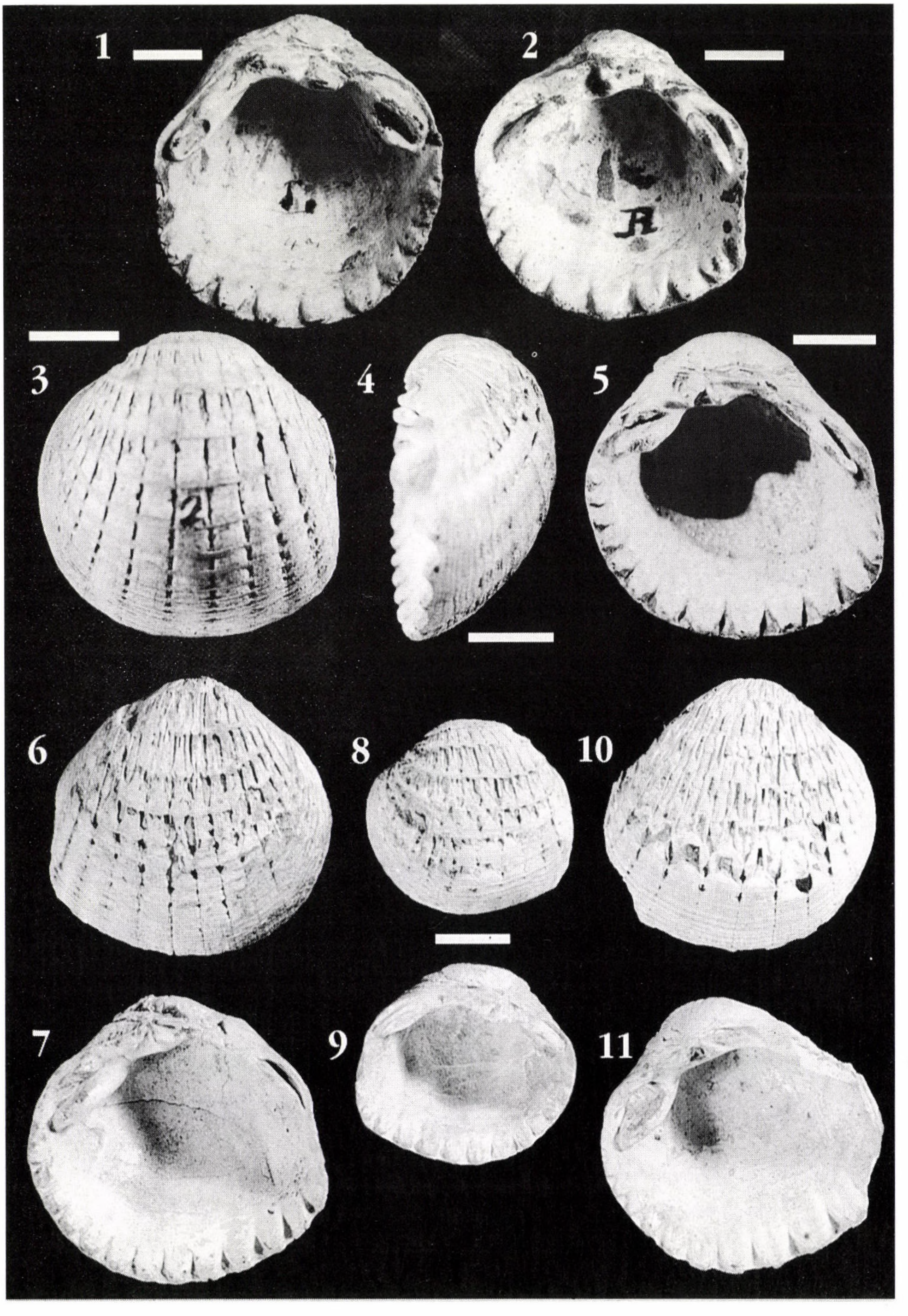


Plate II



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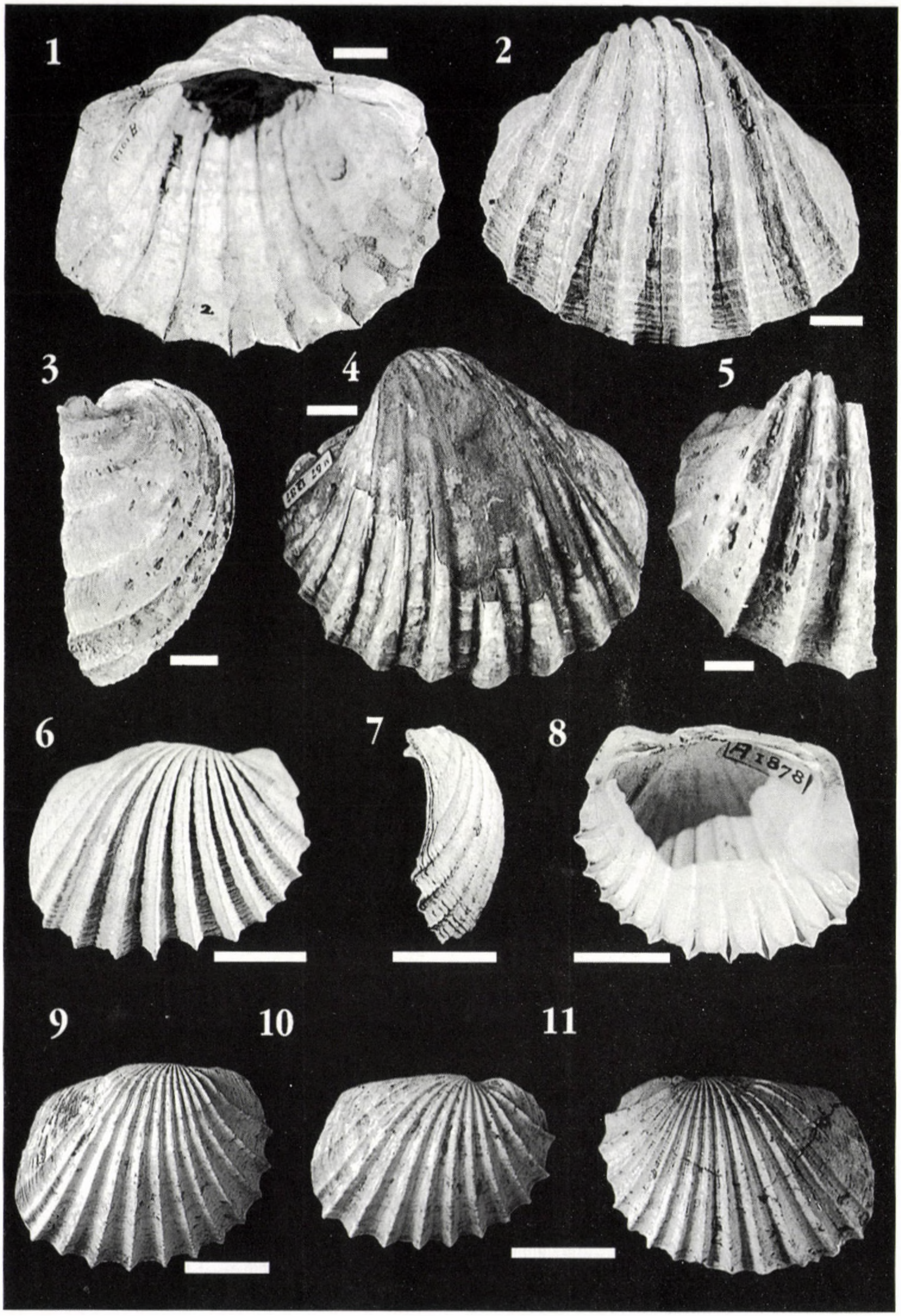

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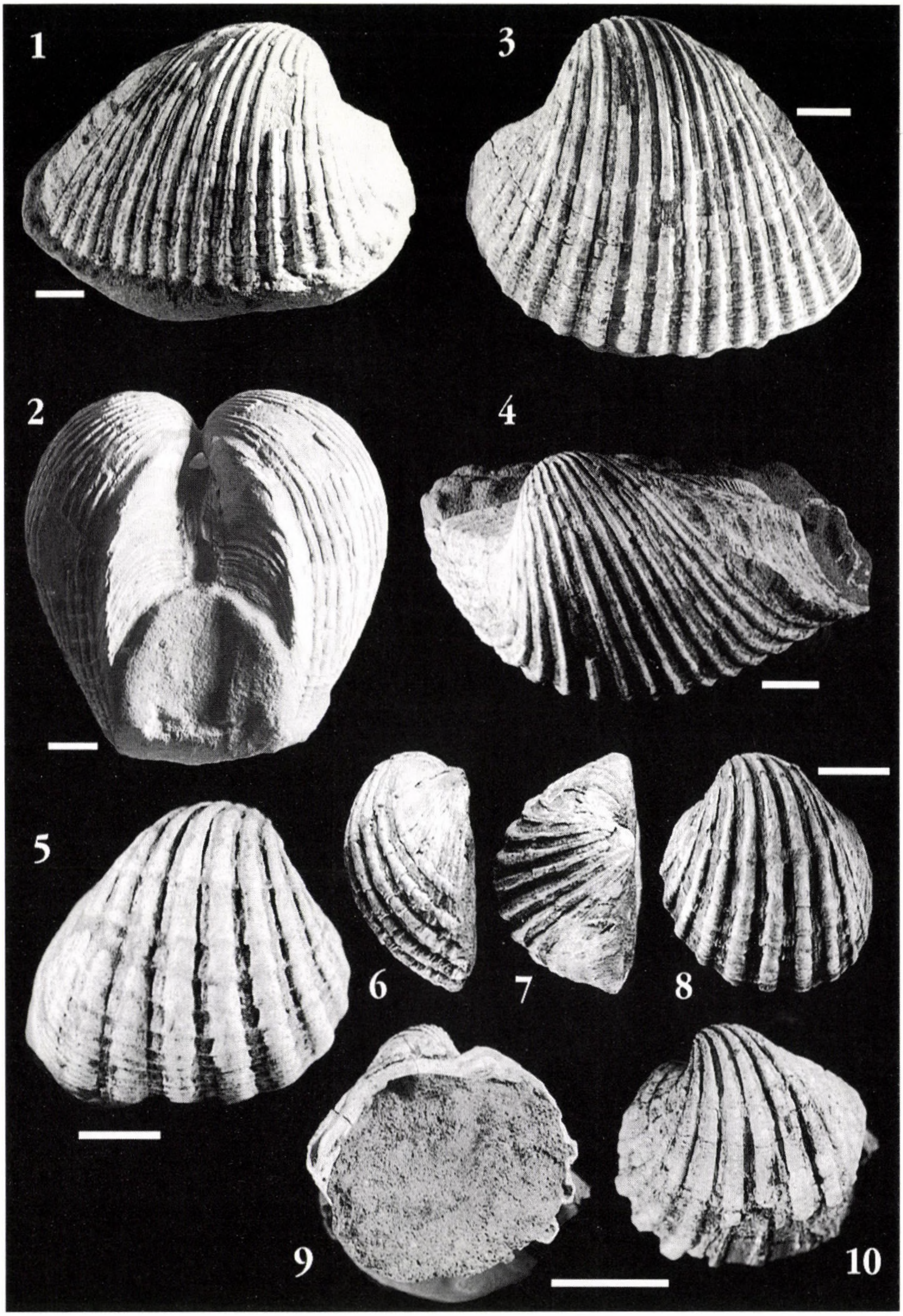

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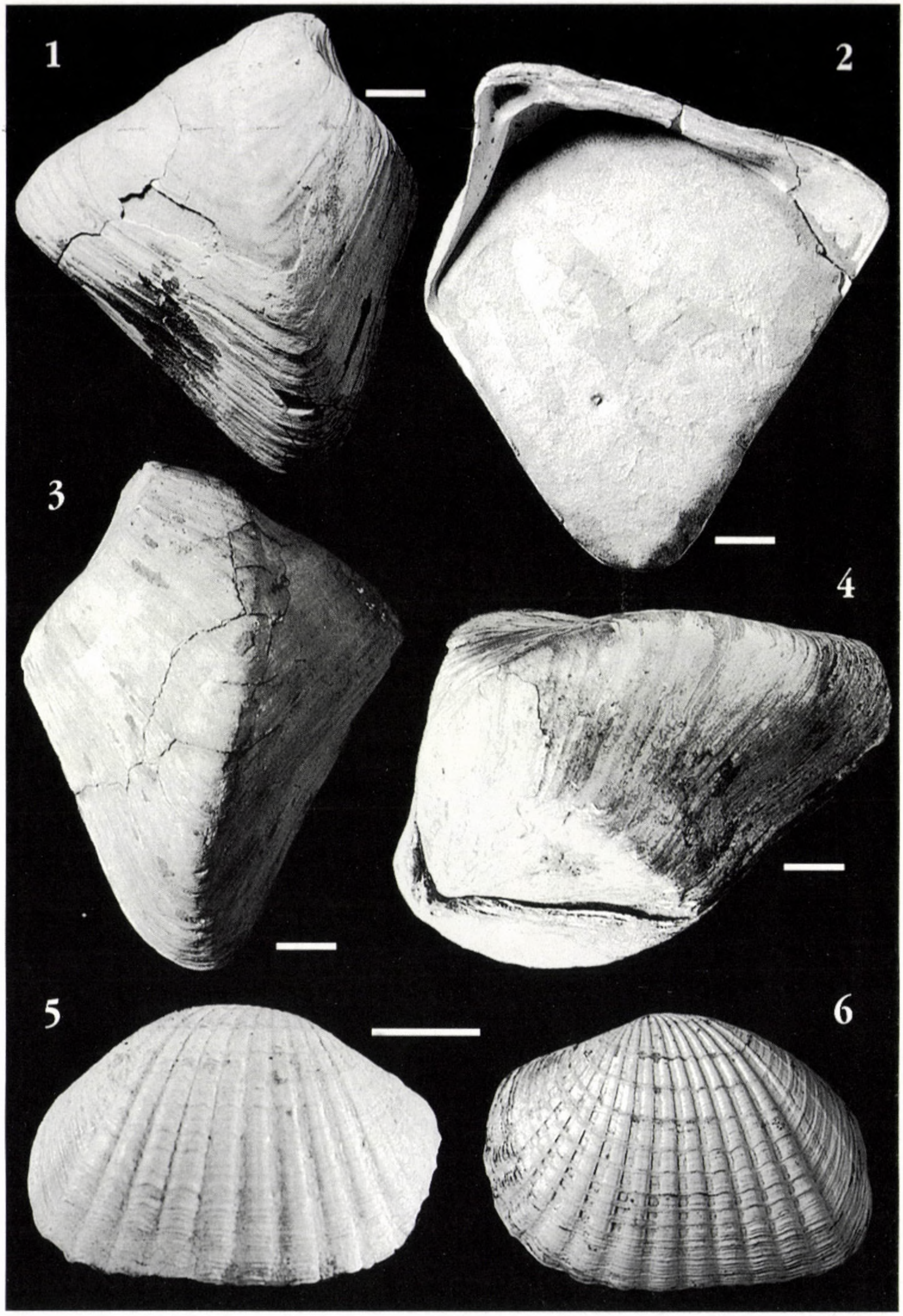

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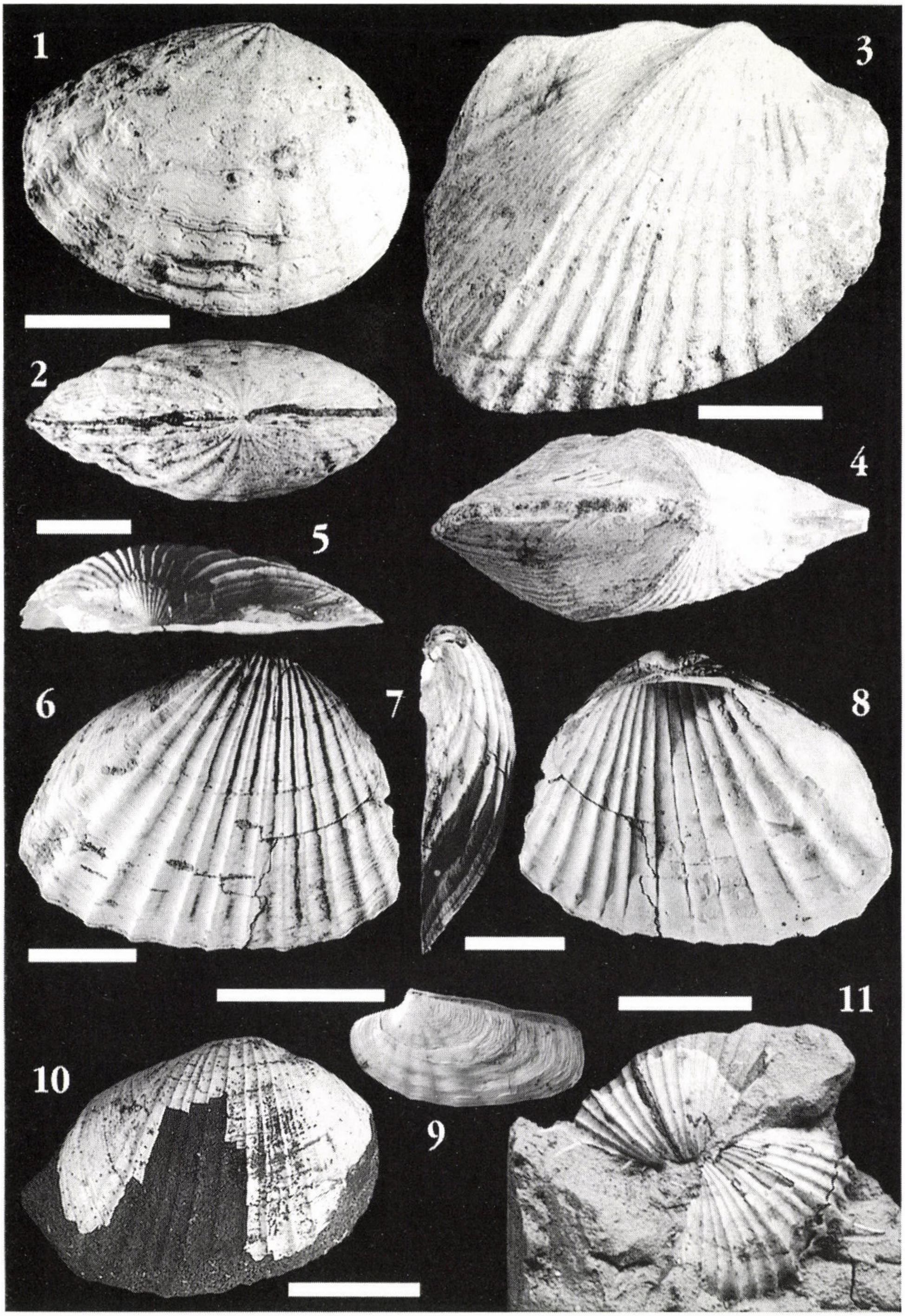

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Fossils and strata of Lake Pannon, a long-lived lake from the Upper Miocene of Hungary

Editors: Imre Magyar and Dana H. Geary

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History of the Carpathian–Balkan Geological Association

Gusztáv Morvai

The "Carpathian Geological Association" was founded in 1922, at the XIIIth International Geological Congress in Brussels, by representatives of Czechoslovakia, Poland, Romania and Yugoslavia.

Between the two World Wars only three congresses were held. It was at the IIIrd Congress in Prague (1931) that Hungarian geologists were present for the first time. At the XXth International Geological Congress in Mexico City (1956) it was decided to reconstitute and enlarge the Association. It was at this moment that Hungary joined the CBGA, along with Bulgaria and the USSR. After the IVth Congress of the CBGA (Kiev-Lvov, 1958) congresses took place every second year, and after 1969 every fourth year only. Hungary hosted a CBGA Congress in 1969. Austria joined the Association in 1977 and Greece in 1989.

At the congresses the geologists of the host country presented the results of the geologic research performed on the pertinent sector of the Carpathian-Balkan Mountain system in the form of summarizing lectures and field trips. The representatives of the other member countries also reported on their latest research results.

In the year 1958 six Standing Commissions were established; six more have been added since then. These committees are supposed to coordinate the joint efforts in the specific fields between one congress and the next. The Standing Commissions on Tectonics, Petrography, Geochronology, Sedimentology, Hydrogeology and Engineering Geology compiled and printed several important generalized maps of the region, while the Commission on Geophysics contributed considerably to the knowledge of the earth's crust by measuring and interpreting geophysical sections.

The CBGA also contributed to the activities of other international organizations. For example at CBGA Congresses there has also been a possibility to coordinate the activities of the European map-compiling commissions of the International Geological Congress (on Geology, Metallogeny, Hydrogeology etc.). The Tectonic Map of the CBGA Region served as a basis for the production of the Metallogenic Map of the COMECON. A considerable number of IGCP projects also organized meetings within the framework of CBGA Congresses.

Key words: science history, Carpathian–Balkan Region

Introduction

The "Carpathian Geological Association" was founded after World War I by the (partly) new states surrounding Hungary. Their combined territories included the entire Carpathian mountain system as well as its inner and outer forelands. However, the Memorandum of Foundation, dated 19 August 1922, set as an additional aim the study of the Sudetes and the Northern Balkan as well.

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After World War II part of the Carpathians and its forelands were incorporated into the USSR (i.e. the Ukrainian SSR), and most of the Balkan mountain system belonged to Bulgaria. Thus, when re-establishing the Association it was inevitable to accept the USSR and Bulgaria as well as Hungary (containing most of the inner foreland), as members of the CGA. It was appropriately renamed CBGA: "Carpathian-Balkan Geological Association". It is evident from the memoirs of the proposers that this extension of the Association in 1956 was not at all exempt of political considerations (nor was its foundation in 1922).

Between the two World Wars the driving forces of the Association were prominent geologists of the four founding countries. Among the initiators of the re-constitution there were several outstanding geoscientists who were "founding fathers" of the Association (such as **R. Kettner, K. Petkovic)** or organizers of the first three congresses **(G. Macovei, D. Andrusov, A. Codarcea)**. The new member countries were also represented by scientists of international reputation (Bulgaria by **E. Bonchev,** Hungary by **E. Szádeczky-Kardoss**, and the USSR by **N. P. Semenenko** and **E. K. Lazarenko**). The ensuing congresses (IV-XV) were organized, apart from the aforementioned, by additional renowned scientists such as **S. Wdowiarz, V. Ianovici, O. Fusan, B. Kamenov, P. Stevanovic, D. Radulescu, V. V. Naumenko, A. Grubic** and **D. Papanicolaou.** The active Hungarian contribution was assured by **G. Pantó, K. Balogh, V. Széky-Fux, A. Rónai, J. Fülöp, G. Hámor, G. Császár, P. Árkai, G. Morvai** and many others. After the IXth Congress (Budapest, 1969) more and more young geologists became involved in the CBGA, giving a new impetus to several fields of research, such as tectonics, sedimentology, mineralogy, isotope geochronology, hydrogeology, etc.

In the following an attempt is made to briefly outline the 75-year history of the C(B)GA.

Between the two World Wars

At the XIIIth International Geological Congress held in Brussels (Belgium) in 1922, representatives of Czechoslovakia, Poland, Romania and Yugoslavia agreed to establish a **Carpathian Geological Association**. The French-language Memorandum was signed by **R. Kettner, J. Nowak, G. Murgoci** and **V. Petkovic.**

The founders defined the **goals and tasks** of the Association as follows:

1. To facilitate geologic research along the state boundaries,
2. To cooperate regularly with geologists of the other member countries,
3. To organize conferences and excursions (field trips),
4. To facilitate the exchange of publications, samples, etc.

The first three congresses took place in:
I. Lwów (Poland), 1925
II. Bucharest (Romania), 1927
III. Prague (Czechoslovakia), 1931

Attendance varied between about 65 and 75, of which about 20 were foreign nationals. Hungarian geologists were present only at the IIIrd Congress in Prague: **H. Böckh** and **Gy. Rakusz**. The Congresses included interesting **field trips**:
I. Oil fields of the Polish Carpathians from Borislaw to the border with Romania.
II. Oil fields of Ploesti and the Carpathian nappes.
III. Bohemian Massif (pre-congress), Slovak Volcanic Range (post-congress).

After the IIIrd Congress (1931) the activities of the Association were interrupted because of the unexpected death of **K. Petkovic** and of the deteriorating political atmosphere in Europe. They were assumed again only ten years after the end of World War II.

After World War II

At the XVIIIth International Geological Congress in London (1948) the idea was brought up to revive the Carpathian Geological Association. But it was only eight years later, at the XXth IGC held in Mexico City (1956) that the leaders of the Czechoslovak and the Romanian delegations (**D. Andrusov** and **A. Codarcea**, respectively) formally proposed to re-establish the activities of the CGA, enlarging it with the territory of the Alps (!) and eventually of the Balkan Range. During the discussion, Hungary was represented by **G. Pantó.** Since no agreement could be reached about the participation of the Alpine countries, the field of membership of the Association was extended to the Balkan only. Bulgaria, Hungary and the USSR became member states of the newly renamed CBGA.

The IVth Congress of the CBGA

The USSR accepted to organize the first congress of the CBGA. In order to underline the legal continuity with the former CGA, it was numbered as the IVth. The USSR was in fact represented by the Ukrainian SSR.

The field trip of the Congress held in Kiev and Lvov (1958) crossed the NE Carpathians. The President of the Congress was **E. K. Lazarenko**. There were altogether 255 participants, including a delegation of six Hungarians headed by **E. Szádeczky-Kardoss**.

The most important decision of the IVth Congress was to create six **Standing Commissions**, namely on Tectonics, Stratigraphy-Paleogeography and

Paleontology, Igneous Petrology, Mineralogy and Geochemistry, Hydrogeology, and Geologic Maps. Each Commission was sponsored by one of the member countries. Hungary was entrusted with the Standing Commission on Igneous Petrology, with **E. Szádeczky-Kardoss** as its chairman.

The Vth Congress of the CBGA was held in Bucharest (Romania) in 1961. Altogether 145 papers were presented in five sections. There were 84 official delegates and 175 other attendants. The Hungarian delegation, headed by **E. Szádeczky-Kardoss,** numbered eight. The Honorary President of the Congress was **G. Macovei**, its President **A. Codarcea,** with **V. Ianovici** as the Secretary General. Three concurrent post-congress field trips were organized to visit the Neogene volcanic rocks and mineralization in the Baia Mare (Nagybánya) region, the Cretaceous flysch of the Eastern Carpathians, and the nappes of the Southern Carpathians, respectively. A joint excursion followed to the Dobrogea and to the delta of the Danube. The closing plenary session took place in the coastal resort of Mamaia.

At this congress two additional Standing Commissions were established: one on Geophysics and another on Sedimentology, as well as three **Subcommissions**: on Absolute Age and Isotope Geology, on Geothermal Energy and on Engineering Geology.

During the Congress, the Commission on the Geological Map of Europe conducted a business meeting chaired by **A. Bogdanov.**

The VIth Congress of the CBGA took place in Poland, in 1963. At the opening plenary session in Warsaw four talks were given about the stratigraphy, the structural setting and the mineral deposits of the Flysch Belt of the Polish Carpathians. Further introductory lectures were delivered in Cracow on the oil and gas accumulations in the Carpathians and their foredeep. The other papers were presented in four thematic sections. The participants of the ensuing field trips got acquainted with the geologic setting of the Carpathians, the Tatra and the Pieniny Mountains.

This Vth Congress had 270 participants, including 147 from abroad. The President was **M. Mrozowski**, the Secretary General **S. Wdowiarz**. Two hundred and ten papers were read (140 by foreign nationals). The Hungarian delegation with only eight-members was headed by **F. Benkő**.

The most important decisions taken by the Standing Commissions were

– to produce a 1:1,000,000 scale Tectonic Map of the Carpathian-Balkan-Dinaric region,

– to begin acquiring eight deep geophysical sections to explore the crust.

M. Mahel, the chairman of the Standing Commission on Tectonics, was charged with the coordination of the first task and **D. Prosen,** the chairman of the Commission on Geophysics, with the organization and implementation of the second.

The VIIth Congress of the CBGA was organized in Sofia (Bulgaria) in 1965. At the opening plenary session **E. Bonchev** spoke on the geologic setting of Bulgaria and **J. Jovchev**, the President of the Congress, on the mineral resources of the country. Altogether 262 papers were presented in six sections. Special sections were devoted for the first time to coal and hydrocarbons. Field trips crossed the Balkan Mountains (Stara Planina) along five sections. Others presented the ore deposits of the Sredna Gora Mountains and of the Varna-Burgas area.

A considerable part of the papers was printed *before* the Congress. There were more than 500 participants; *E. Szádeczky-Kardoss* was the head of the 17-member Hungarian delegation.

During the congress, the Middle and East European sheets of the Metallogenetic Map of Europe were coordinated under the direction of **P. Lafitte**.

The decisions of the Congress granted support to the most important initiatives of the Commissions:

– to produce tectonic, hydrogeological and coal geologic maps,

– to edit tectonic and sedimentological dictionaries,

– to compile a mineralogical encyclopaedia of the region,

– to increase the number of mineralogical, isotope geochronological and paleomagnetic investigations.

The VIIIth Congress of the CBGA was held in Belgrade (Yugoslavia) in 1967. The plenary session was chaired by **M. Lukovic**. An overview of the geologic formations in the Yugoslav sector of the Carpathian and Balkan Mts. was presented by **P. Stevanovic**. Subsequent papers were read in seven sections. The pre-congress field trips went across the Dinaric Mountains, and the post-congress one visited the terrains and mineral deposits of the Carpathian-Balkan Mountain system and of the Macedonian Massif.

The Commission of the Geological Map of Europe held a business meeting chaired by **P. Walter**.

Decisions taken at the Congress urged the summarizing and assessment of the data presented by the lectures. Moreover, they fostered the activities of the Committees: explanatory notes to the hydrogeological, engineering geologic, metallogenetic and paleo-transportation maps, geologic crustal profiles, etc. It was at this occasion that it was decided to produce a Map of Metamorphites of the CBGA region. A number of decisions dealt with problems of geochronology and paleomagnetism.

The IXth Congress of the CBGA was organized by Hungary, in 1969. The Congress held in Budapest hosted 386 people from 16 countries; the number of foreign nationals was 276. In conformity with the decisions of the preceding VIIIth Congress there were seven thematic sections. The keynote speaker of Section I was **E. Szádeczky-Kardoss**, the President of the Congress. He discussed the relationship of metamorphism and tectonics in the Carpathian-Balkan-Dinaric region. In Section II **G. Pantó** gave a lecture about the genesis of post-volcanic rocks and mineralization. In Section III **J. Fülöp**, the Secretary General of the congress, spoke about facies maps and facies studies. In Section IV **M. Földvári-Vogl** presented a paper on the distribution of trace elements in sedimentary rocks, in Section V **V. Dank** on the generation and migration of hydrocarbons, in Section VI **A. Rónai** on the engineering geology and hydrogeology of lowlands, and in Section VII **L. Egyed** on regional geophysics (all of these concerned the Carpathian-Balkan-Dinaric region). All papers were published in vols XIII–XV. and in a special issue of the **Acta Geologica** of the Hungarian Academy of Sciences.

Three symposia were held within the framework of the Congress, namely on Radiometric Geochronology, on Pre-Carboniferous Stratigraphy, and on Geophysics.

Prior to the congress, the participants had the possibility to attend a jubilee session celebrating the 100th anniversary of the foundation of the Hungarian Geological Institute (MÁFI) and the related colloquia on Jurassic, Eocene and Neogene Stratigraphy and Bauxite Geology, with ensuing field trips.

There were three regional post-congress field trips: to the Transdanubian Central Range, to Northern Hungary, and to the Lowlands, as well as two thematic ones: Oil Fields and Geophysics.

The decisions taken at the Congress defined some high priority tasks:

– The Commission on Tectonics, having finished the 1:1,000,000 scale tectonic map, set itself the goal to plot a 1:500,000-scale version;

– The Commission on Mineralogy and Geochemistry urged the preparation of a metallogenetic map and the investigation of the rare metal and microelement contents of igneous and post-volcanic formations;

– The Subcommission on Geochronology undertook to produce an inventory and a Geochronological Map of the Basement of the Carpathians and the Balkan;

– The Commission on Igneous Rocks highlighted the geologic-geophysical interpretation of deep crustal sections (traverses);

– The Commission on Stratigraphy proposed to organize symposia dealing with stratigraphic problems of the Triassic, the Jurassic, the Lower Cretaceous and the Flysch Zone;

– The Commission on Sedimentology continued to prepare paleo-transportation maps of various Cretaceous horizons;

– The Commission on Hydrogeology and Engineering Geology urged the completion of the national drafts of the 1:1,000,000 scale maps by the year 1970, and

– The Commission on Geophysics decided to continue the crustal and paleomagnetic measurements.

The Xth (jubilee) Congress (50th anniversary) of the CBGA was held in Czechoslovakia in 1973. It recorded 584 attendants, including 253 foreign nationals. The Hungarian delegation (50!) was led by **E. Szádeczky-Kardoss.** **D. Andrusov** was the Honorary President, **O. Fusan** the President, and **O. Samuel** the Secretary of the Congress. There were eleven (!) pre-congress field trips, of which eight dealt with the stratigraphy, tectonics and mineral resources of the Western Carpathians.

The opening plenary session took place in Bratislava. **D. Andrusov, O. Fusan,** and **J. Slavik** presented the geologic and structural setting of Slovakia and the achievements in mineral exploration. **M. Mahel** commented on the Tectonic Map of the Carpathian-Balkan-Dinaric Region. Other papers were read in eight sections. A post-congress field trip demonstrated once more the stratigraphy and tectonics of the Western Carpathians.

The Presidium of the Congress (the Council of the CBGA) stated with satisfaction that the activities of the Association were developing satisfactorily. Non-member countries were displaying growing interest. Joint study of scientific problems of the region and involvement in global problems were defined as the new aims after the completion of the thematic maps.

The XIth Congress of the CBGA was organized in Kiev (1977) by the Ukrainian Academy of Sciences and the Ukrainian Ministry of Geology. **N. P. Semenenko** was the President, and **V. V. Naumenko** the Secretary General. There were more than 500 people attending. The Hungarian delegation of 58 was headed by the author of the present paper (**G. Morvai**). At the plenary session **N. P. Semenenko** spoke on the geochemical model of the Earth's geospheres and **O. S. Vialov** on some fundamental geologic problems of the northeastern Carpathians. Altogether 252 papers were read, including 26 by Hungarians.

The five pre-congress field trips had a total of 180 participants. One geophysical excursion introduced the seismicity of the Crimean peninsula, the others were focussed on the stratigraphy and tectonics, the hydrocarbon provinces, the hydrogeology and the solid mineral resources of the northeastern Carpathians and of their foredeeps. **P. Árkai** presented the 1:1,000,000 scale Map of Metamorphites of the CBD region edited by **E. Szádeczky-Kardoss** and himself, **B. Bogdanov** the 1:1,000,000 scale Metallogenetic Map, and **N.P. Semenenko** the 1: 2,500,000 scale Isotope Geochemical Map of the Crystalline Basement.

F. Ronner announced the intention of Austria to join the CBGA.

After the establishment of the International Union of Geological Sciences (IUGS) the idea of affiliation was repeatedly brought up. On the initiative of **Gy. Grasselly,** the Hungarian Vice-President of the IUGS, at an extraordinary meeting of the CBGA Council held in Hungary the representatives of the member countries accepted the proposal that the CBGA become an affiliate of the IUGS. After approval by the National Committees of the CBGA, the Council of IUGS also agreed. The affiliation was ratified by the XXVth International Geological Congress held in Paris (France) in 1980.

The XIIth Congress of the CBGA was held in Bucharest (Romania) in 1981, organized by the University of Bucharest and the Ministry of Geology. **D. Radulescu** was the President, and **M. Sandulescu** the Secretary General. Six pre-congress and three post-congress field trips had been planned, but some of them had to be cancelled or combined, due to lack of sufficient interest. Even then the excursion dealing with the subduction-induced magmatism of the Romanian Carpathians had only nine foreign participants, including four from Hungary. A similarly small group took part in the trip demonstrating the flysch and molasse belts of the Eastern Carpathians. Nevertheless, the Congress itself had nearly 500 participants. The 46-member Hungarian delegation was headed by **G. Morvai**. Altogether 192 lectures were held, including 31 by Hungarians and 53 by other foreigners. However, more than the half of the lectures were held by Romanians. The Bucharest Congress approved the affiliation of the CBGA to the IUGS and accepted the proposal formulated at the previous Congress in Kiev to set up a Commission on Mineral Deposits.

The XIIIth Congress of the CBGA was held in Cracow, 1985. The plenary session was chaired by **Z. Dembowski**. A temporary Scientific Committee had been set up for the preparation of the Congress (Chairman: **S. Wdowiarz,** Secretary: **D. Poprawa**). The introductory keynote lectures dealt with new data about the structure of the Carpathian flysch, the relation between the Inner and the Outer Carpathians, and with new information on the Carpathian foredeep and its basement. The Congress was attended by 410 people, of which 258 were foreign nationals. **G. Morvai** was the leader of the Hungarian delegation of 61 experts.

There were five pre-congress field trips to view the terrains, the structural setting, and the hydrogeology-engineering geology of the Polish Carpathians and the Carpathian foredeep, with special attention to the Quaternary outcrops. The program had listed 368 papers, but due to the absence of the authors more than 100 were not presented. The organizers regrouped the remaining lectures, to general dissatisfaction, since the schedule became fairly chaotic.

The representative of Poland proposed to set up a separate Commission on Quaternary Geology, and that of the Ukrainian SSR the creation of a Commission on Petroleum Geology. Both proposals were, however, rejected by the Council of the CBGA. The Council changed the sponsorship of several Commissions. The chairmanship on the Commission of Stratigraphy was transferred to Hungary, that of the Commission on Igneous Rocks to Romania, and that on Metamorphic Rocks to Austria.

The XIVth Congress of the CBGA was organized by the Bulgarian Academy of Sciences in Sofia in 1989. **M. Zhelyaskova-Panayotova** became the President of the CBGA Council, with **K. Kolcheva** as its Secretary General. At the plenary session **E. Bonchev** could not read his keynote speech, because he was already seriously ill. **M. Mahel** spoke about the evolution and problems of the

Carpathian-Balkan orogen, and **R. D. Dokov** presented the new Metallogenetic Map of Bulgaria.

The Congress registered a record attendance of 795 (of which 289 came from abroad.) The 48-member Hungarian delegation was headed by **G. Morvai.** Three hundred and seventy-two papers were read in ten sections, including 30 by Hungarians. There were also symposia held in conjunction with the Congress. Four of these aroused considerable interest: "Study of flysch sediments and turbidites", "Conditions of the genesis of sedimentary mineral deposits", "Study of the coal formations in the Carpathian-Balkan region", and "Protection of non-living nature". Several IGCP Projects held meetings in the framework of the Congress, as well as the representatives of the IGCP National Committees of the region.

The Congress accepted the application of Greece for membership in the CBGA. In addition, for the first time in the history of the Association there was an observer from Albania.

Yugoslavia had agreed to organize the next (XVth) Congress in Belgrade in 1993. This was, however, rendered practically impossible by the wars and international sanctions that followed the disintegration of Yugoslavia. Belgrade stepped back. This was formally announced by **S. Karamata** at an extraordinary meeting of the CBGA Council held on 19 May 1993 in **Budapest**. Instead, the newest member country, Greece, volunteered to organize the Congress, but only in 1995. The Council accepted the offer with gratitude, being fully aware of the new challenges brought about by the radical political changes in eastern and Central Europe.

The XVth Congress of the CBGA was held in Athens (Greece) in September 1995. Its officers were **D. Papanicolaou** (President), **Th. Markopoulos** (Vice-President), and **G. Migiros** (Secretary General). Beside several institutions of higher education, a large number of research units also took part in the organization, such as the National Center of Marine Research, the Institute of Mining and Geological Exploration (IMGE), the Institute of Oceanography, and the Institute of Earthquake Prevention and Prediction. At the opening plenary session **D. Papanicolaou** spoke about the past and the near future of the CBGA. Among the approximately six hundred participants there were 17 Hungarians. **G. Császár** had been delegated to represent Hungary in the Council of the CBGA. There were twelve sections (corresponding to the 12 Standing Commissions), three symposia and three IGCP project meetings. One hundred and seventy-two papers were presented at the sessions, and 117 at the symposia and the IGCP meetings. The symposia dealt with the tectonostratigraphy, the seismicity, and the energy resources of the Carpathian-Balkan region. IGCP Project 262 dealt with the correlation of the Tethyan Cretaceous. IGCP Project 276 presented the map of terranes of the Alpine-Himalayan Belt, and IGCP Project 356 the plate tectonic relations of Alpine metallogeny in the Carpathian-Balkan region. There were altogether 18 papers by Hungarian authors. This was the first Congress of the CBGA at which nature and environmental protection were the topics of a separate section.

The modified Statutes of the CBGA were ratified, accepting the modifications that had been proposed and discussed at the 1993 Council Meeting in Budapest, and reviewed at Thessaloniki in 1994. Austria accepted the task of organizing the next (XVIth) Congress in 1998 in Vienna.

The Statutes and Their Modifications

The first version of the Statutes were adopted at the closing session of the Ist Congress held in Lvov in 1925. It defined the denomination, the aims, the member states, the official language and the emblem (logo) of the association. Special accent was put on the link with the International Geological Congress. This was reflected by the organizational structure as well (Congress, Council, Delegates).

In 1958 the Statutes were revised and modified. The world "**Balkan**" was inserted into the denomination. Bulgaria, Hungary and the USSR were added to the list of member countries. The structure was enlarged by six Standing Commissions.

At the Vth Congress in Bucharest (1961) the organization was further developed by establishing more Standing Commissions and one Sub-commission. "Engineering Geology" was added to the denomination of the Commission for Hydrogeology.

In 1973, at the Xth Congress held in Bratislava, the aims were completed by adding the research activities in applied geology, such as Petroleum Geology and Metallogeny. Some issues concerning the organization, the functions of the Congresses and the Commissions were formulated more precisely. It was urged to nominate corresponding secretaries to assist the leaders of national delegations, to set up National Committees, and to found a journal for the Association (this latter idea has not yet been realized).

In 1977 the XIth Congress (Kiev) granted membership to Austria. In 1981 the XIIth (Bucharest) Congress ratified the affiliation of the CBGA to the IUGS and decided to create a Standing Commission on Mineral Deposits. In 1989 the XIth Congress (Sofia) accepted Greece as a member country.

At the XVth Congress (Athens, 1995) the Statutes were thoroughly modified in several aspects:

– The new version stressed the international, scientific and non-profit character of the CBGA, and reconfirmed its affiliation to the IUGS;

– English was made the official language of the Association;

– The right of decision-making was transferred from the Congress to the Council. The initiators of this modification pointed out that the majority of the participants of the Congresses is regularly made up by earth scientists of the host country, and this imbalance inevitably leads to injustice in care of voting;

– In any case the old statutes have turned out to be too strict and rigid, rendering difficult any extension of the field of activity or modification of the organizational structure. For instance, the experts engaged in Petroleum Geology, Coal Geology, or Quaternary Geology have not succeeded in obtaining independent Commissions for their respective fields of research. Even the specialists of Isotope Geology, Engineering Geology, Mineral Deposits and Environmental Geology had to struggle through several congresses to obtain separate Commissions;

– Another hot point of debate was the beginning and termination of the President's mandate. Previously the President of the Association was nominated by the country organizing the next Congress, but he/she was formally elected only at the Congress he /she had organized, and his/her term lasted until the following congress. Under the new system the Congress elects the President of the next congress, who should direct and coordinate not only the organization of that congress but also the activities of the CBGA between the two congresses, in his/her capacity as the President of the Association.

Tentative Assessment of the Activities of the CBGA

The organization of the congresses always required enormous efforts on the part the host countries. In all cases only the cooperation of all (scientific, educational and administrative) geologic institutions of the country could assure success. Even then it has happened (for instance at the XIIIth and XIVth Congresses) that only extended abstracts could be published. Furthermore, the Hydrogeological, the Engineering Geologic and several Paleotransportion Maps have never been published. As far as the Geologic Map is concerned, its compilation and plotting has not even begun.

Most of the thematic Commissions were active, but they had also passive periods. Whether the aims were attained or not depended mainly on the economic situation, and in some cases on the personal attitude and organizing skills of the chairpersons and the members. The following Standing Commissions were particularly active: Tectonics (**M. Mahel**, Czechoslovakia), Sedimentology (**A. Slaczka**, Poland), Igneous Petrology (**E. Szádeczky-Kardoss** and **V. Széky-Fux**, Hungary and **M. Borcos**, Romania), Isotope Chronology (**N. P. Semenenko**, Ukrainian SSR), Hydrogeology (**B. Kamenov**, Bulgaria). The Commission on Geophysics was particularly active during the chairmanship of **D. Prosen**, and that on Stratigraphy and Paleontology during the chairmanship of **G. Császár** (Hungary). In spite of all efforts exerted by its Ukrainian leaders (**E. K. Lazarenko, N. P. Shcherbak, V. N. Kvasnitsa**) the activity level of the Commission on Mineralogy and Geochemistry was variable due to the passivity of most members and to the rather inadequate financial support.

The situation of the CBGA was rendered difficult by the growing competition of other international endeavors in the field of earth sciences. These included for example the applied geologic programs organized by the Geologic Standing Committee of COMECON, the projects of the IUGS-UNESCO joint International Geological Correlation Programme (IGCP) relevant to the Carpathian-Balkan region, themes 6 (Geophysics) and 9 (Geology) of the Academies of Sciences of the Socialist Countries. This led in many cases to duplication of activities, waste of human resources, and excessive overloading of the active participants. As an example Problem 10 of COMECON had used the Tectonic Map produced by the CBGA for the compilation of its Metallogenetic Map of the region, which was then presented at the XIth Congress of the CBGA; in another, Problem Committee 6 of the Academies of Sciences of the Socialist Countries (KAPG) played an important role in the continuation of measurements on the deep crustal geological sections initiated by the CBGA. The Council of the Association has stubbornly refused to create Standing Commissions on Coal Geology, Petroleum Geology and Quaternary Geology, precisely in order to avoid duplication or overlapping with research activities being done in the framework of COMECON and INQUA, respectively.

Congresses were held initially every second year, but after the IXth Congress every fourth year only. Poland and Rumania each organized three congresses, Bulgaria, Czechoslovakia and the Ukrainian SSR two, and Greece, Hungary and Yugoslavia one. Austria is the organizer of the present congress (1998), and Slovakia has offered to host the next (XVIIth) one in 2002. The Council of the CBGA used to hold Business Meetings during, and from 1978 on also between congresses, to discuss any arising problems. Between congresses the Commissions used to meet once, but some Commissions met twice, or even three times.

The CBGA requests no membership fee. The participation fee of the congresses and their field trips covers only part of the expenses. The deficit used to be counterbalanced by the financial contribution of the Academies of Sciences (eventually jointly with other state institutions). After the change of political regime in the former "Socialist" countries (1990), the funding of science by the state dropped dramatically. The money needed for the work in Commissions or for attending Congresses had to be raised by the scientists themselves. Alternatively, they had to succeed in making their employer recognize that this kind of activity was an integral part of their job. For this reason successful activity could go on only where the participating scientists were supported and sponsored by the professional institutions of their country or financed from other sources.

Earth scientists of non-CBGA countries have displayed considerable interest in the geologic research conducted in this region. On the other hand, after the disintegration of Czechoslovakia and Yugoslavia the Czech Republic, Croatia and Slovenia did not want to be members of the Association. The negative attitude of the Czech Republic had neither scientific reasons nor historic justification. This opinion is confirmed by the fact that recently the Czech Republic has again applied for membership, as has Slovenia. The application of the Czech Republic was accepted by the Council in May 1998, to be ratified at the present XVIth Congress in Vienna, where the applications of Slovenia and Albania will be discussed.

Interest for the CBGA rose with time, as shown by the increase in the number of participants and papers. Participation was especially high at the congresses held in Sofia (1965, 1989), in Bratislava (1973) and in Athens (1995). After the Xth Congress, however, a regrettable phenomenon occurred. A considerable number of attendants participated only in the Congress proper, missing the field trips because of the increasing scarcity of funds. Consequently, the positive tradition of the first three congresses, namely putting the main accent on the insitu demonstration of geologic formations and phenomena, was only followed by Congresses IV to X. Thereafter, in spite of all efforts exerted by the host countries, participation in the field trips steadily declined.

Hungarian interest for the CBGA began seriously growing after the IXth Congress (Budapest, 1969) which was attended by 110 Hungarian geoscientists. In Congresses X to XIV the average number of participants from Hungary was over fifty, with about thirty papers presented. Even at the XVth Congress (Athens, 1995), the first one after the change of political regime, Hungarian participation was 17. Increasing Hungarian interest is proved by the fact that one of the field trips of the XVIth (Vienna) Congress is planned to visit Western Hungary, and one of the Special Symposia (S1, INHIGEO) has a Hungarian co-convenor (**E. Dudich**, the previous representative of Hungary in the Council of the CBGA, who succeeded the author of the present paper.)

It can be predicted that Hungarian participation and active contribution will continue to grow in spite of the still existing economic problems. This is my wish for **G. Császár**, the present representative of Hungary in the Council of the CBGA, for the Hungarian members of the Standing Commissions, and for all who are interested in the CBGA.

Note

At the XVIth Congress of the CBGA in Vienna, which turned out to be one of the biggest, at least according to the distributed List of Participants (638), there were 52 attendees from Hungary, with 37 oral presentations and 28 lectures formally included in the Program and the Abstracts Volume.

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Annex 1
Congresses of the Carpathian (Balkan) Geological Association

Number	Year	Host country	Town	President
I.	1925	Poland	Lwów	J. Nowak
II.	1927	Romania	Bucharest	G. Macovei
III.	1931	Czechoslovakia	Prague	R. Kettner
IV.	1958	USSR (Ukraine)	Lvov – Kiev	E. K. Lazarenko
V.	1961	Romania	Bucharest	A. Codarcea
VI.	1963	Poland	Warsaw – Cracow	M. Mrozowski
VII.	1965	Bulgaria	Sofia	Y. Y. Yovchev
VIII.	1967	Yugoslavia	Belgrade	M. Lukovic
IX.	1969	Hungary	Budapest	E. Szádeczky-Kardoss
X.	1973	Czechoslovakia	Bratislava	O. Fusan
XI.	1977	USSR (Ukraine)	Kiev	N. P. Semenenko
XII.	1981	Romania	Bucharest	I. Folea
XIII.	1985	Poland	Cracow	Z. Dembowski
XIV.	1989	Bulgaria	Sofia	M. Zhelyaskova-Panayotova
XV.	1995	Greece	Athens	D. Papanicolaou
XVI.	1998	Austria	Vienna	W. Janoschek
(XVII.)	2002	Slovakia	Bratislava	J. Vozár *

* approved at the Vienna Congress

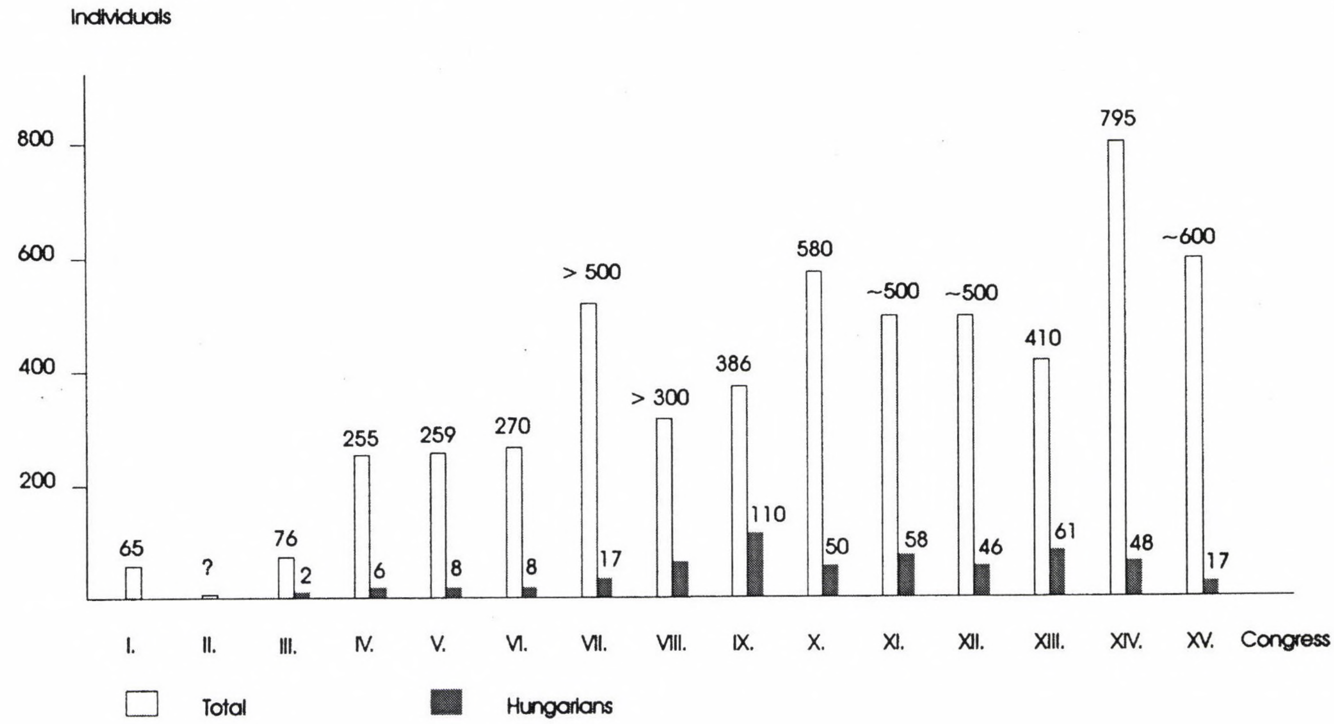
Annex 2
Business Meetings of the Council of the CBGA held between Congresses

Date	Venue	President	Objective
13 Oct. 1979	Budapest	Gy. Grassely	Affiliation to IUGS
2–3 Dec. 1982	Bucharest	I. Folea	Preparation of XIIIth Congress
21–22 Oct. 1987	Cracow	W. Slizewski	Preparation of XIVth Congress
12–15 May 1992	Sofia	M. Zhelyaskova-Panayotova	Preparation of XVth Congress
19 May 1993	Budapest	E. Dudich	Postponement of XVth Congress, discussion of the modified statutes
25–27 May 1994	Thessaloniki	D. Papanicolaou	Preparation of XVth Congress
May 1997	Vienna	W. Janoschek	Preparation of XVIth Congress

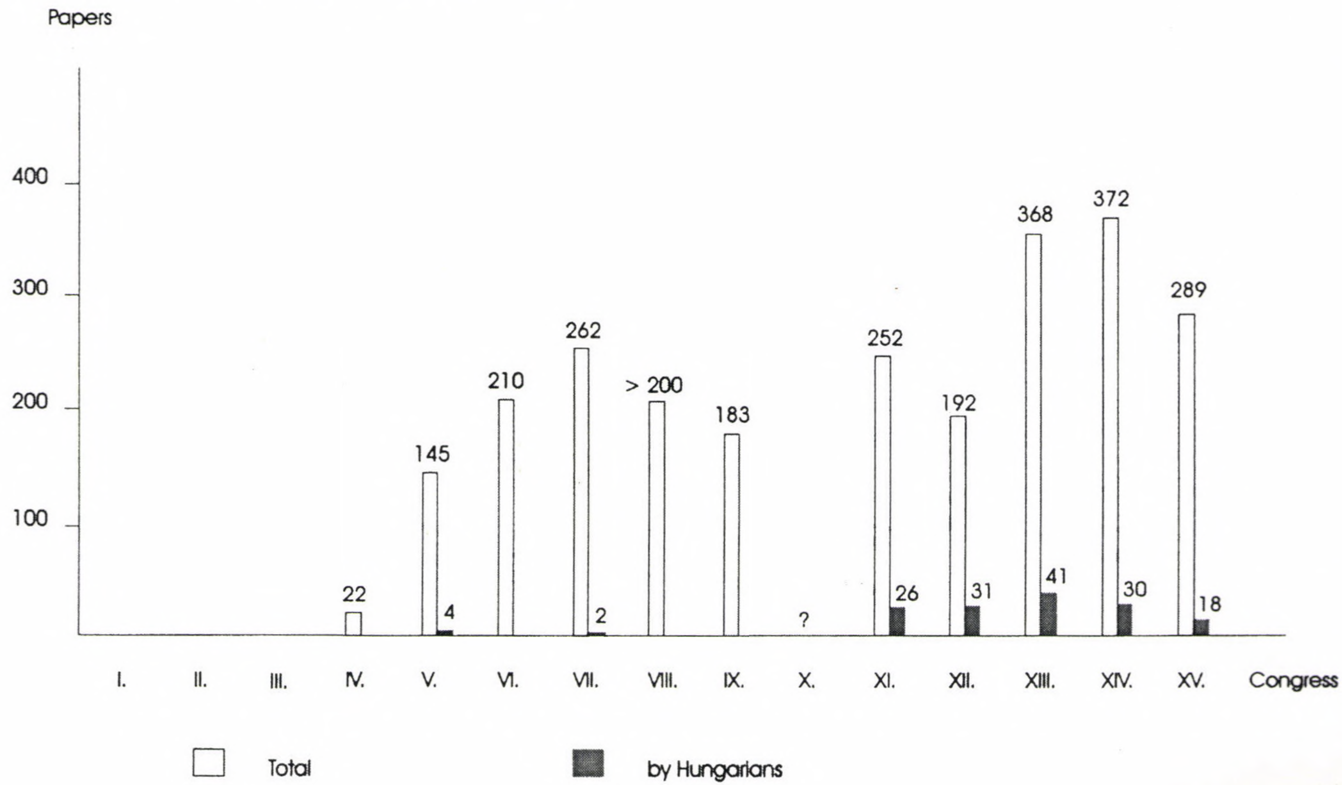
Annex 3
Chairpersons and the Representatives of Hungary in the Commissions of the Carpathian–Balkan Geological Association (as of 1997)

Name of the Committee	Chairman	Country	Hung. representative
1. Engineering Geology	I. Brutchev	Bulgaria	P. Scharek
2. Environmental Geology	I. Mariolakos	Greece	P. Scharek
3. Geologic Maps	M. Sandulescu	Romania	G. Hámor
4. Geochemistry and Isotope Geology	V. Zagnitko	Ukraine	I. Viczián
5. Geophysics	E. Lagios	Greece	Gy. Pogácsás
6. Magmatism	M. Borcos, J. Lexa	Romania, Slovakia	P. Gyarmati
7. Metamorphites	V. Höck	Austria	P. Árkai
8. Mineral Deposits	M. Petkovic, Milankovic	Yugoslavia	vacant
9. Mineralogy	V. N. Kvasnitsa	Ukraine	I. Viczián
10. Sedimentology	A. Slaczka	Poland	J. Haas
11. Stratigraphy, Paleontology and Palaeogeography	G. Császár	Hungary	G. Császár
12. Tectonics	D. Plasienka	Slovakia	L. Fodor
Sub-commission of Hydrogeology	B. Filipovic	Yugoslavia	P. Scharek

Annex 4
Number of Participants at the Congresses of the C(B)GA



Annex 5
Papers Presented at the Congresses of the CBGA



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Pre-Alpine development of the andalusite-sillimanite-biotite-schist from the Sopron Mountains (Eastern Alps, Western Hungary)

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The pre-Alpine metamorphic P-T path of the andalusite-sillimanite-biotite-schist from the Kovács-árok, Brennberg, Sopron area was inferred from mineral equilibria and geothermo-barometry of zoned garnet.

The first step in the P-T history is given by the reaction of the paragonite component of muscovite and quartz, to form sillimanite and plagioclase. This reaction gives a minimum pressure estimate between 320 and 480 MPa at minimum temperatures of about 550–600 °C. Formation of andalusite and the different breakdown reactions of staurolite show a pressure decrease and a temperature increase, which leads to the temperature peak of the metamorphism at the beginning of the granulite facies; this is indicated by the appearance of the K-feldspar+sillimanite assemblage. Mineral equilibria and geothermo-barometry of the garnet core indicates pressure between 240 and 380 MPa and temperature between 650–700 °C.

The thermal peak of the metamorphism was followed by a slight decrease in the temperature and a significant increase in the pressure, up to 1060 MPa. This relatively high pressure event is marked by the Ca-rich zone of the garnet and the relatively albite-rich plagioclase core. Subsequent garnet zones indicate rapid pressure decrease and slow cooling (P=400–480 MPa, T=607–670 °C and P=240–400 MPa, T=564–580 °C).

Key words: micaschist, amphibolite facies, pre-Alpine metamorphism, mineral equilibria, P–T path, geothermo-barometry, Eastern Alps, Hungary

Introduction

The studied rocks are andalusite-sillimanite-biotite schists, outcropping in the Kovács-árok, near Brennberg (Fig. 1) and extend into Austria (Brennberg-Kaltes Bründl Series). This series is considered to be a member of the Grobgneis Series of the Lower Austroalpine Unit, but Draganits and Grasemann (1996) stressed the similarities with the Strallegger Gneisses and with the Dist-Paramorphosenschiefer (Koralpe) of the Middle Austroalpine Unit.

This rock type offers a unique opportunity to gain an insight into the pre-Alpine metamorphic development of the area because Alpine overprint of the rock was not pervasive. The Alpine overprint seems to be confined to the zones of fluid movement in the vicinity of shear zones, quartz veins and gneiss veins. Several works (e.g. Lelkes-Felvári et al. 1984; Kisházi and Ivancsics 1987,

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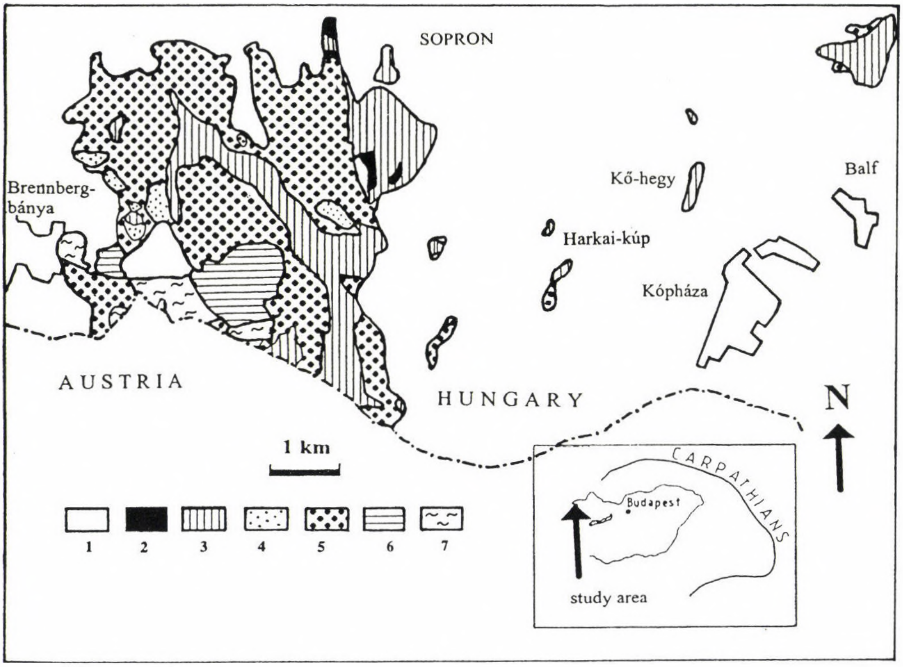

Fig. 1
Geologic sketch map of the Sopron area after Kisházi and Ivancsics (1989). 1. Tertiary and Quaternary sedimentary rocks; 2. Leucophyllite; 3. Gneiss; 4. "Kyanite-quartzite"; 5. Chlorite-muscovite schist; 6. Kyanite/chloritoid-muscovite schist; 7. Andalusite-sillimanite-biotite schist

and references therein) were published discussing the polymetamorphic character of these micaschists and phase relations of the rock-forming minerals as well as their implications for the pressure-temperature conditions of metamorphic events.

In spite of the extensive work published in the literature, quantification of the pre-Alpine metamorphic P–T history is still missing. This work is an attempt to provide data on pre-Alpine pressure-temperature development and the P–T path of this rock on the basis of mineral equilibria and geothermo-barometry.

Geology

Metamorphic rocks of the Sopron area are the easternmost outcropping part of the Austroalpine nappe system and belong to the Grobgneis series of the Lower Austroalpine Unit. The outcropping rocks comprise gneisses, metagranites, micaschists, and leucophyllites. Amphibolite can only be found in boreholes. Most of the rocks in the area show strong Alpine overprint except for the polymetamorphic andalusite-biotite schist. The metamorphic evolution model of the Sopron area is mostly based on this type of rock (e.g. Lelkes-Felvári et al. 1984; Kisházi and Ivancsics 1985, 1987, and references therein).

Lelkes-Felvári et al. (1984) describes a three-stage metamorphic development in the area. The first metamorphic stage is thought to be a pre-Hercynian one, probably Caledonian Barrow-type metamorphism with staurolite, kyanite, garnet and sillimanite. This metamorphic stage is established only on textural grounds and is not confirmed by geothermo-barometry. The second stage is considered to be of Hercynian age and characterised by andalusite, sillimanite and garnet. Draganits and Grasemann (1996) obtained 300–500 MPa pressure and 650 °C temperature for the pre-Alpine mineral assemblage of the Austrian part of the Brennberg-Kaltes Bründl Series. The third stage is considered to be Alpine which developed under greenschist to lower amphibolite facies conditions. However, Török (1996, 1998) has shown some lines of evidence of the existence of an Alpine high-pressure metamorphic phase (about 1.3 GPa pressure and about 550–600 °C temperature) in the metagranitoids and orthogneisses on the basis of phengite barometry and mineral equilibria. Similar P–T conditions were obtained by Demény et al. (1997) for the Mg-chlorite-muscovite-quartzphyllites (leucophyllites), crosscutting the gneisses. Draganits and Grasemann (1996) obtained 550±30 °C temperature and 0.95±0.15 GPa pressure on muscovite-chlorite-quartz schist in contact with the orthogneisses.

The studied samples of andalusite-sillimanite-biotite schist contain a well-preserved pre-Alpine mineral assemblage and thus provide a unique clue to the understanding of the pre-Alpine development of the area. The samples were collected from the classic locality, near Brennberg, in the western side of the Kovács-árok.

Petrography

The rock contains quartz, biotite, andalusite, sillimanite, plagioclase, K-feldspar, muscovite, staurolite and accessory kyanite, garnet, corundum, spinel, ilmenite, tourmaline, zircon, monazite, and xenotime. This complex mineralogy is due to polymetamorphism as was pointed out by the previous workers cited above. The micaschists are sometimes cut by metagranitoid veins which are attributed to the anatexis of the rock at depth.

Two rock types can be distinguished both in the outcrop and in thin sections, which form alternating layers. The most abundant one is a typical dark colored metapelite, rich in andalusite, biotite and sillimanite and without clear preferred orientation. The other is much lighter, contains much more quartz, and less Al_2SiO_5 and biotite, with occasional zircon and tourmaline-rich layers. The Al_2SiO_5 species is mostly fibrolitic sillimanite associated with biotite or muscovite. Rarely some thin sillimanite or andalusite-rich layers can also be observed. These features may imply a psammitic origin for these quartz-rich layers. These alternating bands of pelitic and psammitic origin may be relics of the pre-metamorphic sedimentary layering.

Some parts of the rock have experienced retrograde metamorphism which was most probably due to the Alpine phase. The Alpine overprint seems to be strong in the vicinity of shear zones (e.g. the micaschist of the Oromvég quarry) and gneiss veins (e.g. the micaschist of the left-hand side valley of the Kovács-árok). In these places the Alpine overprint is pervasive. Rare pre-Alpine garnet relics in the cores of some Alpine garnets and pre-Alpine staurolite and corundum relics in Alpine kyanite-staurolite-chloritoid aggregates are the only observable relics of the pre-Alpine mineral assemblage. The andalusite-sillimanite-biotite schist of the Kovács-árok outcrop seems to have suffered only local and less intensive Alpine overprint along some thin gneiss and quartz veins. The less intensive nature of the Alpine overprint means that most of the pre-Alpine minerals can be seen as relics. In these parts the andalusite porphyroblasts have an alteration rim composed of a very fine-grained aggregate of staurolite and kyanite (Fig. 2) or sometimes partial alteration of andalusite to sericite can be observed. The biotite and sillimanite-rich parts altered to fine grained muscovite, biotite and idioblastic staurolite. Sometimes the relics of fibrolitic sillimanite are still observable. New idioblastic-hypidioblastic garnet occurs in these aggregates. New garnet also occurs as corona on big biotite flakes. Albite overgrowth on pre-Alpine oligoclase is an indicator of the Alpine overprint as well. Shear zones, gneiss and quartz veins may have been the pathways of infiltrating fluids, which may have facilitated breakdown of the pre-Alpine mineral assemblage and formation of a new mineral assemblage during Alpine metamorphism. The intensity of this replacement may have been a function of the amount of fluid migrating along the individual pathways. Alpine metamorphism of the andalusite-sillimanite-biotite schist will be discussed in detail in a separate paper.

Two types of andalusite porphyroblasts can be distinguished in those parts of the rock which was not affected by the Alpine overprint. The first one is deformed, has undulose extinction, resembles an aggregate of smaller andalusite domains, and contains abundant inclusions of biotite, sillimanite, corundum, spinel, ilmenite and relic staurolite. The other one is not deformed, and contains only biotite and quartz inclusions. Occasionally this type forms overgrowths around the andalusite of the first type. Sometimes andalusite in contact with K-feldspar can also be observed.

Spinels form xenoblastic crystals mostly in the vicinity of staurolite relics in the andalusite poikiloblasts of the first type. Spinel and corundum often occur together in the same andalusite, and sometimes they can be observed in contact with each other (Fig. 3). A fine-grained paragonite and muscovite rim resulting from retrograde metamorphic reactions can sometimes be observed around spinels. Often new, most probably Alpine staurolites crystallised on this mica-rich reaction product (Fig. 3). In some rare cases, small, rounded sericite aggregates can be observed enclosed in the andalusite porphyroblasts.

Corundum can be found exclusively as inclusion in deformed andalusite and is mostly associated with biotite (Fig. 4) but may also occur without a close

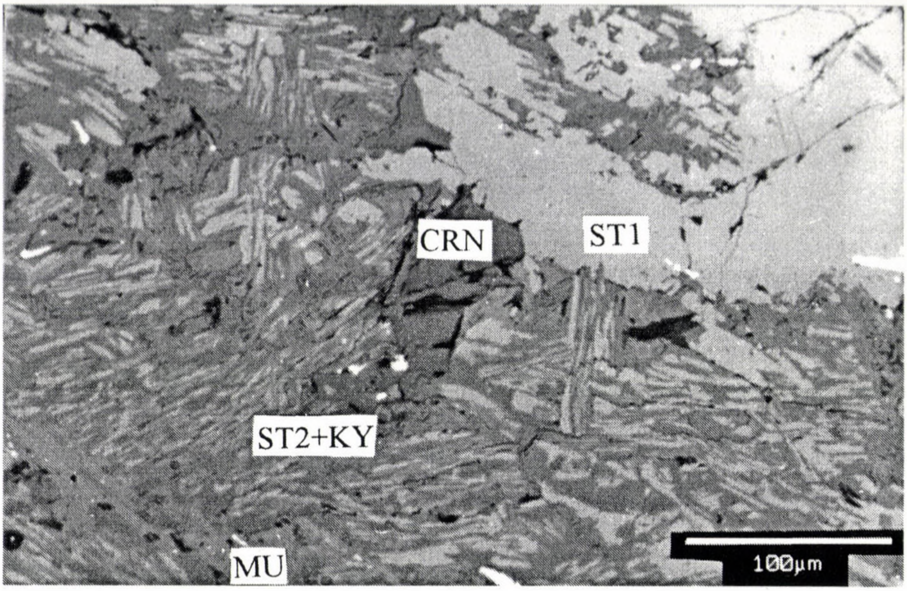

Fig. 2
Backscattered electron image of fine-grained kyanite+staurolite (ST2+KY) intergrowth with muscovite (MU) after andalusite. Relic staurolite (ST1) and corundum (CRN) can also be seen

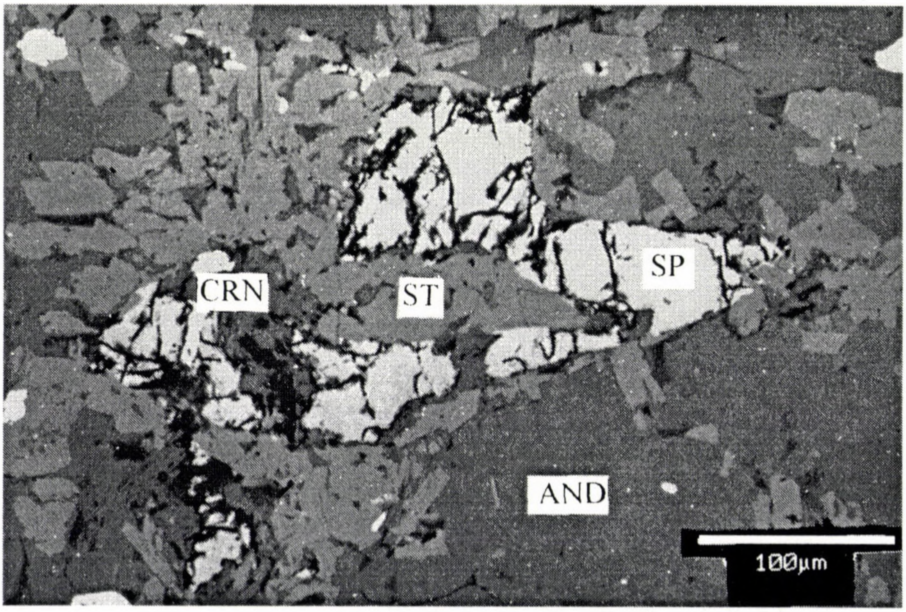

Fig. 3
Backscattered electron image of spinel (SP) and corundum (CRN) inclusions in andalusite (AND). The retrograde replacement of spinel and corundum with new idioblastic-hypidioblastic staurolite (ST) and white mica (muscovite and paragonite) can be observed. The dark patches in staurolite are white micas and the white ones are ilmenite flakes

Fig. 4
Corundum (CRN) and biotite (BI) inclusions around relic staurolite (ST) in andalusite (AND). The scattered opaque minerals are ilmenites. Plane polarized light, horizontal field of view: 0.5 mm

association of biotite (Fig. 5). There are some cases when the corundum is surrounded by a biotite corona. Aggregates of small vermicular corundum crystals with the same extinction can also be observed intergrown with andalusite.

Staurolites can be observed in three textural positions:

1. The first one is when it forms ragged, xenoblastic inclusions in deformed andalusite porphyroblasts, associated with biotite and sillimanite ± spinel ± corundum. These could be the pre-Alpine staurolite relics.

2. Sometimes small idioblastic-hypidioblastic staurolite appears around ragged spinel inclusions in deformed andalusite (Fig. 3). These staurolites may have formed either during the retrograde phase of the pre-Alpine metamorphism or during the Alpine metamorphism.

3. The other occurrences are related to the Alpine metamorphism. New, fine-grained staurolite+kyanite intergrowths replace former andalusite porphyroblasts (Fig. 2) or small idioblasts of staurolite appear with fine grained micas, albite and new garnet as described above.

Sillimanite occurs either enclosed in andalusite (Fig. 6), or as aggregates, associated mostly with biotite and plagioclase (Fig. 7), sometimes with

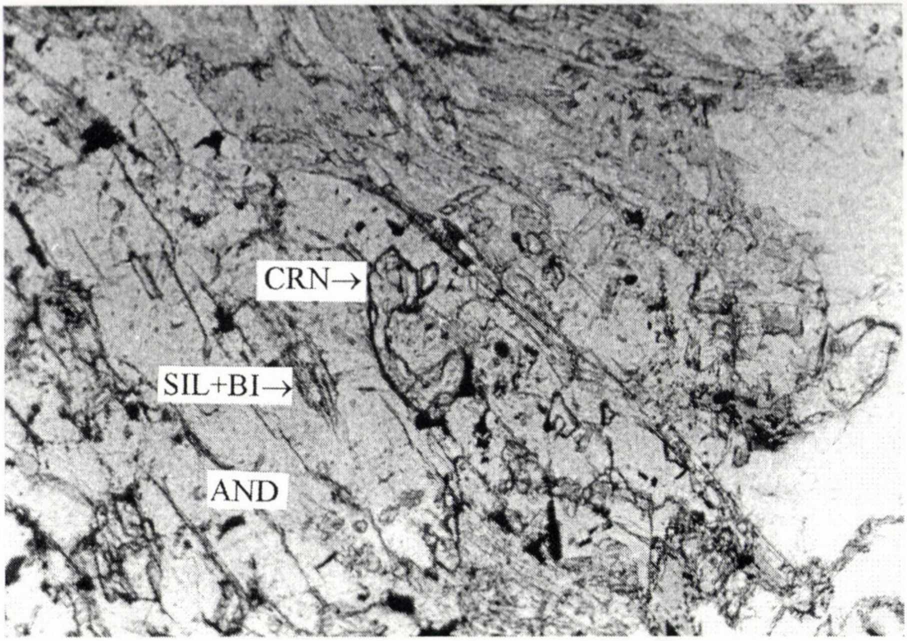

Fig. 5
Corundum (CRN) and sillimanite+biotite (SIL+BI) inclusion in andalusite (AND). Plane polarized light, horizontal field of view: 0.5 mm

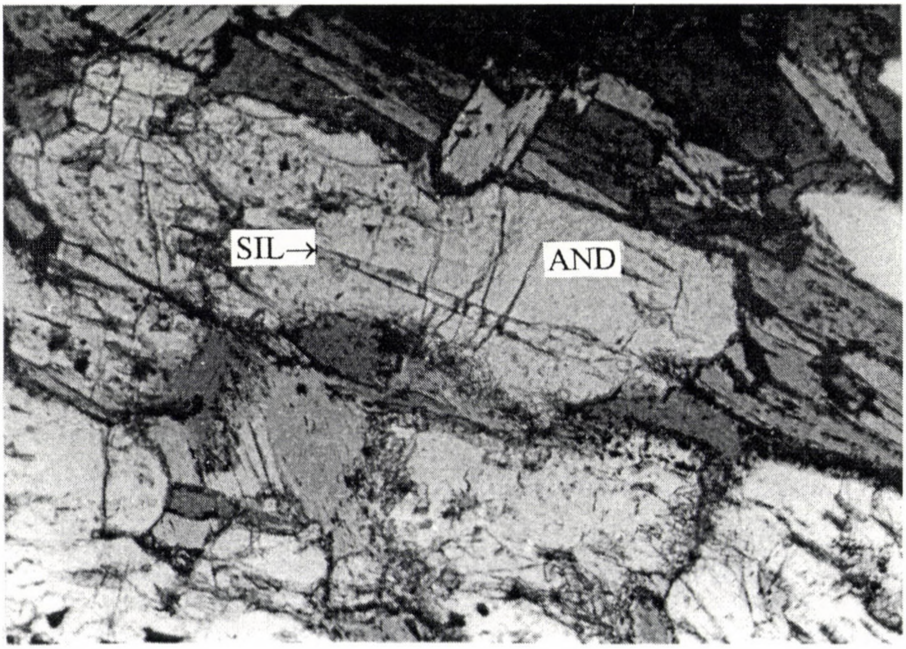

Fig. 6
Sillimanite (SIL) inclusion in andalusite (AND) surrounded by biotite. Plane polarized light, horizontal field of view: 1 mm

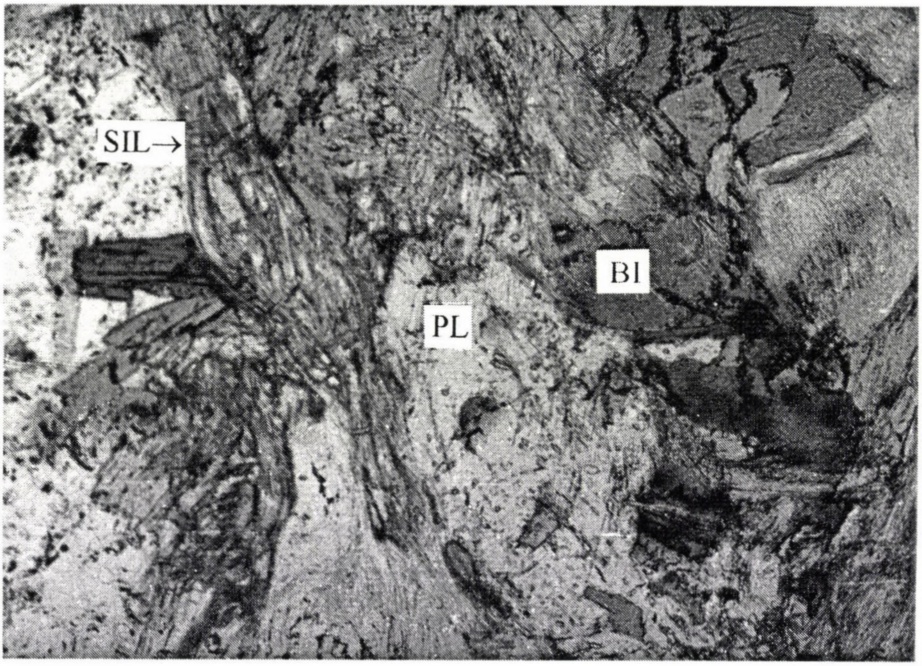

Fig. 7
Sillimanite (SIL), associated with biotite (BI) and plagioclase (PL). Plane polarized light, horizontal field of view: 2 mm

muscovite. Sillimanite needles enclosed in plagioclase can often be observed. The plagioclase-biotite-sillimanite assemblage sometimes forms long layers. These layers may contain K-feldspar with fibrolite needles (Fig. 8) as well. Retrograde reaction of K-feldspar and sillimanite assemblage to muscovite was also observed. Sillimanite enclosed in andalusite porphyroblasts exhibit either a long-prismatic shape or a fibrolitic one, while those associated with micas and plagioclase tend to be fibrolitic with rare prismatic development. Prismatic sillimanite inclusions in andalusite are mostly associated with relic staurolite, spinel and corundum.

Kyanite occurs as small-elongated porphyroblasts only in places where Alpine overprint can be observed, replacing andalusite; it is therefore regarded as related with Alpine metamorphism. Rare kyanite inclusions enclosed in andalusite can also be observed in the same places with Alpine overprint (Fig. 9).

Garnet occurs as rare, relatively large xenoblasts (up to 1–2 mm) in places where the above-mentioned signs of Alpine overprint are missing, associated with andalusite, sillimanite, biotite, plagioclase and sometimes muscovite. Garnets contain quartz, biotite, plagioclase and zircon inclusions. These garnets are quite rare and considered to be pre-Alpine ones.

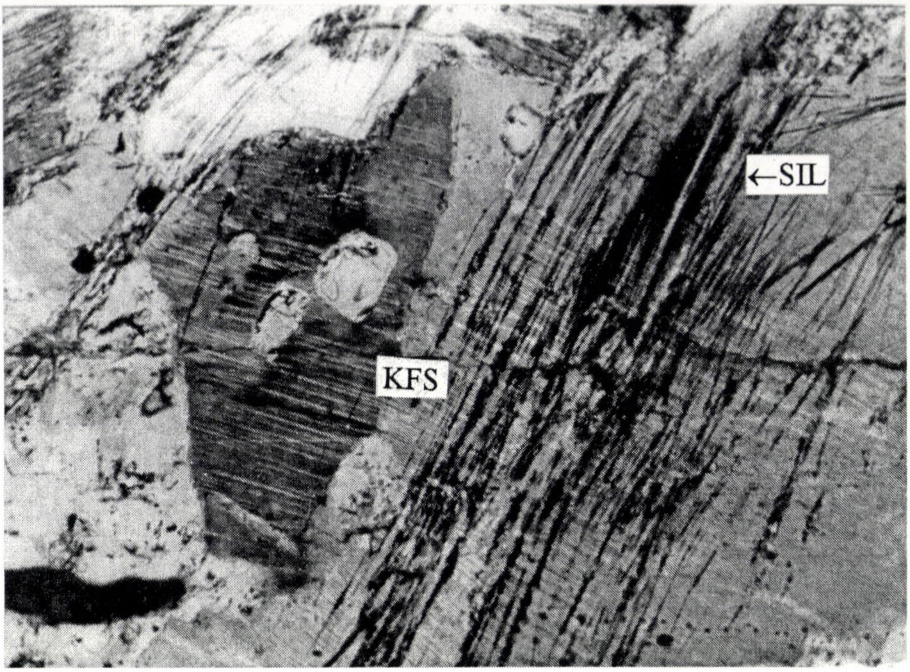

Fig. 8
Sillimanite (SIL) needles in perthitic K-feldspar (KFS). Crossed-polarized light, horizontal field of view: 1 mm

Other parts of the rock where the Alpine overprint is obvious contain small, idioblastic-hypidioblastic garnet associated with new staurolite idioblasts, plagioclase (sometimes of albitic composition), fine-grained biotite and muscovite. Garnet also occurs as corona around biotite. These garnets are thought to be related to the Alpine metamorphism. The Alpine garnets are much smaller in size and much more abundant than the pre-Alpine ones.

Mineral chemistry

The chemistry of some of the rock-forming minerals was determined by an AMRAY 1810I microprobe equipped with EDS detector at the Department of Petrology and Geochemistry, Eötvös University. Natural and synthetic minerals were used as standards and ZAF correction was used during calculation of mineral compositions.

Garnet

Pre-Alpine garnet compositions were measured in samples where the above-mentioned signs of Alpine overprint were not observable and garnet

Fig. 9
Kyanite (KY) inclusions in andalusite (AND) in that part of the rock which shows the Alpine overprint. The rim of the andalusite is partly replaced with fine grained aggregate of kyanite+staurolite. Plane polarized light, horizontal field of view: 2 mm

was associated with sillimanite or unaltered andalusite. Pre-Alpine garnets are rich in almandine (68.01–80.22%) with low to relatively high spessartine (2.5–17.9%), intermediate pyrope (5.9–13.86%) and low grossular+andradite (0.9–11.8%) content.

Alpine garnets can be usually distinguished from the pre-Alpine ones on the basis of their higher grossular component, but there is some overlap of Alpine garnet rim compositions with the core composition of some Ca-richer pre-Alpine garnet. Pre-Alpine and Alpine garnets can be distinguished by textural criteria and mineral association mentioned above and the different zoning trends. In the measured 4 pre-Alpine garnets almandine and spessartine are changing together (i.e. spessartine increases when almandine increases and spessartine decreases when almandine decreases), but changes in grossular are always to the opposite. In the measured 8 Alpine garnets spessartine and grossular are changing in the same way and almandine is changing to the opposite. Moreover the pre-Alpine garnets have typical retrograde zoning, with increasing spessartine content from core to rim, while the Alpine garnets have progressive zoning with decreasing spessartine content from core to rim. Composition of garnets is shown in Fig. 10, together with the field of the Alpine garnets from the micaschists of the left-side valley of the Kovács-árok and Oromvég quarry (Török, in prep.). Figure 10 also summarises the main zoning trends of the Alpine and pre-Alpine garnets as described above.

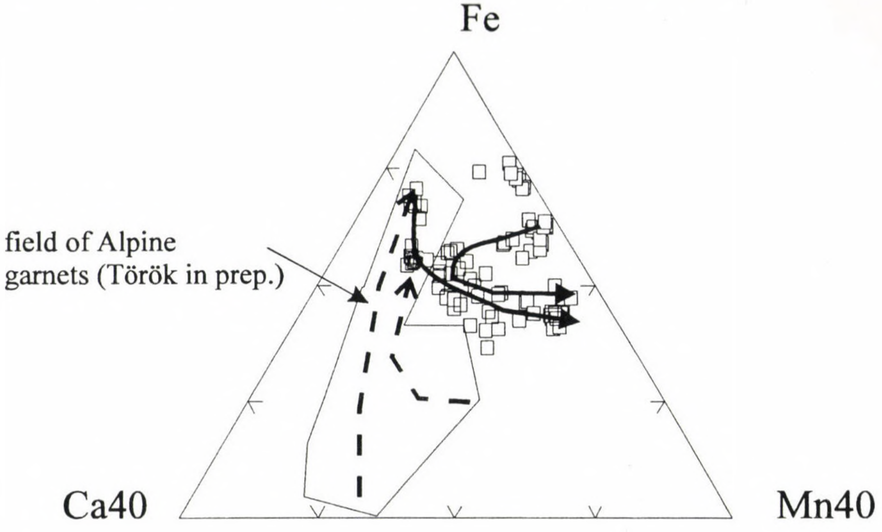

Fig. 10
Composition of pre-Alpine garnets from the andalusite-sillimanite-biotite schist of the Kovács-árok in the Ca40-Fe-Mn40 triangle. The marked area shows the composition of the Alpine garnets (Török, in prep.) for reference. The arrows show the zoning trends of the pre-Alpine garnets and the broken arrows in the Alpine garnet field show the zoning trends of Alpine garnets for comparison. The beginning of the arrows mark the core compositions and the ends mark the rims

Big, xenoblastic garnets are quite rare and they have composite and discontinuous zoning (Fig. 11). The composition of the core is quite uniform with $Alm_{76.1–77.0}Spe_{11.9–13.4}Prp_{8.0–9.4}Gr{+}Adr_{1.5–2.8}$. The core is followed by a relatively Ca-rich and Fe-poor zone with a composition of $Alm_{68.5–71.7}Spe_{9.2–14.1}Prp_{8.5–9.5}Gr{+}Adr_{6.8–10.4}$. After the Ca-rich zone the Ca decreases again with increasing Mn and Fe: $Alm_{71.3–72.6}Spe_{15.8–17.8}$ $Prp_{7.5–9.0}Gr{+}Adr_{0.9–5.0}$. Composition of the Ca-rich zone of the pre-Alpine garnet is close to the field of the Alpine garnets. Zoning pattern of another pre-Alpine garnet with Ca-rich core is presented in Fig 11a. The general zoning trends are similar to the zoning trends of garnet in Fig. 11, but the Ca-poor core is lacking and the Ca-rich part is thicker. Zoning pattern of this garnet shows that the Ca-rich part of the garnet has also some zoning. The composition of the Ca-rich part coincides with the rim composition of the Alpine garnets, which crystallised under high pressure (800–1300 MPa, Török in prep.). Representative garnet analyses are shown in Table 1.

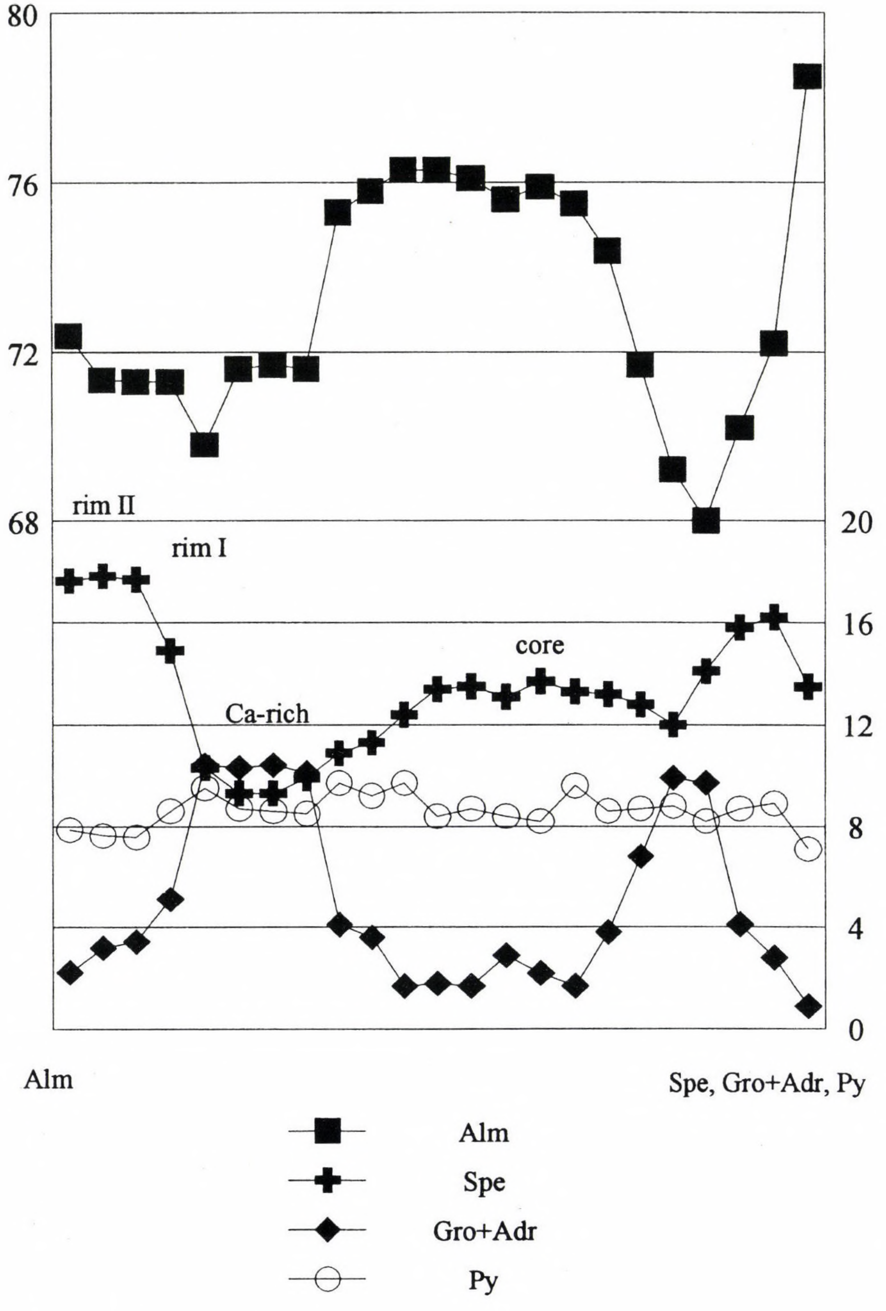

Fig. 11
Compositional zoning of the garnet used for geothermo-barometry

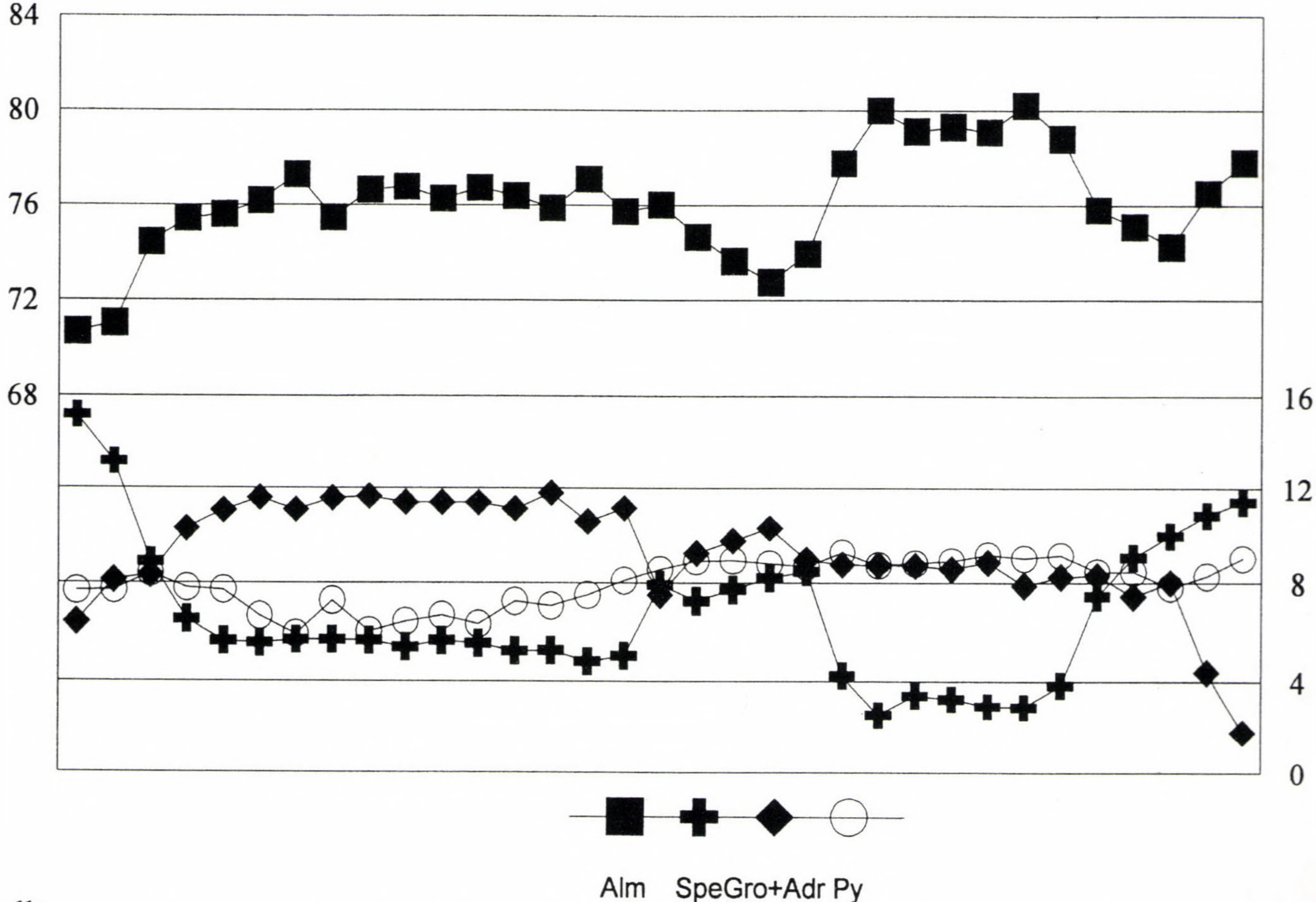

Fig. 11a
Zoning pattern of another pre-Alpine garnet with Ca-rich core. The general zoning trends are similar to the garnet in Fig. 11, but the Ca-poor core is lacking

Table 1
Representative garnet analyses. Cation numbers are on the basis of 12 oxygens

	1.	2.	3.	4.	5.	6.	7.	8.
SiO_2	36.72	36.63	37.23	36.93	36.82	36.82	36.56	36.26
Al_2O_3	20.66	20.61	20.42	20.69	20.36	20.37	20.69	20.50
MgO	2.36	2.33	2.15	2.13	1.88	2.26	1.93	1.86
FeO_{tot}	33.88	33.77	32.63	32.45	32.31	32.27	32.20	31.73
CaO	0.54	0.64	3.56	3.59	1.82	2.22	0.73	1.08
MnO	5.46	5.22	4.08	4.05	6.45	5.52	7.61	7.63
Sum	99.62	99.20	100.07	99.84	99.64	99.46	99.72	99.06
Si	2.994	2.998	3.007	2.987	3.004	2.999	2.985	2.98
Al	1.985	1.988	1.944	1.972	1.958	1.955	1.991	1.986
Mg	0.287	0.284	0.259	0.257	0.229	0.274	0.235	0.228
Fe^{2+}	2.284	2.297	2.162	2.143	2.172	2.151	2.161	2.126
Fe^{3+}	0.026	0.015	0.042	0.052	0.033	0.047	0.037	0.054
Ca	0.047	0.056	0.308	0.311	0.159	0.194	0.064	0.095
Mn	0.377	0.362	0.279	0.277	0.446	0.381	0.526	0.531
Sum cat	8.009	8.005	8.014	8.017	8.011	8.016	8.012	8.018
Alm	76.26	76.58	71.60	71.71	72.09	71.71	72.38	71.35
Sps	12.59	12.07	9.37	9.28	14.92	12.69	17.62	17.82
Gro	0.25	1.12	8.23	7.80	3.67	4.10	0.26	0.46
Prp	9.58	9.48	8.69	8.59	7.66	9.15	7.86	7.64
Adr	1.32	0.75	2.11	2.62	1.66	2.35	1.88	2.73

Analyses in columns 1–2 represent the core composition coexisting with plagioclase and biotite from which geothermo-barometric calculations were made. Analyses in columns 3–4 represent intermediate compositions between garnet cores and rims. Analyses in columns 5–6 and 7–8 represent rims I and II, respectively

Muscovite

There is a good trend in Ti and Al6 content of the muscovites in contact with different garnet zones (Fig. 12) starting from the Ca-rich garnet zone. Muscovite was not stable when the core of the garnet was formed. Muscovites in contact with the Ca-rich garnet zone have the lowest Ti content (0.015–0.027 atoms p.f.u.) and highest Al6 (3.703–3.752 atoms p.f.u.) and those in contact with rim I have the highest Ti content (0.065–0.093 atoms p.f.u.) and lowest Al6 (3.622–3.691 atoms p.f.u.). Muscovites in contact with rim II of the garnet have intermediate Ti and Al6 content (0.051–0.067 and 3.671–3.735 atoms p.f.u., respectively). Muscovites in the matrix are similar in composition to those in contact with the Ca-rich zone of the garnet, they have low Ti (0.009–0.013 atoms p.f.u.) and high Al6 (3.706–3.771 atoms p.f.u.) content. This implies that matrix muscovites and those in contact with the Ca-rich zone of the garnet may have formed from the retrograde reaction of the K-feldspar and sillimanite, while

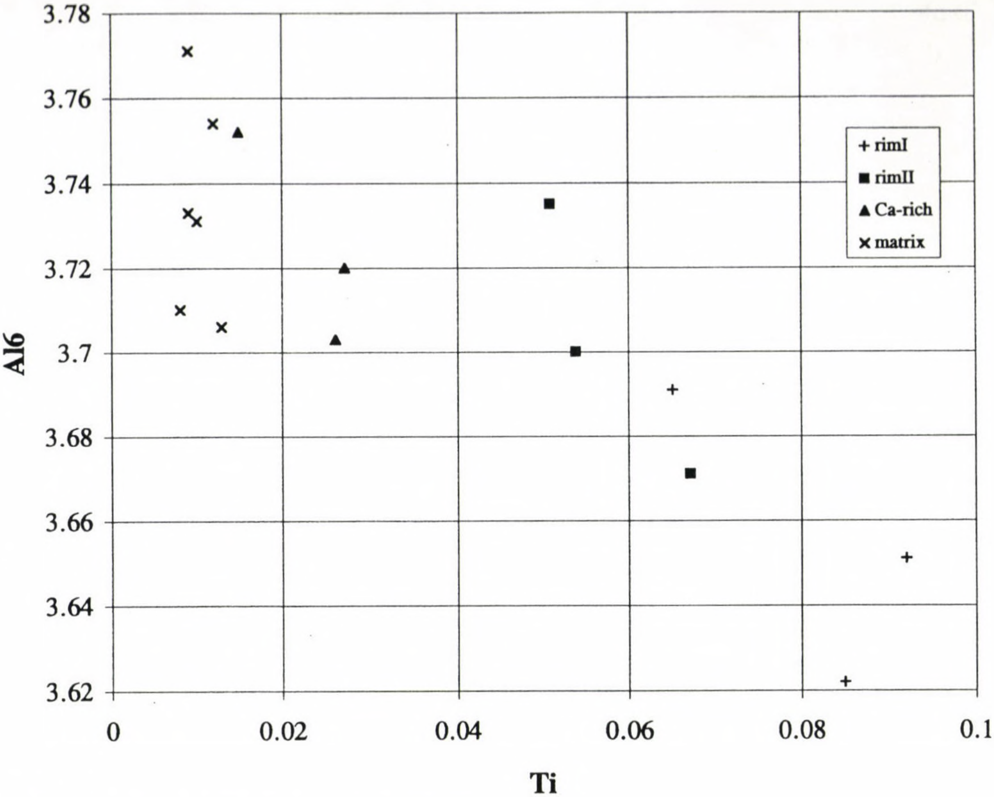

Fig. 12
Ti vs. Al6 plot of different muscovite generations

those in contact with the garnet rims I and II are were modified and reequlibrated with the garnet during its retrograde growths or alternatively they were formed contemporaneously with the subsequent garnet rims. Representative muscovite analyses are shown in Table 2.

Biotite

There is a trend in the composition of biotites as well, similarly to the muscovites. The only difference is, that biotite was stable with the core of the garnet and muscovite was not stable and was missing from the mineral assemblage. The Ti and Al4 contents increase from the inclusions in the garnet core towards the biotites in contact with the edge (rim II) of the garnet (Fig. 13 and Table 2). Biotite inclusions in garnet core have the lowest Ti (0.05–0.11 atoms p.f.u.), and the lowest Al4 content (2.38–2.68 atoms p.f.u.) and biotites

Table 2
Representative microprobe analyses of micas. Cation numbers are on the basis of 22 oxygens

	1.	2.	3.	4.	5.	6.	7.
SiO_2	34.56	34.72	34.24	33.60	46.5	47.13	45.89
Al_2O_3	18.93	20.89	19.83	19.72	35.04	34.99	35.37
TiO_2	0.65	1.83	2.51	2.47	0.26	0.93	0.67
FeO_{tot}	20.87	21.20	22.08	21.45	0.96	0.96	1.07
MgO	9.63	6.88	6.86	7.18	0.93	0.76	0.96
K_2O	9.87	9.79	9.75	9.58	11.25	11.27	11.22
Na_2O	b.d.	b.d.	0.22	0.28	0.35	0.23	0.29
Sum	94.56	95.31	95.49	94.28	95.29	96.27	95.47
Si	5.354	5.328	5.289	5.250	6.198	6.214	6.116
Al	3.456	3.778	3.610	3.632	5.505	5.437	5.555
Ti	0.076	0.211	0.292	0.290	0.026	0.092	0.067
Fe	2.704	2.721	2.852	2.803	0.107	0.106	0.119
Mg	2.235	1.574	1.579	1.672	0.185	0.149	0.191
K	1.950	1.916	1.921	1.909	1.913	1.895	1.907
Na	0	0	0.066	0.085	0.090	0.059	0.075
Sum cat.	15.775	15.529	15.608	15.641	14.024	13.952	14.03

1. Biotite inclusion in garnet core, in contrast with plagioclase; 2. Biotite inclusion in contact with the Ca-rich zone of the barnet, between the core and the rim (columns 3 and 4 in Table 1); 3. Biotite in contact with arnet rim I (columns 5 and 6 in Table 1); 4. Biotite in contact with the garnet rim II (columns 7 and 8 in Table 1); 5. Muscovite in contact with the Ca-rich zone of garnet; 6. Muscovite in contact with rim I garnet; 7. Muscovite in contact with rim II garnet; b.d. – below detection

in contact with the rim II garnet have the highest Ti (0.27–0.31 atoms p.f.u.) and Al4 (2.73–2.75 atoms p.f.u.). Biotites in contact with the Ca-rich garnet zone and with the rim I garnet zone have intermediate Ti and Al4 content (0.21–0.25, 2.68–2.76 and 0.23–0.29, 2.68–2.71 atoms p.f.u., respectively). Matrix biotites tend to be more magnesian than those listed above (Fig. 14).

Plagioclase

The analysed plagioclases (31 points) have quite uniform composition from the inclusions in garnet to those in contact with different zones of the garnet. Their albite content varies from 77.6 to 85.4% with 13.5 to 20.7% anorthite and 0–1.7% K-feldspar. Most of the measured plagioclase do not show any significant zoning. One plagioclase grain out of the six measured ones have shown slight zoning, with $Ab_{80.2}An_{19}Ort_{0.8}$ in the rim and $Ab_{85.4}An_{13.5}Ort_{1.1}$ in the core. This plagioclase is in the vicinity of the garnet used for geothermo-barometry and shown in Fig. 11. Some representative plagioclase analyses are shown in Table 3.

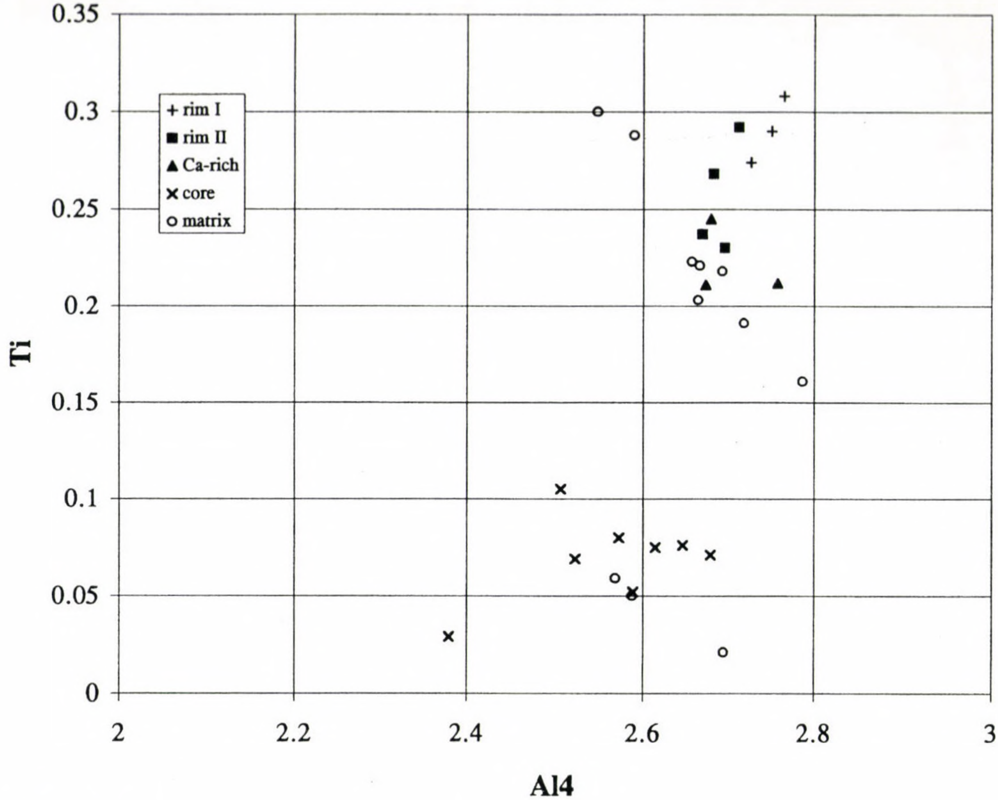

Fig. 13
Al4 vs. Ti plot of different biotite generations

Staurolite

Compositions of the pre-Alpine and Alpine staurolites were measured and plotted on the Mg/Mg+Fe vs. Al4 plot (Fig. 15). Lelkes-Felvári et al. (1984) divided the staurolites to an old and a young generations and found that texturally different staurolites have different compositions. We were not able to make similar distinction on compositional grounds, because the field of Alpine staurolites overlap the field of the pre-Alpine ones on the Mg/Mg+Fe vs. Al4 plot (Fig. 15), although part of the Alpine staurolites are richer in Al4 than the pre-Alpine ones.

We collected several chemical composition data of staurolites from low pressure (Abukuma-type) and medium pressure (Barrow-type) terrains to see if staurolites from these two terrains can be distinguished on compositional grounds. Plotting data of low pressure (Abukuma-type) and medium pressure (Barrow-type) staurolites from the literature makes it clear that there is no

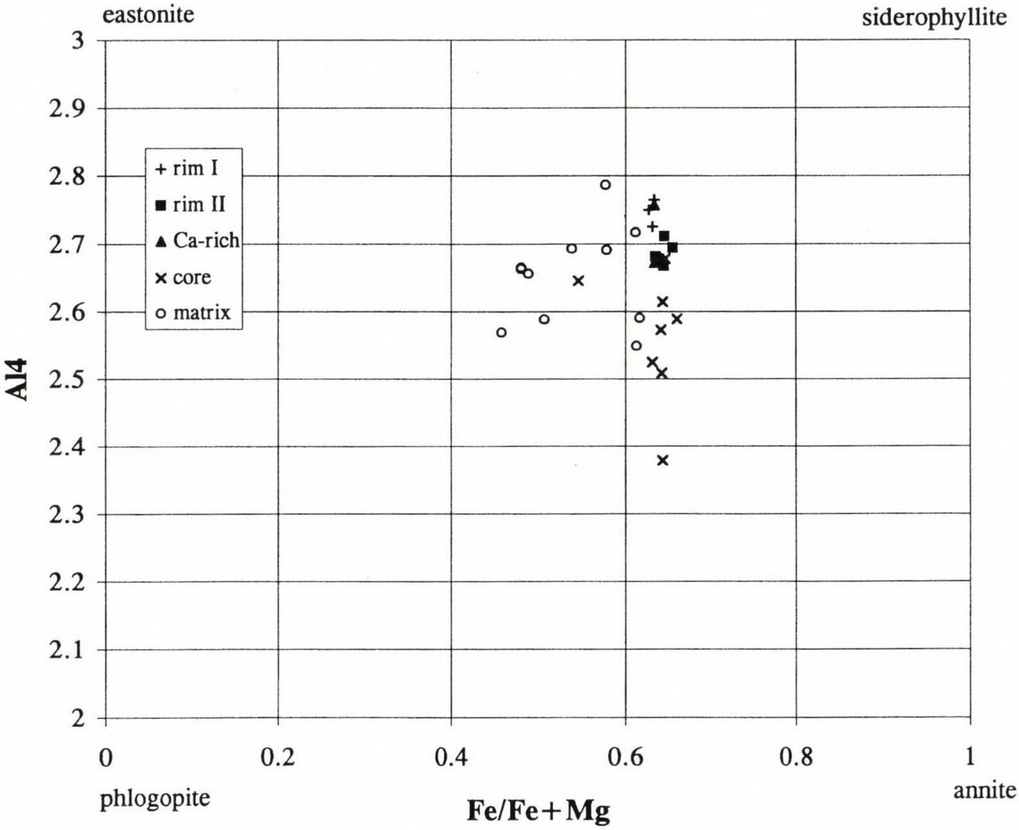

Fig. 14
Fe/Fe+Mg vs. Al4 plot of different biotite generations

systematic distinction between the compositions of low pressure and medium pressure staurolites. This may imply that composition of staurolites is governed rather by the rock chemistry than the pressure. Representative staurolite analyses are shown in Table 4.

Spinel

Spinels can be classified as zincian hercynites, with a considerable amount of spinel ($MgAl_2O$) content. The measured compositional range is quite restricted, but it can be also due to the limited amount of analyses (Table 4).

Geothermo-barometry

Geothermo-barometric calculations were made with TWEEQUE program of Berman (1991) which uses thermodynamic data of Berman (1988, 1990), Chatterjee and Froese (1975), Fuhrman and Lindsley (1988) and McMullin et

Table 3
Representative plagioclase analyses. Cation numbers are on the basis of 8 oxygens

	1	2	3	4	5
SiO_2	63.74	62.24	63.30	63.57	63.26
Al_2O_3	22.87	23.26	23.05	23.05	22.51
CaO	3.98	4.39	4.02	3.94	3.03
Na_2O	9.47	9.11	9.62	9.38	10.55
K_2O	0.13	0.30	0.06	0.15	0.21
Sum	100.19	99.30	100.05	100.09	99.56
Si	2.810	2.777	2.797	2.805	2.813
Al	1.188	1.223	1.201	1.199	1.180
Ca	0.188	0.210	0.190	0.186	0.144
Na	0.809	0.788	0.824	0.803	0.909
K	0.007	0.017	0.003	0.008	0.012
Sum cat.	5.004	5.014	5.016	5.001	5.058
or	0.727	1.675	0.324	0.842	1.116
ab	80.563	77.649	80.980	80.483	85.344
an	18.71	20.679	18.695	18.674	13.540

1. Plagioclase inclusion in the core of the garnet; 2. Plagioclase in contact with Ca-rich zone of the garnet; 3. Plagioclase in contact with rim II of the garnet; 4. Plagioclase in contact with rim I of the garnet; 5. Plagioclase core.

al. (1991). All pressure and temperature data were obtained as averages of three measurements.

Pressure and temperature data were obtained from two rim compositions of garnet with coexisting biotite, muscovite and plagioclase and from a core composition coexisting with inclusions of plagioclase and biotite. This garnet has discontinuous zoning, which exposes the relatively Ca-rich zone and another inner zone with lower Ca and higher Mn content. These zones are also in contact with biotite, muscovite and plagioclase. Timing of the two rim compositions were made on the basis of the garnet profile shown in Fig. 11. The rim composition with smaller amount of spessartine and the higher amount of grossular is the older one (garnet rim I, Table 1) and the one with higher spessartine and lower grossular is regarded to be the younger one (garnet rim II, Table 1). Calculating with the core composition of the garnet and with the enclosed biotite and plagioclase (Tables 2 and 3) we assume that the coexisting Al_2SiO_5 species was sillimanite. In this case we obtain 303 MPa average pressure and 668 °C average temperature (range: 220–380 MPa and 640–710 °C). The calculations were also made with andalusite as coexisting Al_2SiO_5 species, but the changes in the obtained pressure and temperature were insignificant.

Geothermo-barometry of the relatively Ca-rich zone, with the contacting plagioclase, biotite and muscovite yields much higher pressure (769 MPa as average of three measurements), than the low Ca core and rims. The calculated temperature is slightly lower (634 °C, average of three measurements) than that calculated from the core of the garnet. The whole P–T range is between 670–890 MPa and 605–660 °C. The calculations were made also with the same Ca-rich garnet, contacting biotite and muscovite and with the core of the zoned plagioclase with relatively high albite content (see plagioclase composition section), assuming that this plagioclase core was in equilibrium with the Ca-rich garnet zone. In this case the obtained pressure is significantly higher and the temperature is slightly higher (860–1060 MPa, 615–666 °C).

The garnet rim I (Table 1) and the coexisting biotite, muscovite (Table 2) and plagioclase (Table 3) yield an average pressure of 447 MPa and an average

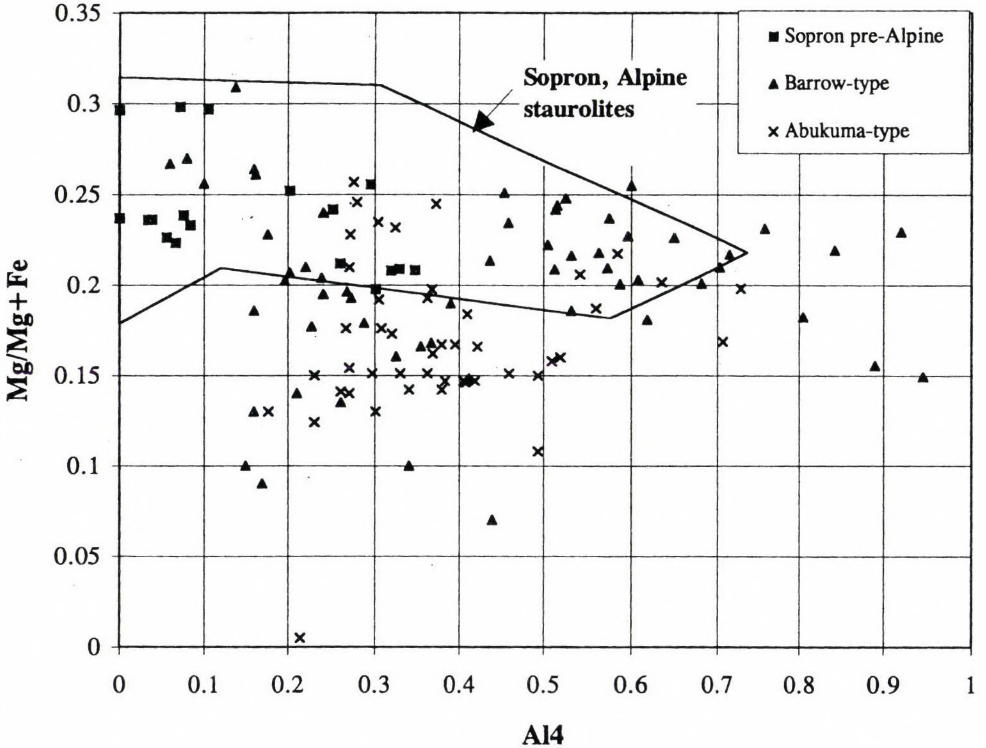

Fig. 15
Mg/Mg+Fe vs. Al4 plot of the pre-Alpine and Alpine staurolites from Sopron and staurolites of the Abukuma-type (low pressure) and Barrow-type (medium pressure) from the literature. For further explanation see text. The composition of the Alpine staurolites is given by the marked field. Staurolite analyses were taken from: Carlson and Nelis (1986), Crawford and Mark (1982), Florence and Spear, 1993), Goodman (1993), Guidotti (1974), Holdaway et al. (1988), Holdaway and Goodge (1990), Jones (1994), Kwak, (1974), Mohr and Newton (1983), Schumacher (1985) Selverstone et al. (1983), Török (1994) and Yardley et al. (1980)

Table 4
Representative spinel (columns 1 and 2) and pre-Alpine staurolite (columns 3–5) analyses. Cation numbers are on the basis of 4 oxygens for spinels and 48 oxygens for the staurolites

	1.	2.	3.	4.	5.
SiO_2	b.d.	b.d.	27.97	28.53	28.31
Al_2O_3	57.84	57.96	52.62	52.81	51.65
TiO_2	b.d.	b.d.	0.63	0.16	0.57
MgO	3.32	3.32	2.38	2.17	2.75
FeO_{tot}	23.87	24.24	12.57	12.52	11.61
ZnO	14.40	14.11	1.37	0.35	3.18
MnO	b.d.	b.d.	0.1	0.15	0.11
Sum	99.43	99.63	97.64	96.69	98.18
Si	–	–	7.798	7.966	7.895
Al	1.972	1.972	17.291	17.378	16.977
Ti	–	–	0.132	0.034	0.119
Mg	0.143	0.143	0.989	0.903	1.143
Fe^{2+}	0.577	0.585	2.931	2.923	2.708
Zn	0.308	0.301	0.282	0.072	0.655
Mn	–	–	0.023	0.035	0.025
Sum cat.	3.000	3.001	29.447	29.311	29.522
Spi	14.31	14.28			
Her	54.99	55.70			
Gah	29.32	28.62			
Fra	1.38	1.41			

temperature of 637 °C (range: 400–480 MPa, and 607–670 °C). The outer, discontinuous garnet rim II (Table 1) yields an average pressure of 340 MPa and an average temperature of 575 °C (range: 240–400 MPa and 564–580 °C).

Discussion

Progressive metamorphic P–T path

Most of the information on progressive metamorphic evolution of the rock can be obtained from two textural domains:

1. The plagioclase+sillimanite+biotite association, where K-feldspar associated with sillimanite can also be observed.

2. The andalusite porphyroblasts and related biotite, sillimanite, staurolite, spinel and corundum association. The formation and decomposition reactions of staurolite are the most relevant for the determination of the prograde pressure-temperature evolution of the rock.

1. The plagioclase-sillimanite-biotite domain

Plagioclase with sillimanite inclusions are thought to be the result of the reaction:

$$paragonite+quartz=albite+sillimanite+H_2O \quad (1)$$

Curve of this reaction in the P-T space was calculated from data of Berman (1988) using the TWEEQUE program (Fig. 16). The sillimanite stability field (after Holdaway, 1971) gives the lower and upper pressure limits between 320 and 480 MPa at minimum temperatures of about 550–600 °C. Timing of this reaction in the prograde metamorphic path relative to the staurolite breakdown reactions to be discussed below is a little ambiguous, because these two domains are rarely in contact with each other. On the basis of sillimanite inclusions in andalusite porphyroblasts (Fig. 6) it seems to be a reasonable approach that this assemblage has formed prior to andalusite formation and the staurolite breakdown. This is also supported by the fact, that this reaction is well within the stability field of the staurolite-muscovite-quartz assemblage. Lelkes-Felvári et al. (1984) have also found sillimanite inclusions in andalusite and point out that sillimanite has also formed prior to the growth of the andalusite porphyroblasts.

Sillimanite needles enclosed in K-feldspar provide the evidence on the further existence of muscovite in this assemblage. This muscovite became poor in paragonite component as the reaction (1) took place and reacted with quartz to form K-feldspar and sillimanite at higher temperature, at a later stage of metamorphism. The reaction curve in Figs 16 and 17 is drawn after Chatterjee and Johannes (1974). However muscovite was observed in this domain, which is considered to be a product of retrograde metamorphism (i.e. the breakdown of K-feldspar+sillimanite assemblage). This observation is supported by the ragged boundary of the K-feldspars in contact with muscovites. Muscovites in the other parts of the rock can also be of retrograde origin, or locally may have persisted metastably. If this latter is true, the temperature may not have exceeded the stability of muscovite+quartz for a long time.

2. The andalusite-rich domain

Two possible explanations can be given for the origin of staurolite inclusions in andalusite porphyroblasts. The first explanation is, that these staurolites belong to an older metamorphic event, which is probably of Caledonian age. This explanation is favoured by Lelkes-Felvári et al. (1984). The other explanation is, that the staurolites were formed during the same low pressure metamorphism, which produced the andalusite. This version of staurolite formation seems to be supported by the P-T conditions of formation of the sillimanite+plagioclase assemblage from the reaction (1), which is within the stability field of the staurolite. However, as it was shown previously, distinction cannot be made between medium and low pressure staurolites on the basis of mineral chemistry. This would imply that neither of the explanations can be proved or disclosed unambiguously so far.

In the course of the progressive metamorphism, staurolites were consumed most probably by andalusite/sillimanite forming reactions, such as:

$$staurolite+muscovite+quartz=Al_2SiO_5+biotite+H_2O. \qquad (2)$$

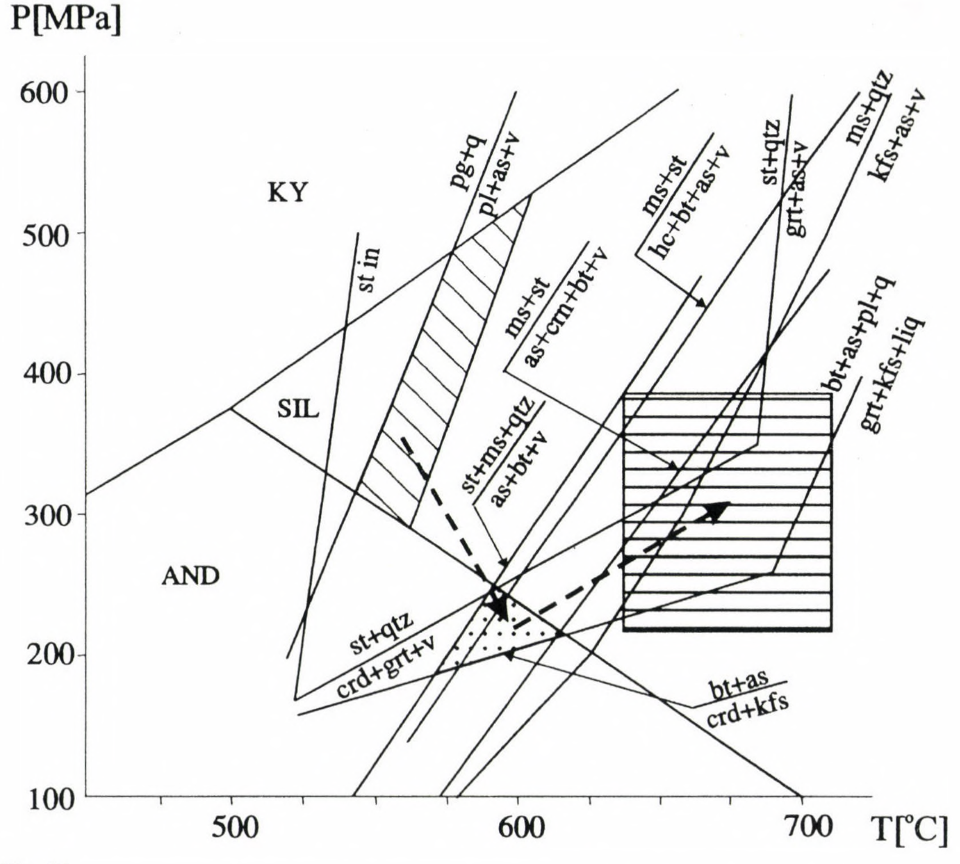

Fig. 16
Progressive metamorphic P-T path inferred from mineral equilibria and from the observed mineral reactions. The hatched box of the temperature peak was obtained from geothermo-barometry of the garnet core and the enclosed biotite and plagioclase inclusions. The ms+st=as+crn+bt+v and the ms+st=hc+bt+as+v reactions were calculated from Berman (1988) database using the TWEEQUE program. The curve of st+qtz+ms=as+bt+v is after Hoschek (1969). Al_2SiO_5 stability fields are after Holdaway (1971). ms+qtz=kfs+as+liq. reaction curve is after Chatterjee and Johannes (1974), staurolite stability field is constructed from data of Richardson (1968), Ganguly (1972) and Rao and Johannes (1979). Biotite out reactions are after Le Breton and Thompson (1988) and Vielzeuf and Holloway (1988). Legend: as – Al_2SiO_5 species (kyanite, sillimanite or andalusite), v – vapour, liq. – liquid (melt), other mineral name abbreviations are after Kretz (1983)

Regarding that andalusite is very common in the rock, this reaction may have occurred mainly in the stability field of the andalusite. However this reaction did not consume all the staurolite. It has stopped most probably because either quartz or muscovite were consumed and staurolite is stable to higher temperatures when these two minerals are absent. Most probably quartz

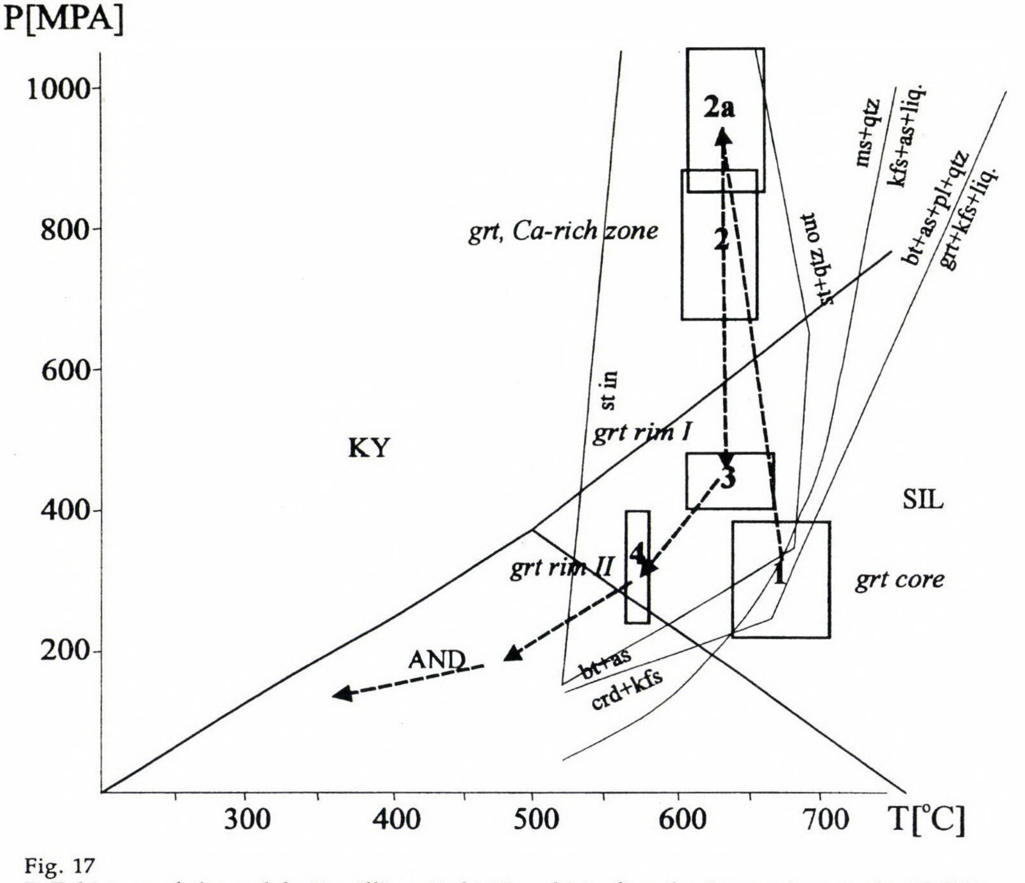

Fig. 17
P–T history of the andalusite-sillimanite-biotite schist after the temperature peak. Al_2SiO_5 stability fields are after Holdaway (1971). ms+qtz=kfs+as+liq. reaction curve is after Chatterjee and Johannes (1974), staurolite stability field is constructed from data of Richardson (1968), Ganguly (1972) and Rao and Johannes (1979). Biotite out reactions are after Le Breton and Thompson (1988) and Vielzeuf and Holloway (1988). P–T box 2 was drawn from the pressure-temperature calculated from Ca-rich garnet zone and plagioclase, biotite, muscovite in contact with it. P–T box 2a was drawn from the pressure-temperature calculated from Ca-rich garnet zone, biotite, muscovite in contact with it and the albite-rich plagioclase core. liq. – melt

was used up first, because reaction of staurolite with quartz results in formation of garnet and cordierite (e.g. Ganguly 1972). Since neither garnet nor cordierite were observed in the andalusites, it can be assumed, that quartz was consumed before the temperature reached the staurolite+quartz out curve. As the temperature grew it gave rise to further staurolite decomposition, forming corundum+biotite, or spinel. It should be noted however, that spinel may have formed from the reaction of staurolite with muscovite and quartz as described by Schumacher (1985) and Török (1994):

zincian staurolite+muscovite+quartz=
andalusite+Zn-bearing spinel+biotite+ ilmenite (3)

The conditions of formation of spinel in this reaction were estimated to be at temperatures of about 600 °C at about 200 MPa pressure (Török 1994). Similar P-T conditions can be applied for the decomposition of staurolite in this case as well (stippled area in Fig. 16). P–T conditions of this reaction are limited by the andalusite stability field and the biotite+andalusite stability field. Since there is no trace of decomposition of biotite+andalusite/sillimanite to cordierite+K-feldspar, the stability curve of the former assemblage provides us with the lower pressure limit. The andalusite stability field gives the upper pressure limit and the upper temperature limit. The low temperature side of the P–T box is limited by the breakdown curve of the staurolite+muscovite+ quartz assemblage (Fig. 16). These reactions and stability fields define well the pressure and temperature conditions between 180–250 MPa and 575–620 °C.

Another possible breakdown reaction of staurolite was calculated in the KFMASH system from database of Berman (1988) using the TWEEQUE program (Fig. 16).

muscovite+staurolite=hercynite+biotite+andalusite/sillimanite+vapour (4)

Curve of this reaction is very close to the first breakdown of the staurolite+muscovite+quartz assemblage and fits well into the P-T box of the spinel+andalusite formation, determined above. The formation of spinel may have proceeded with rising temperature in the sillimanite stability field as well, because there are some observations on the coexistence of these two minerals in the andalusite porphyroblasts.

Formation of corundum may have occurred at somewhat higher temperatures, close to the temperature peak of the metamorphism, which was obtained from geothermo-barometry of garnet core and enclosed biotite+ plagioclase inclusions and from the existence of the sillimanite+K-feldspar assemblage. This is also demonstrated by the curve of a possible corundum-forming reaction calculated in the KFMASH system using database of Berman (1988) and the TWEEQUE program (Fig. 16).

muscovite+staurolite=andalusite/sillimanite+corundum+biotite+vapour (5)

Breakdown of staurolite to sillimanite and corundum was also described by Goodman (1993) from quartz-poor metasediments of the late Proterozoic Dalradian Supergroup, in east Glen Muick, Scotland. This reaction took place at higher pressures (up to 700–800 MPa), and somewhat higher temperatures (up to 750 °C). The higher temperatures recorded by Goodman (1993) seems to be reliable, regarding the slope of the reaction curve (5).

Simultaneous formation of spinel and corundum is demonstrated in some places, where these two minerals coexist in andalusite porphyroblasts (Fig. 3). Formation of spinel and corundum may have stopped when both quartz and muscovite was used up. This is demonstrated by the fact that no quartz or muscovite was observed as inclusions in deformed andalusite in the vicinity of the spinel and corundum.

Another cause of survival of staurolites to higher temperatures can be their elevated ZnO content. Substitution of Fe/Mg to Zn in the staurolites causes expansion of its stability field both to the higher temperature region (e.g. Goodman 1993) and to the lower temperatures (e.g. Soto and Azañón 1994). Other minor elements, such as Li have the similar effect (Dutrow et al. 1986).

The above-described spinel and corundum forming reactions and partial retrograde metamorphic decomposition of spinel and corundum and formation of new, Alpine staurolite show that the andalusite porphyroblasts did not perfectly protect the inclusions from communicating with matrix minerals and fluids. So small portions of fluids, circulating in the cracks of the rock, may have infiltrated into the andalusite and caused retrograde reactions of the inclusions. The sericite aggregates, enclosed in the andalusite porphyroblasts may also be an indication of this process. The amount of these fluids may not have been enough for the complete decomposition of the andalusite porphyroblast itself, and in most cases they left behind relics of the inclusion spinel as well. However in those parts of the rock where the Alpine overprint was stronger the andalusite is fully replaced by a retrograde metamorphic fine grained sericite-muscovite aggregate or by an Alpine staurolite+kyanite+ muscovite assemblage. However corundum was preserved in some places in the latter assemblage as well (Fig. 2). Such reactions of the inclusions within a host mineral are not unusual and were described by e.g. Whitney (1991) in case of garnets.

The peak temperature and pressure conditions can be given by mineral equilibria and geothermo-barometry of the garnet core and the enclosed biotite and plagioclase inclusions. Mineral equilibria gives a quite restricted temperature constraint (Fig. 16). Existence of K-feldspar+sillimanite assemblage gives a lower temperature constraint while the biotite+sillimanite+ plagioclase assemblage provides the higher temperature limit. The lower pressure limit is given by the biotite+sillimanite assemblage. These constraints provide a pressure interval between 240 and 380 MPa at temperatures between 650–700 °C. These pressure-temperature constraints are in good agreement with the geothermo-barometry of the garnet core and the biotite+plagioclase inclusions and show that the temperature was between 640–680 °C at pressures between 200 and 400 MPa.

Summarising the above mineral equilibria and geothermo-barometric results, the prograde metamorphic P–T path can be traced back until the paragonite component of the muscovites reacts with quartz to form plagioclase and sillimanite at pressure constraints between 320 and 480 MPa at the minimum temperature of about 550–600 °C. After this reaction the pressure decreased at slowly increasing temperature and the breakdown of staurolite and the formation of andalusite took place. The minimum pressure conditions are about 180–250 MPa, which is given by the stability of the biotite+andalusite/ sillimanite assemblage, at about 575–620 °C temperature. As the temperature increased with slowly increasing pressure, breakdown of staurolite continued to form spinel, corundum, biotite and sillimanite. The last reaction is the breakdown of muscovite+quartz assemblage to form K-feldspar+sillimanite. This means that the metamorphism reached the granulite facies conditions. This high temperature is also supported by geothermo-barometric calculations from garnet core and the enclosed biotite and plagioclase. The prograde metamorphic P–T path, inferred from the above-described mineral reactions and from the geothermo-barometry of the garnet core and the enclosed biotite and plagioclase is shown in Fig. 16.

Metamorphic P–T path after the temperature peak

The further metamorphic P–T evolution of the rock is documented by geothermo-barometry of the zoned garnet described in the geothermo-barometry section (Fig. 17). After the metamorphic temperature peak, which is marked by the almandine-rich and Ca-poor core of the garnet, the pressure increased drastically. This increase in the pressure is marked by the Ca-rich zone of the garnet and the albite-rich core of the plagioclase in the vicinity of the garnet. This pressure increase means an intense subsidence, which can be attributed to nappe stacking in the area after the high temperature event at relatively shallow depths. Compositional changes in the Ca-rich part of the garnet in the Fig. 11a show that the high pressure history of the rock may be more complicated than described here.

The compositional uniformity of most of the measured plagioclase grains in different textural positions seems to cause some ambiguity in the geobarometry. In case of pressure increase the anorthite component of the plagioclase decreases and grossular component in garnet increases in coexisting garnet and plagioclase. Increase in grossular component in the garnets is clearly detectable, but decrease in anorthite component of plagioclases is not as unambiguous. There can be three reasons for the homogeneity of most of the plagioclase.

1. Plagioclase reequilibrated after the medium pressure metamorphism. In this case the pressure obtained from Ca-rich garnet zone and the contacting plagioclase, biotite and muscovite can be regarded as a minimum estimate.

This is well demonstrated by the calculations where the pressure and temperature was obtained using Ca-rich garnet zone and the relatively albite-rich core of the zoned plagioclase and the contacting biotite and muscovite. In this case the obtained temperature was higher by 9–10 °C, which is insignificant but the estimated pressure increase was between 170–200 MPa. However re-equilibration of plagioclase is not very probable, because cation diffusion in plagioclase is known to be quite slow (Grove et al. 1984). Thus if plagioclase is produced, it cannot change its composition easily except by dissolution and reprecipitation.

2. The composition of plagioclase did not change during the medium pressure event, it remained metastable and was not in equilibrium with garnet during the medium pressure event. In this case the Ca for the grossular in the garnet came from another Ca-bearing phase or from an external fluid. In this case it is hard to assess the pressure conditions for this event. As the Ca-rich garnet is in contact with plagioclase and sillimanite, it can be deduced that the pressure constraints can be given by the sillimanite stability field. This point is anticipated by the lack of any other Ca-bearing phase in the rock and by the lack of visible alteration caused by fluids which might have been the other source for the Ca.

3. Mass balance between plagioclase and garnet. The amount of pre-Alpine garnet relative to plagioclase is very small. Regarding the portion of Ca-rich garnet in the pre-Alpine garnets relative to plagioclase, this amount is even smaller. When the pressure increased the plagioclase dissolved to form grossular component in the garnet. The amount of Ca-rich garnet was so small that it did not need much anorthite component to dissolve and thus the amount of the remaining albite-richer plagioclase is also small. Thus the apparent uniformity in the plagioclase composition may be caused by the small amount of dissolved anorthite component which reacted to form grossular component in garnet. This is supported by the scarcity of zoned plagioclase. Summarising the arguments in the three points the third one seems to be the most reasonable explanation for the uniformity in plagioclase composition.

Although the pressure and temperature conditions of this metamorphic stage are within the stability field of kyanite, there is no conclusive evidence in the mineralogy, which proves this high pressure. There are some places, where kyanite inclusions can be observed in the andalusite (Fig. 9), but the textural relationship between andalusite and kyanite does not give conclusive evidence for the formation of kyanite in this metamorphic event. Lelkes-Felvári et al. (1984) also described kyanite inclusions in andalusite and they regard this observation as an evidence for the former (Caledonian) medium pressure metamorphism. Regarding both the pressure-temperature conditions obtained from the Ca-rich zone of the garnet and the textural observations, there are three possible explanations for the formation of kyanite inclusions in andalusite.

1. The kyanite inclusions in andalusite can be relic phases of the Caledonian metamorphism, as it was assumed by Lelkes-Felvári et al. (1984). However, it

should be noted that compositional zoning of garnet and geothermo-barometry does not provide conclusive evidence on the existence of the Barrow-type metamorphism prior to the low pressure metamorphic event discussed in this paper. Similarly, staurolites also do not provide any line of evidence for that.

2. The kyanite inclusions can be attributed to the higher pressure phase within the Hercynian metamorphic event (Fig. 17) which is described here.

3. The kyanite inclusions in andalusite have formed during the Alpine metamorphism, similarly to Alpine staurolite and staurolite+kyanite within the andalusite porphyroblasts. There are a lot of examples for Alpine breakdown of andalusite to kyanite+staurolite.

However Peindl (1990) described kyanite pseudomorphs after andalusite in a contact aureole from the Grobgneis of the Waldbach valley (Lower Austroalpine Unit), which means a pressure increase after the low pressure event. This pressure increase might be correlated with the pressure increase shown by the Ca-rich zone of the garnet in the andalusite-sillimanite-biotite schists of the Kovács-árok. However the Hercynian P–T path given by Neubauer et al. (1992) places the higher pressure event before the thermal peak.

The first evidence on Hercynian high-pressure event was provided by Dallmeyer et al. (1992) on the basis of Si content of white mica of the Wechsel gneiss (Lower Austroalpine Unit) from the Wechsel Window. They described zoned phengitic white mica, with 3.5 Si atoms p.f.u. in the core and 3.25 Si atoms p.f.u. in the rim. The high-Si white mica core shows a high-pressure phase at about 370–380 Ma. The age of this phase is given by the Rb-Sr method. The relatively high pressure, recorded by the Ca-rich zone of garnet from the Kovács-árok andalusite-sillimanite-biotite schist may be correlated with that described by Dallmeyer et al. (1992).

Ca-poor rim I and rim II compositions mark significant decrease in pressure and slow cooling again approaching the stability field of the andalusite (P=400–480 MPa, T=607–670 °C and P=240–400 MPa, T=564–580 °C). Formation of the clear, not deformed andalusite overgrowth on mosaic-like, inclusion-rich andalusite, which contains the above-discussed staurolite, spinel, corundum and sillimanite inclusions, can be attributed to this metamorphic stage. This second andalusite generation and re-equilibration under low pressure conditions may have obliterated the high pressure mineral assemblage. The subsidence and uplift may have been a rapid process and the low pressure minerals may have metastably survived the higher pressure conditions.

The age of the metamorphism was estimated by Balogh and Dunkl (1994) with the K/Ar method. The oldest age data were measured on separated biotites of he andalusite-sillimanite-biotite schists from the Kovács-árok which are between 319.5±12.1 and 328±12.5 Ma. This age was interpreted as cooling age which can mark an age after the formation of low pressure garnet rims.

Similar P-T path was determined by Schulz (1992) from zoned garnet of biotite-sillimanite gneiss, in the Moldanubian diaphtorite zone, west of Waldthurn in the Bohemian Massif in NE Bavaria. He obtained 550 °C and 300

MPa for the garnet cores, 575–650 °C and 1400–1800 MPa for the Ca-rich garnet zone and 700 °C and 300 MPa for the garnet rims. In spite of the high pressure metamorphism, well within the stability field of the kyanite, the author did not find any trace of kyanite in the samples.

The role of diffusion in the geothermo-barometric results

The high peak temperatures obtained from the geothermo-barometry and mineral equilibria raise the question of how significant can the diffusional processes be and how they influence the results of the geothermo-barometry. Diffusion at high temperatures may modify and finally completely homogenise garnet zoning. Although garnet commonly tends to homogenise its zoning between 650–750 °C (Woodworth 1977; Yardley 1977; Spear 1988) there are some indications when garnet zoning can be preserved at much higher temperatures than reported in this paper, within the granulite facies conditions (e.g. Indares 1995; Nichols 1995; Chen et al. 1998). Though the peak metamorphic temperature obtained from the core of the garnet of the Kovács-árok andalusite-sillimanite-biotite schist fits with the lower temperature limit of homogenisation of garnets, the measured garnets maintained their zoning. This may be due to the decreasing temperature during garnet growth and development of zoning. However the question if the zoning was modified by diffusion during retrograde metamorphism still remains open. Examination of zoning profiles of diffusion and growth zoning in garnets from the literature may be of help to solve this question. Comparing the typical growth zoning patterns and the diffusion zoning patterns of the garnets in the literature (e.g. Tracy et al. 1976; Woodworth 1977; Yardley 1977; Spear et al. 1990), it seems, that the zoning of the garnets in the Kovács-árok andalusite-sillimanite-biotite schist resemble rather to the growth zoning patterns than the diffusional ones, because typical high temperature garnets with diffusional zoning show large homogeneous cores and thin diffusional rims with sharply changing almandine and pyrope profiles.

Assessing the Fe-Mg exchange reaction between biotite inclusion and the host garnet during retrograde metamorphism is very important in estimation the peak metamorphic temperature. As the temperature decreases after the peak of the metamorphic event, the equilibrium Fe/Mg of the garnet increases and that of the biotite decreases. This exchange reaction and these changes in the Fe/Mg ratios in the garnet and biotite result in underestimation of peak metamorphic temperature. However the peak temperature of the Kovács-árok andalusite-sillimanite-biotite schist is quite well constrained by the muscovite+quartz out reaction and the existence of the biotite-sillimanite-plagioclase and the biotite-sillimanite assemblages. The good agreement between the results of geothermo-barometry and the P–T constraints given by mineral assemblage shows that the Fe-Mg exchange reaction between biotite and garnet and the induced diffusion caused only minor effect in the estimation of peak temperature by garnet-biotite thermometry. This minor diffusion may be the cause of the scatter of the temperature data.

Diffusion between garnet rims and the contacting biotite may also cause the underestimation of the temperature of the retrograde metamorphic P–T path. This effect can be minimised if the amount of modal biotite is large and the garnet crystal is also large, because in this case the diffusive flux is small and does not change the composition of biotite appreciably (Spear 1993). This applies to the case of the andalusite-sillimanite-biotite schists of the Kovács-árok, because the modal amount of the biotite is large and the garnet crystals are relatively large. Unfortunately the diffusion caused possible shift of the P–T results obtained from geothermo-barometry of the garnet rim cannot be checked by the mineral assemblage as in the case of the peak temperature conditions, but on the basis of this mass balance consideration, we assume that underestimation of the retrograde metamorphic temperatures can be neglected, i.e. it is too small to have essential effect on the correct interpretation of the P–T path.

Modification of garnet rims by the Alpine overprint may have been also insignificant, because Alpine garnets have much higher Ca content, than the rims of the pre-Alpine ones and the pressure and temperature obtained from the rims of the garnet coincides well with the stability of the andalusite and sillimanite. If garnet rims were affected by the Alpine overprint they should have increasing grossular content at the rims, but it is not the case in the rims of the pre-Alpine garnets of the andalusite-sillimanite-biotite-schist of the Kovács-árok (Fig. 11 and 11a).

Summary

The pre-Alpine development of the andalusite-sillimanite-biotite schist of the Kovács-árok was deciphered from mineral reactions of two textural domains (plagioclase+K-feldspar+sillimanite+biotite and andalusite+sillimanite+staurolite+biotite+corundum+spinel) and from the geothermo-barometry of a zoned garnet. The first detectable reaction was the breakdown of paragonite to form sillimanite and plagioclase (minimum pressure: 320 and 480 MPa at minimum temperatures of about 550–600 °C). The appearance of andalusite and the staurolite breakdown reactions show an uplift and a slight increase in the temperature (P=180–250 MPa and T=575–620 °C). Appearance of spinel, corundum, sillimanite+K-feldspar and the geothermo-barometry of garnet core show an increase in temperature and slight subsidence of the area (P=240–380 MPa and T=650–700 °C). The sillimanite+K-feldspar assemblage indicates the beginning of the granulite facies.

After the thermal peak of the metamorphism the significant increase in the pressure indicates subsidence of the area, which is probably due to nappe stacking. This event might be correlated with the 370–380 Ma old high-pressure event, recorded in the Wechsel gneiss by Dallmeyer et al. (1992).

Geothermo-barometry of garnet rims show rapid exhumation and relatively slow cooling (P=400–480 MPa, T=607–670 °C and P=240–400 MPa, T=564–580 °C). 320-330 Ma K/Ar cooling ages on biotites (Balogh and Dunkl 1994) indicate the age of late, low temperature stage of the pre-Alpine metamorphism.

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Integrated palynostratigraphy of the Senonian formations in the Tisza Unit (South Great Hungarian Plain, Hungary)

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The author provides new data for the paleontological and stratigraphical knowledge of the Upper Cretaceous formations of the Tisza Tectonic Unit. The integrated pollen and dinoflagellate zonation and correlation of the areas of Bácska and along the Körös River, belonging to the Villány Zone, as well as the Izsák Formation of the Mecsek Zone (Fig. 5), are presented. The formations are assigned to the Coniacian–Late Campanian stages. The palynological results are well supported by correlation with the nannoplankton zones (CC12/13–CC22/23). Sedimentation occurred in the tropical-subtropical belt of the Normapolles Plant Geographical Province of the warm-water, marginal areas of Tethys. The youngest Upper Cretaceous shallow marine formations, recognized in the boreholes of Great Hungarian Plain, are of nearly the same age as the Late Campanian deeper-water, open marine sediments of the Transdanubian Central Range (henceforth abbreviated to TCR).

Key words: Tisza Unit, Villány Zone, Mecsek Zone, Upper Cretaceous, integrated palyno- (pollen and dinoflagellate) stratigraphy, nannozones, correlation, paleoenvironmental relations: Normapolles Province, shelf–open marine environment.

Introduction

Senonian formations of Bácska and along the Körös River, belonging to the Villány Zone, as well as the Izsák Formation of the Mecsek Zone of the Tisza Unit (Szentgyörgyi, in Árkai et al. 1996) were investigated with the aim of studying the palynologically little or not known formations and to establish the integrated pollen and dinoflagellate zonation of the Senonian of the Great Hungarian Plain.

The palynological investigation of the Bácska area was carried out on the basis of the study of boreholes Bácsalmás Ba-1, Csávoly-1 and Madaras Ma-5. In the region of the Körös River some boreholes of the surroundings of Komádi were studied (Komádi K-1, Komádi-2, -4, -7, -8, -10, -13). Since the sequence of borehole Izsák-1 proved to be barren the Izsák Formation was characterized only on the basis of borehole Nádudvar-Dk-3 (Fig. 1). The Ba-1 key section is continuous; the other hydrocarbon exploration wells were only partially cored.

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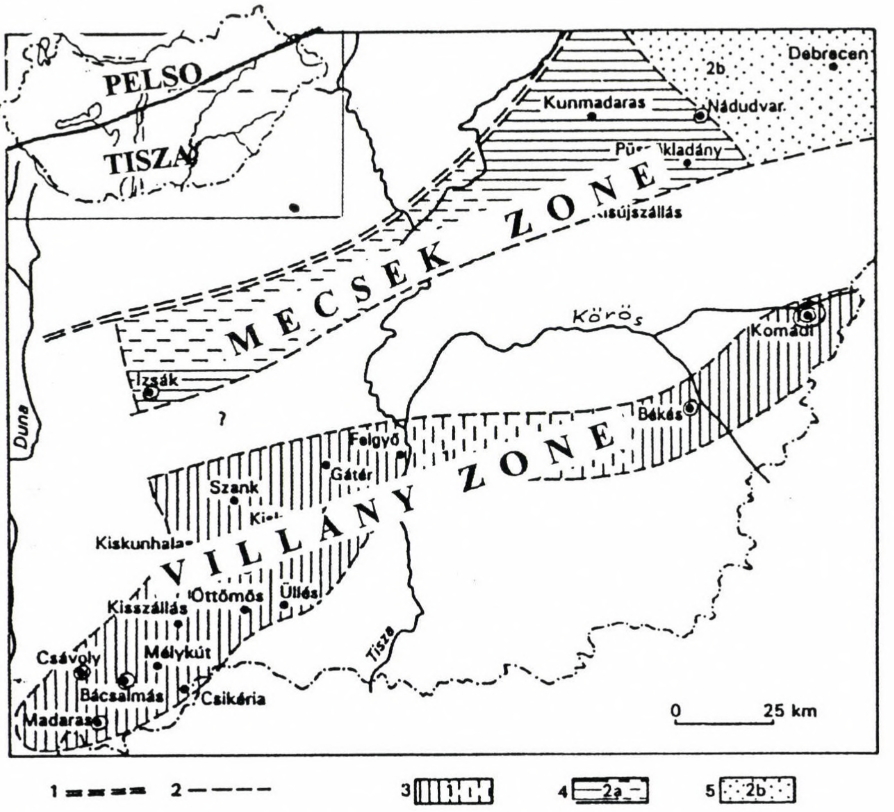

Fig. 1
Upper Cretaceous facies-belts in the basement of the Great Hungarian Plain. 1 main fracture-belt of Hungary; 2. facies borders; 3. epicontinental Senonian; 4. "Puho" Marl; 5. flysch. ○ Studied boreholes and areas (after Szentgyörgyi, in Árkai et al. 1996)

Bácska area

The continuous Senonian shelf formations of the southern part of the Danube–Tisza Interfluve can be assigned vertically into three formations: Szank Conglomerate Formation, Csikéria Marl Formation and Bácsalmás Formation (Szentgyörgyi, in Árkai et al. 1996).

Among the studied boreholes, the Szank Conglomerate Formation (which was formed mainly by the disaggregation of the older formations of the basement) was exposed only by borehole Ba-1 (793.0–816.3 m). The Csikéria Marl Formation was encountered in borehole Ba-1 (741.7–793.0 m) in its entirety; in boreholes Csávoly-1 (1460.0–1587.5 m) and Ma-5 (582.0–600.4 m) only in its upper section. The two latter boreholes stopped in this formation.

The Bácsalmás Formation was penetrated in its most complete development by borehole Csávoly-1 (1018.0–1460.0 m). As opposed to this thickness of more than 400 m borehole Ba-1 exposed the formation in a thickness of little more than 200 m (531.3–741.7 m) and borehole Ma-5 only in a thickness of 140 m (444.0–582.0 m).

On the basis of both the spore-pollen and the dinoflagellate investigations the sequences can be well correlated with each other; this is also confirmed by the nannoplankton investigations.

Borehole Bácsalmás-1

Borehole Bácsalmás Ba-1 was drilled in 1982 within the framework of the National Key Section Program. The multidisciplinary geologic investigation of the borehole was of great importance from the point of view of knowledge of the Senonian formations of the southern Great Hungarian Plain.

Besides the paleontological data (foraminifera, nannoplankton, palynology) chemical, mineralogical-petrologic, and pebble-analytical results contributed to the environmental knowledge and age determination of the Alföld formations of the Senonian basin (Haas 1987). The aim of the present study is the establishment of a pollen and dinoflagellate zonation of the Senonian formations of Bácska and the revision of the previous, slightly contradictory biostratigraphical results (Haas 1987; Siegl-Farkas 1986).

The investigated borehole data are of increased value as a standard because of the continuous coring and serve as a basis for the establishment of the palynozonation of sequences investigated in Bácska (Csávoly, Madaras) and in other areas of the Alföld (Komádi, Nádudvar), and for further correlation of the Senonian formations as well.

Upper Cretaceous formations were encountered by borehole Ba-1 in the depth interval 531.3–816.3 m. They were assigned by Haas (1987) into the following petrologic units (in order of deposition): Szank Conglomerate Formation (793.0–816.3 m), Csikéria Marl Formation (741.7–793.0 m), Bácsalmás Formation (531.3–741.7 m) with the members: marl with calcareous nodules – authigenic breccia (675.9-741.7), Csávoly Member (622.2–675.9 m), Madaras Sandstone Member (531.3–622.2 m). The subcrop of the Senonian sequence is constituted by the Triassic Jakabhegy Sandstone Formation and its cover by Lower Pannonian calcareous marl.

Spore-pollen investigations

Since detailed investigations were carried out after drilling the borehole (Siegl-Farkas 1986) only a brief overview is given here of the results and their up-to-date interpretation.

The Szank Conglomerate Formation is characterized by very scattered sporomorph occurrence. *Hungaropollis, Interporopollenites* and *Longanulipollis* genera, represented by one specimen each, are indicative of Upper Santonian–Lower Campanian sediment accumulation. From the Csikéria Marl and the Bácsalmás Formation a rich, well-preserved sporomorphs could be determined. On this basis the sequence was assigned to the *lenneri-bajtayi* Assemblage Zone of the Lower Campanian (718.4–793.0 m, lower section of the marl with calcareous nodules – authigenic breccia and the Csikéria Marl Formation), the *sahi* (641.1–718.4 m, middle section of the Bácsalmás Formation) and the *Plicapollis-Subtriporopollenites* (531.3–641.1 m, upper section of the Bácsalmás Formation) Subzones of the *Pseudopapillopollis* Assemblage Zone (Fig. 2), respectively.

Dinoflagellate investigations

Samples of the Szank Conglomerate Formation do not contain dinoflagellate remains. The Csikéria Marl can be characterized by rich, the lower two members of the Bácsalmás Formation by medium, and its upper member by scattered dinoflagellate association. The assemblage richest in species and density was preserved in the Csikéria Marl. The largest number of species in the sequence appears in the first marine layers deposited on the Szank Conglomerate. The appearance of some of them is restricted only to this formation. Associations of the Bácsalmás Formation are characterized by gradual impoverishment toward the younger layers; thus, associations of the three members can be well separated from each other. The uppermost sandy section proved to be barren.

Marine formations of the sequence were assigned into the *Isabelidinium microarmum* (728.0–792.4 m) and *Tarsisphaeridium geminiporatum* (542.6–792.4 m) Subzones of the *Odontochitina operculata* Assemblage Zone (Fig. 2).

The genus *Odontochitina*, which occurs only rarely, is represented by one specimen each of the *operculata*, *costata*, and *striatoperforata* species.

Microarmum Subzone (728.7–792.4 m)

The subzone was defined as ranging from the first sample containing phytoplankton to the regular occurrence of *Tarsisphaeridium geminiporatum*. The association of the subzone can be well separated from the younger assemblages. Consequently, in some places (747.8–792.3 m) frequent occurrence of *Isabelidinium microarmum* is the most characteristic feature. Species occurring only here are as follows: *Apteodinium* cf. *deflandrei*, cf. *Chatangiella* sp., *Dinogymnium acuminatum*, *D. microgranulosum*, *Isabelidinium* sp., *Micrhystridium stellatum*, cf. *Pervosphaeridium truncigerum*, *Spiniferites ramosus*, *Spiniferites belfastense*, *S. dentatus*, and *S. scabrosa*. Some of the listed species appear at the lower boundary of the subzone; others appear gradually upward. The entire Csikéria Marl Formation was deposited during the duration of this subzone.

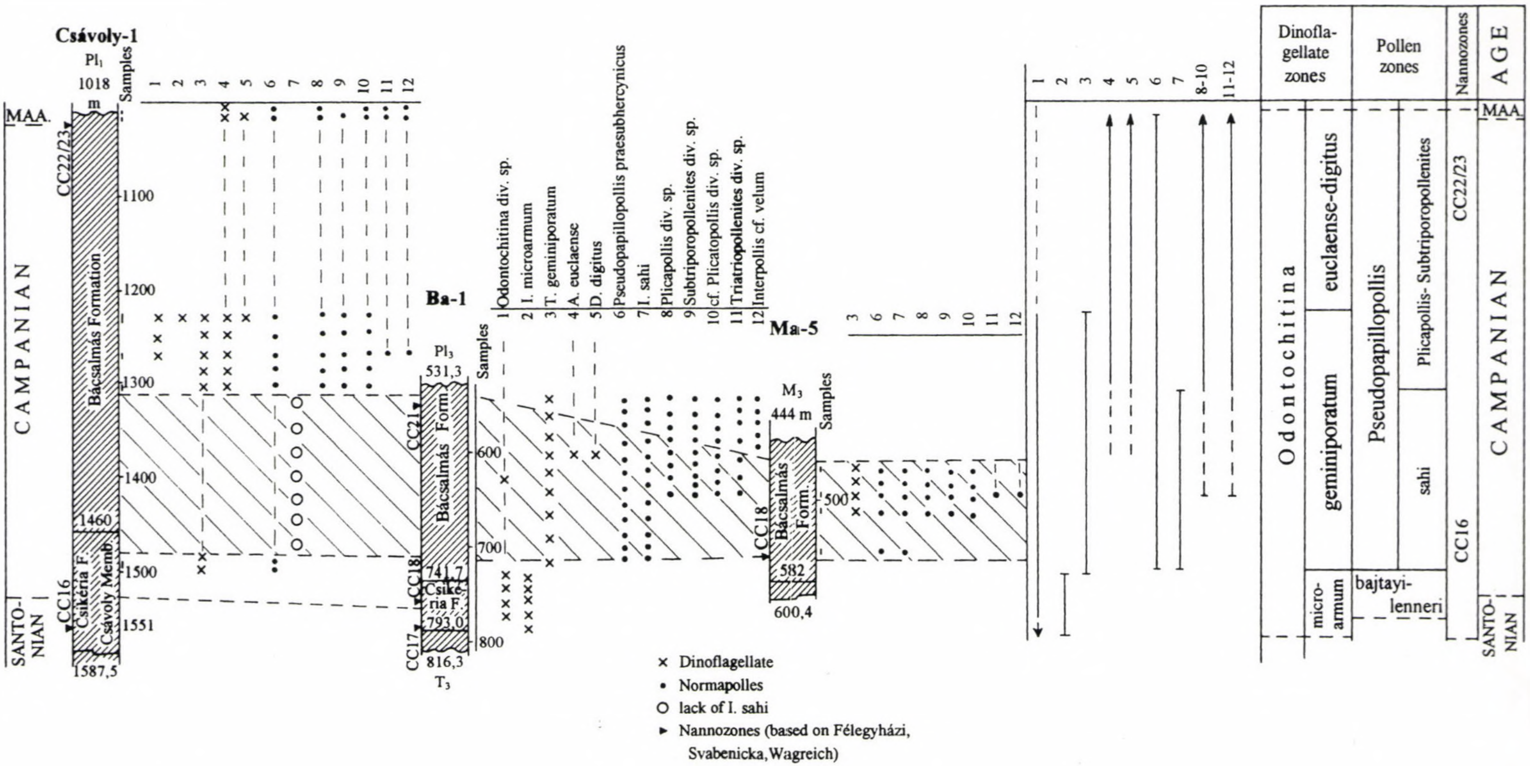

Fig. 2
Correlation of the Senonian formations in the Bácska Area

Geminiporatum Subzone (542.6–728.7 m)

The eponymous species is of regular occurrence in the association of the Bácsalmás Formation. On the basis of the alternations of the appearing and disappearing species as well as the more and more scattered assemblages, the subzone can be divided into four sections: in the deepest section the scattered and regular occurrence of *Alterbidinium varium* is characteristic (694.7–722.3 m, lower member of the Bácsalmás Formation). It is followed by an increasingly impoverished association, then by a section characterized by *Cribroperidinium othoceras*, with the most frequent occurrence of *Tarsisphaeridium geminiporatum* (621.1–659.7 m, upper section of the Csávoly Marl). The association of the section above it becomes poorer than the previous ones; here only the eponym of the subzone and some species of *Dinogymnium* and *Spiniferites* occur. In the uppermost sandy section (531.3–542.6 m) no dinoflagellate was determined. Almost the whole of the Bácsalmás Formation was deposited during the time of this subzone.

Age determination and correlational possibilities

Various previous paleontological and biostratigraphical investigations put the time of deposition of the Bácsalmás Senonian formations in different periods of the Senonian stage, similarly to the chronostratigraphical subdivision of the Senonian sequence of the TCR.

On the basis of the occurrence of *Globotruncana calcarata* as well as the absence of the Maastrichtian species the foraminifer investigations assigned the entire marine section of the Bácsalmás sequence to the Campanian stage of the Senonian (Sidó, in Haas 1987). With the occurrence of *Zygodiscus spiralis, Broinsonia parca* and *Quadrum gothicum* the nannoplankton data gave the age of the sequence as Upper Santonian-Upper Campanian (Félegyházi in Haas 1987). On the basis of the species listed by Félegyházi, using the zonation of Sissingh (1977) and Perch-Nielsen (1985) the sequence can be assigned to the CC17–CC21 Nannozones. By means of palynology, using the zonation and age determination of Góczán (1973), the sequence was assigned to the interval between the Upper Campanian and Upper Maastrichtian (Siegl-Farkas 1986; Siegl-Farkas in Haas 1987).

Since then, the age determination of the palynozones established for the formations of the TCR (Góczán 1964, 1973; Góczán and Siegl-Farkas 1990; Siegl-Farkas 1993a) had to be revised according to the integrated stratigraphical results which were achieved in the middle of the '90s. Correlation of the palynozones and nannozones (Siegl-Farkas 1993b; Siegl-Farkas and Wagreich, 1996) was completed by the introduction of the dinoflagellate zonation (Siegl-Farkas 1997) and the designation of the Santonian–Campanian boundary by means of integrated stratigraphical investigations (Lantos et al. 1996).

On the basis of the correlation of the palynozones and nannozones of the formations of the TCR as well as their alignment with the Santonian–

Campanian boundary, the sequence of the Bácsalmás borehole, correlated with them, should most sensibly be assigned to the Campanian stage.

In the present paper this is reliably proved by the correlation of the palynozones, dinoflagellate and nannoplankton zones established and introduced for the Senonian formations of Bácska (Fig. 2).

Paleoenvironmental conclusions

The facies analysis of the marine sequence of borehole Ba-1, using the terminology of Haas (1987), distinguishes deep, pelagic, slope and neritic environments in chronological order, which indicates rapid transgression, then subsequent slow regression.

This conclusion is also confirmed by the quantitative changes in the proportions of sporomorphs of terrestrial plants and marine phytoplankton. In the deeper section of the sequence the relatively rich dinoflagellate occurrence is associated with a much poorer sporomorph assemblage, while upward the phytoplankton association, which becomes poorer, is replaced by a more and more diversified and abundant sporomorph one (Siegl-Farkas 1986), verifying the regressional tendency. Facies changes shown by Haas (1987) are accompanied by dinoflagellate association changes, which verify them. Formations of the pelagic (*microarmum* Subzone) and slope facies (older section of the *geminiporatum* Subzone) can be characterized by relatively rich but variable associations and the neritic formations by much poorer ones.

In the Campanian, with its tripartite climate subdivision according to the "Williams suite", species of the determined *Isabelidinium, Altebidinium* and *Trithyrodinium* genera are characteristic members of the associations of warm water environments (Kirsch 1991).

Borehole Csávoly-1

The borehole exposed Upper Cretaceous formations in the depth interval of 1018.0–1587.5 m; Bácsalmás Formation: 1018.0–1460.0 m, Csikéria Marl Formation, Csávoly Member: 1460.0–1587.5 m. Its subcrop is constituted by older Cretaceous, while its cover by Miocene formations.

Palynological investigation was carried out on 16 cores of the Senonian section. Among them six samples – whose carbonate content exceeded 90% (Szentgyörgyi 1985) – proved to be palynologically barren. Thus, on the basis of the occurrence of palynomorphs, we can provide information only on the lower (1487.2–1551.0 m), middle (1228.0–1312.5 m) and uppermost (1018.5–1023.0 m) sections of the sequence. In two of the associations of the ten positive samples only spores-pollens were encountered, and in eight dinoflagellates were also determined.

Spore-pollen investigations

The uppermost part of the Csikéria Marl and the Bácsalmás Formation can be characterized by a richer occurrence of pteridophyte spores than the middle section of the section (e.g. *Bikolisporites toratus, Cicatricosisporites div. sp., Gleicheniidites senonicus, Trilites sp., Polypodiaceoisporites div. sp.*); at the same time, upward in the sequence, an increasing number of previously missing genera (eg. *Echinatisporites sp., Taurocusporites maastrichtiensis, Stereisporites sp.*) appear.

Gymnosperm pollen grains, as is characteristic in the Upper Cretaceous in this area of Tethys, occur only sporadically. *Alisporites sp.*, which is relatively frequent in the interval 1228.0–1232.0 m, joins the richest dinoflagellate association of the borehole. Upward in the borehole the number of occurring pollen grains of angiosperms increases, in the same way as in the Bácsalmás sequence, which indicates the emersional tendency of the basin. On the basis of their frequency the lower, middle and upper sections of the borehole can be well separated.

The predominance of the *Normapolles* genera is characteristic throughout. Younger *Normapolles* and *Postnormapolles* genera (*Concavipollis, Labrapollis, Nudopollis, Plicapollis, cf. Plicatopollis, Pseudotrudopollis, Romeinipollis, Subtriporopollenites* etc.) appear in the middle and upper sections of the borehole. The zonal index *Pseudopapillopollis praesubhercynicus* is predominant in the associations of the section between 1232.0–1310.5 m. Higher upsection it is replaced by the *Plicapollis-Subtriporopollenites* association in the samples. *Interporopollenites sahi*, missing from the associations, provided a good correlational possibility between the three Bácska sequences (Fig. 2).

Palmae-related pollen grains (*Monocolpopollenies sp., Gemmamonocolpites sp.*), determined from the associations of the samples taken in the depth intervals 1018.5–1023.0 m and 1487.2–1487.4 m, definitely indicate nearshore sediment deposition of a tropical-subtropical climate. The quartz sand content of these samples (30%) is the highest encountered in the section (Szentgyörgyi, in Árkai et al. 1996).

Formations of the sequence between 1018.0–1487.4 m were assigned to the *sahi* (1310.5–1487.1 m) and *Plicapollis-Subtriporopollenites* (1018.0–1310.5 m) Subzones of the Campanian *Pseudopapillopollis* Assemblage Zone. The section between 1487.4–1551.0 m cannot be assigned into a palynozone; its age is in all probability Late Santonian–Early Campanian.

Note: in this borehole, palynological investigations had already been carried out previously (Kedves 1984), presenting the spore-pollen associations of 1271.5–1490.0 m and 1567.5–1586.5 m and distinguishing them from those of the Bakony region as "Csávoly-type Upper Cretaceous Senonian".

Dinoflagellate investigations

The Csikéria Marl can be characterized by a very poor, the Bácsalmás Formation toward its middle section by an increasingly rich, and in its uppermost part again by a scattered phytoplankton association. Species of genus *Dinogymnium* are of regular occurrence. Richness in species of the middle section assumes an optimum biotope (*Alisogymnium euclaense, Biconidinium reductum, Dinogymnium digitus, D. heterocostatum, Gillinia hymenophora, Odontochitina operculata, Paralecaniella indentata, Tarsisphaeridium geminiporatum*). According to Szentgyörgyi (in Árkai et al. 1996) the mineral composition of this section is as follows: 10% dolomite, 20% quartz and 70% calcite.

On the basis of the determined species the middle section of the borehole (1228.0–1487.4 m) is assigned to the *geminiporatum* and the formations above it (1018.0–1228.0 m) to the *euclaense–digitus* Subzones. The deeper section did not contain dinoflagellate remain suitable for zonation.

Nannoplankton investigations

One sample of the deepest section of the Csikéria Formation as well as the middle and uppermost sections of the Bácsalmás Formation each was studied. According to the investigations, the sequence was deposited in the time interval between the Late Santonian CC16 and the Late Campanian (Early Maastrichtian?) CC22/23 nannozones.

Determined taxa: 1551.0 m: *Lithastrinus grillii Stradner, Reinhardtites anthophorus (Deflandre) Perch-Nielsen, Rucinolithus wisei Thierstein, Lucianorhabdus cayeuxii Deflandre, Biscutum hattneri Wise.* 1228.0–1232.0 m: *Broinsonia parca constricta Hattner et al., Microrhabdulus attenuatus Deflandre.* 1018.5–1023.0 m: *Quadrum sissinghii Perch-Nielsen, Quadrum trifidum (Stradner) Prins and Pearch-Nielsen, Ceratolithoides aculeus (Stradner) Prins and Sissingh, Ceratholithoides quasiarcuatus Burnett, Reinhardtites levis Prins and Sissingh.* This assemblage is characteristic for the CC22 Zone but can be correlated within the range up to the CC23 Zone (the Campanian/Maastrichtian boundary). Determination of the species and the zone classification is the works of L. Svabenicka.

Borehole Madaras-5

The borehole, which stopped in the Csikéria Marl, exposed Upper Cretaceous formations in the depth interval 444.0–600.4 m. Its lithostratigraphical subdivision is as follows: Bácsalmás Formation 444.0–582.0 m, Csikéria Marl Formation 582.0–600.4 m. Miocene layers make up the cover of the sequence.

Sporomorph investigations

In the section between 473.2–576.9 m of the borehole (discontinuously cored) a palynological investigation of 5 samples was completed. The sample taken from the Madaras Member (conglomerate) of the Bácsalmás Formation (575.2–565.2 m) did not contain any sporomorphs. The one taken from 564.4–565.2 m contained only Normapolles, while the sporomorph association of the section between 473.2–521.0 m also contained dinoflagellates in small amounts. The positive samples were found in the limestone and sandy limestone rocks, respectively, of the Bácsalmás Formation.

In the assemblages the Normapolles group is predominant. The zonal index *Pseudopapillopollis praesubhercynicus* is rare in all samples but the subzonal index *Interporopollenites sahi* is of frequent occurrence.

The subzonal index fossil of the dinoflagellate zonation *Tarsisphaeridium geminiporatum* occurred, though in small numbers, but in the association of every sample. In the territory of the TCR *Nelsoniella aceras* (520.0 m) was determined in the same subzone (Gyepükaján-9, upper section of the Polány Marl). *Gillinia hymenophora* (473.2 m) was also distinguished in the formations of same age in southern France (Tercis) (Odin et al. 1998). In Hungarian formations, in the territory of the TCR, it has so far only been found in the Ganna Member (1497.0 m) of the sequence of the borehole Nagygörbö-1 as well as during the present work in the sample, taken at 1228.0 m of the borehole Csávoly-1. The fact that its appearance is linked with sandy formations in all the three sequences may refer to the nearshore (shelf) environmental requirements of the species.

The Middle Campanian age designated by the zonal index palynomorphs is confirmed only approximately by the nannoplankton zonation, according to which the deposition of the formations took place in the time interval between the zones CC18–CC22 (classification of Wagreich, this study).

Correlation of the Senonian formations of the Bácska area

On the basis of both the sporomorph and dinoflagellate investigations the associations of the presented sequences can be well compared and reliably correlated with each other (Fig. 2).

For the correlation by sporomorphs a reliable basis was provided by *Interpollis cf. velum, Pseudopapillopollis praesubhercynicus, Plicapollis serta, P. div. sp., Plicatopollenites plicatus, P. div. sp., Subtripollenites div. sp.,* and *Triatriopollenites sp.,* appearing regularly in all three boreholes, and *Interporopollenites sahi,* missing only from the associations of the borehole Csávoly-1.

The dinoflagellate data are less frequent but can be well fitted into the sporomorph data and complete them. For this case occurrences of *Alisogymnium euclaense, Biconidinium reductum, Dinogymnium digitus, D. sphaerocephalus, Gillinia hymenophora, Isabelidinium microarmum, Odontochitina operculata, O. div. sp.,* and *Tarsisphaeridium geminiporatum* were used. Correlation by means of palynology was supported by nannoplankton zonation in every instance.

The thickest Csávoly sequence, considered as the most basinal one, comprises the greatest time interval. The lower section of the Csikéria Marl Formation, which became known here, is the oldest (Santonian) formation and the upper section of the Bácsalmás Formation is the youngest (Campanian; possibly near the Campanian-Maastrichtian boundary?). The initial stage of the deposition of the Bácsalmás Formation can be put into the nearly same time plane in the three sequences; thus, the Bácsalmás and Madaras sections can be reliably correlated with the Csávoly one. We consider the Szank Formation, penetrated only in the Bácsalmás sequence, coeval with the Santonian-Campanian boundary section of the Csikéria Marl Formation. The basin in the surroundings of Bácsalmás was in a structurally higher position than the other two ones; here marine sedimentation began later. The erosion of the Bácsalmás and Madaras sequences deposited during the Lower Campanian, according to the evidence of the cover formations, took longer than in the surroundings of Csávoly. Since the subcrop(s) of the Madaras and Csávoly boreholes are not known the timing of the transgression cannot be exactly determined here.

Investigation of the Körös Formation

Two types of the formation are distinguished by Szentgyörgyi (1982, 1983, 1985). The reference section of the facies consisting of terrigenous rocks is the borehole Komádi-4, while the type section of the sequences deposited on the basal breccia is borehole Komádi-10.

Terrigenous (sandstone and siltstone) facies: borehole Kom-4, -7, -8, Kom-K-1 (Fig. 3).

The Körös Formation is encountered in its maximum thickness (1000 m) in the Kom-4 borehole. The subcrop of the Senonian formations (2111.0–3125.0 m) is constituted by Lower Cretaceous and their cover by Miocene sediments. In the section of the borehole between 2147.0–3009.0 m, palynological investigation of 19 samples was undertaken, of which 5 proved to be barren. Thirteen samples contained only sporomorphs of terrestrial origin. In the 2183.0–2188.5 m interval only a single *Micrhystridium sp.* indicates a marine environment. According to Szentgyörgyi (in Árkai et al. 1996) the sediments were deposited on shelf slope and outer shelf.

The deepest investigated sporomorph associations of the borehole (2847.0–2853.0 and 3000.9 m) contained no angiosperm pollen except for some pteridophyte spores. The samples above it may be characterized by assemblages becoming increasingly rich upsection. Among the pteridophytes the occurrence of *Bikolisporites toratus* is regular. Of the *Normapolles* genera, appearing in medium amounts, mostly the "primitive" ones (still with a thin wall but already complicated pore structure) with the *Tenerina sp., "Latipollis" sp., cf. Monstruosipollis sp., Minorpollis sp.*, the *Complexiopollis* and *Oculopollis* species are predominant.

So far, an identical association has not been found in the Hungarian Upper Cretaceous formations. Similar assemblages were determined in the Austrian Gosau layers (Gams Basin, Grabenbach Formation: Siegl-Farkas and Wagreich 1996; Weissenbachalm: Siegl-Farkas 1994; Hradecká et al. 1999). These early *Normapolles* taxa, which can be assigned to the *Complexiopollis* genera, are known

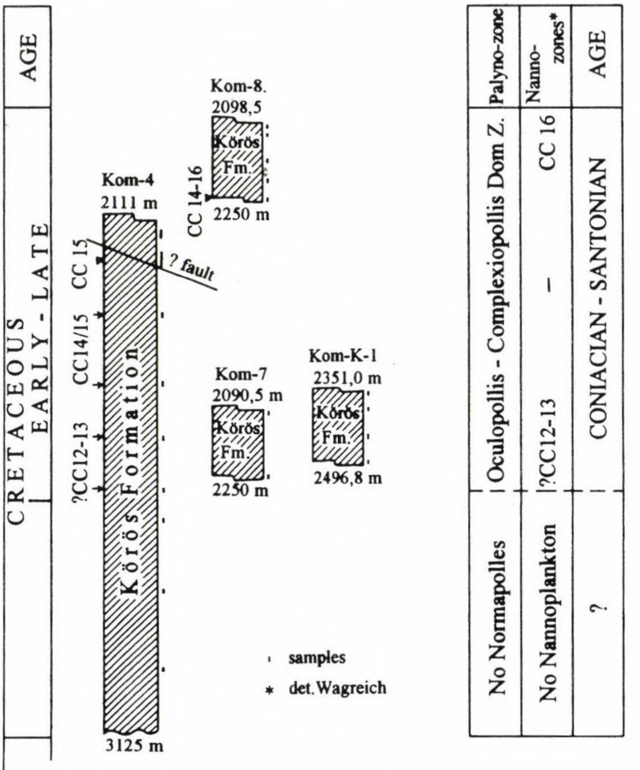

Fig. 3
Correlation of terrigeneous Körös Formation

in the Bohemian (south Czech Republic) and German (Pirna) Turonian–Coniacian formations (Krutzsch 1959; Góczán et al. 1967; Pacltova 1981). In the associations the occurrence of *Oculopollis* species is frequent, which excludes that the formations belong to the Turonian stage. The composition of the association rather indicates a Coniacian–Early Santonian age but on the basis of the occurrence of the older elements the possibility of reworking cannot be excluded. This is also hinted at by the state of preservation of a part of the sporomorphs. The *Complexiopollis*-type grains and the small-sized *Oculopollis*, as well as a part of the pteridophyte spores, show stronger oxidation (thin, broken forms) than the bigger *Oculopollis* and some pteridophyte spores in a good state of preservation. This may mean that part of the Early Senonian formations was reworked during the Late Coniacian–Early Santonian. This possibility seems to be also confirmed by the sedimentation conditions (detrital terrigenous).

In the associations of the borehole, including the assemblages of the uppermost samples as well (where the more frequent sampling was carried out), no younger *Normapolles* could be determined. This contradicts the assignment of the formation up to now to the Campanian–Maastrichtian stage (Szentgyörgyi, in Árkai et al. 1996).

The nannoplankton investigations also mostly make Coniacian-Lower Santonian (CC12–CC15) sedimentation probable, but do not exclude a Turonian possibility. Investigated samples of the depth interval 2713.0–3009.0 m contained no nannoplankton; thus the marine origin of these rocks is uncertain.

On the basis of its nannoflora the section between 2644.0–2713.0 m could be conditionally assigned to the Middle Turonian–Middle Coniacian (CC12–CC13) but the poor state of preservation of the flora may also indicate possible reworking. *Micula decussata*, determined in 2320.0–2456.0 m, indicates the Middle Coniacian (CC14/15), while *Reinhardtites anthophorus*, occurring in 2206.0–2214.0 m, the Santonian (CC15) stage (classification and zonation of Wagreich, this study).

On the basis of the analogies of the results of the nannoplankton and palynological investigations of the Gams Basin (Siegl-Farkas and Wagreich 1996) an assignment to the Upper Coniacian–Lower Santonian stages seems to be the most reasonable.

Taking into account all of the above we assign the terrigenous Körös Formation, exposed in the sequence, to the older section of the *Oculopollis–Complexiopollis* Dominance Zone (Upper Coniacian–Lower Santonian).

Note: one sample of the borehole, taken from the depth interval 2193.5–2197.5 m, does not fit into the above-outlined picture, either on the basis of its rock material or age. The sample contains a rich Paleogene marine (dinoflagellate and nannoplankton) association; no Upper Cretaceous sporomorph could be found in it. If it is not a mixed-in sample (!?), it could be indicative of post-Paleogene tectonic event(s).

Detrital terrigenous formations were also exposed by the boreholes Komádi-K-1, Kom-7 and Kom-8. The investigated formations of boreholes Komádi-K-1 (2351.0–2496.8 m) and Kom-7 (2090.5–2241.3 m) can be correlated with the lower part of the reference section. The section of borehole Kom-8 between 2098.5–2246.2 m is characterized by a very rich sporomorph association. The genus *Oculopollis* is predominant. Some species of *Hungaropollis, Pseudoplicapollis*, occurring beside the scattered *Compexiopollis* as well as the *Interporopollenites* species which appear regularly, give as the time of deposition of the formation a younger part of the Santonian stage. The rich pteridophyte spore occurrence indicates a nearshore and the determined *Spiniferites div. sp.* and *Tanyosphaeridium sp.* dinoflagellates a more frankly marine sedimentation.

Nannoplankton investigation carried out in the 2245.0–2246.2 m interval indicates the Middle Coniacian–Middle Santonian (CC14–CC16) stages (Wagreich 1992).

All the above mean that the sequence of borehole Kom-8 includes formations younger than the reference section (borehole Kom-4).

Basal breccia facies, borehole Kom-2, -10, -13 (Fig. 4)

The borehole Kom-10 exposed Upper Cretaceous formations between 2288.0–2356.0 m. Its subcrop is constituted by Paleozoic and its cover by Pliocene layers. Samples of all the three cores cut in the Senonian section of the sequence (section between 2288.0–2323.0 m) were studied.

The sporomorph associations give evidence of equally rich pteridophyte spore and angiosperm pollen occurrences. Among the pteridophyte spores *Bikolisporites toratus* is predominant. The occurrence of *Devecserisporites campanicus* in several specimens provides good correlation data for the comparison with the formations assigned to the Campanian stage of Bácska area. The possibility of correlation is also well supported by the *Interporopollenites sahi, Plicapollis serta, Subtriporopollenites* and *Semioculopollis* species. Among the *Normapolles* genera *Oculopollis* and *Trudopollis* are predominant. A marine environment is indicated by the remains of *Dinogymnium westralium* and organic foraminifera shells.

On the basis of the above-listed facts the investigated formations were assigned to the Middle Campanian *Interporopollenites sahi Subzone.* Palynologically an assignment to the Maastrichtian stage is not reasonable (Szentgyörgyi, in Árkai et al. 1996).

Formations deposited on basal breccia were also exposed by the boreholes Kom-2 and Kom-13. Sporomorph associations of the Kom-2 borehole (2386.0–2518.0 m) provide a good correlational possibility for the comparison with the Senonian formations of Bácska area. Different coalification stages of the sporomorphs indicate reworking of older Senonian formations. Age of sedimentation is given by the occurrence of the less coalified *Hungaropollis, Krutzschipollis* and *Isabelidinium microarmum,* according to which the investigated section may be correlated with the section between 741.7–793.0 m of the Csikéria Marl Formation of borehole Ba-1 of Bácska region (Fig. 4).

The occurrence of the primitive members of the *Complexiopollis* group, determined in samples between 2557.2–2698.5 m of the Senonian section (2437.0–2690.0 m) of borehole Kom-13, calls attention to the reworking of the Coniacian–Early Santonian formations in the Middle Campanian stage. The assumption of Campanian sedimentation is justified by the occurrence of the *Tarsisphaeridium geminiporatum, Cyclopsiella elliptica* dinoflagellates, the *Devecserisporites campanicus, Taurocusporites sp.* pteridophyte spores, as well as the *Interporopollenites div. sp., Minorpollis sp., Semioculopollis minimus* and *Vacuopollis sp. Normapolles* species.

According to the dinoflagellate zonation the formations can be assigned to the *geminiporatum* Subzone and be correlated with the middle section of the Bácsalmás Formation (Fig. 4).

Szentgyörgyi (in Árkai et al. 1996) also called attention to the petrologic and faunistic similarities of the Körös Formation and the Csikéria Marl, also developed in the surroundings of Komádi.

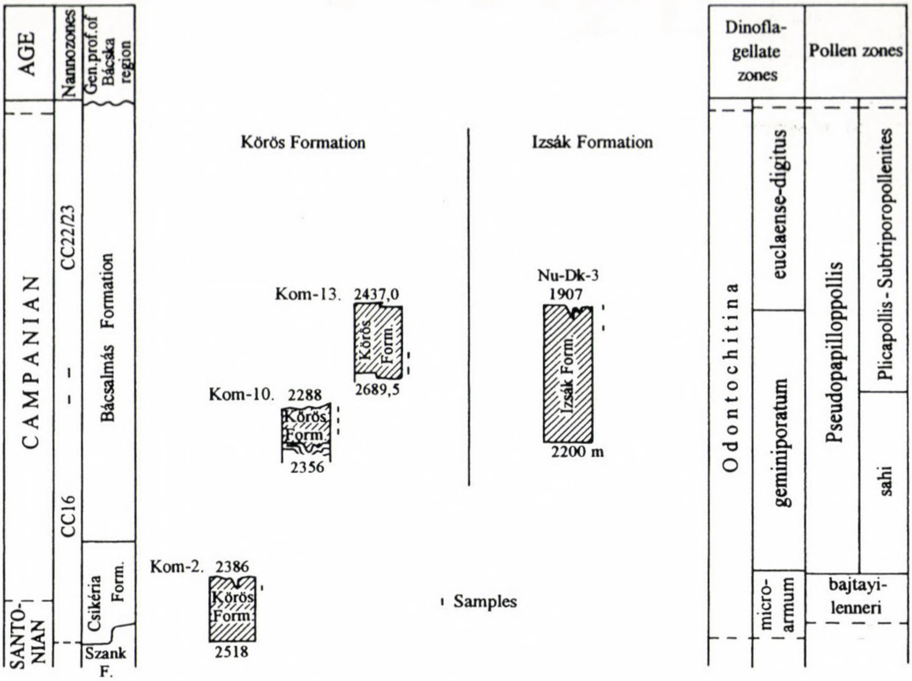

Fig. 4
Correlation of Kőrös- (on basal breccia) and Izsák Formation

According to the palynological investigations the Körös Formation of basal breccia facies shows greater similarity to the Bácska sequences of similarly basal breccia facies (Szank Formation) than to the detrital, terrigenous facies of the Körös Formation.

The palynological investigations indicate a significant lacuna between the two formations, indicating that they belong to two separate cycles (Fig. 5).

Mecsek Zone, Izsák Formation (Szentgyörgyi, in Árkai et al. 1996)

Investigated samples from the type section of the Izsák Formation (borehole Izsák-1, 695.0–819.5 m), developed in the Great Hungarian Plain area of the Mecsek Zone, proved to be barren. Samples of borehole Nu-Dk-3 of Nádudvar (1909.0–1962.0 m), accepted as the reference section of the formation, can be characterized by a very scattered sporomorph association in a bad state of preservation. The determined *Odontochitina operculata, Dinogymnium sphaerocephalum, Biconidinium reductum, Siniferites div. sp., and Veryhachium sp.* indicate a marine environment. On the basis of the occurrence of *Biconidinium reductum* the investigated section of the borehole within the *Odontochitina* Assemblage Zone can be correlated with the middle section of the borehole

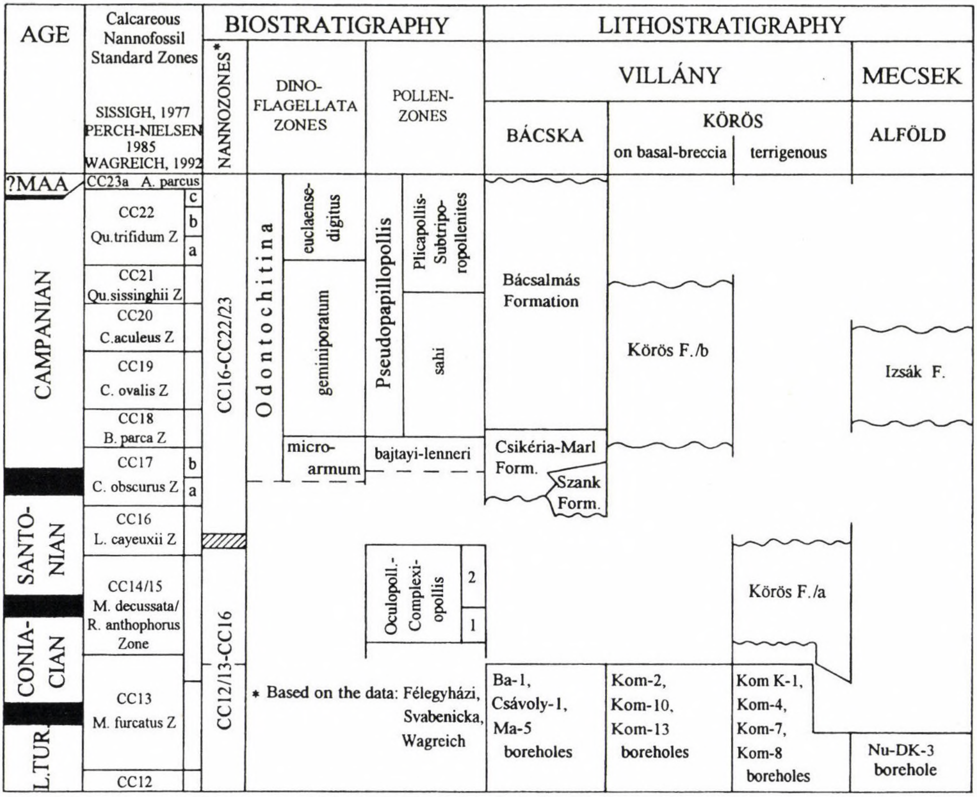

Fig. 5
Biozonation of the Upper Cretaceous formations in the Great Hungarian Plain

Csávoly-1 of the Bácska facies (Bácsalmás Formation, 1228.0–1487.4 m) in all probability (Figs 2, 4).

The presence of organic substance in great quantities assumes strong areal erosion and relief energy as well as strong terrigenous transportation as well. Among the pollen grains of the angiosperms only the occurrence of some species of the *Oculopollis* and *Minorpollis* genera does not contradict a Late Campanian age, which can be designated on the basis of the dinoflagellates. Assignment to the Maastrichtian stage (Szentgyörgyi, in Árkai et al. 1996) is not reasonable on the basis of the present palynological investigations.

Paleoenvironmental relations

Senonian sequences of the Pelso- (Transdanubian Central Range) and the Tisza Unit (Mecsek–Villány Zone) – still situated far from each other during the Lower Cretaceous (Haas and Hámor 1996) – were deposited in sub-basins of Tethys in the course of the Upper Cretaceous transgression. Sometimes, and in some places, they are evidence of similar (Bácska–Bakony), and in other places of different (environs of the Körös River), sedimentary conditions. By means of palynology the Upper Cretaceous formations, of different facies but contemporaneous, can be well correlated with each other. The two areas belonged to the same floral province as proved by common flora elements. Palynologically a greater difference between the two areas appears in the time range of biozones, and their correlation and facies, respectively.

Formations of similar origins as the sequences indicate that during the Senonian, on the northern margin of Tethys, very similar sedimentary environments could be formed, even at great distances from each other.

According to the palynological data the transgression in the Great Hungarian Plain areas of the Tisza Unit can be dated earlier than in the Transdanubian Central Range. While in the Villány Zone marine sedimentation already took place in the TCR paludal limnic layers of the Ajka Coal Formation were deposited. Its characteristic sporomorph association cannot be found in the Tisza Unit at all.

The youngest formations of the Csávoly sequence of the Villány Zone are of nearly the same age as the youngest formations of the TCR near the Campanian–Maastrichtian boundary. This means that in the areas of both tectonic units (Tisza and Pelso) the Senonian formations which are younger than this are equally eroded due to later geologic events.

Both above-mentioned sub-basins of Tethys were situated in the longshore / nearshore region of the area of the Normapolles Phytogeographical Province until the end of the Late Campanian and the beginning of the Maastrichtian stage, respectively. A greater difference between the assemblages of the two areas appears in the dinoflagellate association of the end of the Campanian. While at the end of the Campanian, in the Villány Zone there was still a shallow marine environment, in the area of the TCR already a deeper water, open marine environment of oceanic type was formed. This is proved by the fact that among the largely similar dinoflagellate assemblages the association of the *Pyxidinopsis bakonyensis* Assemblage Zone (Siegl-Farkas 1997) can be shown only in the territory of the TCR. At the same time, this assemblage was also determined from the association of the section designated as the stratotype of the Campanian–Maastrichtian boundary of southern France (Tercis) (Odin et al., in press).

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Plates

Magnification: all figure are X 1000

Plate I

1–2. *Vadaszisporites pseudofoveolatus* (Deák) Deák & Comb., Komádi-13 bh., 3. core. 2557.2–2557.8 m.
3–4. *Taurocusporites maastrichticus* W. Kr., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
5–6. *Cicatricosisporites hughesi* Dett., Madaras-5 bh., 22. core, 492.5–495.5 m.
7–8. *Impardecispora trioreticulosa* (Cooks. & Dett.) Venkat., Kar, Raza, Madaras-5 bh., 22. core, 492.5–495.5 m.
9. *Polypodiaceoisporites* cf. *brejani* Cernj., Csávoly-2 bh., 12. core, 1018.5-1023.0 m.
10. *Cicatricosisporites* sp., Komádi 13 bh., 3. core, 2557.2–2557.8 m.
11. *Echinatisporites* sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
12. *Klukisporites* cf. *tuberosus* Döring, Komádi-13 bh., 3. core, 2557.2–2557.8 m.
13. *Stereisporites psilatus* (Ross) Pf., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
14. *Devecserisporites goczani* Siegl-Farkas, Komádi-10 bh., 3. core, 2314.0–2318.0 m.
15. *Taurocusporites* sp., Komádi-2 bh., 2400.0–2416.0 m.

Plate II

1–5. *Plicatopollis plicatus* (Pf.) W, Kr., 1–2: Csávoly-1 bh., 12. core, 1018.5–1023.0 m, 3: Madaras-5 bh., 18. core, 473.2–474.4 m, 4: Ba-1 bh., 531.3 m, 5: Ba-1 bh., 622.4–623.5 m
6. *Triatriopollenites* sp., Csávoly-1 bh., 1018.5–1023.0 m.
7. *Plicapollis excellens* (Pf.) W, Kr., Madaras-5 bh., 22. core, 492.6–495.5 m.
8. *Pseudoplicapollis* sp., Komádi-13 bh., 2557.2–2557.8 m.
9. *Plicapollis excellens* (Pf.) W, Kr., Csávoly 1 bh., 12. core, 1018.5–1023.0 m.
10. *Plicapollis pseudoexcelsus* Greif., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
11. *Triatriopollenites lubomirovae* (Gladk.) Kds., Csávoly-1 bh., 12. core, 1018.8–1023.0 m.
12. *Plicapollis silicatus* Pf., Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
13. *Triatriopollenites lubomirovae* (Gladk.) Kds., Ba-1 bh., 542.6 m.
14. *Triatriopollenites pseudogranulatus* (Gladk.) Kds., Ba-1 bh., 542.6 m.
15–16. *Plicapollis* sp., Madaras-5 bh., 25. core, 555.7–555.8 m.
17. *Plicapollis conserta* Pf., Ba-1 bh., 555.7–555.8 m.
18–21. *Pseudoplicapollis praesubhercynicus* Góczán, 18–19: Madaras-5 bh., 30. core, 564.5–565.2 m, 20: Madaras-5 bh., 22. core, 492.6–495.5 m, 21: Madaras-5 bh., 25. core, 520.1–521.0 m.
22. *Magnoporopollis krutzschi* Kds, & Herngreen, Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
23–24. *Tetrapollis validus* (Pf.) Pf., Csávoly-1 bh., 18. core, 1271.3–1271.5 m.
25. *Interporopollenites endotriangulatus* Heg., Kds, & Párduc, Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
26. cf. *Papillopollis csavolyensis* Kds., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
27–28. *Papillopollis csavolyensis* Kds., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
29–32. *Interporopollenites sahi* Góczán, 29: Madaras-5 bh., 22. core, 492.6–495.5 m, 30: Madaras-5 bh., 18. core, 473.2–474.4 m, 31: Ba-1 bh., 613.8–614.8 m, 32: Madaras-5 bh., 520.0–532.0 m.
33. *Interporopollenites tenuis* W.Kr., Ba-1 bh., 613.8–614.8 m.
34. *Interporopollenites tenuis* W.Kr., Ba-1 bh., 670.9 m.
35–36. *Interporopollenites sahi* Góczán, Madaras-5 bh., 22. core, 492.6–495.5 m.
37. *Interporopollenites csavolyensis* Kds., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
38. *Interporopollenites giganteus* Góczán, Madaras-5 bh., 22. core, 492.6–495.5 m.
39. *Interporopollenites* sp., Madaras-5 bh., 22. core, 492.6–495.5 m.
40. *Interporopollenites zaklinskaiae* Kds, & Hegedüs, Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

Plate III

1. *Oculopollis* sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
2. *Oculopollis orbicularis* Góczán, Madaras-5 bh., 22. core, 492.6–495.5 m.

3–4. *Trudopollis* sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

5–6. *Trudopollis imperfectus* Pf., Ba-1 bh., 703.9–704.8 m.

7. *Longanulipollis elegans* Góczán, Madars-5 bh., 30. core, 564.5–565.2 m.

8. *Hungaropollis* sp., Madaras-5 bh., 30. core, 564.5–565.2 m.

9. cf. *Nudopollis* sp., Csávoly-1 bh., 1018.5–1023.0 m.

10–11. *Trudopollis imperfectus* Pf., Madaras-5 bh., 22. core, 492.6–495.5 m.

12. *Subtriporopollenites* sp., Madaras-5 bh., 22. core, 492.6–495.5 m.

13–14. *Hungaropollis nodosus* Góczán & Siegl-Farkas, 13: Csávoly-1 bh., 12. core, 1018.5–1023.0 m, 14: Madaras-5 bh., 30. core, 564.5–565.2 m.

15–16. cf. *Romeinipollis hungaricus* Kds., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

17. *Subtriporopollenites constans* Pf., Csávoly-1 bh., 18. core, 1270.0–1273.0 m.

18. *Endoinfundibulapollis distinctus* Tschudy, Csávoly-1 bh., 18. core, 1270.0–1273.0 m.

19–20. *Suemegipollis triangularis* Góczán, Csávoly-1 bh., 18. core, 1270.0–1273.0 m.

21–22. *Rugutriporites balinkaense* (Kds) Kds., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

23. *Subtriporopollenites* sp., Madaras-5 bh., 25. core, 520.0–521.0 m.

24. *Subtriporopollenites constans* Pf, ssp, minor Kds., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

25–26. *Suemegipollis germanicus* W, Kr., Madaras-5 bh., 22. core, 492.5–495.5 m.

27. *Pseudotrudopollis crassiexinus* W, Kr., Csávoly-1 bh., 17. core, 1228.0–1232.0 m.

28. *Oacolpopollenites* sp., Csávoly-1 bh., 17. core, 1228.0–1232.0 m.

29–30. *Semioculopollis minimus* Góczán, Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

31. cf. *Momipites tenuipolus* Andersen, Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

32. cf. *Complexiopollis* sp., Madaras-5 bh., 22. core, 492.6–495.5 m.

33–34. *Complexiopollis csavolyensis* Kds., Csávoly-1 bh., 19. core, 1310.5–1312.5 m.

35. *Boltenhagenipollis magnoporatus* Kds, & Diniz, Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

36. *Gemmamonocolpites* sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

37. *Monocolpopollenites* sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

38. Ilexpollenites sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

Plate IV

1. *Hungaropollis krutzschi* Góczán, Komádi-2 bh., 7. core, 2400.0–2416.0 m.

2–3. *Oculopollis* sp., Komádi 13 bh., 3, core, 2557.2–2557.8 m.

4. *Oculopollis* cf. *orbicularis* Góczán, Komádi-8 bh., 11. core, 2098.5–2098.9 m.

5–7. *Interporopollenites* sp., 5: Komádi-8 bh., 15. core, 2245.0–2246.2 m, 6: Komádi-4 bh., 19. core, 2202.0–2206.0 m, 7: Komádi-8 bh., 15. core, 2245.0–2246.2 m.

8. *Oculopollis parvoculus* Góczán, Komádi-13 bh., 3. core, 2557.2–2557.8 m.

9. cf. *Pompeckjoidaepollenites* sp., Komádi-8 bh., 10. core, 2043.15–2043.4 m.

10. *Interporopollenites* cf. *endotriangulus* Kds, & Párducz, Komádi-8 bh., 15. core, 2245.0–2246.2 m.

11. *Interporopollenites* sp., Komádi-8 bh., 11. core, 2098.5–2098.9 m.

12. *Pseudoplicapollis peneserta* Pf., Komádi-8 bh., 13. core, 2154.6–2155.0 m.

13. *Trudopollis hemiperfectus* Pf., Komádi-13 bh., 13. core, 2557.2–2557.8 m.

14. *Semioculopollis minimus* Góczán, Komádi-4 bh., 14. core, 2152.0–2154.0 m.

15–16. *Pseudoplicapollis peneserta* (Pf.) W, Kr., Komádi-4 bh., 22. core, 2438.0–2456.0 m.

17–18. *Conclavipollis burgeri* Van Amerom, Komádi-4 bh., 20. core, 2206.0–2214.0 m.

19. *Complexiopollis* sp., Komádi-4 bh., 14. core, 2152.0–2154.0 m.

20. *Atlantopollis reticulatus* W.Kr., Komádi-7 bh., 3. core, 2090.5–2091.0 m.

21. *Retitricolpites* sp., Komádi-4 bh., 14. core, 2152.0–2154.0 m.

22–23. cf. *Conclavipollis* sp., Komádi-4 bh., 14. core, 2152.0–2154.0 m.

24–25. *Complexiopollis* sp., Komádi-4 bh., 19. core, 2202.0–2206.0 m.

26. *Complexiopollis* sp., Komádi-4 bh., 18. core, 2197.0–2202.0 m.

27–28. cf. *Atlantopollis* sp., Komádi-4 bh., 25. core, 2713.0–2719.0 m.

29. *Complexiopollis* sp., Komádi-13, 3. core, 2557.2–2557.8 m.
30–33. *Complexiopollis vulgaris* (Groot & Groot) Groot & W, Kr., 30–31: Komádi-7 bh., 3. core, 2090.5–2091.0 m, 32–33: Komádi-4 bh., 15. core, 2183.0–2188.5 m.
34–35. *Complexiopollis labilis* (Góczán) Góczán & W, Kr., Komádi-13 bh., 3. core, 2557.2–2557.6 m.
36. *Complexiopollis* sp., Komádi-4 bh., 23. core, 2542.0–2549.0 m.
37. *Complexiopollis* cf. *helmigii* (Van Amerom) Sole de Porta, Komádi-4 bh., 18. core, 2197.0–2202.0 m.
38. *Complexiopollis vulgaris* (Groot & Groot) Groot & W, Kr., Komádi-7 bh., 3. core, 2090.5–2091.0 m.
39. *Minorpollis* sp., Komádi-4 bh., 25. core, 2713.0–2719.0 m.
40–41. cf. *Complexiopollis* sp., Komádi-4 bh., 19. core, 2202.0–2206.0 m.
42–43. *Complexiopollis* sp., Komádi-4 bh., 19. core, 2202.5–2206.0 m.

Plate V

1. *Odontochitina operculata* (O, Wetzel) Defl, & Cookson, Ba-1 bh., 760.9 m.
2. *Alisogymnium euclaense* (Cooks, & Eis.) Lentin & Vozzh., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
3–4. *Odontochitinopsis molesta* (Defl.) Eisenack, 3: Ba-1 bh., 655.8 m, 4: Ba-1 bh., 634.4 m.
5. *Alisogymnium sphaerocephalum* (Vozzh.) Lentin & Vozzh., Nádudvar-DK-3 bh., 5. core, 1909.0–1914.0 m.
6. *Dinogymnium microgranulosum* Clarke & Verdier, Ba-1 bh., 791.7 m.
7. *Dinogymnium denticulatum* (Alberti) Evitt, Clarke & Verdier, Ba-1 bh., 755.3 m.
8. *Dinogymnium digitus* (Defl.) Evitt et al., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
9. *Alisogymnium euclaense* (Cooks, & Eis.) Lentin & Vozzh., Csávoly-1 bh., 1018.50–1023.0 m.
10. *Dinogymnium* cf. *acuminatum* Evitt et al., Ba-1 bh., 752.5–753.4 m.
11. *Alisogymnium* cf. *euclaense* (Cooks, & Eis.) Lentin & Vozzh., Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
12. *Dinogymnium cretaceum* (Defl.) Evitt et al., Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
13. *Alisogymnium assamicum* (Jain et al.) Lentin & Vozzh., Csávoly-1 bh., 17. core, 1228.0–1232.0 m.

Plate VI

1. *Isabelidinium microarmum* (McIntyre) Lentin & Williams, Ba-1 bh., 782.1 m.
2. *Nelsoniella tuberculata* Cookson & Eis., Madaras-5 bh., 25. core, 520.0-521.0 m.
3. *Isabelidinium microarmum* (McIntyre) Lentin & Williams, Ba-1 bh., 788.0 m.
4. *Tarsisphaeridium geminiporatum* Riegel, Madaras-5 bh., 22. core, 492.6–495.5 m.
5. *Isabelidinium microarmum* (McIntyre) Lentin & Williams, Komádi-2 bh., 7. core, 2400.0–2416.0 m.
6. Scolecodonta (Annelidae), Ba-1 bh., 562.0 m.
7. *Tarsipshaeridium geminiporatum* Riegel, Ba-1 bh., 627.0–628.0 m.

Plate VII

1. *Apteodinium ?cribrosum* Cookson & Eis., Ba-1 bh., 779.5 m.
2. cf. *Gonyaulax cassidata* Eis, & Cookson, Ba-1 bh., 530.2 m.
3. *Chatangiella victoriensis* (Cooks, & Manum) Lentin & Will., Csávoly-1 bh., 19. core, 1310.5–1312.5 m.
4. cf. *Deflandrea phosphoritica* Eisenack, Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
5–6. cf. *Biconidinium reductum* (May) Kirsch, Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
7. cf. *Membranilarnacia polycladita* Eis., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.
8. Scolecodonta (Annelidae), Ba-1 bh., 643.0 m.
9. cf. *Trithyrodinium* sp., Ba-1 bh., 791.7 m.

Plate VIII

1. cf. *Xenascus sarjeanti* (Corradini) Stover & Evitt, Csávoly-1 bh., 19. core, 1310.5–1312.5 m.
2–3. *Gillinia pyriformis* Marshall, Csávoly-1 bh., 17. core, 1228.0–1232.0 m.
4. *Gonyaulax margaritifera* Cooks, & Eis., Ba-1 bh., 530.2 m.
5–8. *Gillinia hymenophora* Cooks, & Eis., Madaras-5 bh., 18. core, 473.2–474.4 m.
9–10. cf. *Trichodinium castanea* (Defl.) Clarke & Verdier, Csávoly-1 bh., 19. core, 1310.0–1312.5 m.
11. cf. *Chytroeisphaeridia* sp., Csávoly-1 bh., 12. core, 1018.5–1023.0 m.

Plate IX

1. *Acritarcha* sp., Ba-1 bh., 770.2–771.2 m.
2. *Alterbidinium varium* Kirsch, Ba-1 bh., 709.3 m.
3. *Acritarcha* sp., Ba-1 bh., 766.7 m.
4. *Acritarcha* sp., Ba-1 bh., 791.7 m.
5. *Micrhystridium stellatum* Defl., Ba-1 bh., 768.3 m.
6. *Spiniferites bulloideus* (Defl, & Cooks.) Sarjeant, Ba-1 bh., 731.0 m.
7. *Spiniferites dentatus* Gocht, Ba-1 bh., 738.0 m.
8. cf. *Palaeohystrichophora infusorioides* Defl., Ba-1 bh., 647.6–648.5 m.
9. *Biconidinium* sp., Nádudvar-DK-3 bh., 6. core, 1955.0–1962.0 m.
10. cf. *Pervosphaeridium truncigerum* (Defl.) Yun, Ba-1 bh., 792.3 m.
11. cf. *Pterodinium cingulatum* (O, Wetzel) Below, Madaras-5, 22. core, 492.6–495.5 m.

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Plate I

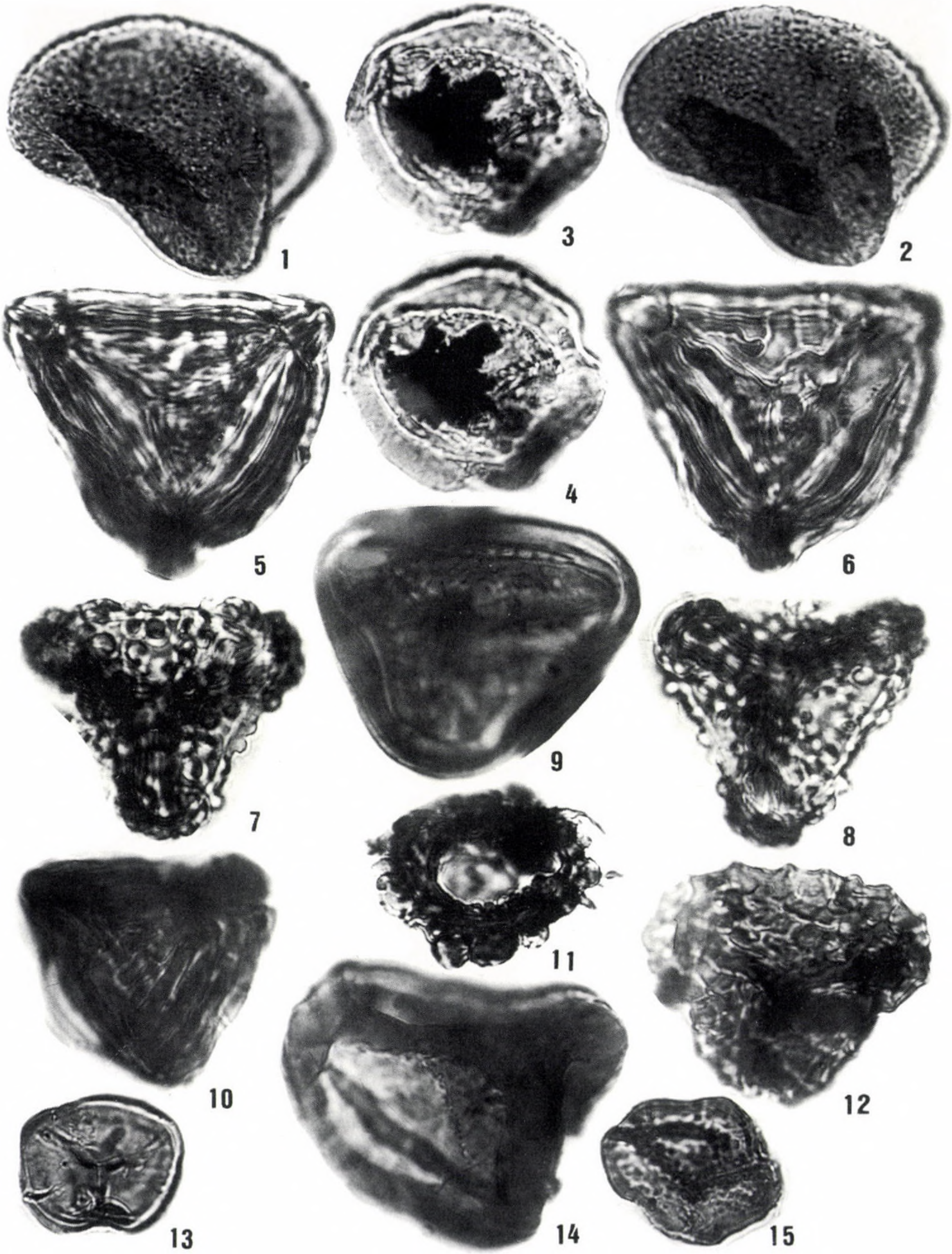

Plate II

Plate III

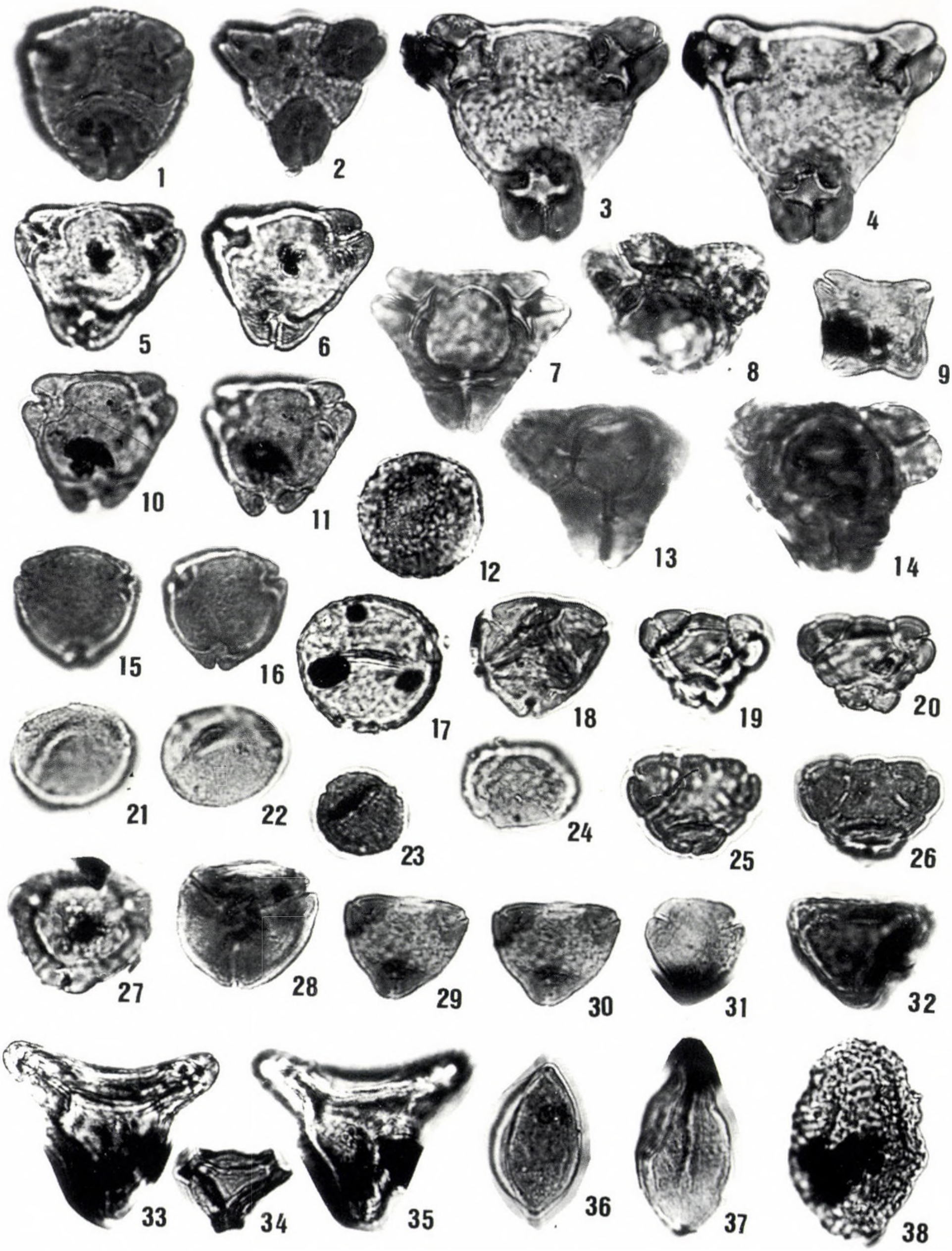
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Plate IV

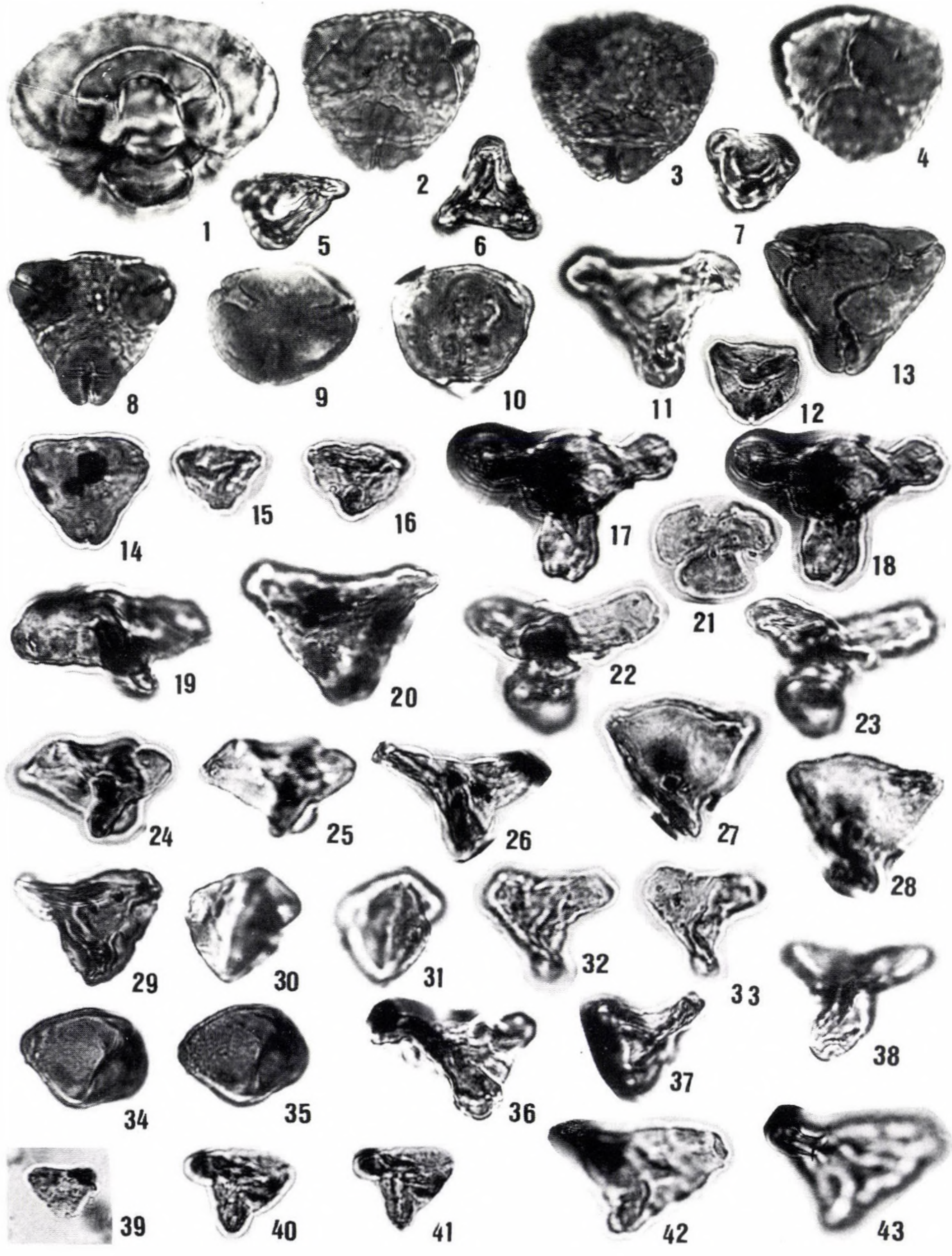
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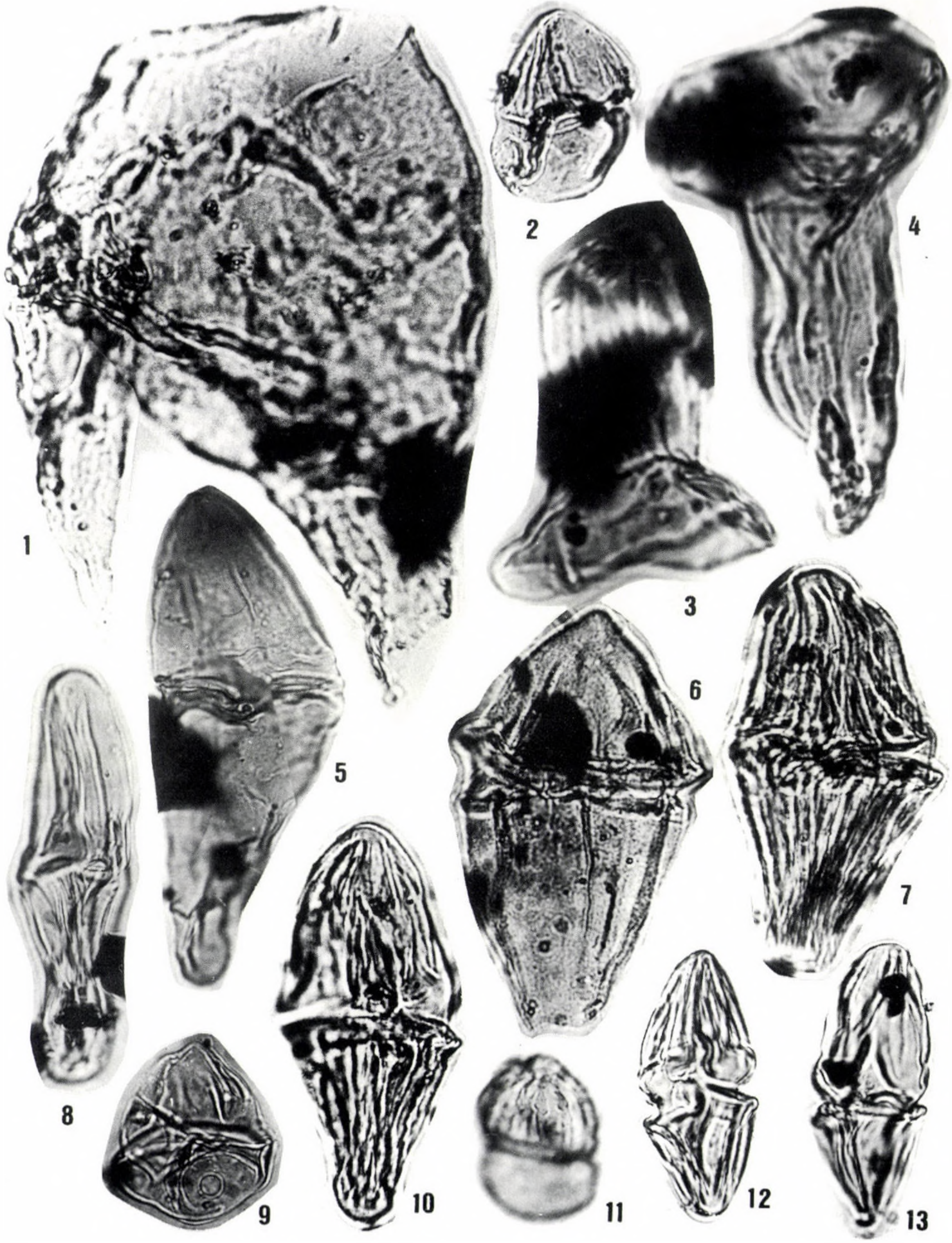
Plate V



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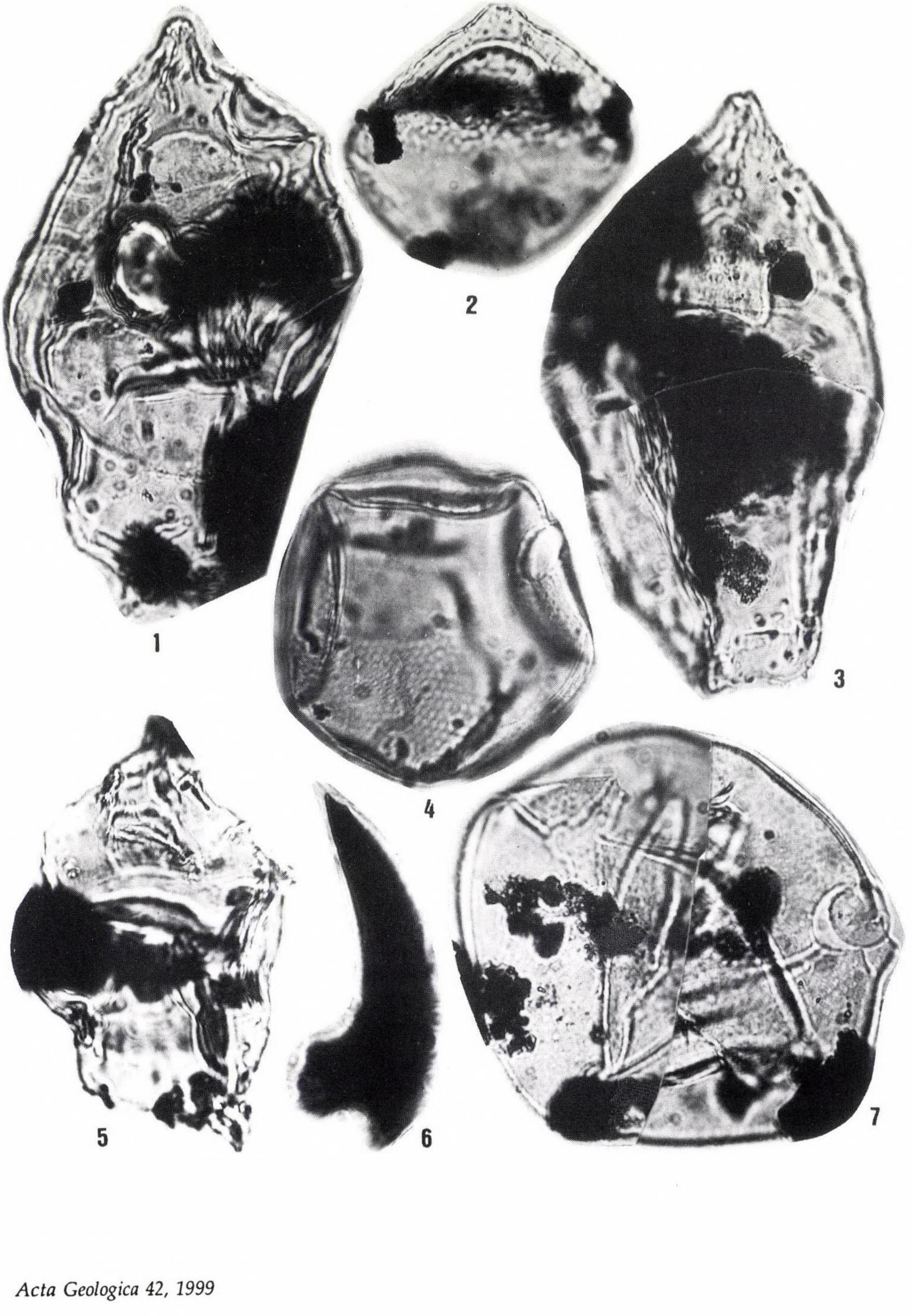

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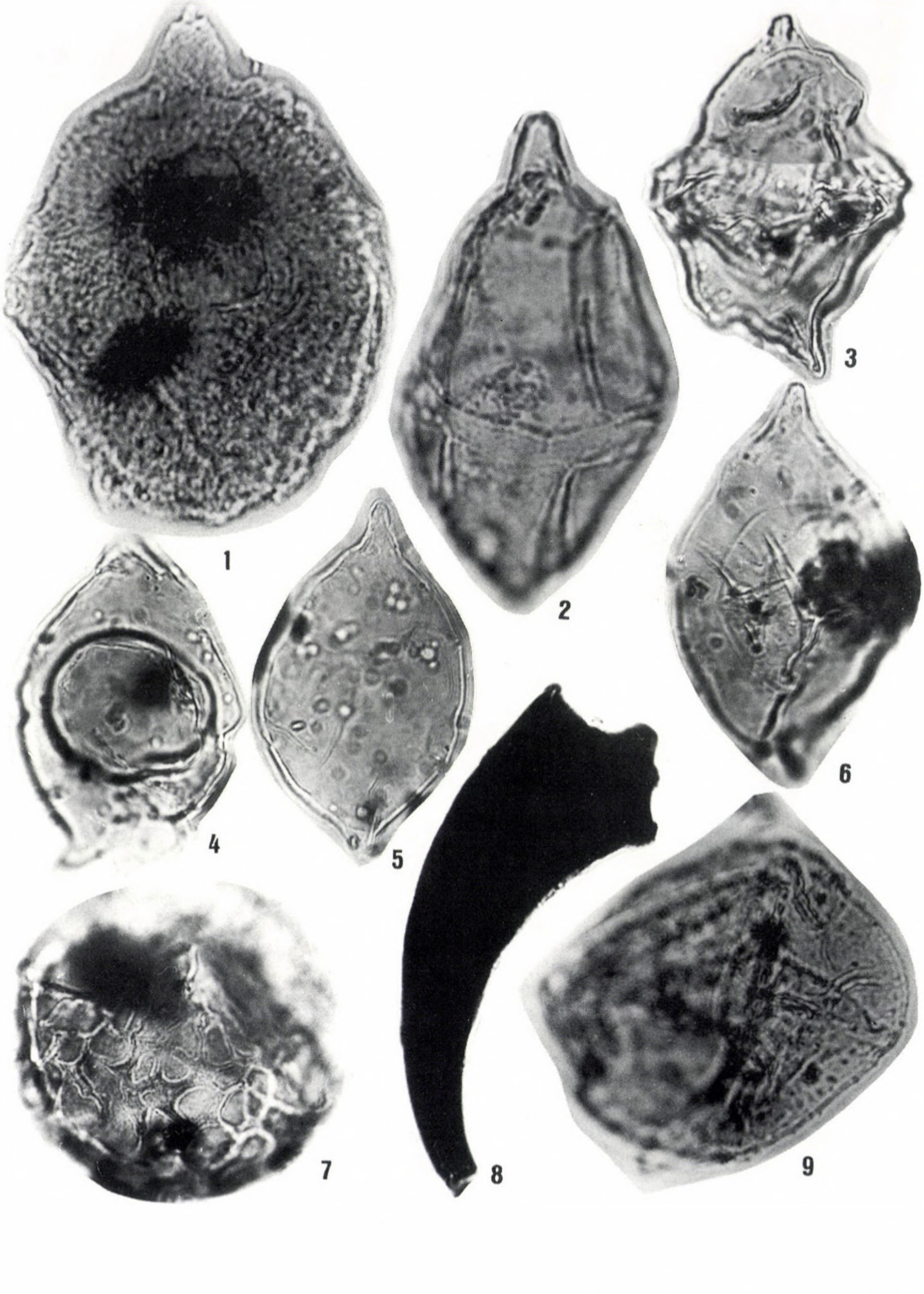

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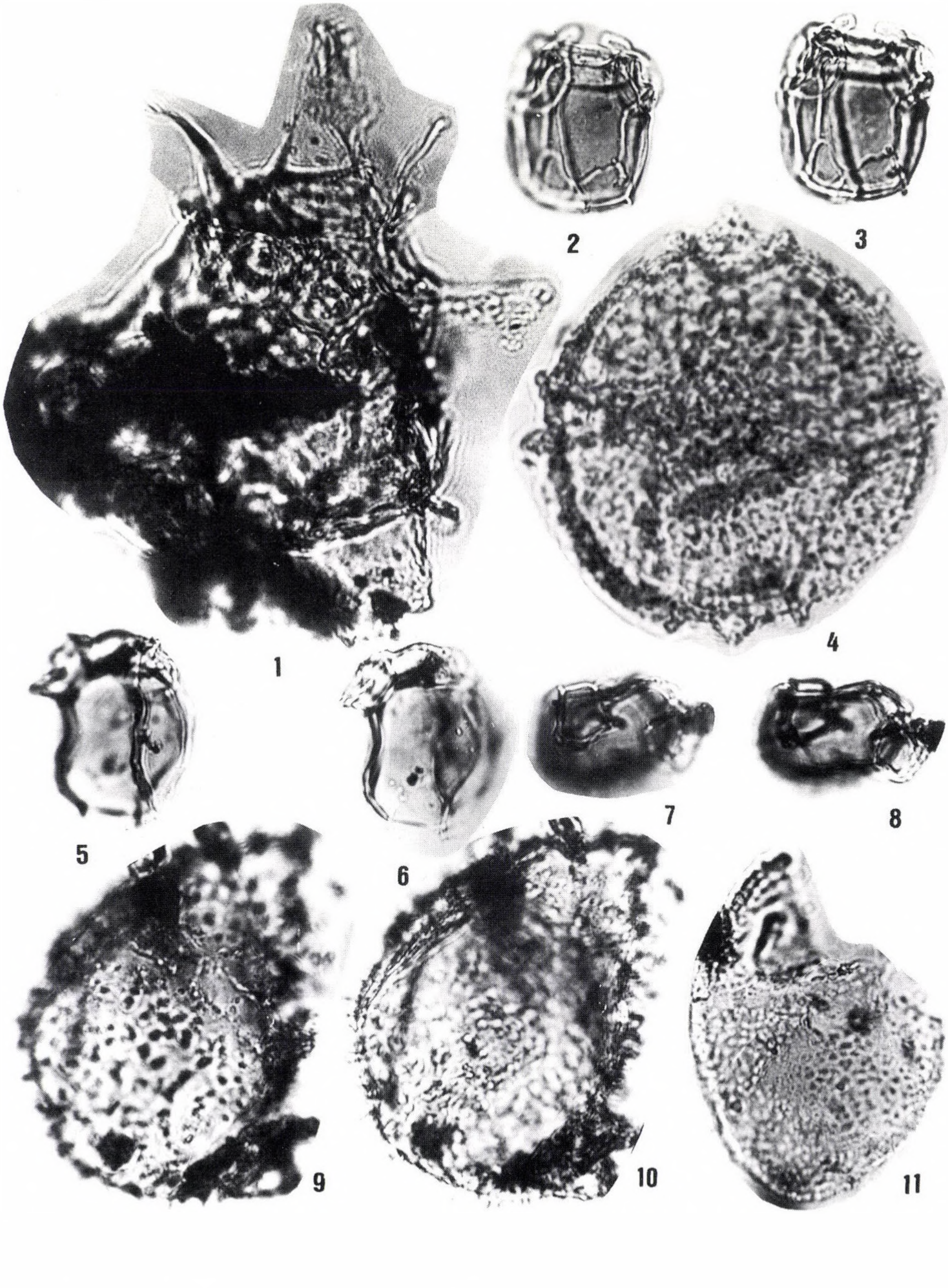

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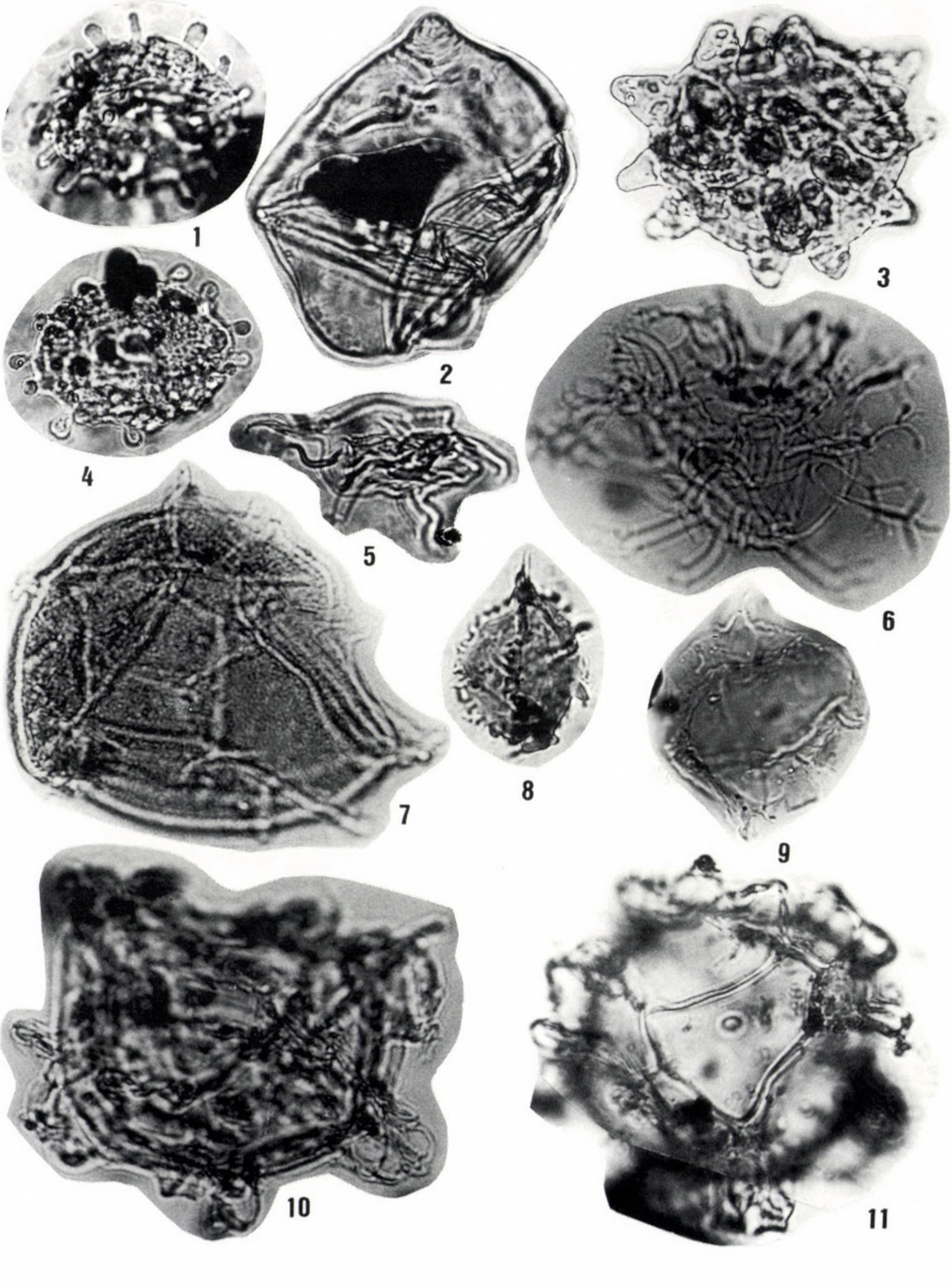

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Taxonomic notes on benthic foraminifera from SW-Hungary, Middle Miocene (Badenian) Paratethys

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The thirty-eight most common (5%) benthic foraminifera species from two boreholes from the Middle Miocene (Badenian – 16.6 Ma–12.9 Ma) of the Pannonian Basin Paratethys (Hungary) are illustrated and documented. In accordance with the applied species concept an effort has been made to compile synonymous names used in the Paratethys, and to record intraspecific morphologic variability. In general we believe that most of the variation is connected to environmental changes. In the case of *B. elongata*, the ecophenotypic nature of the intraspecific variability has been demonstrated on the basis of its correlation with depth, where depth has been reconstructed from the planktic/benthic foraminiferal ratio. This correlation questions the excessive use of *Bulimina* stratigraphy in the Paratethys, as it suggests that time horizons follow the major transgressive and regressive cycles of the region.

Key words: Paratethys, Badenian, Middle Miocene, benthic, foraminifera, taxonomy, intraspecific variability, palaeoecology, palaeobathymetry, Bulimina

Introduction

Previous work

Foraminiferal research concerning the Middle Miocene has a long tradition in the Central Paratethys. The great pioneers like Fichtel and Moll (1798) and especially d'Orbigny (1846) studied material from many different areas of the Central Paratethys. D'Orbigny's monograph on the Vienna Basin provided a firm foundation for later work. Recent revisions of these classical works were published in the mid-eighties (Papp and Schmid 1985; Rögl and Hansen 1984). Furthermore, numerous publications on the fauna of the Vienna Basin (Grill 1941; Marks 1951; Bachman et al. 1963; Rupp 1986), the Carpathians and its foredeep (Cicha and Zapletalova 1960, 1963, 1965; Cicha and Ctyroka 1988; Sutovska et al. 1993; Grigorovich et al. 1995), and the Transylvanian Basin (Popescu 1987; Popescu and Brotea 1994), are available among others.

The Middle Miocene in Hungary is a rather well-studied period. Many publications have been written at the Geological Institute documenting the research carried out on material from full core boreholes (Halmai et al. 1982).

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The Miocene foraminifera fauna was studied by I. Korecz-Laky, whose results were published in many subsequent publications of the Institute (Korecz-Laky 1969, 1982, 1987; Korecz-Laky and Nagy-Gellai 1977, 1985). However, these works present the described taxa mostly without scanning photos, except for Korecz-Laky and Gellai (1985), and without the demonstration of the morphologic variability doubtlessly present in the material.

Species concept

The species concept adopted in this study differs from the one used by most previous workers dealing with Paratethys material. We believe that extreme splitting of existing foraminiferal taxa, especially by describing new species, is not a proper way of conducting taxonomy, as it leads to a chaos of names, preventing the subsequent step of interpreting the fauna (Boltovskoy 1965; Zachariasse 1975; van Morkhoven et al. 1986, and many others). In practice this implies that often the full scale of morphologic variability present in the material is considered without introducing new species names.

Morphologic variability was observed at all taxonomic levels below the genus level. Intergrading species complexes with a full series of transitional forms were encountered many times. In such cases, a line had to be drawn to separate the two taxa. However, no biometric measurements were made to support it. Variability interpreted below the species level was arranged by morphotypes. The recognized morphotypes are treated as "formae". The ecophenotypic nature of such intraspecific variability of foraminiferal species has been demonstrated by Poag (1978), Feyling-Hansen (1972), Miller et al. (1982), Jorissen (1988) and Spencer (1996), among others.

This revision aims to give a full taxonomic account of the foraminiferal fauna by:

1. Compiling a critical synonymy list reflecting the species concept used;
2. Documenting the intraspecific morphologic variability present in the material, by showing the transitional forms, and end members of series;
3. Interpreting the morphologic variability by demonstrating the ecophenotypic nature of the variability, if possible;
4. Illustrating the fauna by scanning photos, which were rarely published from the Badenian of Hungary.

Geologic setting

The two studied cores were cut in SW Hungary, NE of the Mecsek Mts. The Kapos Lineament, a quite prominent tectonic feature of the region, separates the two drilling sites (Némedi-Varga 1986). Tekeres-1 is located somewhat south of the fault, and Tengelic-2 is just to the north of it (Fig. 1).

The sampled sediments were deposited in the Central Paratethys, an enclosed sea of Middle Miocene (Badenian) times. The Badenian is the last fully marine

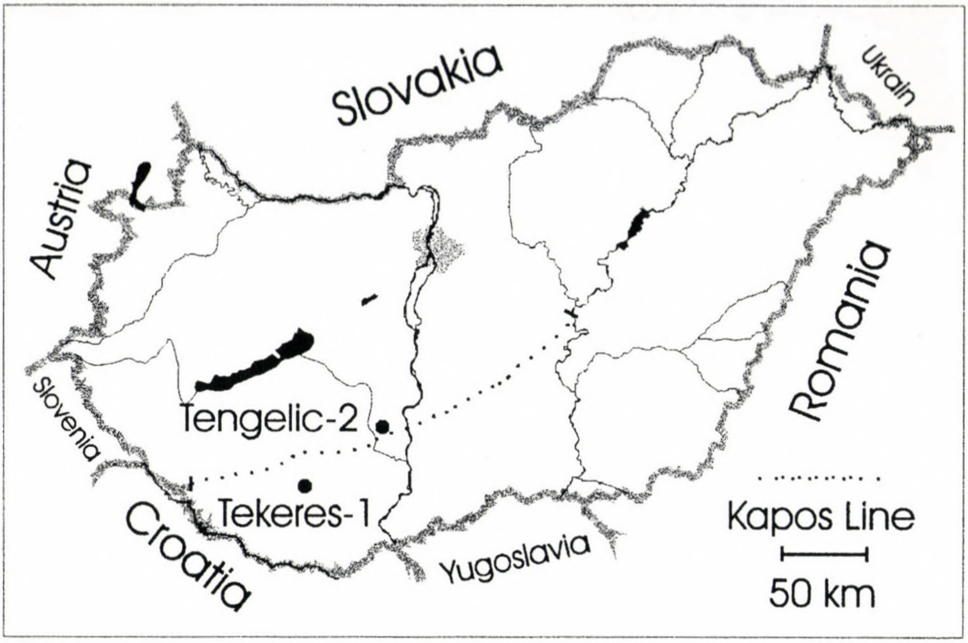

Fig. 1
Location of boreholes Tekeres-1 and Tengelic-2 in Hungary

stage in the history of the basin's development. The Central Paratethys had connections to the Mediterranean in Early Badenian; thus, it is not surprising to find the fauna similar to the Mediterranean (Rögl et al. 1978; Rögl and Steininger 1983; Steininger and Rögl 1984; Báldi 1997).

Material and methods

The material came from two boreholes, Tekeres-1 and Tengelic-2 from SW Hungary, at the foot of the Mecsek Mts (Fig. 1). Dating of the cores is based primarily on nannoplankton stratigraphy (Nagymarosy 1982, 1985). It is believed that the studied material from the two cores covers the entire Badenian (Fig. 2).

The Badenian of Tekeres-1 starts at a depth of 272 m of the core at the Karpatian–Badenian boundary. The NN5–NN6 boundary is reached at 56m. The Sarmatian has been recorded from 28 m onward at the top of the core. The marine sedimentation in the Badenian of Tengelic-2 begins above an erosional surface at 863 m overlying tuff layers; nannofossils date this level as NN5 (Nagymarosy 1982). Samples bearing foraminifera in analytically sufficient numbers are present from 272–47 m in the case of Tekeres-1, and in Tengelic-2

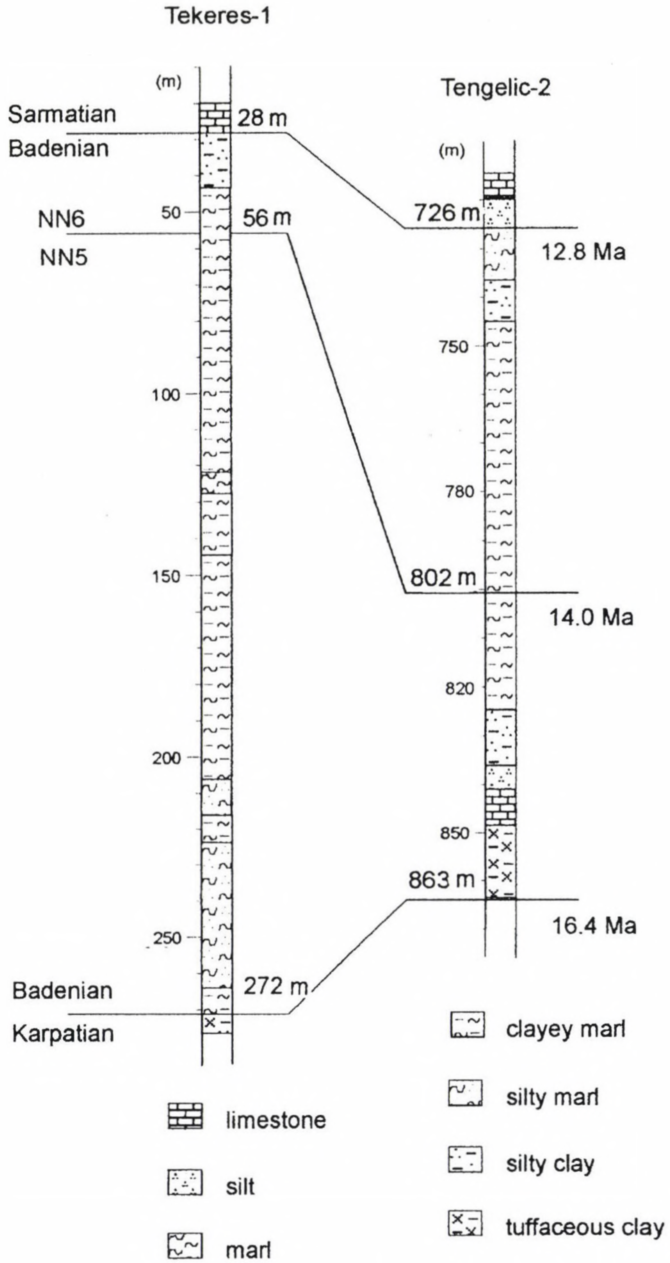

Fig. 2
Biostratigraphic correlation and lithology of core Tekeres-1 and Tengelic-2, after Nagymarosy (1982, 1985) and Halmai et al. (1982). Absolute ages are from Steininger et al. (1990)

from 844 to 738 m. The lithology of the Badenian in both cores can be referred to as "schlier", meaning clayey, silty marls.

The picked foraminifera specimens were mounted on Chapman-slides, making it easy to observe morphologic variability in detail. Foraminifera were picked from the 125 micrometer – 0.6 mm fraction. More then sixty samples were examined from both cores, each sample containing more than 200 specimens of benthic foraminifera. The species presented in this work make up more than 5% of the benthic assemblage in at least one sample of the two cores. The species are presented in alphabetic order of the genera. Where it was considered necessary the basis of separating species or formae is given, and where possible their depth distribution as well. Paleo-water depth is assessed by employing the plankton\benthos ratio (van der Zwaan et al. 1990). The studied foraminifera fauna is fairly well preserved. Only few specimens of foraminifera showed signs of reworking.

Taxonomy

Ammonia beccarii (Linné, 1758)

Plate I, figs 1–2

Nautilus beccarii – Linné, C., 1758; Systema naturae ed 10., tomus 1. p. 710 (fide Ellis & Messina)
Rosalina viennensis – d'Orbigny, 1846; For. Foss. Vienne, pl. p. 177, Pl. 10, figs 22–24
Rotalia papillosa Brady – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, plate 5, fig. 17
Rotalia beccarii (Linné) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, plate 5, fig. 18
Ammonia beccarii (Linné) – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate CLV, figs 1–3
Ammonia beccarii (Linné) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 67, Nr. 129, Pl. 61, figs 1–5
Ammonia beccarii (Linné) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 56, Plate 2, figs 1–3

Most of the observed specimens in our material have pustules filling the umbilical area. In most samples where this species is present there are signs of reworking, except at the top of NN5 in the Tekeres core. There it is common and forms an assemblage with *Fursenkoina acuta.*

Astrononion cf. *italicum* Cushman & Edwards, 1937

Plate I, figs 3–5

Astrononion italicum – Cushman & Edwards, 1937; Contr. Cushman Lab. Foram. Res., vol. 13, pt. 1, p. 35, pl. 3, figs 19–20
Astrononion italicum Cushman & Edwards – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 50, pl. 6, figs 3a–b

Melonis sp. – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 64, Plate 27, figs 5–6

The observed specimens in our material show much resemblance with small-sized flat *Melonis barleeanum*, differing by slightly thickened sutures. This resemblance led Rupp (1986) to determine it as *Melonis* sp., since the actual openings of the supplementary apertures are not always visible. However, here the thickening of the sutures is considered an initial stage of supplementary aperture development, characterizing the genus *Astrononion*.

Bolivina hebes Macfadyen, 1930

Plate I, figs 6–8

Bolivina hebes – Macfadyen, 1930; Geol. Survey Egypt, p. 59, pl. 2, figs 5a–c (fide Ellis & Messina)
Bolivina hebes Macfadyen – Cicha & Zapletalova, 1963; Die Vertreter der Gattung Bolivina ... Sborn. Ustr. Ust. Geol., paleontol., vol. XXVIII. (1961), p. 157, figs 30a–b
Bolivina merecuanai – Sellier de Civrieux, 1976: Estúdio sistemático y ecológico de las Bolivinitidae recientes de Venezuela. Oriente, Univ, Inst, Oceanogr, Cuadernos Oceanogr, Cumana, Venezuela, no. 5, p. 15, plate 9, 1–10, plate 10, 2–9 (fide Ellis & Messina)

Illustration provided for *B. merecuanai* from recent material off the coasts of Venezuela shows forms covering the variation existing in our Miocene material. Thus, it is considered a junior synonym of *B. hebes. B. merecuanai* has been described from a neritic depth.

Bolivina plicatella Cushman, 1930

Plate II, fig. 1

Bolivina plicatella – Cushman, 1930; Florida State Geol. Surv. Bull., vol, 4, p. 46, pl. 8, fig. 10a–b
Bolivina plicatella Cushman – van der Zwaan, 1982, Utr. Micropal. Bull. vol. 25, p. 141, pl. 2, figs 1–5, textfig. 59.

According to Cushman's description the ornamentation forms two longitudinal ridges with connecting transverse bridges. Although in our material the depressions show a slightly more irregular pattern, nevertheless it is considered *B. plicatella.*

Bolivina scalprata (Schwager) var. *miocenica* Macfadyen, 1930

Plate II, fig. 2.

Bolivina scalprata (Schwager) var. *miocenica* – Macfadyen, W., A., 1930, Miocene Foram. from the Clysemic Area of Egypt and Sinai. Geol. Survey Egypt, Taf 4. Abb. 2 a, b, 61 Cairo (fide Ellis & Messina)
Bolivina scalprata miocenica Macfadyen – Cicha & Zapletalova, 1963; Die Vertreter der Gattung Bolivina ... Sborn. Ustr. Ust. Geol., paleontol., vol. XXVIII. (1961), p. 125 figs 6a–d
Bolivina scalprata var. *miocenica* – Macfadyen – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate XCI, fig. 4

Sutures, easily observable under the light microscope, are folded like in *B. viennensis*, but the test is smooth, lacking any ornamentation. Outline of the test has generally the same proportion as *B. spathulata f. dilatata*, with a length/width ratio of 2:1. The periphery is acute, sharp, almost keeled, in perfect accordance with Macfadyen's description.

Bolivina spathulata (Williamson, 1858)

Plate II, figs 3–5

Textularia variabilis Williamson var. *spathulata* – Williamson, 1858, Rec. Foram. G.B., p. 76, pl. 6, figs 164, 165

Bolivina spathulata Williamson – van der Zwaan, 1982, Utr. Micropal. Bul. vol. 25, p. 141, pl. 2, figs 1–5, textfig. 59

Bolivina dilatata – Reuss, 1850; Neue Foraminiferen aus den Schichten des Österreichischen Tertiär Beckens. K. Akad. Wiss.-naturwiss. Denkschr. Wien, Cl., vol. 1, p. 381, figs 15a–c (fide Ellis & Messina)

According to van der Zwaan (1982) the intraspecific variation of *B. spathulata* ranges from short, stout forms with a length/width ratio of 2:1 to elongated slender forms with a ratio of 3:1. Applying these ratios to our material the stout forms dominate, where the real slender forms are missing. Nevertheless, to split the variation here specimens with a length/width ratio smaller than 2:1 are considered *B. spathulata f. dilatata*, with a ratio larger then 2:1 *B. spathulata f. spathulata*. The species is more frequent in Tekeres-1, and in NN5, although present in Tengelic-2 and NN6 as well.

Short, stout forms	Elongated, slender forms
B. spathulata f. dilatata	*B. spathulata* f. *spathulata*

Bolivina viennensis Marks, 1951

Plate II, figs 6–7

Bolivina viennensis – Marks, 1951, Cushman Found. for Foram. Res. vol. 11, part 2, p. 60, pl. 7, figs 1–2

Bolivina viennensis Marks – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 57, Plate 7, figs 10–12

We observed folded sutures, similarly to *B. scalprata var. miocenica* in perfect accordance with the original material and description provided by Marks (1951). The difference between *B. scalprata var. miocenica* and *B. viennensis* is based on the appearance of the wall. The former one has smooth walls, while the later has a rough, porous wall, giving it a "sandy appearance" (Marks 1951).

Bulimina elongata d'Orbigny, 1846

Plate III, figs 1–6

Bulimina elongata – d'Orbigny, 1846, For. Foss. Vienne, Pl. 11. figs 19–20

The species shows considerable morphologic variation. Morphotypes vary from short to elongated forms, both with and without spines. These four easily recognizable characters gave the basis of differentiating the four morphotypes under the light microscope (see table). Some of the specimens counted as "spines absent" perhaps would show a few tiny spines in the SEM (scanning electron microscope), which are unobservable by the magnification of a light microscope. The priority of the name *B. aculeata* d'Orbigny, 1826 has been considered over *B. elongata*, but the phylogenetic relationship existing between the taxa, led us to favor the name *B. elongata*.

	Lhort, stout forms Length/width ratio lower than 2:1	Long slender forms Length/width ratio higher than 2:1
Spines present	*Bulimina elongata f. subulata*	*Bulimina elongata* f. *aculeata*
Spines absent	*Bulimina elongata f. minima*	*Bulimina elongata* f. *elongata*

The four morphotypes were found to show correlation with water depth as estimated by plankton/benthos ratio (Fig. 3). This supports the idea that the morphologic variation is ecophenotypic in nature. This finding would suggest that the *Bulimina* stratigraphy (Cicha and Ctyroka 1988) is an ecostratigraphy, following major transgressive and regressive cycles in the Paratethys.

Bulimina elongata f. subulata Cushman & Parker, 1937

Plate III, fig. 1

Bulimina elongata f. subulata – Cushman & Parker, 1937, Notes on some European Eocene species of Bulimina. Contr. Cushman Lab. Foram, Res. Sharon, Mass., USA, vol. 13, pl. 2, p. 51, figs 6–7
Bulimina subulata Cushman & Parker – van der Zwaan, 1982, Utr. Micropal. Bul. vol. 25, p. 144, pl. 3, figs 4–8, textfig. 61a.
Bulimina elongata subulata Cushman & Parker, 1937 – Cicha & Ctyroka, 1988, The genus Bulimina ... Revue de Paléobiologie. Vol. Spéc. No. 2., Benthos 86', ISSN 0253–6730, p. 502, plate 1. figs 6–7
Bulimina elongata lappa Cushman & Parker, 1937 – Cicha & Ctyroka, 1988, The genus Bulimina... Revue de Paléobiologie. Vol. Spéc. No. 2., Benthos 86', ISSN 0253-6730, p. 502, plate 2. fig. 6–7
?Bulimina insignis Luczkowska, 1960 – Cicha & Ctyroka, 1988, The genus Bulimina ... Revue de Paléobiologie. Vol. Spéc. No. 2., Benthos 86', ISSN 0253–6730, p. 503, plate 1. figs 12–14

This is a short, stout form with spines on initial parts of the test. *B. insignis* is probably a synonym as the significance of the shape of chambers is doubtful. This morphotype was the most common *B. elongata* morphotype in the material. It is found to prefer water depths deeper than 50–100m (Fig. 3).

Bulimina elongata f. aculeata d'Orbigny, 1826

Plate III, fig. 3

Bulimina aculeata – d'Orbigny, 1826, Tabl. méth. Ceph., Ann. Sci. Nat., sér. 1, vol. 7, p. 269 Fornasini, 1902, Mem. R. Accad. Sci. Ist., ser. 5, vol. 9, p. 153, fig. 4.
Bulimina aculeata d'Orbigny – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 98, plate 2, fig. 16

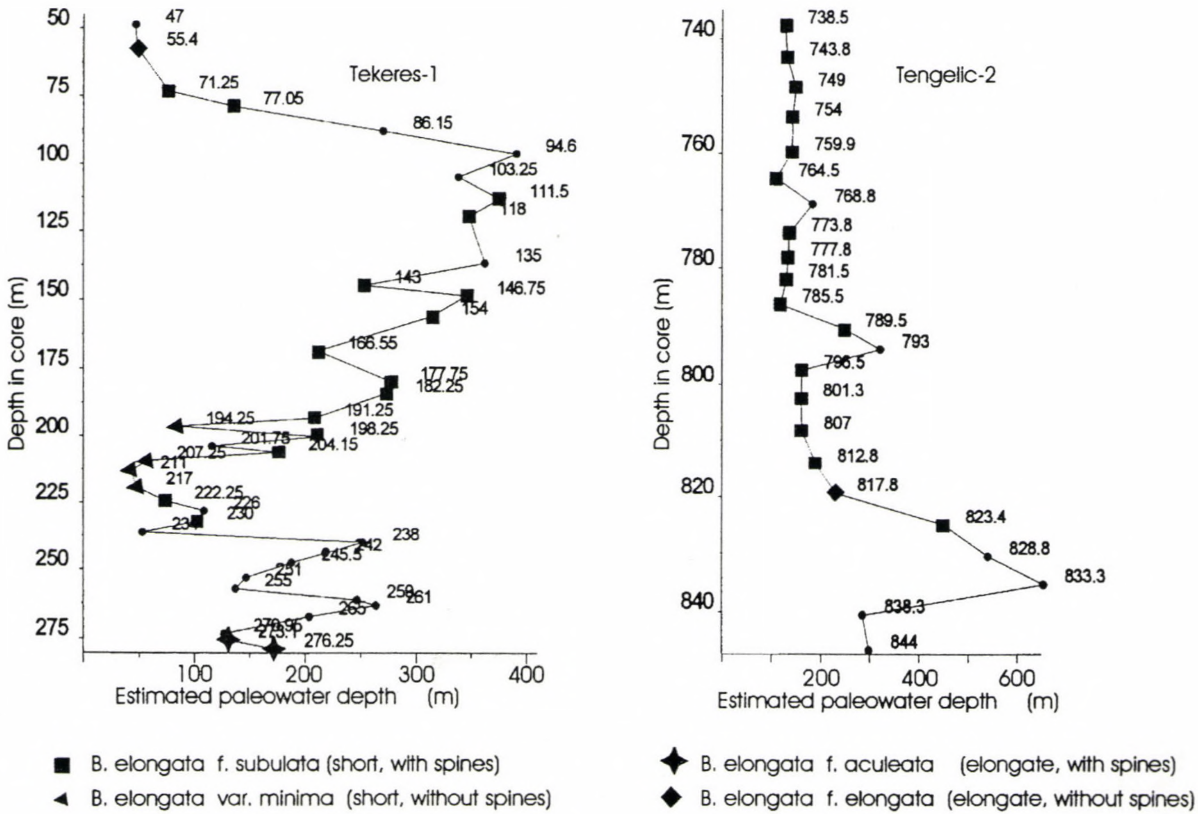

Fig. 3
Dominant morphotype of the *Bulimina elongata* denoted by asterix, square, diamond, triangle where *B. elongata* is present in sufficient number. Otherwise where *B. elongata* is rare or absent, but sampled, it is denoted by small dots. The horizontal axis is the estimated paleo-water depth, based on plankton/benthos ratio.

Bulimina aculeata d'Orbigny – van der Zwaan, 1982, Utr. Micropal. Bul. vol. 25, p. 142, pl. 3, figs 1–2, textfig. 61e, f.
Bulimina aculeata d'Orbigny – Cicha & Ctyroka, 1988, The genus Bulimina ... Revue de Paléobiologie. Vol. Spéc. No. 2., Benthos 86', ISSN 0253–6730, p. 503, plate 1. figs 8–11, 15

It is an elongated form with spines on initial chambers. It is quite rare in the material, and occurs only at the Karpatian–Badenian boundary interval in core Tekeres-1 (Fig. 3).

Bulimina elongata var. *minima* Tedeschi & Zanmatti, 1957

Plate III, fig. 4

Bulimina elongata var. *minima* – Tedeschi & Zanmatti, 1957, Diagnosi di forme nuove. Riv. Ital. Pal. Strat., Milan, 1957, vol. 63, no. 4, p. 249, plate 3a–c (fide Ellis & Messina)

This is a short morphotype without spines or similar ornamentation on the younger part of the test. It is the dominating morphotype of *B. elongata* in the Tekeres core (211–217 m), where it is found together with *Epistominella smithi.*

Bulimina elongata f. elongata d'Orbigny, 1846

Plate III, fig. 6

Bulimina elongata – d'Orbigny, 1846, For. Foss. Vienne, Pl. 11. figs 19–20
Bulimina elongata d'Orbigny – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 99, plate 2, fig. 3
Bulimina elongata d'Orbigny – Papp & Schmid, 1985, Die Foss. Foram. des Tertiären Beckens von Wien p. 73, Nr. 137, Pl. 63, figs 5–9
Bulimina elongata – Cicha & Ctyroka, 1988, The genus Bulimina ... Revue de Paléobiologie. Vol. Spéc. No. 2., Benthos 86', ISSN 0253-6730, p. 502, plate 1. figs 1–4
Bulimina elongata vagina – Cicha & Ctyroka, 1988, The genus Bulimina ... Revue de Paléobiologie. Vol. Spéc. No. 2., Benthos 86', ISSN 0253–6730, plate 1. fig. 5

This is an elongated morphotype without spines. This morphotype occurs only in two samples at the top of the NN5 zone coinciding with the end of a major regression (Tekeres-1, 55.4 m; Tengelic-2, 817.8 m). It is suspected that its occurrence is due to changed nutritional conditions related to the drop of sea level.

Bulimina costata d'Orbigny, 1826

Plate II, fig. 8

Bulimina costata – d'Orbigny, 1826, Tabl. méth. Ceph., Ann. Sci. Nat., sér. 1, vol. 7, p. 269 Fornasini, 1901, Boll. Soc. Geol. Ital., vol 20 p. 174, pl. 1.
Bulimina striata – d'Orbigny, 1826, Tabl. méth. Ceph., Ann. Sci. Nat., sér. 1, vol. 7, p. 269; Fornasini, 1902, Mem. R. Accad. Sci. Ist., ser. 5, vol. 9, p. 372, pl. 1.
Bulimina buchiana – d'Orbigny, 1846, For. Foss. Vienne, Pl. 11. figs 15–18
Bulimina buchiana d'Orbigny – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 98, plate 2, fig. 2

Bulimina costata d'Orbigny – van der Zwaan, 1982, Utr. Micropal. Bul. vol. 25, p. 143, pl. 3, figs 9–11, textfig. 62
Bulimina costata d'Orbigny – Papp & Schmid, 1985, Die Foss. Foram. des Tertiären Beckens von Wien p. 72, Nr. 136, Pl. 63, figs 1–4

This is the only costate Bulimina species present in our material. *B. striata* and *B. buchiana* are considered here junior synonyms of *B. costata*.

Bulimina pyrula d'Orbigny, 1846

Plate IV, figs 1–4

Bulimina pyrula – d'Orbigny, 1846, For. Foss. Vienne, p. 184, Plate 11, figs 9–10
Bulimina ovata – d'Orbigny, 1846, For. Foss. Vienne, p. 185, Plate 11, figs 13–14
Bulimina pupoides – d'Orbigny, 1846, For. Foss. Vienne, p. 185, Plate 11, figs 11–12
Bulimina pyrula (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 58
Bulimina ovata (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2 , p. 57
Bulimina pupoides (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 58
Bulimina pyrula (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien pp. 69–70, Nr. 133, Nr. 134, Nr. 135, Pl. 62, figs 2–10
Praeglobobulimina pyrula (d'Orbigny) – Rupp, 1986; Paloökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 65, Plate 31, fig. 6
Praeglobobulimina pupoides (d'Orbigny) – Rupp, 1986; Paloökologie der Foram. in der Sandschalerzone ... Beitr. Palont. Österr. vol. 12, Wien, p. 65, Plate 31, figs 4–5

This is one of the few species where previous authors followed the species concept extensively exploited here. Papp and Schmid (1985) lumped *B. ovata* and *B. pupoides* under the name of *B. pyrula*, documenting the variability by biometric measurements as well.

Elongated, slender form	Transitional form	Short, globular form
B. pyrula f. ovata	B. pyrula f. pupoides	*B. pyrula f. pyrula*

Cassidulina crassa d'Orbigny, 1839

Plate IV, figs 5–7

Cassidulina crassa – d'Orbigny, 1839; Foram. Amer. Merid., p. 56, pl. 7, figs 18–20 (fide Ellis & Messina)
Cassidulina crassa d'Orbigny – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 68
Cassidulina crassa d'Orbigny – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 113, plate 2, fig. 9
Cassidulina crassa d'Orbigny – Korecz-Laky, 1982; A Tengelic 2. sz. fúrás foraminifera faunája. Ann. Inst. Geol. Hung. vol. LXV p. 175, plate 9, fig. 5.
Cassidulina crassa d'Orbigny – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 11, fig. 4

In spite of the inflated character of the chambers this lacks the sub-globular form of C. *subglobosa*, as it is flatter. The periphery is rounded.

Cassidulina laevigata d'Orbigny, 1826

Plate III, figs 7–8

Cassidulina laevigata – d'Orbigny, 1826, Ann. Sci. Nat., vol. 7, p. 282, tab. 15, figs 4–5
Cassidulina laevigata d'Orbigny – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 10, figs 11–12
Cassidulina laevigata var. *carinata* – Silvestri, 1896, Acc. Pont. Nuovi Lincei, Mem., vol. 12, p. 104, pl. 2, fig. 10
Cassidulina carinata (Silvestri) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 10, figs 8–10

C. *carinata* was not considered a separate species, but a morphologic, most likely ecophenotypic variation of C. *laevigata*. The serrately keeled morphotype, C. *laevigata* var. *carinata* was just as abundant in the material, as the non-keeled C. *laevigata*, which is termed "C. *laevigata f. laevigata*" here.

Cassidulina oblonga Reuss, 1850

Plate V, fig. 1

Cassidulina oblonga – Reuss, 1850; Denkschr. Akad. Wiss., Math.-naturwiss. Cl., vol. 1, p. 376 (fide Ellis & Messina)
Cassidulina oblonga Reuss – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 11, fig. 3

Continuous morphologic variation was observed between C. *oblonga* and C. *subglobosa* (Brady). According to the original descriptions numerous other features can be used to distinguish the species, but here the outline was considered the most important, and inflatedness the second most important character. Thus, C. *oblonga* has an egg-like outline ("eiförmig") and flush sutures. Generally, it is larger in size than C. *subglobosa*. The original descriptions make it possible to use other hierarchies of characteristics to differentiate the two species. Thus, it is believed that a different set of characters might have divided the morphologic continuity in a different way. Nevertheless, the two taxa are considered to be separate and not a continuous plexus.

Cassidulina subglobosa Brady, 1881

Plate V, figs 2–3

Cassidulina subglobosa – Brady, 1881; Notes on some of the reticularian Rhisopoda of the "Challenger" Expedition Part III, Quart. Jour. Micr. Sci., London 1881, n.s. vol. 21, p. 60. Fig. Brady, 1884, Rep. Voy, Challenger, Zool, vol. 9. pl. 54 figs 17a–c (fide Ellis & Messina)
Globocassidulina subglobosa Brady – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 62, Plate 20, figs 9–10

As mentioned before with C. *oblonga* this species shows a morphologic continuity. It is different from C. *oblonga* in its almost globular form and usually inflated chambers, resulting in strongly depressed sutures.

Cassidulina teretis, Tappan, 1951

Plate V, fig. 4

Cassidulina teretis – Tappan, 1951; Northern Alaskan index foraminifera. Cushman Found. Foram. Res., Contr., Washington D.C., 1951, vol. 2, plate 1, fig. 30
? *Cassidulina cruysi* – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 68, pl. 8, figs 3a–4c

An umbilical boss is clearly visible on both sides of the test in accordance with Tappan's description of the species. Similarly built *C. cruysi* has shell material filling the umbilical area, and its chambers are inflated. The observed specimens of *C. teretis* are more keeled, with an umbilical boss. The original descriptions of these species do not provide a differential diagnosis concerning their relationship. This might be due to the short time between publishing the descriptions of these species. Here they are regarded as synonyms.

Cibicides lobatulus (Walker & Jacob, 1798)

Plate V, figs 5–8; Plate VI, figs 1–6

Nautilus lobatulus – Walker & Jacob, 1798, p. 642, fig. 36. (fide Ellis & Messina)
Cibicides refulgens – Monfort, 1808, Conchyliologie systematique et classification metodique des coquilles. Paris, France, F.Schoell, 1808, tome 1, p. 123, fig. on p. 122
Truncatulina letkésiensis – Franzenau, 1894; Adatok Letkés faunájához. Math. Term. Közl., Budapest, vol. 26, nr. 1, p. 7, pl. 1, figs 11a–c
Cibicides lobatulus (Walker & Jacob) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 73
?*Cibicides lobatulus* (Walker & Jacob) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 189, plate 9, figs 16–17
Cibicides lobatulus (Walker & Jacob) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien Nr. 120, Pl. 56, figs 1–5 and Nr. 122, Pl. 57, 1–3
?*Planulina austriaca* (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 65, Nr. 124, Pl. 58, figs 4–9
Cibicides lobatulus (Walker & Jacob) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 11, figs 8–10

The species is abundantly present in our samples providing enough material for the recognition of its full morphotypic variation. It is believed that the variation is partly due to different ontogenetic stages of the species, and partly due to differences in microhabitat.

Juvenile form	Sharply keeled, with irregularities	Lumping category
C. lobatulus f. letkesiensis	C. lobatulus f. lobatulus	*C. lobatulus f. refulgens*

C. lobatulus f. letkesiensis (Franzenau) is a distinct, easily recognizable phenotype of juvenile *C. lobatulus*. The most prominent feature of the taxon is easily observed on the umbilical side, where a central initial chamber is embraced by 5–6 others forming the last whorl (Plate VI. figs 4–6).

C. lobatulus f. lobatulus in the present work stands for *C. lobatulus sensu stricto*. This implies a recognizable irregularity of the last chambers, planoconvexity and coarse pores (especially on the spiral side). The last chamber often stands off from the previous coil, leaving a gap between the two coils. Mostly fully grown, large sized specimens were included here (Plate V. figs 5–8).

C. lobatulus f. refulgens (Monfort) is the "lumping" category, including all *C. lobatulus sensu lato*, making *C. lobatulus f. refulgens* one of the morphologically most variable taxa in the material. This comes partly from the practice of using it as a lumping category, where all specimens determined *C. lobatulus* were put, except for typical *C. lobatulus f. lobatulus* and *C. lobatulus f. letkesiensis*. Generally the observed specimens were fully grown, but small-sized individuals lack the irregularity of *C. lobatulus f. lobatulus* (Plate VI. figs 1–3).

Cibicides ungerianus (d'Orbigny, 1846)

Plate VI, figs 8–10

Rotalina ungeriana – d'Orbigny, 1846, For. Foss. Vienne, p. 157, plate 8, figs 16–18
Truncatulina pseudoungeriana – Cushman, 1922, Prof. Pap. US Geol. Surv., vol. 129, p. 97, pl. 20, fig. 9
Cibicides ungerianus (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 73, pl. 8, figs 2a, b
Cibicides ungerianus (d'Orbigny) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 190, plate 11, figs 8–9
Cibicides ungerianus (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 60, Nr. 111, Pl. 51, figs 7–11
Cibicidoides ungerianus (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 12, figs 4–6

Cushman (1922) differentiated *C. pseudoungerianus* and *C. ungerianus* primarily on the visibility of the spiral suture. If the early whorls on the dorsal side were observed, it was named *C. ungerianus;* in those cases where they were obscured by shell material, it was designated *C. pseudoungerianus*. Following this criteria, a wide variation is present in our samples, where *C. pseudoungerianus* is more frequent.

Dentalina elegans d'Orbigny, 1846

Plate VI, fig. 7

Dentalina elegans – d'Orbigny, 1846, For. Foss. Vienne, p. 45, plate 1, figs 52–56
Dentalina pauperata – d'Orbigny, 1846, For. Foss. Vienne, p. 46, plate 1, figs 57–58
Dentalina elegans d'Orbigny – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 160, plate 8, fig. 13
Dentalina pauperata d'Orbigny – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 160, plate 8, fig. 14
Dentalina elegans d'Orbigny – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 28, Nr. 21, Pl. 10, figs 1–5, Nr. 22, Pl. 10, figs 6–8

The species name *D. pauperata* described in d'Orbigny's classical monograph on the Vienna Basin has been found to be synonymous to *D. elegans* (Papp and Schmid 1985), thus we did not follow the practice of Korecz-Laky, who differentiated them.

Epistominella smithi (R.E. & K.C. Stewart, 1930)

Plate VIII, figs 1–3

Pulvulinella smithi – R.E. & K.C. Stewart, 1930, Post-Miocene foram. from the Ventura Quadrangle. Jour. Pal., Menasha, Wis., USA, vol. 4, no. 1, p. 70, pl. 9, fig. 4.
Epistominella pulchella – Husezima & Maruhasi, 1944, A new genus and thirteen new species of foraminifera ... Sigenkagaku Kenkyusyo (Res. Inst. Nat. Resources), Journ. Tokyo, 1944, vol. 1, no. 3, p. 398, plate 34, figs 10a–c
Epistominella pulchella Husezima and Maruhasi – Loeblich & Tappan, 1988, Foraminiferal genera and their classification, p. 574, plate 627, figs 1–6

E. pulchella and *E. smithi* are considered synonyms, as the illustrations of the original descriptions suggest. No differential diagnosis has been provided concerning the relationship of the two species. In our material two different phenotypes are present; the common form is delicately built, strongly keeled, with transparent walls, in perfect accordance with the descriptions of *E. smithi* and *E. pulchella.* The few specimens of the other morphotype are less keeled, robustly built and strongly calcified. The interior-marginal aperture curves away from the periphery toward the umbilicus.

Fursenkoina acuta (d'Orbigny, 1846)

Plate VII, figs 1–2

Polymorphina acuta – d'Orbigny, 1846, For. Foss. Vienne, p. 82, pl. 75, figs 1–6
Fursenkoina acuta (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiren Beckens von Wien p. 82, Nr. 165, Pl. 75, figs 1–6
Fursenkoina acuta (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 62, Plate 19, figs 6–8
Virgulina schreibersiana Czjek – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 176, plate 5, fig. 4

We agree with Papp and Schmid (1985) in finding *F. schreibersiana* synonymous with *F. acuta.* The illustration of *V. schreibersiana* given by Korecz-Laky (1969) working with the same material is definitely identical with *F. acuta.*

Gavelinopsis praegeri (Heron-Allen & Earland, 1913)

Plate VII, figs 3–4

Discorbina praegeri – Heron-Allen & Earland, 1913; Clare Isl. Surv., Roy. Irish Acad. Proc., vol. 31, sec. 3, p. 122, pl. 10, figs 8–10
Gavelinopsis praegeri (Heron-Allen & Earland) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 62, Plate 20, figs 8–10
Gavelinopsis praegeri (Heron-Allen & Earland) – Loeblich & Tappan, 1988, Foraminiferal genera and their classification, p. 560, plate 608, figs 6–12

This species is always easily recognizable by its umbilical plug on the ventral side. In our material the test is biconvex and this results in a resemblance to lenticular *Oridorsalis umbonatus,* which is strengthened by the fact that both species have 2 to 2 1/2 slowly enlarging last whorls on the spiral side.

Gyroidina soldanii d'Orbigny, 1825

Plate VII, figs 5–6

Gyroidina soldanii – d'Orbigny, 1825, p. 122, Nr. 5. fide Ellis & Messina
Rotalina soldanii – d'Orbigny, 1846, For. Foss. Vienne, p. 155, Plate 8, figs 10–12
Gyroidina soldanii var. *altiformis* – R.E. & K.C. Stewart, 1930, Post-Miocene foram. from the Ventura Quadrangle. Jour. Pal., Menasha, Wis., USA, 1930, vol. 4, no. 1, p. 67, pl. 9, fig. 2 fide Ellis & Messina
Gyroidina soldanii d'Orbigny – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 64
Gyroidina soldanii d'Orbigny – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate CLIV, figs 1–3
Gyroidina soldanii d'Orbigny – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 60, Nr. 109, Pl. 50, figs 4–9
Gyroidinoides soldanii d'Orbigny – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 63, Plate 22, figs 3–5

Gyroidina soldanii var. *altiformis* is differentiated on the basis of backward tilting dorsal faces resulting in oblique sutures on the dorsal side (R. E. Stewart and K. C. Stewart 1930). According to Stewart and Stewart (1930) it is a "deeper form", referring to its deeper marine habitat, than that of the typical *G. soldanii. G. soldanii* var. *altiformis* was often observed in our material, indicating that some of our samples were likely to come from the lower limits of the depth range of *G. soldanii.*

The most important features used in distinguishing *G. soldanii* from *G. umbonata* is that *G. soldanii* has: 1. more chambers in the last whorl, 2. a larger size, 3. last chambers which are raised over the spiral plain, causing the last whorl to stand off from the rest of the test.

Gyroidina umbonata (Silvestri, 1898)

Plate VII, figs 7–9

Rotalia soldanii (d'Orbigny) var. *umbonata* – Silvestri, 1898, Foraminiferi pliocenici della Provincia di Siena, Parte II, Accad. Pont. Nuovi Lincei, Mem., Roma, Italia, 1898, vol. 15, p. 329, plate 6, fig. 14
Gyroidinoides umbonatus (Silvestri) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 63, Plate 22, figs 6–8

In comparison with *G. soldanii, G. umbonata* is smaller and has only 5 chambers in the last whorl, in accordance with the original description given by Silvestri (1898). For further details see differential diagnosis for *G. soldanii.*

Hanzawaia boueana (d'Orbigny, 1846)

Plate VIII, figs 4–6

Truncatulina boueana – d'Orbigny, 1846, For. Foss. Vienne, p. 169, Plate 9, figs 24–26
non *Nonionina boueana* – d'Orbigny, 1846, For. Foss. Vienne, p. 108, Plate 5, figs 11–12
Cibicides boueanus (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 72, pl. 8, figs 9a, b

Cibicides boueanus (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 64, Nr. 121, Pl. 56, figs 6–9
non *Hanzawaia boueana* (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 45, Nr. 76, Pl. 35, figs 1–5
Hanzawaia boueana (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 63, Plate 23, figs 1–3

H. boueana is considered here as a strongly keeled plano-convex form. *Nonionina boueana* (d'Orbigny) and its present-day designation to *H. boueana* by Papp and Schmid (1985) lacks these characteristics. On the other hand, *Truncatulina boueana* (d'Orbigny) and Papp and Schmid's designation of *C. boueana* has a sharp keel, and thus are synonyms. Generally, specimens considered *H. bouana* in the present work are having transparent walls and finely pored under the light microscope. This species intergrades with *C. ungerianus.*

Heterolepa dutemplei (d'Orbigny, 1846)

Plate IX, figs 1–6; Plate X, figs 1–2

Rotalina dutemplei – d'Orbigny, 1846, For. Foss. Vienne, p. 157, Plate 8, figs 19–21
Rotalina haidingerii – d'Orbigny, 1846, For. Foss. Vienne, p. 154, Plate 8, figs 7–9
Rotalina kalembergensis – d'Orbigny, 1846, For. Foss. Vienne, p. 151, Plate 7, figs 19–21
Cibicides dutemplei (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 72
Cibicides dutemplei (d'Orbigny) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 189, plate 10, figs 1–2
Eponides haidingerii (d'Orbigny) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 181, plate 11, figs 21–22
Heterolepa dutemplei (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 61, Nr. 112, Pl. 52, figs 1–6; p. 59, Nr. 108, Pl. 50, figs 7–9 and p. 57, Nr. 103, Plate 46, figs 5–9
Heterolepa dutemplei (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 63, Plate 24, figs 7–9
?*Cibicidoides austriacus* (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 59, Plate 12, figs 7–9
Heterolepa dutemplei (d'Orbigny) – Holcov & Maslowsk, 1995, Phenotypic variability of genera Heterolepa and Gemellides from Central Paratethys Oligocene and Miocene. Rev. Esp. de Micropal. vol. 27, 1, pp. 51–69
non *Planulina austriaca* (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 65, Nr. 124, Pl. 58, figs 1–6 and p. 66, Nr. 125, Pl. 59, figs 1–6
non *Cibicides kullenbergi* Parker, 1953 – Phleger, F.B., Parker, F.L. & Peirson, J.F., 1953, Swedish Deep-Sea Expedition 1947–48, Reports, 7(1), 49, pl. 11, figs 7–8
non *Cibicidoides mundulus* (Brady, Parker and Jones, 1888) – van Morkhoven, F.P.C.M., W.A. Berggren & A.S. Edwards (1986): Cenozoic cosmopolitan deep-water benthic foraminifera. ed.: H. J. Oertli, Bulletin des Centres de Recherches exploration-production Elf-Aquitaine. Mem. 11, pp. 65–67, plate 21

In our material *Heterolepa dutemplei* showed high phenotypic variability similar to the rest of the Cibicides group. We agree with Papp and Schmid (1985) considering its ecophenotypic nature. It is suspected that this variation is depth related, where the flatter and more lobate morphotype has a shallower habitat, while the more compact robust form has a deeper habitat. The first morphotype is the *R. haidingerii* and *R. kalembergensis* (inner neritic 100 m) (Plate X. figs 1–2), followed by the typical *H. dutemplei* morphotype (outer neritic, upper bathyal 100–600 m) which we found dominant in our material (Plate IX, figs 1–3). *C. kullenbergi* – *H. dutemplei* transitional forms were observed in one of the samples (Tengelic-2, 844 m) from greater bathyal depth (Plate IX, figs 4–6). *C. kullenbergi* is treated as a junior synonym of *Cibicidoides mundulus* by van Morkhoven et al. (1986), which was found to be a "bathyal-abyssal" species in accordance with our observation. Holcova and Maslowska (1995) were able to separate the shallow morphotypes from typical *H. dutemplei* by biometrical methods (discriminant analyses). However, in the overall picture drawn in their work they interpret this phenotypic variability as time-related, based on the results of the cluster analyses correlating morphometric clusters with Paratethys stages. We think these different time intervals represent different stages of basin development corresponding to different cycles of transgression, with different water depths.

C. austriacus in Rupp (1986) is regarded as a transitional form between *C. lobatulus* and *H. dutemplei*, definitely different from the much flatter *P. austriaca*. In the observed material such morphotypes were assigned to *C. lobatulus s.l.*

Hoeglundina elegans (d'Orbigny, 1826)

Plate VIII, figs 7–8

Rotalina elegans – d'Orbigny; 1826, Tabl Meth. p. 272 no. 6, nomen nudum, fide Ellis & Messina
Rotalina partchiana – d'Orbigny, 1846, For. Foss. Vienne, p. 153, Plate 7, figs 28–30 and Plate 8, figs 1–3
Epistomina elegans (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 64
Epistomina elegans (d'Orbigny) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 182, plate 10, fig. 10
Hoeglundina elegans (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 59, Nr. 106, Pl. 49, figs 1–6

A series of slit-like apertures parallel to the peripheral keel characterizes the species. In the case of earlier chambers, these secondary apertures may be covered by shell material.

Hoeglundina elegans has an aragonitic wall (Loeblich & Tappan, 1988, p.446). The presence of well-preserved specimens of such aragonitic species means extremely good preservation.

Nonion commune (d'Orbigny, 1825)

Plate X, figs 4–5

Nautilus scapha – Fichtel & Moll 1798; Testacea Microscopica aliaque minuta ex generibus Argonauta et Nautilus ad naturam delineata et descripta, Wien, A. Pichler p. 105, Pl. 19, figs d–f
Nonionina communis – d'Orbigny, 1825; p. 128, Nr. 20 (fide Ellis & Messina)
Nonionina communis – d'Orbigny, 1846, For. Foss. Vienne, p. 106, Plate 5, figs 7–8
Nonion scaphum (Fichtel & Moll) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 49, pl. 5, figs 16a–b
Nonion boueanum (d'Orbigny) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 168, pl. 5, fig. 16, pl. 6, fig. 16

Nonion boueanum (d'Orbigny) – Korecz-Laky, 1982; A Tengelic 2. sz. fúrás foraminifera faunja. Ann. Inst. Geol. Hung. vol. LXV, p. 174, plate 9, fig. 6
Nonion commune (d'Orbigny) – Rögl & Hansen, 1984; Foraminifera described by Fichtel & Moll in 1798 – A revision of Testacea Microscopica – N. Denkschr. Nat. Hist. Mus. Wien, 3, p. 66,
Nonion commune (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 45, Nr. 74, Pl. 34, figs 1–5
Nonion commune (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 65, Plate 27, figs 11–12

We agree with Marks in finding *N. scaphum* (Fichtel & Moll) and *N. commune* to be synonyms. Priority should have been given to the name of *N. scapha,* but inadequate topotypes and tradition in the Paratethys led us to stick to the name *N. commune.* In the observed material under the light microscope there appears to be granular shell material in the umbilical area and radially along the sutures. However, scanning photographs revealed that this granular shell material is made of small round cylindrical grains, similarly to illustrations given in Rupp (1986).

Oridorsalis umbonatus (Reuss, 1851)

Plate XI, figs 3–4

Rotalina umbonata – Reuss, 1851; Über die fossilen Foraminiferen und Entromastraceen der Septarienthone der Umgegend von Berlin. Deutsch. Geol. Ges., Zeitschr., Berlin, Deutschland, 1851, vol. 3, p. 75, pl. 5, fig. 35 (fide Ellis & Messina)
Oridorsalis umbonatus – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 65, Plate 30, figs 4–6

According to Loeblich and Tappan (1988, p. 630), small secondary openings are present on the spiral side. This cannot be observed in our material, due to poor preservation. It is quite a rare species in our material.

Pullenia bulloides (d'Orbigny, 1825)

Plate XI, figs 1–2

Nonionina bulloides – d'Orbigny, 1825, p. 127, Nr. 2 (fide Ellis & Messina)
Nonionina bulloides – d'Orbigny, 1846, For. Foss. Vienne, p. 107, Plate 5, figs 11–12
Pullenia bulloides (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 69
Pullenia bulloides (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 45, Nr. 75, Pl. 34, figs 6–9
Pullenia bulloides d'Orbigny – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate LXXVI, figs 1–2
Pullenia bulloides (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 66, Plate 32, figs 4–5
Pullenia quinqueloba (Reuss) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 186, plate 4, fig. 6

This species is characterized best by its unique globular form, resulting from involute planispiral growth. The number of chambers visible on the last whorl may vary from four to five (Papp and Schmid 1985), or three to six (Loeblich and

Tappan 1988). The dominant form observed here, had five chambers in the last whorl. Perhaps this resulted in the excessive use of the name *P. quinqueloba* (Korecz-Laky 1969).

Siphonina reticulata (Czjek, 1848)

Plate X, fig. 3

Rotalina reticulata – Czjek, 1848, Haid. Nat. Abh., vol. 2, p. 45, pl. 3, figs 7–8
Siphonina fimbriata – Reuss, 1850, Neue Foraminiferen aus den Schichten des Österreichischen Tertiärbeckens. K. Akad. Wiss. Wien, Math.-Nat. Denkschr., Wien, Österreich, 1850, Vol. 1. p. 372
Siphonina reticulata (Czjek) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 65, pl. 8, figs 8a–c
Siphonina reticulata (Czjek) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 182, plate 10, fig. 12
Siphonina reticulata (Czjek) – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate CII, fig. 15 a–b

There is a general agreement that *S. reticulata* (Czjek) and *S. fimbriata* (Reuss) are synonyms. This species is an open marine mud dweller. It is present in greater numbers (just exceeding the cut-off level) in the lowermost Badenian.

Trifarina angulosa (Williamson, 1858)

Plate X, figs 6–7

Uvigerina angulosa – Williamson, 1858; Rec. Foram. G. B., p. 67, pl. 5, fig. 140
Angulogerina angulosa (Williamson) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 63, pl. 7, fig. 16
Trifarina angulosa (Williamson) – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate CLL, figs 1–4
Angulogerina angulosa (Williamson) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 56, plate 4, figs 2–3

T. angulosa is characterized by tri-serial growth throughout the test. In cross-section it is triangular with carinate angles. The aperture is similar to Uvigerina. It is more frequent in core Tekeres-1.

Uvigerina

Determining specimens of the genus *Uvigerina* the following hierarchy of criteria were used after van der Zwaan et al. (1986).

1. chamber arrangement, seriality
2. ornamentation ranging from absence to strong, and from costate to hispid.

Pursuing these criteria, three morphologic groups of *Uvigerina* were recognized in the material. These are the *U. semiornata,* the *U. peregrina,* and the *U. bononiensis* group.

The *U. semiornata* group has a generally tri-serial chamber arrangement throughout the whole test, the aperture is on a short neck, standing in a depression, while chambers overlap the preceding ones.

The *U. peregrina* group is characterized by its inflated chambers without much overlapping, due to the rather freely standing chambers the basal sutures become straight. This group is distinguished by a quite long apertural neck, and a tendency for reduced seriality.

The last group of *Uvigerina* is the *U. bononiensis* group. The most important criterion of the group is the pronounced reduction of seriality in the adult stage and the usually costate ornamentation.

The following table summarizes the recognized species in the material:

Group	Ornamentation		Species
U. semiornata	strong	costate hisp.-cost. hispid	*U. acuminata*
	weak	striate hisp.-cost. hispid ornam. absent	*U. semiornata* *U. semiornata* **U. sem. karre.*
U. peregrina	strong	costate hisp.-cost. hispid	**U. venusta* *U. aculeata* *U. aculeata*
	weak	striate hisp.-cost. hispid	**U. pygmea* *U. romaniaca*
U. bononiensis	strong	costate hisp.-cost. hispid	
	weak	striate hisp.-cost. hispid	*U. bononiensis*

* *U. semiornata* karreri, *U. pygmea* and *U. venusta* are present in the material in amounts <5%, thus a synonym list is not presented.

Uvigerina acuminata Hosius, 1893

Plate XI, figs 5–6

Uvigerina aculeata – Hosius, 1893; Beitrage zur Kenntnis der Foraminiferenfauna des Miozns, Tl. 1. Verh. naturhist. Ver. Reinl. Wesf., 49, p. 108, pl. 2, fig. 9.

Uvigerina acuminata – Hosius, 1895; Beitrage zur Kenntnis der Foraminiferenfauna des Ober-Oligozäns vom Doberg bei Bænde. J. ber. naturwiss. Ver. Osnabræck, 10, p. 167 (footnote)

Uvigerina acuminata Hosius – von Daniels, 1986; Uvigerina in the NW European Neogene, In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 92, plate 5, figs 1–8

Uvigerina acuminata Hosius – Cicha et al., 1986: Oligocene and Miocene Uvigerina from the Western and Central Paratethys. In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 144, plate 8, figs 2–3

Uvigerina macrocarinata – Papp & Turnovsky, 1953: Die Entwicklung der Uvigerinen in Vinbodon (Helvet und Torton) des Wiener Beckens. Jb. Geol. B., A., 96, p. 123, pl. 5, B, figs 1–3

Uvigerina macrocarinata Papp & Turnovsky – von Daniels, 1986; Uvigerina in the NW European Neogene, In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 94, plate 6, figs 1–6

Uvigerina macrocarinata Papp & Turnovsky – Cicha et al., 1986: Oligocene and Miocene Uvigerina from the Western and Central Paratethys. In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 154, plate 11, figs 5–7

According the differential diagnosis given by Cicha et al. (1986), the difference between *U. macrocarinata* and *U. acuminata* lies in whether the costae cross the sutures. However, in our material many transitional forms exist, thus based on this criteria the separation of the two species is impossible. Therefore, the two species are considered here to be synonyms.

Uvigerina semiornata d'Orbigny, 1846

Plate XII, figs 1–2

Uvigerina semiornata – d'Orbigny, 1846, For. Foss. Vienne, p. 189, Plate 11, figs 23–24

Uvigerina urnula – d'Orbigny, 1846, For. Foss. Vienne, p. 189, Plate 11, figs 21–22

Uvigerina semiornata d'Orbigny – Cicha et al., 1986: Oligocene and Miocene Uvigerina from the Western and Central Paratethys. In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 144, plate 8, figs 1–7; plate 9, figs 1–7, plate 10, figs 1–5

Here all the numerous subspecies described in Cicha et al. (1986) are not considered separately. Nevertheless, it might be worthwhile to mention that *U. semiornata semiornata* dominates the assemblage. *U. semiornata karreri*, a slender form almost completely lacking ornamentation was also encountered in the material.

Uvigerina aculeata d'Orbigny, 1846

Plate XII, figs 6–7

Uvigerina aculeata – d'Orbigny, 1846, For. Foss. Vienne, p. 191, Plate 11, figs 27–28

Uvigerina asperula (Czjek) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 177, pl. 2, fig. 19

Uvigerina aculeata d'Orbigny – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate CXLV, figs 1–4

Uvigerina aculeata d'Orbigny – Cicha et al., 1986: Oligocene and Miocene Uvigerina from the Western and Central Paratethys. In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 159, plate 14, figs 1–7

The illustration of *U. asperula* given in Korecz-Laky (1969) is well within the variation observed in *U. aculeata*, thus here the two are considered synonyms.

Uvigerina romaniaca Papp & Schmid, 1978

Plate XII, fig. 5

Uvigerina romaniaca – Papp & Schmid, 1978: Die Entwicklung der Uvigerinen im Badenien der Zentralen Paratethys. Chronostratigraphie und Neostratotypen, 6: M 4, Badenian, p. 283, plate 11, figs 15–17

Uvigerina romaniaca Papp & Schmid – Cicha et al., 1986: Oligocene and Miocene Uvigerina from the Western and Central Paratethys. In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, p. 164, plate 16, figs 1–4

This species differs from *U. aculeata* by its weaker ornamentation, although a continuous variation exists between these two species.

Uvigerina bononiensis Fornasini, 1888

Plate XII, figs 3–4

Uvigerina bononiensis – Fornasini, 1888, Tavola Paleo-protistografica Soc. Geol. Ital., Boll., Roma, vol. 7, fasc. 1, p. 48, plate 3, figs 12–12a

Hopkinsina bononiensis Fornasini – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 178, plate 4, fig. 7

Uvigerina szaklensis Majzon – Korecz-Laky & Nagy-Gellai, 1985, A Börzsöny-hegység oligocén és miocén képződményeinek foraminifera faunája (Foraminiferal fauna from the Oligocene and Miocene in the Börzsöny Mountains). Ann. Inst. Geol. Publ. Hung. vol. LXVIII, plate CXLIX, figs 1–4

Uvigerina bononiensis Fornasini – Cicha et al., 1986: Oligocene and Miocene Uvigerina from the Western and Central Paratethys. In: van der Zwaan, et al. (eds) 1986: Atlantic-European Oligocene to Recent Uvigerina. Utrecht. Micropal. Bull. vol. 35, pp. 175–178, plate 20, figs 9–11, plate 21, figs 1–3

Small-sized, slightly compressed specimens, with reduced seriality. This is one of the very few species found only in one of the cores (Tekeres-1), but almost completely missing from the other (Tengelic-2).

Spiroplectammina carinata (d'Orbigny, 1826)

Plate XIII, figs 1–5

Textularia carinata – d'Orbigny, 1826, Tabl. méth. Ceph., Ann. Sci. Nat., sér. 1, vol. 7, p. 269 Fornasini, 1902, Mem. R. Acad. Sci. Ist., ser. 5, vol. 7, p. 263, fig. 2

Textularia carinata – d'Orbigny, 1846, For. Foss. Vienne, p. 247, Plate 14, figs 32–34

Spiroplectammina carinata (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2 , p. 35, pl. 6, figs 2 a–b

Spiroplectammina carinata (d'Orbigny) – Korecz-Laky, 1969, A keleti Mecsek hegység miocén foraminiferái. Ann. Inst. Geol. Publ. Hung. vol. LII, 1, p. 146, plate 3, fig. 16

Spiroplectammina carinata (d'Orbigny) – Korecz-Laky, 1982; A Tengelic 2. sz. fúrás foraminifera faunája. Ann. Inst. Geol. Hung. vol. LXV, p. 174, plate 9. fig. 2.
?*Spiroplectammina spinosa* (d'Orbigny) – Korecz-Laky, 1982; A Tengelic 2. sz. fúrás foraminifera faunája. Ann. Inst. Geol. Hung. vol. LXV, p. 174, plate 9. fig. 3
Spiroplectinella carinata (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 86, Nr. 176, Pl. 80, figs 1–4
Spiroplectammina carinata (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 67, Plate 36, fig. 2

This species shows an enormous range of morphologic variation, which ranges from slender elongated forms to short stout forms. The periphery of *Spiroplectammina* in our material ranges from subacute to serrately keeled. According to the genus definition in Loeblich and Tappan (1988), the margin is supposedly rounded, while Spiroplectinella might possess a marginal keel. As the margin of the test was found to be a highly variable characteristic, we do not accept differentiation based on it. Occasionally, the chambers can end in a spine-like fashion, making the outline of the test serrate, as in the case of *Spiroplectammina spinosa,* which we found a synonym of *S. carinata.*

Valvulineria complanata (d'Orbigny, 1846)

Plate XIII, figs 6–7

Rosalina complanata – d'Orbigny, 1846, For. Foss. Vienne, p. 175, plate 10, figs 13–15
Valvulineria complanata (d'Orbigny) – Marks, 1951; Cushman Found. for Foram. Res. vol. 11, part 2, p. 64, pl. 6, figs 13 a–c
Valvulineria complanata (d'Orbigny) – Papp & Schmid, 1985; Die Foss. Foram. des Tertiären Beckens von Wien p. 66, Nr. 126, Pl. 59, figs 7–11
Valvulineria complanata (d'Orbigny) – Rupp, 1986; Paläoökologie der Foram. in der Sandschalerzone ... Beitr. Paläont. Österr. vol. 12, Wien, p. 69, Plate 41, figs 1–6

We observed the same variation as described by Rupp (1986) from the Vienna Basin. Generally, sutures on the spiral side are flush, while small-sized specimens tend to have thick, elevated spiral sutures. This suggests that the variation is most likely ontogenetical in nature.

The species is most common in samples with much pyrite, probably indicating organic-rich sediments, a preferred habitat of *V. complanata.*

Conclusion

The list of synonyms given for the thirty-eight most frequent (5%) species revealed thirty-four synonymous names, excluding designations of new genera. These synonyms are commonly used for the same species in different parts of the Paratethys.

Intraspecific variability, where considered significant enough for mentioning here, has been observed in almost a quarter of the species. The ecophenotypic nature of such morphologic variability has been successfully demonstrated for one of the *Bulimina* species. The distribution of *B. elongata* morphotypes was found to be depth-controlled, where depth estimates were based on the planktic/benthic ratio.

Foraminifera discussed here are common in the work of Korecz-Laky, especially Korecz-Laky 1982, dealing with material from one of the cores examined here. In spite of this, not many species presented here can be found in her work. This is partly because Korecz-Laky used the full residue, giving large-sized foraminifera more attention, and partly because of differences in species concept.

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Plate I

1–2. *Ammonia beccarii* (Linné, 1758)
1. spiral view, × 148, Tekeres-1 71.25 m
2. umbilical view, × 145, Tekeres-1 71.25 m

3–5. *Astrononion* cf. *italicum* Cushman & Edwards, 1937
3. spiral view with clearly visible supplementary apertures, × 239, Tekeres-1 259 m
4. spiral view, with thickened sutures easily observable under the light microscope, and small supplementary apertures, × 259, Tekeres-1 259 m
5. apertural view, × 329, Tekeres-1 259 m

6–8. *Bolivina hebes* Macfadyen, 1930
6. × 178, Tekeres-1 204.2 m
7. side view, × 218, Tekeres-1 154 m
8. apertural view of same specimen as fig. 7, × 984, Tekeres-1 154 m

Plate II

1. *Bolivina plicatella* Cushman, 1930, × 384, Tengelic-2 759.9 m
2. *Bolivina scalprata* (Schwager) var. *miocenica* Macfadyen, 1930, × 298, Tekeres-1 47 m

3–5. *Bolivina spathulata* (Williamson, 1858)
3. *B. spathulata f. dilatata*, × 274, Tekeres-1 273.1 m
4. transitional form between the two morphotypes, × 174, Tekeres-1 273.1 m
5. *B. spathulata f. spathulata*, × 174, Tekeres-1 166.4 m

6–7. *Bolivina viennensis* Marks, 1951
833.3 m
7. side view, × 350, Tengelic-2 833.3 m
8. *Bulimina costata* d'Orbigny, 1826; × 230, Tengelic-2 759.9 m

Plate III

1–6. *Bulimina elongata* d'Orbigny, 1846
1. *Bulimina elongata f. subulata*, × 270, Tengelic-2 833.3 m
2. Transitional form between *Bulimina elongata* f. *subulata* and *Bulimina elongata* f. *aculeata*, × 252, Tengelic-2 759.9 m
3. *Bulimina elongata* f. *aculeata*, × 164, Tengelic-2 833.3 m
4. *Bulimina elongata* var. *minima*, × 333, Tekeres-1 211 m
5. Transitional form between *Bulimina elongata* var. *minima* and *Bulimina elongata* f. *elongata*, × 298, Tengelic-2 759.9 m
6. *Bulimina elongata* f. *elongata*, × 248, Tengelic-2 759.9 m

7–8. *Cassidulina laevigata* d'Orbigny, 1826
7. × 299, Tekeres-1 201.75 m
8. side view, × 334 Tekeres-1 201.75 m

Plate IV

1–4. *Bulimina pyrula* d'Orbigny, 1846
1. *B. pyrula* f. *ovata*, × 165, Tekeres-1 211 m
2. *B. pyrula* f. *pupoides*, × 177, Tengelic-2 738.5 m
3. *B. pyrula* f. *pyrula*, × 181, Tengelic-2 738.5m
4. high magnification picture of *B. pyrula* showing elongated pores, × 1380, Tengelic-2 764.5 m

5–7. *Cassidulina crassa* d'Orbigny, 1839
5. × 187, Tengelic-2 812.8 m
6. × 150, Tengelic-2 812.8 m
7. apertural view × 186, Tengelic-2 812.8 m

Plate V.

1. *Cassidulina oblonga* Reuss, 1850; × 236, Tengelic-2 764.5 m

2–3. *Cassidulina subglobosa* Brady, 1881
2. × 330, Tekeres-1 201.75 m
3. × 332 Tekeres-1 201.75 m
4. *Cassidulina teretis* Tappan, 1951; × 246, Tengelic-2 759.9 m

5–8. *C. lobatulus* f. *lobatulus* (Walker & Jacob), 1798
5. spiral view of an extremely irregular specimen, × 174, Tengelic-2 833.3 m
6. side view, × 171, Tekeres-1 47 m
7. spiral view, × 218, Tekeres-1 259 m
8. umbilical view, × 182, Tekeres-1 259 m

Plate VI

1–3. *C. lobatulus* f. *refulgens* (Monfort, 1808)
1. spiral view, × 224, Tekeres-1 259 m
2. umbilical view, × 257, Tekeres-1 146.75 m
3. side view, × 244, Tekeres-1 146.75 m

4–6. *C. lobatulus* f. *letkesiensis* (Franzenau, 1894)
4. spiral view, × 332, Tekeres-1 259 m
5. umbilical view, × 332, Tekeres-1 259 m
6. side view, × 332, Tekeres-1 259 m
7. *Dentalina elegans* d'Orbigny, 1846, × 55, Tengelic-2 844 m

8–10. *Cibicides ungerianus* (d'Orbigny, 1846)
8. spiral view, × 126, Tekeres-1 251 m
9. umbilical view, × 138, Tekeres-1 251 m
10. side view, × 175, Tekeres-1 251 m

Plate VII

1–2. *Fursenkoina acuta* (d'Orbigny, 1846)
  1. 759.9 m
  2. × 122, Tengelic-2 759.9 m

3–4. *Gavelinopsis praegeri* (Heron-Allen & Earland, 1913)
  3. spiral view, × 185, Tengelic-2 738.5 m
  4. umbilical view, × 212, Tengelic-2 738.5 m

5–6. *Gyroidina soldanii* d'Orbigny, 1825
  5. spiral view, × 154, Tengelic-2 777.8 m
  6. side view, × 216. Tengelic-2 789.5 m

7–9. *Gyroidina umbonata* (Silvestri, 1898)
  7. spiral view, × 334, Tekeres-1 259 m
  8. umbilical view, × 334, Tekeres-1 259 m
  9. side view, × 334, Tekeres-1 259 m

Plate VIII

1–3. *Epistominella smithi* (R. E. & K. C. Stewart, 1930)
  1. spiral view, × 196, Tengelic-2 759.9 m
  2. umbilical view, × 202, Tengelic-2 759.9 m
  3. side view, × 254, Tekeres-1 211 m

4–6. *Hanzawaia boueana* (d'Orbigny, 1846)
  4. spiral view, × 162, Tengelic-2 759.9 m
  5. umbilical view, × 184, Tengelic-2 759.9 m
  6. side view, × 264, Tekeres-1 47 m

7–8. *Hoeglundina elegans* (d'Orbigny, 1826)
  7. X160, Tengelic-2 759.9 m
  8. apertural view, × 256, Tengelic-2 759.9 m

Plate IX

1–3. *Heterolepa dutemplei* (d'Orbigny, 1846)
  1. spiral view, × 116, Tengelic-2 838.3 m
  2. umbilical view, × 90.5, Tengelic-2 838.3 m
  3. side view, × 145, Tengelic-2 838.3 m

4–6. *H. dutemplei*, deep water morphotype, probabily transitional form to *C. kullenbergi*
  4. spiral view, × 163, Tengelic-2 838.3 m
  5. umbilical view, × 216, Tengelic-2 838.3 m
  6. side view, × 222 Tengelic-2 838.3 m

Plate X

1–2. *H. dutemplei*, shallow water morphotype
  1. spiral view, × 157, Tengelic-2 838.3 m
  2. umbilical view, × 110, Tengelic-2 838.3 m

3. *Siphonina reticulata* Czjek, 1848, × 174, Tengelic-2 833.3 m

4–5. *Nonion commune* (d'Orbigny, 1825)
  4. spiral view, × 128, Tengelic-2 738.5 m
  5. apertural view, × 260, Tengelic-2 777.8 m

6–7. *Trifarina angulosa* (Williamson, 1858)
  6. × 222, Tekeres-1 204.2 m
  7. apertural view, × 444, Tekeres-1 204.2 m

Plate XI

1–2. *Pullenia bulloides* (d'Orbigny, 1825)
1. spiral view, × 188, Tengelic-2 738.5 m
2. apertural view, × 240, Tengelic-2 738.5 m

3–4. *Oridorsalis umbonatus* (Reuss, 1851)
3. spiral view, × 153, Tekeres-1 251 m
4. umbilical view, × 156, Tekeres-1 251 m

5–6. *Uvigerina acuminata* Hosius, 1893
5. × 160, Tengelic-2 838.3 m
6. × 183, Tekeres-1 134 m

Plate XII.

1–2. *Uvigerina semiornata* d'Orbigny, 1846
1. × 159, Tengelic-2 759.9 m
2. × 162, Tengelic-2 749 m

3–4. *Uvigerina bononiensis* Fornasini, 1888
3. × 201, Tekeres-1 166 m
4. side view, × 229, Tekeres-1 166 m

5. *Uvigerina romaniaca* Papp & Schmid, 1978, × 166, Tekeres-1 111.5 m

6–7. *Uvigerina aculeata* d'Orbigny, 1846
6. × 193, Tengelic-2 807 m
7. × 136, Tengelic-2 807 m

Plate XIII

1–5. *Spiroplectammina carinata* (d'Orbigny, 1826)
1. × 105, Tengelic-2 838.3 m
2. × 96.5, Tengelic-2 838.3 m
3. × 77, Tengelic-2 838.3 m
4. × 92, Tengelic-2 838.3 m
5. × 74.5, Tengelic-2 738.5 m

6–7. *Valvulineria complanata* (d'Orbigny, 1846)
6. spiral view, × 234, Tengelic-2 738.5 m
7. umbilical view, × 150, Tengelic-2 738.5 m

Plate I

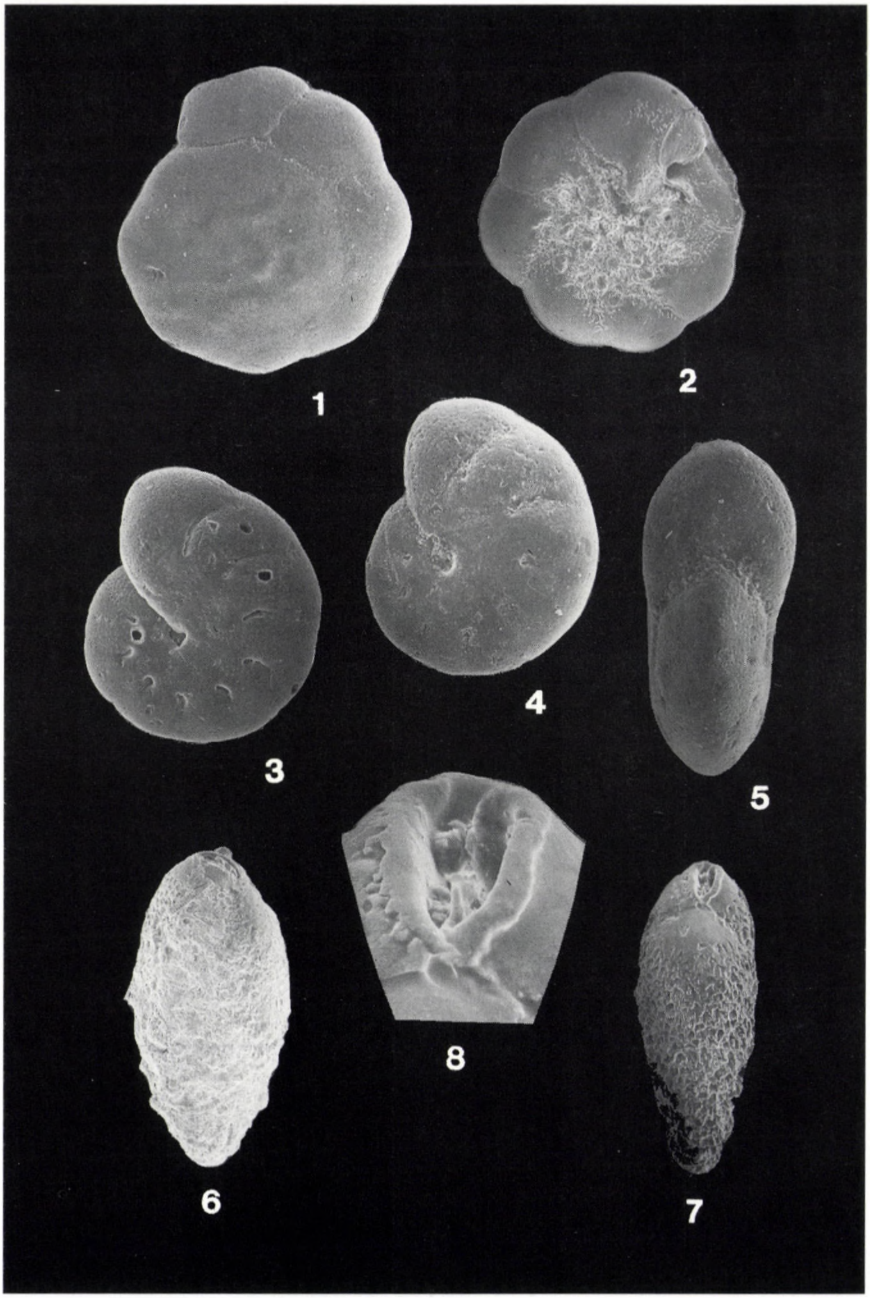


Plate II

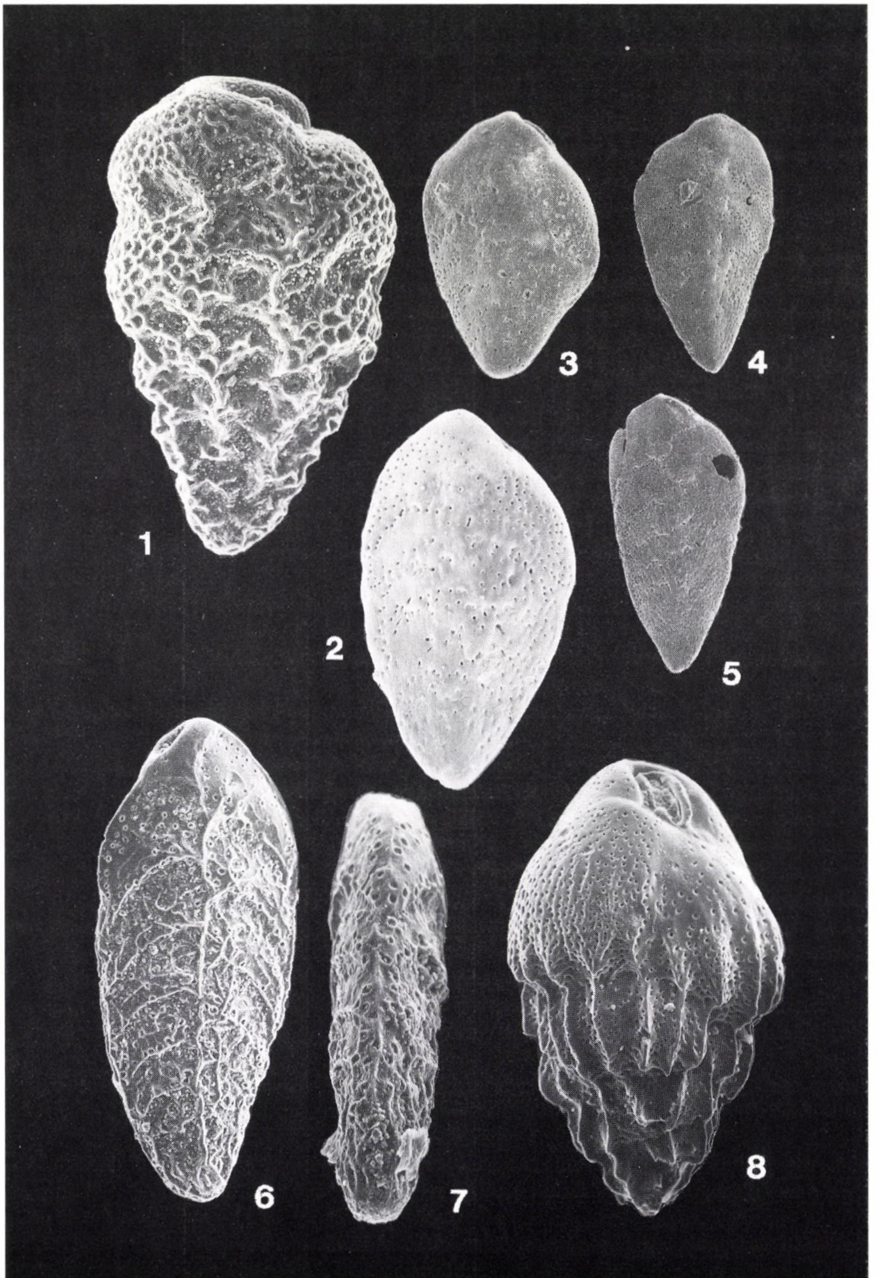


Plate III

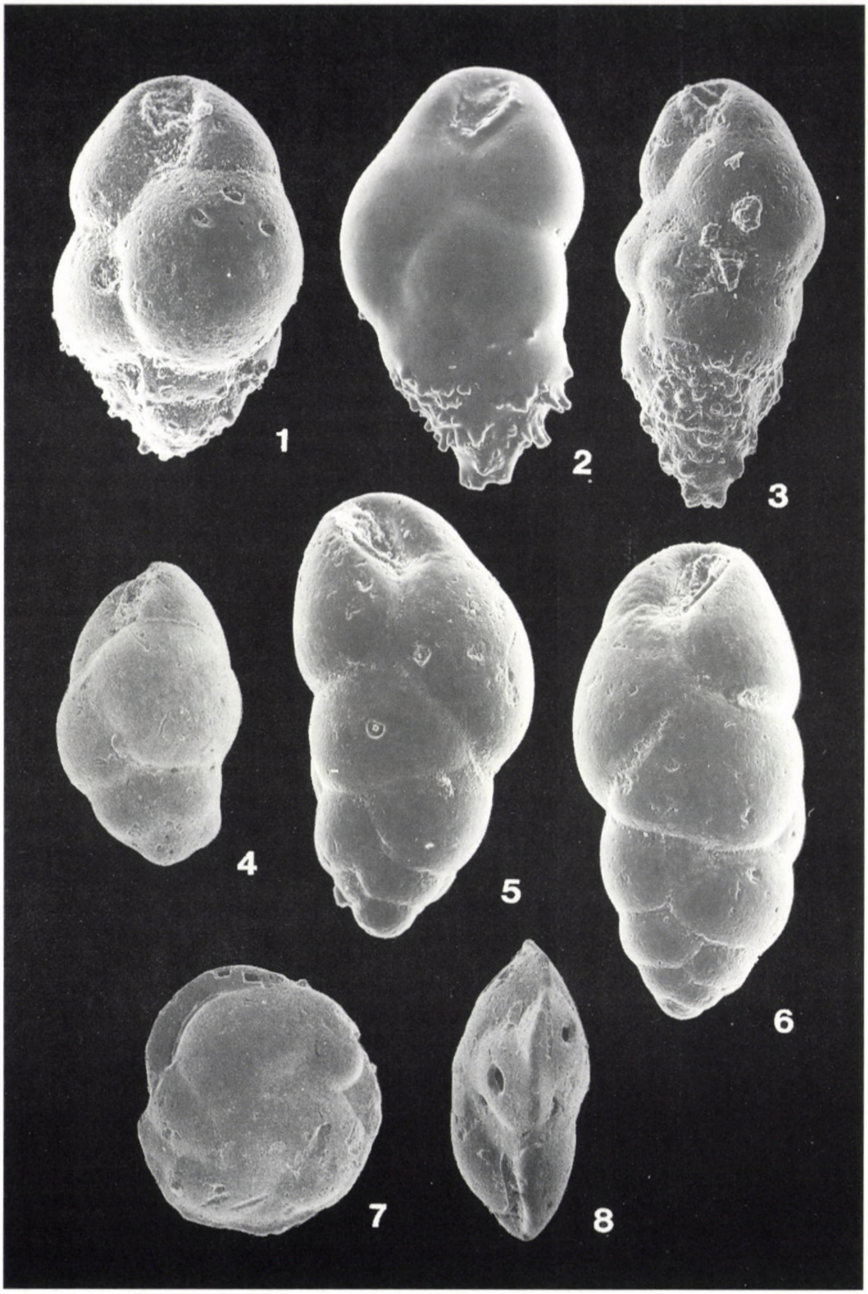
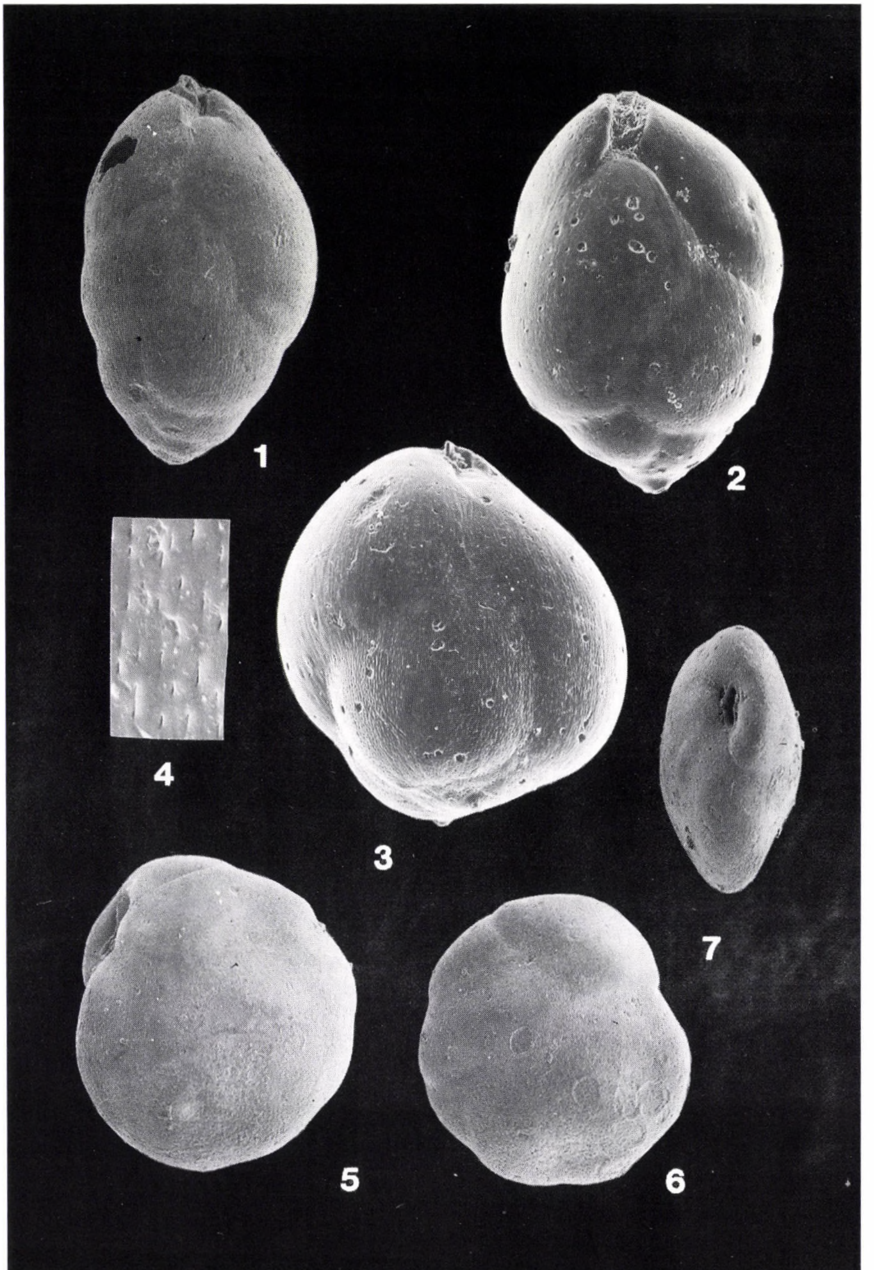


Plate IV



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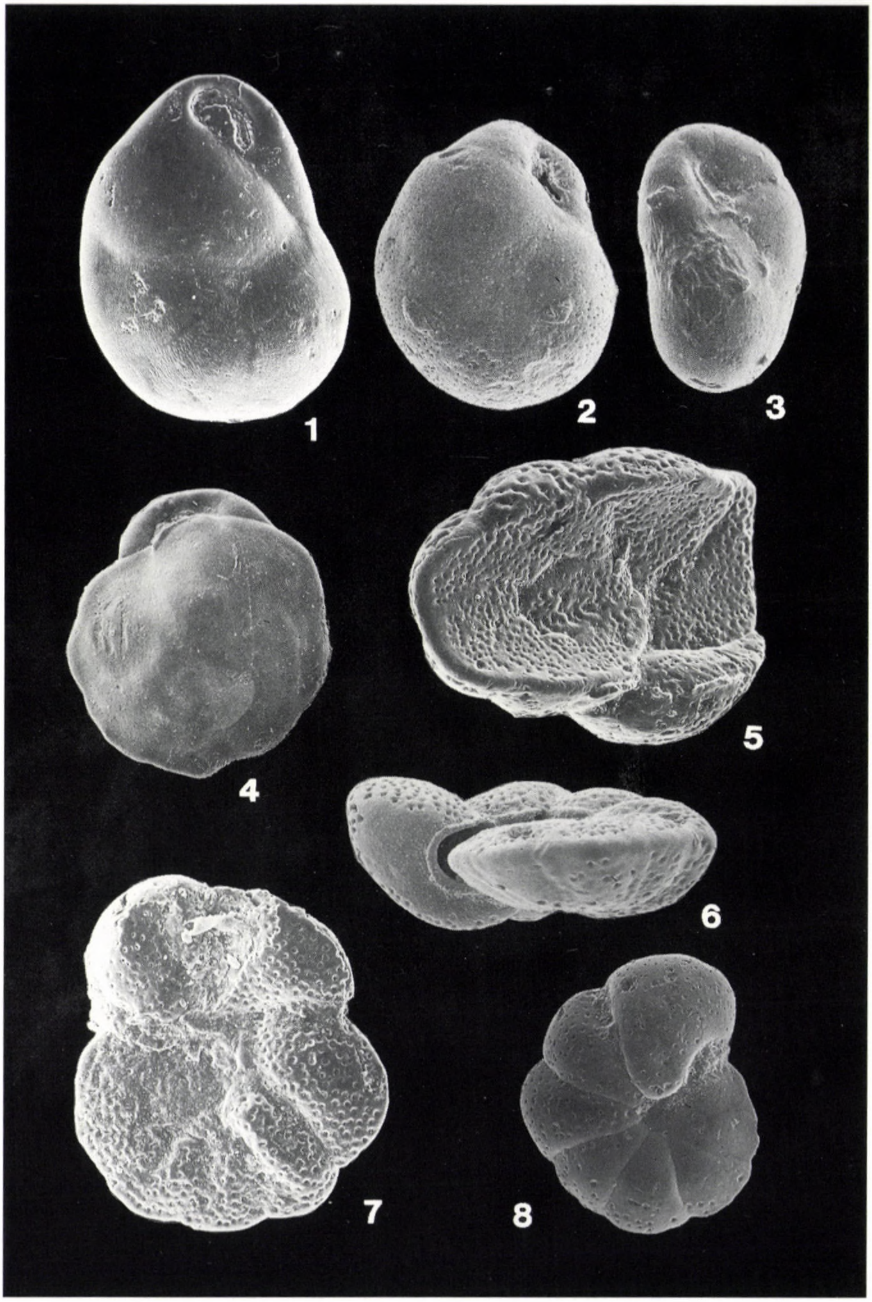

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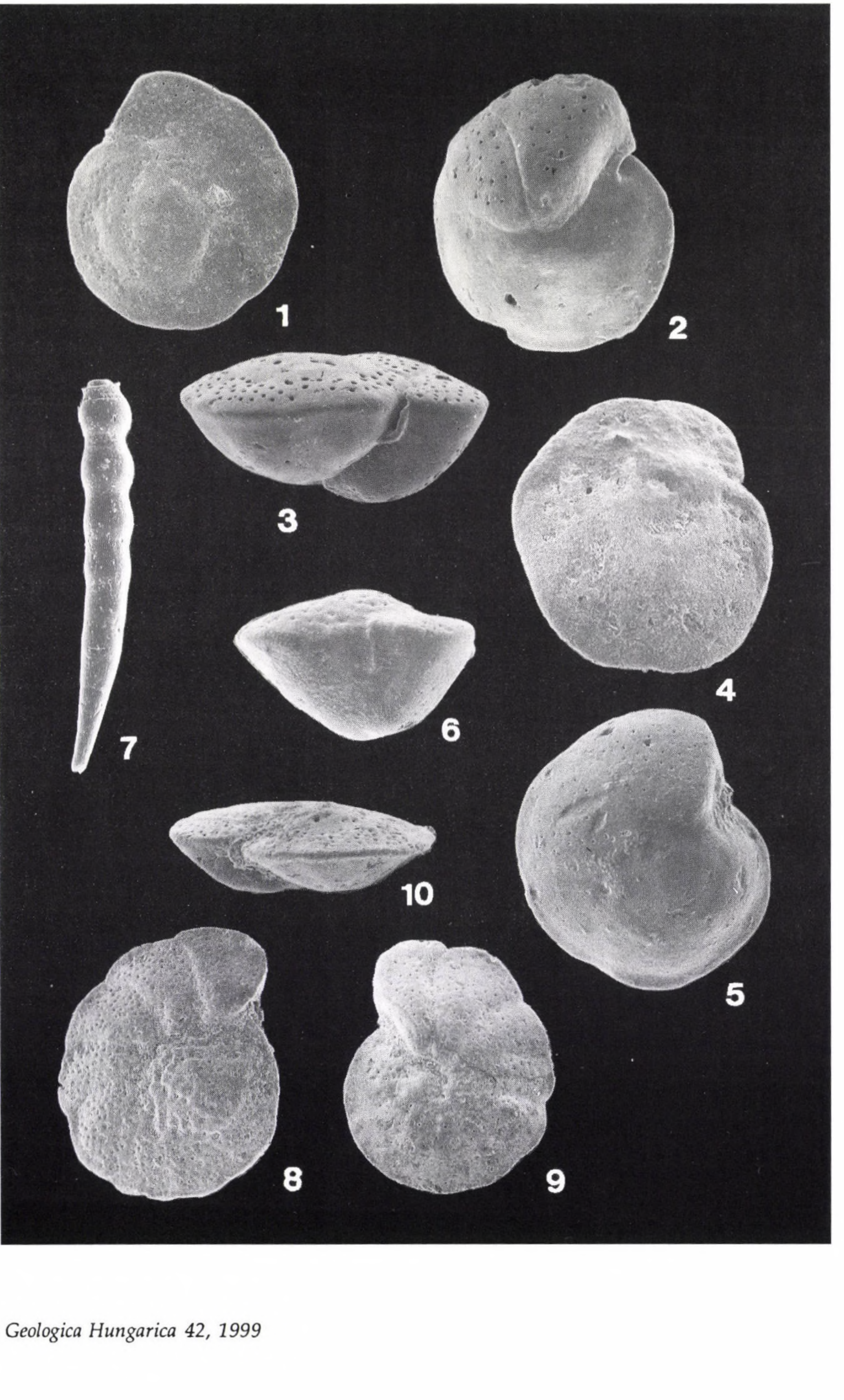

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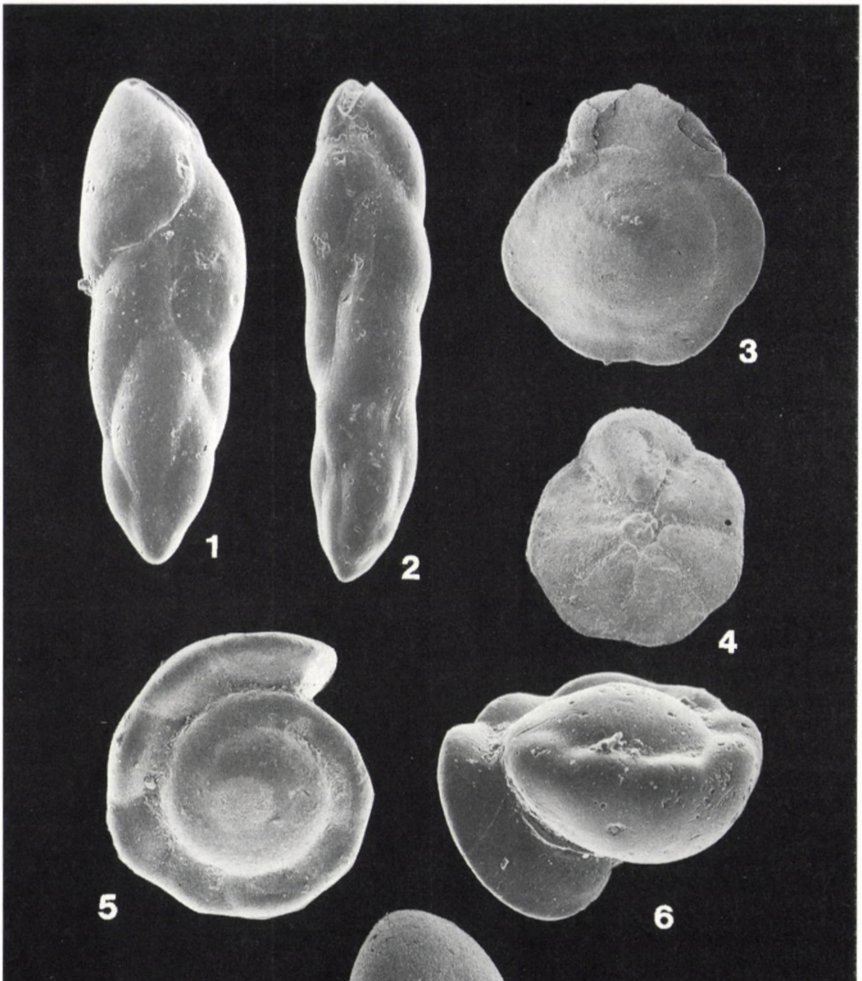

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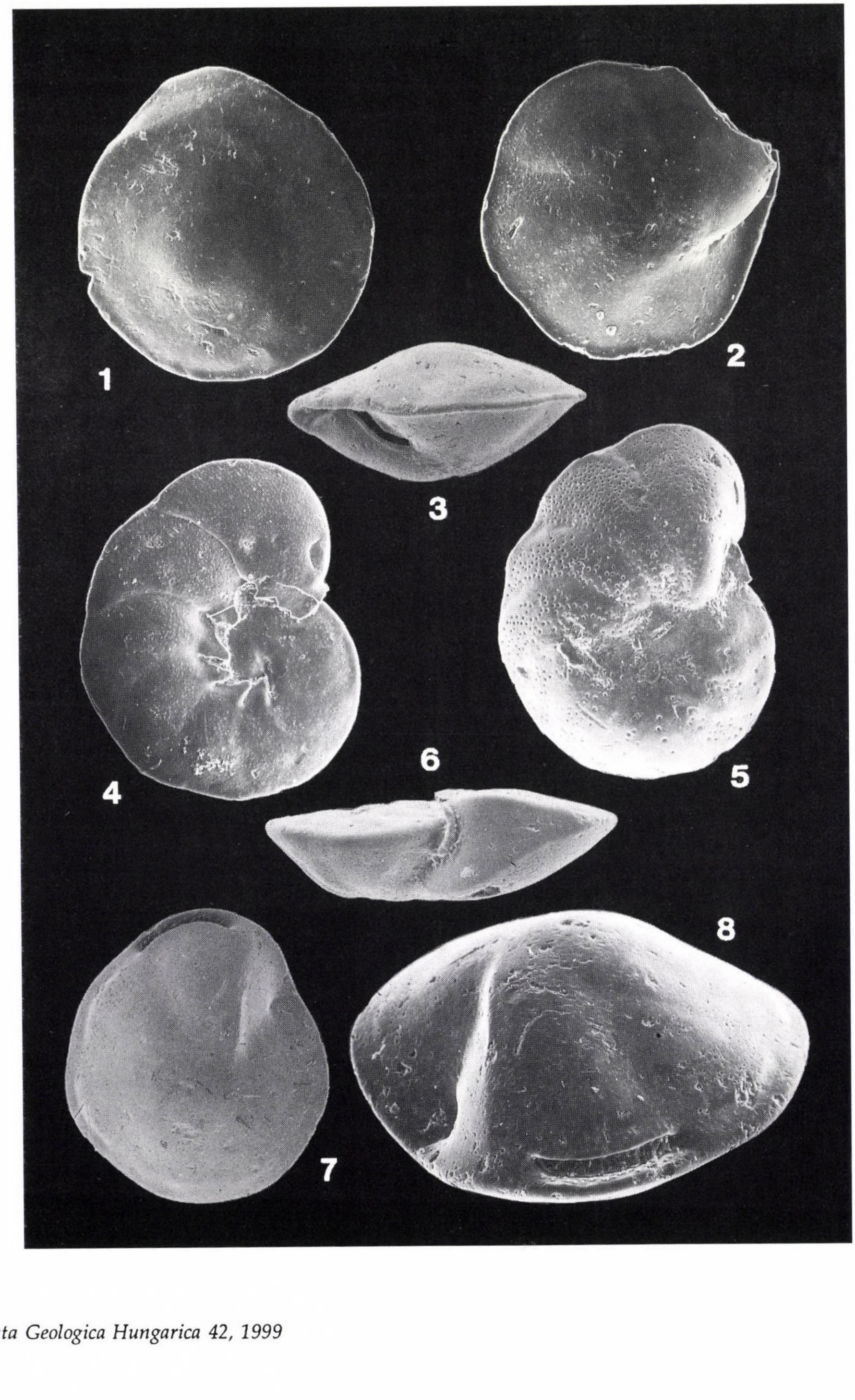

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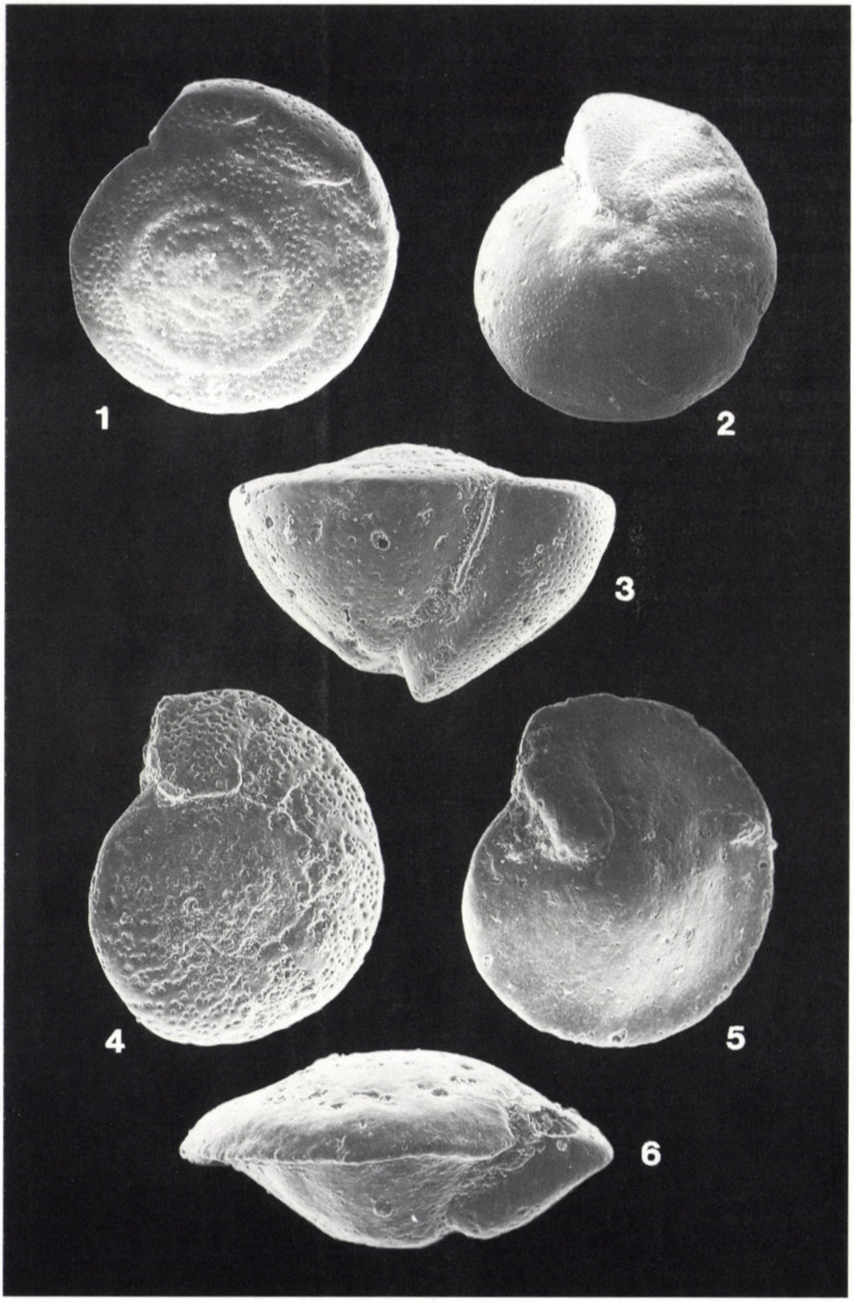

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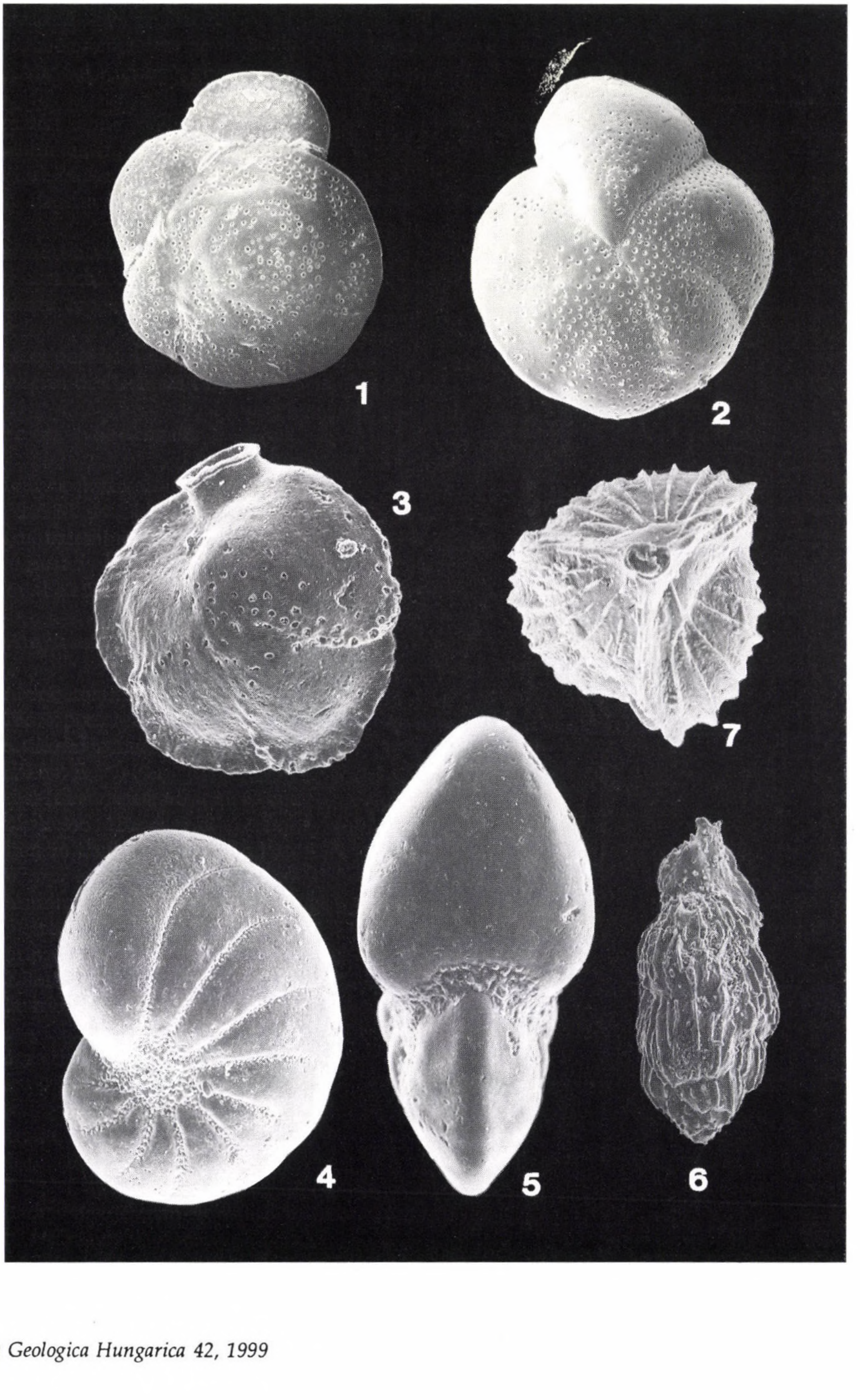

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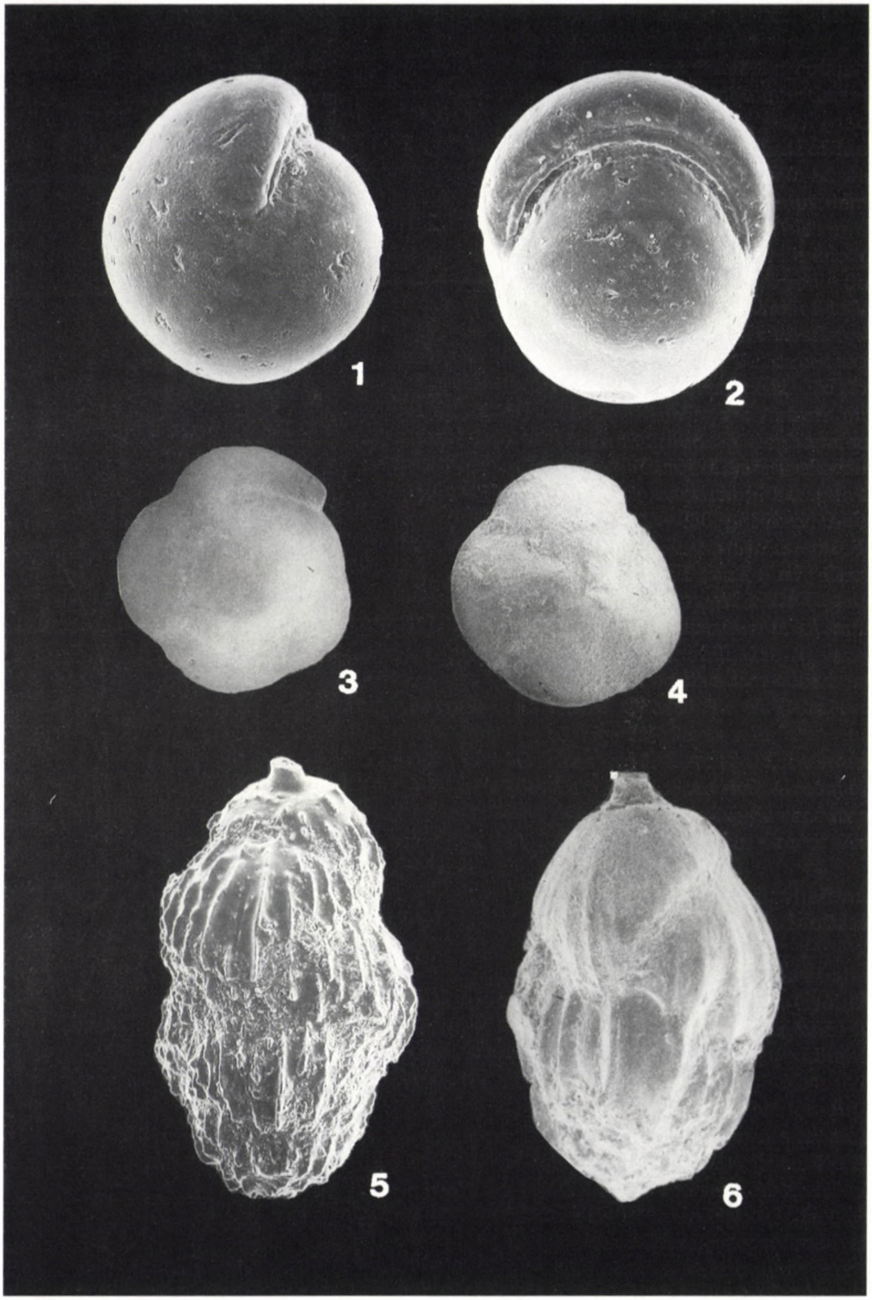

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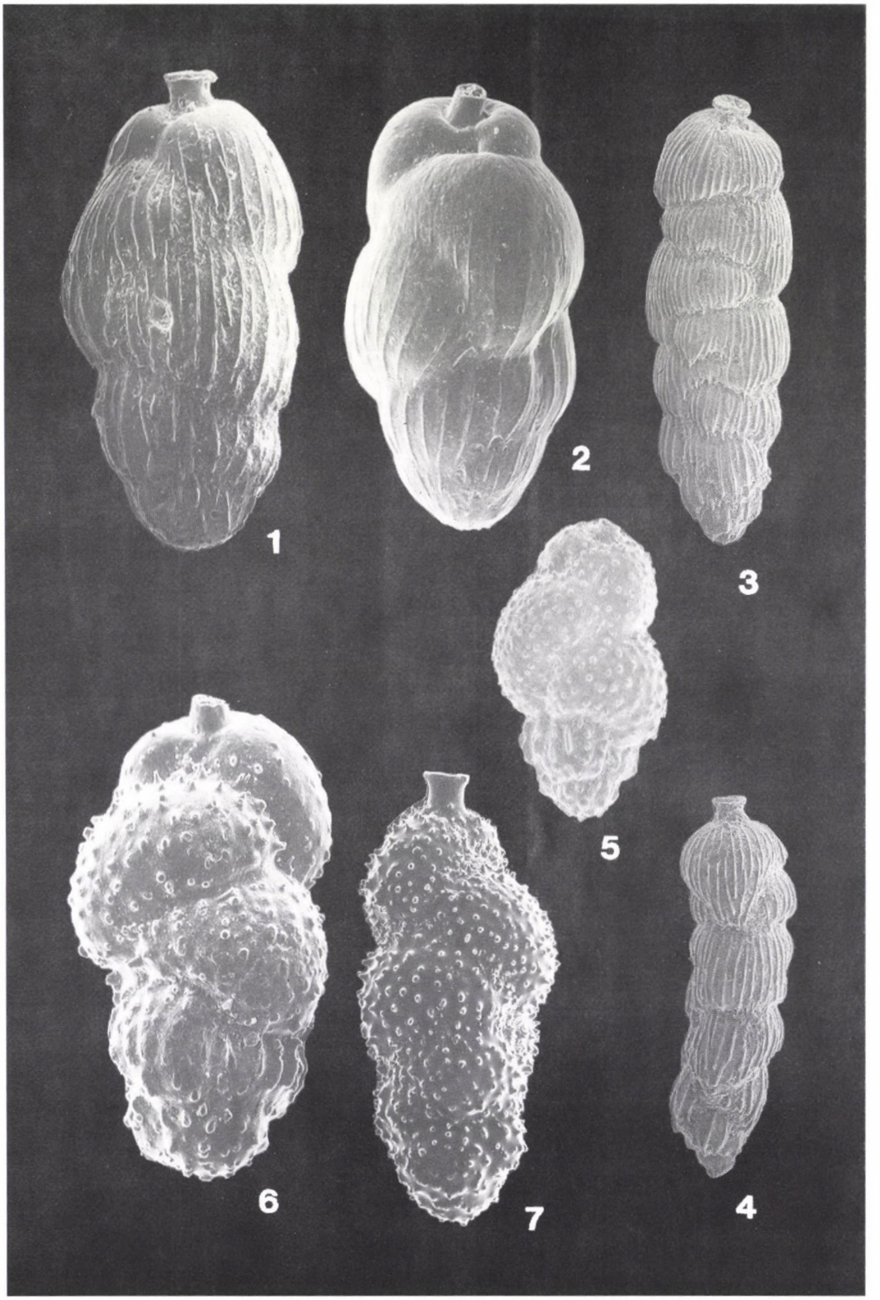

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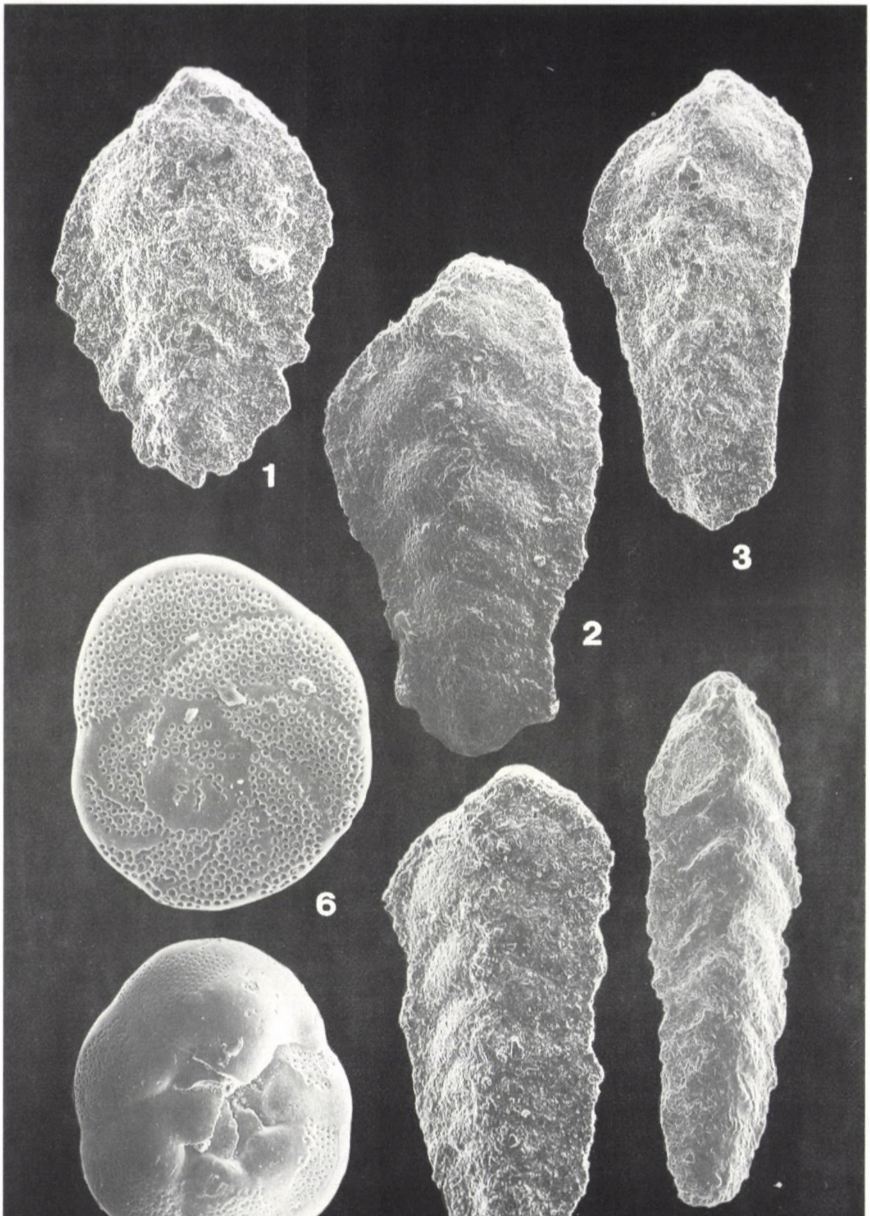

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Karst collapse phenomena in the Upper Miocene of Mallorca (Balearic Islands, Western Mediterranean)

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The tabular carbonate deposits that outcrop in the eastern (Llevant) and southern (Migjorn) lowlands of the island of Mallorca (the so-called Marines) are Upper Miocene in age. They were not affected by the Alpine orogeny and their deposits consist mainly of littoral and inner platform carbonate facies with scarcely any terrigenous input. Within this Upper Miocene sequence three depositional units are traditionally described: the Calcisiltite Unit with Heterostegina (corresponding to an open platform), the Reef Unit representing the progradation of a coral reef platform and finally the Santanyi Limestones characterized by shallow marine carbonate platform deposits with sand bars, mangrove swamps and stromatolitic facies.

The contact between the Reef Unit and the Santanyi Limestone in the southeastern part of Mallorca shows paleokarst features which include dissolution and collapse structures with breccia formation. Several types of collapse structures have been recognized and different types of karst-brecciation related to them according to the different stages of the karstification processes. Breakdown processes on previously lithified basal Santanyi Limestone beds and V-shaped plastic deformation in their upper levels are the result of karstic voids and the subsidence developed in the underlying Reef Unit. Different stages of development and the spatial distribution of these paleokarst features can be related mainly to the facies distribution of the Reef Unit. Breccia formation processes are involved in the cavity collapses and stress the great diversity of breccia types occurring in karstic environments.

Key words: karst collapse, breccia formation, karst infillings, Upper Miocene, Mallorca

Introduction

The morphological and morphodynamic aspects of the study of the paleokarst cannot be assessed separately, as they should be framed within the geological context in which they once developed. Within this context the stratigraphical, sedimentological, petrologic, tectonic and hydrogeologic aspects are remarkable.

The lithology and structural setting of the Upper Miocene deposits which form part of the island of Mallorca are magnificent bases on which paleokarstic phenomena can be observed. Mainly limestones and locally dolomites built up a kind of landscape locally known as marines at the eastern and southern coasts of the island. The geological processes that affected the western Mediterranean during the Upper Miocene led to events of emersion of these carbonates, giving

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way to the development of important paleokarstic phenomena. Examples of paleokarst phenomena are visible in the uppermost Miocene deposits in Mallorca.

In spite of the growing importance of paleokarst studies in the last decade (James and Choquette 1987; Bosak et al. 1989; Esteban and Klappa 1983), those on the Majorcan paleokarst are scarce considering both the abundant occurrences of the paleokarst horizons in the stratigraphic record of the island and the good exposures. There are only isolated studies dealing with regional problems and morphological aspects and geometries of the deposits, both of the paleokarst phenomena of the Jurassic limestones (Fornós et al. 1986/87; Fornós et al. 1995) and the Miocene reef deposits (Fornós et al. 1989, 1995). In the present paper we summarize our previous studies and discuss the stratigraphical, sedimentological, petrologic, morphological, and genetic aspects of the paleokarst processes which affected the Upper Miocene reef deposits (Fornós et al. 1989).

Geologic setting

From a geologic point of view (Fornós and Gelabert 1995) the island of Mallorca (Fig. 1) shows a complex tectonic pattern framed within the context of the plate tectonics of the Western Mediterranean (an area located between the European and the African plates) which resulted in a strong orography.

The geologic structure as well as the lithology strongly influence the geomorphology of the island. Three principal zones can be differentiated. The steepest and most important range, the Serra de Tramuntana or Serra Nord, is constituted by well-structured deposits of mainly Late Triassic to Jurassic age. It is located on the northwestern side of the island and strikes in a NE–SW direction, and is where the main peaks are located (e.g. Puig Major, 1445 m). It has a rough relief which delimits the island on its NW edge. The Serres de Llevant Mountains, on the opposite side (in the east), show a smoother topography despite the fact that they are also composed of Late Triassic to Jurassic deposits (being, however, more dolomitic) and show the same Alpine tectonic influence reflecting in a NNE–SSW alignment. The ranges are separated by the El Pla area and surrounded by the "marines" (Marina de Llucmajor and Marina de Llevant) formed by tabular postorogenic deposits of Neogene (mainly Upper Miocene) age.

Stratigraphical and sedimentological aspects: Upper Miocene stratigraphy

The geologic history of Mallorca encompasses an interval from the Carboniferous to the Quaternary with an important gap at the base of the Cenozoic. The most remarkable common feature of this formations is the predominance of the carbonates in their composition and the scarce amount of siliciclastics. The sedimentology of the outcropping formations is highly

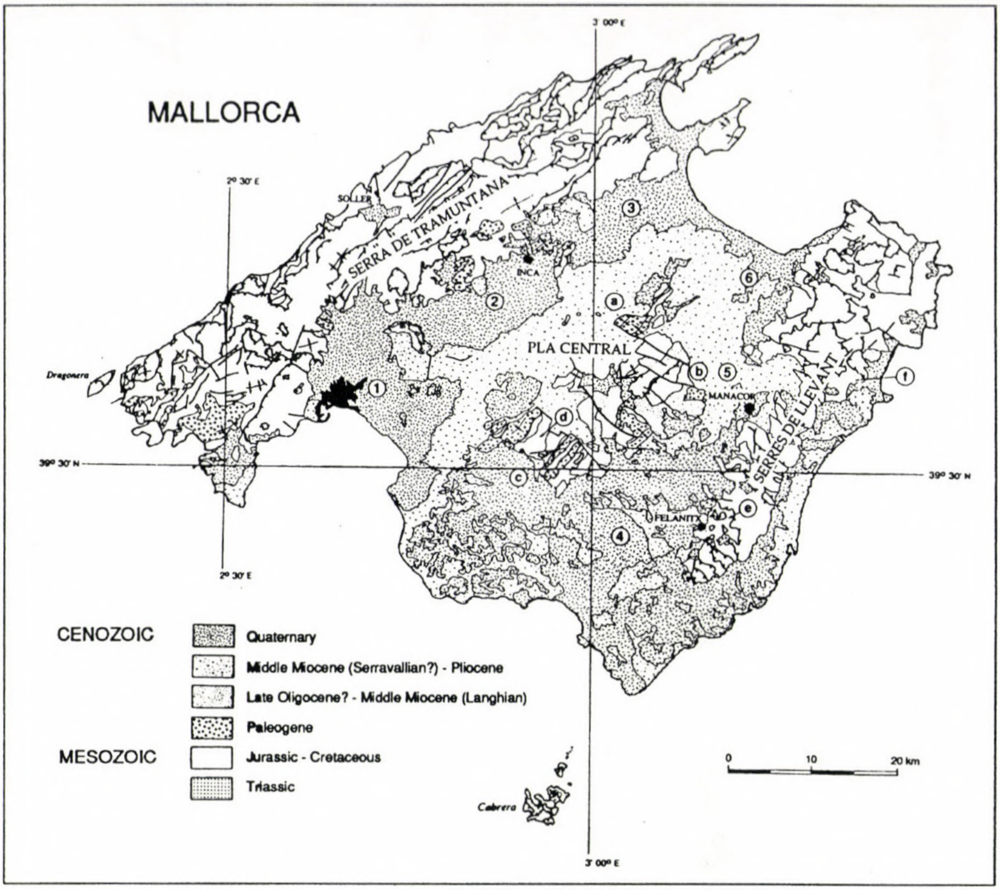

Fig. 1
Geologic map of the Island of Mallorca (Roca 1992). 1. Palma Basin; 2. Inca Basin; 3. Sa Pobla Basin; 4. Campos Basin; 5. Manacor Basin; 6. Sta. Margalida Basin; a – Sineu; b – Bonany; c – Massís de Randa; d – Puig de Randa; e – Son Macià-Sant Salvador syncline; f – Son Servera

complex and reflects the principal facies zones from littoral to the deep pelagic environments.

The Cenozoic is widely represented in Mallorca, having a thickness of more than 1500 m (Ramos-Guerrero et al 1989). Two main units can be distinguished: a pretectonic and syntectonic unit (Middle Eocene to Lower Miocene) and a post-tectonic one (from the Middle Miocene to the present). The Cenozoic deposits are located in the central lowlands of the island as well as the marines in the Migjorn region.

The most important limestone outcrops and water reservoirs of Mallorca are Upper Miocene and Liassic in age, respectively (Fig. 2). They are made up of limestones and calcarenites, and crop out widely as a series of tabular platforms surrounding the older formations (mostly Liassic limestones). The succession

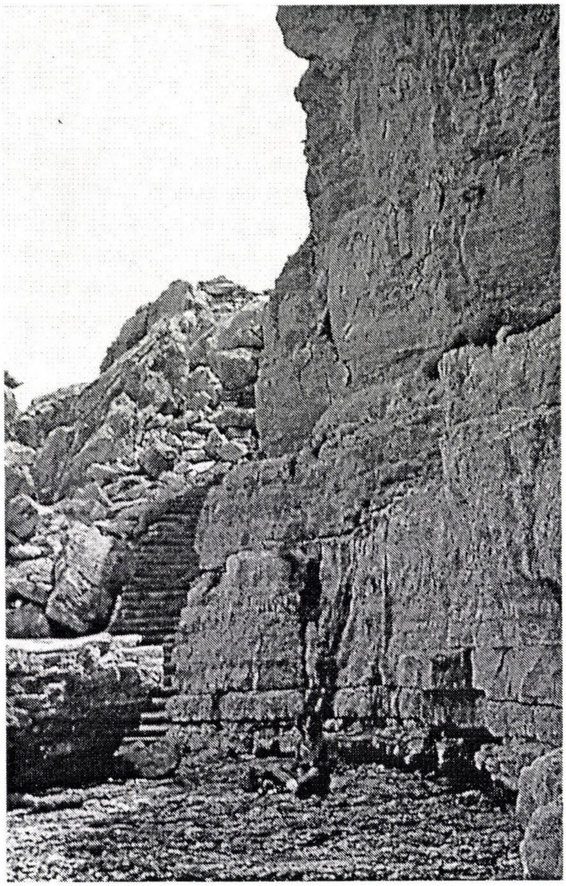

Fig. 2
Well-stratified Upper Miocene outcrop in the cliffs of the Marina de Llevant (southeastern part of Mallorca) showing the Santanyí Limestone sequence

consists of alternating calcarenites and calcisiltites at the base evolving into massive reefal limestones and calcarenites and ending with oolites and stromatolithic limestones. Such tabular outcrops occur all along the seacoast. The height of the fairly continuous littoral cliffs may exceed 30 m. The platforms which are made up of these post-orogenic deposits are furrowed by streams which, when reaching the coast line, form the typical littoral entrance morphologies known as "cales" (pocket bays). The continuity and size of the coastal outcrops allow accurate sedimentological and paleokarst studies. Nevertheless, advancing inland these possibilities decrease exponentially and dramatically.

As described by Esteban (1979/80) the Upper Miocene of Mallorca, as in many Mediterranean areas, consists of platform carbonate with well-developed reefs. Classically three depositional units have been defined in Mallorca (Pomar et al. 1983, 1985; Pomar 1991). They are clearly differentiated by their lithology and depositional environment (Fig. 3). At the base a unit named "Calcisiltites

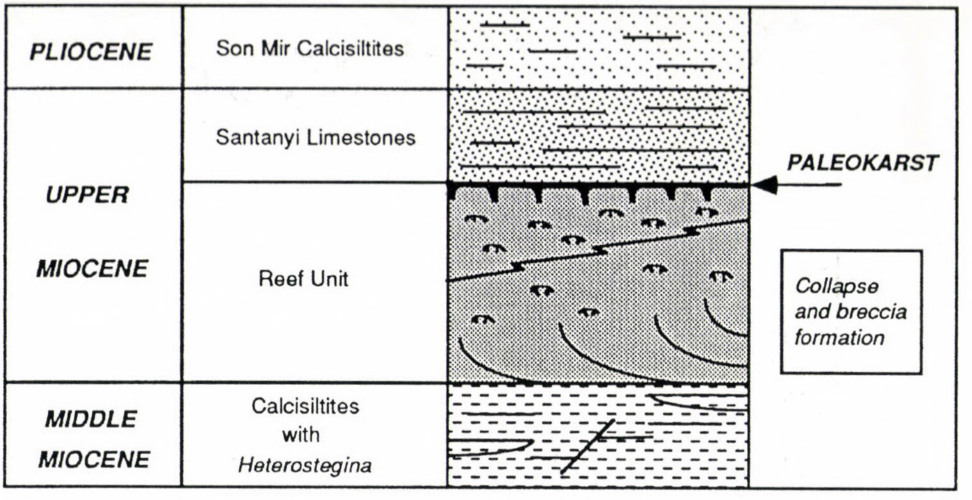

Fig. 3
Stratigraphic chart of the Upper Miocene from Mallorca showing paleokarst features

with *Heterostegina*" occurs which discordantly covers the pretectonic and syntectonic deposits. This unit consists of calcarenites and calcisiltites and contains an abundant foraminifer fauna (mainly the characteristic *Heterostegina*) and echinoids. It shows strong bioturbation and corresponds to an open carbonate platform facies. This unit is overlain slightly unconformably by the "Reef Unit" (Pomar 1991). It is composed of bioclastic calcarenites and coral reefs interlayered with levels of oolitic limestones and stromatolites. It corresponds to the development of a coral reef of barrier type with wide lagoons. An erosive surface, partially karstified is visible at the top of the reefal unit. In the southern and eastern zones of Mallorca the Reef Unit is covered by the "Santanyí Limestone Unit" the so-called Terminal Complex (Fornós and Pomar 1984). It is made up of carbonate deposits of a littoral type involving a great variety of facies including oolitic limestones (mangroves, sand shoals, etc.). Stromatolitic limestones occur in the uppermost levels which indicate a shallowing-upward trend with an important regressive event (Fig. 4). The top of the unit appears to be truncated by a relevant erosion surface covered by deposits of Pliocene age that scarcely outcrop in the marines, and by the Plio-Quaternary and Quaternary which are basically calcarenitic with dune episodes related to Pleistocene glacial events (Butzer and Cuerda 1962; Cuerda 1975).

The geometry of karstic phenomena

Features related to subsidence phenomena caused by karstic dissolution processes are frequently found on the marines in the Migjorn region (southern

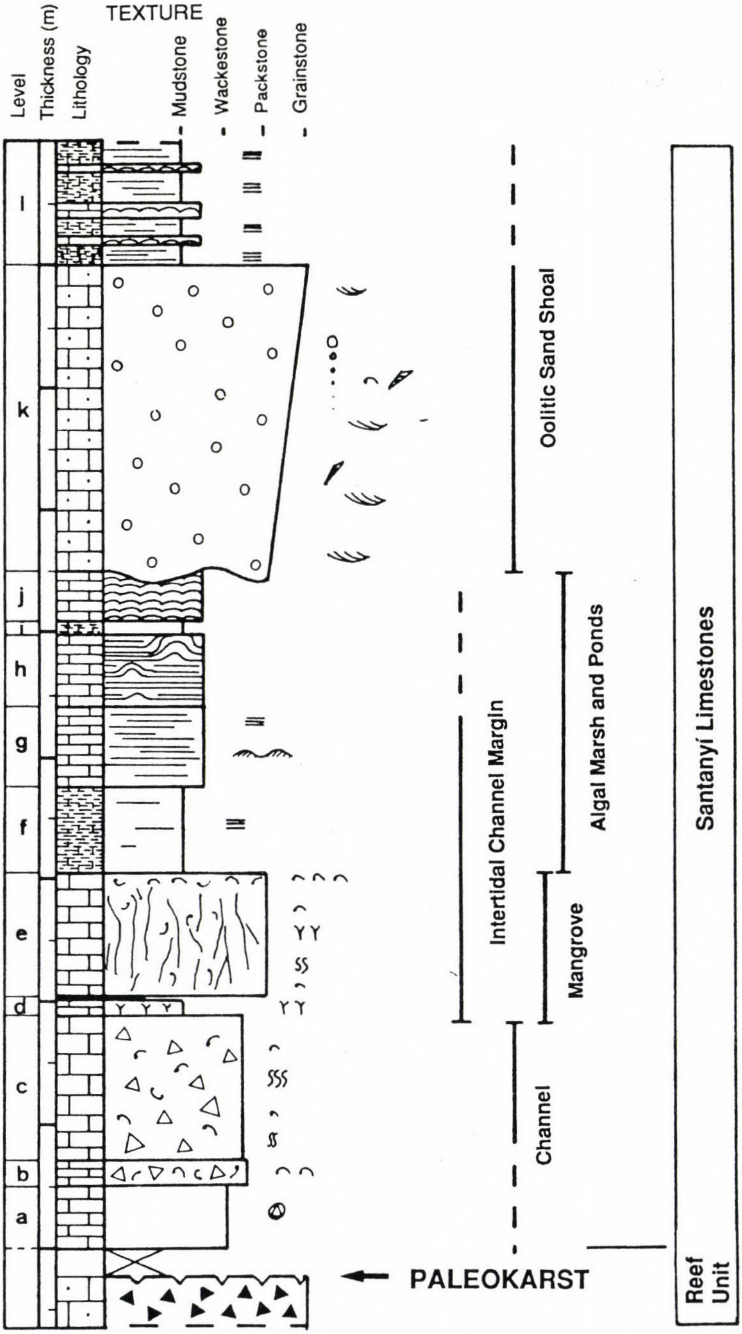

Fig. 4
Stratigraphical series and sedimentological interpretation of the Santanyí Limestone sequence near Port de Manacor

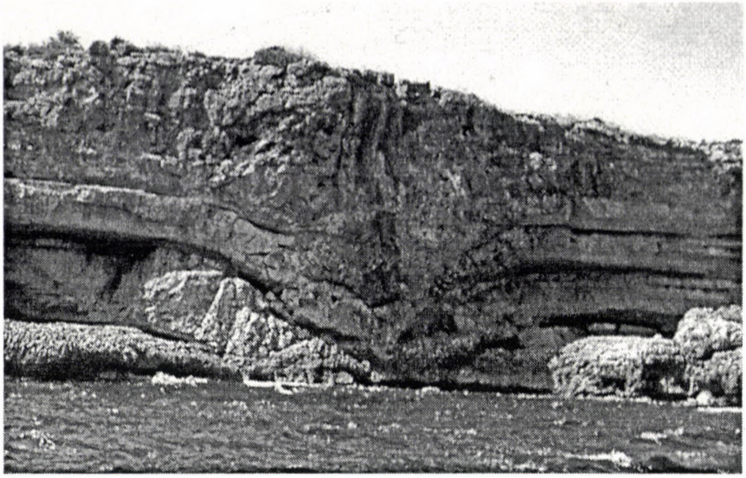

Fig. 5
Limestones from the Upper Miocene outcrop all along the Migjorn region in Mallorca, where collapsing karstic phenomena took place. The basal materials at sea level correspond to the Reef Unit whereas the well-stratified levels correspond to the Santanyí Limestone Unit

and southeastern Mallorca), especially near the cliffs surrounding Porto Colom and Santanyí (Fig. 5). They can be followed continuously from Cala Llombards (South of Santanyí) to the North of S'Algar, in Portocolom. A series of nearly continuous deformations follow the contact between the Reef Unit and the Santanyí Limestones. Collapse deformations can often affect the entire assemblage of Santanyí Limestones. These deformations are mostly located at the lower levels (the more plastic ones) corresponding to the mangrove-related sediments (Fornós and Pomar 1982) but they also affect the upper levels interpreted as oolite sand shoal. The Reef Unit has only been affected in its upper levels.

The general morphology of the collapse structures resembles that of an hourglass but with an asymmetric shape, as the upper cavity is wider and smoother than the lower one (Fig. 6). In terms of their morphology three different types of karstic deformations can be distinguished from the base to the top. An irregularly shaped cavity or several connected cavities of various dimensions are visible at the base which are always located within the Reef Unit. Their geometries are related in an very irregular manner with the stratification. The voids are filled with angular and subangular fragments originating both from the Reef Unit itself and the Santanyí Limestone. In the middle there is a narrow, chimney-like conduit whose walls are highly rectilinear and almost vertical, and which connects the lower cavity with a sinusoidal depression which was formed by plastic warping of the lower levels of the Santanyí Limestone. These levels merge towards the chimney in a concentric way. Locally this upper structure appears to be represented by a

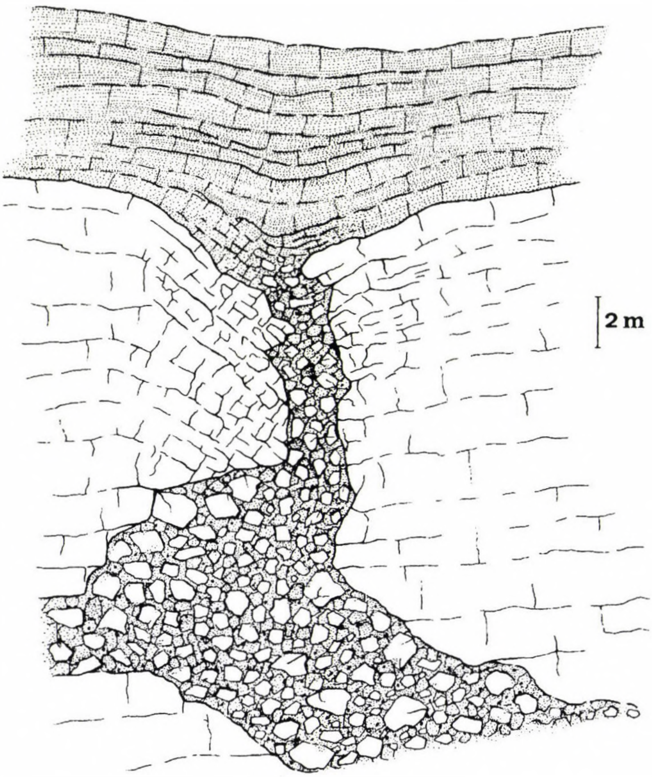

Fig. 6
Morphological outline of the collapses affecting the Upper Miocene. Their morphology is similar to that of an hourglass

confined area of fallen blocks corresponding to the upper-level of the Santanyí Limestones. After careful examination of eighteen representative localities of these collapse structures some morphometric data can be inferred. The diameter ranges from 25 to 90 m, although very small ones (<5 m wide) can also be found, while their maximum depth ranges from 3.5 to 14 m; the angle of the walls of the depressions is between 20° to 30°. The angle may exceed 45° and vertical walls also occur in the chimney. In some areas the sinkings may coalesce forming a more complex structure of greater dimension.

Characteristics of karstic deposits: breccia types and breccia formation processes

Although clays and, to a greater degree, reddish silts may also form in the latest stage of the cavity infillings, breccias are the most characteristic deposits in the karstic cavities.

The formation of breccia deposits is the result of collapse mechanisms which followed the formation of dissolution cavities. The breccias show very varied

Fig. 7
Collapse breccias formed by angular clasts of oolite limestones from the Santanyí Limestones, which fill Reef Unit cavities

compositional and textural characteristics depending on the genetic process, the involved Upper Miocene levels and the evolutionary stage of the karstic collapse.

We can differentiate several types of breccias. The most obvious are the collapsing ones generated by the falling off and filling of rocks and clasts (mainly coming from the Santanyí Limestone). The interparticle pores are filled by carbonate cement of various types (phreatic and vadose).

If the upper levels of the Santanyí Limestone oolitic grainstones were also involved in the collapse, the resulting breccia deposits were oligomictic, containing unrounded clasts of sizes ranging from a few centimeters to more than one meter without any matrix (Fig. 7). In some cases the clasts show plastic deformation. In the cases where the materials only came from the lower levels (mangrove facies), the matrix-supported breccias contain subangular to subrounded clasts of lesser size (centimetric) and the matrix is composed of calcarenites and calcisiltites.

The basal cavities (which are always located within the Reef Unit) are filled with polymictic breccias (with clasts derived from the Santanyí Limestones as well as the Reef Unit). They contain clasts of various size (from 1 cm to over 1 m) and shape (from angular to subrounded) and a significant amount of calcarenitic matrix (Fig. 8).

Fig. 8
Polymictic breccias (with clasts coming from the Santanyí Limestones as well as the Reef Unit) filling the basal cavities which are always set within the Reefal Unit and with an important presence of a calcarenite matrix

The different types of breccias described above correspond to different developing stages of the karstic collapses. The spatial distribution of the various breccia types correlates with the collapse growth stages which began with the dissolution of the Reef Unit, was followed by penecontemporaneous infilling of the cavities by semi-plastic breaking of the basal levels of Santanyí Limestones and completed by the fragile breaking of the upper layers of the oolite limestones.

Paleokarst development and evolution

In the marina de Llevant on the Migjorn area the Reef Unit is mainly composed of wackestones and packstones with very abundant benthonic foraminifera and mollusks and framestones with corals. They represent a characteristic lagoonal facies with alternating deposits of calcarenites and calcisiltites, and the presence of isolated coral colonies as patch reefs. The high porosity (formed as a result of the preferred dissolution of aragonitic coral skeletons) might have provoked the flux of plastic sediments towards those voids leading to an increase in the porosity of the upper levels and finally breaking, brecciation, collapse and also infilling of the secondary cavities (Fig. 9). Moreover, the differences in the texture of the Reef Unit and the Santanyí Limestones respectively may have favoured to subterranean drainage throughout the Reef Unit.

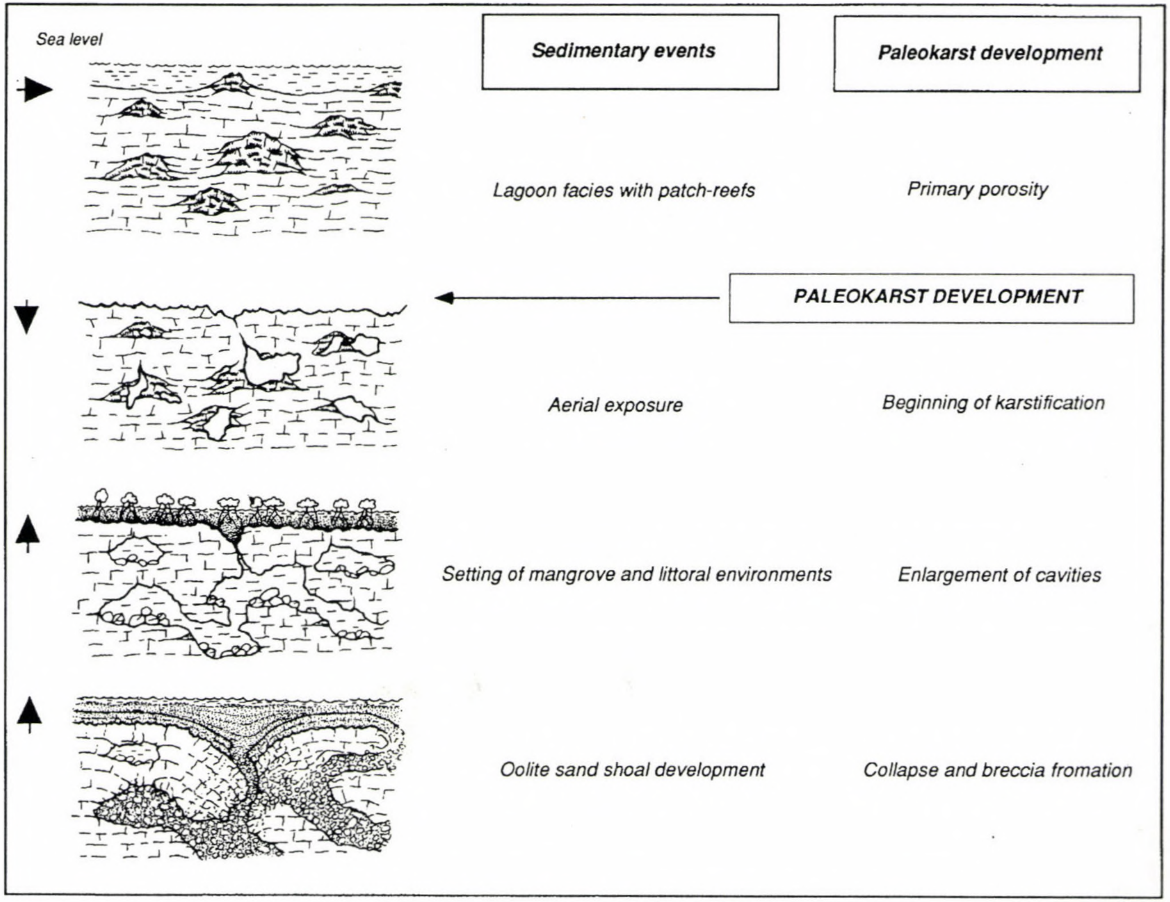

Fig. 9
Evolving outline of paleokarst development affecting the Upper Miocene of Mallorca

Although the general morphology of the collapses is very similar to that of a doline, firm evidence is missing that this kind of paleokarstic depressions performs the functions of real dolines. No direct connection with the surface can be found as a rule, although in extreme cases the collapses may reach the surface. Clear morphologies of subaerial exposure (karren) have not been found in this area either, although in other places such as the Marina the Llucmajor (Fig. 10) this seems quite possible. Consequently, the model appears to be in accordance with the notion of subsidence in a subjacent karst assumed by Jennings (1985) or by the Caribbean karst model (James and Choquette 1984).

Dating of paleokarst development is difficult. Initiation of the process seems to take place just after the deposition of the Reef Unit. The growth of the subsurface cavities must have continued during the deposition of the Santanyí Limestone, at least until its lithification, because the deformation or breaking (brecciation) would have been impossible in non-consolidated sediments. Nevertheless the karstification processes might have continued in some places during the Pliocene and the Pleistocene ages.

Fig. 10
Possible morphologies of subaerial exposure (littoral karren) in the Marina de Llucmajor (southern Mallorca)

Conclusions

Three depositional units build up the Upper Miocene of Mallorca. From base to top: "Heterostegina-bearing Calcisilt Unit", Reef Unit, and Santanyí Limestone Unit. The Santanyí Limestone Unit (also called Terminal Complex) is made up of stromatolitic and oolitic facies and it overlies a wide Late Tortonian-Messinian coral reef platform. It was formed in a shallow marine carbonate platform with sand bars and mangrove swamps.

Subsidence features of solutional origin are conspicuous in the rocky shore around Santanyí, in the southeastern part of Mallorca. Repeated deformations are outlined by the lower beds of the Upper Miocene rocks. These deformation-outcrops show also breakdown debris and setting features frequently affecting all of the Santanyí Limestone Unit. Karstification of the coral patch reef must be involved in the development of such collapse features. Voids which were formed mainly by solution of aragonitic coral skeletons may have triggered the flow of plastic materials and in some cases promoted the formation of secondary voids in the overlying beds. When these secondary cavities reached a size of several meters, sudden breakdown may have occurred causing the deposition of chaotic piles of blocks and boulders which filled the evolving void.

No clear evidence has been found so far of any direct connection between the funnel-like depressions and exokarstic features (dolines), although in some exceptional cases the hollows may reach the surface. The absence of noticeable subaerial exposure morphologies points to a genetic model closely related with the concept of subjacent karst subsidence.

Surface observations show a similar pattern of collapse processes causing the deformation of the Upper Miocene levels. They are always characterized by the presence of funnel-shaped depressions as their most distinctive features. These bedding-parallel hollows are associated with setting subsidence of materials towards underlying voids developed in the Reef Unit. Chimney-like voids appear to be located below the depressions and indicate vertical migration of the plastic layers as well as subsidence of broken blocks and slabs from more the resistant beds.

Breccia formation processes are related to solution cavity collapses resembling those occurring in other karstic environments. The types of breccias observed are dependent on both genetic processes and features of the host rock. Although there are several breccia types the most common are those collapse breccias which were formed by rock fall and by rock infilling. Interparticle voids coated by calcium carbonate cement with several growth episodes are common in the breccias. In rare cases red silts and clays make up the final cavity filling, which is not necessarily contemporaneous with the breccia formation.

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Eroded porous-media aquifer controlled hydrovolcanic centers in the South Lake Balaton Region, Hungary: The Boglár Volcano

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The volcanic centers next to Balatonboglár township represent 3.5 Ma old products of post-extensional alkaline basaltic volcanism in the Pannonian Basin (eastern Central Europe). They are small, eroded volcanic centers located on the southern shore of Lake Balaton and genetically related to the Bakony–Balaton Highland Volcanic Field eruptive centers. The relatively small area (500 m × 500 m) contains at least 2 eruptive centers, which are probably related to each other and have built up a complex volcano, called the Boglár Volcano. The volcanic rocks overlie the older Pannonian clastic sedimentary sequence and represent the topographic highs in this area. The areas of lower elevation around the eruptive centers are covered by Pleistocene to Holocene swamp, lake and river clastic sediments, which strongly suggest intense erosion during the last few million years. All volcanic rocks around Balatonboglár are volcaniclastic. There is no evidence of lava flow occurrence. The volcaniclastic sediments have been divided into two lithofacies associations. The largest amount of volcaniclastic rocks is located in the center of the local hills and has been interpreted as a phreatomagmatic crater fill lapilli tuff. They contain large amphibole megacrysts and small olivine crystals. The second lithofacies association is interpreted as lahar deposits. This sequence contains an unusually large amount of fossil tree trunks, which are identified as *Abies* species. Within a small area in the western hills small outcrops show evidence of maar-lake clastic sediment occurrence. On the hilltops debris shows intimate interaction processes between clastic sediments and basaltic melt. We interpret this to mean that the eruptive centers of Boglár Volcano were formed under subaerial conditions, with explosions fueled by intensive interaction between water-saturated Pannonian sand and uprising basaltic magma.

Key words: phreatomagmatism, hydrovolcanism, diatreme, maar, strombolian, Pannonian Basin, Bakony–Balaton Highland Volcanic Field

Introduction

A wide variety of volcanic activities characterizes the Neogene in the Pannonian Basin. Following Miocene to Quaternary subduction-related calc-alkaline volcanism in the northeastern margin of the Pannonian Basin and syn-extensional Middle Miocene intermediate to acidic and potassic to ultrapotassic magmatism, widespread post-extensional alkaline basaltic volcanism occurred between the Late Miocene and the Quaternary. It is known

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that alkaline basaltic and calc-alkaline volcanism was also contemporaneous in several places, e.g. Harghita (Harangi, in press). The alkaline basaltic volcanoes can be found either near the calc-alkaline stratovolcanic complexes or far from them, at the central part of the Pannonian Basin (Szabó et al. 1992). The Bakony–Balaton Highland Volcanic Field (BBHVF) is located about 170 km southwest of Budapest (Fig. 1) and comprises more than 50 basaltic volcanoes (e.g. Jugovics 1968; Jámbor et al. 1981) including shield volcanoes, tuff rings and maars. The Boglár Volcano is genetically part of the BBHVF but is geographically separated from it by Lake Balaton. Only two areas are known as volcanic eruptive centers on the southern shoreline of Lake Balaton, Fonyód and Boglár. The few studies of the underlying thick Pleistocene and Holocene sediments in the Lake Balaton Basin show no evidence of buried volcanic edifices in the region nor is there evidence of other surface eruptive centers in the southern shoreline region. The Boglár Volcano eruptive centers probably lie on the north–south trending Tapolca Basin tectonic line (Borsy et al. 1986) and which forms elongate, young, sediment-filled basins (Tapolca Basin in the north; Nagyberek Region to the south). Fonyód and the Boglár Volcanoes represent erosional remnants above this basal geomorphologic structure. The studied area consists two of main morphological highs, Várdomb (Castle Hill) (up to 165 m a.s.l.) and the smaller Temetődomb (Graveyard Hill) (up to 145 m a.s.l.). There is a small morphological high next to the Temetődomb, called Sándordomb (Alexander Hill) (up to 128 m a.s.l.). All of the three major elevated areas are covered by volcanic rocks. K/Ar age determination of rock samples from Boglár give an age of around 3.5 Ma (Borsy et al. 1986; Balogh, pers. com.). Early geologic mapping showed this region as explosion breccia-buried hills with small-scale lava flows (e.g. Lóczy 1894, 1913).

Physical volcanology

The volcaniclastic deposits of the Boglár Volcano can be divided into three main lithofacies (associations) based on the relative amount of accidental lithic clasts, style of sedimentary structures and textural characters (both at microscopic and macroscopic scales). A fourth lithofacies can be identified as volcanogenic clastic sediment from Holocene talus debris. These lithofacies are:

1. *Phreatomagmatic crater filling coarse-grained lapilli tuff lithofacies association (PXLT)*
2. *Reworked volcaniclastic lapilli tuff (lahar?) lithofacies association (LT)*
3. *Peperite lithofacies association, mixture of lava and Pannonian clastic sediments (MX)*
4. *Maar lake sandstone lithofacies association (ML).*

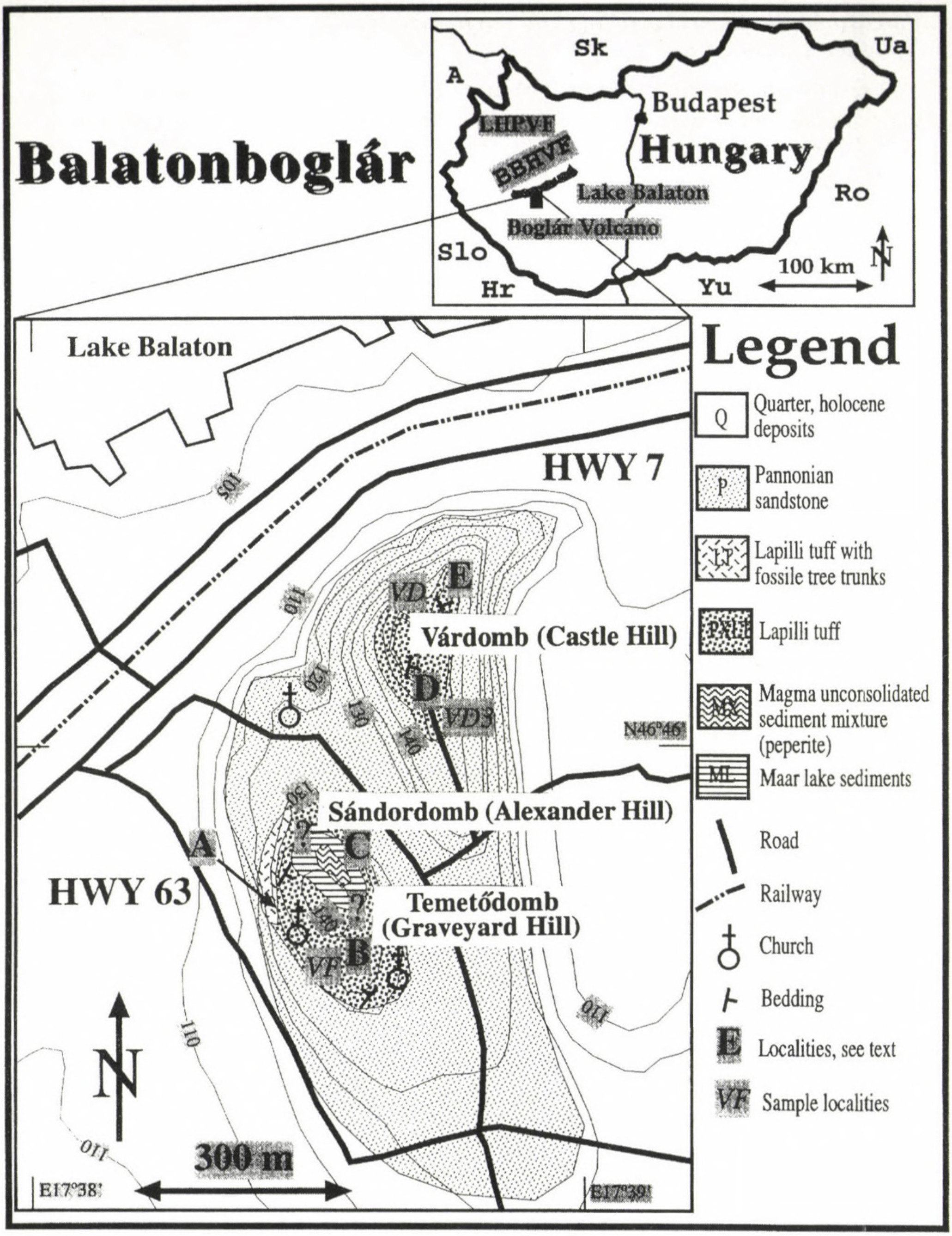

Fig. 1
Geologic map of the Boglár Volcano

Phreatomagmatic crater filling coarse-grained lapilli tuff lithofacies association (PXTL)

Description – This is the main volcanic lithofacies association in the region; it occurs on all three hilltops. The PXLT is characterized by poorly sorted, slightly bedded, brown lapilli tuff. In none of the measured sites is there any well-defined bedding. Where the bedding is relatively well developed (Várdomb, Castle Hill) measurements are always very diverse without any well-defined characteristic orientation. The dip is always steep, around 25°. In the bedded part of the sequence there is no evidence of impact sags, cross-bedding or well-defined scour fill structures. The individual beds are usually undulating with no characteristic upper and lower contact. The lithofacies contains high amounts of juvenile fragments (70 vol.-%) as unsorted vesiculated basaltic lapilli fragments which are usually palagonitized or show palagonitization processes occurring inside the clast center (Fig. 2a). Juvenile fragments can be separated into two groups. Clasts in the first group are characterized by black, highly vesiculated, usually large, rounded and semi-rounded scoriaceous lapilli. In the second group clasts are mostly light brown to yellowish, less vesiculated sideromelane fragments with small, mostly feldspathic, microliths. The vesicles are usually rounded and slightly asymmetric in shape. Often vesicles of both types of juvenile fragment are filled by secondary minerals – mostly calcite, minor zeolite – or fine grained altered sandstone fragments. There is not more than 20 vol.-% of accidental lithic fragments. Accidental lithic fragments consist mainly of sand fragments, probably from the Pannonian sandstone formations (75 vol.-% of the total accidental lithics) (Fig. 2a and Fig. 2b). There is also a low proportion of carbonate fragments (marl, limestone), probably of Mesozoic age, from depth (10–15 vol.-% of the total accidental lithics). Occasionally there are small schist fragments (probably Silurian phyllite), and a few fragments of crystalline basement rock (gneiss), but their origin is unclear because of lack of data on the basement stratigraphy of Boglár Volcano. Usually the accidental lithics are angular to semi-angular. The matrix of the samples contains strongly to slightly altered volcanic glass. Often small sandstone and silt aggregates are visible which represent intimate mixtures with the palagonitized glassy matrix. The matrix usually contains small, rounded, ovoid glauconite fragments, probably from the underlying Pannonian sandstone beds. In several samples strong carbonitization is visible. The large elongate calcite crystals occur along margins of the large clasts and show pore-fill structures. Fine-grained sand, silt and altered glass-coated lapilli are common. Free crystal proportions in the PXLT are relatively high (few vol.-% of total). They are usually brown amphibole (up to 2 cm in diameter) and altered olivine (up to 0.5 cm in diameter) in mostly idiomorphic crystals (Fig. 3).

The juvenile glass shards are very fresh, and their composition is trachy basalt from each locality (Table 1), according to electron microprobe analyses

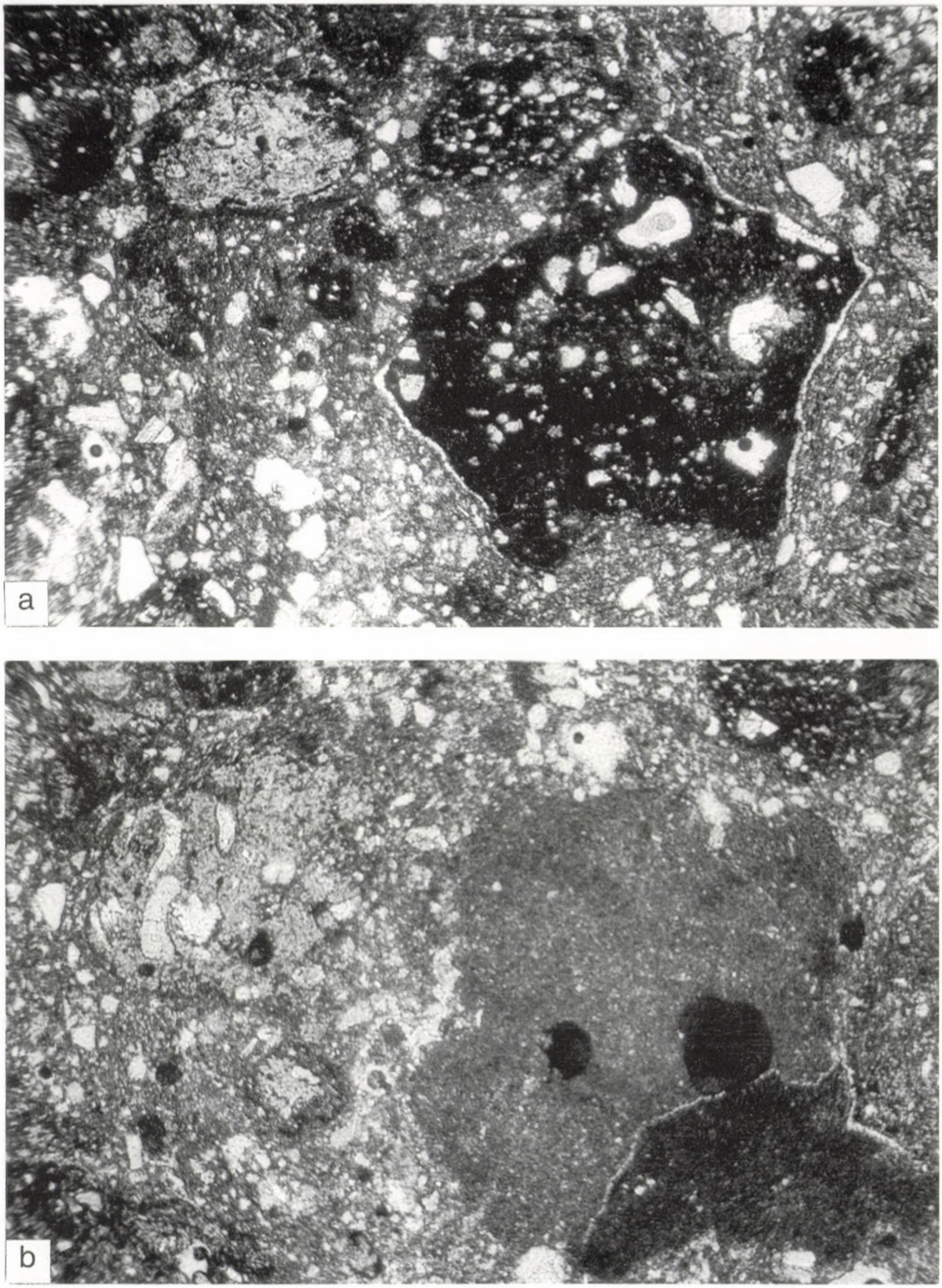

Fig. 2a, b
Photomicrographs of the PXLT lithofacies from locality D (see Fig. 1). In both pictures the shorter side of the picture is 2 mm. Note the bright, white-colored, small, angular fragments. They are quartz fragments derived from the Pannonian clastic deposits. In Fig. 2a note the difference between sideromelane (small, microvesiculated, slightly rounded, elongate lighter colored shard in the top/left corner) and tachylite glass shards (large, angular black fragment in the middle of the picture)

Table 1

Geochemical composition of volcanic glass shards from pyroclastic rock samples from three different localities. Locality names are shown in Fig. 1. Electron microprobe analyses were carried out on JEOL 8600 Superprobe, 15 kV acceleration voltage, 10–20 μm beam diameter on polished thin sections, OXIDE9 standard. $Fe_2O_3/FeO = 0.3$ using Middlemost (1989) classification

VD		CIPW Norm		Fe_2O_3/Fe) = 0.3			
SiO_2	49.53	**Q**		**ns**		**cs**	
TiO_2	3.03	**C**		**ks**		**mt**	3.09
Al_2O_3	19.19	**Z**		**di**	9.99	**cm**	
Fe_2O_3	2.13	**or**	21.39	*wo*	5.10	**il**	5.75
FeO	7.11	**ab**	18.36	*en*	2.89	**hm**	
MnO	0.13	**an**	31.93	*fs*	2.00	**tn**	
MgO	3.79	**lc**		**wo**		**pf**	
CaO	8.90	**ne**		**hy**	3.66	**ru**	
Na_2O	2.17	**kp**		*en*	2.16	**ap**	
K_2O	3.62	**hl**		*fs*	1.50	**fr**	
P_2O_5	n.m	**th**		**ol**	5.42	**pr**	
H_2O	n.m	**nc**		*fo*	3.07	**cc**	
Total	99.60	**ac**		*fa*	2.34	**H_2O**	
							Trachy-basalt
VD3		CIPW Norm		Fe_2O_3/Fe) = 0.3			
SiO_2	49.29	**Q**		**ns**		**cs**	
TiO_2	3.26	**C**		**ks**		**mt**	3.41
Al_2O_3	17.48	**Z**		**di**	18.03	**cm**	
Fe_2O_3	2.35	**or**	19.86	*wo*	9.14	**il**	6.19
FeO	7.82	**ab**	19.31	*en*	4.77	**hm**	
MnO	0.12	**an**	26.82	*fs*	4.11	**tn**	
MgO	3.38	**lc**		**wo**		**pf**	
CaO	9.82	**ne**	0.72	**hy**		**ru**	
Na_2O	2.44	**kp**		*en*		**ap**	
K_2O	3.36	**hl**		*fs*		**fr**	
P_2O_5	n.m	**th**		**ol**	4.98	**pr**	
H_2O	n.m	**nc**		*fo*	2.56	**cc**	
Total	99.32	**ac**		*fa*	2.43	**H_2O**	
							Trachy-basalt
VF		CIPW Norm		Fe_2O_3/Fe) = 0.3			
SiO_2	49.87	**Q**	0.18	**ns**		**cs**	
TiO_2	3.24	**C**		**ks**		**mt**	3.28
Al_2O_3	18.17	**Z**		**di**	13.65	**cm**	
Fe_2O_3	2.26	**or**	18.26	*wo*	6.94	**il**	6.15
FeO	7.55	**ab**	18.11	*en*	3.78	**hm**	
MnO	0.15	**an**	30.85	*fs*	2.92	**tn**	
MgO	3.59	**lc**		**wo**		**pf**	
CaO	9.57	**ne**		**hy**	9.17	**ru**	
Na_2O	2.14	**kp**		*en*	5.17	**ap**	
K_2O	3.09	**hl**		*fs*	4.00	**fr**	
P_2O_5	n.m	**th**		**ol**	5.42	**pr**	
H_2O	n.m	**nc**		*fo*	3.07	**cc**	
Total	99.63	**ac**		*fa*	2.34	**H_2O**	

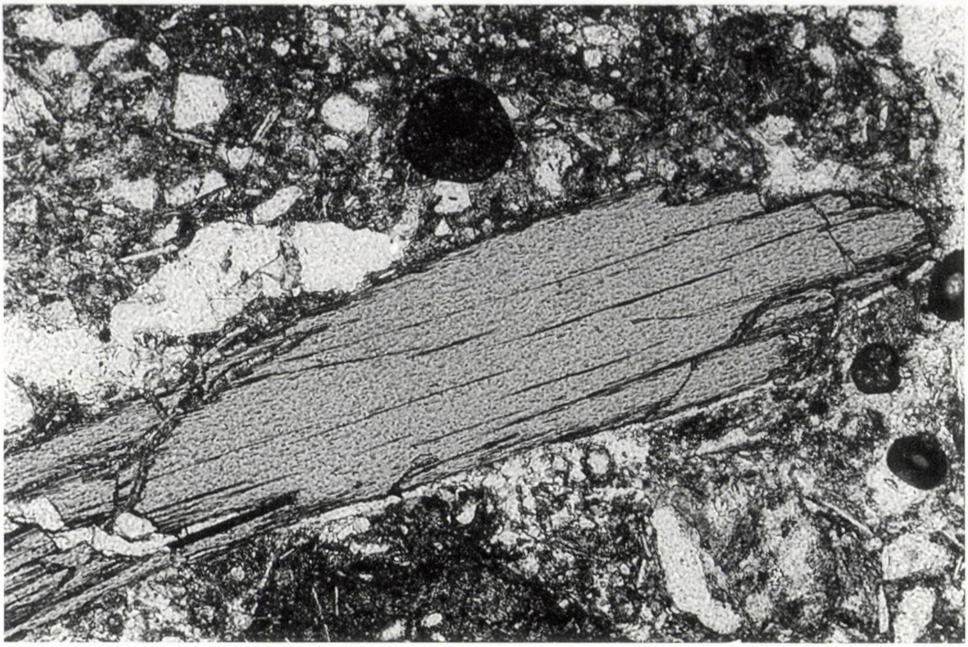

Fig. 3
Photomicrograph of the amphibole crystals in the pyroclastic sample from the PXLT lithofacies association at locality D (see Fig. 1). The shorter side of the picture is 2 mm

(Jeol 8600 Superprobe, 15 kV acceleration, 10–20 μm beam diameter, polished thin section, OXIDE9 standard) of fresh glass shards (98 vol.-% of total).

Interpretation – The PXTL lithofacies association is interpreted as crater-filling, coarse-grained lapilli tuff. The two major kinds of juvenile fragment (sideromelane and tachylite glass) show evidence of phreatomagmatic explosive processes during the eruptive history with strong magmatic influence. The lighter colored volcanic glass (sideromelane) originated from the magma-water interaction (Wohletz and Sheridan 1983; Sheridan and Wohletz 1983; Houghton and Hackett 1984; Houghton and Schmincke 1985; Fisher and Schmincke 1994) but the black scoriaceous fragments are more likely to be Strombolian magmatic explosive products (Houghton and Schmincke 1985; Houghton and Hackett 1984). The steeply dipping beds, the wide variety of dip direction in a small area, the very steep contact zone with the underlying Pannonian sandstone beds with small microfaults and the circular distribution of the lithofacies strongly suggests a near vent situation. This agrees with the descriptions of Lorenz (1986), Cas and Wright (1987) and White (1991b) of near vent deposits related to small monogenetic alkaline basaltic volcanic centers. The high amount of sandstone, siltstone fragments in the matrix, and intimate mixing with the altered glass matrix infers intimate mixing of probably water saturated unconsolidated sand and uprising basaltic magma, leading to phreatomagmatic explosions. The presence of large amounts of vesiculated sideromelane glass and scoriaceous tachylite fragments is suggestive of subaerial conditions during eruption according to Cas and Wright (1987), Fisher and Schmincke (1984) and McPhie et al. (1993). The diffuse bedding and geometrically well-defined altered zones in single rock fragments indicate some thermal, and probably hydrothermal, disturbance which could also be interpreted as a near vent depositional environment. The very strong carbonitization and palagonitization could be related to wet and probably hot environments, such as the vent zone of eruptive centers. The low proportion of deep-seated lithic fragments indicates that the Pannonian sand column was high, resulting in the explosive process controlled by the water content of the porous-media aquifer, producing normal maar volcanic centers as described by Lorenz's (1986) model. Following continuous drying of the porous-media aquifer, Strombolian magmatic explosive activity took place inside the local maar basin(s), a process well known in other maar volcanic fields where the maars are developed on thick clastic (porous media aquifer) beds (e.g. Lorenz 1986; White 1991b). In these zones strong reworking occurred, mixing primary hydrovolcanic explosion-generated fragments (sideromelane glass shards) with magmatic ones (tachylite glass shards), a similar process to that described by Houghton and Smith (1993).

The PXTL lithofacies association is located on the upper slopes of each hillside. There are no significant petrographic, petrological, textural and compositional differences between samples from these three localities. Because two of these hillsides are separated from each other by Pannonian clastic deposits outcropping in-between they can be interpreted as remnants of individual eruptive centers, but there is not enough field evidence to reconstruct the exact locality and number of eruptive centers. The similar characteristics of the deposits on each hillside also suggest that there was a volcanic complex with individual vent sites which, following erosion, now represent the hills around Boglár.

Reworked volcaniclastic lapilli tuff (lahar?) lithofacies association (LT)

Description – It is a max. 10–15 m thick (visible thickness) lithofacies in the western corner of the Temetődomb (Graveyard Hill) which consists of a very chaotic, structureless, light brown, sandy lapilli tuff. This lithofacies contains, high proportion of unsorted fine to coarse-grained, light yellowish sand matrix (70 vol.-% of total volume), comprised of a mixture of altered volcanic glass and Pannonian sands. There are no signs of well-defined bedding or of any systematic distribution of large clasts. The matrix of the lithofacies is weakly cemented and friable. The upper and lower contacts are not currently exposed, but there is strong evidence for identifying this lithofacies as that described as chaotic explosion breccia by Lóczy (1894, 1913). That description reports that the volcaniclastics have a sharp contact with the Pannonian sandstone, with abundant microtectonic features. The lithofacies association is dominantly composed of accidental lithic fragments (visual estimation: 90 vol.-% of total volume). The lithofacies association contains a high proportion of rounded, semi-rounded, thermally affected Pannonian sandstone pebbles (visual estimation: 75 vol.-% of the total volume of large – 1 cm clasts, up to 10 cm in diameter). The Pannonian sandstone pebbles have no impact character or string orientations. A few large (up to 10 cm in diameter) altered, palagonitized, micro-vesicular basaltic fragments also occur in this lithofacies. The tree trunks are usually covered by greenish, strongly altered tuffs (Fig. 4). The high amount of tree trunk fossils have a wide size range, from 1–2 cm in length to pieces up to 40–50 cm long and 10–15 cm wide (Fig. 5). The larger, concave shape juvenile or lithic fragments also contain altered fine-grained rims of varying thickness.

Interpretation – The chaotic structure of different lithologies in a very altered sandy matrix, and presence of altered tuff rimmed larger fragments, strongly suggests reworking processes, for instance deposition by lahars. The characteristic

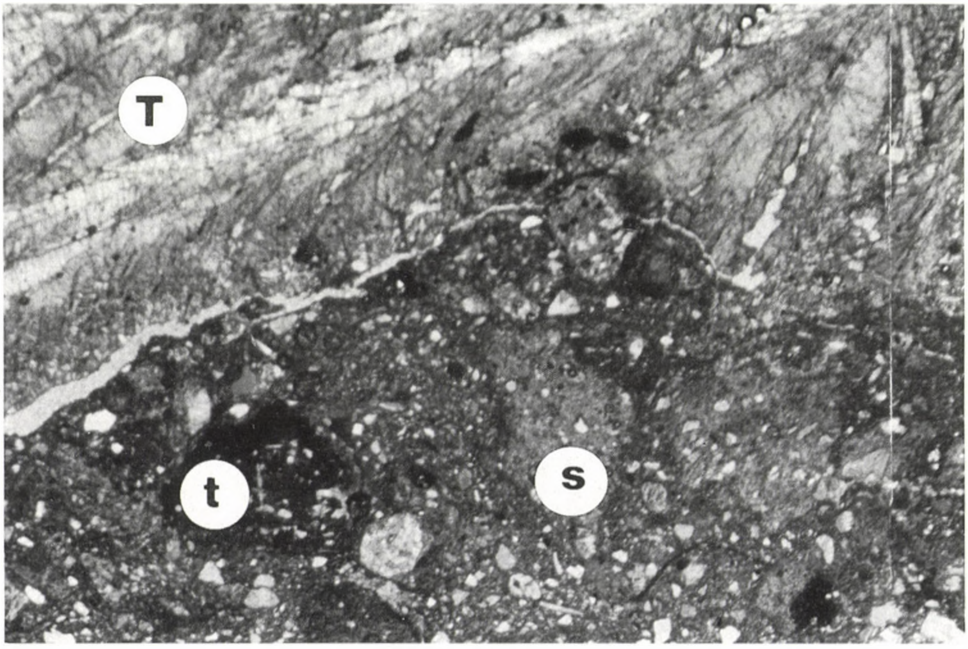

Fig. 4
Photomicrograph of the tree trunk and host rock in thin-section. The shorter side of the picture is 2 mm. T – tree fossile, t – trachylite, s – sideromelane

Fig. 5
Photo of the lahar deposit with tree trunks at locality A (see Fig. 1)

sharp contact of the Pannonian clastic sediments result from valley fill deposition of the sediment, or of a small volume of strongly reworked-recycled, vent-filling deposits of a local basically phreatic explosion center. The large amount of sandy matrix and matrix-supported large clast-bearing character is interpreted as a cohesive debris-flow deposit in which the massive, matrix-supported pebbly mudstone (thermally affected Pannonian sandstone fragments) and tree trunk fragments were suspended in and supported by the matrix according to Lowe's (1982) model. In this interpretation we describe this lithofacies as a lahar deposit, which means rapid water-supported flows of volcaniclastic particles generated on volcanoes (Smith and Lowe 1991; McPhie et al. 1993).

Peperite lithofacies, mixture of lava and Pannonian clastic sediments (MX)

Description – In a very small area on the top of the Temetődomb (Graveyard Hill) several unusual rock samples (with 10–35 cm diameter) were found (Fig. 6). The original source of the samples is unknown but is probably not far away from the locality where they are found, because this is locally the highest elevated area in the region. The samples consist of a mixture of fine-grained sand and scoriaceous juvenile fragments. The juvenile fragments are black, highly vesiculated, and very glassy. The vesicles reach 1 cm in diameter and are usually rounded with a smooth inside surface. The contact between the original sand/silt region and the original juvenile fragments is diffuse. Several alignments are clearly visible as a string of small vesicles and fresh sandy or muddy, "untouched" relict sand fragments. The transition between the juvenile fragments and the sandstone is usually continuous. Next to the large juvenile fragments the original sandstone/siltstone fragments are melted and probably mingled together with the original basaltic melt.

Interpretation – The wide range of characteristic fragments on top of Temető-domb (Graveyard Hill) are interpreted as peperite fragments. Peperites form when hot lava intrudes into wet unconsolidated sediments (Fisher 1960; Williams and

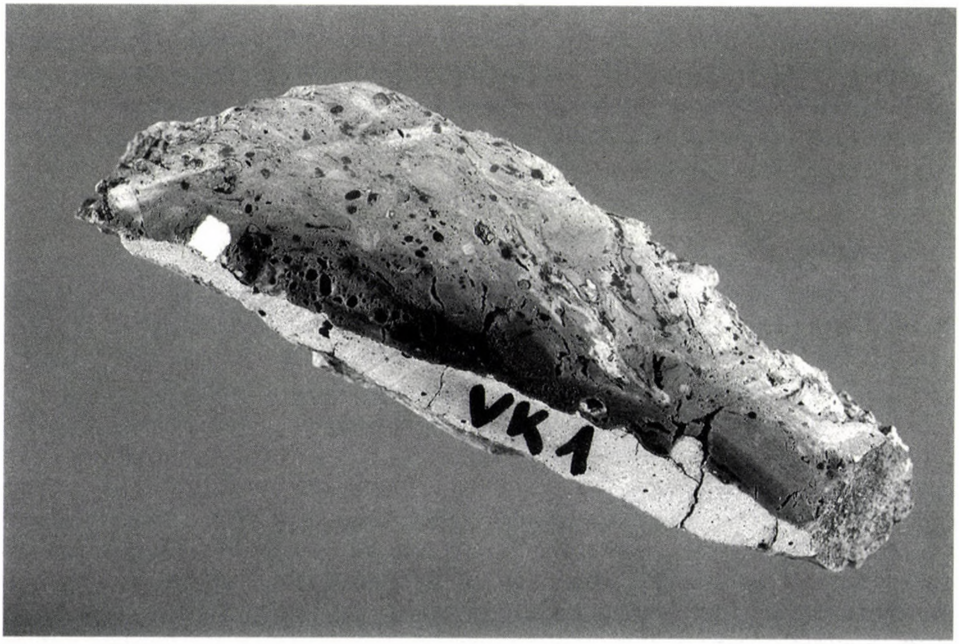

Fig. 6
Picture of the peperite. The shorter side of the picture is 15 cm. Note the irregular shaped dark, scoriaceous fragment in the middle of the sample in a light color sand matrix

McBirney 1979; Busby-Spera and White 1987). They are characterized by a clastic texture in which either component may form the matrix. This description clearly fits the observed character of samples from Temetődomb (Graveyard Hill). Using the classification of Busby-Spera and White (1987) we can classify the peperite from Boglár as globular-type peperite, where the host sediment was probably unconsolidated water-saturated and fine-grained, as the Pannonian sand must have been during the eruptive history of the Boglár Volcano.

Maar lake sandstone lithofacies (ML)

Description – There are several intercalated, thin (5–25 cm), well-cemented sandstone beds in the northern sector of the Temetődomb (Graveyard Hill) area. The sandstone beds are more abundant toward the upper part of the theoretical stratigraphic column. There is no well-exposed section where the stratigraphic section is easily measurable. The deposits in several cases contain large amounts of carbonate matrix and become laminated, coarse-grained, and unsorted. The sandstone is pink to light gray. The matrix of the sandstone contains small altered volcanic glass shards with feldspathic microliths, free broken brown amphibole, feldspar fragments and several small rounded tuff fragments. Rounded glauconite clasts are also present.

Interpretation – The intercalated sedimentary layers, which contain several clasts of volcanic origin, suggest that the sediment is a post-eruptive, probably maar lake sediment. The large amount of quartz fragments implies an abundant sand source from the Pannonian sand beds, which makes up a large part of the volcaniclastic products.

Paleobotany

The fossil trees are unsorted and unoriented in the volcaniclastic matrix. The studied sample is angular in shape, with numerous cracks filled either by iron hydroxides or matrix of the host rock. It is a decimeter size fragment of secondary xylem, strongly deformed. Hence, its shape does not allow any inferences about the original size of the tree. It has been studied under the number MP 917. The preservation is poor. The wood is epigenized by hyperblastic calcite. Locally however, in the cracks, some tracheids have been epigenized by pyrite and subsequently oxidized into iron hydroxides. These processes preserve the original wood structure quite well. However, pits were observed only in a small area. Some other tracheids have been cast by the matrix, but not delicately enough to preserve the tracheid pits. It has been studied with Collodion casts. The Tracheidoxyl (i.e. an isolated secondary xylem compound unique to tracheids) has faint but marked growth rings. Radial pitting is typically abietinean with either spaced round pits or biseriate opposite ones. Quite frequently pits are in contact, somewhat flattened. Rays are heterogeneous including tracheids at their edges. The ray cell walls are quite thick and the transversal ones are pitted. The cross-fields are occupied by two oculipores. Axial parenchyma and resin channels have not been observed. Although observations were possible in only a very small area (2 x 1 mm), because of the poor preservation, the general pattern indicate quite safely the genus *Abies*. The genus *Abies* is a frequent component of paleoarctic biota, particularly fresh and temperate settings. It has been reported from the entire Neogene. It is thus not a very good indicator, either from a paleoecological or a stratigraphic point of view. Extant Abies, at least in Europe, prefer temperate to fresh temperate climates with a regular water supply. They are often found in mountains. Although they can grow on boggy grounds, they usually avoid wet areas. There is no hint of charcoal on the surface of the wood, nor any burns (Fig. 4 and Fig. 5). This suggests a low temperature deposition history of the host sediment. On the other hand, the original epigeny through pyrite implies fossilization under reductive conditions.

Conclusion

Our study is based on several key field observations and detailed descriptions of the different lithofacies (associations) around the hill area of Balatonboglár township. We have concluded that the main volcanic lithofacies on the three hilltops show similar sedimentological and petrological characteristics. We interpret these volcaniclastic rocks as the products of a phreatomagmatic explosive eruption. The large amount of quartzofeldspathic fragments in the matrix in the volcaniclastic sequences strongly suggest the importance of magma and unconsolidated water-saturated sediment mixing processes during the eruptive history. This conclusion is supported by the occurrence of globular peperite in the region. Probably peperite was formed when magma intruded into wet maar lake sediment and/or Pannonian clastic sediment. The eruptions took place under subaerial conditions (high vesicularity of juvenile fragments, no pillow structures), in which the main water supply for the phreatomagmatic explosions was the water saturated clastic Pannonian sediments (Fig. 7a).

The steeply dipping and chaotic structural position of pyroclastic deposits suggest that they were in a near vent position where syn-eruptive movements (Fig. 7b and c), thermal effects, and hot springs were important during and after the eruptions, as reported by White (1991a, b) from near vent deposits of Hopi Buttes volcanic field. Repeated explosions might be responsible for the intensive reworking/recycling process of juvenile and sedimentary fragments in the vent zone similar to the model of Houghton and Smith (1993). The phreatomagmatic explosions probably caused maar depression(s) which was/were filled by water and in which maar lake sedimentation could have occurred (Fig. 7b). With decreasing pore-water supply from the porous-media aquifer, Pannonian sand, dryer explosive events took place, building up small Strombolian scoria cones and adding more magmatic juvenile shards (tachyilite)

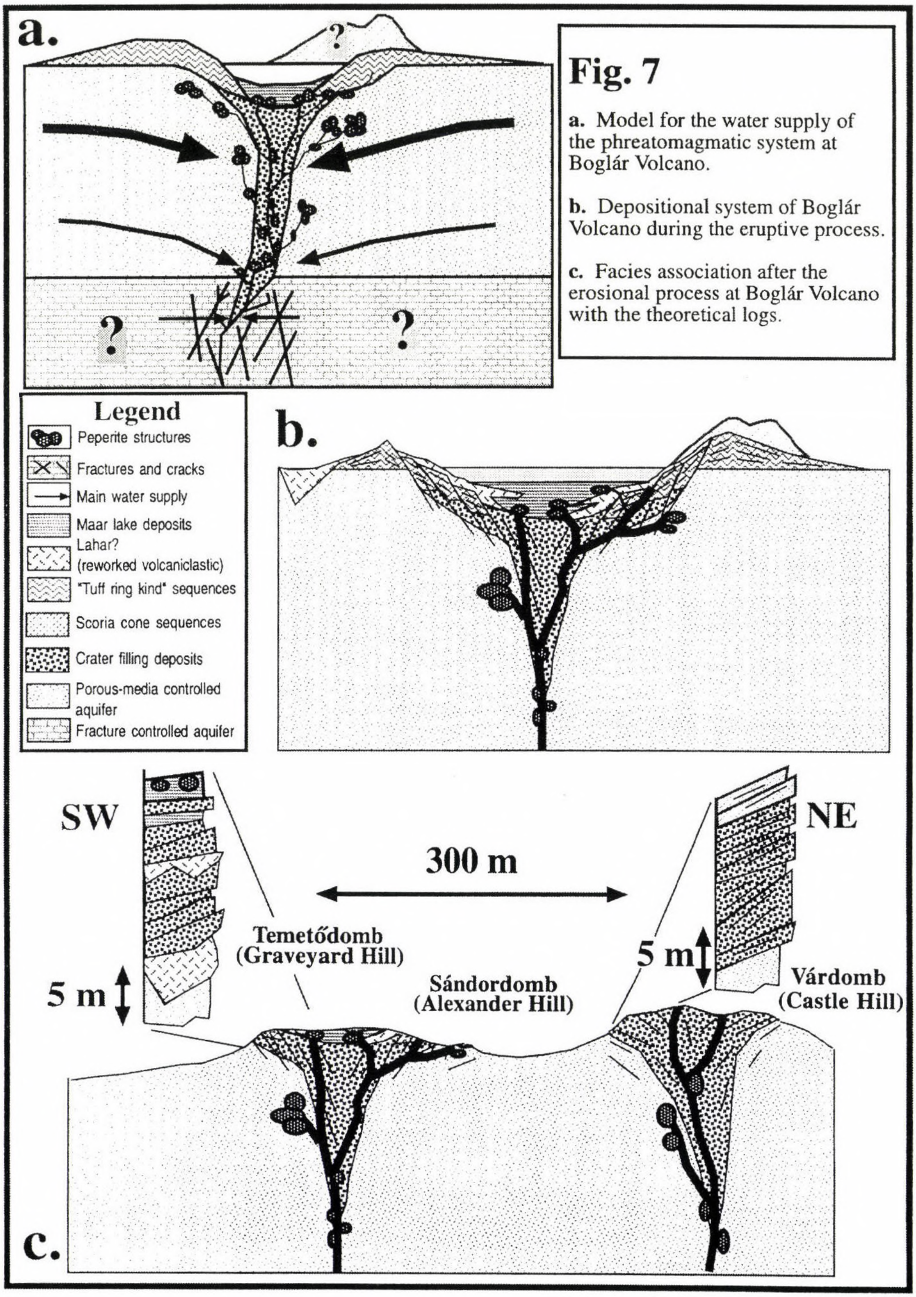

Fig. 7
Composite picture of the eruptive mechanism (a. and b.), theoretical stratigraphic columns (c.) and reconstruction of lithofacies relations after erosion (c.) of the Boglár Volcano

to the deposits. This process is well known from other volcanic fields where the main water source of the phreatomagmatic explosions was porous-media aquifer (e.g. in Hopi Buttes, Arizona, White 1991a; 1991b, Western Snake Rivers, Godchaux et al. 1992; Eifel, Germany, Büchel 1993). The new tree fossil suggests that the eruptions took place under subaerial conditions, and that probably all the volcanic processes were low temperature, which fits with the low-temperature phreatomagmatic, phreatic explosive history.

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Platform-basin transition and depositional models for the Upper Triassic (Carnian) Sándorhegy Limestone, Balaton Highland, Hungary

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The Upper Julian–Lower Tuvalian (Carnian, Upper Triassic) intraplatform basin successions in the Balaton Highland (Hungary) are represented by alternating limestone and marl sequences formed in changing periods of enhanced terrigenous input and platform translocations. The last phase of basin infilling was studied by means of a detailed microfacies investigation in three measured sections. Based on sedimentological, lithological and paleoecological characteristics sixteen facies types, representing six paleoenvironments, were defined. A generalized depositional model was set up to illustrate the relations of environments in different time intervals.

In the studied case, ecological stress, i.e. adverse conditions, may have been a vital factor controlling the carbonate production during relative sea level rises. The platform retreated by the effect of the influx of fine terrigenous material which reduced its growth potential. During highstand platform progradation resulted in a shallowing-upward facies pattern. During the Late Julian lowstand organic-rich laminites with a specific ostracod assemblage, indicating a hypersaline, oxygen-depleted environment, were generated.

Key words: microfacies analysis, intraplatform basin, Sándorhegy Formation, Carnian, Balaton Highland

Introduction

Based on modern ideas of carbonate sedimentology and sequence stratigraphy (Kendall and Schlager 1981; Sarg 1988; Schlager 1991; Vail et al. 1991), some remarkable papers have dealt with Triassic sedimentation of marginal and offshore regions of the Tethys (Gnaccolini and Jadoul 1990; Bechstädt and Schweizer 1991; De Zanche et al. 1993; Budai and Haas 1997). Previous studies of Carnian basin evolution revealed the basis for the concept of the Sándorhegy Formation (Budai 1991; Csillag 1991; Haas 1994; Nagy 1998). Although Carnian sediments of the Balaton Highland were studied according to various aspects in the last decades (Góczán et al. 1983; Gyalog et al. 1986; Góczán and Oravecz-Scheffer 1996a, b) these studies did not focus on the

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evolution and significance of the final phase of basin accretion. In this work an attempt is made to eliminate this gap in our knowledge by detailed sampling and thin section analysis of three key sections (Fig. 1). Particular emphasis was placed on the description of different facies and microfossils assemblages (foraminifers, ostracods and calcareous algae). Based on the observations, sedimentological models were established to explain the relationships of the defined facies-types and the basin evolution.

Geologic setting

The Balaton Highland (as part of the Transdanubian Range) was located between the Northern Calcareous Alps and the Southern Alps at the early stages of Alpine evolution (cf. Kázmér and Kovács 1985, fig. 4; Haas 1987, fig. 5). Based upon the evidence of foraminifer and sporomorph assemblages the Zsámbék Basin (sensu Góczán and Oravecz-Scheffer, 1996b) shows paleogeographic similarities to the German Basin, whereas the Veszprém Basin shows good correlation with the St. Cassian (Italy) one (Haas 1994; Góczán and Oravecz-Scheffer 1996b).

Triassic geologic history of the Balaton Highland within the Transdanubian Range was determined by the paleogeographic setting after Variscan tectogenesis and Middle Triassic rifting (Budai and Vörös 1992, 1993) of Tethys (cf. Haas and Budai 1995; Haas et al. 1995; Haas and Budai, in press). In the Middle Anisian–Lower Ladinian (*"Buchenstein Epoch"*), due to westward progression of the expanding ocean, extensional type basins began to form, accompanied by intense magmatism (Harangi et al. 1996). Sedimentation in the deep basins is represented by pelagic cherty limestones and volcanic tuffs (Fig. 2); however, in areas between the easternmost part of the Balaton Highland and the Buda Mts., carbonate platforms (Budaörs platform) survived (fig. 7a. in Haas 1994). In the Carnian (*"Reingraben"* or *"Lunz Epoch"*) carbonate platform evolution was mainly controlled by sea-level fluctuations and changing intensity of terrigenous influx. During the Early Julian global sea-level fall (Haq et al. 1987) two isolated platforms grew (the Ederics reef in the west and Sédvölgy reef in the east the latter is chronologically the direct continuation of the Ladinian platforms, being underlain by the Budaörs Dolomite F.; see fig 7b. in Haas, 1994) and between them an intraplatform basin was established. In the Early Julian–Middle Tuvalian that basin was filled up due to terrigenous input and platform progradation (Haas 1994, fig. 7c–d). The last phase of this period is represented by the Sándorhegy Formation in the Balaton Highland. From the latest Carnian (*"Dachstein Epoch"*) the Fődolomit (phrase "Fődolomit" is analogous to Hauptdolomite or Main Dolomite and used hereinafter) accumulated on a leveled topography (Fig. 2). The remainder of the Mesozoic sequence in the Balaton Highland was eroded. Eoalpine overthrusts (NW–SE direction) resulted in a few km of shortening (Litér Overthrust, Fig. 1; cf. Balla and Dudko 1993).

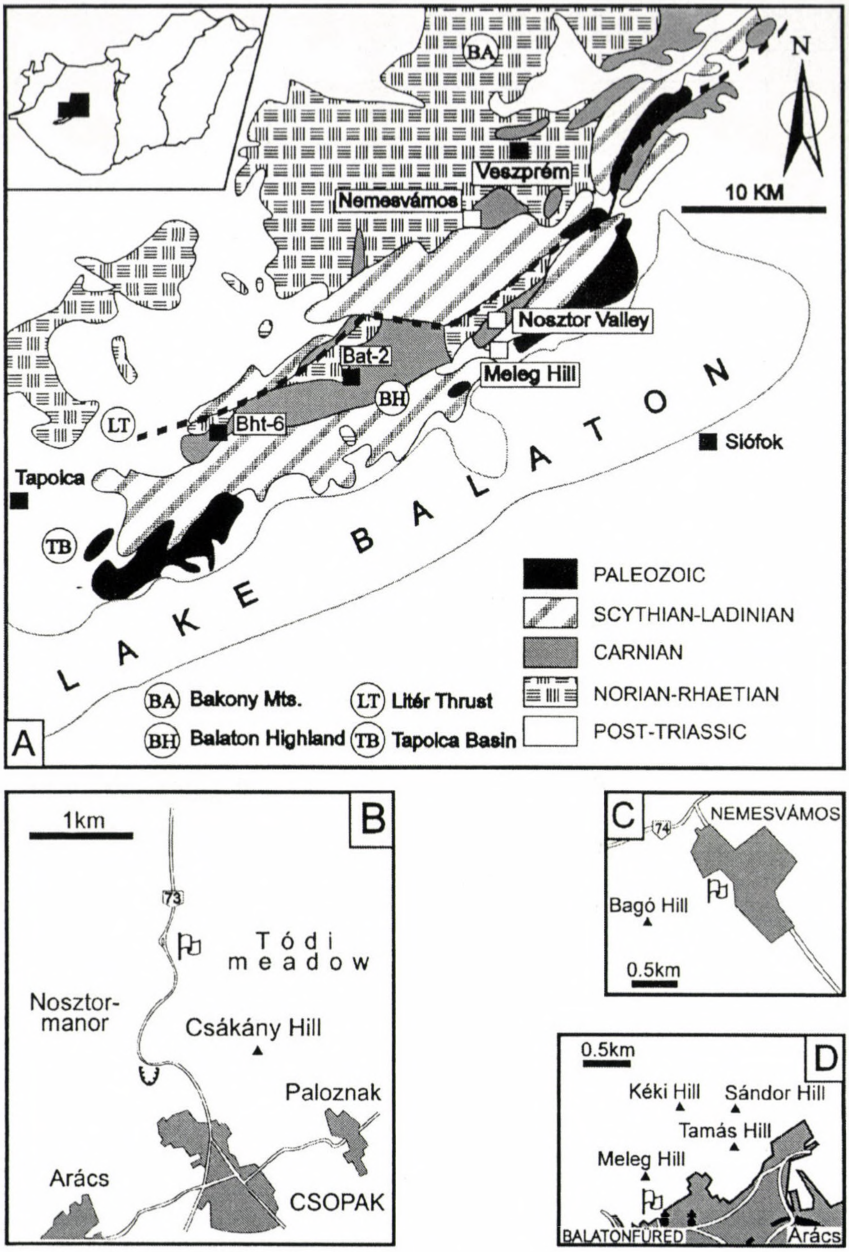

Fig. 1
A-Location of the studied successions in the Balaton Highland, Hungary. Empty squares show sampling points. Geology after Fülöp (1984). Abbreviations: Bat-2: borehole Barnag-2, Bht-6: borehole Balatonhenye-6. B, C, D-sketch maps of the particular studied areas, flags for illustrating the studied sections (for accurate itinerary see Nagy 1998)

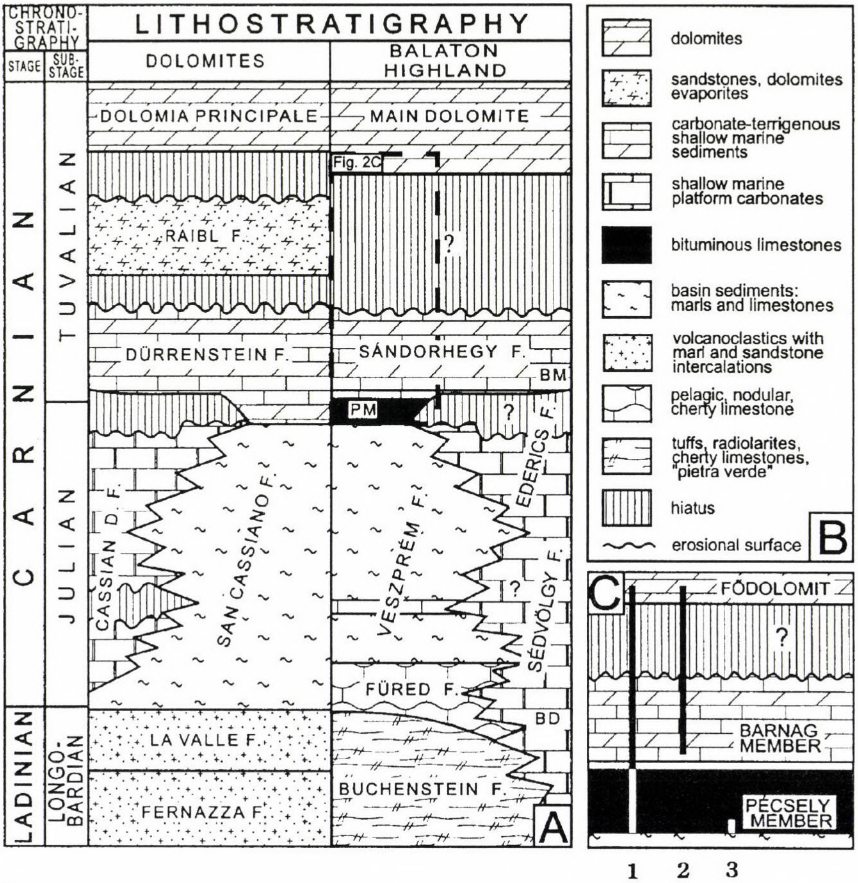

Fig. 2
A-Carnian lithostratigraphy of the Balaton Highland (Hungary) tentatively correlated with that of the Dolomites. Compiled from Bosellini (1991) and De Zanche et al. (1993) for the Dolomites, and Budai and Haas (1997 – slightly modified) for the Balaton Highland. Abbreviations: BM – Barnag Mb., PM – Pécsely Mb., BD – Budaörs Dolomite F. B – Legend for the lithology. C – Inset from Fig. 2A indicating the range of the studied successions. 1. Nosztor Valley, 2. Meleg Hill, 3. Nemesvámos

Facies description and types

Three sections exposing parts of the Sándorhegy Formation were measured and investigated in detail. The location of the measured sections in Noszor Valley (Csopak), Meleg Hill (Balatonfüred) and Nemesvámos are shown in Fig. 1. The following descriptions focus on the key section in Nosztor Valley but additional information was also provided for the others. The section in Nosztor Valley is the most complete one, representing the entire formation, i.e.

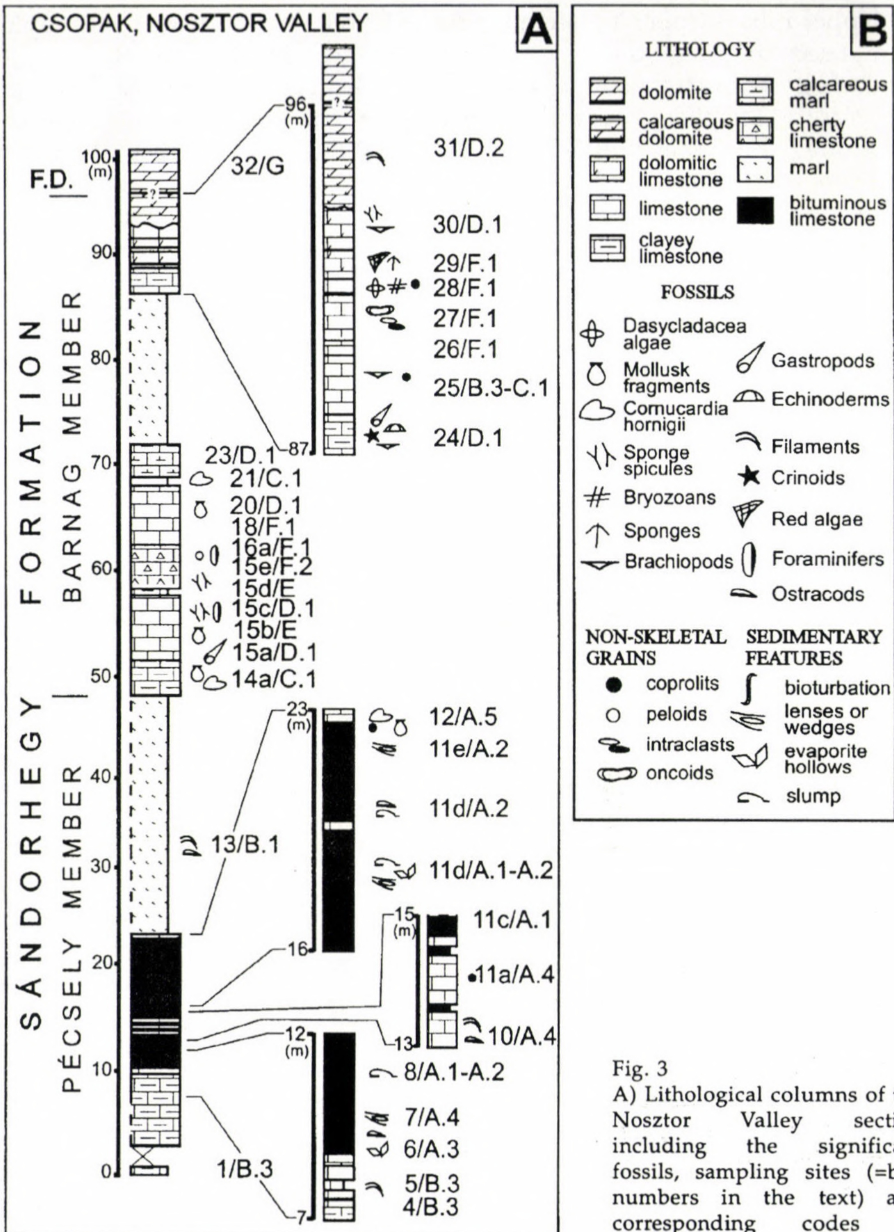

Fig. 3
A) Lithological columns of the Nosztor Valley section including the significant fossils, sampling sites (=bed numbers in the text) and corresponding codes of microfacies. For facies codes see the text and also Fig. 3B for fossils. B) Legend for lithology, fossils and sedimentary features. Abbreviation: FD – "Fődolomit"

the Upper Julian–Upper Tuvalian interval. The others, which are less complete, were used to study the lateral facies changes. In the studied area, the Sándorhegy Formation is represented by two distinct members: the lower one, the Pécsely Member, is made up mainly by marls and black bituminous limestones; the upper one, the Barnag Member, is constituted by marls and bioclastic limestones and dolomites (Figs 3, 11, 12).

The lowermost part (0–10 m) of the Nosztor Valley road-cut section exposes yellow-brown, medium-bedded marls to clayey marls (environment code B.3 in Fig. 3). On weathered surfaces microslumps and microslides can be observed. The unit generally contains very few allochtonous mollusk fragments (1–3%) but in some beds it is densely packed with filaments showing definite orientation, parallel to the bedding (mudstone-wackestone, beds 1–5). Euhedral gypsum pseudomorphs (calcitized) and crustacean coprolites also occur in some beds (Fig. 4A).

The marls are followed (10–23 m, beds 6–12) by black bituminous laminites (code A in Fig. 3) capped by a single limestone bed rich in pelecypods and crustacean coprolites (Fig. 4B). Based on the types of laminites and the amount of intraclasts and bioclasts five facies types could be distinguished: A.1) laminites made up of an alternation of very thin (0.1–2 mm) bituminous limestone and clayey limestone laminae (Fig. 4B, C); A.2) "ribbon" limestones (0.5–5 cm thick layers) with up to 15–20 cm-thick shale interlayers and tempestite beds (ostracod and mollusk coquinas – bed 11e). Synsedimentary deformation structures such as small-scale slumps, slides and cracks are common (Fig. 4D). It can be assumed that slumping of the thin layers may have been caused by a lubricious marly layer at the base; A.3) wackestone-packstone intercalations or lenses in the laminites containing peloids (or pellets) and filaments or often a great amount of mollusk fragments; A.4) dolomite beds with vugs, probably after dissolution of evaporites (Fig. 5A), with algal-mat rip-ups and various intraclasts. Pores under larger reworked mollusk shells preserved sediments of different origin (wackestone-grainstone) in the sheltered area (Fig. 5B). Cavities after dissolved gypsum or salt crystals were subsequently filled by sparry calcite. The shape of the open cavities shows angular forms. A.5) a single limestone bed (12) densely packed with large (1–8 cm) bivalves (*Cornucardia hornigii* Bittner) and crustacean, Favreina-type coprolites and few intraclasts closes the lowermost part of the section. This bed is overlain by a poorly exposed and intensively bioturbated marl and clay series (23–47 m). This lithofacies-type (code B.1 in Fig. 3) is characterized by a low percentage of allochthonous fragments of mollusks and echinoderms in micritic matrix (bed 13, Fig. 5C, mudstone).

The marl interval is covered by medium-bedded limestones representing the lower part of Barnag Member (47–72 m). The limestone series begin with thin to medium-bedded, yellowish-brown, light-gray, clayey limestones and calcareous marls (bioclastic wackestones, code C.1 in Fig. 3). The clayey limestones are nodular and thin-bedded with clay intercalations in bed 14

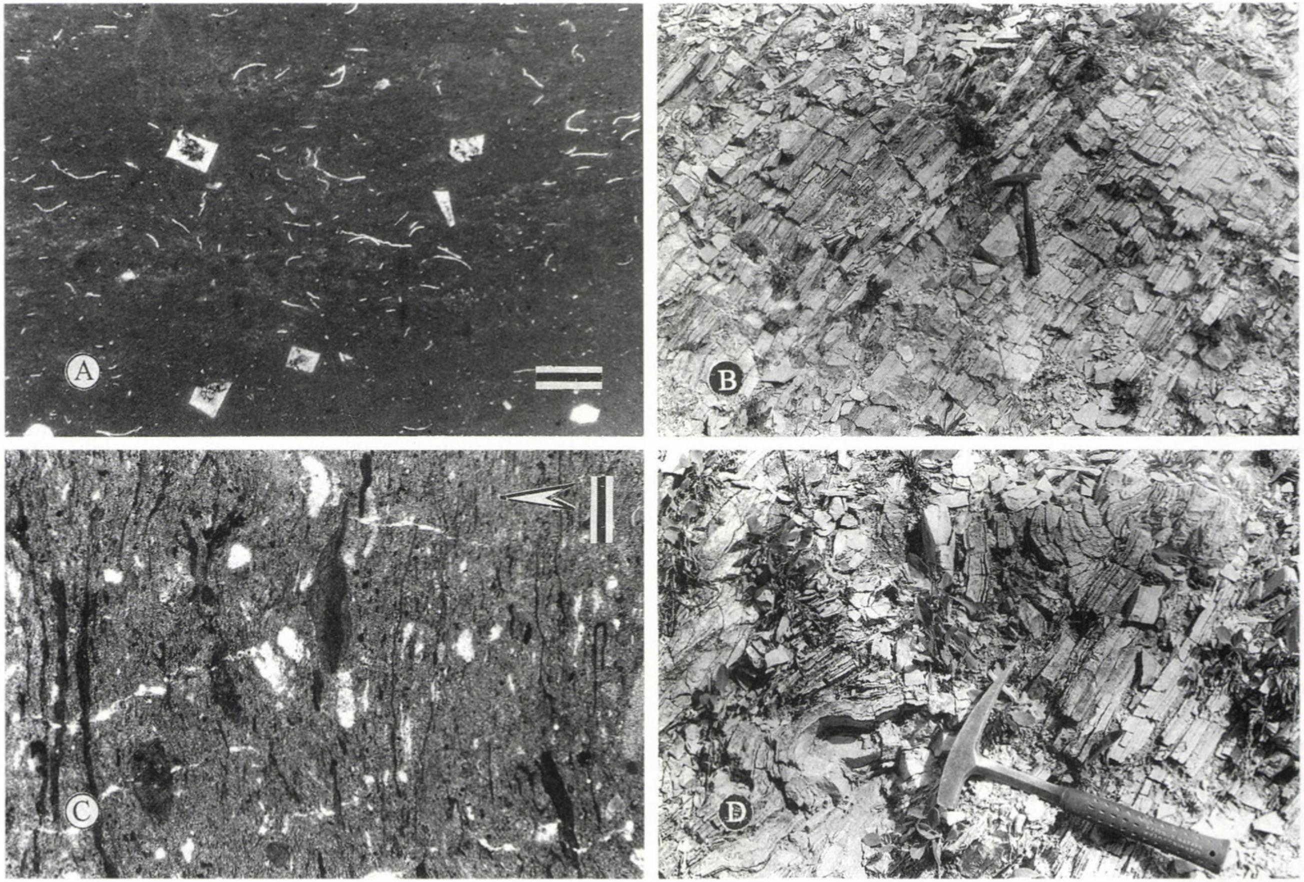

Fig. 4
Photos (A–D) showing characteristic features of Pécsely Member in the section at Nosztor Valley. A) tiny filaments and sparite pseudomorphs after evaporites; × 13 2; B) bituminous laminites (lower part of the Pécsely Mb.); C) subtidal algal-limestone from the anoxic laminites; × 13, 7A; D) forms of gravity mass flow, i.e. slumps in the laminites. Hammer, coin and black bars for Fig. scale, the length of a bar is 1 mm. After semicolon (;) magnification, sampling sites

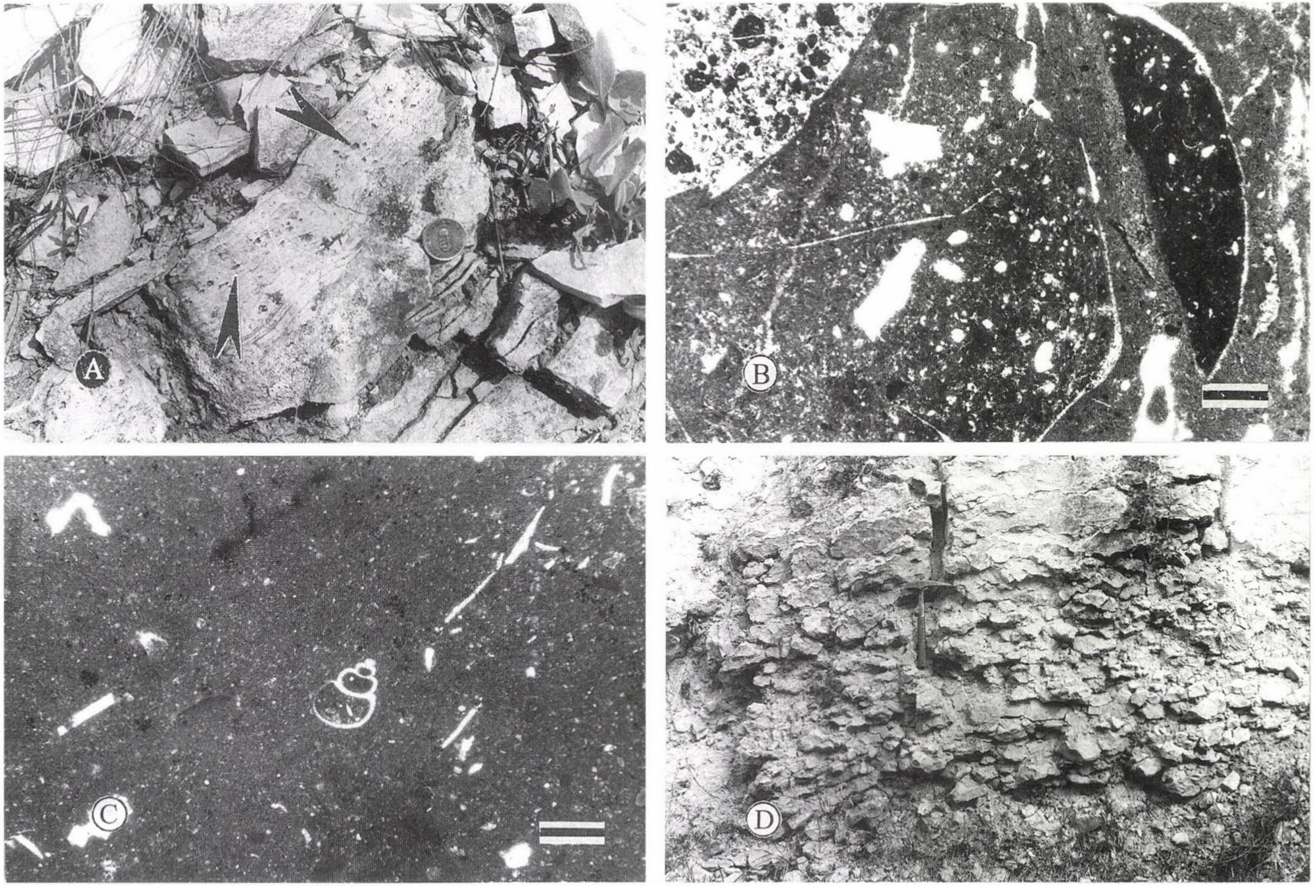

Fig. 5
A) dissolved vugs (arrows) after evaporites in the laminites in the lower part of Pécsely Member. B) intraclasts in the laminites, peloidal grst in the upper left corner and pelitic filled shell fragment in the right and the center; × 13, 6. C) gastropod-echinoderm bioclastic mdst/wkst from marl series of the Barnag Member; × 13, 23. D) nodular limestone with clay parting continuously grading to medium bedded limestone (lowest part of Barnag Mb.). Hammer, coin and black bars for scale, the length of a bar is 1 mm. After semicolon (;) magnification, sampling sites

(Fig. 5D), and grade upward to medium to thick-bedded limestone. The rocks are made up predominantly of allochtonous fossil fragments (20–30%) such as large gastropods, bivalves, echinoderms, few ostracods, holothuroids and a specific foraminifer assemblage (*Gsollbergella-Agathammina* Association). The overlying medium to thickly bedded gray limestone beds (bed 14b–15a, Fig. 6A) contain a few small, probably autochtonous megalodontids (*Cornucardia hornigii* Bittner, *Neomegalodon carinthiacus* Hauer). Poorly sorted, fairly rounded allochtonous bioclasts (40–60%) along with brachiopod remnants are common and in some beds large oncoids occur (Fig. 6B, C, bioclastic wackestone-packstone, code D.1 in Fig. 3). It is worth mentioning that large epigenetic calcite framboid pseudomorphs (up to 10 cm in diameter) after gypsum were found in bed 15. Beds 15 and 16 show distinct lithological and microfacies features. They are light to deep gray or brown limestones, often with chert nodules, in 15–20 cm thick layers with parallel or undulated bedding planes (code E in Fig. 3). They contain white, gray or brown elongated chert nodules, 10–20 cm in size. The bioclasts are much better preserved than in the previously described facies. Foraminifers (*Aulotortus-Triadodiscus-Nodosaria* Assemblages) and calcitic molds of siliceous sponge spicules (Fig. 6D, foraminiferal and sponge spicule biomicrite-wackestone) are dominantly characteristic. The cherty limestones are overlain by clayey limestones showing upward-increasing clay content. It is important to mention that bed 21 consists of *Cornucardia* and *Neomegalodon* lumachella in a clayey matrix.

The upper part of the section (87–96 m) is made up of medium to thick beds of fossiliferous limestones and clayey limestones which are locally dolomitized (beds 24–30). These layers are underlain by marls but that interval is not exposed in the section. An erosional surface (between beds 30–31) closes the upper carbonate interval. In bed 24, the great amount of fossil fragments (many brachiopod fragments and crinoid ossicles) suggest the same microfacies (D.1) as observed in bed 15–16 (wackestone). In bed 25 there are a very few remnants of fossils such as mollusk and brachiopod fragments (mudstone). Beds 26–29 are 20–40 cm thick, light brown or gray limestones or partially dolomitized limestones, with parallel bedding planes. The rocks are made up mainly of bioclasts but intraclasts and oncoids also occur (packstone-grainstone, code F.1 in Fig. 3). Mollusks, echinoderms and foraminifers predominate, although crinoid ossicles and green and red algae are also common and some poorly preserved fragments of bryozoans (?) were also recognized (Figs 7A–8D). Skeletal debris was commonly encrusted by microorganisms such as blue-green algae or cyanobacteria and sessile foraminifers (Tolypammina). Most of the grains are well rounded (abraded) and sorted and have micritic envelopes. Around the grains a fine, fibrous, isopachous calcite fringe is visible and the pores are filled with sparry calcite. Large gastropods are often filled with micritic internal sediments. Beds 27–30 are partially dolomitized, with that process becoming more intensive upward in the section (Figs 8A–D). Beds 27–28 are made up of fine xenotopic (-A) dolomite and in Bed 29 idiotopic (-C)

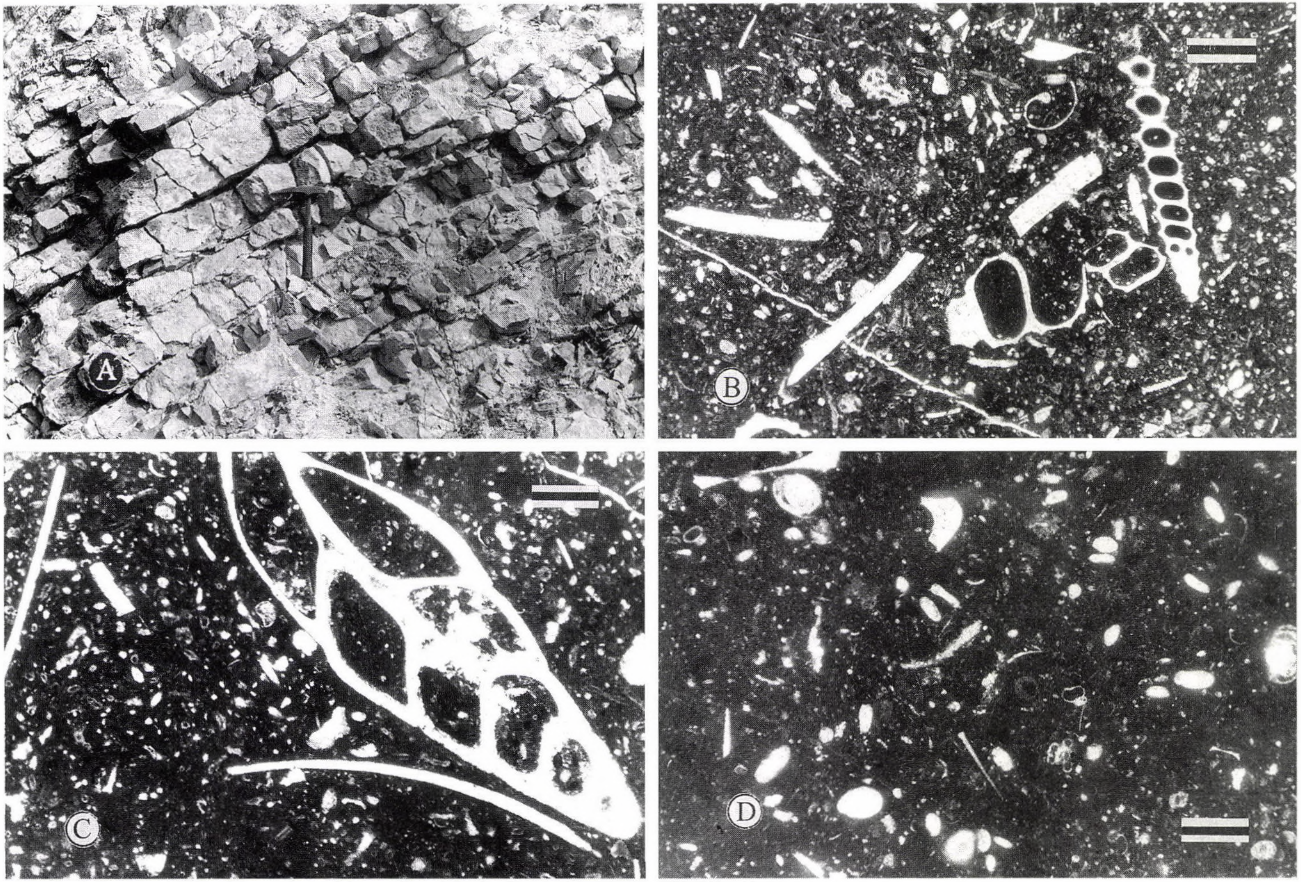

Fig. 6
Photos showing characteristics of slope and terrace facies in Nosztor Valley (NV) and Meleg Hill (MH). A) medium-thickly bedded bioclastic limestone, Barnag Mb. (NV). B) mollusk, echinoderm bioclastic pkst/wkst from slope facies; × 13, 1; (MH). C) bioclastic limestone with a huge gastropod and thin shell fragments from slope facies; × 13, 15, (NV). D) foraminiferal-sponge spiculine wkst/pkst, (*Aulotortus*?-*Triadodiscus*? in the top left and in the right, spicule in the bottom center); 13X, 2, (MH). Hammer, black bars for scale, the length of a bar is 1 mm. After semicolon (;) magnification, sampling sites

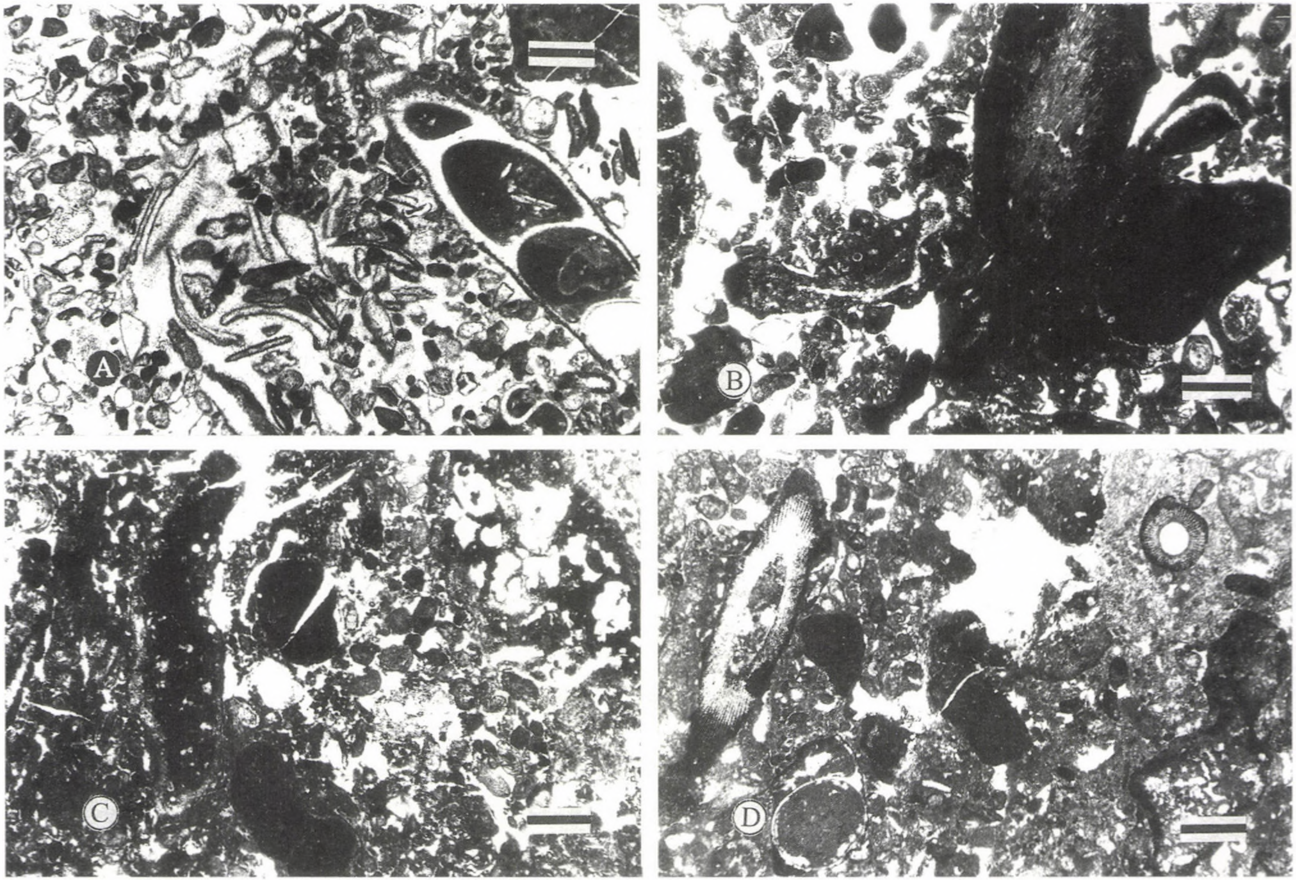

Fig. 7
Photos showing characteristics of wave-agitated winnowed platform edge sand facies from Nosztor Valley (NV) Meleg Hill (MH). A) intraclastic-peloidal-bioclastic grst with pelitic intrasediments in a gastropod; × 13, 26, NV. B) encrusted red algae in peloidal-intraclastic grst; × 13, 6E, MH. C) pelitic lumps encrusted by microorganisms in peloidal-intraclastic grst; × 13, 6B, MH. D) bioclastic, partly micritized grst with echinoderm spines (left and right), gastropod cross-section (bottom left) and micritic envelopes around pelitic lumps (center); × 13, 6C, MH. Black bars for scale, the length of a bar is 1 mm. After semicolon (;) magnification, sampling sites

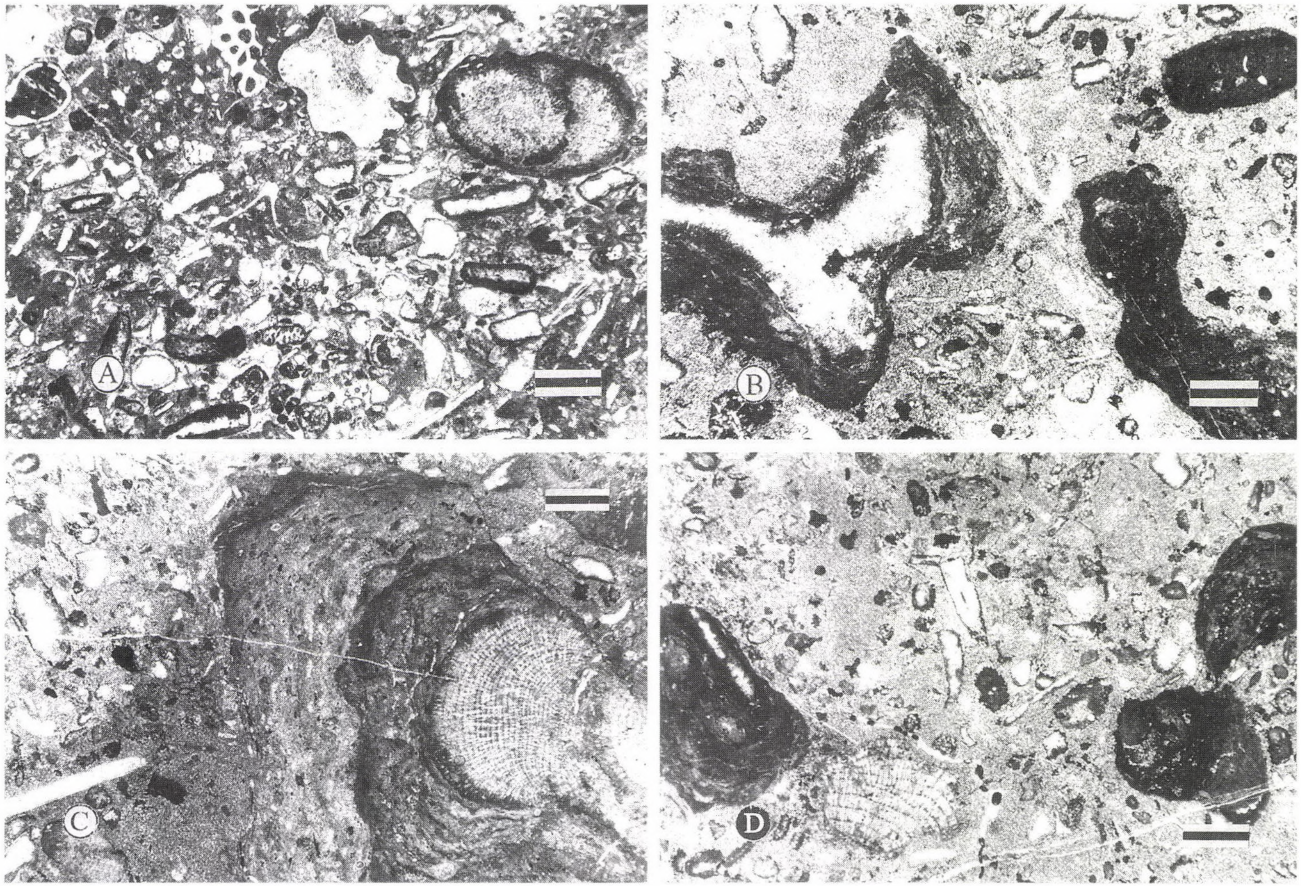

Fig. 8
Photos showing high-energy facies (A–D) which were moderately dolomitized. A) slightly dolomitized bioclastic grst/pkst with echinoderm plates (top), totally calcitized clast after microborings (top right); × 13, 26, NV. B) considerably dolomitized bioclastic grst with red algae encrusted by sessile foraminifers or algae; × 13, 27, NV. C) red algae encrusted by sessile foraminifers or algae in considerably dolomitized grst; × 13, 29, NV. D) red algae, oncoids and encrusted lumps in partly dolomitized grst; × 13, 29, NV. Black bars for scale, the length of a bar is 1mm. After semicolon (;) magnification, sampling sites

dolosparites fill the voids and biomoldic pores. In bed 30, a few relict texture elements could be observed (D.1) and small dolomite crystals and patches (porphyrotopic) appear (mudstone–wackestone).

Over a sharp erosional surface (93–95 m) reddish-lilac or yellowish-brown dolomitized limestones and dolomites occur (bed 31). The contact with the Fődolomit is not perfectly clear. In the rocks of reddish color no fossils were found due to total dolomitization, while mollusks, echinoderm fragments and algal-mat remnants could be recognized in the yellowish rocks (D.2). This rather problematic interval is covered by the typical Fődolomit (bed 32). It consists mainly of dolomites, partly dolomitized limestones or clayey dolomites (idiotopic, code G Fig. 3). Their color varies from light yellow to gray, even to white. The thickness of the beds ranges between 10–25 cm with parallel bedding planes. Particles and matrix are heavily dolomitized (Fig. 9D): bioclasts (dasycladacea and mollusks ghosts), breccias (deep gray to black lithoclasts or algal-mat fragments in dolosparitic matrix) are typical. The facies-types recognized in the other sections (Meleg Hill and Nemesvámos) could be correlated with the ones in Nosztor Valley, although some additional microfacies types could also be observed. These are as follows (Figs 11, 12).

The lowermost part of the section at Meleg Hill (beds 1a–3 and 8a–13), as well as the beginning of the section at Nemesvámos (beds 1–3), are characterized by facies-type D.1. The lower-middle part of the Meleg Hill section can be matched lithologically to the middle part of Nosztor Valley one (cherty limestones were applied as markers), although facies-type F.1 reaches greater thickness and F.2 is more dominant in Balatonfüred. In bed 8, large oncoids could be observed with many fragments of mollusks and echinoderms. In bed 9, large pectinids were found in intensively bioturbated mudstones. The upper part is made up of various limestone/clayey-marl and dolomitized limestone/dolomitized marl alternations (C.1–D.1).

In the Nemesvámos section beds 6–8 have been emplaced in the lowermost part of Pécsely Member (Csillag and Haas 1993). This lowermost part of the member was exposed only in this section (code C.2 in Fig. 12) beneath the bituminous laminitic limestones. In the yellowish-brown, gray, medium to thickly bedded lithoclastic and bioclastic peloidal limestones (wackestone) redeposited and encrusted lithoclasts and bioclasts, i.e. oncoids, prevail. Bioclasts appear both in the peloidal micritic matrix and as particles within the lithoclasts. In the matrix a few filaments, ostracods, encrusted fragments of mollusks, green algae, foraminifers (*Tetrataxis* Association), gastropods, echinoderms and brachiopods are typically present. The lithoclasts contain different forams (*Aulotortus* sp.), sponge spicules, echinoderm and mollusk fragments (Fig. 9B, C). Lithoclasts can reach cm size and are angular. The rocks are intensively bioturbated.

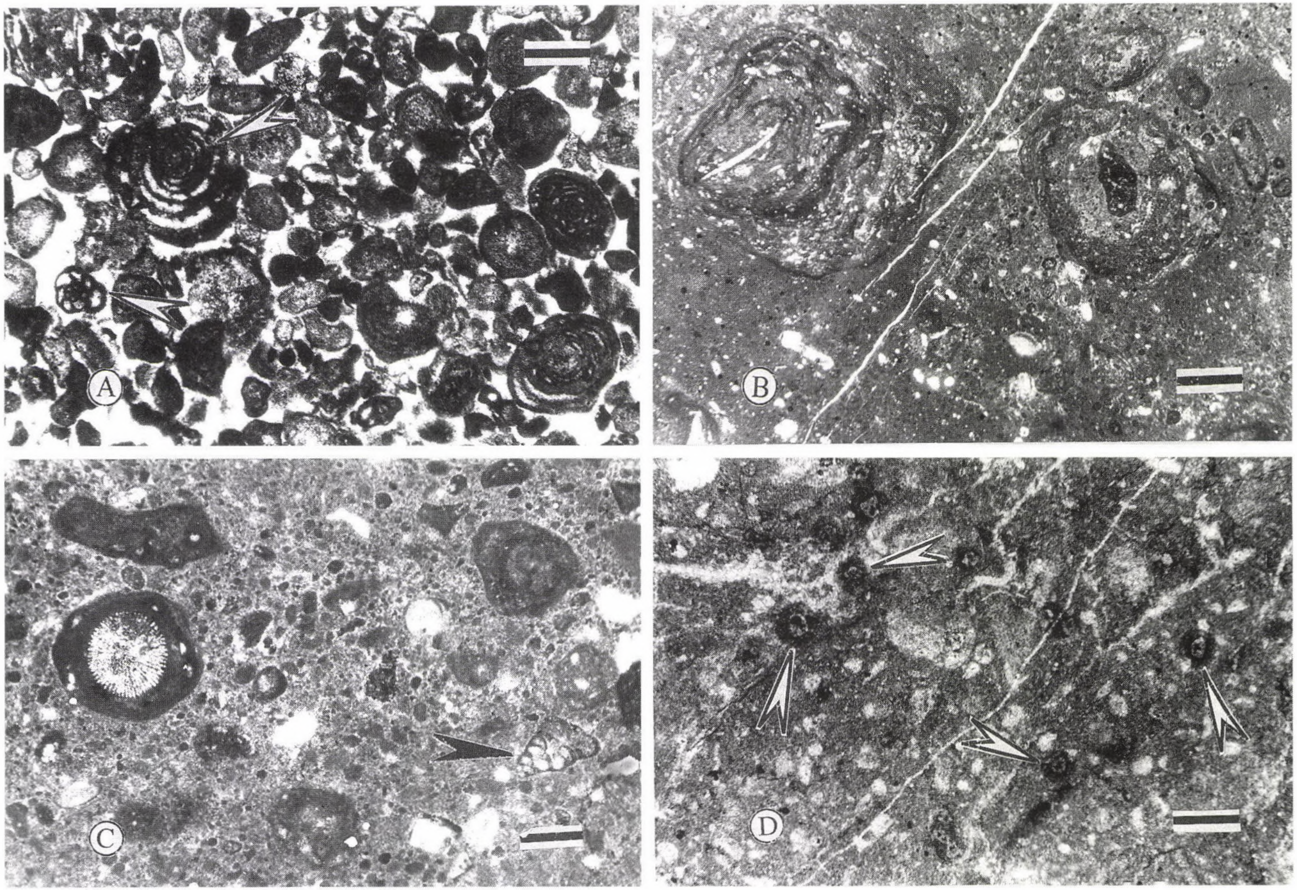

Fig. 9
High-energy, slope facies type at Nemesvámos (N) and tidal flat microfacies from Nosztor Valley (NV), Meleg Hill (MH). A) foraminiferal-intraclastic peloidal grst, *Glomospirella* (upper left) and *Trochammina* (lower left); × 26, 5B, MH. B) oncoidal wkst/pkst; × 13, 8, N. C) oncoidal-peloidal wkst/pkst with *Tetrataxis* foraminifer (arrow) and encrusted echinoderm ossicle (left); × 26, 7, N. D) heavily dolomitized algal limestone or dolomite ("Fődolomit"?), arrows show Dasycladacea algae; × 13, 15, MH. Black bars for scale, the length of a bar is 1 mm. After semicolon (;) magnification, sampling sites

Characterization and evaluation of microfossil assemblages

A number of previous studies summarized the characteristics of the Julian–Tuvalian microfossil assemblages of the Balaton Highland area (Oravecz-Scheffer 1987; Góczán et al. 1991; Kovács 1991; Monostori 1994; Góczán and Oravecz-Scheffer 1996b). The present study has yielded some new microfossil data, mainly about foraminifer and algae which were determined from thin sections by A. Oravecz-Scheffer and O. Piros, respectively (Figs 10–12). These data significantly contributed to a better understanding of the relationship between biotopes and depositional environments. Based on the determinations of Oravecz-Scheffer, the following four foraminifer associations could be defined in the studied sections:

– *Aulotortus-Triadodiscus-Nodosaria* Association with sponge spicules (code E). Fossils are well preserved and there is no indication of any reworking (Fig. 6D). Fragile nodosarids also require unagitated conditions to be well preserved. The fauna characterizes the shallow subtidal environments (e.g. backreef or shallow subtidal upper zone of a gentle foreslope); however, they may have been transported to deeper regions as well (Meleg Hill beds 4a–4c, Nosztor Valley beds 15b and 15d). In the section at Nemesvámos, the lithoclasts containing foraminifers and sponge spicules were encountered in basin facies (*Tetrataxis* ass.) suggesting basinward redeposition (Fig. 9B, C, beds – 8, code C.2);

– *Gsollbergella-Agathammina* association was found in marl-clayey lime- stone beds (Meleg Hill 8–9a, Nosztor Valley 15a and 18 beds). Both genera indicate an undisturbed, muddy environment (C.1 facies);

– *Glomospirella-Trochammina-Ophthalmidium* association characterizes a wave-agitated, shallow subtidal environment (code F.2);

– *Tetrataxis* association with a great number of encrusting sessile foraminifers, echinoderms, and a few filaments were encountered in the matrix of lithoclastic wackestone at Nemesvámos.

Earlier ostracod studies revealed that hypersaline conditions prevailed during the deposition of the black laminites forming the lower part of Pécsely Member (Monostori 1994). Green algae found in some layers of the Barnag Member indicate a high-energy environment (code F.1 in beds 26–28, Nosztor Valley; bed 6, Meleg Hill). Conodonts found in a marl bed (Kovács 1991) in the lower part of the Barnag Member (probably beds 15–18 in Nosztor Valley) suggest a deeper water slope-basin environment.

Depositional Setting

In the investigated Upper Julian–Upper Tuvalian sections (Nosztor Valley, Meleg Hill, Nemesvámos), within five paleoenvironments sixteen facies types (A.1-G) have been recognized. Recognition of the facies was based on microfacies characteristics and evaluation of microfossil (foraminifer, ostracod)

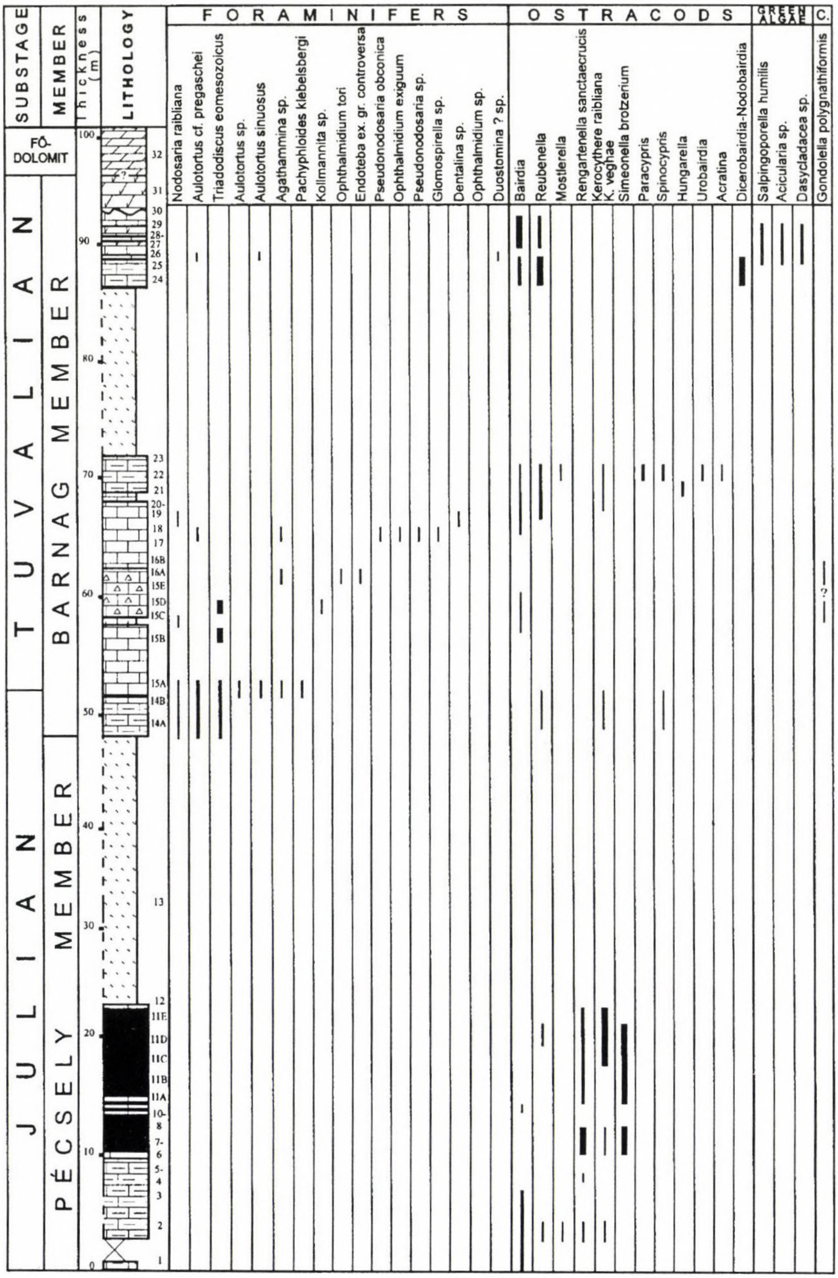

Fig. 10
Evaluation of foraminifers, ostracods (after Monostori 1994), green algae and conodonts (abbr. C; after Kovács 1991) from the section at Nosztor Valley, Csopak. Thicknesses of lines indicate quantity. See Fig. 3B for lithology

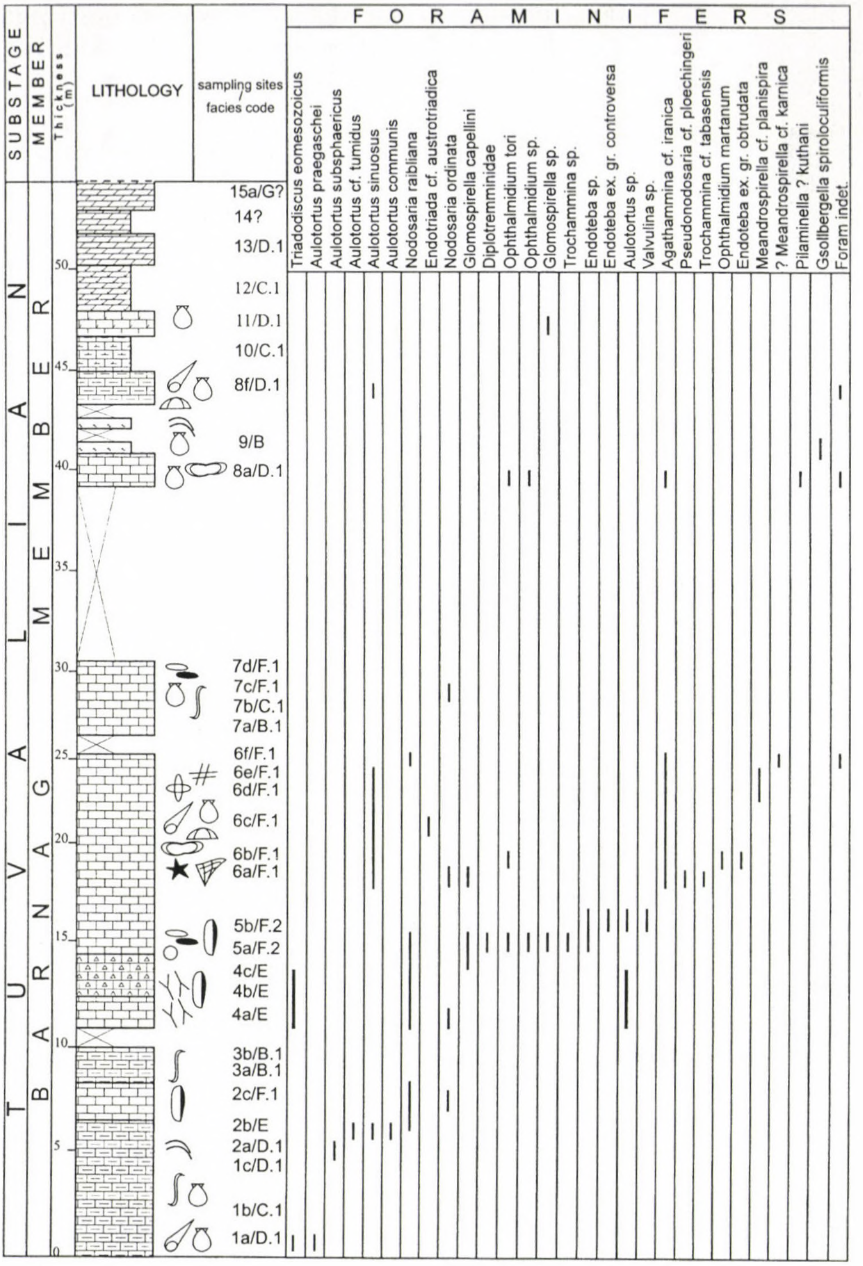

Fig. 11
Lithological column including the significant fossils with sampling sites, corresponding facies types and evaluation of the characteristic foraminifer taxa of the Meleg Hill section, Balatonfüred. Thicknesses of lines indicate relative quantity. For facies codes see the text and also Fig. 3B for fossils

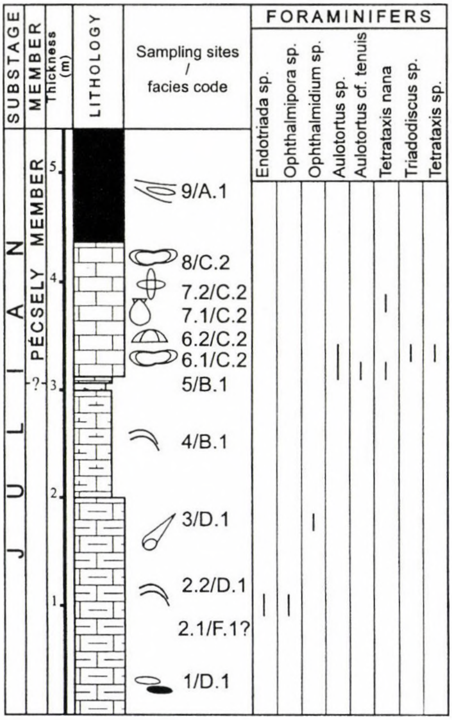

Fig. 12
Lithological column including significant fossils with sampling sites and corresponding facies types and evaluation of foraminifers from the section at Nemesvámos. Thicknesses of lines indicate quantity. For facies codes see the text and also Fig. 3B for fossils

assemblages. A concise description of environments and facies units is given in Table 1.

Shallow basin facies unit. Two sub-units (A and B) have been differentiated within this unit according to lithological and paleoecological features. They both indicate a low energy level, beneath the normal wave base, although there is some small interbedding with normal grading which may reflect either tempestites or simple sliding on gentle slopes towards the basin. The lithofacies and biofacies of the *anoxic-dysoxic basin* (A) can be characterized by a) various types of laminites with a high content of organic carbon and without any indication of bioturbation, and b) specific vagile ostracod assemblages (Monostori 1994) on the bottom indicating hypersalinity (beds 6–12 in Nosztor Valley; bed 9 at Nemesvámos). Besides the ostracods, subtidal microbial mats forming by a special bacterial biocenosis could be outlined in the laminites (M. Hámor-Vidó, pers. comm.). In bed 12, directly covering the laminites, the fossil content indicates normal salinity, suggesting the termination of the oxygen-depleted and hypersaline conditions. In the *normal saline shallow basin* (B) autochtonous gastropods, bivalve-ostracod filaments, and pectinids are the typical forms and the intensive bioturbation indicates a well-oxygenated bottom water (beds 1–5, 13, and between 23–24 in Nosztor Valley; beds 3, 9 at Meleg Hill; bed 4 at Nemesvámos). In beds 3–5 in Nosztor Valley, calcite pseudo-morphs of small euhedral evaporite crystals indicate increasing salinity upsection.

Slope facies unit (C and D). This facies unit can be characterized by varying amount of fossil fragments. They are mostly mollusks, echinoderms, foraminifers, holothuroids, ostracods and brachiopods but typical shallow marine platform elements (e.g. thalamid algae, sponges, corals) are absent. According to the quantity of redeposited bioclasts, two sub-units were distinguished. Fine, heavily crumbled bioclasts indicate relatively long transportation on the slopes (*lower* *foreslope*-C), while the large number of poorly sorted and moderately rounded skeletal debris suggest short transportation distance (*middle and upper foreslope*-D). In the Nemesvámos section, encrusted bioclasts and lithoclasts were transported and redeposited in the lower slope facies. The characteristics of some lithoclasts were very similar to those which were considered as a typical feature of facies unit E (next paragraph). That fact can raise the suspicion of juxtaposition of the two paleoenvironments. The uppermost part of the Nosztor Valley section is pervasively dolomitized but from the relict texture elements it can be assumed that the slope facies reappears over the erosional surface. Slope facies characterize the lowermost and also the upper part of Barnag Member in Nosztor Valley (beds 14a–15b and 17–23, 24 and 30–31) and in Meleg Hill (beds 1a–3 and 8a–13) as well as the beds beneath the lowermost part of Pécsely Member at Nemesvámos (beds 1–3, 6–8).

Slope terrace facies unit (E). Good preservation of fossils (foraminifers and spicules) and low percentage of reworked sediments in micritic matrix indicate a low energy environment beneath the normal wave base. This foraminifer assemblage usually inhabits the shallow subtidal zone (Góczán and Oravecz-Scheffer 1996b). Lithoclasts originating from this facies unit may have been transported to deeper regions (to slope or basin) as was mentioned above. This facies could be traced in the Barnag Member in Nosztor Valley (beds 15–16) and Meleg Hill (beds 4–5).

Winnowed platform edge sand facies unit (F). The predominantly well-rounded (abraded) calcarenite grains indicate a high energy level and wave agitated bottom in the shallow, euphotic zone. Micritic infilling in some of the intraparticle pores may indicate poor winnowing, i.e. a temporally fluctuating energy level of the depositional environment. Based on the microfossil assemblage two facies-types have been recognized within this unit. *Skeletal grainstones with oncoids and intraclasts* (F.1) developed in Nosztor Valley (beds 16b and 26–28) and in Meleg Hill (bed 6). Beside a few autochtonous elements (dasycladacea, gastropods) it is made up mostly of fragments originating from the upper slope. *Foraminiferal intrapelsparites* (F.2) have been recognized in a few beds in Nosztor Valley (beds 15d–15e) and Meleg Hill (beds 5a–5b) sections. It is important to note that in this facies typical reef elements are also absent.

Tidal flat facies unit (G). This environment is typical in the Fődolomit but it was also observed in the uppermost part of the Barnag Member which can hardly be differentiated from the Fődolomit due to late diagenetic processes. It could be recognized both in the Nosztor Valley (bed 32) and Meleg Hill sections (beds 14). The following evidence could be diagnostic: i) intertidal algal-mats, and ii) black pebbles.

The conspicuous absence of reef-building organisms such as corals, sponges and calcareous algae in the high energy environments suggests that there were no such intensively colonizing forms in the shallow, euphotic zone on the platform

Table 1
Summary of the platform-basin lithofacies based on the studied outcrops.

Facies units Facies code	Shallow basin A	 B	Slope C
Sub-unit	Restricted basin with hypersaline, stagnant, anoxic-dysoxic bottom water	Shallow basin with normal salinity and open circulated water	Lower foreslope
Facies types	A.1) cyclic black laminites: bituminous or clayey lmst, A.2) ribbon lmst with interbedded shales and tempestites: normal graded ostracod and mollusk lumachella, A.3) dolomite beds with vugs after evaporites and fragments of algal-mats and pkst/grst intraclasts, A.4) lmst (wkst) forms lenses or interlayers in laminites, A.5) bivalve coquina in fecal pelletal lmst	B.1) skeletal mudstone with allochtonous, poorly preserved fragments of forams and echinoderms, B.2) mdst with pectinids, B.3) mdst/wkst with well-preserved fragments of autochtonous microfossils: ostracod and mollusk filaments and coprolites	Thin to medium-bedded clayey lmst, calcareous marls and bioclastic wkst/pkst C.1) clayey lmst, nodular and thin-bedded with clay intercalations, C.2) medium-bedded lmst, peloidal basinal matrix with reworked exotic lithoclasts-bioclasts (from envir. E), wkst
Color	Black to light gray	Brown, yellowish-brown	Yellowish-brown, light gray
Bedding and sedimentary structures	Laminites often form small-scale slumps, slides or cracks. It often encircles skeletal wkst lenses. Small-scale submarine erosional surfaces occur at the base of tempestites. Ribbon lmst has dissolved surfaces	Thin-medium bedded marls, calcareous marls or clayey lmsts. Usually intensively weathered and poorly exposed. Parallel or often nodular bedding. Bedding contacts are mainly abrupt with clay parting	Parallel bedding and often undulated thin layering. C.1) transitional contacts to less marly suites upsection In C.2) internal bedding structure is homogenized by bioturbation
Grains and biota	Hypersaline ostracods, plankton algae, submarine algal-bacterial mat, scarce fish-prints	Normal saline autochtonous ostracods, thin bivalve filaments. Very low percentage of resedimented fragments of echinoderms, mollusks	C.1) a few allochtonous, well-rounded and sorted bioclasts (ostracods, pelecypods, gastropods and echinoderms), some holothuroids, forams: *Gsollbergella-Agathammina* Association C.2) coated intraclasts and encrusted foraminifers, echinoderms, green algae and brachiopods. *Tetrataxis* foraminifer association.
Occurrence	Lower part of the Pécsely Mb. at Nosztor Valley and Nemesvámos	Lowest part of Pécsely Mb. at Nemesvámos, several beds in Nosztor Valley, in lower and upper parts of Meleg Hill section	C.1) in all three sections C.2) only in the upper part of the Nemesvámos section

Abbreviations used in the text: bdst – boundstone, mdst – mudstone, wkst – wackestone, pkst – packstone, grst – grainstone, lmst – limestone

Slope D	Slope terrace E	Winnowed platform edge sand unit F	Tidal flat G
Middle and upper foreslope			
D.1) medium to thick bedded lmst, clayey lmst, D.2) reddish-lilac dolomitized lmst-calcareous marl and dolomite	Medium bedded skeletal wkst/pkst. Often contains large chert nodules (10-20 cm)	Thin to medium-bedded skeletal grst/bdst. Often contains large chert nodules (10-20 cm). F.1) intraclastic-bioclastic-oncoidal grst/pkst. Well\ rounded grains, calcitized. Microbial envelopes and coated grains. F.2) intraclastic-peloidal grst	Medium to thickly bedded dolomitized lmst or dolomites with hardground surfaces and algal mats
Grey and reddish-lilac	Light to deep gray	Light brownish-gray	Gray to white
Parallel bedding, bedding planes in D.2) deposited over an erosional surface	Parallel or undulating bedding surfaces	Massive, parallel bedding contacts	Parallel bedding. Bed surfaces show hardground and loferite (birdseye) structure
Abundant allochtonous bioclasts (echinoderms, mollusks, forams), in some parts with large oncoids. Poorly sorted, moderately rounded. In D.2) the fragments are partly dolomitized	Well-preserved sponge spicules and foraminifer association: *Aulotortus-Triadodiscus-Nodosaria*	In general: forams-mollusks-echinoderms. In F.2) special foram association: *Glomospirella-Ophthalmidium-Trochammina*	A few algal-mats, black-pebble breccia, dolomitized pelloids (?) with subtidal elements, as green-algae (Dasycladacea) fragments. Pelecypods: Neomegalodon
In all three sections and facies D.2) only bed 31 at Nosztor Valley	Middle part of Nosztor Valley and lower and middle part of Meleg Hill sections	Middle part of Barnag Mb. at Nosztor Valley and lower and middle part of Meleg Hill sections (Barnag Mb.)	Uppermost part of sections at Nosztor Valley and Meleg Hill, "Fődolomit" Formation

margins during the deposition of the Barnag Member in the Early Tuvalian basin. This observation is in harmony with the fact that there are only a few occurrences of Tuvalian frame builders in the area of the Balaton Highland ("Ederics reef" and "Sédvölgy reef") described by previous studies (Góczán et al. 1983; Gyalog et al. 1986; Oravecz-Scheffer 1987; Csillag et al. 1995).

In the Southern Alps the estimated difference in the depth between the Cassian platforms and basins may have been 500–600 m (cf. Bosellini 1984). The rimmed shelf (Wendt and Fürsich 1980; Flügel 1982) was connected to the basin by a steep slope (at least 25–35; cf. Kenter 1990). A similar setting can be assumed for the Balaton Highland in the Middle Julian between the Ederics reef–Sédvölgy reef and the Veszprém basin (see Haas 1994; fig. 7c). Fine bioclasts and lithoclasts of platform origin were accumulated at the toe of the slopes (cf. Wendt and Fürsich 1980; Csillag et al. 1995). On the platform the sedimentation rate probably matched or exceeded the relative rise of sea level causing progradation; thus, shoaling upward sequences were formed in the basin (cf. Russo et al. 1992). At the end of the Julian, shallow basins were connected to the isolated platforms by gentle slopes, where a 100–150 m difference of depth can be assumed (Fig. 13). In the present case, a topography without significant Tuvalian reef organisms can be postulated which may have responded to the environmental changes of a widespread mild declinity. The organisms inhabiting the platform margin must have been predominantly binding and encrusting forms. This assumption may explain the very small amount of redeposited reef-colonizing elements on the slopes. The shallow, flat subtidal terraces may have hosted autochtonous foraminifers and siliceous sponges. The relative position of the slope terraces is ambiguous because they may have been located on both sides (inner and outer) of the sand shoals. The setting closer to the outer zone (basinward) is favored by the fact that a typical constituent of the E facies (*Aulotortus*, spicules) could frequently be found in the slope and basin facies. Moreover, in the shallowing upward succession the E facies is always overlain by the platform edge sand (facies F), which also supports its basinward setting. Taking into consideration the broad tidal flat was initiated in the Late Tuvalian the depositional setting may have been unified by that time (Fig. 13), but it was actually preceded by several complex processes, i.e. subaerial erosion, soil formation, etc., in the Middle Tuvalian.

Depositional models

Characteristics of terrigenous and carbonate depositional systems are significantly different. While terrigenous transport is generally enhanced by the fall of the sea level (drop of the base level), shedding of fine carbonate particles are intensified during highstands (Schlager 1991). The fundamental discrimination between them is the bioproductivity of the carbonate environments producing great amount of lime sand and mud. Moreover, the organisms inhabiting the shallow water struggle for light, resulting in rapid vertical expansion of build-ups.

During lowstands the carbonate platforms may have been subaerially exposed, leading to reduced bioproduction. The exposed areas may undergo subaerial weathering processes (Fig. 14A). During the subsequent transgression the reviving reef colonies cannot always keep pace with the sea-level rise, which resulted in drowning. However, according to observations in Holocene environments, the growth potential of the reef nucleus can generally match the rate of sea-level rise and if there is no unfavorable ecological offset (e.g. terrigenous influx) then carbonate accumulation is practically continuous in the nucleus (Schlager 1981).

Regarding the paleogeographical position in the Julian–Tuvalian, fine terrigenous sediments may have been transported as suspended particles from the Southern Alps to the area of the Veszprém Basin (Garzanti 1985; Haas 1994). The growth of the isolated platforms (Ederics and Sédvölgy platforms) could not keep pace with the sea level rise due to adverse conditions which reduced the growth potential (the transparency of the water was weakened by the contemporaneous siliciclastic supply). In most cases, platforms (or parts of them) drowned and pelagic marls accumulated above shallow marine carbonates (Fig. 14B). Continuous shallow marine carbonate deposition could be observed only on certain uplifted areas that were not affected by such negative factors (certain areas of the Keszthely Mts. – Ederics Limestone – and eastern Bakony Mts. – Sédvölgy Dolomite; cf. Haas and Budai 1995). When the stand of the sea level reached its climax during the transgression/highstand periods most of the carbonate sediments were shedded onto the slopes. The vertical growth of the buildups matched or exceeded the rate of sea level rise and the platform prograded basinward (Fig. 14C). During that process the fragments of platform fossils were spread out to the slope and basin (cf. Reijmer and Arnout 1993).

In the opinion of the author the black laminites of the Pécsely Member in the Nosztor Valley section were deposited in a shallow basin which became restricted during the lowstand in the Late Julian, although other models suggested by several authors (Budai 1991; Budai and Haas 1997; Budai and Csillag 1998) cannot be excluded either. These previous studies interpreted their deposition during the previous highstand which resulted in the lower sequence boundary being assigned to a different stratigraphic horizon (fig. 3 in Haas et al. 1998). The model suggested by the present paper is based on the following assumptions: 1) the general bottom topographic setting was inherited from the Late-Anisian–Ladinian rifting, 2) the relative shallowness of the remnant basin at the beginning of the deposition of the Pécsely Member was a result of an advanced upfilling, 3) the fact that a similar specific ostracod assemblage indicating hypersalinity occurs in beds which relate to a lowstand period (based on sedimentological and lithological features in other parts of the Mediterranean area – Gerry et al. 1990), 4) assumed water stratification in a

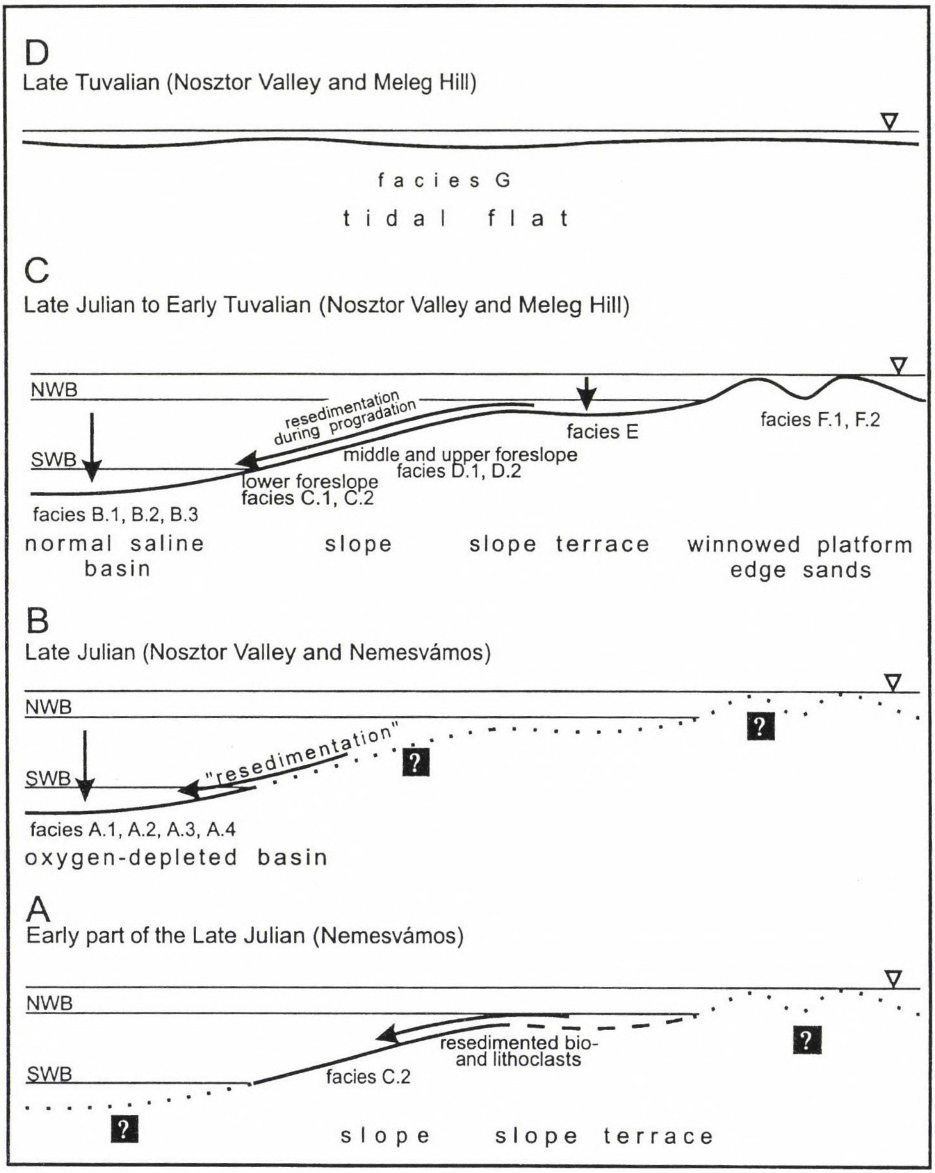

Fig. 13
General facies distributions of the Late Julian–Early Tuvalian intraplatform basin based on lithology and characteristic microfossil assemblages indicating energy level. Facies definitions were used in the sense of Wilson (1975) and Read (1985). Depositional setting schemes represent different chronological periods. Abbreviations: SWB – storm wave base, NWB – normal wave base. For facies code see Table 1. Dotted lines indicate probable distribution. Vertical arrows in B and C indicate mainly autochthonous sedimentation

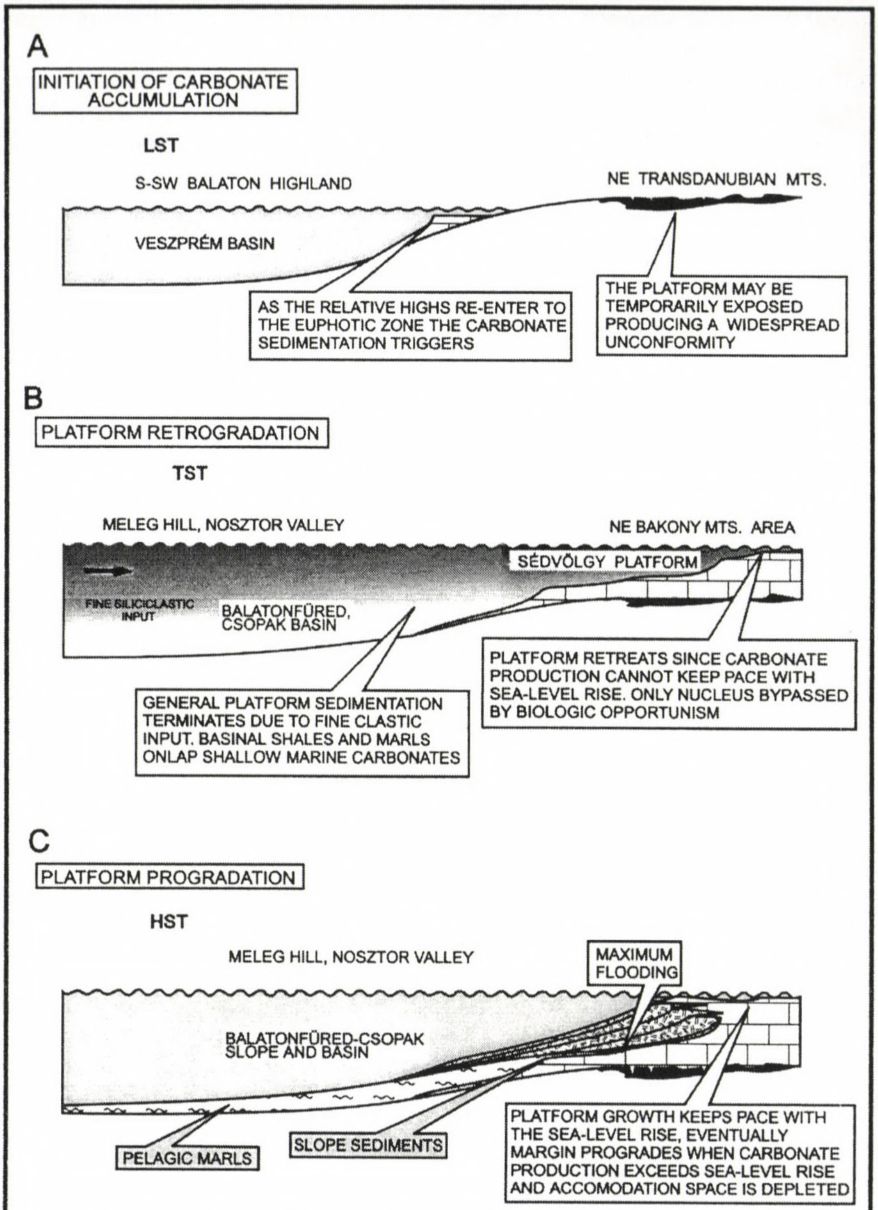

Fig. 14
General sedimentological model for the terrigenous effect introduced by Haas (1994) after Schlager (1991). This model also incorporates platform areas which were not a subject of the present paper but suggested by previous works (Haas and Budai 1995). For detailed explanation see text

stagnant basin, 5) subaquatic evaporite precipitation and 6) the high organic content of the laminites.

The Carnian sediments were deposited on a considerably dissected bottom evolved from the Ladinian extensional tectonics. Carbonate sedimentation prevailed on the highs emerging into the euphotic zone whereas pelagic marls were accumulated in the depressions. In the transgression and early highstand stages, the normal saline water could easily pass over the morphological sills (barriers) creating suitable conditions for benthos and plankton (Fig. 15A). The exact location of the barriers is still in debate; moreover, in some areas (western part of Balaton Highland) the laminitic carbonates are substituted by marls (borehole Bht-6: Csicsó and Pécsely Member; see Fig. 1 for location). In the Veszprém basin (with its depositional center at Balatonfüred) there was an open balanced circulation, i.e. the outflow equalized the inflow (Demaison and Moore 1980). In the Late Julian, a fall of the sea level resulted in the separation of certain parts of the basin (Balatonfüred and Nemesvámos area, Nosztor Valley–Sándor Hill outcrop, Barnag-2 borehole; see Fig. 1) and in these restricted basins water circulation ceased (Fig. 15B). The ambient platforms might have been subaerially exposed at the same time. In an arid climate the rate of evaporation would exceed the inflow, causing negative water balance, hypersalinity and water stratification. In the basin circulation terminated, accounting for stagnant bottom water and introducing anoxic-dysoxic conditions (def. Tyson and Pearson 1991). Hypersalinity is reflected in the specific benthonic ostracod assemblages (Monostori 1994). Contemporaneous deposition of evaporite crystals with carbonate particles can be interpreted by applying Schmalz's model (1969); however, the process could not develop up to the final phase (i.e. complete filling of the basin with evaporite salts) because the subsequent sea level rise in the Early Tuvalian opened the basin and the influx of normal saline water replaced the volume of that lost through evaporation.

Control of sea-level change on the deposition of the Sándorhegy Formation

Based on concepts and definitions of Sarg's (1988) and Van Wagoner's et al. (1988) basic works on sequence stratigraphy the following system tracts could be recognized in the three studied sections (Fig. 16).

Bituminous laminites in the lower part of the Pécsely Member were formed during a lowstand systems tract (beds 6–11 in the Nosztor Valley). Drop of sea level resulted in a restriction of the basins leading to anoxic-dysoxic conditions and hypersaline bottom water. The underlying clayey limestones and marls exposed in the basal part of the Nosztor Valley section can be considered as HST/TST of the previous sequence (beds 1–5). The contact is interpreted as a Type-2 sequence boundary (Van Wagoner et al. 1988) because no subaquatic hiatus can be detected. The base of the bivalve coquina bed exposed over the laminites (Nosztor Valley – bed 12) is defined as a transgressive surface, which can be related to the re-opening of the restricted basin. The abrupt rise of sea level with contemporaneous enhancement of terrigenous influx caused

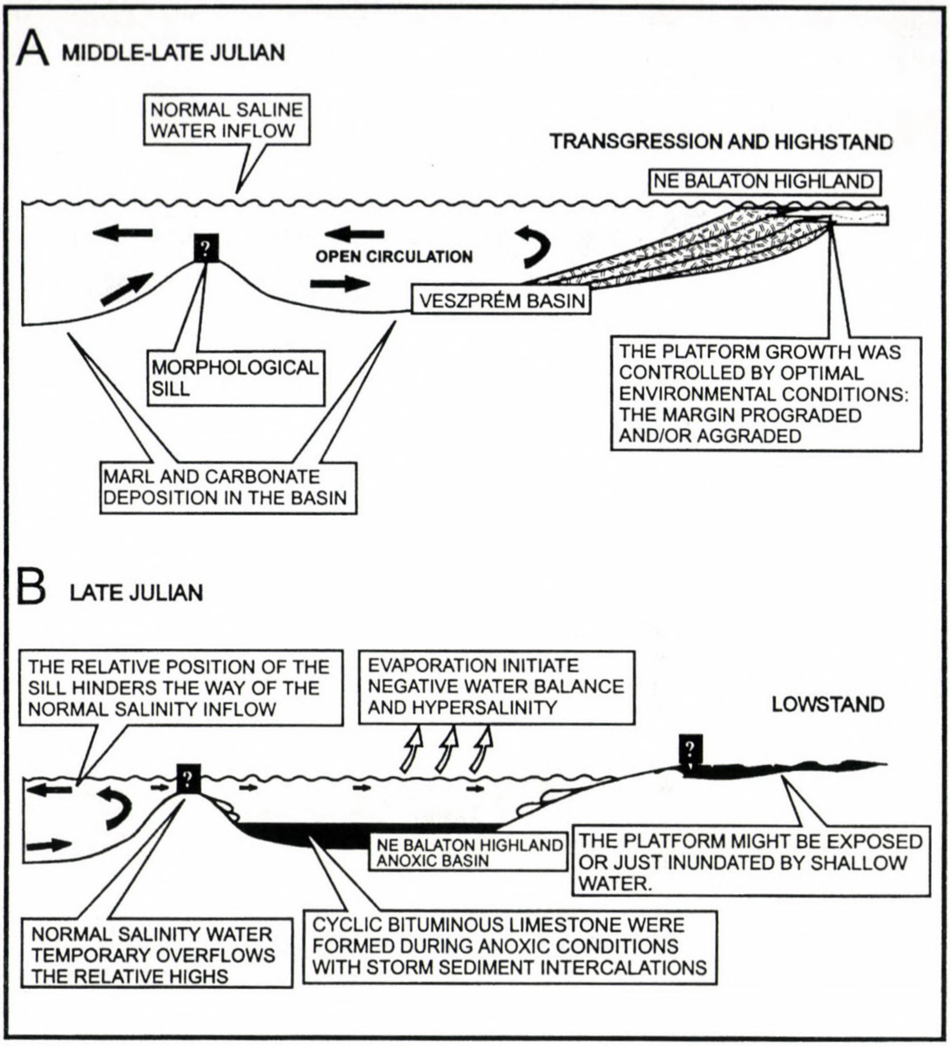

Fig. 15
General sedimentological model for the origin of the laminitic limestones in the lower part of the Pécsely Member. This model also incorporates dissected basin bottom morphology suggested by previous works (Budai and Vörös 1992, 1993). For detailed explanation see text

backstepping of the ambient platform (TST – bed 13). The subsequent highstand led to platform progradation (HST – beds 14–30). Between beds 23–24, another terrigenous suite occurs which can be attributed to higher order (4th–5th) cyclicity (Budai and Haas 1997). The upper boundary of the sequence is marked by a sharp erosional surface. The stratigraphic and genetic interpretations of the peculiar reddish-violet rocks between the erosional surface and the

Fődolomit are very doubtful. The author can only make inadequate assumptions about the depositional setting of these considerably dolomitized and poorly exposed layers. In thin sections, relicts of mollusk and algal-mat fragments could be observed which may indicate deposition during a progradational period and subsequently underwent terrestrial weathering process (Csillag 1991; Budai and Haas 1997).

The small section at Nemesvámos represents the Late Julian HST/TST (C.2) preceding the LST of the Pécsely Member (A). In the section at Meleg Hill the Mid-Tuvalian HST can be observed beds 1a–6f) but effects of higher order cyclicity in the upper part of the member (beds 7a–7d and 8a–12) can be assumed. The contact between the Fődolomit and the uppermost part of the Barnag Member appears to be undefinable in the Meleg Hill section due to pervasive dolomitization of the topmost part of the Barnag Member (Fig. 16).

In the Dolomites, the Upper Julian–Lower Tuvalian interval is represented by the Dürrenstein Formation (Pisa et al. 1980; Bosellini 1991). In contrast to the Sándorhegy Formation, it has been classified as a complete 3rd-order cycle (De Zanche et al. 1993) and interpreted as an aggradational, basin-filling mixed carbonate-terrigenous unit (Bosellini 1984) or simply the interior platform facies of the upper Schlern Dolomite (Biddle et al. 1992). It is mostly made up of dolomites, sands and arenites. The differences between the lithology of the Sándorhegy Limestone and the Dürrenstein Formation can be explained in the case of the latter by the proximity to the alluvial sedimentary systems (Schlager and Nicora 1979; Garzanti 1985; Bosellini 1991; De Zanche et al. 1993).

Conclusions

(1) Based on microfacies studies on three sections of a Carnian intraplatform basin in the Balaton Highland area the following paleo-environments have been reconstructed: A) shallow restricted basin, B) shallow basin with open circulation, C) lower foreslope, D) middle and upper foreslope, E) slope terrace, F) winnowed platform edge sand unit.

(2) Four characteristic foraminifer associations have been distinguished reflecting the biotope conditions: i) *Aulotortus-Triadodiscus-Nodosaria*: low energy, slope terrace; ii) *Gsollbergella-Agathammina*: low energy, unagitated basin; iii) *Glomospirella-Ophthalmidium-Trochammina*: high energy, wave-agitated winnowed sands; iv) *Tetrataxis*: slope or basin facies, low energy.

(3) Because of the huge amount of siliciclastic sediments transported from the hinterland the deep Ladinian basin evolved to shallow intraplatform basin with wide gentle slopes by the end of the Julian. Sea level fluctuation caused significant facies shifts on the gentle slope. Due to the progradation of the carbonate platform a shallowing upward facies trend, i.e. basin–slope–terrace–sand shoals, can be recognized in the basin succession.

(4) The bituminous limestone of the Pécsely Member was formed during a lowstand period. The morphological inequalities of the bottom (sills) may have

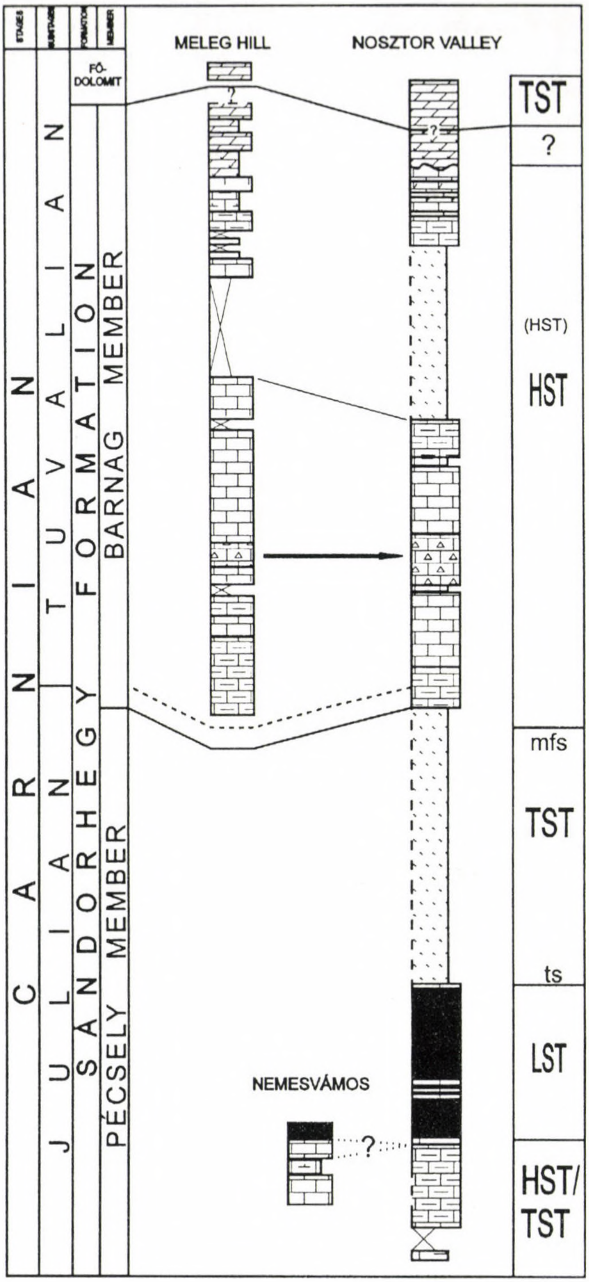

Fig. 16

Sequence stratigraphic interpretation of the three successions. Arrow shows horizon of lithological correlation (cherty limestone). Solid lines indicate lithostratigraphic boundaries defined by Csillag & Haas (1993), jagged line indicates the probable Julian-Tuvalian boundary defined by Góczán et al. (1991), dotted line indicates non-exposed or missing beds. Abbreviations: ts-transgressive surface, mfs-maximum flooding surface. For lithology see the legend on Fig. 3B played an important role in the restriction of the basin. The impact of the rapid sea-level rise (TST) and the ecological stress produced by the input of the fine siliciclastics resulted in retreating of the ambient carbonate platforms. In the basin, relatively thick marl with clayey-limestone intercalations represents the transgressive systems tract. Progradation of the platforms indicates the next HST while the following sea-level fall (LST) is clearly marked by an erosional surface (sequence boundary).

(5) The characteristics of the Sándorhegy Formation can be explained by four system tracts which can be correlated with the sequence of the Dürrenstein Formation in the Dolomites (Italy), although their lithology differs significantly.

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A Variscan monazite age from the Zemplin basement (eastern Western Carpathians)

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Monazites from a quartz-micaschist from the Zemplin Unit in southeastern Slovakia were dated chemically by means of the electron microprobe. The measured Th, U, and Pb contents suggest that the monazites in this rock formed during the Variscan period at 338±22 Ma. The most likely interpretation is that monazite growth occurred in connection with the amphibolite facies regional metamorphism, which is documented throughout the Zemplin Unit. Therefore, the obtained Variscan monazite age can be taken as a further argument to tentatively correlate the Zemplin Unit with the medium to high-grade Variscan basement of the Western Carpathians.

Key words: Western Carpathians, Zemplin Unit, Amphibolite facies metamorphism, Monazite, Chemical Th-U-Pb dating

Introduction

The Zemplin basement forms a small window (Fig. 1) within the Upper Carboniferous/Permian cover sequences in the southeastern part of the Slovak Republic, at the border with Hungary. It is located at the contact zone between the Western and Eastern Carpathians, where the tectonic relationships are still fairly unclear (Grecula et al. 1981a, b; Kisházi and Ivancsics 1988; Kobulsky et al. 1989; Lelkes-Felvári et al. 1996). Rb-Sr whole rock data from metasediments of the Zemplin Unit have been interpreted in favor of a Precambrian age of the basement (Pantó et al. 1967). Muscovites from the same samples gave ages of 394±52 and 450±130 Ma (Rb/Sr) and 260±10 Ma (K/Ar). The presumed Proterozoic age was one of the main reasons to tentatively correlate the Zemplin Unit with the Eastern Carpathian basement (Grecula et al. 1981a, b). Other authors (Slávik 1976; Rudinec and Slávik 1971; Mahel 1986) discussed a correlation with Variscan basement units in the central Western Carpathians (Cierna Hora or Branisko, Fig. 1). From the point of view of lithology and metamorphic conditions, the Zemplin Unit is indeed well comparable with other Western Carpathian units (Vozárová 1991; Faryad 1995). Judging from the sedimentary facies of the Upper Carboniferous/Permian cover sequences, the Tatric and Veporic units of the Western Carpathians and the Zemplinicum were part of an extensional basin at the end of the Variscan orogenic cycle, and

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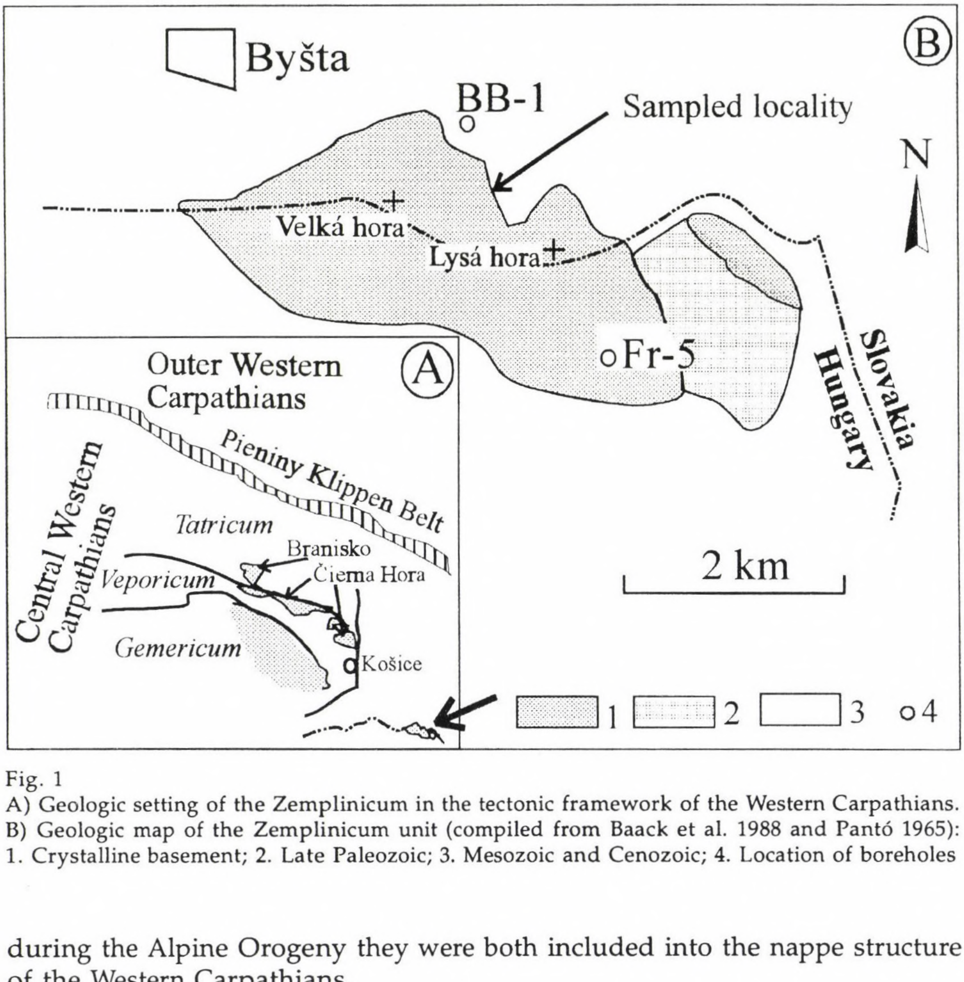

Fig. 1
A) Geologic setting of the Zemplinicum in the tectonic framework of the Western Carpathians.
B) Geologic map of the Zemplinicum unit (compiled from Baack et al. 1988 and Pantó 1965):
1. Crystalline basement; 2. Late Paleozoic; 3. Mesozoic and Cenozoic; 4. Location of boreholes

during the Alpine Orogeny they were both included into the nappe structure of the Western Carpathians.

In order to include the Zemplin basement in a reliable fashion within the regional tectonic concept, more comprehensive geochronological information is needed. We contribute to this goal by presenting here electron-microprobe monazite model ages for a quartz micaschists from the Zemplin Unit.

Petrographic background

The Zemplin basement comprises mainly metapelites, i.e. (quartz-) micaschists to muscovite-gneisses, biotite–amphibole gneisses, amphibolites and metagranites. Detailed petrographic and mineralogical data of these rocks are given in Vozárová (1991) and Faryad (1995).

The conditions of regional metamorphism in the Zemplin Unit were constrained at 0.6–0.8 GPa and 600–700 °C using garnet-plagioclase-amphibole and plagioclase-garnet-biotite-muscovite thermobarometry in amphibolites and gneisses, respectively (Faryad and Vozárová 1997). Some gneisses are partly migmatized. A later phase of mylonitization occurred under lower to middle greenschist facies conditions. It also affected the Permian cover sequences and is therefore considered an Alpine event.

The micaschists/muscovite gneisses suffered a particularly strong mylonitic overprint. They consist mainly of plagioclase, quartz and muscovite, but may contain pseudomorphs of biotite, sillimanite, staurolite or kyanite. Kyanite from muscovite gneiss was reported by Pantó (1965) and Kisházi and Ivancsics (1988) from boreholes, and by Magyar (1969) from the Lysá hora area (Fig. 1B).

The monazite-bearing sample of a quartz micaschist (ZEM-G1) that has been used for dating was collected from a small outcrop along a forest path on the northern flank of the Velká and Lysá hora hill. The rock is strongly mylonitized and consists of quartz, white mica, chlorite, some plagioclase and accessory tourmaline, zircon, monazite and Fe–Ti phases. Quartz shows undulatory extinction, deformation bands, deformation lamellae and incipient boundary recrystallization. Muscovite forms fish-like flakes, which are re-deformed and partly replaced by fine-grained white mica. Some muscovites contain numerous Fe–Ti phases and seem to have replaced earlier biotite. The fine-grained micas are intergrown with small amounts of chlorite and Fe–Ti phases and follow the schistosity in the rock. Tourmaline forms up to 0.5 mm zoned grains with small green cores and relatively wide brown rims. Fairly large yellow rutile, mantled by opaque phases, was also found. In contrast to zircon, which is oval to elongate with rounded edges and very probably detrital, monazite is isometric and partly of angular shape. The monazite grains were found as inclusions in recrystallized quartz, or adjacent to muscovite relics and appear to belong to the pre-mylonitic high-grade metamorphic paragenesis. About 10 grains were detected by BSE imaging. However, most grains were very small (< 5 μm).

Results of monazite dating

Monazite dating was carried out based on Th, U, and Pb analyses with the electron microprobe. Several recent studies have shown that this method commonly provides reliable monazite crystallization ages with moderate errors (e.g. Rhede et al. 1996; Cocherie et al. 1998). The theoretical basis of this chemical dating technique, its advantages and major risks are addressed in Suzuki et al. (1991) and Montel et al. (1996). The analytical procedure used in the present study follows the working routine presently established in the Salzburg laboratory, which is described in detail in Finger and Helmy (1998).

Four larger, isometric to slightly oval monazite grains with a size between 10 and 20 μm were selected for analysis. The spot (5 μm) was always placed into the grain centers. The analytical results are given in Table 1. The compositions of all four grains are similar. Th contents are moderate, and mainly

Table 1
Chemical data of the analyzed monazites from sample ZEM-G1 (grain centers)*

	m1	m2	m3	m4
SiO_2	0.50	0.23	0.37	0.57
P_2O_5	29.09	29.46	29.02	28.90
CaO	1.03	0.97	0.93	0.58
Y_2O_3	0.93	0.70	0.36	0.60
La_2O_3	13.52	14.05	13.89	14.36
Ce_2O_3	29.19	30.10	30.44	30.73
Pr_2O_3	3.31	3.24	3.32	3.56
Nd_2O_3	12.29	12.45	12.66	12.79
ThO_2	4.85	4.71	4.81	4.49
UO_2	0.75	0.74	0.66	0.61
PbO	0.11	0.10	0.10	0.09
Total	95.56	96.75	96.55	97.29
Si	0.020	0.009	0.015	0.023
P	0.996	1.001	0.993	0.985
Ca	0.045	0.042	0.040	0.025
Y	0.020	0.015	0.008	0.013
La	0.202	0.208	0.207	0.213
Ce	0.432	0.442	0.450	0.453
Pr	0.049	0.047	0.049	0.052
Nd	0.178	0.178	0.183	0.184
Th	0.045	0.043	0.044	0.041
U	0.007	0.007	0.006	0.005
Pb	0.001	0.001	0.001	0.001
A[9]	1.016	1.010	1.008	1.008
tetr.	0.977	0.983	0.988	0.988
Mo	90.19	90.72	90.85	92.74
Br	7.76	8.33	7.63	4.92
Hu	2.05	0.95	1.52	2.34

* Mineral formulas were calculated on the basis of 4 oxygens. The analyses have a slight deficit in the totals and slightly too low cation sums in the A[9] position, because Sm, Gd and the HREEs were not determined. Br and Hu are the theoretical percentages of brabantite and huttonite solid solution in monazite as recast from the Ca and Si formula units

represented by a brabantite component ($CaTh[PO_4]_2$). U concentrations are relatively high, as often observed in the case of metapelites. The LREEs show moderately steep patterns and Y contents are rather low.

Calculated model ages and errors are given in Table 2, including those obtained from the Salzburg laboratory standard F-5 (Finger and Helmy 1998). Using a defocused beam of 5 μm diameter, an accelerating voltage of 15 kV, a probe current of 250 nA and counting times of 240 s (Pb), 50 s (U) and 30 s (Th), the following precisions (2-sigma) could be obtained for each analysed point: 0.011 wt.% for Pb, 0.025 wt.% for U, and 0.04 wt.% for Th.

In Fig. 2, the data are graphically presented in a Pb vs. Th* isochrone diagram (Suzuki et al. 1991). All four analysed monazite grains gave Variscan model ages between 325 and 353 Ma. The weighted average age calculated from all analyses is 338±22 Ma and is interpreted as the age of monazite crystallization in the rock.

Discussion and conclusion

We believe that the obtained Variscan monazite model age approximately dates the amphibolite facies regional metamorphism in the Zemplin Unit. This assumption seems reasonable because it is known from many studies worldwide that monazite typically forms during the prograde phase of amphibolite facies metamorphism of metapelites (e.g. Franz et al. 1996).

Table 2

Th, U, Pb contents (wt.% elements), Th* values and model ages (MA) of the analyzed monazites from sample ZEM-G1 and from laboratory standard F-5

STANDARD F-5					
m1	5.738	0.517	0.113	7.422	**341 ± 36**
m2	2.350	0.547	0.065	4.131	**354 ± 65**
m3	0.863	0.222	0.025	1.585	**347 ± 170**
m4	1.252	0.232	0.030	2.007	**333 ± 134**
m5	5.764	0.526	0.111	7.474	**333 ± 36**
					339 ± 23
SAMPLE ZEM-G1					
m1	4.264	0.660	0.101	6.413	**353 ± 42**
m2	4.141	0.651	0.094	6.260	**336 ± 43**
m3	4.227	0.583	0.089	6.123	**325 ± 44**
m4	3.946	0.539	0.085	5.698	**335 ± 47**
					338 ± 22

Model ages were calculated after the method of Montel et al. (1996), Th* values have been recast with these model ages using the equation given in Suzuki et al. (1991). F-5 monazites are from a fraction dated by U/Pb mass spectrometry at a concordant age of 34±12 Ma (Friedl 1997)

Low yttrium abundances in monazites, as observed also in the analyzed grains from sample ZEM-G1 (Table 1), are often taken as an argument for low-T (upper greenschist facies) monazite formation, according to the geothermometer of Heinrich et al. (1997). However, because of the lack of coexisting xenotime in the sample, this thermometer cannot be reasonably applied in the present case. XRF data indicate very low Y whole rock contents of only 8 ppm for sample ZEM-G1, what may have been the actual limiting factor for Y incorporation into the monazite lattice, independent of the formation temperature. Of course we cannot completely rule out the possibility that the monazite ages date a phase of Variscan retrogression of an older high-grade metamorphic rock. However, in the absence of textural evidences for such a process, we consider this rather unlikely. As mentioned, the retrograde mylonitic fabrics in the rock are very probably Alpine and not Variscan in age (see above).

There can be little doubt that the dated rock is a true part of the Zemplin basement, although the quartz-micaschists occur essentially in the upper parts of the Lysá and Velká hora hills, and graphite-bearing gneisses were found along a tectonic zone between these two hills (Fig. 1). Close relationships between quartz micaschists and biotite gneisses and amphibolites are documented in other parts of the Zemplin Unit. For example, similar mica and quartz-rich mylonitized rocks were also found under Badenian sediments in borehole BB-1 (Vozárová 1991), which was drilled about 500 m west of the sample locality of this study. They contain there intercalations of biotite gneiss, and in the deeper part of the borehole amphibolites occur. Likewise, partly mylonitized quartz-muscovite gneisses/micaschists were described by Kisházi and Ivancsics (1988) from the Hungarian side of the Zemplin Unit in borehole Fr-5 (Fig. 1B).

In our opinion, the new Variscan monazite age provides a further argument to tentatively correlate the Zemplin basement with other Western Carpathians

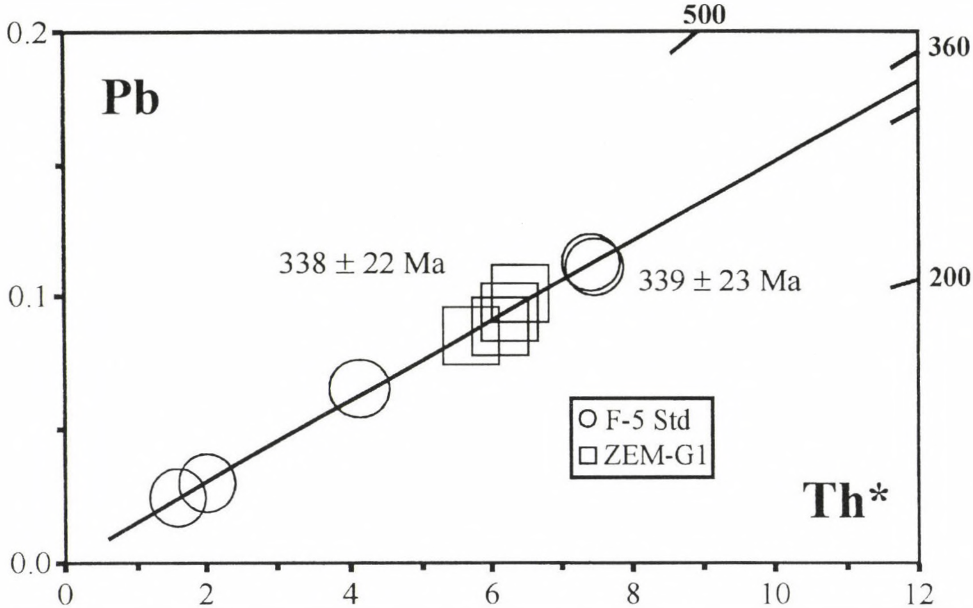

Fig. 2
Total Pb vs. Th* isochrone diagram after Suzuki et al. (1991) using the values given in Table 2. Time scale shown is based on the position of zero-intersect isochrones. Drawn isochrone refers to the calculated weighted average of 338 Ma for sample ZEM-G1. Size of symbols corresponds approximately to the 1σ error of the Pb analysis

basement units. Nevertheless, more geochronological data are needed to fully understand the metamorphic evolution of the Zemplin Unit before a reliable decision about the tectonic relations between Zemplinicum and Eastern and Western Carpathians can be made.

Acknowledgements

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Is there any proof for Quaternary tectonism in the Paks area, Hungary? (In connection with the publication of the volume[1] "Seismic safety of the Paks nuclear power plant")

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Data on the Quaternary tectonism of the Paks NPP area have not been included in the seismic hazard assessments; nevertheless, they are important factors of the realm of scientific and non-scientific discussions on the subject. After intensive studies, geophysicists believed that only fault-related Quaternary tectonism had been proved. Ráner et al. (1997) do not go beyond the statement itself whereas Tóth and Horváth (1997) support their opinion by structural analysis of seismic sections. In those sections, however, where the possibility of fault-related Quaternary tectonism appears, the presence of Quaternary sediments is doubtful. In turn, in those sections where Quaternary sediments are definitely present and their stratification is clear, they are definitely not faulted. The boundaries of the significant Quaternary depressions are located above flexures in underlying Pannonian (Late Neogene) sediments. This is probably due to the fact that in time sections the lower velocity of the Quaternary sediments "pushes down" the stratification of the Pannonian sediments. It can be concluded that there is no real proof of fault-related Quaternary tectonism in any of the seismic sections. Other studies (of joints in outcrops and trenches, of boreholes, of geomorphology, etc.) did not reveal traces of such tectonism either (Balla et al. 1997; Marosi and Schweitzer 1997) so that it seems possible that there was no fault-related tectonism of Quaternary age at all in the area.

Key words: depressions, faults, flexure folds, fluvial sedimentation, interpretation, meander, Quaternary, seismic risk, seismic profiles

*

Defining the seismic hazard for the Paks nuclear power plant is a critical question for Hungarian – and not only for Hungarian – society; its study has lasted for about two decades. Since earthquakes are related to faults, information on the activity of the faults and their distribution is needed for the hazard assessment. Therefore, the Hungarian seismic hazard studies had two main objectives: distribution and parameters of *earthquakes*, and of *recent tectonic movements*.

[1] Marosi, S., A. Meskó (Eds): Akadémiai Kiadó, Budapest, 1997, 193 p.

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So far the relationships between earthquakes in the Carpathian area and concrete faults have not yet been elucidated in a model convincing enough to be integrated into seismic hazard assessments; hence these assessments have been performed independently of any fault network. In such a situation, any considerations of the recent fault-related tectonic movements have remained a coloring element with no real effect. However, parameter values which *quantitatively* describe the seismic hazard of the nuclear power plant for the engineers who plan and perform the technical protection are one thing. A completely different thing is any conclusion on Quaternary tectonic activity which only shows the seismic hazard to be *qualitatively* higher or lower, thus affecting primarily the atmosphere of scientific and lay discussions.

From both points of view, the publication of the volume "Seismic safety of the Paks nuclear power plant" is of great importance. In the presented papers, however, only studies on the recent fault activity will be addressed. Two of them (Balla et al. 1997; Marosi and Schweitzer 1997) do not see any convincing traces of such activity, two others (Ráner et al. 1997; Tóth and Horváth 1997) believe that geophysical studies have provided sufficient proof of this activity. The three other works (Chikán et al. 1997; Szeidovitz and Varga 1997; Tóth and Mónus 1997) did not express any opinion on this question.

Among the geophysicists, both Ráner et al. (1997) and Tóth and Horváth (1997) believe that the high-resolution seismic sections prove Quaternary fault activity. Balla et al. (1997) disagree with that, also based on high-resolution seismic sections. In this question, instead of the "vehement debate" mentioned by Gúthy (in Ráner et al. 1997), all that has happened is a firming of opinions, mostly in manuscripts unavailable to the public. The authors of the studies in the volume did not know of each others' most recent results, and in the studies cross-references are practically absent. In order to fill this knowledge gap, the author of this paper thinks it necessary to say some words about how convincing the proofs of Quaternary tectonic activity are in the two mentioned studies.

In the Summary[1] by Ráner and Szabó, one can read that "the geophysical investigations verified the strike slips associated with Quaternary compression" although "the geophysical investigations have failed to clarify the age of the youngest movements". In the entire study, however, only the description of the high-resolution ("shallow") seismic survey[2] given by Gúthy can be related to this statement. In this description, following the declaration that the question is unresolved, one can read: "The task was to separate the effects of ground roll, lithological changes within the Pleistocene–Holocene sequence and structural elements. An unambiguous solution of the problems associated with the structure of the Pleistocene–Holocene sequences is still being awaited". Although there is no unambiguous reference to this, in Fig. 1 a fault affecting

1 Ráner et al. 1997, p. 74.
2 Ráner et al. 1997, pp. 70–71.

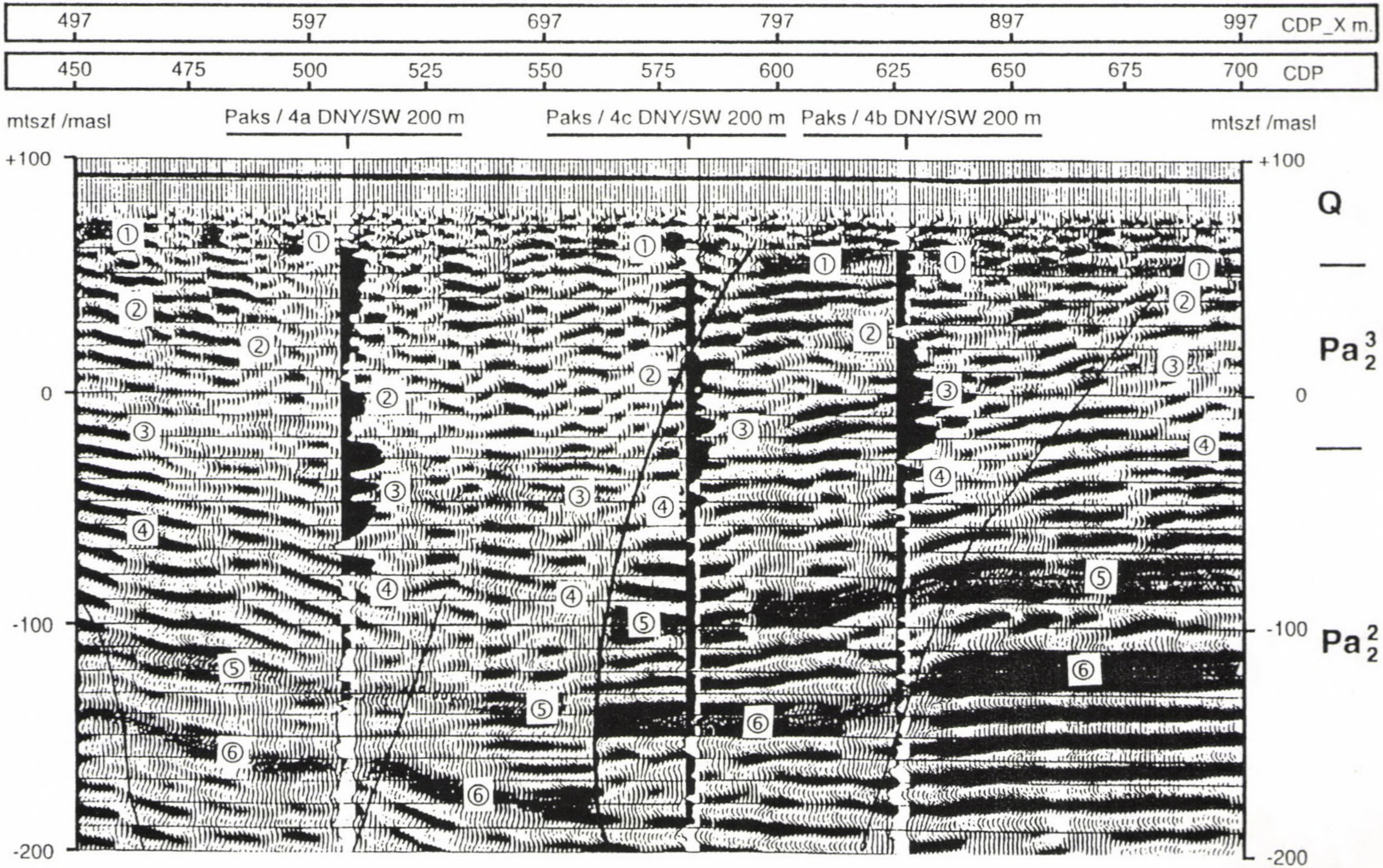

Fig. 1
Fragment of the seismic depth section Pa-15 with the well logs of boreholes Paks-4a, Paks-4c and Paks-4b (Ráner et al. 1997, Figure 21). Some refractors are marked in colors (with no captions) as follows: 1 = magenta, 2 = light blue, 3 = greenish yellow, 4 = yellow, 5 = green, 6 = violet

the base of the Quaternary is indicated. As is especially well seen in Fig. 1, both the base of the Quaternary and the offset in it are rather unconvincing; moreover, in the northwestern[1] limb of the "fault", a poor signal zone and the trough-like depression are visible. They appear to follow on the surface an ox-bow of the Danube (Fig. 2) which in turn brings phenomena of non-tectonic

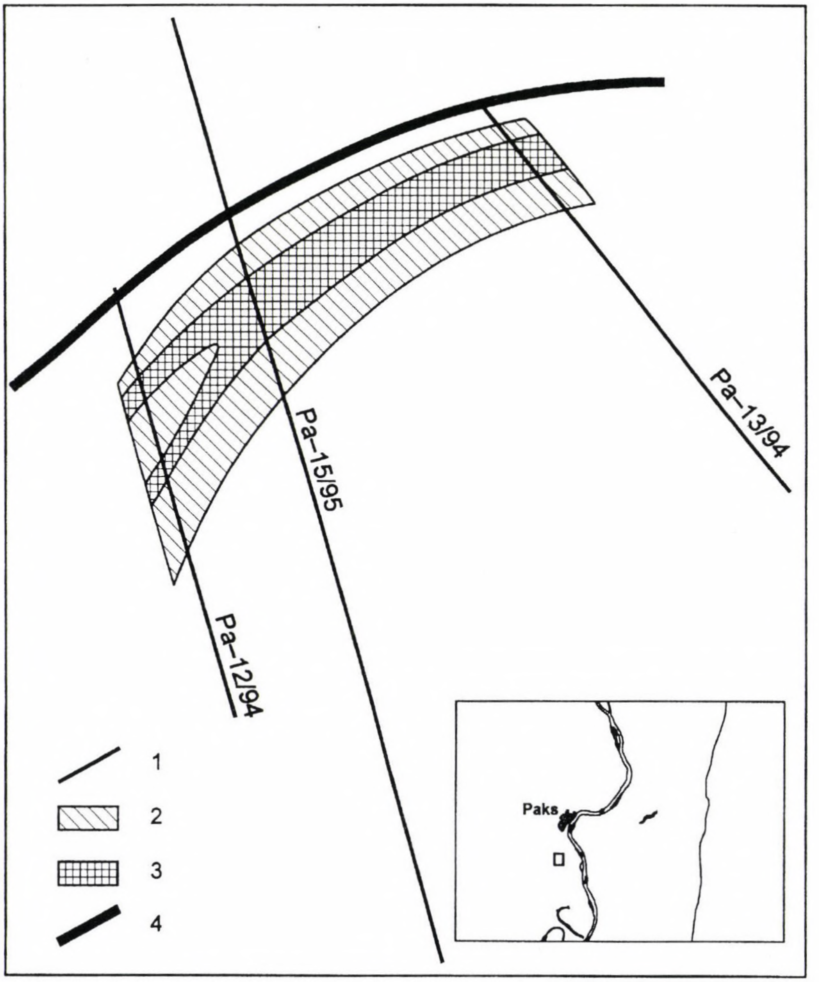

Fig. 2
Connection of the trough-like depression in the reflection pattern of the Quaternary sediments and poor signal zone below it, with a Danube ox-bow (Balla et al. 1997, Figure 10). For location, see insert. 1. high-resolution seismic section; 2. poor signal zone; 3. trough-like depression; 4. outer edge of the Danube ox-bow (Balla et al. 1997, Fig. 10)

1 Left in the section.

origin into the section. Thus, the material demonstrated in the study in question proves neither "Quaternary compression" nor "associated strike slips".

The study by Tóth and Horváth (1997) expounds with imposing thoroughness the recent seismic survey (Fig. 3) performed with modern techniques and its processing as well as the reprocessing of earlier sections and interpretation of the results. Therefore, it is easy to recognize which conclusions are sufficiently verified and which are not. Faults in the sections have been identified in the high-quality seismic image of the Pannonian (Late Neogene) sequence, and

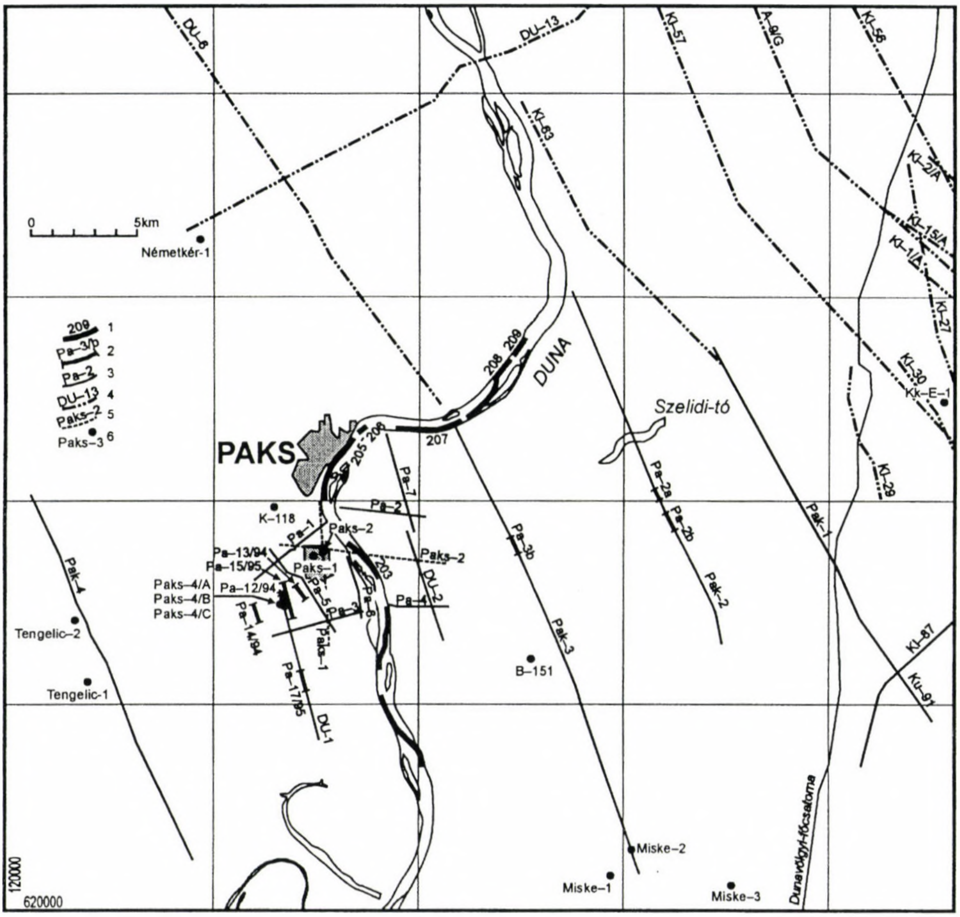

Fig. 3

Location map of seismic lines (Tóth and Horváth 1997, Figure 3). 1. multichannel reflection seismic section measured on the Danube; 2. high-resolution reflection seismic section measured by ELGI; 3. reflection seismic section measured by ELGI; 4. reflection seismic section measured for the oil industry; 5. refraction section; 6. borehole tectonic phenomena in the Quaternary sediments have been related to these faults.

It is pointed out that the fault-related Quaternary movements do not necessary result in faults crossing the Quaternary sediments and approaching the surface. This is probably true, and causes difficulties in recognizing recent fault activity. Following that way of thinking one results in assuming that Quaternary rejuvenation is possible for *all the existing faults*. This is reflected in the recommendation by the IAEA not to locate nuclear power plants above known faults. However, when the Paks NPP was sited this requirement did not yet exist, and the faults below it were unknown. It became clear much later that a fault, definitely crossing Pannonian sediments[1], is situated below the nuclear power plant. As a consequence, the question here is not whether movements are *possible* or not, but whether movements in the recent geologic past can be *proved* or not and whether they resulted in faults reaching the surface or not (beside the shaking connected with earthquakes, these faults can be a hazard for the nuclear power plant since the shear on them can destroy buildings, pipelines, etc.). Therefore, in the following, proofs, not possibilities, will be discussed.

Tóth and Horváth (1997) consider two groups of tectonic features: faults and bends. They discuss the tectonic image of the base or lower part of the Quaternary sequence separately for each section. From the statements on the *faults* (for an overview, see Appendix) one can conclude that faults in the Pannonian sediments do not cross the base of the Quaternary in sections Danube-202, Danube-203 (Fig. 4), Danube-205 (Fig. 5), Danube-207 (Fig. 6), Danube-208 (Fig. 7), Pa-2a and Pa-2b but may cross it in sections Pa-3b, Pa-12 (Fig. 8a), Pa-13 (Fig. 8b), Pa-14 and Pa-15. Such a "crossing" is unconvincing in all cases since the faults indicated in the Pannonian sequence indeed *continue* out of the sections upward, but the presence of Quaternary sediments in those levels is far of being obvious although of course it cannot be excluded either.

It should be noted that in sections Pa-12 (Fig. 8a), Pa-13 (Fig. 8b) and Pa-15[2] push down of the near-surface reflectors is visible after reprocessing, and that in the first two of them the accompanying poor-in-signals, which in maps follow a strongly-bent Danube ox-bow (Fig. 2), i.e. surface or near-surface unevenness also was involved in their generation. Since the corresponding effects were not removed during reprocessing, the existence of faults in deeper horizons remains doubtful.

It should be mentioned that, in a study on neotectonics with attempts to prove the existence of *Quaternary*, moreover, *Late Pleistocene* faults, the "young faulting" in the captions of the drawn summary of tectonic features[3] is deceptive since it gives the impression that "young" is the synonym of *Quaternary* or *Late Pleistocene*. There are many faults of this type in the figure although most of the points marked as faults remain within the Pannonian even according to authors' identification.

Beside the faults, *bends* are visible in sections as well. Tóth and Horváth (1997) write[4] that "the Danube-203 and Danube-205 sections show that these young sediments are slightly deformed. The tectonic origin of this deformation

1 See in Figure 6 by Balla et al. (1997) faults B and B′; the latter is seen in Figure 18 by Tóth and Horváth (1997) as well.
2 Tóth and Horváth (1997), Figures 13, 14 and 16.
3 Figure 18 by Tóth & Horváth (1997)
4 Page 136, second last paragraph 4.

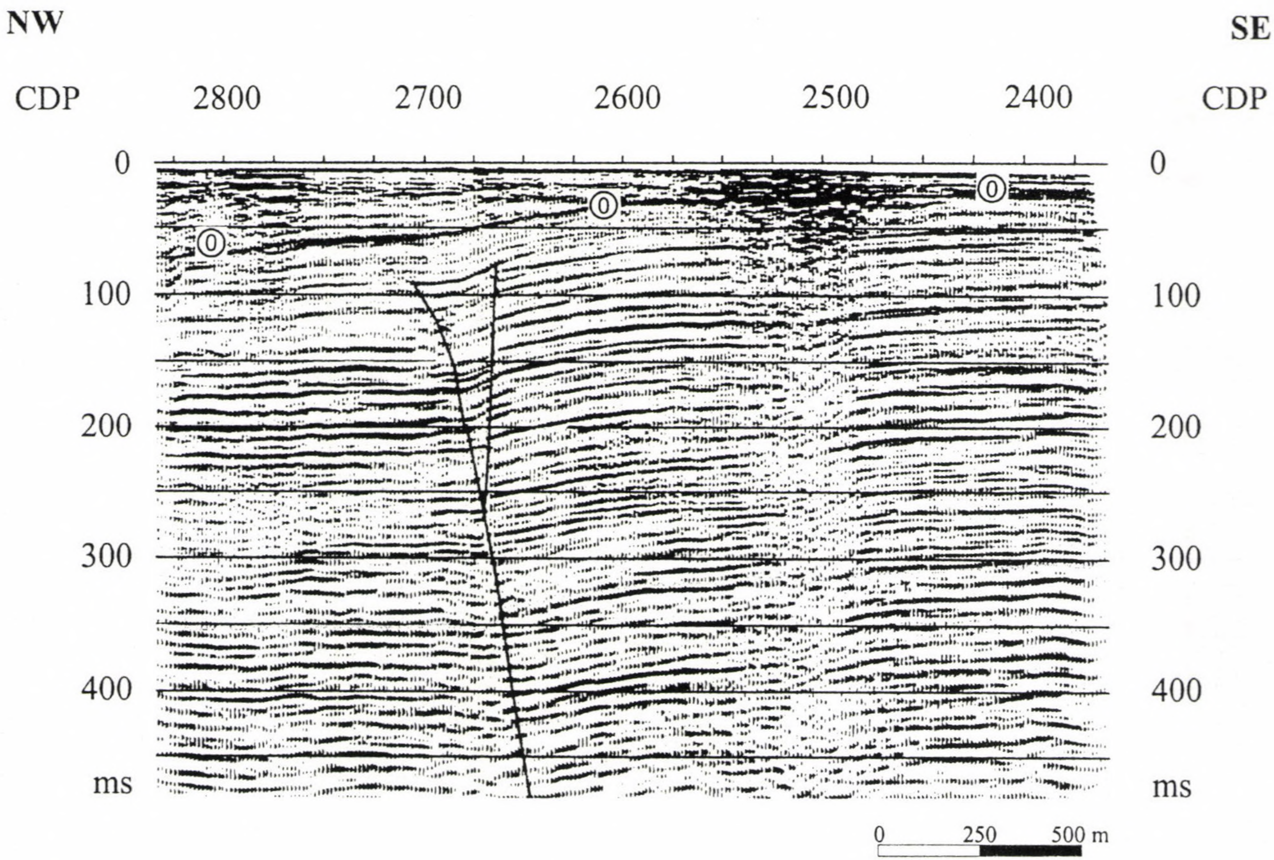

Fig. 4
Danube-203 interpreted time section (Tóth and Horváth 1997, Figure 6). ◎ = base of Quaternary (green in the original Figure)

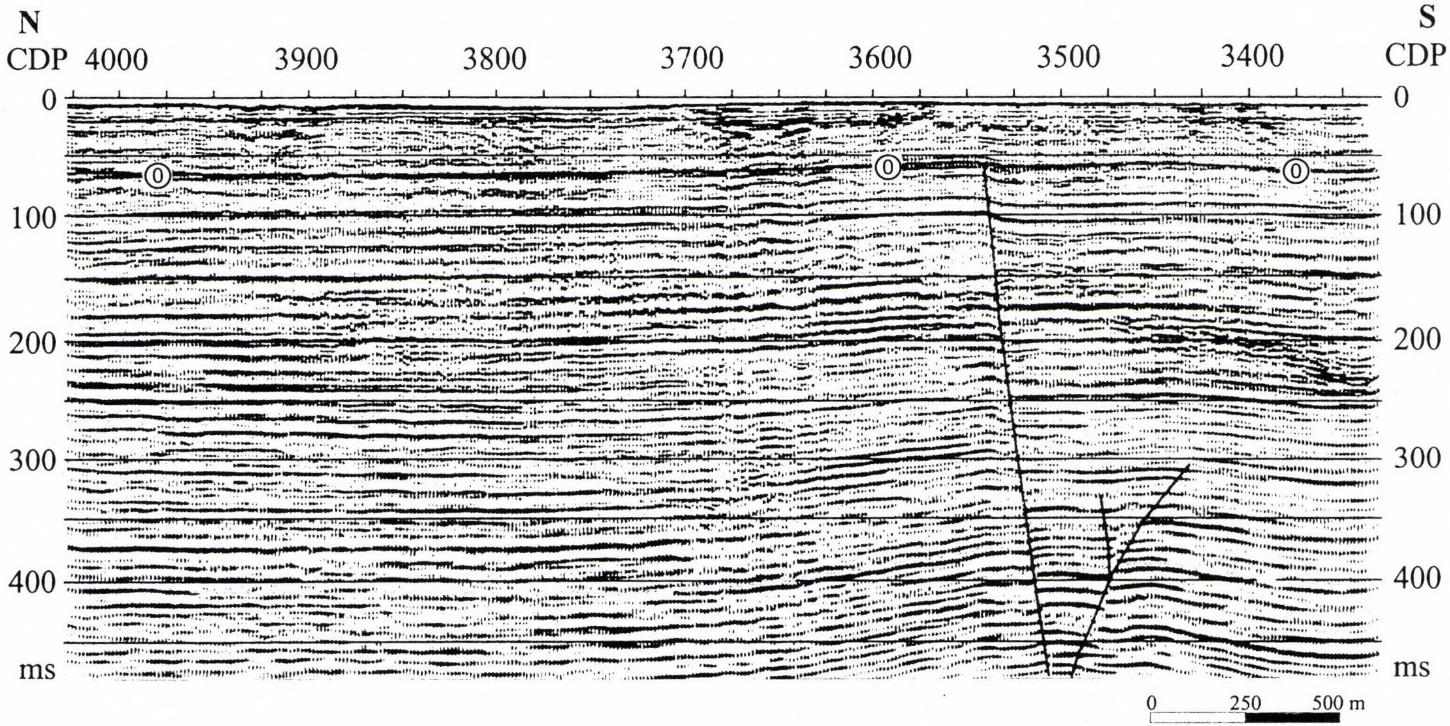

Fig. 5
Danube-205 interpreted time section (Tóth and Horváth 1997, Figure 7). ⓪ = base of Quaternary (green in the original Figure)

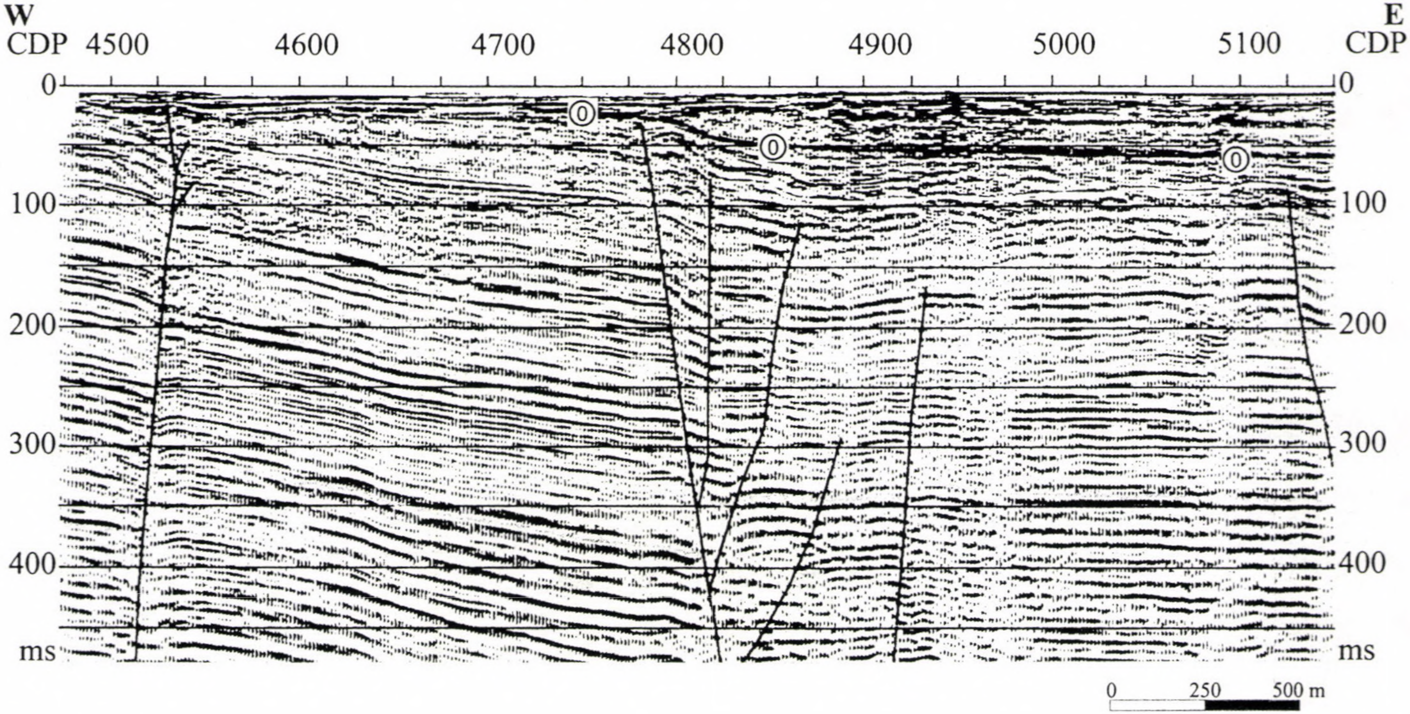

Fig. 6
Danube-207 interpreted time section (Tóth and Horváth 1997, Figure 8). ◎ = base of Quaternary (green in the original Figure)

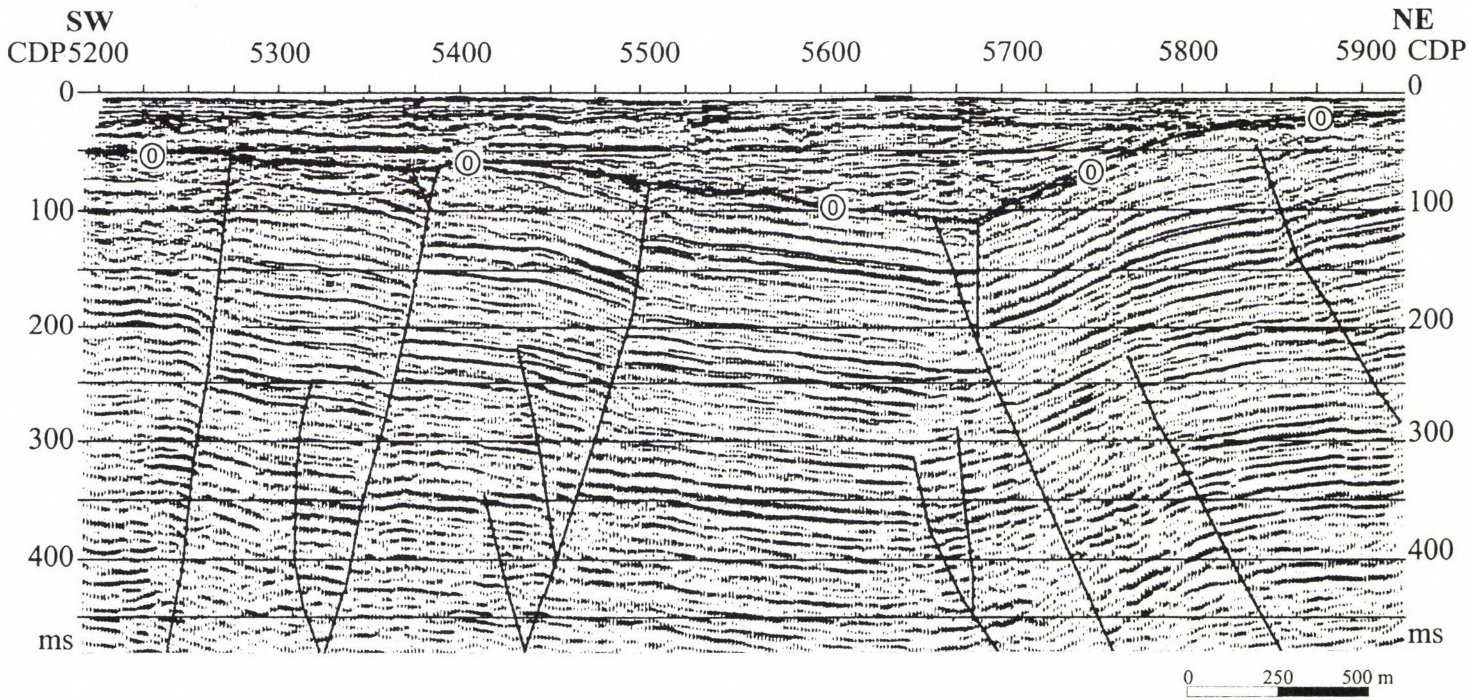

Fig. 7
Danube-208 interpreted time section (Tóth and Horváth 1997, Figure 9). ◎ = base of Quaternary (green in the original Figure)

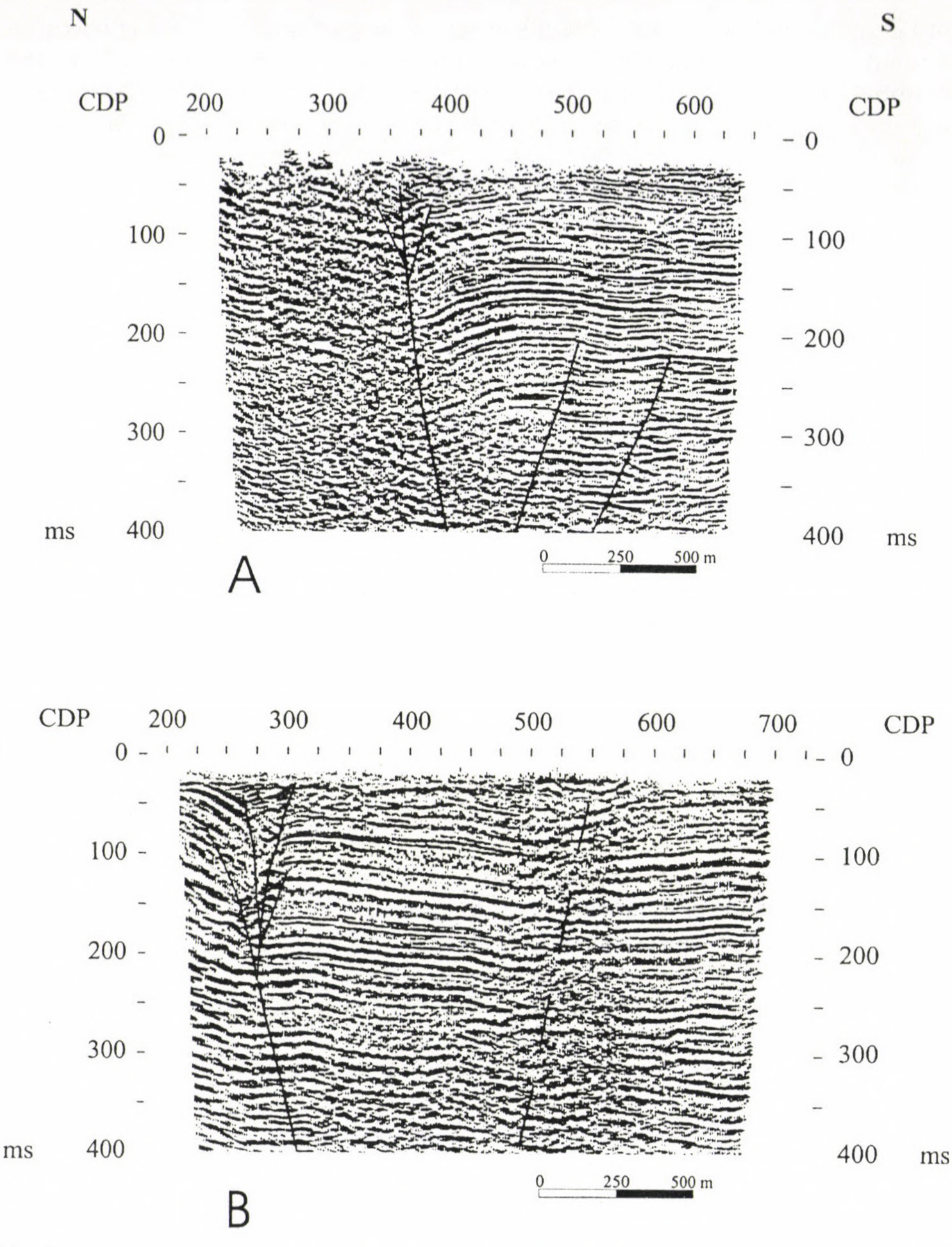

Fig. 8
Interpreted high-resolution time sections, A) Pa-12 (Tóth and Horváth 1997, Figure 13), B) Pa-13 (Tóth and Horváth 1997, Figure 14)

and genetic relationship to the faults observed in the Pannonian strata becomes evident on the Danube-208 section". Of these sections, however, in the Danube-203 one (Fig. 4), just above the fault indicated within the Pannonian sequence (and not reaching the base of the Quaternary) disturbances are visible in the Quaternary sequence which make it impossible to judge on the quiet or deformed shape of the stratification. In the Danube-205 section (Fig. 5), in turn, the base of the Quaternary sediments above the fault seems to be as horizontal and quiet as those parts farther away from the fault. One can believe that there were deformations in the original sections which disappeared in the course of reducing them in size, but the fact is that in the published sections even slight deformation is not recognizable.

In Fig. 7 an asymmetric depression is conspicuous, with the base of the Quaternary following the stratification of the Pannonian sequence. The authors were probably referring to this phenomenon.

It seems possible that they thought about stratification disturbances as well; the latter, in the Danube sediments, however, are not only visible above the faults in the Pannonian but also far from them (*e.g.* around CDP ≈ 5530, 5610 or 5780), and their image cannot be interpreted in terms of tectonic origin.

The relationship between the *flexural bend* in the Pannonian strata and shape of the base of the Quaternary is not a peculiarity of Fig. 7 (CDP ≈ 5690–5780); a similar feature is visible in Figs 5 (CDP ≈ 2680–2600) and 7 (CDP ≈ 4780–4860), as well. In turn, there is no sharp Quaternary thickness change in the Danube sections with any underlying flexures in the Pannonian. At the same time, it is an important fact that the Quaternary sediments are *horizontally* layered in each section, even above their bent base.

For the geologic interpretation it should be assumed that any flexure in the Pannonian strata resulted in a similar flexural bending of the earth's surface, which was present as a depression during Quaternary accumulation but did not evolve any further (Quaternary sediments were not deformed and remained horizontal). This interpretation seems to be only acceptable with a caveat: *there was not enough time* between the generation of the flexure, i.e. of the depression, and filling it with sediment, for the cutting of the topographical bend by *fluvial erosion*. The probability of this point needs to be verified.

In Fig. 14 of Marosi and Schweitzer (1997) it is well seen that the Holocene Danube in the Paks area formed meanders in a strip at least 10 km wide, whereas the width of the depression in Figure 7 does not exceed 2 km (corresponding sizes in the two other sections are not larger, but one flank of the depression only is visible in any of them). If one regards the flexures facing each other in Figs 6 and 7 as two flanks of the same depression (and the depression mentioned in Fig. 7 being part of it), the entire width of the latter does not exceed 5 km remaining completely within the meander strip. One can add that, at Paks, the Holocene Danube undercut the loess wall (and the present-day Danube would also undercut it if it were not regulated) with high levels of the Upper

Pleistocene (not separated from the Danube valley by any fault or flexural bend[1]), and this gives a feeling of what fluvial erosion is in space and time.

The age of the deeper levels of the Quaternary sequence in seismic sections is still an open question, but it is clear that they originated approximately under the same conditions as at present day (Jaskó and Krolopp 1991; Marosi and Schweitzer 1997). Therefore, the geologic history outlined above generates deep doubts: in the Danube valley, a surface depression is hardly imaginable with its lateral borders not immediately cut by the erosion accompanying the fluvial accumulation. On the other hand, sedimentation synchronous with the flexure generation should have resulted in bends within the deepest Quaternary horizons.

Consequently, it is necessary to return to the section and to ask whether indeed no mistake can be found in those elements of the seismic image which led to the above conclusion. The sections in question are *time sections* whereas geologic interpretation is only valid if performed upon *depth sections*. Thus, one must question whether the time sections perfectly reflect the depth conditions. The answer depends on velocity distribution; unfortunately, although Tóth and Horváth (1997) performed velocity analyses[2], they did not publish the results. Therefore, only general comments can be made here.

There is a big difference in mechanical properties between the Quaternary fluvial and Pannonian deposits. The former provides loose, disintegrated cores and the second, solid, constant-shaped ones (Balla et al. 1997, p. 41); it is to be expected that seismic velocities are much lower in Quaternary fluvial deposits than in Pannonian sediments. This may result in reduced depths (in meters) where the base of the Quaternary is at a deep level, e.g. 50–60 ms (Figs 4 and 6), unlike where it only is at a 20 ms level, since the higher velocity in the Pannonian sediments at the 20 to 50–60 ms range increases the depths in meters. In other words, where their thickness is greater the lower velocity of the Quaternary sediments may "push down" the deeper reflectors.

The probable velocities in the Quaternary sequence can be estimated from *thickness* data. The maximum thickness of the fluvial sequence in different sections is as follows: 50 ms in Danube-202, 70 ms in Danube-203 (Fig. 4), 65 ms in Danube-205 (Fig. 5), 55 ms in Danube-207 (Fig. 6) and 110 ms in Danube-208 (Fig. 7).

The first two sections fall within the Kalocsa Depression. Near the Danube-202 section is located one of borehole sections in Marosi and Schweitzer[3] (1997). The mean Danube water level in it is around 87 m a.s.l., and the base of the Quaternary fluvial sediments approximately 47 m below it. In the map (Fig. 9) the base of the fluvial Quaternary is at 45 m a.s.l. on

1 Marosi and Schweitzer (1997), Figure 13.
2 Page 129, item 8.
3 Their Figure 8.

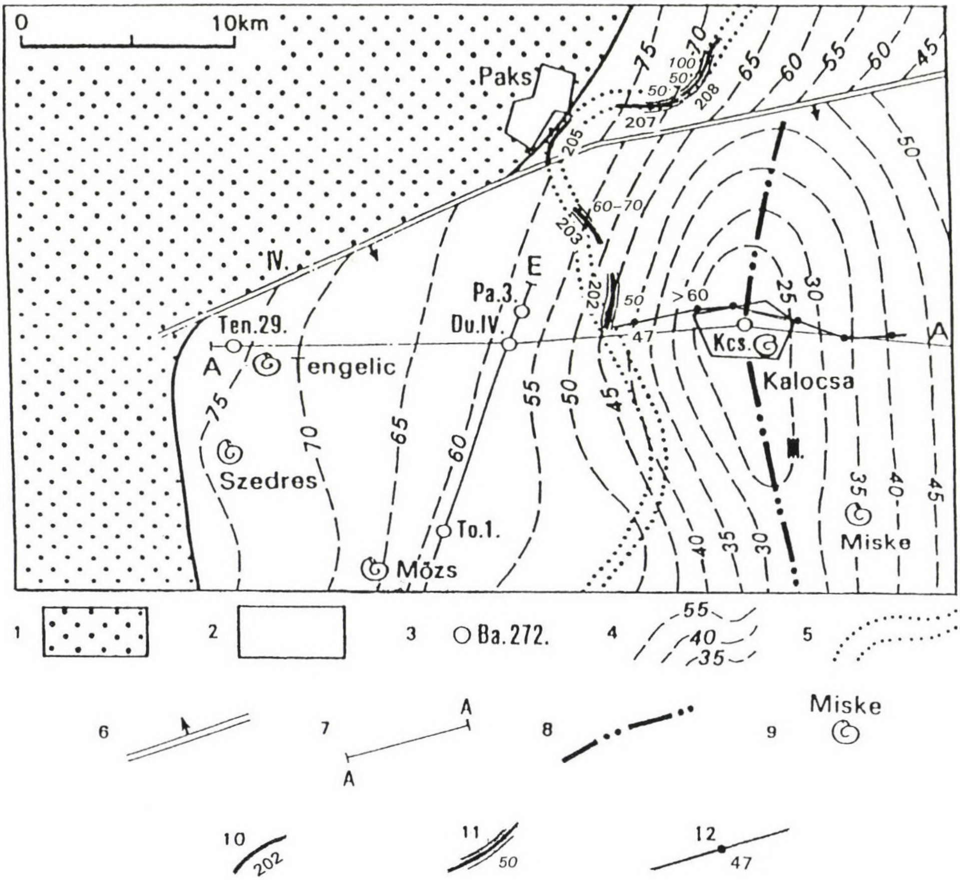

Fig. 9

Contour line map of the base of the northern Kalocsa Depression with seismic and borehole sections (fragment of Figure 3 by Jaskó and Krolopp 1991, completed with seismic sections from Figure 3 by Tóth and Horváth 1997, and with borehole section in Figure 8 by Marosi and Schweitzer, 1997. The Danube bend in the area of seismic sections Danube-202 and Danube-203 is displayed differently in the two compared maps). *1–9.* captions in the original map (updated): 1. loess and associated Pleistocene sediments, 2. fluvial sequence, 3. borehole with legend, 4. contour lines of the base of the fluvial sequence (m a.s.l.), 5. Danube, 6. fault and downthrow, 7. section with legend, 8. axis of the Kalocsa Depression, 9. fossils; *10–12*: additional captions: 10. seismic section with legend (Danube), 11. part of section with great thickness of Quaternary and its value in ms, 12. section of Figure 8 by Marosi and Schweitzer (1997) with boreholes (identified from borehole coordinates by Gy. Maros since there are no location sketches for any of the sections in the original) with Quaternary thickness in meters below the mean Danube level. III, Kalocsa Depression, IV, Paks Fault (I and II are outside of the figure) section Danube-202 and 55 m a.s.l. on section Danube-203; this indicates 42 and 32 m sediment thickness respectively below the 87 m a.s.l. water level. Both data series come from drilling so that they are sufficiently firm. They are correlatable with the travel times in the sections, resulting[1] in (42 × 2)/(50× 1000) = 1680 m/s for the Danube-202 section and in (32 × 2)/(60 × 1000) = 1067 m/s for the Danube-203 one.

It should be noted that the picture in the Danube-202 section is in harmony with the running of the contour lines in the map (position of the base of Quaternary is constant) whereas the Danube-203 section shows a controversial picture (thickness rises toward the NW, not SE). This indicates that the topography of the base of the fluvial sequence is much more complicated than shown in the map (Fig. 9).

As seen, an indirect comparison with drilling data shows that the seismic velocities in Quaternary sediments are only about two-thirds of those usual in the upper part of the Pannonian sequence in the Paks area (≈ 2000 m/s). The velocities obtained are partly less and partly more than that of 1450 m/s for *pure water*[2]; the estimate is, however, too rough to take for certain the deviation from the pure-water velocity which is expected for water-saturated soft sediments. In any case, from the velocity decrease one can suppose that the Pannonian stratification is "pushed down" in time sections below the thick Quaternary[3]. This may result in the "generation" of flexures in the Pannonian sequence on those parts of the section with rapid Quaternary thickness changes.

The question whether this phenomenon is sufficient to generate flexures in the sections (i.e. the flexures disappear in depth sections after recalculation with true velocities) or not could only be answered if such depth sections were constructed. To date these flexures do not serve as a *firm* basis for assuming Quaternary tectonism.

Thicknesses in Figs 6 and 7 are about the same as to their south. In the base Quaternary map (Fig. 9), one can read off thicknesses of 70–80 m at the locations of these sections. When assuming the Danube surface to be about 90 m a.s.l. this would mean a 10–20 m thickness of the fluvial sequence. These values, however, do not seem to be supported by drilling data, and can therefore be left out of consideration. In the Danube-208 section the maximum thickness is even greater. It is about 100 ms (which would be about 60–70 m with velocities obtained in the south), i.e. as great or not much less than in the axial zone of the Kalocsa Depression (drilling data are not available either). The depression axis from Kalocsa, where the last borehole in the map is situated, can be drawn in a straight line towards the Danube-208 section so that the axis bend depicted by Jaskó and Krolopp (1991) disappears. In other words, the Danube-208 section falls into the northern continuation of the axis of the Kalocsa Depression; therefore, the "Paks Fault" cutting the depression in the north (Fig. 9) does not exist, since it was introduced by Jaskó and Krolopp (1991) only to explain that *cutting*. The problem of the northern continuation of the Kalocsa Depression, however, is beyond the scope of this work, which is why here only attention is called to it.

1 Taking into consideration that Tóth & Horváth displayed two-way travel time on the vertical axis.

2 Page 132, section 5.1.1.

3 The "push down" does not influence the deeper horizons of the Quaternary sediments since the velocity is constant along them, and the main condition of the "push down" is velocity change along a distinct time horizon.

From the above analysis it appears that "the tectonic origin of this deformation and genetic relationship to the faults observed in the Pannonian strata" is not at all "evident", either in the Danube-208 section or in any other one. Thus, the seismic sections, otherwise of good quality and of high-level processing, do not serve as *proofs* either for faults or for other deformations in the Quaternary. The conclusion of the authors[1], namely that "evolution of the observed fault zones was ongoing during late Pleistocene time as well, that is 40,000–50,000 years ago. This is so near in time to the present that future activity of the fault cannot be excluded. A most conservative conclusion is, therefore, that the fault is to be considered active" can by no means be accepted.

On the contrary: in the Paks area there is not so far any sort of convincing *proof* of fault-related Quaternary tectonism. Although it is always much more difficult to prove the absence rather than the existence of something, the fact that none of the very extensive studies has *proved* fault-related Quaternary tectonism (Balla et al. 1997) increasingly allows one to suppose that such a tectonism might have not existed at all.

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1 Page 136, last paragraph 5.

Appendix

Conclusions on the faults at the base or lower part of the Quaternary in Tóth and Horváth (1997)

Section code	*Original statements on the faults in the Quaternary*		*Comments*
	at the base	*in the lower part*	
Danube-202	"The bottom of the Quaternary is thought to be located at 50 ms."	"The bottom of the Quaternary ... is fairly ... continuous showing no offset by faults."	–
Danube-203	"The bottom of the Quaternary can be found at about 70 ms twt at the northwestern and rises to 25 ms at the southeastern end of the section."	"The unconformity is apparently not influenced by faulting."	–
Danube-205	"The bottom of the Quaternary can be identified at about 70 ms."	"The fault terminates upward at about 100 ms and folding can be seen further up." Its "few branches terminate even below 200 ms."	The fault is traced in the section up to the base of the Quaternary (50 ms), and it is unclear where "further up" with folding but no fault is. Quaternary is obviously not affected by faults or folding.
Danube-207	"The bottom of the Quaternary is well defined, almost horizontal at about 20 ms twt in the western part of the section. ... However, in the eastern part Pannonian strata are tilted and likely to crop out at the water bottom."	The fault at the boundary between the two parts of the section "does not disturb the topmost 30 ms of the section."	Consequently, the Quaternary is not affected by faults.
Danube-208	"The Quaternary/ Pannonian unconformity is distinct."	"Pannonian strata are both faulted and folded but this deformation ceases below 35 ms twt."	The base of the Quaternary is at 50 ms in the SW, 100 ms in the middle, and 25 ms in the NE. Thus, the 35 ms level falls within the Quaternary, except for the northeastern fifth of the section. The statement on the ceasing level of deformation obviously relates to the NE part of the section. Other faults indicated are cut by the base of the Quaternary.
Pa-2a	"The Quaternary/ Pannonian unconformity can be found at 70 ms at the northern end of the section, and around 80 ms at the southern end. The dip is seemingly constant."	"Faults offsetting the Pannonian strata do not reach this horizon", i.e. the Quaternary/ Pannonian unconformity.	–

Section code	*Original statements on the faults in the Quaternary*		*Comments*
	at the base	*in the lower part*	
Pa-2b	"The Quaternary/ Pannonian unconformity can be found at 100 ms."	"The Quaternary/ Pannonian unconformity ... is seemingly not faulted."	–
Pa-3b	"The Quaternary/ Pannonian unconformity is most probable between 60 and 70 ms ..."	"The base of Quaternary can be faulted."	The base of the Quaternary is only well visible south of the fault; north of the fault both the Quaternary and its base are completely uncertain.
Pa-12	"Due to poor imaging of the upper part of the section it is only a guess that the Quaternary/ Pannonian unconformity can be at about 40 ms."	"The reflection at 60 ms is probably faulted", but "the upper termination of fault cannot be determined precisely due to bad quality of the section above 40 ms."	There are no reflectors above the base of Quaternary guessed at 40 ms, but the fault does not even reach this horizon.
Pa-13	"At the southern end of the section the Quaternary/ Pannonian unconformity is identified as the strong reflector at 45 ms twt time."	At the northern end of the section there is a fault which "is well defined and most probably reaches the bottom of the Quaternary strata."	There are no more reflectors above 45 ms, and the base of the Quaternary in the northern part of the section has been identified along the upper boundary of the section (thus, a presence of the Quaternary is not justified). The position of the base of the Quaternary is uncertain along the entire section.
Pa-14	"Due to poor imaging of the upper part of the section it is just a guess that the Quaternary/ Pannonian unconformity can be at about 40 ms."	In the middle of the section "a fault ... is obvious which reaches the reliably imaged topmost layers at about 45 ms."	Thus, the base of the Quaternary is uncertain, but the fault does not even reach it.
Pa-15	"The base of the Quaternary ... can be assumed at 40 ms (northern end) and 50 ms (southern end) twt."	"The fault zone" in the middle of the section "disturbs the shallowest imaged layers as well. In this strongly-faulted zone the base of the Quaternary cannot be identified."	In the south, above 40 ms, there are practically no reflectors. In the north, above 50 ms, reflections are poor and uncertain. Thus it is unclear whether Quaternary is present in the section or not.
Pa-17	"The Quaternary/ Pannonian unconformity can be located at about 50 ms."	"All the reflectors can be correlated ... without assuming any faulting."	–

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Evidence for Quaternary tectonic activity in the Paks area, Hungary

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The existence of a fault zone near the town of Paks has been suggested by several studies (Szilárd 1955; Rónai 1973; Némedi Varga 1977; Kőrössy 1982; Pogácsás et al. 1989; Kókai and Pogácsás 1991) since the first indication on the structural map of Lóczy Jr. (1939). These studies, based on different geologic data and concepts, arrived at the same conclusion; there is an approximately SW-NE trending regional fault zone which crosses the river Danube in the vicinity of Paks, where the Hungarian Nuclear Power Plant (NPP) is located. Rónai (1973) described this zone as a primary lineament of Quaternary tectonism, while Pogácsás et al. (1989) characterized it as a Pliocene–Quaternary strike-slip fault.

More than ten years of intensive research supported by the Paks NPP verified and mapped in detail this fault zone of Middle Miocene age. However, no consensus has been reached in the question of fault reactivation during recent times. Balla et al. (1997) claim that there is no evidence of Quaternary fault activity, but our studies contradict this opinion.

Seismic profiling carried out on the river Danube imaged faulted Pliocene strata (Torony Formation), but no tectonic deformation could be observed in the unconformably overlying latest Quaternary (0 to 50 ky) fluvial beds. This remarkable time gap between the Torony Formation and the alluvium of the river Danube in the Paks study area is the reason why the exact time period of fault reactivation may be the subject of debate. A more favorable situation can be found, however, around and below the Tisza River in the middle and southern part of the Great Hungarian Plain. Here the Pliocene–Pleistocene strata are apparently complete and their contact is conformable. A long seismic profile acquired in 1996 on the river Tisza documents late Pleistocene (310 ky) fault activity at the Martfű bend of the river. The structural connection between the two fault segments imaged below the rivers Tisza and Danube, at Martfű and Paks respectively, has been documented by Pogácsás et al. (1989) using petroleum industry seismic sections and borehole data. Accordingly, it is concluded that Quaternary reactivation of the fault zone passing by Paks cannot be questioned.

Key words: Quaternary tectonics, high resolution seismic, structural analysis

Introduction

We were pleased to learn that our paper entitled "Neotectonic investigations by high resolution seismic profiling " (Tóth and Horváth 1997) published in the "Seismic safety of the Paks Nuclear Power Plant" (Marosi and Meskó 1997) has raised interest. It was even more challenging that one of the leading geologists of the seismic safety reevaluation project, Zoltán Balla, was so much interested in our results that he found the time to analyze every detail of our study and summarize his observations in a research article (Balla 1999). The

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editorial board of the Acta Geologica Hungarica is thanked for offering us the possibility to comment on the most important points of Balla's evaluation.

Earlier views on Quaternary tectonic activity

Let us start with the first sentence of Balla's (1999) summary. He alleges that "Data on the Quaternary tectonism of the Paks NPP area have not been included in the seismic hazard assessments, nevertheless, they are important factors of the realm of scientific and lay discussions on the subject.". If this statement were true many scientists would have worked in vain. Fortunately this was not the case. The final report on the Nuclear Safety of Paks NPP has been prepared by Ove Arup and Partners International (London) and, as a matter of fact, they included the possibility of active faulting in their probabilistic seismic risk assessment (Ove Arup 1997). They assigned active faulting in the vicinity of the Nuclear Power Plant a 10% probability. One may discuss whether this value is underestimated or overestimated, but it cannot be questioned that it has indeed been incorporated into the probabilistic assessment. This shows that the possibility of Quaternary tectonism, which is active even today, was not considered negligible.

We have to correct another factual mistake of Balla (1999). He states that "This is reflected in the recommendation by the IAEA not to locate nuclear power plants above known faults. However, when the Paks NPP was sited this requirement did not yet exist, *and the faults below it were unknown*" (the italics are ours). The second part of the second sentence (in italics) contains an erroneous statement. A fault in the vicinity of Paks town was already known in the 1930s, many years before siting and construction work on the Nuclear Power Plant began. The tectonic map prepared in 1938 by two acknowledged geologists of that time, Lajos Lóczy Jr. and Ferenc Szentes, clearly indicates a SW–NE and a NW–SE trending fault, crossing each other in the vicinity of Paks. Existence of the SW–NE trending fault zone has not only been proven, but the fault zone has been mapped in detail by now (Tóth and Horváth 1997). The NW–SE trending fault zone has not yet been documented convincingly, although it cannot be ruled out (Tóth and Horváth 1997).

The tectonic map of Lóczy Jr. and Szentes was republished a year later by Lóczy Jr. (1939) in his paper entitled "Geomorphology of the Hungarian basin system, with special emphasis on petroleum exploration". This is shown in Fig. 1.

Furthermore, it should be recalled that the first seismotectonic map of the Hungarian basin (Simon 1939) accepted this tectonic map as a basemap and clearly indicated that the earthquake at Kecskemét in 1911 was related to the fault zone of Paks. This means that not only the existence of the Paks fault zone but also its recent seismic activity was actually documented in the geologic literature of the late thirties. It is worthwhile to cite the prophetic words of Simon (1939) commenting on his seismotectonic map:

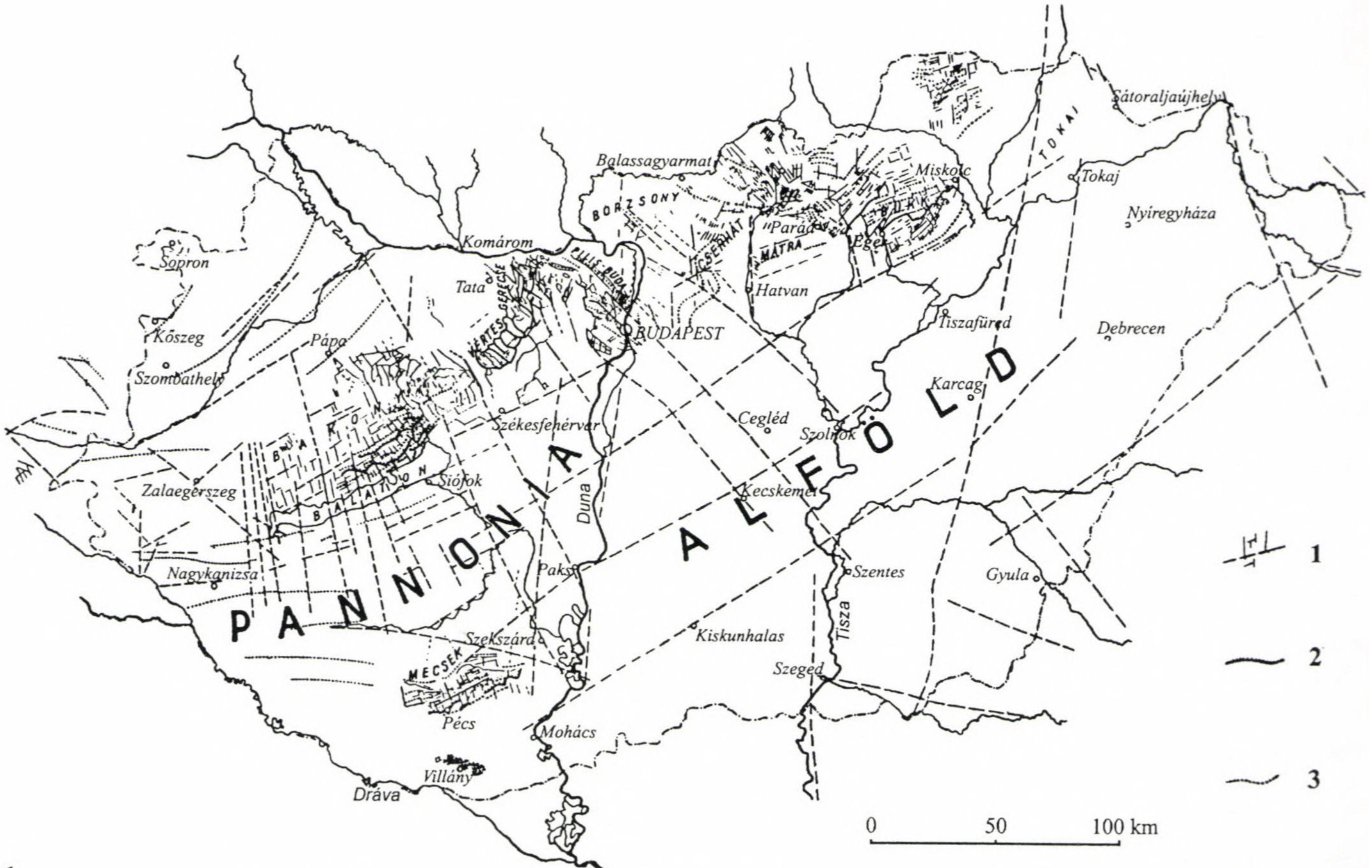

Fig. 1
Structural map of Hungary (Lóczy Jr., 1939). Note the two fault zones intersecting in the vicinity of Paks. 1. Faults and fractures; 2. Overthrusts; 3. Folds

"When in the future one will have to locate a public utility of vital importance or a new settlement, the published earthquake map will give advice on the expected seismic risk and on the relationship between the structural setting of the Hungarian subsurface and the seismic activity. In this respect the amplification induced by unconsolidated soils and the effect of the faults cutting through the basin are striking."

This message was very well received by the next generation of geologists and this is probably best illustrated by the works of András Rónai. He had been interested for decades in the Quaternary movements of Hungary and their relationship to earthquakes. Studying the uplift of Pannonian strata and river terraces in mountainous areas and the depth of the uppermost Pliocene strata and sedimentary cycles in the basin areas, he derived a map of Quaternary vertical movements (Rónai 1973). On his map the SW–NE trending zone below the river bend at Paks is described as a primary lineament of Quaternary tectonism. Other geomorphologic studies argued for the existence of a fault zone in the Paks region as well. Szilárd (1955) noted the tectonic influence on the river course of the Danube and mentioned the joint effect of a NE–SW and a NW–SE trending fault zone. The same opinion was shared by Pécsi (1959) in his Danube monograph. The existence of the fault zone has been concluded by studies based on different geologic and geophysical data as well (Némedi Varga 1977; Kőrössy 1982).

One can see that the presence (or at least the possibility) of a significant fault in the vicinity of Paks was known by geoscientists well before the planning of the Nuclear Power Plant began, and even the possibility of its seismic activity had been suggested as early as 1939.

Effect of the weathering layer on the seismic image

Let us now discuss some of Balla's (1999) specific comments on our paper (Tóth and Horváth 1997). One can see from his analysis of the shallow seismic reflection profiles that the key point of his criticism is the distorting effect of the near surface weathering layer. This is a layer at or near the surface in which seismic velocities are substantially lower than in the substrate and often show high lateral variation. This lateral variation can cause so-called "push down" (a lens with smaller velocities than the surrounding strata) or "pull up" (a lens with higher velocities than the surrounding strata) effects in the deeper part of the seismic time section. However, it is important to note that the most significant velocity inhomogeneities are caused by unconsolidated and not fully saturated lenses and *are located mostly above the water table.* That is why the geophysical weathering layer (which is defined by the seismic velocities) is often taken to be the zone above the water-table and may not correlate exactly with a layer of geologically weathered materials. In our case, when profiling on the river Danube, obviously both the source and the receivers were towed below the water table. This is a key factor contributing to the higher resolution and more coherent image of the river seismic sections.

The effect of the weathering layer is a well-known fact among processing and interpreting geophysicists and geologists and it can be accounted for. Removal of the distorting effect of the shallow layer velocity inhomogeneities is tedious work based on detailed analysis of the direct and refracted waves. This analysis has been carried out for all of the high-resolution land seismic sections discussed by Tóth and Horváth (1997). In most cases, however, simple calculations can help to decide whether a structure present on a seismic section is a true geologic structure or an artifact caused by a shallow layer in- homogeneity.

This important decision has to be carefully made in every case, as the final tectonic conclusion can be very different if one demonstrates that the seismic image represents a true geologic structure and not a "push down" caused by a low velocity inhomogeneity in the weathering layer. Let us see an example based on a particular high resolution land seismic profile and note again that the discussed distorting effect is usually much larger on land than on river seismic profiles.

Balla et al. (1997) allege that the "push down" observed on land shallow seismic sections *Pa-12, Pa-15* and *Pa-13* can be explained by the distorting effect of an abandoned and filled up channel of the river Danube. They claim that, of the two lenses observed on geoelectrical profiles (Stickel and Zalai 1994) only the shallower (between 84 and 90 m a.s.l.) had been corrected for, but not the deeper (between 63 and 75 m a.s.l.). Balla (1999) suggests that this latter, uncorrected lens is the cause of the observed "push down".

Considering that the depth of the "push down" observed on the *Pa-15* seismic section is about 70 ms and the velocity of the strata below the weathering layer is about 1600 m/s one can calculate the velocity within the deeper lens necessary to cause the "push down". One arrives at an acoustic velocity of 290–300 m/s, which is less than the acoustic velocity in air. As the deeper lens (between 63 and 75 m a.s.l.) is below the water table such a low acoustic velocity can be obviously excluded. This simple calculation shows that the observed "push down" cannot be explained as the effect of a shallow layer inhomogeneity. We suggest, therefore, that it is a real feature related to a negative flower structure connected to a strike-slip fault zone.

This is confirmed by the fact that several independent seismic profiles imaged the same fault structure, although different sources and different acquisition parameters were used. In addition they were processed by independent groups of geophysicists: *Du-1*: seismic section measured and processed by the oil industry using explosive sources; *Pak-3*: seismic section measured and processed by the Eötvös Geophysical Institute (ELGI) using explosive sources; *Pa-13*: shallow seismic reflection profile measured by ELGI using vibroseis sources; *Pa-15*: shallow seismic reflection profile measured by ELGI using explosive sources; *Duna-203* and *-208*: high resolution seismic section measured on river Danube by the Department of Geophysics, Eötvös University (ELTE) using watergun source. *Therefore we conclude that the fault pattern observed on the seismic profiles are images of real structures and not artifacts created by shallow layer inhomogeneities.*

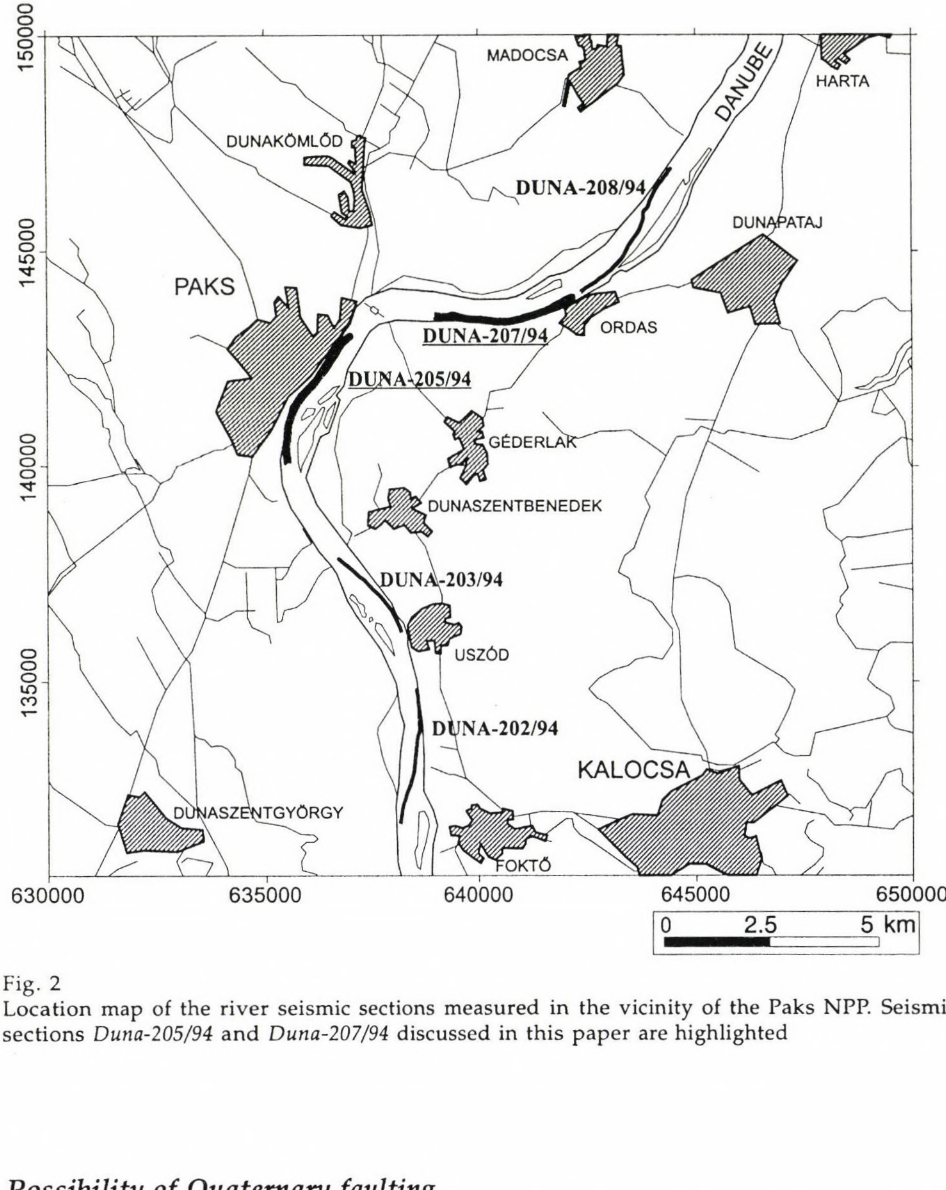

Fig. 2
Location map of the river seismic sections measured in the vicinity of the Paks NPP. Seismic sections *Duna-205/94* and *Duna-207/94* discussed in this paper are highlighted

Possibility of Quaternary faulting

In spite of the disagreement on some details most Hungarian earth scientists and involved foreign specialists agree on the existence of a fault zone in the vicinity of Paks. This fault zone can be derived from seismic, gravity, magnetic and well data and is scientifically well established. The true question of the debate – as indicated by the title of Balla's paper – is whether fault activity during Quaternary times can be proven.

The traditional "rule of thumb" of structural geology says that fault activity *postdates* the deposition of the youngest faulted strata and *predates* the deposition of the overlying unfaulted strata. This "rule of thumb" is, however, not perfect. Both the logic and the methodology of the rule can fail.

The logic fails when a fault-related earthquake does not create an offset at the surface, thus leaves no footprint in the youngest (topmost) strata. This is the case for many earthquakes which, according to their location and focal mechanism, can be connected to active faults. These are called blind or hidden faults.

The methodology is imperfect too, as in applying the rule one usually cannot precisely determine the age of the termination of fault activity. One can only specify a time interval which is as long as the time gap associated with the unconformity between the faulted and non-faulted strata (time discordance). If for example the youngest strata below the unconformity is 10 Ma and the oldest strata above it is 1 Ma old one can only state that faulting ceased *sometime* between 10 Ma and 1 Ma. In many important cases this uncertainty in timing leaves wide ground for subjective interpretations and preconceptions. To avoid this new data and further geologic constraints must be considered.

The above theoretical consideration is essential to bring the question of Quaternary fault activity in the vicinity of Paks to a decision. Let us analyze the river seismic profile *Duna-207/94* (Fig. 3a, b). It was acquired about 7–9 km to the northeast of the Nuclear Power Plant (see location map in Fig. 2). Strongly tectonized Late Miocene through Pliocene strata is evident on the section. It is unconformably overlain by the apparently not tectonized, fluvial deposits of river Danube. The age of the strata below and above the unconformity is well constrained.

Shallow borehole *Paks-881* reached the base of the fluvial deposits of the river Danube at 27 m depth. At the depth of 20.5 m a piece of driftwood was encountered. Radiocarbon dating resulted an age of 40 000 year BP for this piece of drift (Hertelendi et al. 1989). Extrapolating this value one can estimate the age of the oldest fluvial deposits in the Paks area to be about 50 ky. Age of the topmost faulted strata can be estimated with confidence from the data of *Paks-2, -2a, -3, 4a* and *4b* boreholes drilled in the vicinity of the Nuclear Power Plant. All five boreholes crossed 70–110 m thick deposits of the Torony Formation (dune sand with intercalation of 1–2 m thick clay and silt layers) below the unconformity. Hungarian stratigraphers agree on the heteropic relationship between the Torony and Nagyalföld Formations, both of them being Pliocene in age according to biostratigraphic and magnetostratigraphic data (Jámbor et al. 1988; Elston et al. 1990). The latest results of Juhász et al. (1996) also support the Pliocene age of the strata found below the unconformity.

Along the high resolution seismic profile *Duna-207/94* the approximately 50 ky old alluvium of the river Danube unconformably rests on the Pliocene Torony Formation. It is clear in the seismic section that faults propagate up to the unconformity and offset the Torony Formation. Applying the discussed rule of

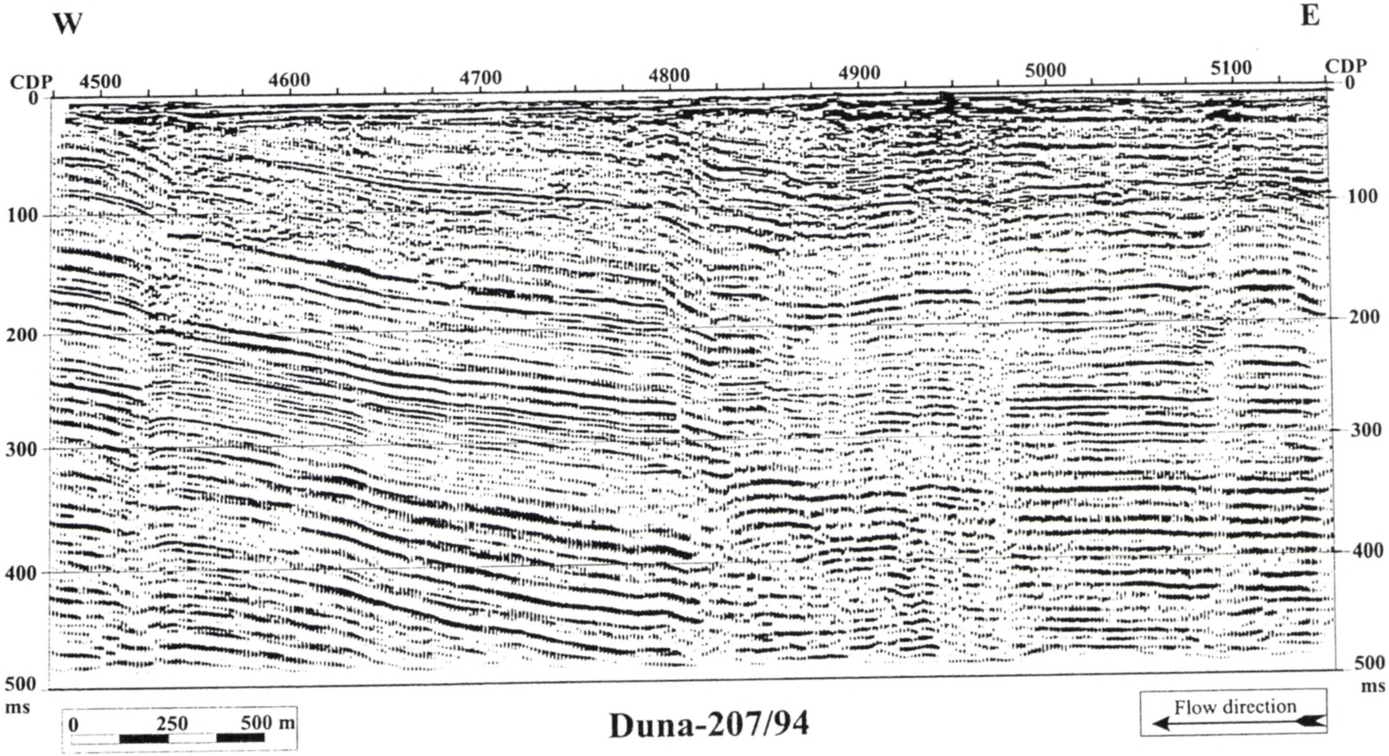

Fig. 3a
Duna-207/94 high-resolution multi-channel seismic section measured at the Paks bend of the Danube. Location of the section can be seen in Fig. 2

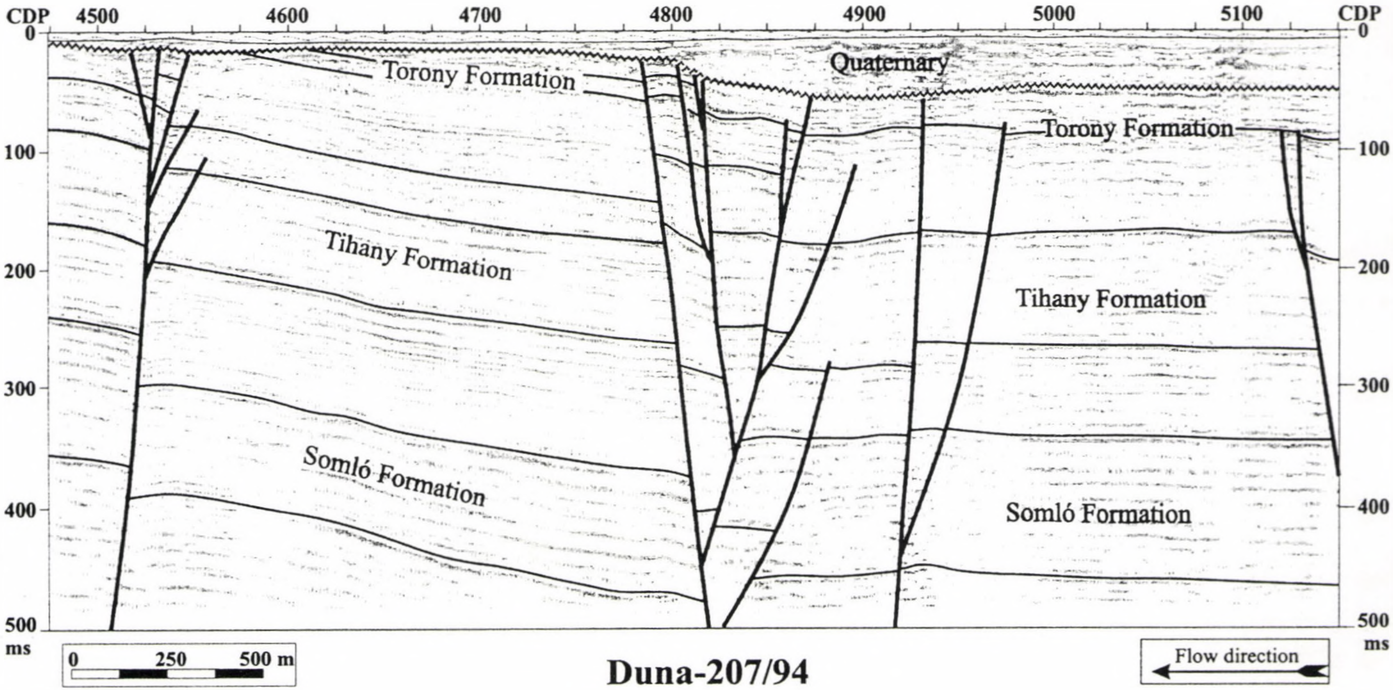

Fig. 3b
Interpretation of the *Duna-207/94* seismic section. The Pliocene Torony Formation and the underlying Late Miocene formations are clearly faulted; however, tectonic deformation cannot be observed in the unconformably overlying late Quaternary strata

structural geology we can conclude that the fault structure at Paks postdates the Torony Formation.

This is even more obvious on the high-resolution seismic profile *Duna–205/94*, and on an ultra-high resolution seismic profile measured in 1996 at the same location (*Duna_Seistec–4/96* in Fig. 4a and 4b). This ultra-high resolution seismic profile was measured with a single-channel system using a boomer source. Although only the topmost about 5 meters of the sediments below the riverbed are imaged, a very detailed analysis of the faulted structure can be performed. The resolution of the *Duna_Seistec–4/96* profile is 0.1 m! This ultra-high resolution seismic profile measured along the *Duna-205/94* multi-channel profile clearly images approximately 1.5 m thick alluvium of the river Danube and the underlying Torony Formation. The boundary between the two is a marked unconformity clearly visible on the seismic section. It is clear from the new data that the fault is continuous right up to the unconformity between the Torony Formation and the Danube alluvium (see Fig. 5 zoomed section). Two important refinements can be made to the interpretation of *Danube-205/94* (Tóth and Horváth 1997) based on this new ultra-high resolution seismic section:

i) Base Quaternary is actually in a higher position than previously estimated. We can clearly see that less than 2 meters of river deposits are present below the Danube in front of the town of Paks. This is significantly less than the approximately 40 m (about 55 ms two-way travel time on the seismic section) thick alluvium indicated by Tóth and Horváth (1997) when interpreting the *Duna–205/94* multi-channel seismic profile.

ii) Fault penetration is proven apparently to the bottom of the river.

The first refinement is a confirmation of Balla's (1999) remark saying that "This indicates that the topography of the base of the fluvial sequence is much more complicated than shown in the map (Figure 10)." The second point however strongly contradicts his observation saying that "In the Danube–205 section (Figure 6), in turn, the base of the Quaternary sediments above the fault seems to be as horizontal and quiet as those parts farther away from the fault. One can believe that there were deformations in the original sections which disappeared in the course of reducing them in size, but the fact is that in the published sections even slight deformation is not recognizable." On the contrary, *Duna_Seistec–4/96* ultra-high resolution seismic profile leaves no doubt about signs of tectonic deformation even as shallow as 1 meter below the river-bottom.

Fault activity postdating the Torony Formation is so obvious that specialists analyzing the data of the *Paks-2* borehole arrived at the following conclusions well before any seismic profile had been measured on the river Danube (Bruckner-Wein et al. 1982):

"... during the deposition of the Torony Formation a practically fresh water fluvio-lacustrine environment was present. Significant *structural evolution during early Pleistocene times* resulted in strong erosion; however, this structural evolution already affected the Pannonian strata..." (the italics are ours). Simply speaking we can say that the high-resolution seismic sections measured on land and on the river Danube only confirmed this earlier observation. This chapter can be concluded with the statement that *based on the borehole and seismic data collected in the vicinity of Paks the possibility of Quaternary (Early and Middle Pleistocene) tectonic activity in the area cannot be questioned.*

Evidence of Quaternary faulting

The aim of the previous chapter was to argue for the *possibility* of Quaternary tectonism in the Paks area. As it has been shown with the help of recent geologic and geophysical data the possibility of Quaternary faulting cannot be excluded from conservative seismic risk evaluation. As mentioned earlier this has been actually done by Ove Arup (1997), although in the light of most recent data the 10% probability of this event appears to be underestimated.

Since the seismic survey on the river Danube in 1994 new data have been collected on the river Tisza as well and we present in this paper, for the first time, the most relevant of them. We hope that this will suit Balla's requirement that "... proofs, not possibilities will be discussed.".

Direct *evidence* for Quaternary fault activity can be found in an area where the time gap between Pliocene and Quaternary strata is smaller, or ideally non-existing, because sedimentation has been continuous. This favorable situation can be found below the Great Hungarian Plain. It is still a subsiding part of the Pannonian basin (Rónai 1986) and in the central part no observable time gap is present between the Pliocene and the Quaternary strata. This could be demonstrated by the "classic" magnetostratigraphic sections of Dévaványa and Vésztő, which documented the presence of every magnetic polarity reversal in the almost 1200 m thick late Pliocene and Pleistocene section (Rónai 1981). Taking into consideration more recent magnetostratigraphic logs measured in other boreholes Pogácsás et al. (1990) verified the chronostratigraphic correlations with the help of seismic sections. The age of the formations at the bottom of the Dévaványa and Vésztő boreholes have been reinterpreted to 4.25 Ma from the previously suggested 5 Ma. This revision further supports the idea that no time gap is present at the Pliocene/Pleistocene boundary in this part of the Pannonian basin.

An almost 200 km long continuous, high resolution, multi-channel seismic profile was acquired between Szeged and Kisköre on the river Tisza in 1996. A portion of this section can be seen in Fig. 6a. Acquisition and processing parameters of this seismic section are very similar to the ones used on river Danube and discussed by Tóth and Horváth (1997). A significant difference, however, is that the seismic section in Fig. 6 is a depth-migrated section, excluding all possibilities of the problems discussed by Balla (1999) in connection with unmigrated time sections. This 1100 m long section was recorded just north of the river bend at Martfű. The location map is included

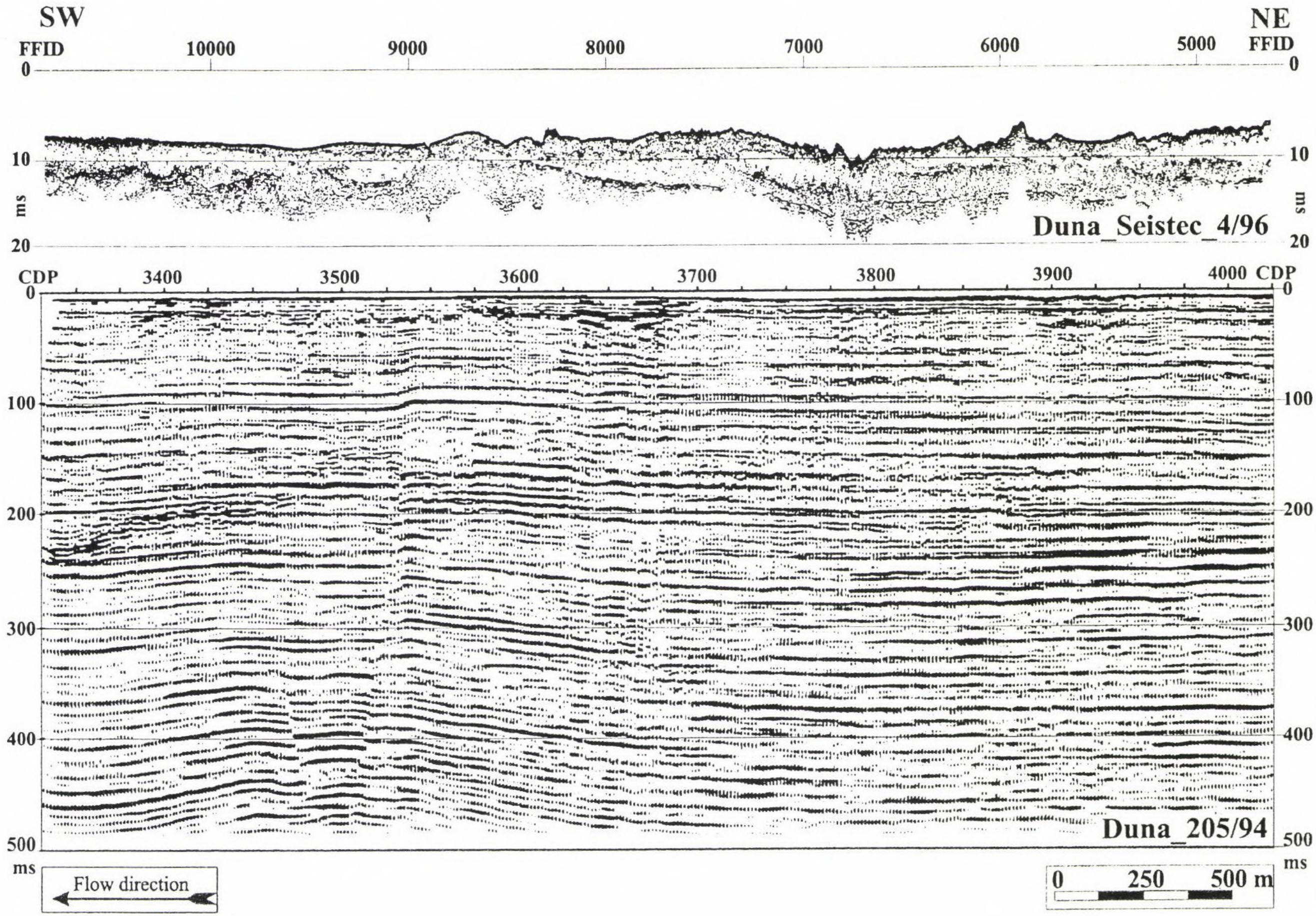

Fig. 4a
Duna–205/94 high–resolution multi-channel section and the corresponding *Duna_Seistec_4/96* ultra–high resolution single-channel seismic section. Location of the sections can be seen in Fig. 2

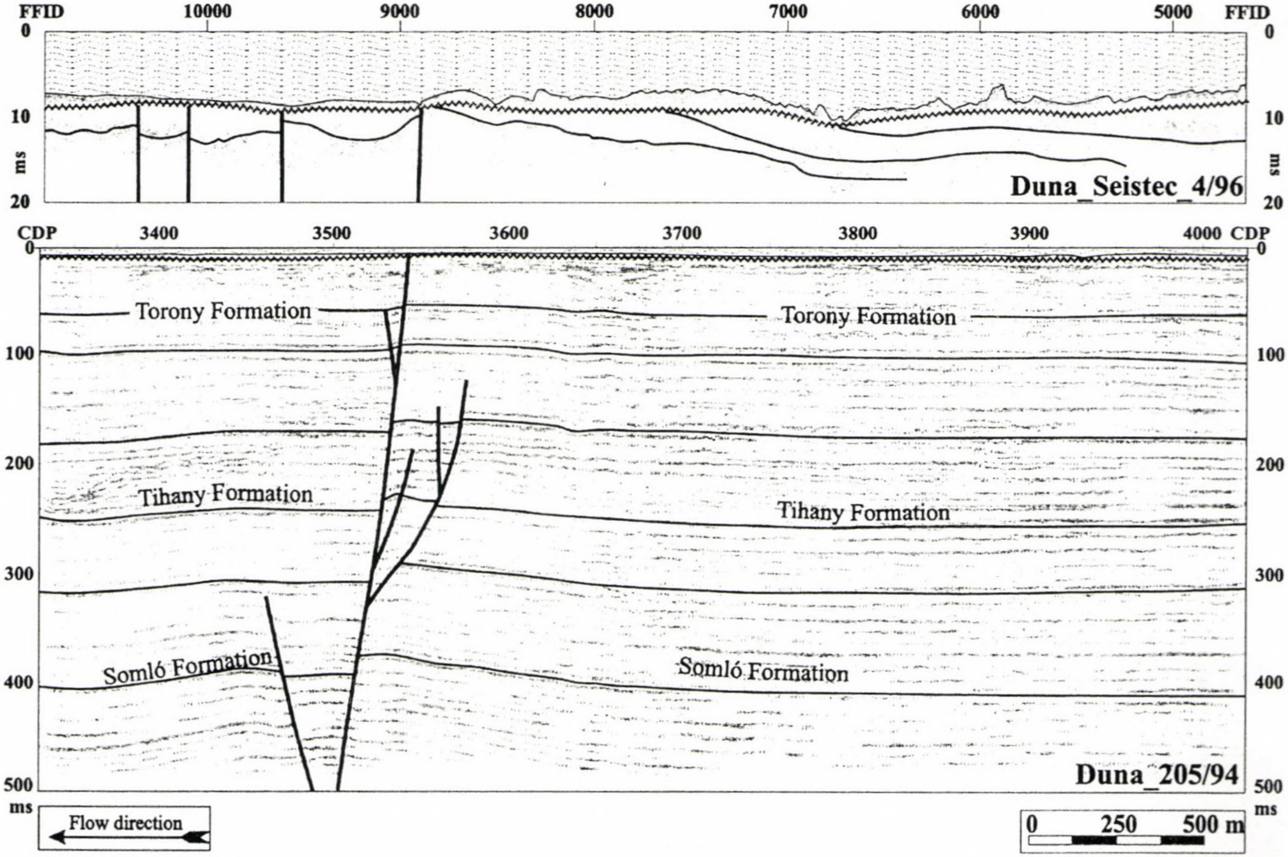

Fig. 4b
Interpretation Duna–205/94 high-resolution multi-channel section and the corresponding *Duna_Seistec_4/96* ultra-high resolution single-channel seismic section. Note the fault propagating virtually up to the river bottom. For details of the ultra-high resolution seismic section see Fig. 5

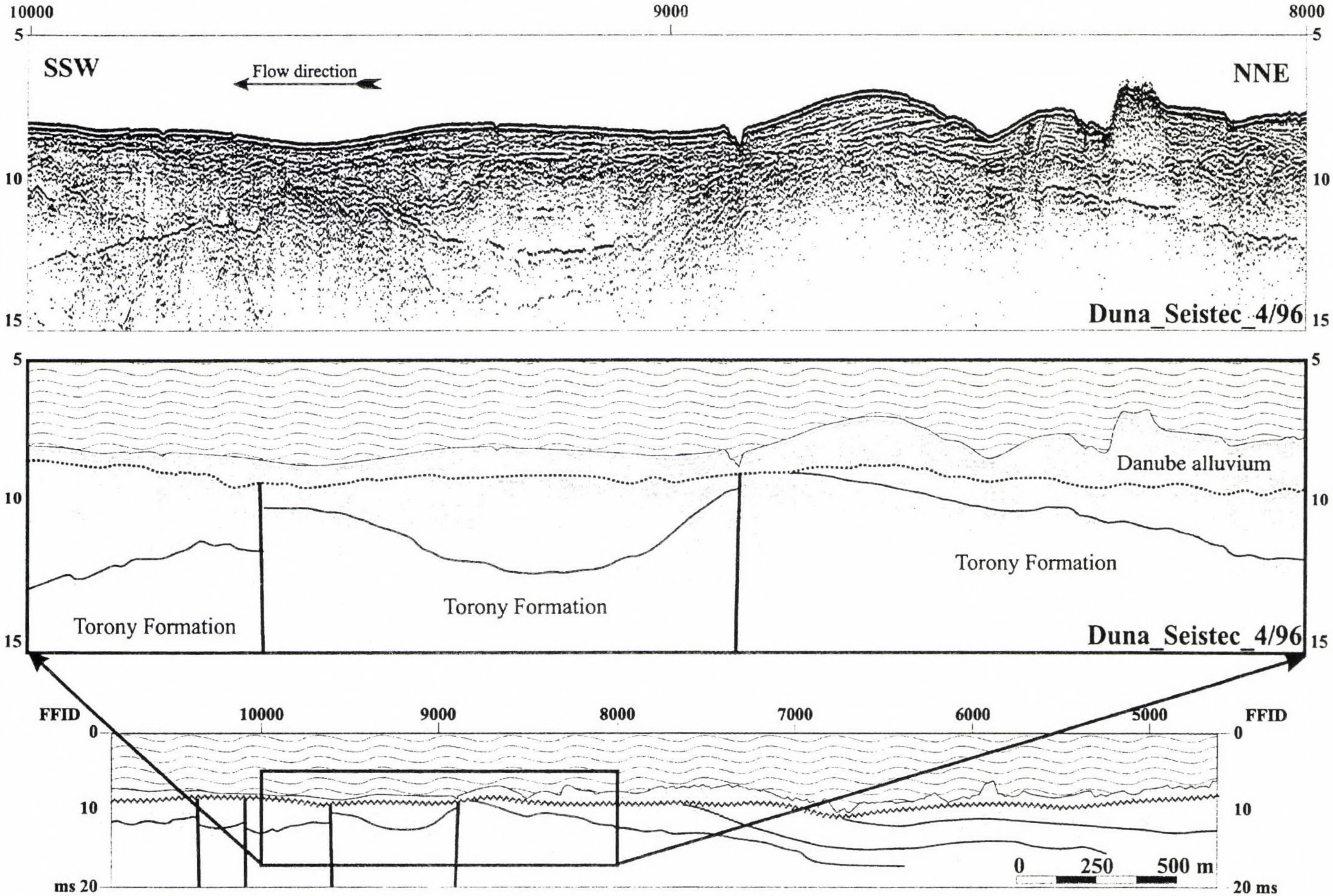

Fig. 5
Enlarged detail of the ultra-high resolution single-channel seismic section shown in Fig. 4. Note the clear image of the Danube alluvium and the penetration of the fault branches to the very top of the Pliocene strata (Torony Formation)

in Fig. 6b, and an arrow marks the exact location of the section. The seismic section images the strata below the river right from the river bottom down to a depth of 500 m. A negative flower structure is clearly observable on the section as part of a transtensional strike-slip zone. The resolution of the seismic section is less than 2 meters in the topmost part; therefore, vertical offsets exceeding 2 meters are perfectly imaged. The fault branch at CDP 2185 can be traced up to 45 m depth. This indicates that even at this shallow depth the fault has a vertical offset of about 2 m.

What can be the age of the strata at 45 meters depth? We can easily make a good estimation. D. Lőrincz (1997) published the Quaternary thickness map of the area. With the help of this map we can mark the base of Quaternary strata at around 350 m depth along the *Tisza-40/96* seismic section shown in Fig. 6b. Assuming 2.4 Ma for the age of the continental Pliocene/Quaternary boundary and continuous sedimentation during Quaternary times, 45 meters depth corresponds to less than 310 ky. Therefore *310 ky ago (late Pleistocene) the fault zone was still active.* It is important to note that vertical offsets along the fault branches increase with depth. This indicates that strike-slip movement took place not in one step, but repeatedly during Quaternary (synsedimentary fault).

In the map insert in Fig. 6b the fault zone mapped by D. Lőrincz (1997) has been indicated as well. D. Lőrincz and Szabó (1993) and D. Lőrincz (1997) also suggest that this fault zone was active during the Quaternary. It has been confirmed that the fault structure imaged by the seismic section *Tisza–40/96* is a branch cutting through Quaternary deposits of the fault zone mapped by D. Lőrincz. The same fault zone is shown in Fig. 7 together with the Paks–Kiskőrös–Kisújszállás fault zone mapped by Kókai and Pogácsás (1991) on the Great Hungarian Plain and the fault zone mapped by Tóth and Horváth (1997) at Paks. It is clearly seen that three distinct segments of the same *Paks fault zone* have been mapped by three independent groups of researchers using different datasets. *Accordingly, this fault zone is a regional feature, most probably a strike-slip, which shows evidence of Quaternary activity.*

Closing remark

In the light of the above discussion one can raise the question: What about the seismic safety of the Paks Nuclear Power Plant? The answer is fortunately reassuring! In their final report Ove Arup (1997) calculated the response spectra corresponding to peak ground acceleration. In these calculations all data available at that time and many different scenarios have been incorporated and the results served as strict guidelines for the reinforcement of the Paks Nuclear Power Plant.

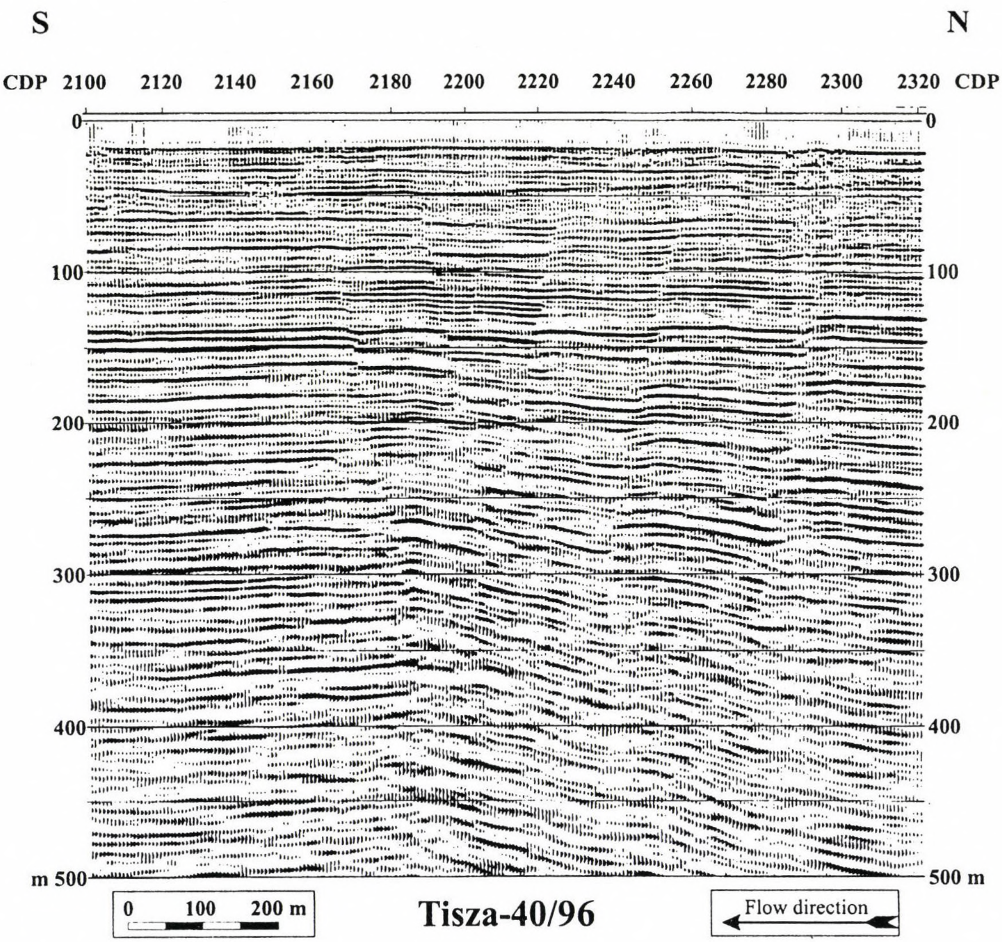

Fig. 6a
Tisza-40/96 high-resolution multi-channel seismic section imaging a fault zone below the Tisza at the river bend near Martfű

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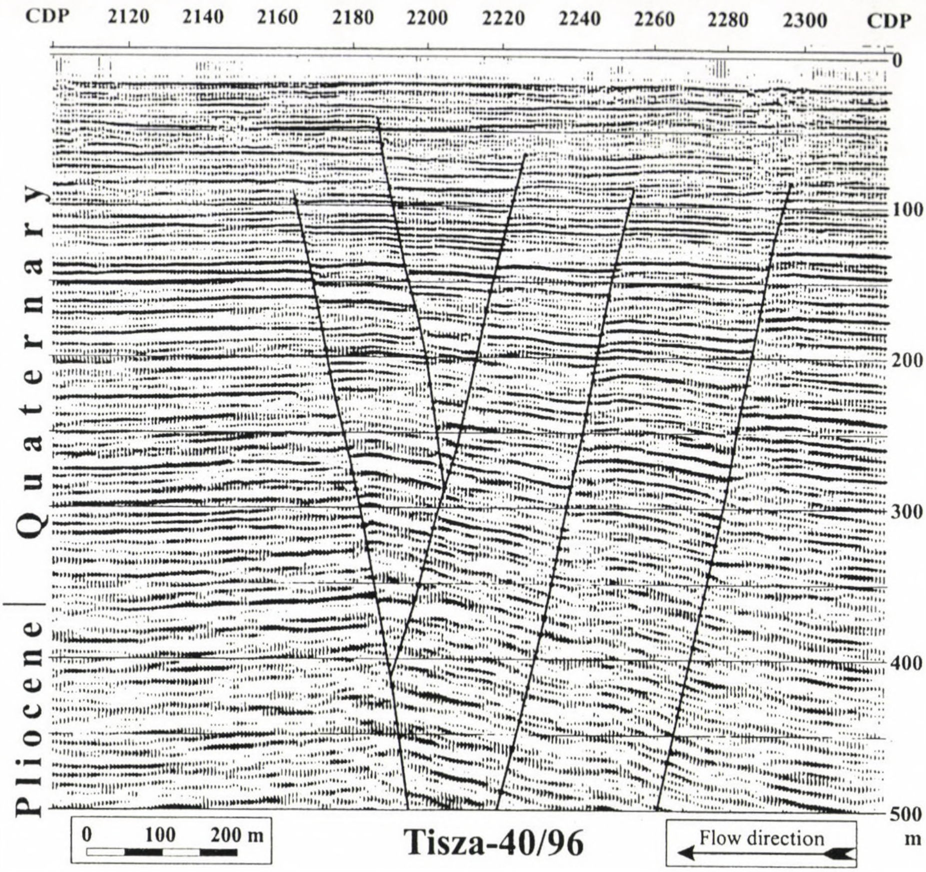

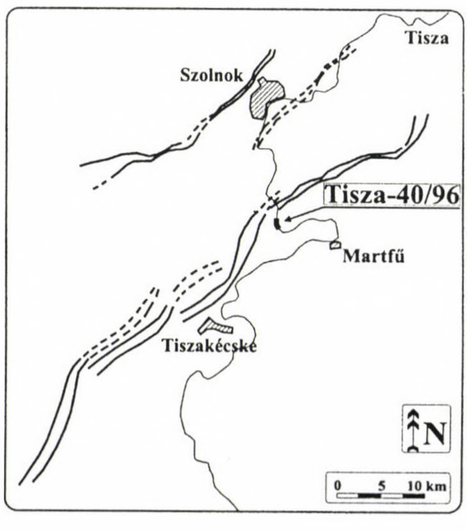

Fig. 6b
Interpretation of the *Tisza-40/96* seismic section. The sketch map below indicates the location of the section, and shows Quaternary fault zones mapped by D. Lőrincz (1997) using land seismic sections. The base of Quaternary strata has been identified with the help of the Quaternary thickness map published by D. Lőrincz (1997). Faulting of the Quaternary strata up to a depth of 45 m (below the surface) is obvious, which clearly suggests that the fault was still active during the late Pleistocene

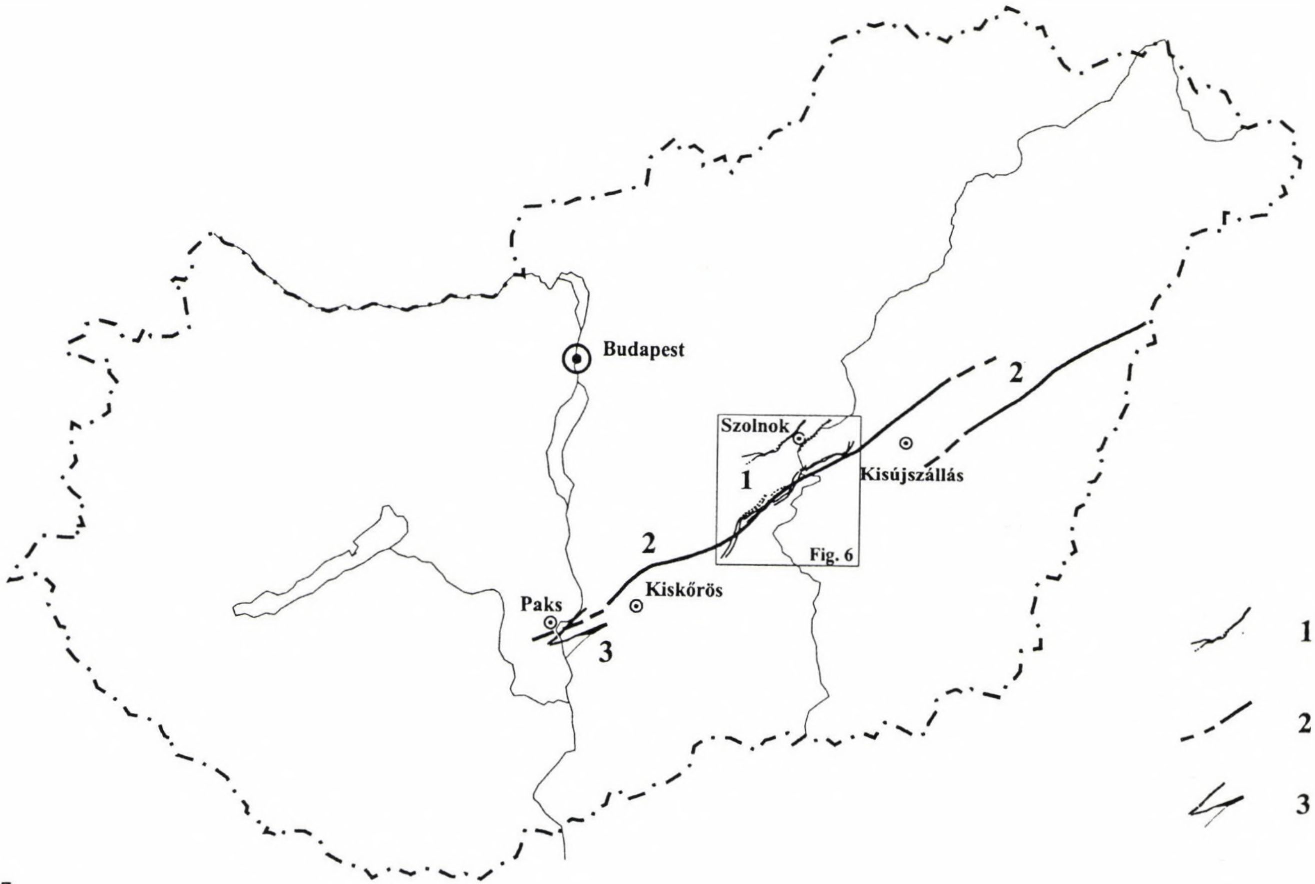

Fig. 7
Structural relationship of the Paks fault zone with the Paks–Kiskőrös–Kisújszállás fault zone mapped by earlier studies. This fault zone was crossed and imaged by the 1996 Tisza river seismic survey and late Pleistocene activity of the fault below the Tisza was documented (see. Fig. 6b). 1. Fault zone mapped by D. Lőrincz (1997); 2. Fault zone mapped by Kókai and Pogácsás (1991); 3. Fault zone in the vicinity of Paks mapped by Tóth and Horváth (1997)

discussion and his constructive comments are highly appreciated. The research has been partly founded by the OTKA T-019393 research project of the Hungarian National Science Foundation. Last but not least all those participating at the Danube and Tisza seismic surveys are thanked for their indispensable work.

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Conference Report

"Mineral Deposits, Economy, and Culture" - GEO '99

The Fourth Meeting of Hungarian Geoscientists, 1999, and HUNGEO-2000

In 1995, the idea of organizing a meeting of Hungarian geoscientists living in different countries of the world was conceived, in connection with the celebration of the 1,100th anniversary of the Hungarian Conquest, the founding of Hungary in the Carpathian Basin.

The meeting, under the name of HUNGEO'96 and held in Budapest, was attended by geologists, geophysicists, geographers end cartographers from all over the world. It was a remarkable success. As a result its Organizing Committee was transformed into the Standing Committee of HUNGEO-TOP (Hungarian Geoscientists' Scientific and Educational Program), hosted by the Hungarian Geological Society. It was decided to organize similar meetings every year until 2000.

The 1997 Meeting took place at Csíkszereda (Miercurea Ciuc in Transylvania, Romania) and the 1998 one again in Budapest, on the occasion of the 150th anniversary of the Hungarian Geological Society.

The third meeting was held in Eastern Slovakia (Upper Hungary until 1920) and Transcarpathian Ukraine (Hungarian Subcarpathia until 1920), between 18 and 23 August 1999. The 56 participants from six European countries (Austria, Germany, Romania, Slovakia, Sweden and the United Kingdom) visited several metallic and non-metallic mineral deposits in the Slovakian (former Szepes-Gömör) Ore Mountains (Ag, Au, Pb-Zn, Cu, Fe, asbestos, magnesite, opal, talcum) and the cultural highlights of this more than 700 year-old mining district, such as the towns of Kassa (Košice), Rozsnyó (Rožňava), and Eperjes (Prešov), the castles of Betlér (Betliar) and Krasna Horka, Jászó (Jasov) Abbey, the gothic church in Csetnek (Štitník) and the Ránkfüred (Herl'any) thermal spa (geyser). In Transcarpathian Ukraine the historic castles of Ungvár (Užhorod) and Munkács (Mukačeve) were viewed. The climax of the trip, however, was the visit to the Hungarian Teachers' College (founded in 1994) at Beregszász (Beregove).

At the two evening technical sessions held at Aranyida (Zlatá Idka) and Hrádok (Hradok), respectively, a total of 4 invited keynote talks and 11 other lectures were delivered, mostly on geologic topics but also on problems of earth science education.

At the business meeting of the Standing Committee held at Kassa (Košice) the Preliminary Program Proposal of the next year HUNGEO-2000 meeting was distributed, discussed and modified. Accordingly, it will take place near Budapest, on the occasion of the 1000-year jubilee of the Hungarian state, between 15 and 19 August 2000, under the title: **"The Earth Sciences on the Evolution of the Carpathian Basin".** On this occasion the science of Meteorology will join Geology, Geophysics, Geography and Geodesy-Cartography, including remote sensing and GIS.

The First Circular is scheduled for December 1999.

For further information on **HUNGEO-2000** please contact the

Hungarian Geological Society (MFT)
H-1027 Budapest, Fő utca 68., I/102.
Phone: (36 1) 201 9129
Fax: (36 1) 201 9129
E-mail: mail.mft@metesz.hu

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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, 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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- a footnote with the postal address of the author

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.

Figures and tables

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. 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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Miocene maar/diatreme volcanism at the Tihany Peninsula (Pannonian Basin): The Tihany Volcano

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Department of Petrology and Geochemistry, Budapest

The Tihany Volcano (TV) represents one of the earliest (6–7.5 Ma) volcanic products of post-extensional alkaline basaltic volcanism in the Pannonian Basin, eastern Central Europe. It is a complex volcano located in the Balaton Highland. Volcanic activity began with explosive phreatomagmatic eruptions followed by magmatic explosive events. There is no evidence of any lava flows. The explosive volcanic activity is characterized by loss of mechanical energy due to the continuously decreasing water/magma ratio. Eruption of basaltic magma could have been enhanced by activation of NW–SE and N–S trending faults. The rising basaltic magma interacted with variable amounts of groundwater and possibly with a small amount of surface water resulting in initial hydrovolcanic explosive eruptions. During this stage, the interaction between magma and unconsolidated, water-saturated sediment was the major control of the explosions. Wet base surge and fall deposits, with large amounts of deep excavated lithic fragments, dominate the early volcanic successions. These are interpreted as products of a new type of maar/diatreme volcanism. The occurrence of large amounts of country rock and matrix-supported, massive beds indicates that phreatomagmatic eruptions were characterized by high mechanical energy and probably a continuous water supply from deep karst aquifers, during the main phase of the eruptive history. We propose that the explosive volcanic activity of Tihany occurred under subaerial conditions in a fluvial basin. The hydrovolcanic activity was followed by a Strombolian and minor Hawaiian magmatic explosive period, indicating a termination of water supply from the basement rocks. The Strombolian scoria cones were built up inside the tuff rings and spatter cones were developed in the northern part of the volcanic complex. Following the eruption, lacustrine deposition occurred in the local maar basins.

Key words: explosive volcanism, phreatomagmatic, hydrovolcanic, maar, diatreme, Pannonian Basin

Introduction

During the Neogene various types of volcanic activity took place in the Pannonian Basin (central Eastern Europe; Downes and Vaselli 1995). Miocene to Quaternary subduction-related calc-alkaline volcanism occurred along the northern to eastern margin of the Pannonian Basin. Behind this volcanic arc, Middle Miocene syn-extensional, intermediate to acidic and potassic to

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ultrapotassic magmatism was followed by widespread, Late Miocene to Quaternary, post-extensional, alkaline basaltic volcanism (Harangi 1999). The alkaline basaltic volcanoes can be found either adjacent to the calc-alkaline stratovolcanic complexes or in the central part of the Pannonian Basin (Szabó et al. 1992). Although many detailed petrologic and geochemical studies on the alkaline basalts have been carried out during the last few years (e.g. Embey-Isztin et al. 1993; Downes et al. 1995; Harangi 1999), only a few volcanological studies of these rocks have been published (e.g. Harangi and Harangi 1995; Harangi et al. 1995; Németh 1997; Németh and Martin 1999). Basaltic volcanoes of the Little Hungarian Plain Volcanic Field (LHPVF, western Pannonian Basin) are characterized by initial hydrovolcanic products, followed by Strombolian and Hawaiian deposits (Harangi and Harangi 1995). Subsequent lava flows usually fill the inner parts of the tuff rings. The Bakony–Balaton Highland Volcanic Field (BBHVF, central Pannonian Basin) is located about 100 km east of the LHPVF (Fig. 1) and comprises around 50 basaltic volcanoes (Jugovics 1968; Jámbor et al. 1981), including shield volcanoes, tuff rings and maars (Németh and Martin 1999; Németh and Csillag 1999). The TV represents the earliest manifestation of the basaltic volcanism in the BBHVF. K/Ar radiometric dating indicates that volcanism began at 7–7.5 Ma (Balogh et al. 1986; Harangi et al. 1995; Balogh 1995), preceding the formation of most of the other volcanoes in the BBHVF. The TV is made up entirely of pyroclastic products, most of them formed by explosive hydrovolcanic eruptions. The volcaniclastic deposits always contain a high proportion of deep-seated lithic fragments; most of them were transported as ballistic blocks. The ratios among the different deep-seated lithic fragments are related to the stratigraphic origin of the fragments, as Lorenz (1986) indicated for the maar-forming process. The most common lithic fragments are Permian Red Sandstone (PRS) fragments. There is an apparent paradox in the sedimentological structures of beds: usually the bedded sequence indicates a high water-content depositional environment, while deep excavation of maar-diatreme volcanoes during eruption is not typically associated with large water flows (Wohletz and Sheridan 1983). In this paper tuff ring-type and tuff cone-type of beds are discussed. The tuff ring bed type refers to layers which were deposited in a dry environment, mostly by dry base surge (usually related to normal maar volcanism, e.g. Waters and Fisher 1971; Heiken 1971; Moore 1967; Wohletz and Sheridan 1983; Fisher and Schmincke 1984; Cas and Wright, 1988; Keller 1974; Lorenz 1986; Chough and Sohn 1990; Sohn and Chough 1989). Tuff cone bed type mean beds which were deposited in a wet environment, mostly by wet base surge (usually related to tuff cone-forming, Surtseyan, emergent volcanism; e.g. Fisher and Schmincke 1984; Cas and Wright 1988; Wohletz and Sheridan 1983; Verwoerd and Chevallier 1987; Godchaux et al. 1992; Sohn and Chough 1992, 1993; White 1996).

Geologic setting and previous studies

The TV is located on a small peninsula along the shore of Lake Balaton, and belongs to the Bakony–Balaton Highland Volcanic Field (BBHVF, Fig. 1). The field lies within the Transdanubian Central Range unit, which is correlated with the Upper Austroalpine nappes of the Eastern Alps (e.g. Majoros 1980; Kázmér and Kovács 1985; Tari 1994). The volcanic products are underlain by Paleozoic to Neogene sedimentary sequences, which are exposed in the surrounding areas (Bakony Mts). The Silurian Lovas Schist (SS) crops out on the northern shore of Lake Balaton. It is a 400–600 m-thick formation that contains alternating, very low-grade metamorphosed psammitic and pelitic beds (phyllites, Lelkes-Felvári 1978). The upper boundary of the Lovas Schist is usually an erosion surface. It is overlain by the Upper Permian Red Sandstone (PRS) series, which can be found almost continuously along the northern shore of Lake Balaton. This is a thick (400–600 m), continental alluvial formation (Majoros 1983). The Mesozoic formations (MF) are represented by various Triassic limestone and dolomite formations, which are also typical of the Eastern Alps. The Neogene sediments were deposited on an erosion surface. They consist predominantly of lacustrine sandstone (Pannonian Sandstone Formation – PS) formed in the brackish Pannonian Lake (Müller and Szónoky 1989). These fine-grained clastic sediments are found along the eastern and southern parts of the Tihany peninsula. The Late Miocene basaltic volcanic rocks of Tihany are part of the widespread, post-extensional, Late Miocene to Quaternary alkaline basaltic series of the Pannonian Basin (Szabó et al. 1992). The volcanic eruption in Tihany occurred in a marshy, alluvial environment. The juvenile fragments are slightly nepheline-normative alkaline basalts, which show slight to strong secondary alteration (carbonatization). They show a slight Large Ion Lithophile Element (LILE) enrichment compared to average Ocean Island Basalt (OIB) and are close to the Gough-type OIB end-member (Harangi et al. 1995). Termination of the basaltic volcanism was followed by intensive hot spring activity. Hot spring pipes are found throughout the peninsula. Although geologic investigations of the Tihany peninsula have a history of more than 100 years, most of these were carried out before 1970. They focused on geologic mapping of the area, and on the identification of the main eruptive centers (Lóczy 1894, 1913; Vitális 1911; Papp 1931; Hoffer 1941a, b). It is noteworthy that Hoffer (1941a and b) already recognized that external water or water-saturated sediment may have played a significant role in the explosive eruptions. Based on detailed geologic mapping of the peninsula, Varrók (1957) proposed a volcanotectonic origin for Lake Külső, Lake Belső and the Rátai-csáva, as well as subaqueous deposition of the pyroclastic products. Geophysical investigations of the peninsula showed a strong positive geomagnetic anomaly and a negative Bouguer anomaly in the Lake Külső region (Bender et al. 1965).

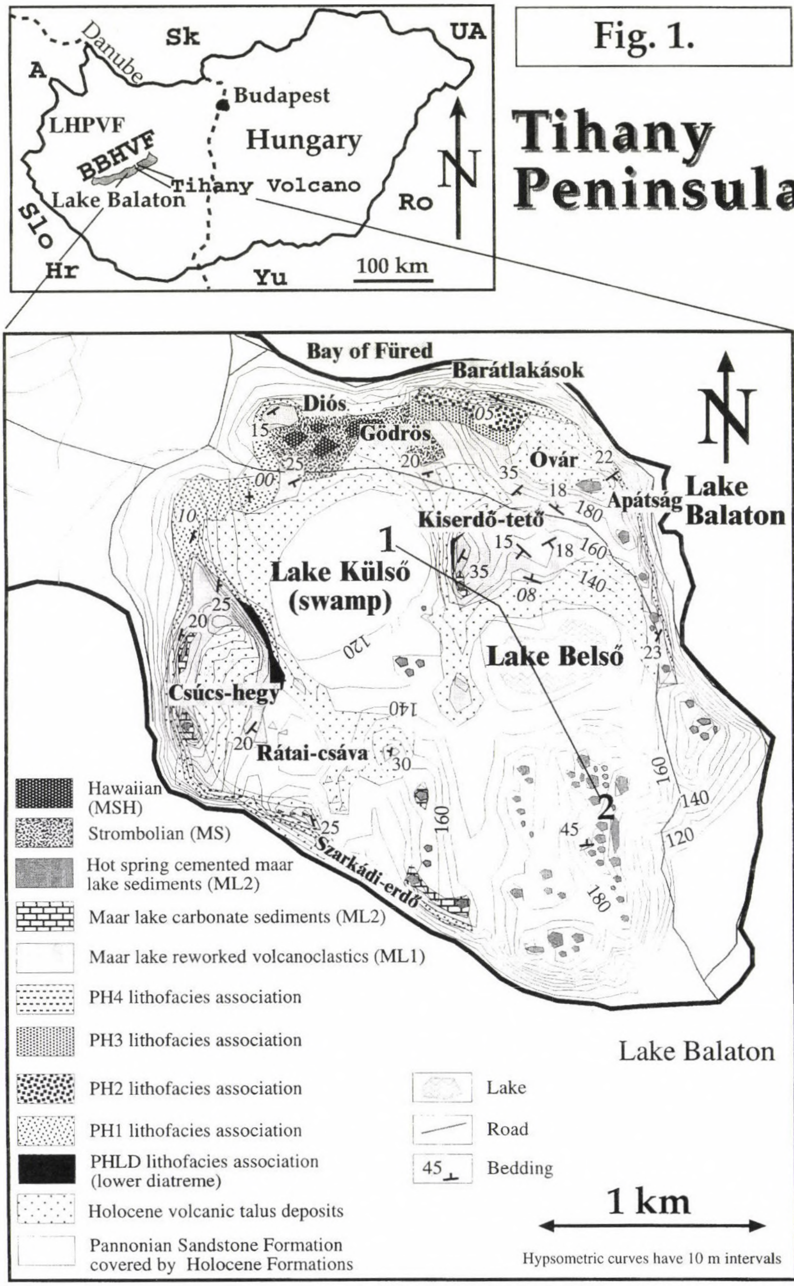

Fig. 1
Geologic map (Hungary, and detailed Tihany Peninsula with topographic contours and cross-section lines)

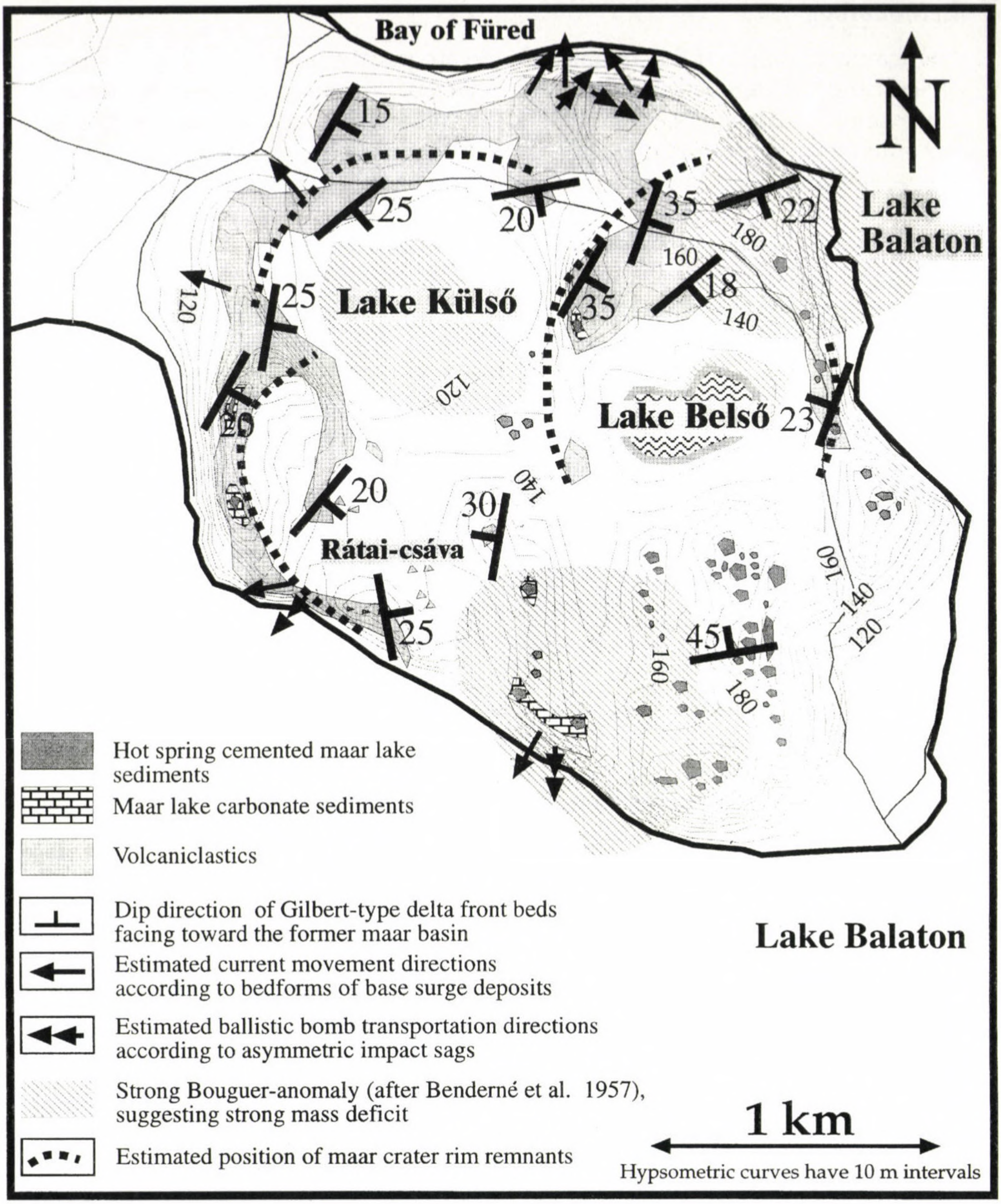

Fig. 2
Reconstruction of positions of different eruptive centers according to measurements of base surge bedforms (calculation of current movement), ballistic trajectory directions from impact sags, and bedding planes of Gilbert-type delta fronts. Note the correlation with the geophysical data after Benderné et al. (1965)

Methodology

Volcanic products of the Tihany volcano are poorly exposed due to intensive agriculture; the detailed field study of selected sites was enhanced by air and space photo analysis (Németh 1997). The pyroclastic beds are generally well cemented with secondary minerals (predominantly calcite), often filling vesicles of juvenile fragments and also the interstitial space between the pyroclasts. The work focused on detailed outcrop description of different sites; detailed centimeter-scale descriptions and sedimentary facies analysis of the volcaniclastic deposits in several sections provided a strong base for interpretation of the eruptive evolution through time and eruptive mechanisms of the TV. In this paper facies observations following the method of Sohn and Chough (1989, 1992, 1993) and Chough and Sohn (1990) are presented. Classification of pyroclasts is based on the scheme of Wright et al. (1980). Granulometric classification and nomenclature of the sedimentary structure are after Sohn and Chough (1989) and McPhie et al. (1993). The bed thickness categories are from Ingram (1954).

Physical volcanology

The pyroclastic deposits of the peninsula can be divided into two main lithofacies associations based on the relative amount of accidental lithic clasts and the sedimentary structures. 1. The initial phreatomagmatic lithofacies associations (PH) consist of different types of lithic-rich block-bearing tuffs and lapillistones. 2. The late magmatic lithofacies association (M) consists of different types of juvenile clast-rich tuffs, lapillistones and pyroclastic breccia. The PH lithofacies association can be found in the northern part of the peninsula, around Lake Külső, and particularly in higher elevation sites close to Lake Balaton (Fig. 4). Beds of the M lithofacies association usually overlie the phreatomagmatic sequences, and crop out in the topographic depression of Lake Külső. In addition, two post-volcanic lithofacies associations can be recognized in Tihany. 1. Maar-lake reworked tephralithofacies association (ML1) 2. Maar-lake carbonate and hot spring deposits lithofacies association (ML2) The ML1 lithofacies usually covers the PH series and is overlain by the ML2 lithofacies. The areal distribution of the separated lithofacies is presented in Fig. 4.

Initial phreatomagmatic lithofacies associations (PH)

PH1 – Alternating lapilli tuff and tuff lithofacies

Description: This lithofacies association, in which several well-bedded lapilli tuff and tuff lithofacies alternate, is more than 10 m thick (Fig. 3). It is characterized by the occurrence of abundant lithic fragments. Antidune structure and cross-bedding are commonly present. Ballistic bombs with impact sags are also common. In general, more deep-seated lithic xenoliths, mainly

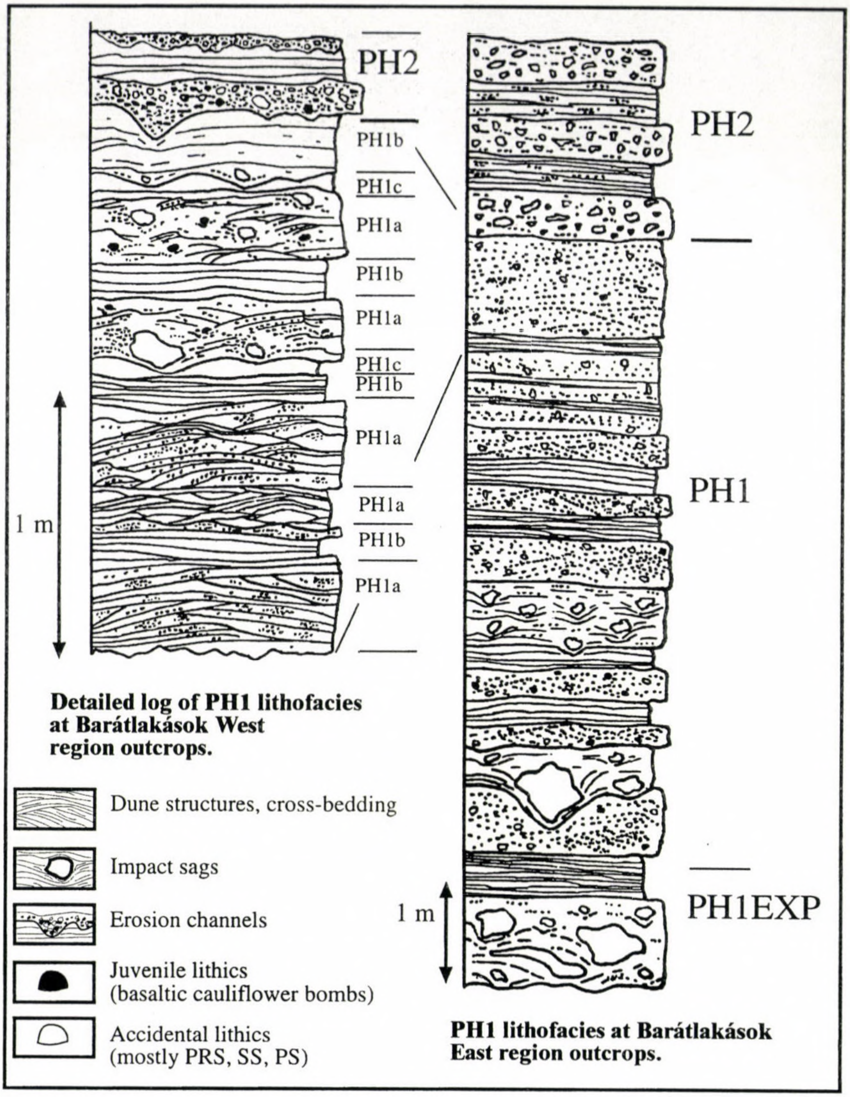

Fig. 3
Stratigraphic column of PH1 lithofacies
PRS - Permian Red Sandstone
SS - Silurian Schist
PS - Pannonian Sandstone

Fig. 3
Simplified stratigraphic column of the PH1 lithofacies association

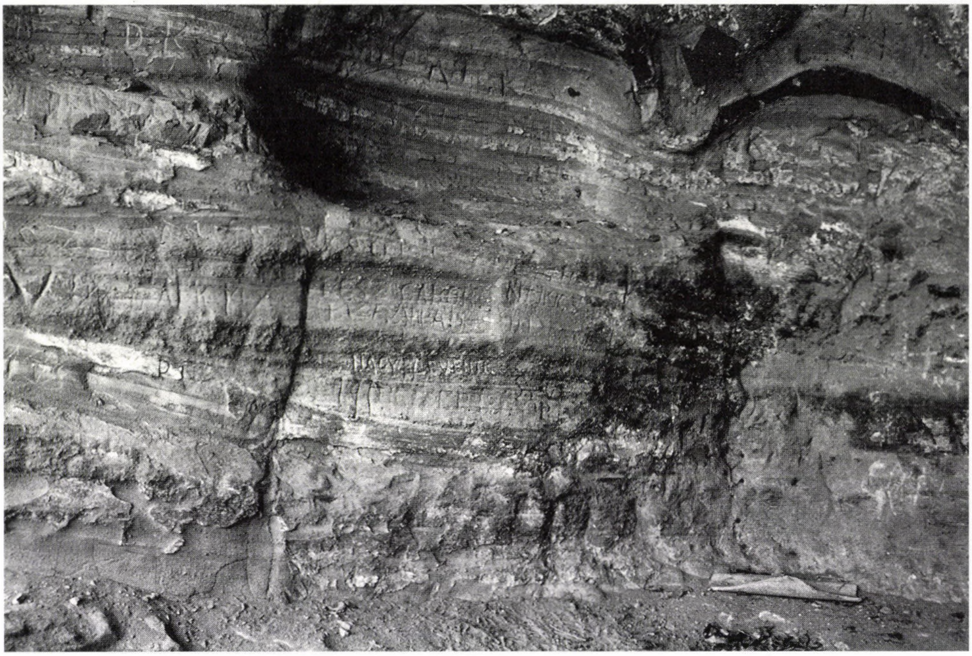

Fig. 4
Overview of the PH1 at eastern side of the Barátlakások outcrop. Note the well-developed dune structures, bedding. The large country rock is Permian Red Sandstone (arrow)

PRS, occur higher up in the stratigraphic column. The lithofacies usually contains very gently dipping beds (max. 15°). Within PH1, several lithofacies can be distinguished.

PH1EXP – Unstructured block-bearing lapilli tuff lithofacies. A light brown-yellowish, poorly sorted lithofacies, characterized by relatively abundant, large (up to 400 mm in diameter), irregularly shaped PS fragments. These fragments are usually elongated and do not show any evidence of thermal contact. In some places, the original sedimentary structure of the sandstone fragments may be recognized. PS clasts are abundant, even in the matrix of the lapilli tuff. The uncommon (max. 5 vol.%), large (up to 100 mm diameter) juvenile fragments are highly vesiculated and usually show cauliflower surface texture. The lithofacies is only present in the lower part of the PH1 lithofacies association with a maximum visible thickness of 2 m.

PH1a – Cross-bedded tuff lithofacies. The average thickness of this lithofacies is 20–25 cm. It is repeated nine or more times, and consists of a grayish-brown, well-bedded, medium-sorted sub-unit comprising more than 90 vol.% of accidental lithic clasts up to 500 mm in diameter. The maximum size of the juvenile clasts is 20–40 mm. They are moderately or highly vesiculated. The upper part of the lithofacies is more strongly palagonitized. In thin section at least 70% of the juvenile fragments are seen to be glassy sideromelan, or slightly palagonitized fragments, with irregular vesicles. The other juvenile fragments are usually black, tachylitic, highly vesiculated, scoriaceous clasts. The vesicles are usually oval. Large (500 mm diameter), ballistic blocks of country rock, with asymmetric impact sags (maximum depression is 15 cm) occur commonly. They are mainly PRS clasts (Fig. 4) and less common SS in the upper part of the lithofacies. There is no evidence of thermal alteration along block contacts.

PH1b – Parallel bedded tuff lithofacies. This is a light or yellow-brown, parallel-bedded, mantle-bedded, laminated and moderately sorted lithofacies. It has a thickness of 5–15 cm-thick and usually overlies the cross-bedded PH1a lithofacies; it is repeated several times. There is a high proportion of accidental lithics, mainly PS fragments and PRS fragments. Some Silurian schist fragments (max. 5 vol.% of the total accidental lithics) occur. The juvenile fragments are made up of a light brown, slightly palagonitized sideromelan glass. Rare black, tachilitic, scoriaceous clasts also occur (max. 20–30 vol.% of the juvenile fragments).

PH1c – Massive tuff lithofacies. This is a massive, well-cemented, unsorted lithofacies. It has a highly undulatory geometry ("boudinage structure") and is 5–15 cm thick. A minimum of four occurrences of this lithofacies is apparent. It is made up mainly of accidental lithic clasts (about 75 vol.%). The juvenile fragments are angular and dense or slightly to strongly palagonitized sideromelan shards. Tachilitic glass is less common. The lithofacies usually contains asymmetrically deformed, rim-type accretionary lapilli (max. 2 mm diameter).

PH1d – Well-bedded lapilli tuff lithofacies. This is a well-bedded sequence of accidental lithic-rich lapilli tuff showing low-angle cross-bedding. It usually overlies the PH1a lithofacies in the eastern part of the Barátlakások locality. Strings of highly vesiculated scoria clasts (clasts have a diameter up to 10 mm and the strings are 500 mm long on average) and well-developed scour fill structures (in 200–400 mm long and 100 mm thick lenses) occur commonly. In the lower part of the lithofacies more PS fragments occur, while in the upper part the deeper seated lithics become more prominent (SS and PRS).

Interpretation: The PH1 lithofacies association was built up by alternating base surge (PH1a, PH1d) and pyroclastic ash fall (PH1b) deposits, with some basal explosion breccia layers (PH1EXP). The high proportion of irregularly shaped PS fragments in the PH1EXP lithofacies indicates that the first explosions occurred at a shallow level, in unconsolidated, water-saturated sediments. The first violent explosions disrupted the PS beds, the fragments of which became the main components of the initial pyroclastic products. The subsequent cross-bedded PH1a lithofacies was formed by base surges. The low-angle cross-bedding, the unsorted, fine-grained character, the adjacent ballistic blocks in the beds, the high proportion of accidental lithic fragments, and the fine bedding support the base surge origin of PH1a lithofacies (Fisher

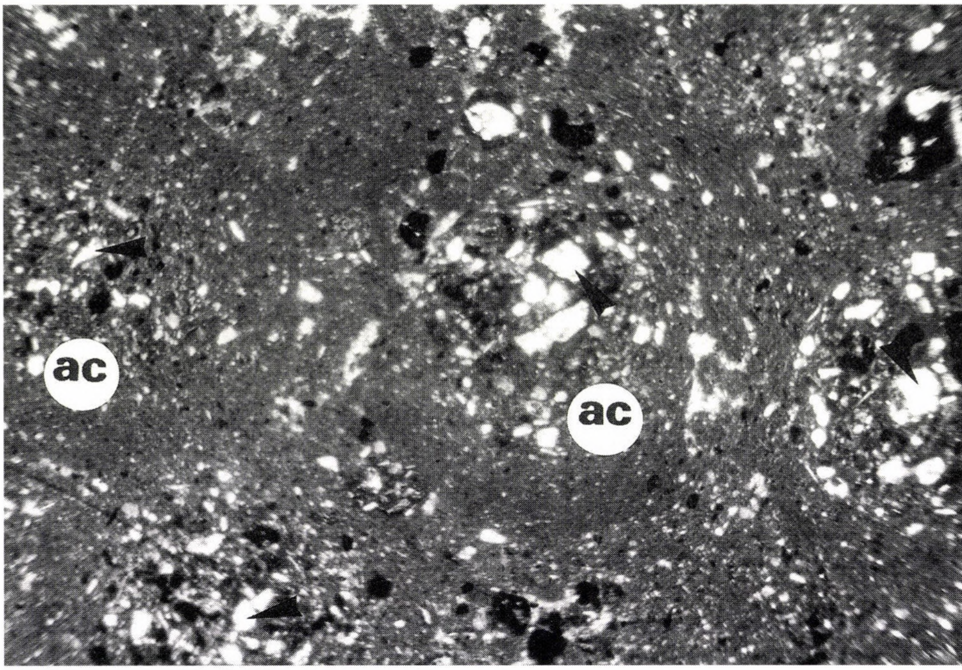

Fig. 5
Microphotograph of fine grained beds of PH1b lithofacies at Barátlakások outcrop showing rim-type accretionary lapilli (ac). Note the homogenous distribution of angular, quartzo-feldspatic fragments (arrows) in the rim and core zone of the accretionary lapilli as well as in the groundmass of the tuff. These fragments derived from the Pannonian sand beds. The longer side of the picture is 2 mm

and Schmincke 1984). The very fine-grained beds of the PH1b lithofacies were deposited directly from the eruption column as ash-fall deposit and covered the PH1a base surge-deposited beds. The water-saturated, fine-grained pyroclastic sediment beds were compacted by the process of water loss (e.g. Rose 1992). This process resulted in the very compact, undulated beds of the PH1c lithofacies. The large amount of deformed accretionary lapilli (rim-type) (Fig. 5) provides strong evidence for a water-rich depositional environment (e.g. Rosi 1992). The large country rocks with underlying impact sags (Fig. 4) originated by ballistic ejection from the vent area (e.g. Keller 1974; Fisher and Schmincke 1984; Cas and Wright 1988). The distribution of these ballistic blocks and bombs and the geometry of bedsags imply a transportation direction mainly from the inner side of the peninsula (from the west), but there are a few samples which show an opposite derivation, i.e. from Füred Bay (from the east). The cross-bedded geometry of several base surge beds in the PH1 lithofacies association supports the same transportation distribution as that of the ballistic blocks and bombs. Therefore, it is suggested that at least two eruption centers could have been active at the same time during the initial stage of the volcanic activity of Tihany. The main eruptive center is inferred to have been located around the Lake Külső region, and another one in the present-day Füred Bay region. The general trend of the distribution of accidental lithic clasts in the lithofacies indicates that the explosion focus must have occurred at increasing depth during the eruptive history (Lorenz 1986 model). Therefore, there is slight evidence for the "inverse" distribution of the accidental lithics compared to the pre-eruptive stratigraphic column. The deep excavated accidental lithics are evidence for a maar type of volcanic activity (Fisher and Schmincke 1984; Cas and Wright 1988; Aranda-Gomez and Luhr 1992, 1996; Büchel and Lorenz 1993). The "wet" bedding characteristic (accretionary lapilli beds, palagonitization, and plastic deformations below the impact sags) shows several lines of evidence for a continuous water supply during the eruptive process. The water supply was probably the water-rich, unconsolidated PS (porous media aquifer). The more common deep-seated xenolites in the upper part of the beds of the lithofacies association suggest continuous drying and downward-migrating explosion focus (Lorenz 1986).

PH2 – Alternating block-bearing lapillistone and lapilli tuff lithofacies association

Description: This 15 m-thick lithofacies is best exposed in the northern region of the peninsula (Barátlakások). It is in erosional contact with the underlying PH1 association, and is built up predominantly of massive block-bearing lapillistone and lapilli tuff beds with a very low angle of dip (max. 15°) (Fig. 6 and 7). Between these massive, unstructured beds, thin (5–15 cm) cross-bedded sequences can be found, containing fewer and smaller lapilli. The main part of the lithofacies association (min. 75 % of the total beds) is a strongly indurated tuff breccia.

PH2a – Massive block-bearing lapilli tuff and lapillistone lithofacies. This is a brown/light brown and gray/light gray, massive or coarsely stratified, 15–85 cm thick lithofacies with several individual beds. It is repeated several times and makes up the principal part (75%) of the PH2 lithofacies association. Impact sags are commonly observed. Behind them, grain-supported lapilli lenses often occur as scour fills. Bombs and blocks (30 vol.% of total) have a maximum diameter of 700 mm and consist of PRS (50 vol.%), SS (25 vol.%), PS (15 vol.%), MF (5 vol.%) and crystalline basalt or moderately vesiculated scoria (5 vol.%). The ratio of deeper-seated accidental lithics of the total volume dramatically increases upward in the lithofacies. The PRS and SS fragments are usually angular, but the PS fragments often show plastic deformation. The elongated SS fragments are generally imbricated. The often fusiform-shaped juvenile bombs present chilled margins around 1 cm thick. They contain large amounts of vesicles, which are usually elongate and have a maximum diameter of 1–2 cm, and in several places are filled by calcite or sand. The blocks and bombs are usually associated with impact craters (75% of the all blocks and bombs), although some of them occur without any impact sags. The matrix of

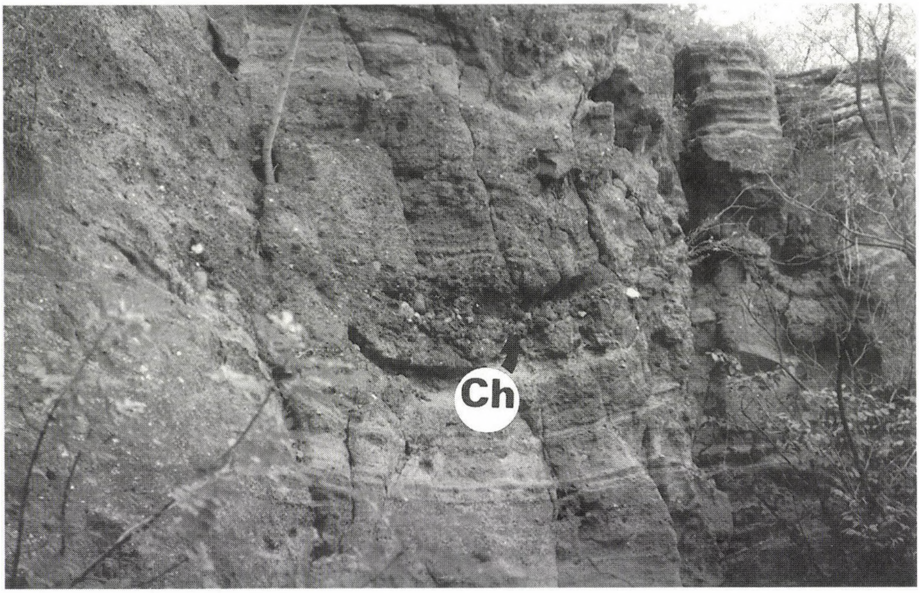

Fig. 6
Overview of the PH2 lithofacies association. Note the channel structure (Ch) in the middle of the picture built up by mostly accidental lithic fragments (Permian Red Sandstone and Silurian schist bombs and blocks)

this unit is composed mainly of fine-grained PS fragments and less highly vesiculated, angular juvenile fragments (mainly altered sideromelan glass shards).

PH2b – Cross-bedded lapilli tuff lithofacies. This sequence is a light gray to light brown, coarsely stratified, well-sorted 15–35 cm-thick lithofacies. Low-angle cross-bedded layers are buried by fine-grained, parallel-bedded tuff lamina. The lithofacies is rich in accidental lithic fragments such as PRS, SS and PS (about 80 vol.% of total volume). The cognate lithics and juvenile basaltic fragments are poorly rounded and incipiently vesiculated. Scoriaceous fragments occur in grain-supported lenses, which are found frequently as scour fills behind impact sags.

Interpretation: The PH2 lithofacies association is a succession of phreatomagmatic base surge and fallout deposits with interbedded, reworked volcaniclastic debris flow layers. The block-bearing, massive, unstructured units may represent near-vent sedimentation of a highly concentrated, turbulent, cohesive flow (Sohn and Chough 1990). The imbrication of schist fragments is consistent with high-density, laminar current movement. The near-vent origin is supported by the presence of large ballistic blocks and bombs (Fisher and Schmincke 1984). The plastic deformation of PS clasts may indicate the water-saturated, unconsolidated state of this rock during the eruptions. Thus,

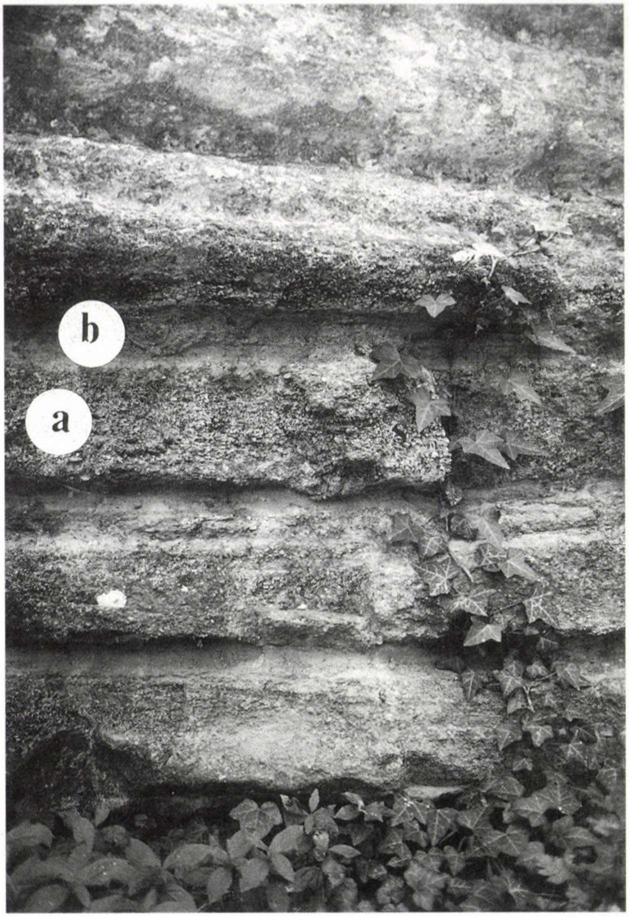

Fig. 7
Close up view of the PH2 lithofacies at Barátlakások west outcrops where PH2a and PH2b is alternating. The PH2b in this site is matrix rich, finer grained lapilli tuff. Most of the lapilli fragments in both PH2a and PH2a are Silurian schist (s) fragments (min.70 %, visual estimation) and chilled, fine grained basaltic lapilli

it is suggested that the interaction between hot magma and water-saturated sediments played an important role in forming the rocks of PH2. The alternation of block-bearing lapillistone and lapilli tuff beds implies a change of explosive energy and/or the geographical position of the explosive centers during the volcanic activity. The block-bearing deposits could have originated by large, intense, discrete phreatomagmatic eruptions similar to those described by Houghton and Hackett (1991) in White Island, New Zealand. The fine-grained lapilli tuff beds could have been connected to more diluted pyroclastic surges. The plane-parallel, very thin layers which cover the cross-stratified beds in the PH2b unit are interpreted as pyroclastic fall deposits (Fisher and Schmincke 1984). The erosional contact between the PH1 and PH2 lithofacies association and their geographical distributions could imply that the PH2 lithofacies association represents a deposit from a different vent than that which produced

PH1. The extremely high proportion of deep-seated accidental lithics, and the strongly indurated, diffusely bedded character of the lithofacies is interpreted as the result of a deep excavated maar/diatreme explosive process with strong interaction between water-saturated sediments and magma. The indurated character is a sign of the high water-supply rate during the explosive process. The high water supply and the deep excavated fragments suggest that during the eruption there must have been a rich underground water supply. Several beds of the PH2a lithofacies show a high calcite content (spathic and microcrystalline) and a reworked character. This is interpreted as interbedded debris flow deposits in the near-vent primary volcaniclastic sequences. Since no information of the exact locality of different vents is available an exact reconstruction is problematic.

PH3 – Cross-bedded scoria-rich lapilli tuff and lapillistone lithofacies association

Description: This is a well stratified, gently dipping (max. 15°) maximally 5 m-thick lithofacies association comprising alternating grain-supported and matrix-supported lapillituff and lapillistone units, with sedimentary structures such as dunes, grain-supported lenses and channels. It is in continuous sedimentary contact with the PH2 association.

PH3a – Grain-supported lapillistone and lapilli tuff lithofacies. This is a dark gray, 35–45 cm-thick lithofacies, which occasionally shows low-angle cross-stratification. The angular juvenile fragments (max. 25 vol.%) contain elongated vesicles filled with calcite and zeolite. Chilled margins are occasionally observed. In thin section the scoriaceous tachilytic shards are more abundant than the yellowish glassy sideromelan fragments (the ratio is around 3:1). The unit is rich in accidental lithics. The basic elements of the accidental lithics are PS and PRS fragments (in the same proportion). There is only a reduced amount of small schist fragments (max. 5v%).

PH3b – Alternating fine-grained tuff lithofacies. This 15–35 cm-thick lithofacies is built up by alternating tuff beds with well-developed cross stratification. Plane-parallel, 1–2 cm thick, accretionary lapilli rich beds bury the dune structures. The lithofacies contains a maximum of 10 vol.% of PRS bombs (100–250 mm diameter) associated with relatively shallow (max. 20–50 mm) impact craters below them.

Interpretation: The PH3 lithofacies association represents a series of pyroclastic surge and fall deposits. The higher concentration of scoriaceous juvenile fragments and the absence of chilled margins imply the increasing role of magmatic volatiles in the explosive process (e.g. Fisher and Schmincke 1984; Houghton and Schmincke 1986, 1989). The increasing magmatic explosivity may imply a continuous drying of the system (e.g. Lorenz 1986; Büchel 1993). The alternating of two kinds of lithofacies (PH3a and PH3b) in this lithofacies association could have originated by the changing of explosive focus depth and/or the horizontal position of the erupting vents. The unit richer in juvenile fragments (PH3a) could be related to a shallower explosive focus. The cross-bedded unit (PH3b) probably represents a more phreatomagmatic effect during the explosions (e.g. Houghton and Schmincke 1986, 1989).

PH4 – Alternating laminated tuff and block-bearing lapilli tuff lithofacies

Description: This lithofacies occurs in the western area of the peninsula near Csúcshegy and Szarkádi-erdő, overlying the PH1 lithofacies association. It is a 10 m-thick unit composed of alternating fine-grained and block-bearing tuff beds. The fine-grained units are built up of several gray or brown tuff beds. Occasionally very low-angle cross-stratification may be observed in this part of the lithofacies. All of the beds exhibit very gentle dips (max. 5–10°). The well-bedded tuffs alternate with massive, block-bearing, 15–25 cm-thick beds. The clast population consists of PS (80 vol.%, of max. 600 mm diameter), PRS (15 vol.%, of max. 100–250 mm diameter) and juvenile scoriaceous basaltic fragments (5 vol.% of max. 50–100 mm diameter). The juvenile fragments usually show mud-coated margins and very fine vesiculation. The PS fragments usually have thermally altered margins (10–20 mm wide). The blocks are not associated with deep impact craters or plastic deformations. Locally, several microfaults may be observed around the large country rocks. The matrix of the lapilli tuff contains a very high amount of PS debris (70 vol.% of the matrix).

Interpretation: This lithofacies can be correlated with the PH2 lithofacies association. The main difference between them may be the water content of the sedimentary pile where the explosive eruption occurred. The PH4 lithofacies provides evidence of a dry environment, such as micro-faulting around impact sags, no plastic deformation, burnt surfaces of fragments and well-bedded tuff and lapilli tuff beds (Fisher and Schmincke 1984; Cas and Wright 1988; Godchaux et al. 1992). The higher amount of PS clasts in the PH4 lithofacies could be explained by a shallower level of explosion, or by thicker PS beds beneath the eruptive center. The PH4 lithofacies is associated with a different vent than the PH1, PH2, or PH3 lithofacies associations.

PHLD – Lower diatreme, crater filling bomb and block-bearing lapilli tuff

Description: In two locations small outcrops of coarse-grained, matrix-supported pyroclastic breccia with massive structure and fine, quartz-feldspathic, sandy matrix is known (Kiserdő-tető and Csúcshegy, lower level). The pyroclastic breccia is unsorted, non or very weakly bedded in several well-distinguished places. They are abundant in large clasts, as PRS blocks up to 50–70 cm in diameter; smaller (up to 10 cm diameter) PS fragments and quartzite (probably from the schist beds) are also present. The blocks are usually sub-rounded and occasionally cause impact sags, but mostly occur floating in a fine-grained matrix. The larger accidental lithics form a vertical alignment. Small, coarse-grained, scoriaceous fragments form grain lenses with varying dip angles. The matrix of the lithofacies is rich in quartz-feldspathic sand with palagonitized juvenile lapilli. The juvenile lapilli have a moderately (light brown sideromelan) to completely (dark, black sideromelan) palagonitized feature. In the lithofacies tachilityc lapilli also occur with higher vesicularity. The sideromelan lapilli are non or just slightly vesicular. Cored lapilli of PRS, PS and LR are common, some of them with a wide, fine-grained rim of palagonitized volcanic glass and/or quartz-feldspathic sand and silt (up to 3 cm in thickness). Irregular fluidization zones with accompanied accretionary lapilli are common. In both localities the PHLD lithofacies is covered by reworked maar lake tephra deposits.

Interpretation: The colorful lithology of the pyroclastic breccia with poor sorting characteristics suggests a vent-filling position as a lower diatreme deposit (term by White 1990), where the fall back tephra was emplaced en masse at the base of the funnel-shaped vent. The fluidization zones with accretionary lapilli and the non or weakly bedded character of the deposit suggest reworking in wet conditions (White 1990). The occurrence of cored lapilli and vertical transportation features represent strong reworking in a phreatomagmatic vent (Lorenz 1988, 1986; White 1991b, 1996). The distribution of the PHLD lithofacies and its relationship with the overlying maar lake reworked tephra beds suggest phreatomagmatic vent locations for two sites in the field. The earlier core descriptions from the Tht-2 borehole in the Lake Külső basin shows similar features as those described above: tuff breccia with a non-bedded, sandy matrix accompanied by a larger number of accidental lithics (PRS, PS) and scoriaceous fragments (Hung. Geol. Surv. Data Base). This description, together with the negative Bouguer anomaly in the area, suggests a maar crater fill origin of the sediment. The higher scoria fragment content of the deposits represents late scoria infill in the former maar basin, as a remnant of upper diatreme structures in a former maar basin (White 1991a and b).

Late magmatic deposits (M) lithofacies association group

MS – Strombolian scoria lithofacies association

Description: Strombolian scoria cone remnants occur in the northern part of the peninsula, around the Kiserdő-tető area and along the shoreline of Lake Külső. Occasionally, Strombolian scoria beds occur as beds a few centimeters thick or lenses in the PH3 lithofacies, as a sign of penecontemporaneous magmatic and phreatomagmatic explosion activity. The average thickness of the scoria bed sequences is around several meters (max. 10 m), usually dipping at 25–35°. The succession of the scoria cone deposits can be divided into two main alternating lithofacies. One of them consists of 20–40 mm-thick beds with fine-grained, well-bedded, gray to light gray deposits. This lithofacies is usually 100–150 mm thick. The other one consists of 50–100 mm-thick, grain-supported beds. This unit is generally 300–400 mm thick. The deposits are usually well sorted. The pyroclasts are predominantly moderately to highly vesiculated juvenile fragments, showing irregular or fusiform shapes. Breadcrust rind bombs occur in the Stormbolian deposits, mainly at the Gödrös – Diós region. Maximum grain size in this area is 400 mm; average fragment size is typically a few cm diameter (10–50 mm). Accidental lithics are usually visible but the amount of this kind of fragment in the entire sediment is never higher than 10 vol.%. The most common accidental lithic is the PRS (10–150 mm diameter). The scoria lapilli may show a burnt surface, or a red, fine-grained cover. The vesicles of scoria fragments are filled with white, quartz-feldspathic fragments (peperite structure?). In the Gödrös – Diós area the juvenile fragments contain large (10–20 mm diameter) pyroxene and olivine crystals, peridotite or lherzolite xenoliths (20–50 mm diameter). In several places it is possible to find dense, angular, volcanic (probably basaltic) fragments. Locally the scoria beds contain thin (10–20 mm thick), reddish, fine-graded beds.

Interpretation: The entire scoriaceous lithofacies association is interpreted as a magmatic (Strombolian) explosive product. In these explosions the role of external water was not very important, but the vesicle-filling sand indicates that the unconsolidated PS, or PS and crater-filling tephra slurry, occurred at this time near the Strombolian eruptive centers. Dense blocks could have been derived from degassed magma, which had remained in the vent for a long time (Houghton and Hackett 1984). The Strombolian scoria beds in the phreatomagmatic series may imply periodic disturbance in the magma/water ratio, or penecontemporaneous Strombolian and phreatomagmatic activity, such as Houghton and Schmincke (1986, 1989) described in a similar situation in the Rothenberg Volcano (East Eifel Volcanic Field, Germany). The numerous red, very thin, fine-grained layers may be signs of small interruptions between the eruptions. These layers may be interpreted as paleosoil levels. The geographically very fragmented and separated distribution of the beds of this lithofacies could be related to a strong erosional process, leaving only the root near the vent zones.

MSH – Strombolian – Hawaiian spatter lithofacies association

Description: In the northern part of the peninsula, around the Diós area, above the scoria deposits several spatter deposit beds occur. The deposit consists exclusively of elongated, moderately vesiculated, black, juvenile pyroclasts with characteristic chilled rims. The vesicles are usually large (10–30 mm) and elongated, filled with secondary minerals or altered lherzolite xenoliths. There is no clear evidence for any clastogenic lava flows. The deposit occurs around a highly irregularly-shaped feeder dyke-system near Diós. The deposit is rich in highly vesicular, large spindle bombs.

Interpretation: This lithofacies may represent the most recent events in the volcanologic history of peninsula. The sedimentological characters of this sediment show Hawaiian-type lava spatter deposition (Fisher and Schmincke 1984; Godchaux et al. 1992). The chilled margin and a small amount of

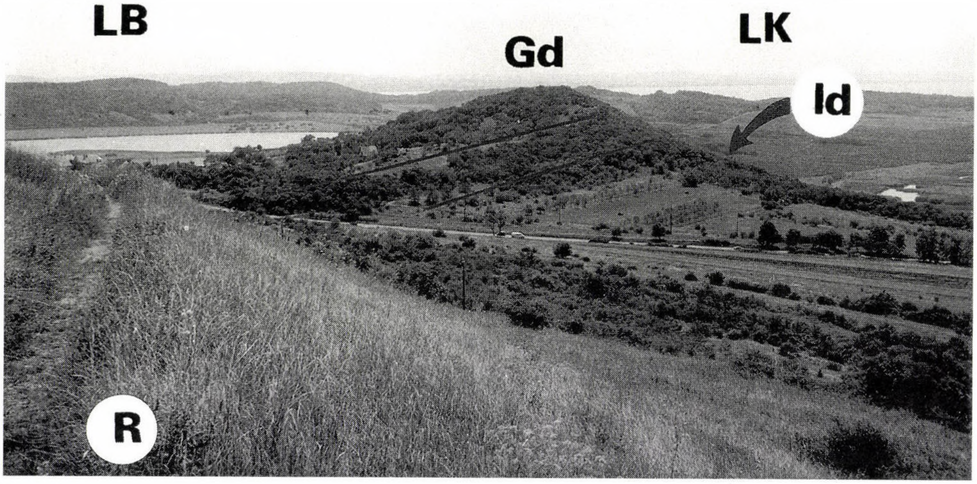

Fig. 8
Overview of a remnant of a Gilbert-type delta front (Kiserdő-tető). View from the Barátlakások region looking toward southwest. The lake in the lefthand side is the Lake Belső (LB). The swampy basin in the righthand side is the Lake Külső (LK). The hill in the middle of the picture (Gd) built up by reworked tephra, as maar lake deposits representing remnant of former Gilbert-type delta front in former maar basin. In small outcrops below the reworked tephra beds (on the picture, behind the hill where the bended arrow point) lower diatreme beds occur (ld). The left bottom corner of the picture is the Barátlakások region, representing crater rim (R) base surge and fall out successions (PH1, PH2, PH3 lithofacies associations). The lines on the hill side represent dipping of beds

vesiculation of pyroclasts can be related to shallow water ponds (inside the maar crater region), or a slight phreatomagmatic influence during fragmentation. The MSH localities represent local magmatic explosion centers.

ML – Maar lake sediments lithofacies association

ML1 – Maar lake reworked tephra lithofacies

Description: This lithofacies is encountered in the more elevated areas around Lake Külső and the northern part of Lake Belső (Fig. 8). The beds always dip steeply (25–35^{o}) and are usually graded (inversely to normally-graded). The originally open framework structure is occluded by secondary mineral fillings (Fig. 9) (mostly calcite or zeolite). The samples contain a large proportion of vesiculated, rounded, semi-rounded, and scoriaceous juvenile fragments. Non-volcanic lithics occur in small proportion (less than 15 vol.%) and they are usually PRS or small sandy fragments from the underlying PS. The lithofacies contains a relatively high proportion (15 vol.%) of broken crystal

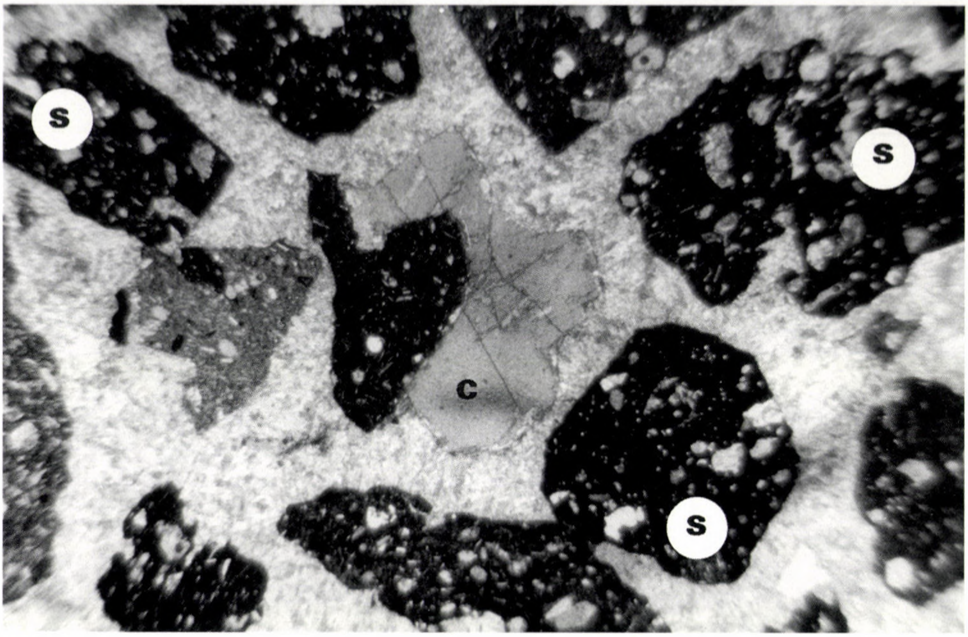

Fig. 9
Microphotograph of calcite cemented scoriaceous (s) fragments from coarse grained reworked tephra bed (Kiserdő-tető). Note the broken crystal attached to scoria lapilli in the middle of the picture (c). The longer side of the picture is 10 mm

fragments such as olivine, pyroxene and quartz. The lithofacies can be separated into two sub-lithofacies, which alternate cyclically. One of them is usually inversely graded and coarse-grained, with a mostly open framework (or, originally open framework but now calcite or zeolite-filled). The other one is usually finer-grained and in several places shows low-angle cross-stratification. Alga rims coat the grains in several places.

Interpretation: The lithofacies is interpreted as reworked maar lake volcaniclastic sediments (Németh 1999), which were probably transported by grain flow (inverse-to-normal graded, coarse-grained sub-lithofacies) and/or turbidity flow currents (fine-grained, bedded, cross-bedded sub-lithofacies) into the maar lake (White 1989; 1991b; Büchel 1993; Drohmann and Negendank 1993). The reworked origin is supported by the large amount of different types of volcanic glass (tachilityc and sideromelan), rounded clasts, the high amount of calcite in the matrix, and the bedded, graded character of the sediment.

ML2 – Maar lake carbonate sediments with hot spring pipes lithofacies

Description: The lithofacies overlies the ML1 one discordantly. It consists of fine-grained, laminated, micro-laminated carbonatic beds (Fig. 10). Hot spring pipes usually disturb the laminae. There are a wide variety of soft sediment

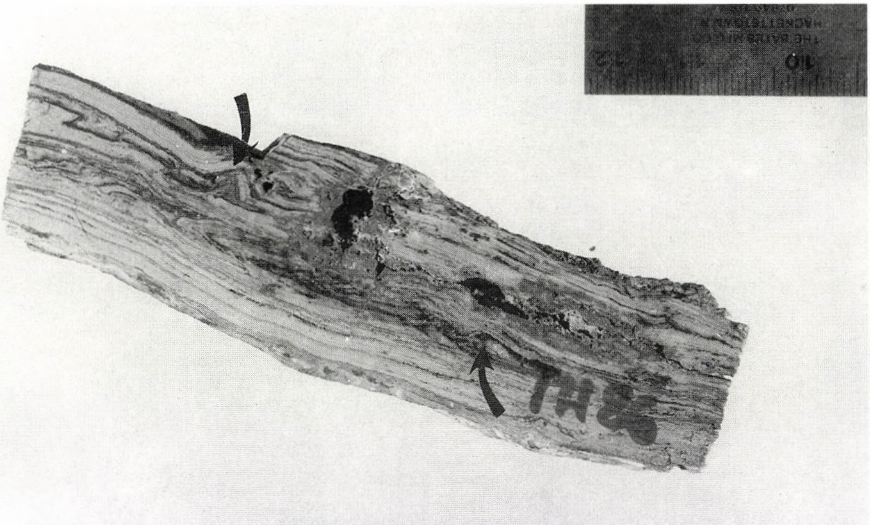

Fig. 10
Hand speciment of a laminated maar lake carbonate bed sequence (Csúcs-hegy). Note the soft sediment structures (arrows). Scale bar is appx. 5 cm

deformations around the hot spring pipes. The carbonate sediments are strongly cemented by quartz around the pipes, and usually contain a small (less than 5 vol.%) proportion of very fine-grained volcanic ash lamina.

Interpretation: The carbonate sediments always overlap the redeposited volcaniclastic lithofacies (ML1), which was interpreted as maar lake sediment. The contact between the two lithofacies is discordant. This is assumed to be a sign of the strong change in the depositional environment. This change might be related to the total erosion of the tuff ring rims and scoria cones around the maar lake(s) when the depositional environment of the maar lake became a quiet lacustrine one. The quiet depositional environment is supported by the occurrence of fine laminae. The disturbance, in the form of soft sediment deformations in the fine lamina layers, might be related to the hot spring activity at the bottom of the maar lake, or penecontemporaneous volcanic explosions close to the actual maar. Small-scale earthquake occurrences around the active maar deposition area also could cause soft sediment slumping or deformation.

Evidence of maar/diatreme volcanism at Tihany and the location of their explosive centers

Maars represent small monogenetic volcanoes and are essentially craters cut into pre-eruptive surfaces (Lorenz 1985). Their pyroclastic ejecta consist of a mixture of pyroclasts of juvenile origin and a large proportion of pyroclasts of country rock origin (Lorenz 1985, 1986). Ejection of large amounts of fragmented country rocks causes a mass deficit at their depth of origin and a consequent collapse of the wall rocks surrounding and overlying the explosion chamber (Lorenz 1985, 1986). Usually the deep maar basin is surrounded by a thin, low-angle dipping tephra rim, in several cases of asymmetric shape (e.g. Lorenz 1985, 1986; White 1991a, b). Tuff rings are small volcanoes with a very wide central crater and a low, broad cone (or rim) (e.g. White 1991a). Tuff rings produce abundant, very fine-grained, blocky pyroclasts (Wohletz 1986). Maar ejecta is much like tuff-ring ejecta (Wohletz 1983) but also contains abundant non-juvenile debris disrupted from the pre-eruptive substrate (e.g. Lorenz 1975; Leys 1982, 1983). Usually tuff rings also have a small collapse structure but it is smoothly mantle-bedded by the tuff ring's own ejecta (e.g. Lorenz 1985). Tuff cones are high-standing, cone-shaped volcanoes that are genetically related to tuff rings (Fisher and Schmincke 1984), although often composed of somewhat coarser-grained, more strongly palagonitized tephra and a very small amount of disrupted country rock content (Wohletz and Sheridan 1983). In general, magma erupting in wetter parts of a monogenetic volcanic field will encounter and mix with water to produce violent "hydrovolcanic" eruptions (Heiken 1971; Wohletz and Sheridan 1983; Fisher and Schmincke 1984; Cas and Wright 1988), which build tuff cones, tuff rings and maars along streams and in lowlands (Hamilton and Myers 1963; Heiken 1971; Lorenz 1973; 1986; Nairn 1979; Kienle et al. 1980). In dry, well-drained settings, monogenetic volcanoes most commonly form a combination of pyroclasts and fluid lava, producing lava flows, steep-sided scoria cones, and widespread layers of fallout scoria (White 1991a). In general the pyroclastic rocks in Tihany are rich in accidental lithic fragments from the pre-volcanic substrata. The lower pyroclastic units usually contain in several cases more than 75% of accidental lithic fragments alongside mostly microcrystalline, blocky to slightly elongated, non- or slightly vesiculated, microvesiculated sideromelan glass fragments. The above- mentioned characteristics of the juvenile fragments establish the events of phreato- magmatic explosive activity because juvenile clasts of phreatomagmatic origin of many volcanoes are usually angular, chilled, vesicle-poor magma fragments and were believed to originate by chilling and fragmentation of magma coming into contact with groundwater or surface water (Heiken 1972; 1974; Lorenz 1975).

The extremely large amount of lithic fragments in the groundmass of the pyroclastic rocks, as well as the large country rock fragments, and the mostly ballistic bombs and blocks are clear evidence of the formation of deep excavated maar craters (Lorenz 1986; White 1991a, b). The localization of the former maar craters was based on previous gravimetric investigations at Tihany, which showed three main negative Bouguer anomaly fields, strongly suggesting mass deficit caused by explosive disruption in those areas (Benderné et al. 1965). Large, Gilbert-type delta fronts, prominent among the high-level deposits, are made up of steeply-dipping beds of reworked tuff that are exposed in large cliffs along Kiserdő-tető and Csúcshegy that define the edge of former maar crater (ML1 lithofacies) (e.g White 1989, 1990). Usually in this area the reworked tephra beds are covered by late, usually strongly silicified maar lake carbonate beds, representing quiet, closed-basin lacustrine sedimentation (ML2). The occurrence of crater-filling fall-back tephra deposits in two locations in surface outcrops and from a core from the eastern Lake Külső basin suggests that a minimum of three maar basins was formed during the TV activity. It is possible to reconstruct an eastern maar around the Lake Belső – Óvár area, another one around the Rátai-csáva – Szarkádi-erdő area, and possibly a third, major maar basin, probably of multiple evolution, in the Lake Külső basin. According to the bedform characteristics of base surge deposits and the ballistic transportation directions of large country rocks, almost all the calculations indicated transportation directions from within the recent peninsula, which consists of the previously established minimum of three eruptive center locations, in outward directions. Most of the primary volcaniclastic deposits are wet base surge, high-concentration, turbulent density currents. A relatively small proportion of the total beds represent fallout tephra or low-density, dry base surge deposits. These two facts suggest that the eruptions in Tihany did not form high eruption clouds. The eruption columns mostly moved radially away from the individual eruptive centers and their areal distribution must have been relatively small due to their high concentration.

Volcanological history of Tihany Volcano

The latest K/Ar data show that the volcanic rocks in the Tihany Peninsula are at least 6–7.56 million years ago. This implies that their development is the oldest in the volcanic area of the BBHVF (Balogh et al. 1986; Harangi et al. 1995; Balogh 1995). At that time a marshy shoreline area of the receding, low-salinity Pannonian Lake is presumed to have been where the Tihany Peninsula is now located. The first volcanic explosions may have taken place in such a paleogeographic environment. Although exposures of a transition of PS beds to pyroclastic rock are not known in the area of the peninsula, it can be stated that PS occur, since they exhibit characteristics of becoming increasingly shallow and limnic with time, indicating a change in paleoenvironment (Kázmér 1990; Müller and Szónoky 1989). To summarize, it can be suggested that the area might have been a marshy, alluvial plain at the time of the first explosions. The pyroclastic formations in a stratigraphically lowermost position show (in view of their large additional lithic grain content) that the explosions might have taken place at a greater depth beneath the surface, but in the PS. The basaltic melt intruding the PS beds with high water content caused high-energy phreatomagmatic eruptions. The first eruptions may possibly have taken place at the center of the peninsula. Beneath the subsurface, the uprising melted basalt reached the water-rich beds and phreatomagmatic, Taalian-type explosions occurred due to abrupt steam development. The explosions propagated out from the bed found over the explosion center; therefore, the formations belonging to this phase are abundant in rocks that are located at a great depth beneath the surface. During the explosion, the eruption cloud is not likely to have been too high, and base surges are assumed to have played a major role in the processes (PH1 lithofacies association). Simultaneously with the main eruptions in the central part of the peninsula, an eruption similar to the previously-described one is likely to have taken place in the area of the Füredi-öböl (Bay of Füred), not far from the present-day shore of Lake Balaton. This is indicated by ballistic bombs and blocks indicating different throw directions found in outcrops of the Barátlakások region (PH1 lithofacies association). Following the first eruptive cycle a new vent was opened, probably within the Óvár region, which produced pyroclastic deposits above the PH1 lithofacies association. The product of this event can be observed at the western side of the Barátlakások outcrop as a PH2 lithofacies association and above it the PH3 lithofacies association. The PH2 association shows an erosional contact with the underlying PH1; it was deposited over the previous volcanic product. The PH2 lithofacies association represents a high-energy mass flow deposit as a near vent volcaniclastic series. The PH3 lithofacies is in continuous contact with the PH2 association. It may be represented as a continuous transition from the phreatomagmatic explosive events to the more magmatic ones. In the next stage, a new vent was opened on the western shore of the peninsula (Csúcshegy) which created pyroclastic beds representing an explosion (probably occurring at a shallower depth) and which led to base surge (PH4 lithofacies). This phase already indicates a decrease in the water content of subsurface beds and the absence of external water recharge. During the volcanism, in the period of phreatomagmatic activity, tuff rings were formed by the influence of subsurface, high-energy phreatomagmatic explosions around the current explosion centers. In this approach the volcanic activity of the TV is classified as an intra-basin maar/diatreme volcanic complex with *Taalian-type* explosions, using Kokelaar's (1986) terminology. When a tuff ring, and probably related maar structures, were produced, the local depressions (maar craters) functioned as local, small-scale sedimentary basins (e.g. White 1991a). In these basins reworked volcaniclastic deposits and lake carbonates were created (Fig. 11). The ML1 lithofacies association probably represents a grain flow and/or turbidity current depositional process, which reworked the eroded tephra from the crater rims. On the flanks of the maar basins the unconsolidated tephra fragments moved down into the deep region of the basin. Following these events, very fine-grained, carbonate, laminated lake sediments were deposited (ML2 lithofacies association). These sediments show considerable evidence of soft sediment movement in the form of slides, slumps and compacted beds. These sediments are truncated by several hot springs, which were probably active during sedimentation. In several places violent degassing caused truncation and gas escape channels; as a result hot spring chimneys could be identified. Large-scale and widespread sediment disturbance may have been related to earthquakes, which probably occurred next to the explosive vents of other

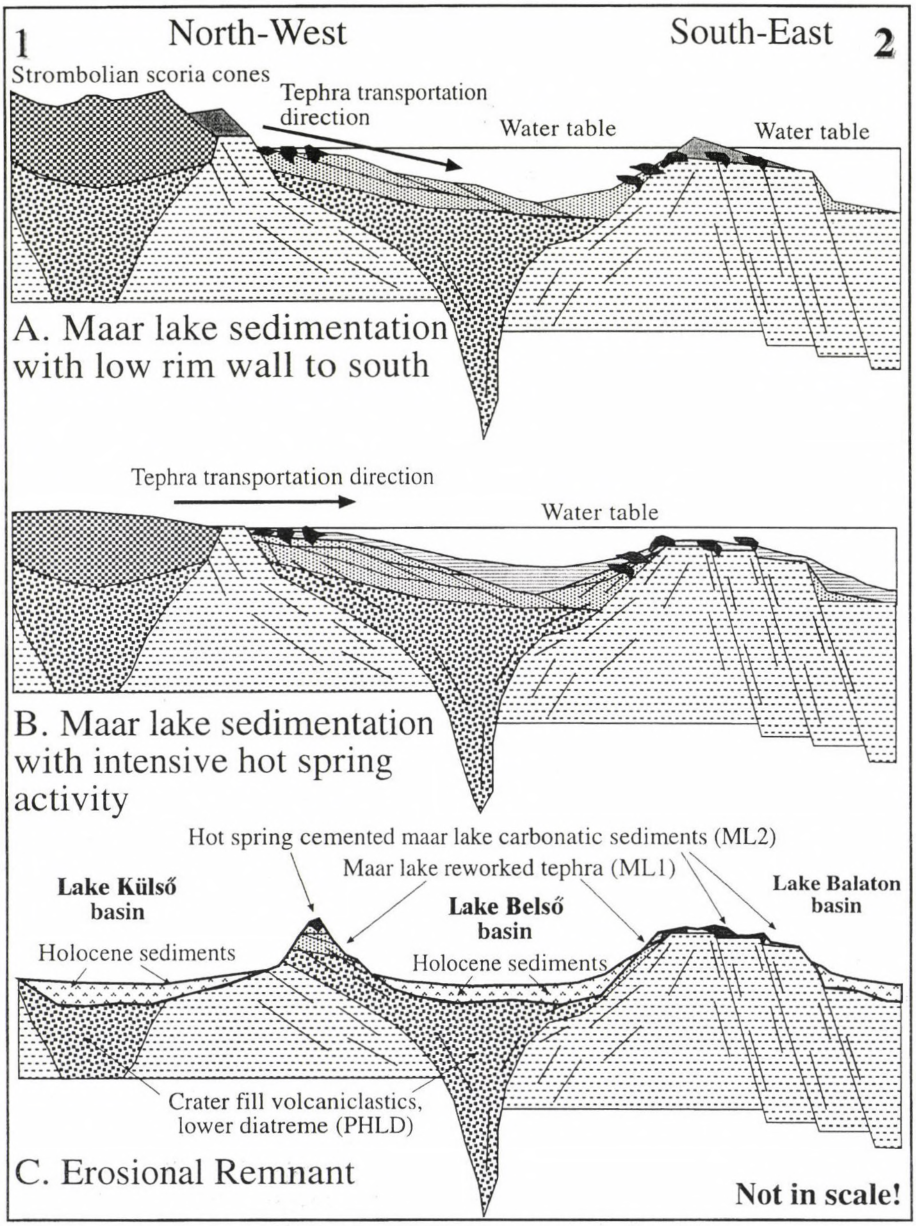

Fig. 11
Facies relationships of volcanic lithofacies associations and the simplified evolution the east maar of Tihany. Maar lake depositional processes with a maar lake basin open to the south, according to the east maar facies relationships

eruptive centers in the TV region. Simultaneously with the last phreatomagmatic events of the active vents, igneous gases became the major moving force behind the eruptions, and a great number of small, Strombolian scoria cones may have been created in the northern part of the peninsula, basically in the area of the recent depression of Lake Külső. It can be suggested that parallel with the Strombolian activity several Hawaiian-type lava fountains were active, especially in the Diós - Gödrös region. No traces indicating lava flows have been detected so far. The TV is an erosional remnant of a maar/diatreme volcanic complex (Fig. 11). The eruption mechanism of the different centers appears to be similar throughout the eruptive history.

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Clay mineralogy of the Komló Calcareous Marl Formation, Bajocian, Mecsek Mountains, Hungary

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Clay mineral assemblages, determined by X-ray diffraction, in the Komló Calcareous Marl Formation (consisting of alternations of carbonate-rich and carbonate-poor layers) of the Mecsek Mountains (Tisza Terrane) record alterations during burial diagenesis and climatic conditions of the continental source area during sedimentation. The assemblages result from erosion of emergent source areas exposed during rifting phases.

In the Bajocian hemipelagic/pelagic sediments exposed north of Püspökszentlászló, the clay fraction is dominated by illite and illite/smectite mixed-layer phases. Kaolinite is rarely found. Mixed-layer illite/smectite is characterized by 40–70 percent illite content and mostly random or 1/4 type-ordered interstratification (according to the nomenclature of Srodon 1980) reflect preferential replacement of smectite by illite during burial diagenesis. This group of clay minerals documents erosion of smectite-rich soils developing under warm and seasonally humid climate and indicates 100–130 °C maximum heating temperature during burial. Discrete illite is abundant and appears not to be altered by burial diagenesis. The sparse occurrence of kaolinite, chlorite and abundant mixed-layer phases besides illite suggest a relatively distant source area during deposition. The abundance of the clay mineral types does not correlate with the lithologies, suggesting that processes forming the alternation of carbonate-rich/carbonate-poor semicouplets did not directly affect the formation of clay minerals.

This conclusion appears to be in agreement with the formerly proposed model, in which the limestone/marl alternation was generated by productivity or composite productivity/dilutional cycles caused by different rates of freshwater runoff, i.e. alternating anti-estuarine and estuarine circulation in an intrashelf basin.

Key words: Bajocian, clay mineralogy, diagenesis, limestone/marl alternation, Mecsek Mountains, paleoenvironment, X-ray diffraction

Introduction

Jurassic formations crop out in the eastern part of the Mecsek Mountains in a relatively extended area. These formations form a mostly marine sequence. This paper deals with two profiles located near the village Püspökszentlászló (Fig. 1a, b and c).

In the Jurassic the area was part of the sedimentary basin of Tethys and participated in its evolution. Differentiation of the initial carbonate shelf into extensional half-graben structures, submarine highs and plateaus started during

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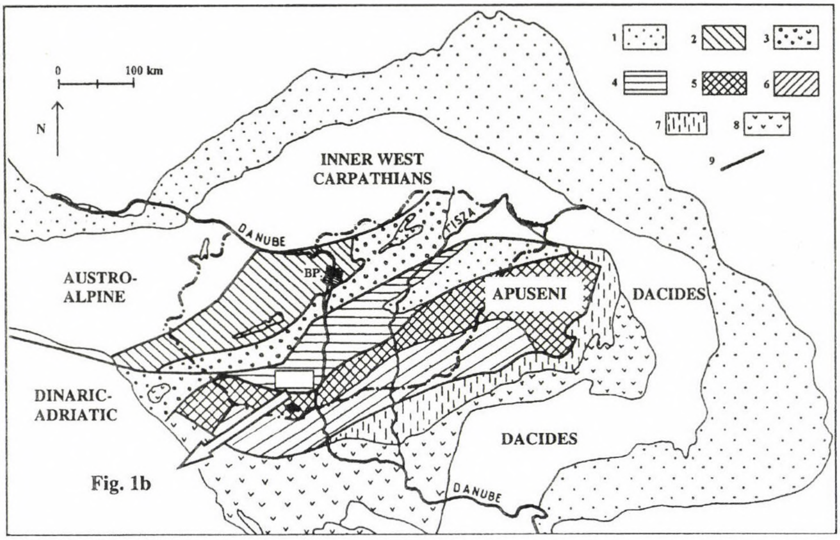

Fig. 1a
The location of Mecsek Mountains in the Mesozoic tectonofacial units of Hungary, modified after Török (1997a). 1. Foredeep and flysch units; 2. Transdanubian and Drauzug units; 3. Bükk and Inner Dinaric unit; 4. Mecsek unit; 5. Villány-Bihor unit; 6. Papuk-Lower Codru unit; 7. N. Backa-Upper Codru, Persani unit; 8. Oceanic nappes (Vardar, Meliata, Mures and Olt); 9. Boundaries of tectonofacial units

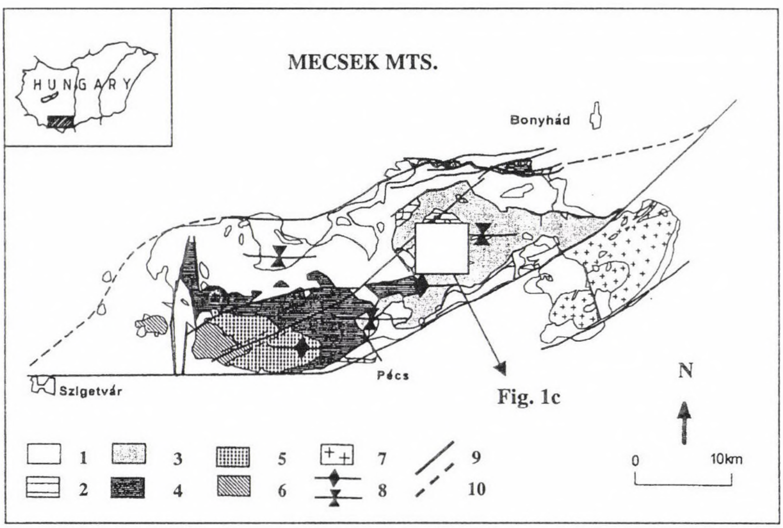

Fig. 1b
Geological map of the Mecsek Mountains, simplified after Török (1997b). 1. Neogene; 2. Cretaceous; 3. Jurassic; 4. Triassic; 5. Upper Permian (sandstone); 6. Upper Permian (siltstone); 7. Carboniferous; 8. anticline, syncline; 9. fault; 10. fault supposed

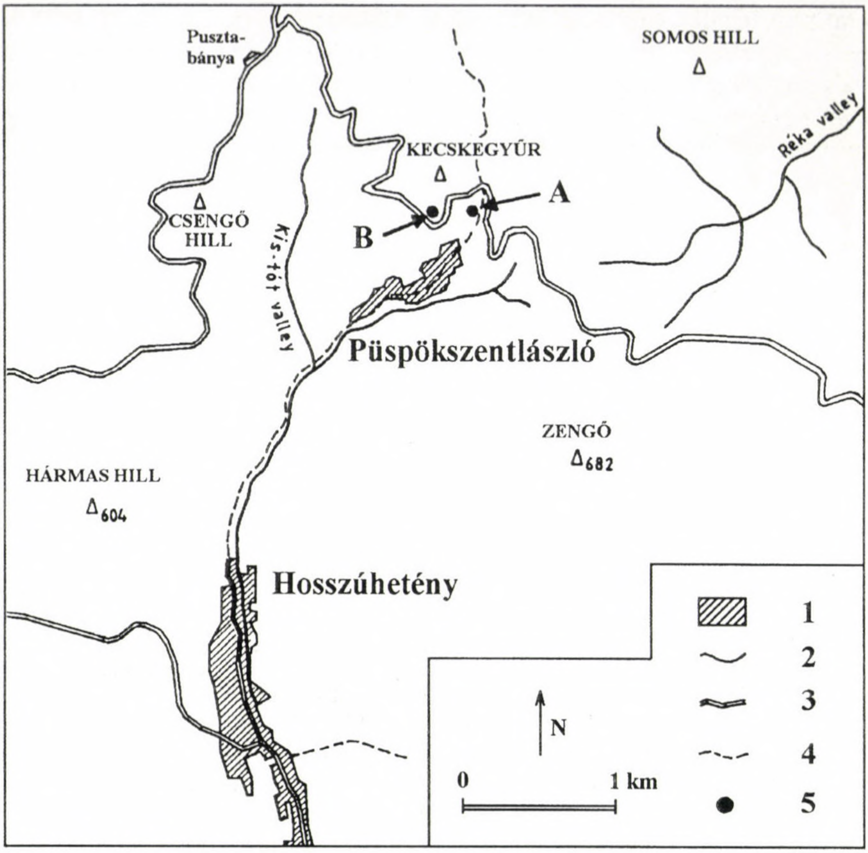

Fig. 1c
Location map of the examined sections. 1. settlement; 2. creek; 3. road; 4. path; 5. location of the examined sections A: section Püspökszentlászló II, B: section Kecskegyűr road cut

the Upper Triassic (Nagy 1969). This process was more enhanced in the Jurassic in connection with rifting of the Penninic Ocean. At the end of the Ladinian, the sedimentation on the outer zones of the European shelf became siliciclastic due to more intensive continental erosion related to eustatic sea level lowstand and/or climatic changes. The Early Liassic in the Mecsek Mountains is characterized by coal and arkose-bearing continental and shallow marine siliciclastic sequence (Gresten facies). In the upper part of the Sinemurian (in accordance with the composite effect of global sea level rise and more intensive subsidence of Mecsek zone) this facies was converted into a deeper water hemipelagic/pelagic basin one with mixed siliciclastic-carbonate lithologies. This sedimentation prevailed until the Upper Bajocian. The intensely bioturbated hemipelagic/pelagic marl facies ("spotted marl") is characteristic for the European margin of the Tethys (Allgäu facies) (Haas 1994).

Stratigraphically the two studied profiles represent an Early to Middle Bajocian succession, i.e. the upper part of the spotted marl sequence (Komló Calcareous Marl Formation). The ammonite *Kumatostephanus* sp. indicates the Sauzei Zone (Galácz 1997, pers. comm.) found in layer P-1a of the section Püspökszentlászló II., i.e. at the base of this sequence.

Raucsik (1997) proposed a model to explain stable isotopic composition of the samples collected on the same locations. The heavy carbon and oxygen-enriched character of the carbonate-rich semicouplets suggests that they were deposited during relatively "drier" and/or "cooler" periods with enhanced bioproductivity. However, this process may have been superimposed by variations in continental runoff during deposition of the carbonate-poor semicouplets, resulting in composite productivity/dilutional cycles. The original isotope record was overprinted to some degree by diagenesis but burial diagenesis seems not to have been significant.

This paper considers mineralogical composition of the clay fraction of the insoluble residue and explains the evolution of the clay mineral assemblages of hemipelagic syn-rift sediments.

Lithology, geological setting

Two sections north of Püspökszentlászló were studied in detail (Figs 2, 3). These sections consist of an alternation of carbonate-rich and carbonate-poor layers (couplets) with centimeter–decimeter scale bed thickness. The thickness of carbonate-rich layers alternates between 12 and 78 cm. The carbonate-poor layers have a minimum of 9 and a maximum of 106 cm bed thickness. Due to upward increase of average carbonate content, in the deeper part of succession the dominant lithologies are marl and calcareous marl alternating with clay marl and marl within a given couplet. In the Kecskegyűr section, however, limestones and argillaceous limestones alternating with calcareous marl become the prevailing rock types. In many cases within a given couplet the "carbonate-rich" and the "carbonate-poor" sediments proved to be petrographically the same carbonate rock type (see Figs 2 and 3). It must be emphasized, however, that the couplet-like appearance of the beds is obvious in the outcrops. All of the rock samples contain 1–5 percent quartz and muscovite silt grains. The beds are gray and greenish gray (with some yellowish shade on the weathered surface) with abundant darker gray spots. The carbonate-poor clayey beds are slightly darker than the limestones. Sharp and continuous bedding contact can also be observed. The carbonate-poor beds with the highest clay contents are thin bedded, while all of the carbonate-rich semicouplets are massive. All of beds are macroburrowed. Microburrowing and lamination are absent. Rock Eval pyrolysis did not detect significant organic matter (TOC values between 0.04 and 0.17%, Hetényi 1997, pers. comm.). These facts indicate oxidizing conditions during the deposition of the Komló Calcareous Marl Formation. Locally, sharp bedding contacts can be observed. This bedding phenomenon

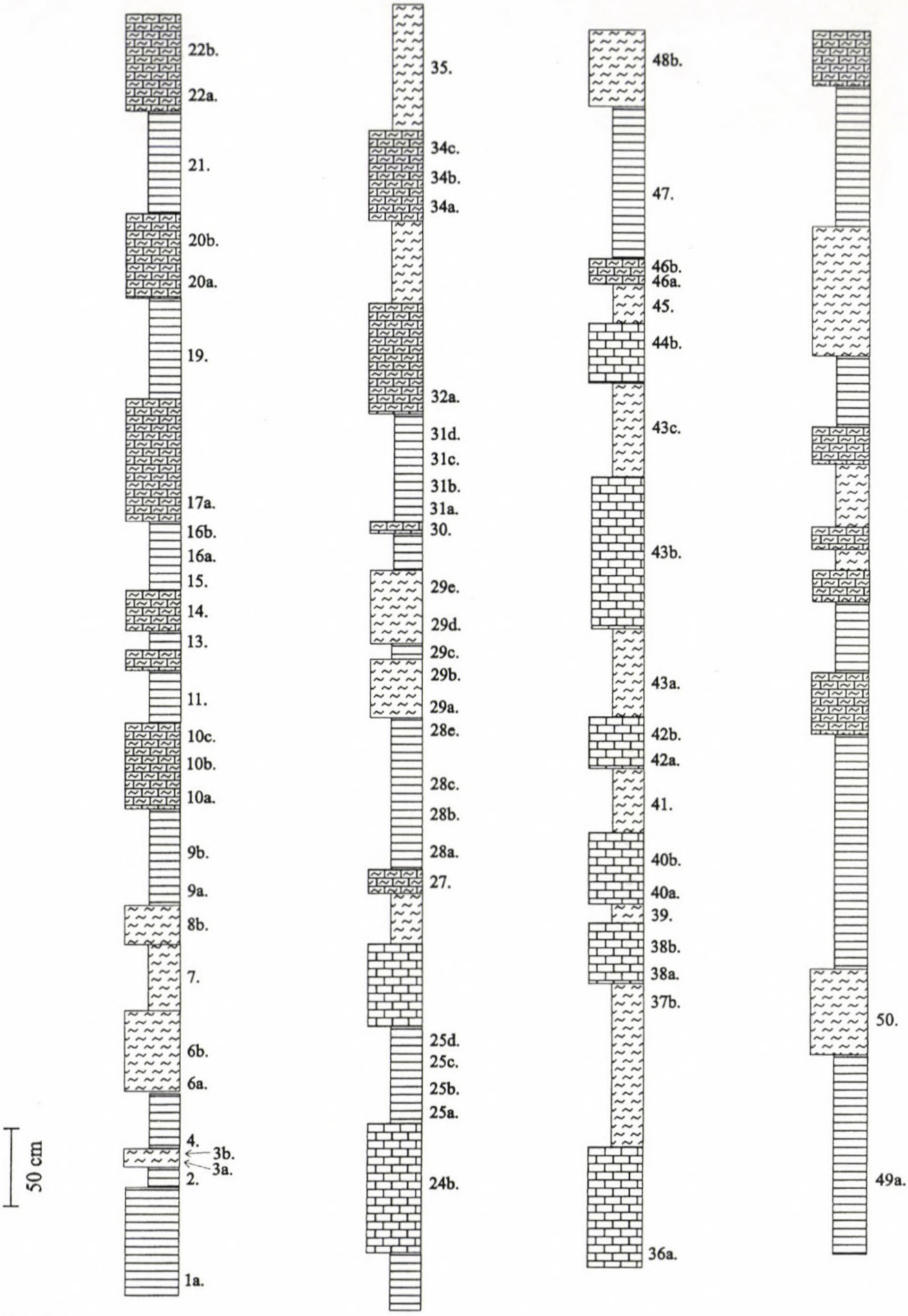

Fig. 2
Lithological column of the section Püspökszentlászló II. For legend see Fig. 3

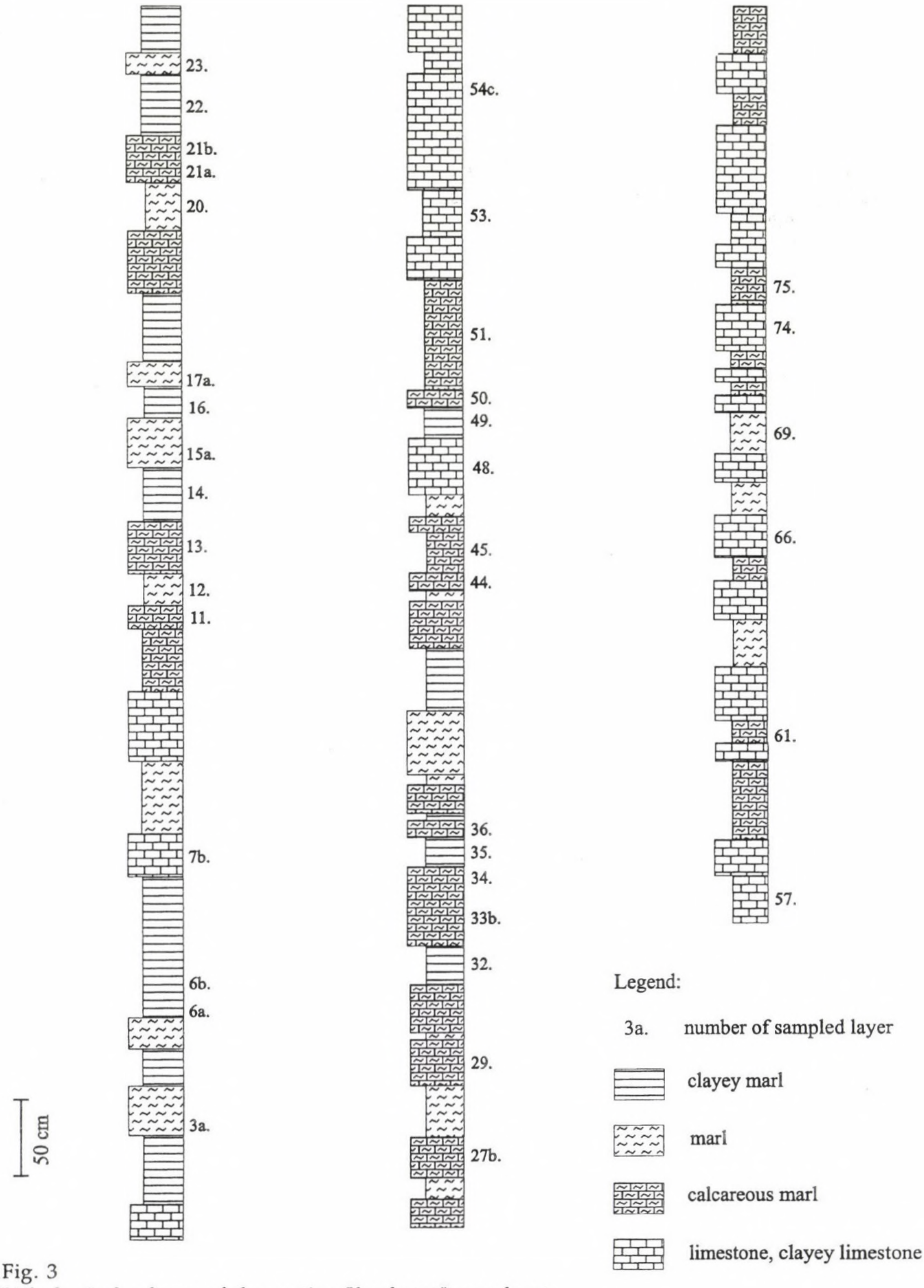

Fig. 3
Lithological column of the section Kecskegyűr, road cut

may indicate punctuated sedimentation (Tucker and Wright 1990). Sedimentary structures, however, indicative of redeposition and erosion by gravity mass flows (including existence of fine-grained turbidites) or contourites (e.g. erosion surfaces, cross lamination, normal or inverse graded bedding, fine lamination, lenticular bedding, fine, obscure silt lenses, tool marks on underlying bedding planes, complete or incomplete Bouma-sequence) were not observed (Stow and Lovell 1979; Tucker and Wright 1990; Piper and Stow 1991). Hardgrounds – indicating submarine dissolution and sediment starvation – were not detected on the sharp bedding contacts.

In thin section the limestones and marls are classified as bioclastic packstones and wackestones. The most abundant biogenic components are radiolarians and siliceous sponges (commonly recrystallized), and filaments. Echinoid fragments represent 1–5 percent of the total bioclast assemblage. One to five percent of unrounded terrigenous quartz silt and very fine sand grains (with 0.03–0.09 mm average maximum measurable diameter in thin section) are present as well. Intensive bioturbation is conspicuous in all samples. Pelagic foraminifera are present in some thin sections. These species (*Lenticulina* sp., *Spirillina* sp., *Garantella* sp.) cannot be used for precise biostratigraphic dating, but they suggest normal marine salinity (Resch 1997, pers. comm.).

A few marl samples show fitted fabric structures. This fact and the partial dissolution of some of biogenic constituents indicate carbonate dissolution and reprecipitation during burial diagenesis. In outcrops wavy bedding surfaces are widespread; nevertheless carbonate concretions did not form, i.e. carbonate redistribution in the sense of Hallam (1964) probably was not significant enough to cause a limestone-marl alternation.

In essence, the studied profiles represent basinal facies, dominated by hemipelagic processes. Sedimentation was presumably continuous. Consequently, this succession is a good candidate for detailed analysis to examine the origin of rhythmic bedding and its possible connection with the distribution of clay minerals.

Methods

The X-ray measurements were made at the sedimentological laboratory of the Department of Geology and Paleontology of University of Innsbruck. The clay fraction under 2 μm was separated by sedimentation after dissolution of carbonates by 3% acetic acid and deflocculation with de-ionized water. X-ray diffraction analysis was performed on oriented pastes, with a Siemens D-500 instrument using CuK_α radiation with Ni filter. Two X-ray diagrams were made: one under natural conditions and one after saturation with ethylene glycol.

Clay minerals were identified primarily by the position of their basal reflections. Specific values were used for characterization of clay mineral types. For estimation of values of illite/smectite ratio in mixed-layer structures and for estimating the ordering of interstratification the standard methods of

Watanabe (1981), Srodon (1980, 1984), and data of Reynolds and Hower (1970) were used. The relative abundance of clay minerals was determined by the peak area ratio of the 001/001 reflection of mixed-layer illite/smectite and the 001 reflection of illite and kaolinite after glycolization. Peak areas of mixed-layer illite/smectites were corrected by factors published by Rischák and Viczián (1974). Mixed-layer phases close to pure illite were corrected by multiplying by a factor of 2, while those close to smectite were multiplied by a factor of 0.5. Peak area of discrete illite was corrected in a similar manner by a factor of 2.

Results

Table 1 shows the semi-quantitative clay mineralogical composition of the studied sections. Illite and mixed-layer illite/smectites of various degrees of expandability are always the dominant clay minerals. Illite is generally of poor crystallinity (values of the Kübler index vary between 0.6–1.0° 2Θ). In most of the samples Kübler indexes cannot be measured exactly due to a shoulder on the high-angle side of the illite 001 peak, probably caused by the expanded mixed-layer phases. Intermediate stages of the transition from smectite to illite clearly appear on the XRD diagrams of most samples where illite/smectite mixed-layers with various proportions of both component layers can be observed.

The estimated semi-quantitative composition of clay fraction of the measured samples is shown in Figs 4 and 5. The relative abundance of the mixed-layer phases in function of the $CaCO_3$-content of the samples of section Püspökszentlászló II. are plotted in Fig. 6a. The proportion of mixed-layer clay minerals in function of the $CaCO_3$-content of the samples of the section of the Kecskegyűr road cut are shown in Fig. 6b. It is obvious that the proportion of mixed-layer illite/smectite in the insoluble residue is independent of the $CaCO_3$-content of the measured samples (see the "*r*" values).

Table 2 shows the results of measurements on mixed-layer structures. In Table 2 the $\Delta 2\Theta_1$ and the $\Delta 2\Theta_2$ values are expressed by Watanabe's terminology (1981). However, most of the samples cannot be precisely fixed on Watanabe's diagram, so for the determination of their illite content and ordering another standard method published by Srodon (1980) was applied. If a reflection occurred between 5.3 and 8.7 2Θ in the diffraction pattern of an illite/smectite affected by ethylene glycol solvent, the examined illite/smectite was considered as a clay mineral with interstratification ordered to some degree (Srodon 1980). For randomly interstratified mixed-layer structures, the most problematical method ("a/b ratio method") using intensity data published by Reynolds and Hower (1970) was used. The 001 peak of illite has a shoulder on the high-angle side in most cases, which also indicates the appearance of illite/smectite, beside a shifting of the 001/001 peak of mixed-layer phases after glycolization.

In Figs 7a, b, 8 the illite percentage intervals in the mixed-layer minerals are shown estimated by Srodon's method. There seems to be no correlation between

Table 1

Semi-quantitative composition of the measured samples. a) Section Kecskegyűr, road cut. b) Section Püspökszentlászló II.

a)

Number of samples	Mixed-layer illite/smectite (%)	Discrete illite (%)	Kaolinite (%)
K-75	**66**	**34**	–
K-74	57	3	–
K-69*	**69**	**31**	–
K-66	79	21	–
K-61	**90**	**10**	–
K-57	**86**	**14**	–
K-54C	100	–	–
K-53	**80**	**20**	–
K-51	**58**	**42**	–
K-50*	86	14	–
K-49	**85**	**15**	–
K-48	100	–	–
K-45	**79**	**21**	–
K-44	79	21	–
K-36*	85	15	–
K-35*	**100**	–	–
K-34	100	–	–
K-33B	72	28	–
K-32	**75**	**25**	–
K-29	61	39	–
K-27B*	82	18	–
K-23*	82	18	–
K-22*	**100**	–	–
K-21B	68	32	–
K-21A*	100	–	–
K-20	**79**	**21**	–
K-17A	85	15	–
K-16	**78**	**22**	–
K-15A*	100	–	–
K-14*	**75**	**25**	–
K-13	82	18	–
K-12	**71**	**29**	–
K-11	90	10	–
K-7B	69	31	–
K-6B	**83**	**17**	–
K-6A	**81**	**19**	–
K-3A	85	15	–

Bold numbers: samples from carbonate-poor semicouplets. Normal numbers: samples from carbonate-rich semicouplets. *: samples with 6.2 2Θ peak

b)

Number of samples	Mixed-layer illite/smectite (%)	Discrete illite (%)	Kaolinite (%)
P-50	59	41	–
P-49A	**72**	**28**	–
P-48B	82	14	4
P-47	**83**	**17**	–
P-46B	82	18	–
P-46A	90	10	–
P-45	**93**	**7**	–
P-44B	78	14	8
P-43C	**94**	**6**	–
P-43B	88	12	–
P-43A	95	5	–
P-42B	92	8	–
P-42A	72	28	–
P-41	**76**	**24**	–
P-40B	61	29	10
P-40A	71	29	–
P-39	**86**	**14**	–
P-38B	26	54	20
P-38A	91	9	–
P-37B	**80**	**20**	–
P-36A	56	30	14
P-35	**73**	**27**	–
P-34C	51	49	–
P-34B	66	24	10
P-34A	53	47	–
P-32A	46	54	–
P-31D	**75**	**25**	–
P-31C	**83**	**17**	–
P-31B	**48**	**52**	–
P-31A	**71**	**29**	–
P-30	100	–	–
P-29E	60	26	14
P-29D	95	5	–
P-29C	**68**	**32**	–
P-29B	91	9	–
P-29A	29	31	40
P-28E	**80**	**20**	–
P-28C	**69**	**31**	–
P-28B	**61**	**39**	–
P-28A	**52**	**48**	–
P-27	74	26	–
P-25D	**81**	**19**	–
P-25C	**75**	**25**	–
P-25B	**53**	**47**	–
P-25A	**80**	**20**	–
P-24B	51	35	14
P-22B	73	27	–
P-22A	70	30	–
P-21	**61**	**39**	–
P-20B	73	27	–
P-20A	75	25	–
P-19	**76**	**24**	–
P-17A	56	35	9
P-16B	**66**	**34**	–
P-16A	**70**	**30**	–
P-15	**32**	**68**	–
P-14	63	37	–
P-13	**63**	**37**	–
P-11	**78**	**22**	–
P-10C	66	34	–
P-10B	77	23	–
P-10A	70	30	–
P-9B	**80**	**20**	–
P-9A	**71**	**29**	–
P-8B	70	30	–
P-7	**75**	**25**	–
P-6B	54	32	14
P-6A	76	24	–
P-4	**77**	**23**	–
P-3B	73	27	–
P-3A	73	27	–
P-2	**83**	**17**	–
P-1A	83	17	–

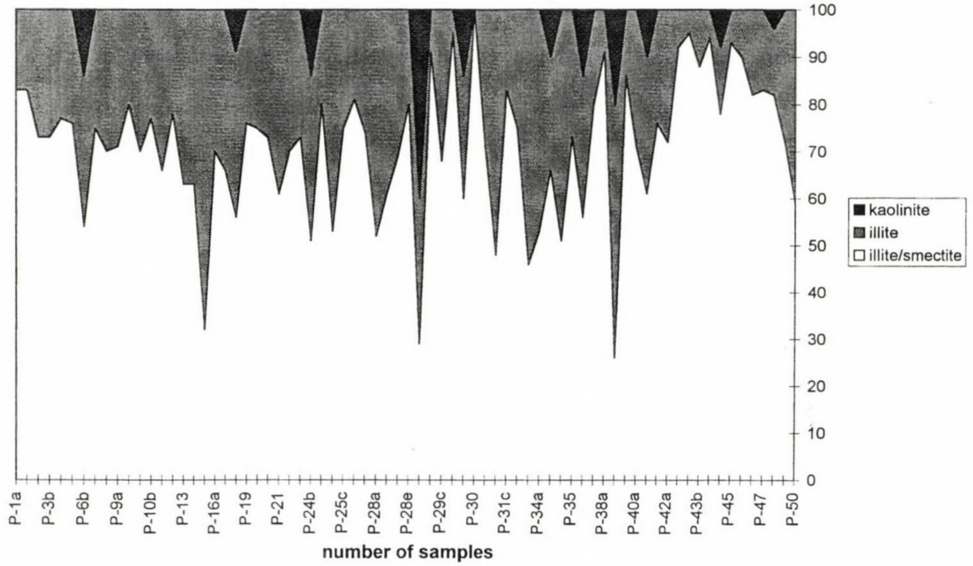

Fig. 4
Semi-quantitative composition of the clay fraction. Section Püspökszentlászló II

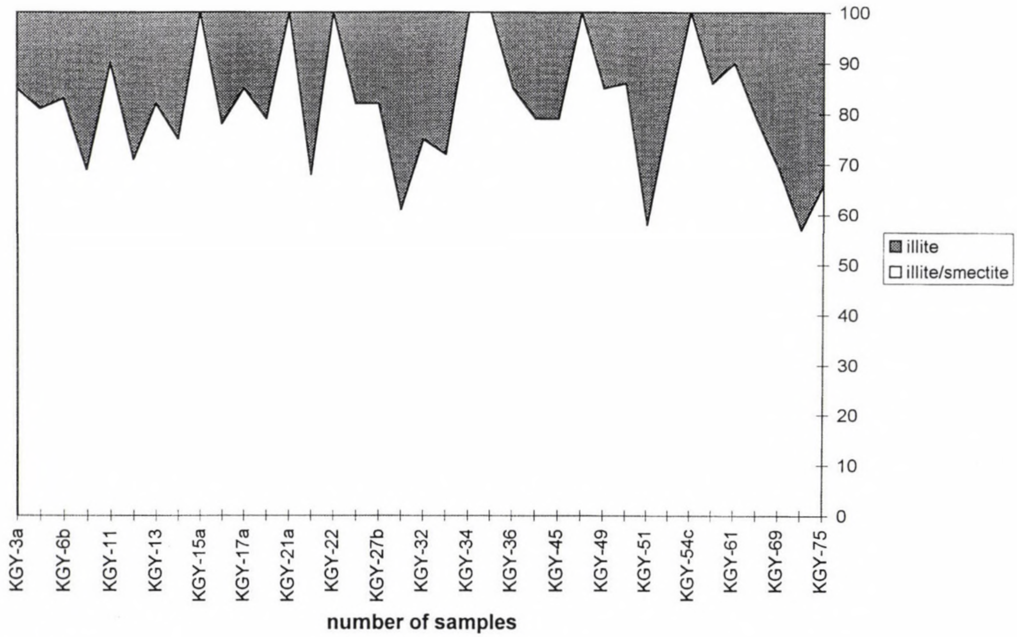

Fig. 5
Semi-quantitative composition of the clay fraction. Section Kecskegyűr, road cut

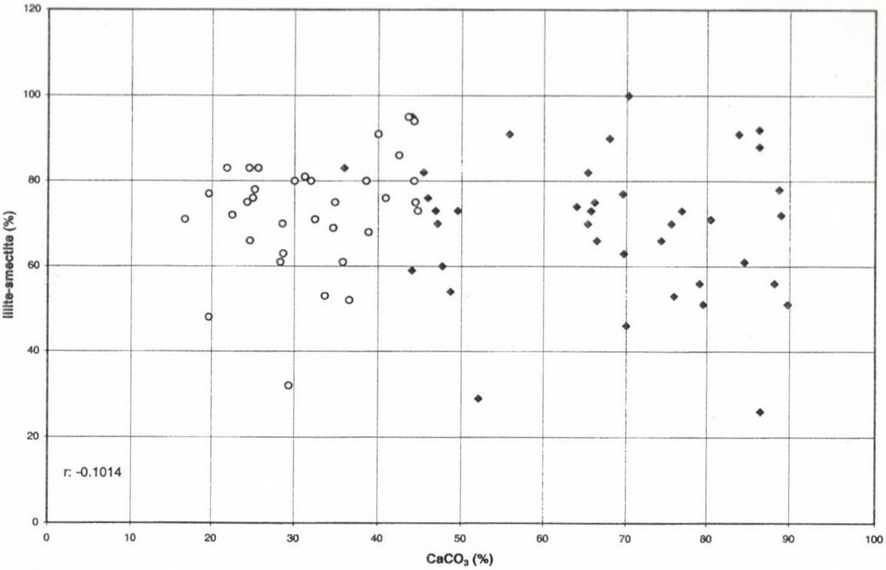

Fig. 6a
Proportion of illite/smectite mixed-layer minerals in function of the $CaCO_3$-content. Section Püspökszentlászló II. Legend: circles: carbonate-poor semicouplets, diamonds: carbonate-rich semicouplets

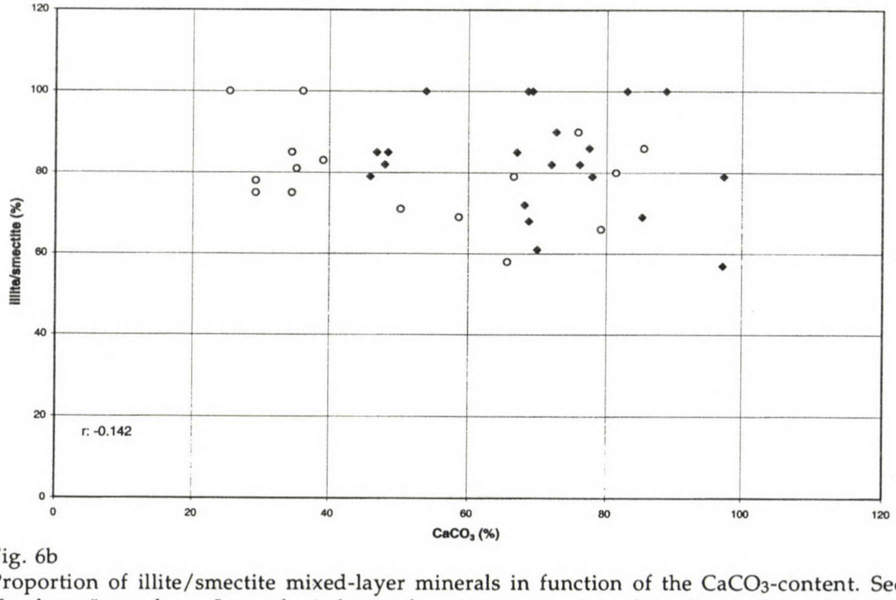

Fig. 6b
Proportion of illite/smectite mixed-layer minerals in function of the $CaCO_3$-content. Section Kecskegyűr road cut. Legend: circles: carbonate-poor semicouplets, diamonds: carbonate-rich semicouplets

Table 2a
Results of measurements on mixed-layer illite/smectites. Section Kecskegyűr, road cut

Number of samples	$\Delta 2\Theta_1$	$\Delta 2\Theta_2$	Position of 002/003 reflexion (2Θ°)	Illite proportion in the mixed-layer structure (%) Środoń (1984)	"a/b ratio" Reynolds and Hower (1970)	Illite proportion (%) in randomly interstratification	Ordering
K-75	3.43	7.59	16.35	40-60	-	-	1/4, 1/2 or IS
K-74	3.42	7.58	16.32	40-60	-	-	1/4, 1/2 or IS
K-69	4.07	7.61	16.38	45-67	0.64	40	random
K-66	3.33	7.54	16.50	50-70	-	-	1/4, 1/2 or IS
K-61	3.37	7.57	16.31	40-50	-	-	1/4 or 1/2
K-57	3.28	7.61	16.32	40-50	-	-	1/4 or 1/2
K-54C	3.58	7.62	16.30-	40-50	-	-	1/4 or 1/2
K-53	3.64	7.66	16.45	50-70	0.81	49	random
K-51	3.82	7.62	16.42	50-70	0.63	39	random
K-50	4.09	7.60	16.40	40-70	0.63	39	random
K-49	3.97	7.55	16.32	45-65	0.72	44	random
K-45	4.15	7.60	16.47	50-70	0.69	42	random
K-44	4.12	7.61	16.46	50-70	0.70	43	random
K-36	3.22	7.70	16.44	50-72	-	-	1/4, 1/2 or IS
K-35	4.11	7.60	16.43	50-70	0.71	44	random
K-34	3.41	7.62	16.49	45-70	-	-	1/4, 1/2 or IS
K-33B	4.09	7.65	16.45	50-70	0.64	40	random
K-32	3.73	7.43	16.14	22-58	0.50	32	random
K-29	3.64	7.22	16.15	22-58	0.60	37	random
K-27B	3.42	7.62	16.43	45-65	-	-	1/4, 1/2 or IS
K-23	3.25	7.70	16.50	50-70	-	-	1/4, 1/2 or IS
K-22	3.70	7.43	16.15	25-60	0.72	44	random
K-21B	3.72	7.43	16.17	25-60	0.75	46	random
K-21A	3.71	7.45	16.16	25-60	0.67	41	random
K-20	3.14	7.70	16.42	50-65	-	-	1/2 or IS
K-17A	3.80	7.59	16.32	40-60	-	-	1/4 or 1/2
K-16	3.62	7.54	16.33	40-60	-	-	1/4 or 1/2
K-15A	3.63	7.24	16.14	25-60	0.77	47	random
K-14	3.67	7.22	16.15	25-60	0.64	40	random
K-12	3.33	7.64	16.49	50-70	-	-	1/4, 1/2 or IS
K-11	3.45	7.61	16.51	50-70	-	-	1/4, 1/2 or IS
K-7B	3.52	7.42	16.21	30-63	0.59	37	random
K-6B	3.41	7.64	16.47	50-70	-	-	1/4, 1/2 or IS
K-6A	3.78	7.57	16.35	45-65	0.68	42	random
K-3A	3.42	7.63	16.50	55-70	-	-	1/4, 1/2 or IS

the lithology ($CaCO_3$-content) and the illite proportions. The majority of the examined mixed-layer illite/smectite minerals is characterized by 40–70 percent of illite layers and by mostly random interstratification or 1/4 partial ordering according to the terminology of Srodon (1980). There are samples with 6.2 2Θ peak after glycolization. These peaks – samples in Table 1 are indicated by an asterisk – seem to indicate little chlorite.

Kaolinite appears seldom (10% of all samples), mostly in low amounts (14.3% average abundance; maximum value = 40% in sample P-29A) and only in the carbonate-rich semicouplets (see Table 1). However, there is no correlation between the $CaCO_3$-content of the samples and the abundance of this clay species (r= –0.2071) (Fig. 9).

Table 2b

Results of measurements on mixed-layer illite/smectites. Section Püspökszentlászló II

Number of samples	$\Delta 2\Theta_1$	$\Delta 2\Theta_2$	Position of 002/003 reflexion (2Θ°)	Illite proportion in the mixed-layer structure (%) Środoń (1984)	"a/b ratio" Reynolds and Hower (1970)	Illite proportion (%) in randomly interstratification	Ordering
P-50	3.23	7.52	16.20	30-60	0.78	48	random
P-49A	3.52	7.54	16.40	45-65	0.71	44	random
P-48B	2.67	7.56	16.41	45-65	-	-	IS
P-47	3.65	7.53	16.41	45-65	0.56	35	random
P-46B	2.90	7.52	16.35	40-60	-	-	1/2 or IS
P-46A	2.73	7.70	16.48	50-70	-	-	IS
P-45	2.69	7.48	16.44	45-65	-	-	IS
P-44B	2.82	7.61	16.37	40-60	-	-	IS
P-43C	3.10	7.58	16.44	45-65	-	-	1/2 or IS
P-43B	2.73	7.72	16.46	45-65	-	-	IS
P-43A	2.98	7.60	16.43	45-65	-	-	1/2
P-42B	2.85	7.67	16.42	45-65	-	-	1/2 or IS
P-42A	3.46	7.62	16.47	48-68	-	-	1/2 or IS
P-41	3.55	7.63	16.40	50-65	0.54	34	random
P-40B	3.49	7.60	16.40	55-60	-	-	1/4 or 1/2or IS
P-40A	3.31	7.54	16.36	40-60	-	-	1/2 or IS
P-39	2.86	7.55	16.38	40-60	-	-	1/2 or IS
P-38B	3.55	7.59	16.43	50-65	0.88	53	random
P-38A	3.01	7.47	16.40	50-65	0.88	53	random
P-37B	3.72	7.43	16.20	30-60	0.82	50	random
P-37A	3.69	7.54	16.20	30-60	0.78	47	random
P-36A	3.54	7.63	16.48	48-68	-	-	1/2 or IS
P-35	3.63	7.45	16.30	40-65	0.62	38	random
P-34C	3.04	7.55	16.30	40-65	0.91	54	random
P-34B	3.59	7.67	16.34	42-65	0.81	49	random
P-34A	3.64	7.46	16.46	50-70	0.80	47	random
P-32A	2.60	7.59	16.19	40-50	-	-	IS
P-31D	3.44	7.64	16.45	42-65	-	-	1/2 or IS
P-31C	3.30	7.61	16.42	40-65	-	-	1/2 or IS
P-31B	3.40	7.58	16.24	35-55	-	-	1/2 or IS
P-31A	3.31	7.59	16.42	40-65	-	-	1/2 or IS
P-29E	3.34	7.61	16.45	42-65	-	-	1/2 or IS
P-29D	3.10	7.42	16.35	40-60	-	-	1/2
P-29C	3.48	7.59	16.38	40-60	-	-	1/2 or IS
P-29B	2,85	7,55	16.39	40-60	-	-	1/2 or 1/4
P-29A	3.56	7.64	16.43	50-70	0.75	46	random
P-28E	3.60	7.35	16.20	30-60	0.57	36	random
P-28C	3.69	7.56	16.45	50-70	0.69	42	random
P-28B	3.39	7.62	16.30	40-60	-	-	1/2 or IS
P-28A	3.48	7.52	16.40	40-62	-	-	1/2 or IS
P-27	3.74	7.57	16.40	48-68	0.80	48	random
P-25D	3.40	7.63	16.42	42-64	-	-	1/2 or IS
P-25C	3.52	7.30	16.10	20-58	0.63	39	random
P-25B	3.57	7.46	16.30	40-64	0.83	50	random
P-25A	3.61	7.59	16.40	48-68	0.58	36	random
P-24B	3.66	7.52	16.35	42-65	0.82	50	random

Table 2b (cont)

Number of samples	$\Delta 2\Theta_1$	$\Delta 2\Theta_2$	Position of 002/003 reflexion (2Θ°)	Illite proportion in the mixed-layer structure (%) Środoń (1984)	"a/b ratio" Reynolds and Hower (1970)	Illite proportion (%) in randomly interstratification	Ordering
P-22B	3.50	7.55	16.42	42-64	-	-	1/2 or 1/4
P-22A	3.67	7.43	16.30	40-64	0.79	48	random
P-21	3.48	7.58	16.40	48-68	0.80	48	random
P-20B	3.68	7.63	16.45	50-70	-	-	random
P-20A	3.63	7.65	16.50	53-70	0.72	44	random
P-19	3.57	7.61	16.42	48-68	0.88	53	random
P-17A	3.65	7.46	16.33	40-65	0.74	45	random
P-16B	3.64	7.52	16.40	48-68	0.62	38	random
P-16A	3.46	7.62	16.44	42-65	-	-	1/2 or 1/4
P-15	3.69	7.53	16.36	40-70	-	-	1/4
P-14	3.51	7.48	16.40	48-68	0.74	45	random
P-13	3.55	7.57	16.40	48-68	0.86	52	random
P-11	3.63	7.53	16.35	50-60	-	-	1/4
P-9B	3.61	7.45	16.30	40-64	0.52	33	random
P-10C	3.56	7.48	16.30	40-64	0.77	47	random
P-10B	3.53	7.64	16.40	48-68	0.69	42	random
P-10A	3.43	7.64	16.43	42-65	-	-	1/2 or IS
P-9A	3.66	7.49	16.40	40-62	-	-	1/4
P-8B	3.74	7.58	16.47	45-67	-	-	1/4
P-7	3.57	7.58	16.40	48-68	0.71	44	random
P-6B	3.66	7.44	16.32	50-60	-	-	1/4
P-6A	3.59	7.56	16.42	42-64	-	-	1/4
P-4	3.68	8.80	16.30	50-60	-	-	1/4
P-3B	3.35	7.21	16.40	40-62	-	-	1/2
P-3A	3.62	7.55	16.40	48-68	0.80	48	random
P-2	3.62	7.14	16.00	5-52	0.54	34	random
P-1A	3.61	7.57	16.40	40-62	-	-	1/4

Discussion

In most regions of the world ocean, clay detrital assemblages reflect the combined influences of land petrography and continental climate (Biscaye 1965). The common clay minerals which, as environmental indicators, require a brief discussion here are kaolinite, mixed-layer illite/smectite and illite.

In the modern oceans kaolinite abundance increases toward the Equator in all oceanic basins and therefore expresses a strong climatic dependence controlled by the intensity of continental hydrolysis. In modern deep-sea sediments kaolinite tends to increase in relative abundance in regions of tropical continental weathering (Biscaye 1965). The strong increase in kaolinite, goethite, gibbsite and fibrous clays could reflect more intensive weathering and soil production in the source area (Chamley 1989). For Jurassic formations in the Mecsek Mountains, Nagy (1969), Viczián (1987) reported high amounts of kaolinite in the Mecsek Coal Formation.

Mixed-layer illite/smectites have long been believed to be formed in diagenetic environments through the alteration of smectite (Hower 1981). Evidence does exist, however, that mixed-layer illite/smectites may form in a weathering environment through the leaching and degradation of a precursor

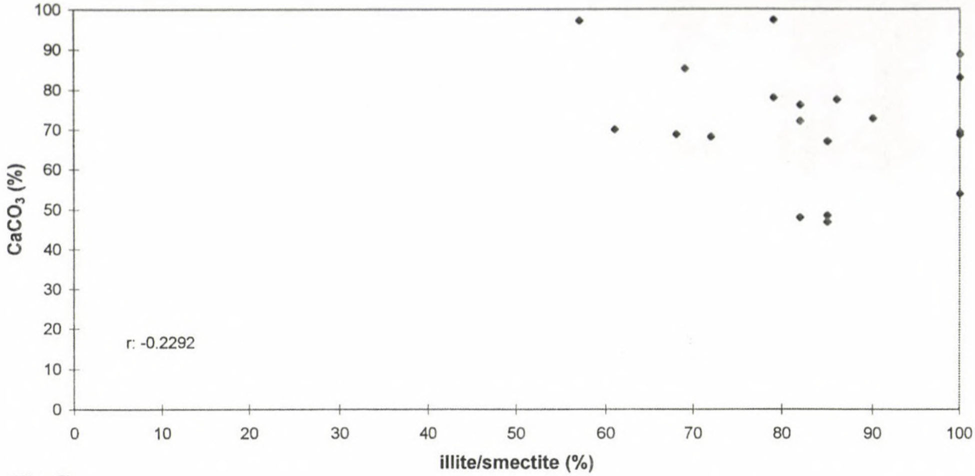

Fig. 7a
Estimated illite proportion of mixed-layer illite/smectite clay minerals. Section Püspökszentlászló II, carbonate-rich semicouplets

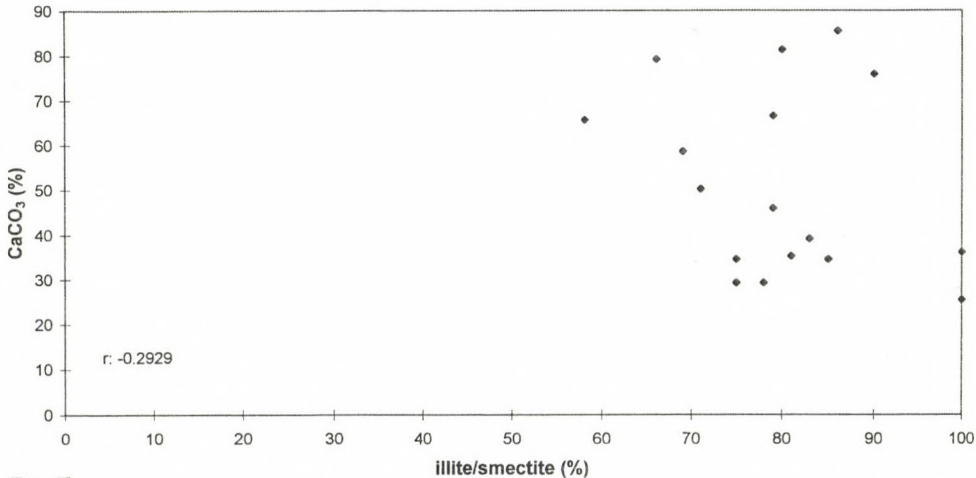

Fig. 7b
Estimated illite proportion of mixed-layer illite/smectite clay minerals. Section Püspökszentlászló II, carbonate-poor semicouplets

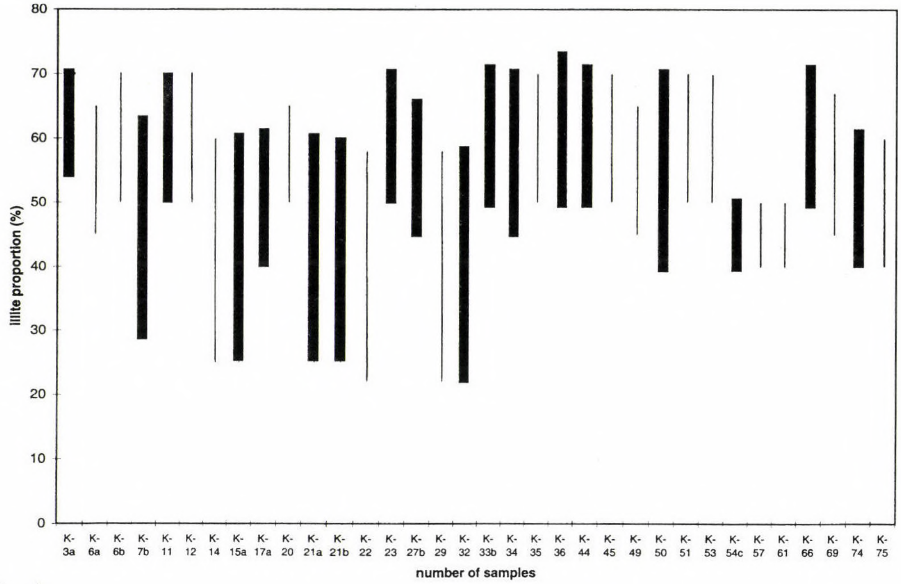

Fig. 8
Estimated illite proportion of mixed-layer illite/smectite clay minerals. Section Kecskegyőr, road cut. Legend: thin lines: carbonate-poor semicouplets, thick lines: carbonate-rich semicouplets

illite (Chamley 1967, 1989). Probably the most favorable climatic conditions are those in which warmth is combined with strong seasonal contrasts in humidity (Paquet 1970). Although smectite and illite/smectite mixed-layer minerals presently form under a variety of climates the most important type appears to be the one in which a pronounced dry season alternates with a less pronounced (or shorter) wet season (Singer 1984). Robinson and Wright (1987) have suggested that some mixed-layer illite/smectite could be produced from smectite during pedogenesis. It must be noted that smectite may be entirely volcanogenic in origin, being derived directly from the weathering of lava or ash and thus having nothing to do with climate. In this case, however, distinctive accessory minerals must occur, such as biotite, sphene, cristobalite, zeolites and, rarely, relict glass shards (Pacey 1984). The distribution of mixed-layer minerals in the present-day oceans have strong geographic controls. This indicates continental sources rather than in-situ diagenetic origin (Biscaye 1965).

The relative abundance of smectite generally shows a distribution that does not parallel the zonal distribution of the main weathering processes. This indicates the accessory control of climate and the dominance of other processes. The increase of Al-Fe-rich smectites and their abundance do not depend on deposition in marine environment but are chiefly attributed to the reworking of continental peneplanation of gently sloping and poorly drained areas (Chamley and Debrabant, 1984).

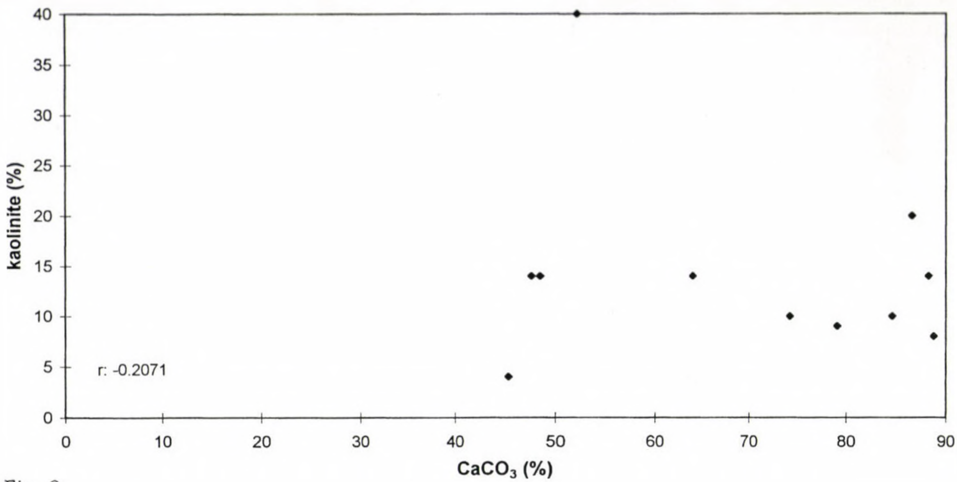

Fig. 9
Kaolinite content of the samples in function of the calcite content

The occurrence of discrete illite in sediments probably has no particular climatic significance (Hallam et al. 1991) but Singer (1984) claims that illite exhibiting high crystallinity signifies formation in either cold or dry conditions with minimum hydrolyzation. Because of low crystallinity of illite in the Komló Calcareous Marl Formation these conditions can hardly be considered as the explanation of their presence in the analyzed samples.

Differential transport plays an important role in determining the distribution of clay minerals in marine deposits. Clay mineral segregation by differential settling represents a common phenomenon on certain continental margins characterized by a simple influx from river to ocean. The distance over which clay sorting acts does not seem to exceed a few hundred kilometers and often appears to represent a few kilometers only. As early diagenesis does not seem (from available data) to account for such a strong differentiation, one may expect that clay sorting was reinforced both by a hydrolyzing climate providing abundant kaolinite and smectite to the sea, and by sudden variations of turbulence between shelf and basin environments. The mechanisms responsible for clay changes recorded on continental margins appear to be dominated by grain size sorting. In situ chemical variations cannot account for such phenomena since they should result in the authigenic formation of illite rather than that of smectite and are not supported by any clear arguments. The influence on sedimentation of direct and rapid subvertical sinking of clay aggregates should not be overestimated, since horizontal transport and resuspension of individual particles and aggregates occur widely in the ocean under the action of surficial, deep, and bottom currents (Chamley 1989). Kaolinite tends to increase in abundance in nearshore facies, probably reflecting its coarse-grained nature and its strong tendency to flocculate compared to most other clays (Parham 1966). Clay sorting usually determines the farther transportation of smectite and fibrous clays relative to most of the other clay species. Thus, the influx may be explainable in terms of a continental margin or delta progradation and retreat, or as a result of shifting depositional patterns.

The sparse occurrence of kaolinite is restricted to few limestone semicouplets which can be observed in the samples of Komló Calcareous Marl Formation. This observation could indicate some resedimentation from the neighboring shelf areas. According to micropaleontological data (Monostori 1997, pers. comm.) there are ostracods in samples of the Komló Calcareous Marl Formation known from other locations of Mecsek Mountains which indicate shallow water environment and certainly were resedimented from the platforms. Nepheloid plumes could cause this type of resedimentation (Tucker and Wright 1990). The presence of morphological barriers to the transportation of the fast-settling kaolinite toward the basin, or the presence of a well-developed river-fed marginal or intracontinental basin, which would prevent the supply to the ocean of minerals pedogenically formed in the upstream zones, however, cannot be excluded.

Another complication concerns diagenesis. Under conditions of increased temperature, due to burial, smectite tends to transform to illite via an intermediate stage of mixed-layer minerals (Hower et al. 1976). The same transformation may also take place at surface temperatures under conditions of repeated wetting and drying (Singer 1988). Viczián (1994) has reviewed the application of the series smectite / illite as a geothermometer. According to his paper the intensity of smectite to illite transition depends on the variables time, temperature, K^+ availability and K^+ concentration in the depositional environment and activation energy. He has presented a trend line calculated from data of Pannonian Basin. Part of this line corresponding to 40–70% smectite in mixed-layer indicates 100–130 °C heating temperature during burial diagenesis. According to fission track data, 100–175 °C temperature of heating during burial was calculated by Dunkl (1992). According to Viczián (1990) illite contents in the mixed-layer phases of the underlying Mecsek Coal and the Vasas Marl Formations are 70–80%. These higher illite contents – and thus higher crystallinity – are due to deeper burial and a higher degree of "ripening" of illite/smectites in accordance with the deeper stratigraphic position of these Lower Jurassic formations.

Paleoclimate simulations using three-dimensional general circulation models have been concentrated on two geologic intervals: the Pleistocene Epoch (Rind 1987; COHMAP Members 1988) and the Cretaceous Period (Rind 1986; Oglesby and Park 1989). Parrish et al. (1982), Hallam (1984), and Kutzbach and Gallimore (1989) declared in agreement with each other that both modeling and empirical research suggest that zonal winds were probably much less important on the Jurassic supercontinent than monsoonal winds. It seems evident that temperature was higher than at present, dry and wet seasons alternated during these monsoon-controlled times. According to Hallam et al. (1991) the increase in smectite abundance of the latest Jurassic rocks from England and France indicates a climate with a more pronounced and extended dry season in contrast with the Cretaceous. Similar to Cretaceous climate simulations Chandler et al. (1992) gave a model for the Early Jurassic. Results from their simulations of the Jurassic climate show that increased ocean heat transport may have been the primary force generating warmer climates. Three major features of the simulated Jurassic climate include the following: (1) Global warming, comparable to the one occurring at present. (2) Decreases in albedo, linked to reductions in sea ice, snow cover, and low clouds, and increases in atmospheric water vapor are the positive climate feedbacks that amplify the global warming. (3) High rainfall rates were associated primarily with monsoons that originated over the warm Tethys Ocean. These systems are found to be associated with localized pressure cells whose positions are controlled by topography and coastal geography. Weissert and Mohr (1996) studied the carbon isotope composition of large amount of limestone representing Oxfordian–Tithonian stages from the Helvetic nappes of Switzerland. They concluded that the climate in the northern Tethyan realm was characterized by a high atmospheric CO_2 level and by monsoonal rainfall pattern. No observation has been made so far that in the Bajocian stage climatic conditions in the Mecsek sedimentary basin were different.

Conclusions

Dominance of illite/smectite mixed-layer phases indicate seasonally alternating monsoon-like climatic conditions during Bajocian in the source area of the Mecsek sedimentary basin. Under these conditions pedogenic smectite and/or disordered mixed-layer illite/smectite were generated and carried into the basin. 40–70% illite proportion in mixed-layer and the moderate ordering is due to diagenesis and indicates 100–130 °C heating temperature during burial. Discrete illites were not influenced by heating to such a degree.

Kaolinite content was poor and found only in limestone samples. The following explanations for this can be envisaged: (1) the climatic (and/or lithologic) conditions were not favorable for generation of abundant kaolinite and/or (2) morphological barriers or a well-developed river-fed marginal basin existed which would prevent the transportation into the ocean of many of the minerals pedogenically formed in the upstream zones, and/or (3) the kaolinite was resedimented by sporadic nepheloid plume activity from the neighboring shallow-water shelf areas. Diagenetic alteration of kaolinite to illite or to chlorite (Chamley 1989) can be excluded because of the 100–130 °C burial temperature suggested by mixed-layer illite/smectites (Huang 1993). The sparse occurrence of kaolinite and abundant mixed-layer phases beside illite and the absence of chlorite suggest a relatively distant source area during deposition.

The clay mineral types do not correlate with the lithologies. This observation suggests that processes forming the limestone/marlstone alternation could not directly affect the formation of the clay minerals.

The above-mentioned observations seem to be in agreement to the model of Raucsik (1997) in which the limestone/marl alternation was explained by productivity or composite productivity/dilutional cycles caused by alternating anti-estuarine and estuarine circulation in an intrashelf basin.

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Ammonite biostratigraphy of the Tata Limestone Formation (Aptian–Lower Albian), Hungary

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The condensed basal layer of the Tata Limestone Formation yielded a rich ammonite fauna preserved as phosphatized and partially glauconitized internal molds. The age of the ammonite assemblage was previously determined as belonging to the Late Aptian nodosocostatum Zone. In the light of recent preliminary investigations the revision of the fauna has become necessary. There are 1443 identified specimens belonging to approximately 45 taxa. Most of the species were first described from Hungary, and this is the first record of the genus Ephamulina and Paracheloniceras outside of Madagascar. Despite the reduced thickness of the embedding sediment the fauna ranges over a wide temporal interval. The assemblage includes some interesting heteromorph specimens such as three new Hamites species. These are the oldest known sure Hamites. Five Aptian ammonite zones (furcata, subnodosocostatum, melchioris, nolani, jacobi) and the Lower Albian tardefurcata Zone were distinguished. The sedimentation and fossilization processes were remarkable: there might have been intermittent sedimentation and/or dissolution with subsequent rough transportation of the fossils. According to the ongoing investigations the ammonite assemblage shows a close Tethyan relationship.

Key words: ammonite stratigraphy, Aptian–Lower Albian, Hungary

Introduction and previous research

The Tata Limestone is one of the most significant Cretaceous formations of the Transdanubian Range. The limestone itself contains only a few, poorly preserved macrofossils – so the age of the formation can only be determined by the fossil-rich lens-like basal layers or by microfossils. These lenses can be found as small pockets filling surface irregularities of the underlying limestone (Fig. 1). The best occurrence of this fossiliferous horizon is situated on the top of the open-air geologic museum of Kálváriadomb (Calvary Hill) in the city of Tata, the type locality.

The Kálváriadomb is a classical locality of Mesozoic formations and fossils in Hungary. The sequence begins with the Upper Triassic Dachstein Limestone and continues up to the Aptian crinoidal limestone (Tata Limestone Formation, TLF). For further stratigraphic details see Fülöp (1975, 1976 in English). The late Prof. J. Fülöp collected a rich fauna from the basal lenses and beds of the Tata Limestone which serves as a basis of the present study. In his monograph

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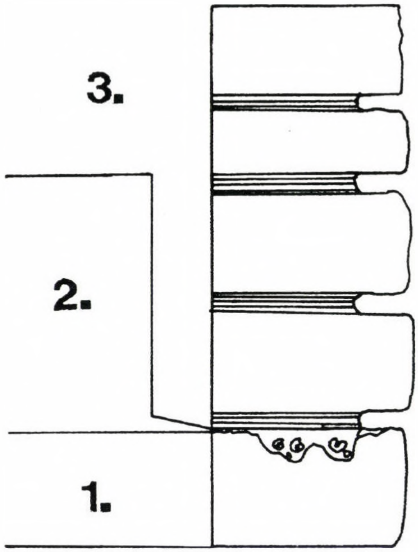


Fig. 1
Schematic stratigraphy of the studied section. 1. Tithonian–Barremian hardground on the top; 2. Basal lenses containing the studied fauna; 3. Tata Limestone Formation

(Fülöp 1976) he devoted a chapter to the TLF and its fauna but without giving systematic descriptions. Previously Fülöp (1954) placed the assemblage in the Upper Aptian *nodosocostatum* Zone (this is not valid name in subsequent zonal schemes) indicating the presence of the whole Clansayesian substage. He analyzed the geologic features of the basal layer and his opinion was the following: "These littoral sediments (were) deposited on the one-time shoreline or not far offshore...".

Since the publication of Fülöp's monograph nobody else worked on the Kálváriadomb except to make some field trip guides. Lelkes (1990) investigated the microfacies types of the Tata Limestone in the Northern Bakony Mountains; he distinguished three microfacies indicating near-shore to offshore regions.

According to Leereveld dinocyst investigations (1992a, 1992b) from boreholes, the age of the overlying Vértessomló Siltstone Formation is Middle Albian, the age of the Tata Limestone is Lower Albian in the upper part of the Vst-8 borehole, and the sedimentary environment was a slightly but gradually deepening middle neritic shelf.

Localities

The fossils were collected from four sites, all in the town of Tata (Fig. 2). There are two artificial trenches in the basal lenses in the Kálváriadomb geologic open-air museum itself, and the Kékkőbánya (Bluestone Quarry). The first locality is the main one – most of the fossils came from here. In the town there were two other temporary outcrops; unfortunately, these places have been buried. One of them is called Vájáriskola (Miners' School), the other is Fazekas utca (Fazekas Street). The assemblage collected from the latter place is quite rich and the proportion of species is fairly similar to that of the Kálváriadomb. For a detailed description of the localities, see Fülöp (1954, 1976).

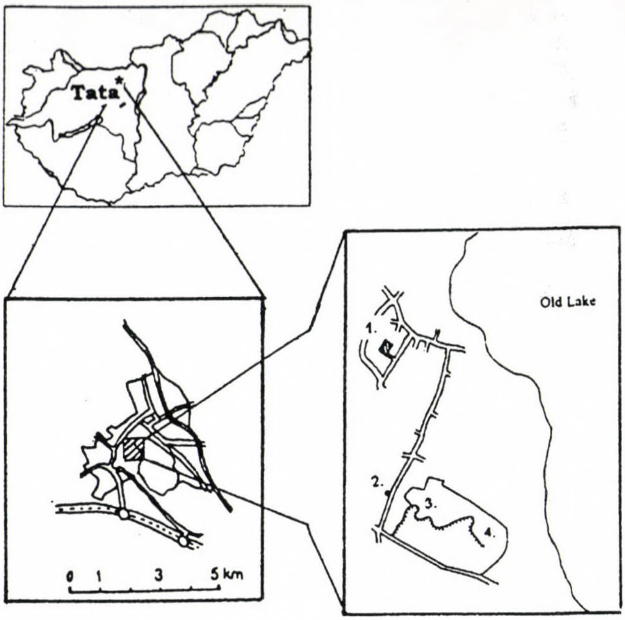


Fig. 2
Sketch map of Hungary and Tata, showing the outcrops of the Tata Limestone Formation cited in the text 1. Vájáriskola (Miners' School), 2. Fazekas utca (Fazekas Street), 3. Kálváriadomb (Kálváriadomb), 4. Kékkőbánya (Kékkőbánya)

The biostratigraphy of the basal bed

The total number of ammonite specimens is approximately 2300 but some of them are poorly preserved. The number of the identified specimens is 1443, ranged approximately in 45 taxa (the systematic description of the ammonite fauna is not yet completed). Fourteen species were already recognized by Fülöp (1976).

All of the material, called the Fülöp Collection, is housed at the Department of Palaeontology, Eötvös Loránd University of Sciences, in Budapest. The ammonites are partially numbered. The assemblages collected from two localities – namely Kálváriadomb and Fazekas Street – are rather similar both in terms of species and of infill material. The ammonites are preserved as phosphatized and glauconitized internal molds, the infill and the matrix are of the same material: grayish-yellow fine-grained sediment. The other two outcrops – Kékkőbánya and Vájáriskola – are a bit different. Here the specimens are worn and filled with red, coarse-grained crinoidal biosparite which is rather different from the other localities. The faunal elements of the TLF came from a highly condensed basal bed and suffered rough transportation; therefore a good preservation of sutures is very rare. The basal beds of the crinoidal limestone yielded the following ammonite fauna. The systematics follows Wright's system (Wright et al. 1996), except for the superfamilies Ancylocerataceae and Turrilitacea of the suborder Ancyloceratina; here the author follows Monks' (1999) system.

Suborder PHYLLOCERATINA Arkell, 1950

PHYLLOCERATIDAE Zittel, 1884

Phylloceras (Hypophylloceras) aptiense (Sayn, 1920)
Phylloceras (Hypophylloceras) velledae (Michelin, 1834)
Partschiceras baborense (Coquand, 1880)
Holcophylloceras guettardi (Raspail, 1831)

Suborder LYTOCERATINA Hyatt, 1889

TETRAGONITIDAE Hyatt, 1889

Tetragonites duvalianus (d'Orbigny, 1840)
Tetragonites heterosulcatum (Anthula, 1899)
Jauberticeras jaubertianum (d'Orbigny, 1851)

Suborder AMMONITINA Hyatt, 1889

DESMOCERATIDAE Zittel, 1895

Melchiorites melchioris (Tietze, 1872)
Valdedorsella getulina (Coquand, 1880)
Puzosia cf. *mayoriana* (d'Orbigny, 1841)
Silesitoides aff. *escragnollensis* (Jacob, 1907)
Beudanticeras (Pseudorbulites) convergens (Jacob, 1907)
Uhligella balmensis (Jacob, 1907)
Uhligella clansayensis (Jacob, 1905)
Desmoceras latidorsatum Michelin, 1838

SILESITIDAE Hyatt, 1900

Neosilesites aff. *balearensis* Breistroffer, 1951

LEYMERIELLIDAE Breistroffer, 1951

cf. *Leymeriella recticostata* Saveliev, 1973

Suborder ANCYLOCERATINA Wiedmann, 1966

ANCYLOCERATIDAE Gill, 1871

cf. *Toxoceratoides honnoratum* (d'Orbigny, 1841)
Tonohamites n. sp.

PTYCHOCERATIDAE Gill, 1871

Ptychoceras laeve (Matheron, 1842)

ANISOCERATIDAE Hyatt, 1900

Protanisoceras acteon (d'Orbigny, 1850)
cf. *Ephamulina trituberculata* Collignon, 1949
Ephamulina curvata Collignon, 1962

HAMITIDAE Gill, 1871

Hamites praegibbosus praegibbosus Spath, 1941
Hamites praegibbosus n. ssp.
Hamites n. sp. 1.
Hamites n. sp. 2.
Hamites n. sp. 3.

DOUVILLEICERATIDAE Parona & Bonarelli, 1897

Cheloniceras cornuelianum (d'Orbigny, 1841)
Cheloniceras (Epicheloniceras) tschernyshewi (Sinzow, 1906)
Cheloniceras (Paracheloniceras) rerati Collignon, 1962

DESHAYESITIDAE Stoyanow, 1949

Deshayesites sp. Kazansky, 1914

PARAHOPLITIDAE Spath, 1922

Diadochoceras nodosocostatum (d'Orbigny, 1840)
Diadochoceras spinosum Mikhailova, 1963
Parahoplites uhligi Anthula, 1899
Parahoplites grossuvrei (Jacob, 1905)
Acanthohoplites bigoureti (Seunes, 1887)
Acanthohoplites aschiltaensis (Anthula, 1899)
Acanthohoplites cf. *andranomenensis* Besaire, 1936
Nolaniceras nolani (Seunes, 1887)
Hypacanthoplites jacobi (Collet, 1907)
Hypacanthoplites cf. *sigmoidalis* Casey, 1965

Hypacanthoplites cf. *milletioides* Casey, 1961
Hypacanthoplites cf. *milletianus* (d'Orbigny, 1841)
Hypacanthoplites cf. *subrectangulatus* (Sinzow, 1908)
Hypacanthoplites elegans (Fritel, 1906)
Hypacanthoplites clavatus (Fritel, 1906)

Table 1
Ammonite zonal scheme for the Aptian (MULA Workshop, 1992 in Hart, 1996) and Lower Albian (Owen 1992).

stage	substage	zones
ALBIAN	Lower Albian	***Leymeriella tardefurcata***
APTIAN	Upper Aptian (Clansayesian)	***Hypacanthoplites jacobi*** ***Acanthohoplites nolani***
	Middle Aptian (Gargasian)	***Parahoplites melchioris*** ***Epicheloniceras subnodosocostatum***
	Lower Aptian (Bedoulian)	***Dufrenoyia furcata*** *Deshayesites deshayesi* *Deshayesites weissi* *Deshayesites tuarkyricus*

Zones known to be present in the basal lenses are shown in bold.

Lower Aptian

The presence of *Deshayesites* sp. is noticeable, because this genus disappeared at the end of the Early Aptian. The huge numbers of *Cheloniceras cornuelianum* d'Orbigny prove that there was sedimentation during the Early Aptian.

Middle Aptian

Tetragonites duvalianus d'Orbigny, *Cheloniceras (Epicheloniceras) martini* d'Orbigny, *Cheloniceras (Epicheloniceras) subnodosocostatum* Sinzow, *Cheloniceras (Paracheloniceras) rerati* Collignon, *Parahoplites* sp., *Diadochoceras nodosocostatum* d'Orbigny, *Diadochoceras spinosum* Mikhailova and *Acanthohoplites aschiltaensis* Anthula are familiar elements of mid-Aptian assemblages. The index ammonite for *subnodosocostatum* Zone is present as well.

Upper Aptian

The nolani Zone is indicated by *Nolaniceras nolani* Seunes, *Tetragonites heterosulcatum* Anthula, *Acanthohoplites andranomenensis* Besaire and *A. bigoureti* Anthula. *Hypacanthoplites jacobi* Seunes, *H. sigmoidalis* Casey, *H. clavatus* Fritel, *H. subrectangulatus* Sinzow well define the *jacobi* Zone.

Lower Albian

Hypacanthoplites cf. *milletianus* d'Orbigny, *H. elegans* Fritel, *Puzosia* cf. *mayoriana* d'Orbigny and *Neosilesites* aff. *balearensis* Breistroffer already appear in the uppermost Late Aptian. The presence of *Beudanticeras (Pseudorbulites) convergens* Jacob, *Phylloceras velledae* Michelin, *Uhligella balmensis* Jacob, two species of *Ephamulina* (only known from the Middle Albian of Madagascar; Collignon, 1962), a huge amount of *Silesitoides* aff. *escragnollensis* Jacob, *Hamites praegibbosus* Spath and *Hamites* species prove the presence of the *tardefurcata* Zone. The determination of the only fragment of cf. *Leymeriella recticostata* Saveliev is somewhat risky due to the poor preservation (no suture preserved). The morphological similarity to *Dufrenoyia lurensis* Kilian is clear and without suture lines and a certain stratigraphic position the identification is not satisfactory. There are several fragments of a *Hamites*-like ribbed form without tubercles. The appearance of the genus *Hamites* (in the fossil record) was at the base of the mid Albian (Spath, 1939). Some uncertain *Hamites* fragments were described by Jacob and Tobler (1906). The investigations of the suture lines and the morphology suggests that the fragments surely belong to the genus Hamites as the oldest known fossil record of the genus.

Conclusion

The age of the revised ammonite assemblage of the Tata Limestone Formation *at the type locality* (Kálváriadomb) can be determined as Lower Albian (other localities as Várhegy at Sümeg are surely of Aptian age as is well determined by ammonite assemblages and foraminifera (Bodrogi 1991)). The ammonite assemblages of this study of five Aptian (*furcata, subnodosocostatum, melchioris, nolani* and *jacob*i Zones) and one Lower Albian (*tardefurcata*) ammonite zones are described in the paper. The presence of a huge amount of *Cheloniceras* and some *Deshayesites* sp. indicate the Lower Aptian but in the absence of the index fossils it cannot be determined precisely which ammonite zones are represented. Most of the preserved ammonites are the familiar elements of the Middle Aptian and the index ammonite of the *subnodosocostatum* Zone is preserved as well. The presence of Upper Aptian is felt to be sure according to its faunal elements described above. The Lower Albian fossils of the basal layer are interesting; however, in some cases they are the oldest representatives of their type. According to its significant Lower Albian assemblage, the age of the basal layers of the basal layers of the Tata Limestone Formation at the type locality is *no younger* than Lower Albian.

Acknowledgements

The author would like to thanks Dr. A. Galácz, Dr. G. Császár, I. Szente and Dr. B. Géczy for their valuable consultation and significant help. The author also thanks Dr. W. J. Kennedy (Oxford University Museum), Dr. H. G. Owen and Dr. N. Monks (both at British Museum of Natural History, London) for their critical review and to L. GY. Mészáros for his technical help. The Hungarian State Eötvös Scholarship and the Soros Foundation Office Hungary supported this study through a research grant to the Oxford University Museum.

Plate I

1. *Holcophylloceras aptiense* Sayn – lateral view of the internal mold
2. *Valdedorsella getulina* (Coquand) – lateral view of the internal mold
3. cf. *Neosilesites* n.sp. – lateral and ventral view of the internal mold
4. *Tetragonites heterosulcatum* (Anthula) – lateral and ventral view of the internal mold
5. *Nolaniceras nolani* (Seunes) – lateral view of the internal mold
6. *Acanthohoplites aschiltaensis* Anthula – lateral view of the internal mold
7. cf. *Leymeriella recticostata* Saveliev – lateral view of the internal mold
8. *Diadochoceras spinosum* Mikhailova – lateral and ventral view of the internal mold
9. *Cheloniceras cornuelianum* (d'Orbigny) – lateral and ventral view of the internal mold

All internal molds figured in natural size. Photographs figured as Figs 3, 4, 7, 8 made by the late Fülöp.

Plate II

1. *Silesitoides* aff. *escragnollensis* Jacob – lateral and ventral view of the internal mold
2. *Ptychoceras laeve* (Matheron) – lateral view of the internal mold
3. *Cheloniceras (Paracheloniceras) rerati* Collignon – lateral view of the internal mold
4. *Hamites* sp. indet – lateral view of the internal mold
5. *Uhligella balmensis* Jacob – lateral view of the internal mold
6. *Hypacanthoplites sigmoidalis* Casey – lateral view of the internal mold

All internal molds figured in natural size except from Fig. 4 (x2). Photographs figured as Figs 1, 3 made by the late Fülöp.

Plate I

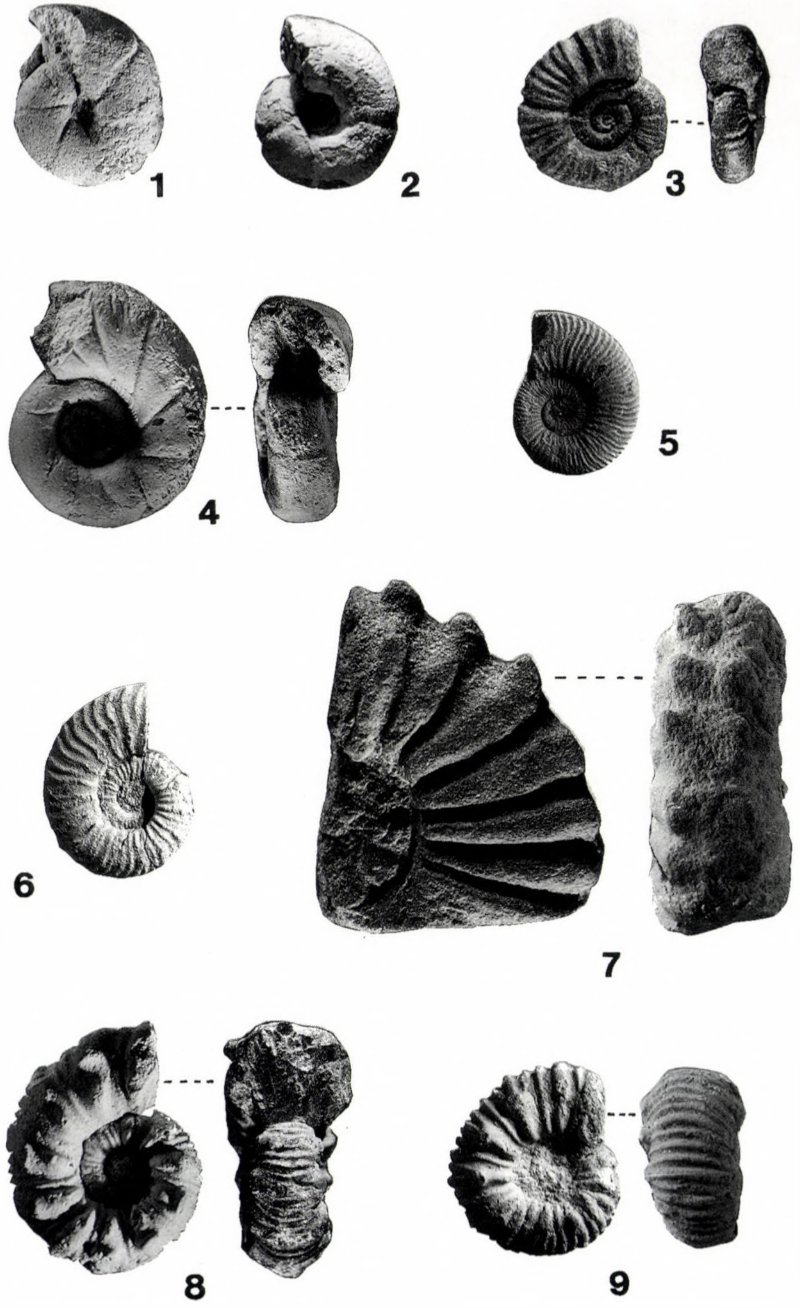
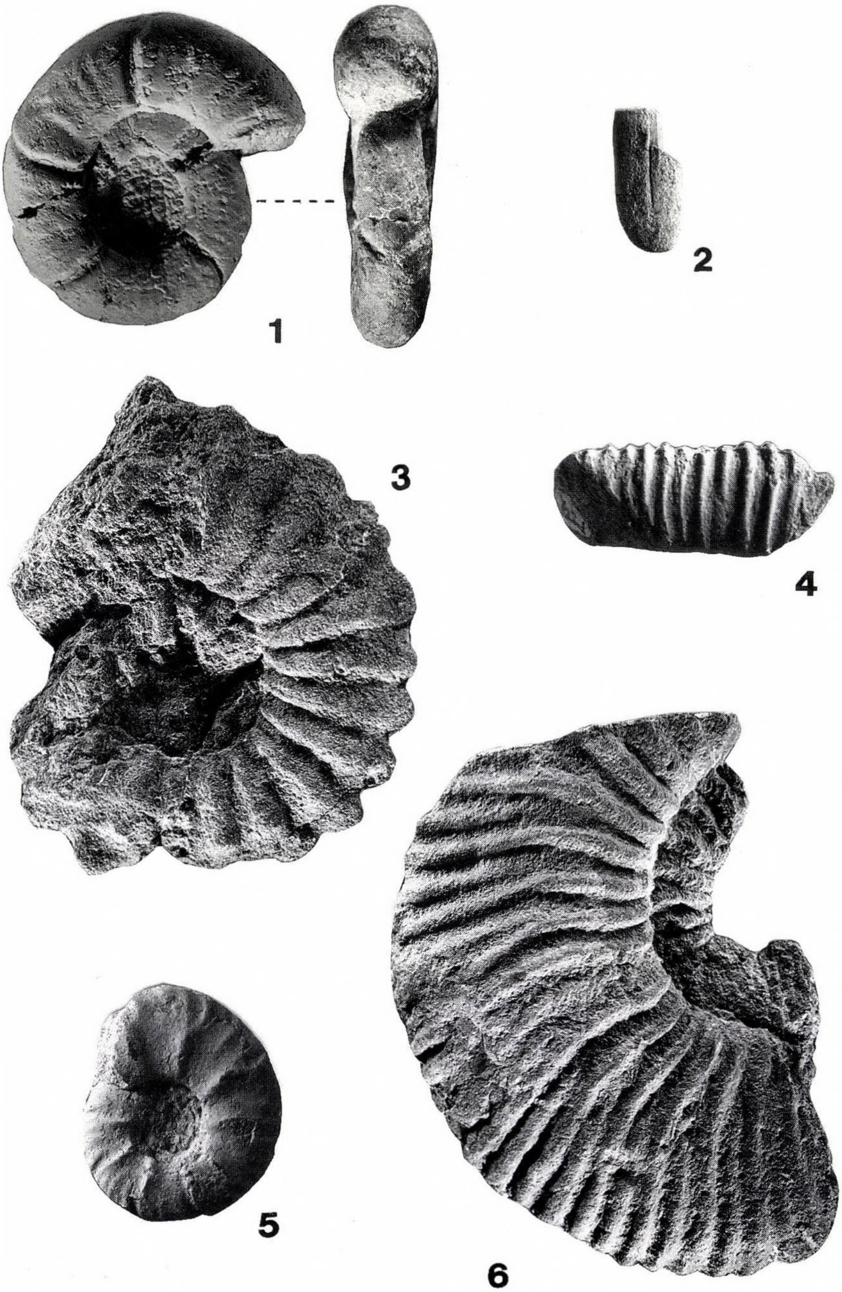


Plate II



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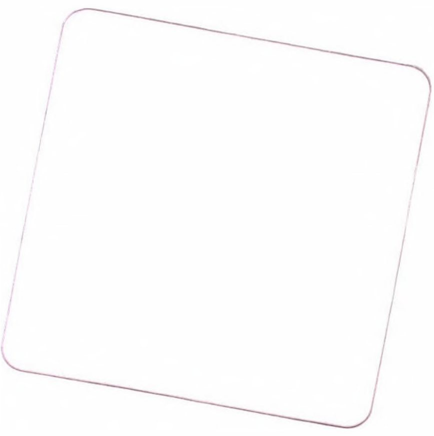

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$\delta^{34}S$ systematics in polymetallic ore deposits in the Mátra and Börzsöny Mts, Inner Carpathian volcanic arc

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The subduction-related calc-alkaline magmatic complexes of the Mátra and the Börzsöny Mts contain significant Pb-Zn ore deposits. Earlier studies on the trace element and isotope geochemistry of the volcanic rocks indicated crustal contamination of the original magmas that also appears in the host rock $\delta^{34}S$ data (ranging from +0.9‰ to +5.6‰). Most of the sulfide minerals studied (galena, sphalerite, chalcopyrite, pyrite) reflect these magmatic compositions. However, in veins formed close to sedimentary country rocks significant $\delta^{34}S$ deviations occur. The presence of both very negative (to -20.7‰) and positive (to +18.1‰) compositions indicate mobilization of biogenic sulfide and seawater-derived sulfate. An attempt was made to apply the data for geothermometric purposes. However, most of the data show disequilibrium fractionations that can be related to the combined effects of contamination and closed-system mineral precipitation.

Key words: calcalkaline volcanic rocks, Pb-Zn ore deposits, sulfides, stable sulfur isotope composition

Introduction

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

Mixing of fluids during ore formation stages has been suggested by Gatter (1987) on the basis of fluid-inclusion microthermometric studies of the Mátra

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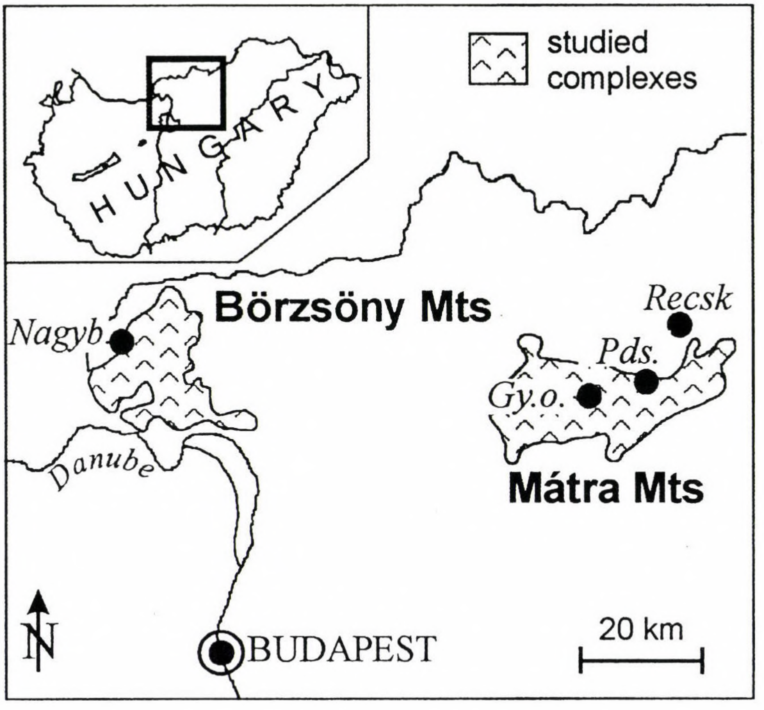

Fig. 1
Location map of the studied ore deposits and host volcanic systems. Nagyb. – Nagybörzsöny, Gy.o. – Gyöngyösoroszi, Pds. – Parádsasvár

Pb-Zn deposit. Subsequently, stable carbon, oxygen and hydrogen isotope compositions of carbonates and inclusion fluids trapped in calcites related to the ore deposits of the Mátra and the Börzsöny Mountains were measured by Pantó et al. (1998) in order to assess the origin and evolution of ore-forming fluids. The complex evaluation of these data revealed that fluid mixing was characteristic for both ore deposits. In the Mátra deposit meteoric water, formation water of meteoric origin and magmatic water were mixed during the ore formation. In the Börzsöny deposit magmatic water and formation water were also present in significant amounts, whereas no direct addition of meteoric water is observed. An additional component with very negative $\delta^{13}C$ and δD values also appears that can be attributed to oxidation of organic matter.

Sulfur isotope compositions of sulfides disseminated in the volcanic rocks and formed in veins might reflect the processes of subduction-related source contamination, mobilization of country-rock materials and isotope fractionations among fluid species. The well-studied volcanic complexes and ore deposits of the Mátra and the Börzsöny Mts might provide insight into the processes that produce sulfur isotope variations in Pb-Zn ore deposits. Since these complexes belong to the Carpathian volcanic chain, the results elucidate the ore formation processes in subduction-related arc systems.

General $\delta^{34}S$ systematics

Ohmoto and Rye (1979) and Ohmoto (1986) have summarized our knowledge of the fractionations of sulfur isotopes in magmatic and hydrothermal systems. Using average sulfur contents and isotope compositions of sulfur reservoirs Ohmoto (1986) and Jambon (1994) have modeled the geochemical cycling of sulfur. However, recent debates on S characteristics of mantle-derived rocks (Ionov et al. 1992; Ionov et al. 1993; Lorand 1993) elucidated the uncertainties of average sulfur contents and isotopic compositions of one of the most important reservoirs, the mantle. Meteorites have S isotopic compositions of around 0‰ (Nielsen 1978), whereas the primary mantle composition seems to be slightly more positive (0 to +2‰, Sakai et al. 1982; Sakai et al. 1984; Chaussidon et al. 1989; Ionov et al. 1992; Eldridge et al. 1991). Mass balance would require the existence of a reservoir with negative $\delta^{34}S$ values. Negative $\delta^{34}S$ values have been found in mantle-derived massive peridotites and xenoliths (Chaussidon and Lorand 1990; Ionov et al. 1992) and have been interpreted by partial melting that results in preferential removal of ^{32}S into the liquid (Chaussidon et al. 1989; Ionov et al. 1992). Significantly more positive $\delta^{34}S$ values (up to +20.7‰) are found in arc volcanic rocks (Woodhead et al. 1987; Alt et al. 1993); these have been explained by invoking sediment recycling (Woodhead et al. 1987; Eldridge et al. 1991) and mixing of mantle-derived S and metasomatic seawater (Alt et al. 1993), since seawater is enriched in the ^{34}S isotope compared to other reservoirs (with a present-day $\delta^{34}S$ value of +20‰, Holser and Kaplan 1966).

Geologic background

A synthesis of the geology of the Mátra Mts has been published by Varga et al. (1975). The geology and the ore deposits of the Mátra Mts have attracted the attention of many researchers, among whom the most important reviews are given by Vidacs (1957) and Kubovics and Pantó (1964).

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

There are two main areas of ore formation in the central and western parts of the Mátra Mts: Parádsasvár and Gyöngyösoroszi–Mátraszentimre. Both deposits are situated in the Lower Badenian "varying andesite". The covering andesite was not affected by the hydrothermal activity. Among the 25 ore veins known in the Gyöngyösoroszi area, 10 veins of 500–1000 m length and 0.8–1.2 m width were mined. Most of the veins are striped with quartz as the dominant mineral. The main ore minerals (Koch 1985; Nagy 1986) are pyrite, gel pyrite, melnikovite, galenite, sphalerite, calcopyrite, marcasite, wurtzite and antimonite formed in several stages (Fig. 2). The average metal contents are: Pb 0.6%, Zn 4.0%, Au 0.8 ppm, Ag 75 ppm. Gatter (1987) and Vető (1988) determined 200–250 °C and 140–170 °C for two main stages on the basis of microthermometric measurements on fluid inclusions in calcites that can be used as estimations of formation temperatures.

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

The ore formation is related to the caldera collapse and took place in two main stages. The first stage is massive breccia-pipe type ("Rózsabánya-type") characterized by pyrrhotite, Fe-rich sphalerite and calcopyrite with sphalerite inclusions. The fluids responsible for the formation of the second stage ores contained significant amounts of Au and Ag. The ore paragenesis of this stage is characterized by arsenopyrite, galenite and Bi-minerals. The mineral paragenesis of the Nagybörzsöny deposit is very complex; its description can be found in the comprehensive book of Koch (1985) (see also Fig. 2). The formation temperatures have been estimated by Vető-Ákos (1982) by means of fluid inclusion microthermometry. The results indicate that the first ore formation stage took place at 260 °C followed by later stages at temperatures down to 110–150 °C.

Analytical techniques

Sulfur isotope compositions of sulfides were determined in the stable isotope laboratory of the University of Calgary. The sulfide samples were first converted to Ag_2S according to the method of Ueda and Sakai (1983). Approximately 1 mg of sulfur was converted to SO_2 by oxidation of Ag_2S at 1000 °C in the presence of a V_2O_5–SiO_2 mixture to assure the SO_2 produced has a uniform oxygen isotope composition. Laboratory standards were typically run with every batch of samples including one with a $\delta^{34}S$ value near CDT (Canon Diablo Troilite) and a second close to NBS 127 international standards. $^{34}S/^{32}S$ ratios

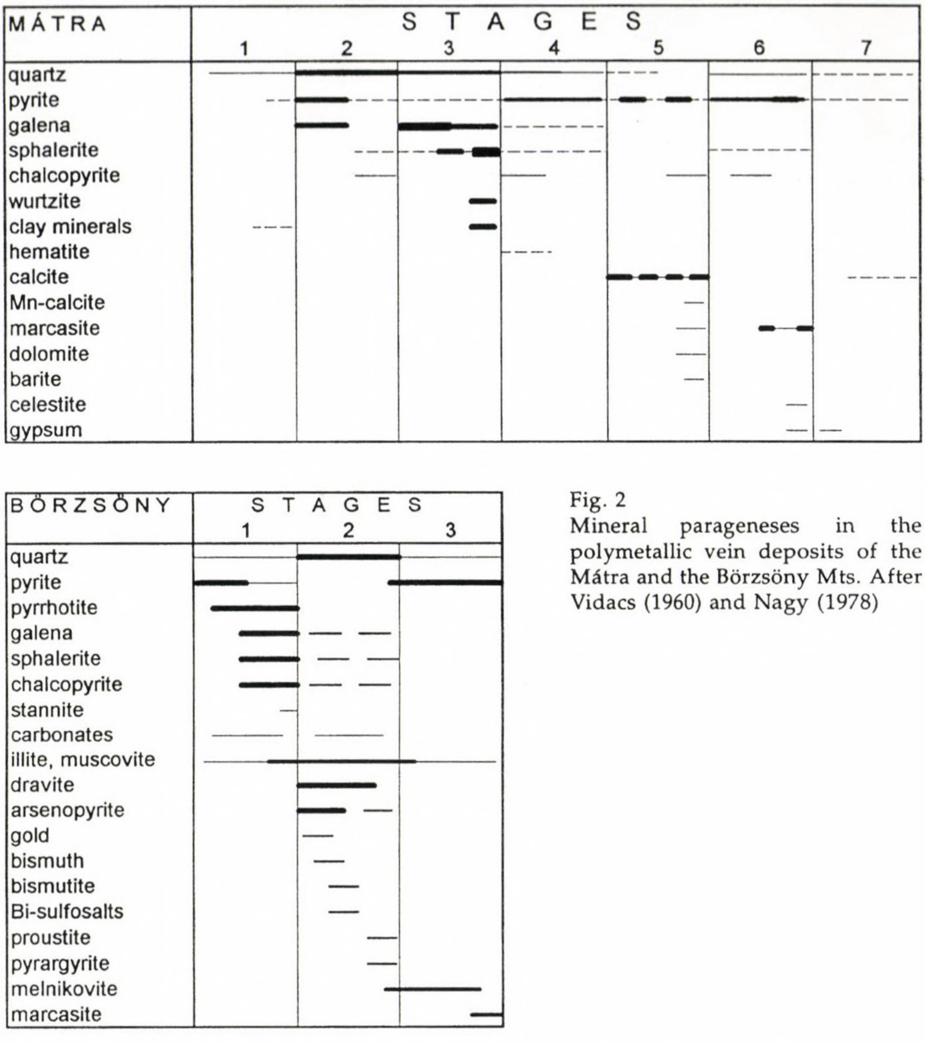

Fig. 2
Mineral parageneses in the polymetallic vein deposits of the Mátra and the Börzsöny Mts. After Vidacs (1960) and Nagy (1978)

were determined in SO_2 gas using a VG602 dual inlet mass spectrometer. The analytical precision is ±0.1‰.

Results

Figure 3 shows the distribution of all $\delta^{34}S$ data. Although the data are widely scattered from −20.7 to +18.1‰, a strong maximum between +1 and +2‰ and a smaller peak at +5 to +6‰ appear. Within the observed range there are distinct areal and mineralogical differences (Figs 4 and 5, Table I).

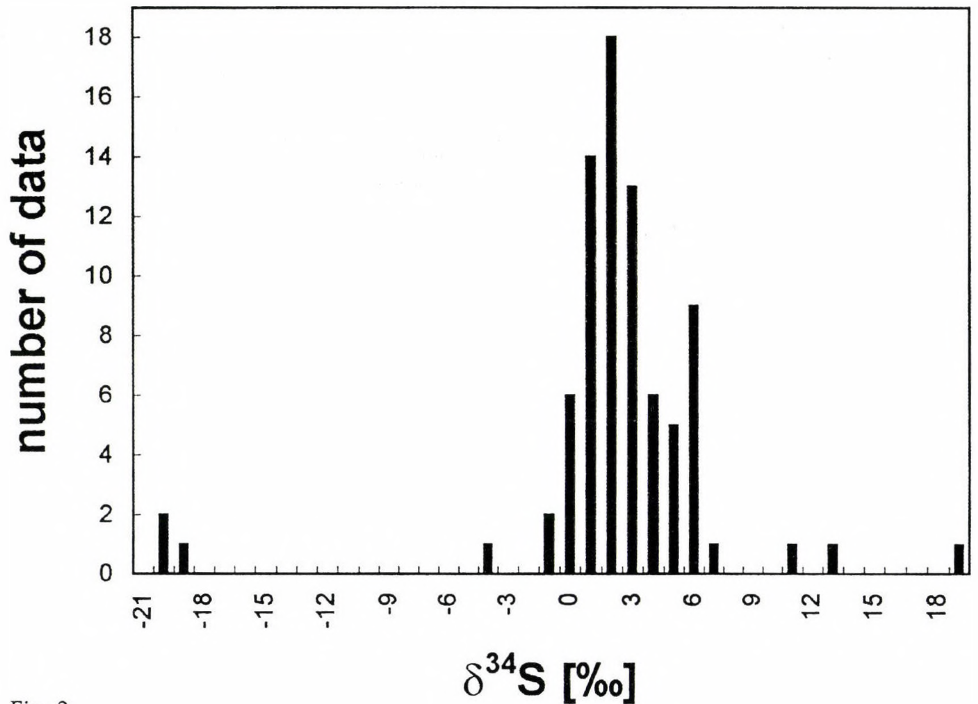

Fig. 3
Histogram of sulfur isotope compositions in ore deposits and related rocks in the Mátra and Börzsöny Mts. The $\delta^{34}S$ data are relative to CDT

As shown in Fig. 4 the host rock compositions are similar in the Mátra and the Börzsöny Mts with average $\delta^{34}S$ values at +3.0‰ and +3.4‰, respectively; thus, these data were plotted together. The data range for the Nagybörzsöny deposit is similar to that of host rocks, whereas the Mátra ore field shows a much greater $\delta^{34}S$ scatter. The sulfides from the central part of the Western Mátra ore mineralization ("Gyöngyösoroszi-1") determine the strong maximum at +1 to +2‰; the large positive and negative deviations appear in the sulfides from the marginal localities of this area ("Gyöngyösoroszi-2"). The smaller peak at +5 to +6‰ is mainly related to the Parádsasvár locality. Beside the areal variations, the $\delta^{34}S$ data depend on the mineralogical composition as well (Fig. 5).

In the Börzsöny Mts galenites, sphalerites and pyrites show increasingly positive $\delta^{34}S$ values in accordance with the theoretical fractionations (Ohmoto and Rye 1979). The $\delta^{34}S$ values of galenites ($\delta^{34}S_{average}$ = +2.4 ‰), sphalerites ($\delta^{34}S_{average}$ = +3.2‰) and pyrites ($\delta^{34}S_{average}$ = +3.6‰) are close to the host rock compositions ($\delta^{34}S_{average}$ = +3.4‰) that can represent the bulk composition of the andesitic magma.

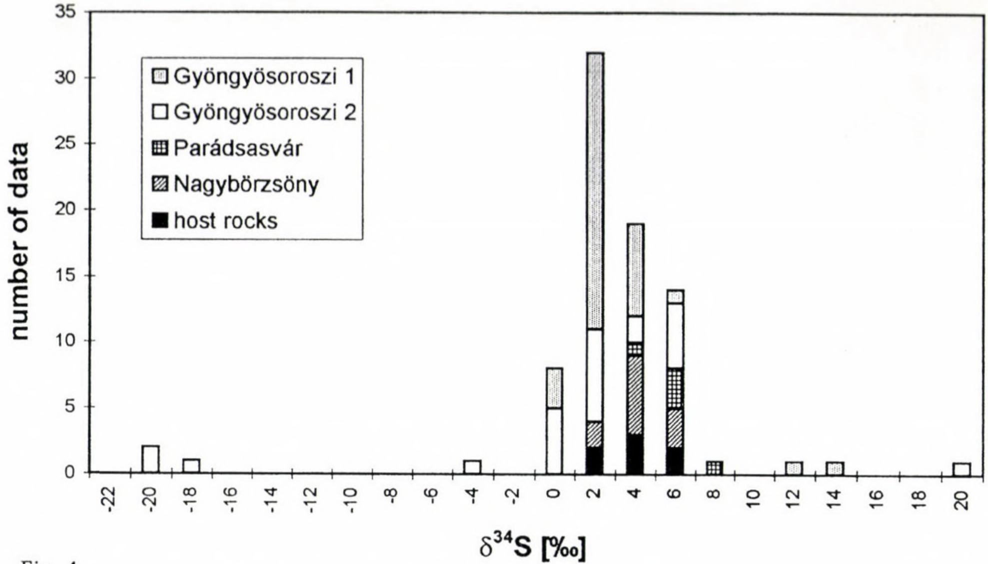

Fig. 4
Sulfur isotope compositions in ore deposit localities in the Mátra and Börzsöny Mts. The $\delta^{34}S$ data are relative to CDT

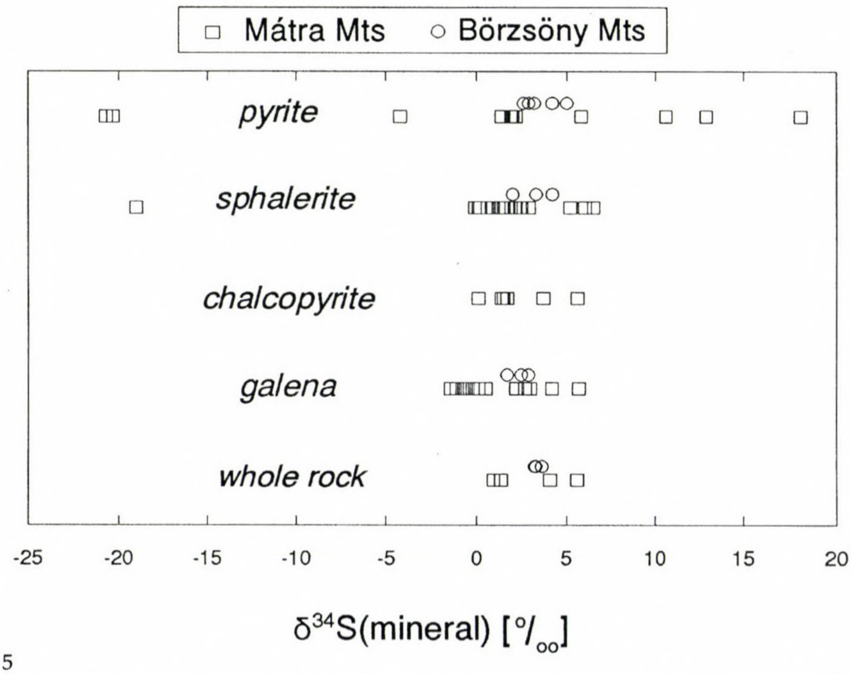

Fig. 5
Sulfur isotope compositions of sulfide minerals in the ore deposits of the Mátra and Börzsöny Mts. The $\delta^{34}S$ data are relative to CDT

Table I

Sulfur isotope compositions ($\delta^{34}S$ relative to CDT in ‰) of sulfide minerals and whole rocks in the polymetallic vein deposits of the Mátra and the Börzsöny Mts, Hungary

No.	locality	level (m)	galena	sphalerite	chalcopyrite	pyrite
Gyöngyösoroszi 1						
1	Károly	150	-1.1			1.9
2	Károly	200		0.2		2.0
3	Károly	250		2.1	1.4	
4	Károly	250		2.9		1.4
5	Károly	460	-1.4	0.6		
6/1	Károly	510		0.8		
6/2	Károly	510		1.9		
7	Károly	510	2.2		0.1	
8	Péter-Pál	460		0.2		
9	Aranybánya II	150				12.9
10	Aranybánya II	150	0.5	1.2	3.7	5.8
11	Aranybánya	350		1.4		
12	Arany-Péter	200	0.1	2.5		1.9
13	Arany-Péter	400		0.6		1.9
14	Arany-Péter	460		1.2		2.2
15/1;2	Hidegkúti		3.0	0.2; 1.2		
Gyöngyösoroszi 2						
17	Szákacsurgó, ***m***			-0.1		*(gp)* 10.6
18	Malombérc	150	-0.7			
19	Malombérc	150	-0.2	0.6	0.1	
20	Malombérc	150	-0.5	1.2	1.7	
21	Mátraszentimre, ***m***	515	5.7	5.2	5.6	
22	Mátraszentimre, ***m***	515				*(gp)* 18.1
23	Mátraszentimre, ***m***	563		6.0		
24	Mátraszentimre, ***m***	608		2.4		*(gp)* -20.7
25	Mátraszentimre, ***m***	608	2.2	5.2		
26	Vereskő, ***m***					*(gp)* -4.2
27	Katalin, ***m***			-19.0		*(gp)* -20.3
28	Jávorkút	350	-0.8	0.1	1.5	
29	Jávorkút	350	-0.6	1.5		
Parádsasvár						
30	Béke shaft		4.2	6.5		
31	Béke shaft, Teodor vein		2.7	6.0	5.6	
Nagybörzsöny						
32	Rózsa adit					2.6
33	Rózsa adit		2.9			3.2
34	Rózsa adit		2.5			
35	Rózsa adit			2.0		
36	Rózsa adit		1.7			
37	Ludmilla shaft					4.2
38	Felső-Fagyosasszony			4.2		5.0
39	Ludmilla shaft			3.3		2.9
Andesitic host rocks						
Mátra Mts						
Gyöngyöstarján bore-hole 4 (110 m)			pyrite-bearing K-trachyte		5.6	
Gyöngyöstarján bore-hole 4 (171.8 m)			pyrite-bearing K-trachyte		1.4	
Gyöngyöstarján bore-hole 4 (87.5 m)			pyrite-bearing K-trachyte		0.9	
Nagylápafő			andesite		4.1	
Börzsöny Mts						
Nagybörzsöny bore-hole 22 (31.5 m)			pyrite-bearing dacite		3.6	
Perőcsény bore-hole 7 (377-378 m)			pyrite-bearing dacite		3.3	
Nagybörzsöny bore-hole 24 (66.2 m)			pyrite-bearing dacite		3.3	

m – marginal veins close to coal-bearing sediments; (gp) – gel pyrite

The data of the Mátra ores show a much larger scatter, although the number of data (n=67) is also much greater than in the Börzsöny Mts (n=14). Regarding the observed scatter, there is no sense in comparing the average compositions of minerals without discussing the effects of possible processes that might be responsible for the deviations.

Discussion

Disseminated sulfides of the andesitic rock as represented by the whole rock sulfur isotope compositions could provide an estimate of the magmatic composition, which is supported by the similarity of the host rock compositions in the Mátra and the Börzsöny Mts. The observed value around +3‰ is significantly higher than the primary mantle composition (0 to +2‰, Sakai et al. 1982; Sakai et al. 1984; Chaussidon et al. 1989; Ionov et al. 1992; Eldridge et al. 1991). Such positive sulfur isotope compositions can be produced by melting of very depleted mantle source rocks that have already suffered extensive partial melting (Chaussidon et al. 1989; Ionov et al. 1992), contamination by subducted crust and seawater recycling (Woodhead et al. 1987; Ionov et al. 1992; Alt et al. 1993), or by H_2S degassing (Sakai et al. 1982; Sakai et al. 1984; Chaussidon et al. 1989).

During an O, Sr, Nd, Pb isotope study on the magmatic rocks of the Mátra and the Börzsöny Mts Downes et al. (1995) concluded that the original mantle-derived magma was contaminated by upper-crustal rocks. According to the oxygen isotope data reported by Downes et al. (1995) the degree of contamination might have been about 3–30%. Using our $\delta^{34}S$ data, average mantle and crustal isotope compositions and sulfur contents, the degree of contamination responsible for the observed positive $\delta^{34}S$ shift can be estimated by mass balance calculation. The input data are: $\delta^{34}S_{mantle}$ = +0.5‰, $Scontent_{mantle}$ = 200 ppm (Sakai et al. 1982; Sakai et al. 1984; Chaussidon et al. 1989; Ionov et al. 1992; Eldridge et al. 1991), $\delta^{34}S_{seawater}$ = +20‰, $\delta^{34}S_{crust}$ = +7‰, S $content_{crust}$ = 780 ppm (Holser and Kaplan 1966), $\delta^{34}S_{this\ study}$ = +0.9 to +4.1‰. Regarding the lack of data on sulfur contents and isotope compositions in basement rocks, the obtained boundary values of contamination (0.3 to 14%) should be taken as a rough estimate. However, the result of mass balance calculation does not contradict the oxygen isotope data.

Taking into consideration that subaerial volcanic systems are studied in this paper, SO_2-degassing cannot be excluded. This influence, however, would more likely produce local variations, explaining the scatter in the rocks of the Mátra Mts.

Similarly to the host rock data, most of the sulfide minerals have $\delta^{34}S$ values between –2 to +7‰ (Fig. 4). Most of the outlying data derive from pyrites of the marginal ore veins of the Gyöngyösoroszi area. The explanation of this feature can be found in the presence of coal deposits that are mined in the northern part of the Mátra Mts. These coals might likely contain both biogenic pyrite and seawater-derived sulfate, which means a large $\delta^{34}S$ variability (see for example Hoefs 1987). The pyrites of the early ore generations found in the central ore field of the Gyöngyösoroszi area have $\delta^{34}S$ values in a narrow range (+1.4 to +2.2‰), whereas the gel pyrite related to the latest generation and formed in the marginal veins have either very negative or very positive $\delta^{34}S$ values. The Aranybánya II vein is at the deepest position among the veins studied and is situated nearest the country rocks. Thus, the observed negative and positive deviations can be attributed to the mobilization of sedimentary sulfur.

These results should be compared with the earlier studies on stable isotope compositions of calcites and fluid inclusions trapped in them (Pantó et al. 1998). In the Mátra deposit mixing of meteoric water, formation water of meteoric origin and magmatic water was presumed, whereas in the Börzsöny deposit the presence of magmatic water and formation water was suggested by the data. An additional component with very negative $\delta^{13}C$ and δD values also appear that can be attributed to oxidation of organic matter.

After comparing the data on calcites with the $\delta^{34}S$ results the former ideas have changed slightly. The calcites and the sulfides of the veins of the central part of the Gyöngyösoroszi area have $\delta^{13}C$ and $\delta^{34}S$ values around 0‰ and +2‰, respectively, whereas the marginal veins of the area show positive $\delta^{34}S$ and negative $\delta^{13}C$ shifts. The $\delta^{13}C$ data around 0‰ and -4‰ were interpreted as indications of mobilized sedimentary carbonate and magmatic CO_2 by Pantó et al. (1998). However, the present $\delta^{34}S$ dataset would rather suggest that the fluids with magmatic $\delta^{34}S$ contained dissolved sedimentary carbon, and the lower $\delta^{13}C$ values originate from oxidation of organic materials. In the case of the Börzsöny deposit the $\delta^{34}S$ data suggest mainly magmatic sulfur; thus, it could not be correlated with the variations in the C and O isotope data. Hence the presence of magmatic CO_2 cannot be excluded.

More than 90% of the sulfide data fall in the range of –2 to +7‰ (Fig. 4). The $\delta^{34}S$ trend in the galenite-chalcopyrite-sphalerite-pyrite series indicate the possibility of geothermometric evaluation of isotope fractionations. Since galenite is the most ^{32}S-rich mineral among those studied, the $\Delta^{34}S$(mineral-galenite) pairs should be investigated. Due to the large scatter observed in pyrites, pyrite was excluded from the evaluation.

In the $\delta^{34}S_{mineral}$-$\delta^{34}S_{galenite}$ plot the equilibrium isotope fractionations at different temperatures are represented by straight lines (Fig. 6). The data points of minerals in equilibrium would be scattered along these lines, whereas data fields cross-cutting the lines indicate disequilibrium. Figure 6 shows apparent disequilibrium fractionations. About 1/3 of the data points fall in the range of the fractionation at the presumed formation temperature (200–300 °C), whereas the other data would yield unreasonably high temperatures. The reason for this observation might be the contaminating effect of mobilized sedimentary sulfur and/or fractionation during continuous precipitation of galenite. Since galenite is enriched in ^{32}S relative to the ore-forming fluid, its precipitation

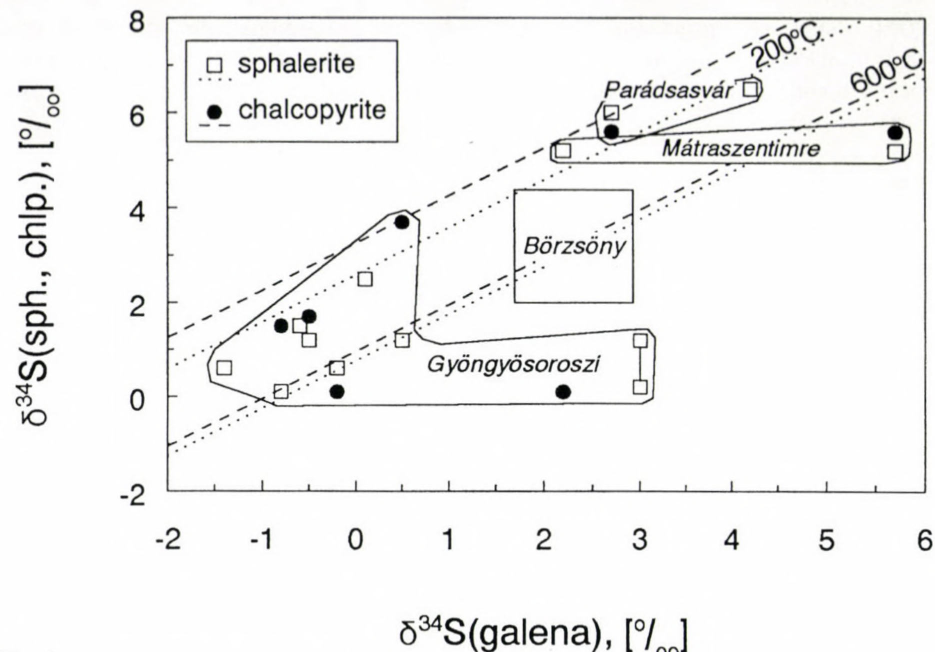

Fig. 6
Sulfur isotope compositions of sphalerites and chalcopyrites as a function of $\delta^{34}S$ values of galena (all data are relative to CDT). Equilibrium isotope fractionations are from Ohmoto and Rye (1979)

depletes the fluid in ^{32}S resulting in increasing $\delta^{34}S$ in the subsequently formed sulfide minerals. As chalcopyrite-galenite and sphalerite-galenite pairs were not studied in the Börzsöny samples, only the data ranges are shown in Fig. 6. These data would also indicate irrealistically high temperatures; thus, these fractionations are regarded as disequilibrium ones. The reason for this disequilibrium can also be the joint influence of contamination and continuous precipitation.

Conclusions

During the investigation of sulfur isotope compositions of ore deposits in the calc-alkaline volcanic suites of the Börzsöny and Mátra Mts the following conclusions were drawn:

1. Based on host rock sulfur isotope compositions the magma might have suffered significant crustal contamination which is in compliance with the results of earlier O, Sr, Nd and Pb isotope studies.

2. The sulfide minerals that belong to the earlier ore generations have $\delta^{34}S$ values close to the presumed magmatic compositions. The presence of mobilized sedimentary sulfur is most significant in the marginal veins of the

Gyöngyösoroszi deposit that belong to the latest ore generations and are closest to the coal-bearing country rocks. Comparing the sulfur isotope compositions of the present study and earlier results on C, O and H isotopes in calcites and inclusion fluids, significant amounts of magmatic CO_2 in the ore-forming fluids of the Mátra Mts. can be excluded. The most likely interpretation is that the magmatic water carrying magmatic sulfur was mixed with solutions derived from the country rocks.

3. The fractionations between sulfide minerals are not useful for geothermometric evaluation due to the combined effects of contamination by mobilized sedimentary sulfur and fractionation during mineral formation.

Acknowledgements

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Geologic setting of the Pre-Tertiary basement in Vojvodina (Yugoslavia) Part II: The north part of the Vardar zone in the south of Vojvodina

Miruna Čanović, Ruža Kemenci

Part I of this study was published in Acta Geologica Hungarica, Vol. 40, in 1997. In that part were described the geologic formations at the base of the Tertiary in those northern parts of the Vojvodina that belong to the southern Tisza Mega-unit.

Part II focuses on the geologic formations of the Vardar Zone, particularly the ophiolite and flysch occurrences found in many wells that penetrated the Pre-Tertiary basement in the south of the Vojvodina and in outcrops in the Fruška Gora. According to their stratigraphic and facial characteristics and magmatism these litho-structural units can be correlated with similar units in the Vardar zone to the south (Figs 1, 2).

A zone of gabbro and diabase present with metamorphites along the margin of the Serbo-Macedonian Massif (Fig. 1), analogous to the geology in central Serbia. To the west is a belt composed of serpentinite, rare Jurassic and widespread Lower Cretaceous sediments. This part of the southern Banat is identical in distribution of units to the main belt of the Vardar Zone.

Along the Danube and the Tisza rivers a ridge in the Pre-Tertiary basement extends as a continuation of the Kosmaj – Avala Ridge at the south.

To the west, in the southern Backa and in the Srem with the Fruška Gora, the geology of the Pre-Tertiary basement is different and very complex. This area consists of Triassic unmetamorphosed and metamorphosed sediments in greenschist facies, Jurassic (?) basinal anchimetamorphosed facies (DCF), serpentinite bodies, diabase and gabbro blocks, as well as Senonian limestone. The geologic setting of this chaotic complex, according Karamata (1999, pers. comm.), points to a trench.

The first unit covering all older, separated units is the Karadjordjevo Formation of Senonian age. The Karadjordjevo Formation is rich in andesitic and associated volcanics, similar to subduction-related ones. This unit shows an analogy with the upper levels of Turonian(?) – Senonian andesite-bearing sediments south of Belgrade and with identical formations in the western part of the South Apusenides described by Lupu (1991). Finally the Maastrichtian Torda Flysch Formation overlies the former units as a belt 15 to 25 km wide extending from the Danube in the west to the Romanian border in the east. At the north its boundary is the transcurrent fault system to the south of Tisza Mega Unit.

In the southern Srem Carboniferous and Lower Triassic deposits belonging to the Jadar Block were found beneath the Tertiary.

Key words: Pre-Tertiary basements, Vojvodina, Fruška Gora, geologic formations, litho-structural units, Vardar Zone

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Introduction

This paper continues the description of the geologic setting of the Pre-Tertiary basement in Vojvodina, the general features of which were described in the monograph by Čanović and Kemenci (1988).

Part I of this study described the geologic formations at the base of the Tertiary in those northern parts of Vojvodina that belong to the southern Tisza Mega-unit, published in Acta Geologica Hungarica (Vol. 40) in 1997.

Part II focuses on the geologic formations of the Vardar Zone, particularly the ophiolite and flysch occurrences encountered in many wells that penetrated the Pre-Tertiary basement in the South of the Vojvodina and in outcrops in the Fruška Gora.

Since 1995, when the first part of this study (Kemenci and Čanović 1997) was submitted, new data on the geology of central parts of Northern Serbia have been collected and summarized. For this reason it has been possible to obtain a better insight into the geology of the Pre-Tertiary units in the Vojvodina, to the north of the Sava and Danube Rivers and the south of the major fault system of the southern border of the Tisza Mega-unit.

The new data of the Pre-Tertiary basement in the Vojvodina south of the Tisza Mega-unit is presented in Figs 1, 2.

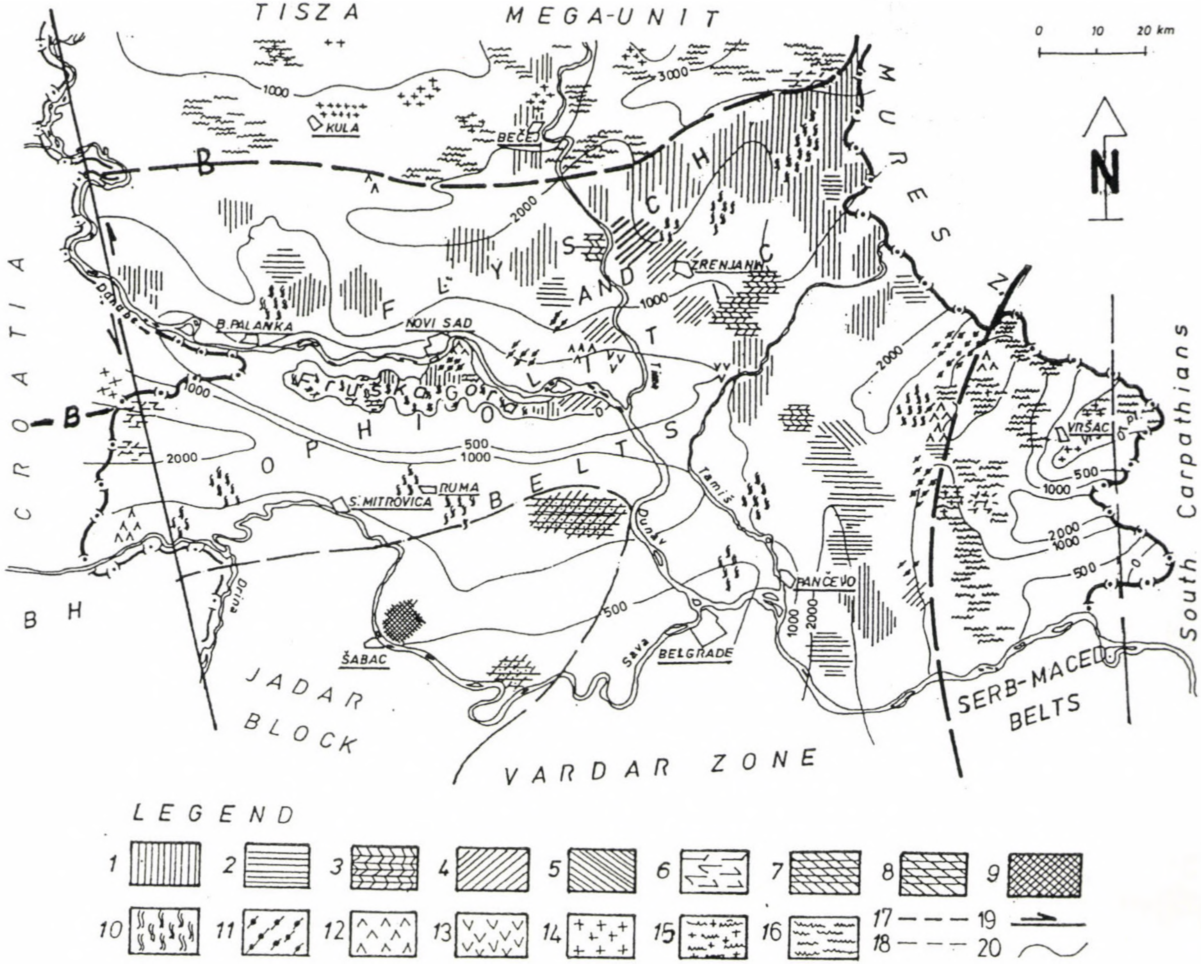
In this paper, first the data on the Fruška Gora, where outcrops are present, some with continuous sections, will be presented, followed by the data on Paleozoic and Mesozoic units encountered in the wells follow.

Geology of the Vardar Zone of South Vojvodina

Brief review of the geology of the Fruška Gora

Interest in gathering geologic information of the Fruška Gora began more than 160 years ago. A monograph on the Fruška Gora was published by Petković et al. (1976) describing in full detail all the phases of the early efforts to recognize the geologic features of the range. This study will be restricted to a brief review of the Pre-Tertiary geologic formations (Fig. 3).

Fig. 1 →
Geologic Map of the Pre-Tertiary Basement in the southern Vojvodina (according to well data and outcrops in the Fruška Gora) – From Čanović, M. and Kemenci, R. (1988), Čanović (1991). Legend: 1. Upper Cretaceous basinal facies (clastic, carbonate, and flysch); 2. Lower Cretaceous basinal facies; 3. Lower Cretaceous shallow-water and reef facies (Urgonian type); 4. Middle–Upper Jurassic basinal facies; 5. Liassic basinal facies; 6. Upper and Middle Triassic, various basinal limestones, recrystallized, partly marbled calcschist; 7. Middle Triassic shallow-water carbonate facies; 9. Carboniferous (Bashkirian) shallow-water facies; 10. serpentinite; 11. gabbro; 12. diabase, spilite; 13. rhyolite; 14. granite, granodiorite; 15. migmatite; 16. crystalline schist (in general); 17. megastructural unit boundary; 18. structural unit boundary; 19. transcurrent fault; 20. isobaths, approximated from well data


TISZA MEGA-UNIT
MURES Z.
CROATIA
BH
JADAR BLOCK
VARDAR ZONE
SERB-MACED. BELTS
South Carpathians
FLYSCH AND C
OPHIOLITS BELTS
Fruška gora
KULA
BEČEJ
ZRENJANIN
B. PALANKA
NOVI SAD
S. MITROVICA
RUMA
VRŠAC
PANČEVO
ŠABAC
BELGRADE
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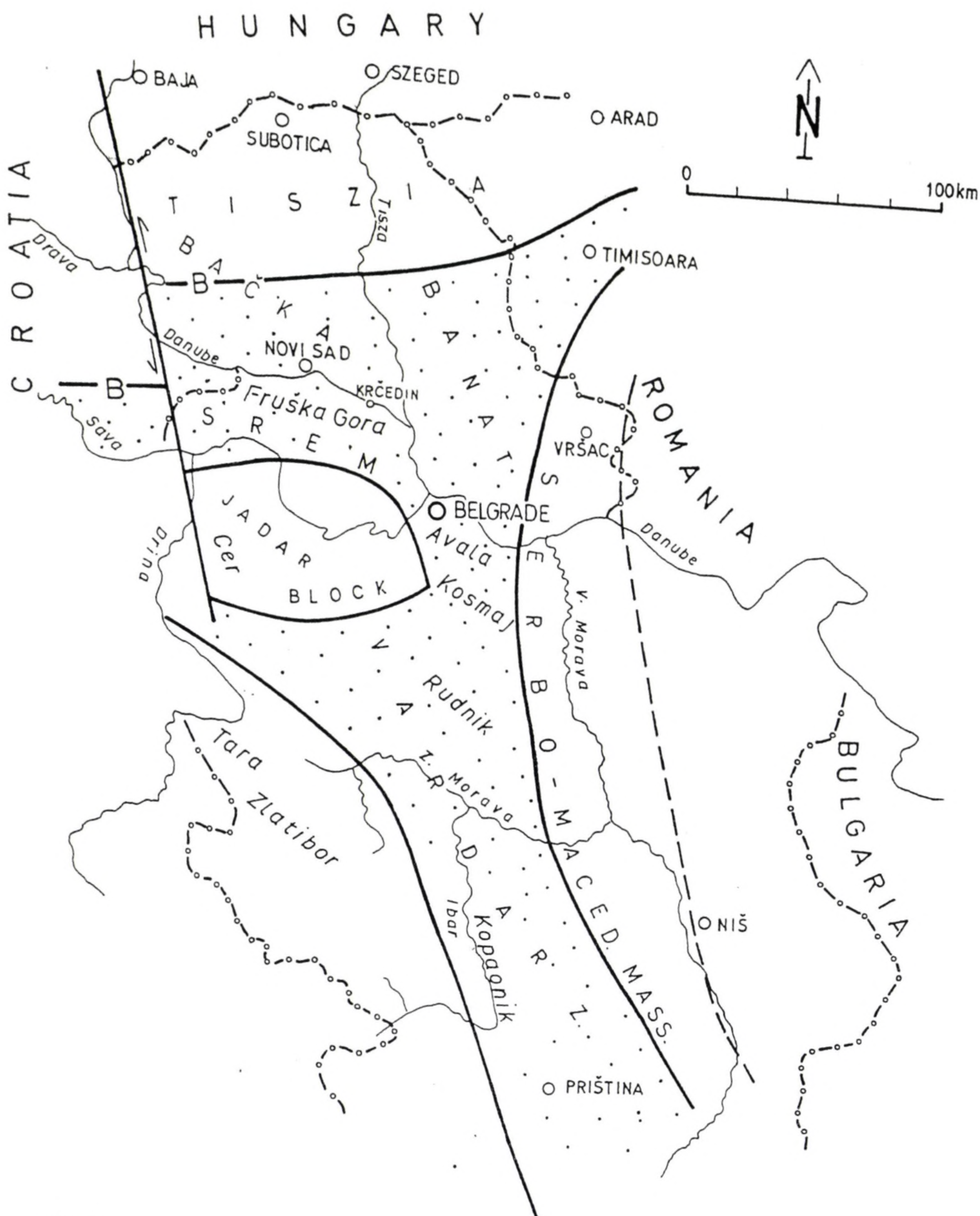

Fig. 2
Sketch map of geotectonic units distribution in Serbia

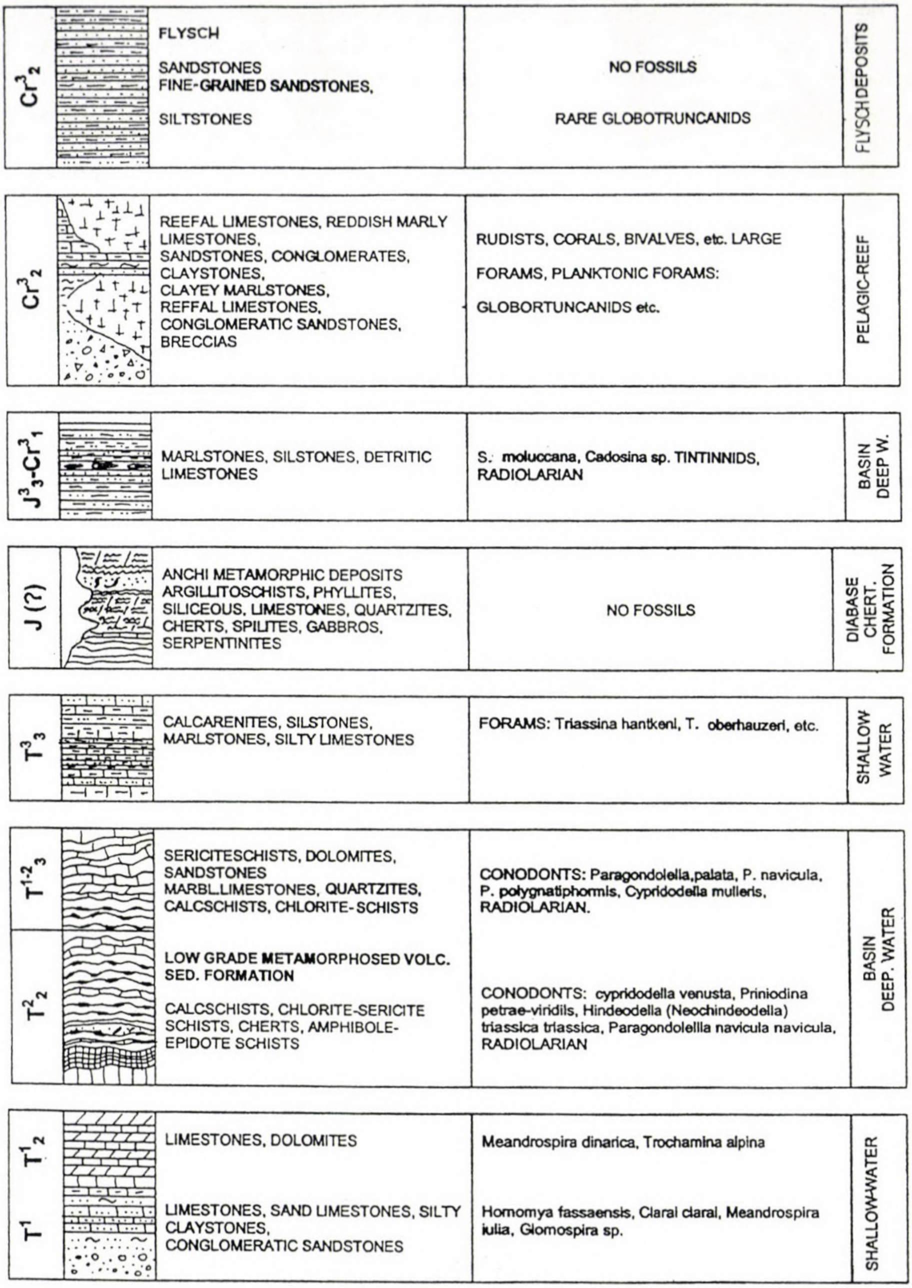

Fig. 3
Mesozoic complex of the Fruška Gora (according to Stojadinović, Čanović, Kemenci). Modified by Čanović, Kemenci (double line indicates different geologic environments)

The first data on the metamorphites of the Fruška Gora were provided by Kišpatić (1883, 1886, 1887) and Koch (1896). Koch considered the metamorphic rocks Azoic; Cvijić (1924) dated them as younger Paleozoic, Djurdjanović (1971) found Ladinian, Carnian and Norian conodonts in recrystallized limestones from this complex metamorphosed in greenschist facies. Čiculić and Aleksić (1971) state the presence of either allochthonous or autochthonous glaucophane schist and consider them as post-Middle Triassic; however, Milovanović et al. (1995) dated such schist near Sr. Karlovci as Lower Cretaceous (123 Ma by the K-Ar method). It is still an open question if this metamorphic complex on the west flanks of the Fruška Gora also contains rocks younger or older than Triassic, or if there are two groups of metamorphic rocks differing in age and origin, but also occurring in different belts.

Triassic deposits are exposed in two belts; Lower to Middle Triassic shallow-water deposits occur along the southern flanks of the Fruška Gora and Middle to Upper Triassic low-grade metamorphites build the belt up to the ridge.

The Lower Triassic is developed in clastic and carbonate facies with foraminifera (*Meandrospira iulia* (Prem.-Sil.)) and bivalves, such as *Homomya fassensis* (Weiss) and *Claraia clarai* Emmr. (Ciculic and Pantic (1969)). The thickness of Lower Triassic in the Fruška Gora ranges from 50 to 100 m (Djurdjanovic et al. 1971).

Anisian formations, developed in shallow-water carbonate facies with foraminifers (*Meandrospira dinarica* Koch.-Dev., Pant., *Trochamina alpina* Krist.-Toll.) occur close to the former unit, at the southern flanks of the Fruška Gora Čiculić and Pantić 1969).

In the calcschist and siliceous limestone sequences of the low-grade metamorphic rocks complex the deep-water facies of the Ladinian was proved by conodonts (Djurdjanović 1971). Along with calcschist and siliceous limestone the Ladinian stratigraphic column contains various metamorphosed (in greenschist facies) sedimentary and volcanic rocks (alternation of quartz-albite-chlorite schists, amphibolite schists, quartz-albite-epidote schists). The thickness of the Middle Triassic package has been estimated to range between 300 and 500 m.

Within the complex of metamorphic rocks, Djurdjanović (1971) also proved with conodonts the existence of Carnian and Norian metasediments. In the Norian the basin began to shallow. The Carnian and Norian deposits are 100 to 200 m thick.

Čanović et al. (1977) proved the presence of Rhaetian Kössen Beds (clastic and carbonate deposits). The shallow-water limestone exhibits a wealth of foraminifera and other faunal remnants: associations with *Triassina hantkeni* Mjz., *Triassina oberhauseri* Koehn.-Zenn., *Tetrataxis inflata* Krist.-Tollm., etc. The Rhaetian limestone also contains alga, ostracod, echinoid, brachiopod and mollusk remnants. Šikić et al. (1975) found similar Rhaethian facies in the adjacent Slavonian area (Papuk). The thickness of the Rhaethian in the Fruška

Gora is estimated to range between 100 and 200 m. These deposits are not metamorphosed.

The tectonized assemblage in the central part of the Fruška Gora near the Rhaethian deposits possibly belongs to the Jurassic. These assemblage consists of anchimetamorphosed mafic rocks, slate, limestone, chert, metasandstone, metabasalt and gabbro. The relationships among lithological members are unclear because of thick cover and can be traced only locally, where more resistant members crop out. Long lenses of serpentinite are probably members of the same geologic unit. These rocks have a comparatively large extension in the Fruška Gora. They are always tectonically associated with other formations. Three east-west trending zones may be distinguished. The serpentinite is highly tectonized and changed by hydrothermal agents. This unit is analogous to the "Diabase-chert formation" or the ophiolite melange of the Vardar Zone, metamorphosed under very low to low-grade metamorphic conditions.

Another example of the evolution of the youngest Jurassic and the oldest Lower Cretaceous deposits was proved by Canovic and Kemenci (1974) on the easternmost flanks of the Fruška Gora near Krcedin. These deposits consist of basin clastites, referred to in literature as the Jurassic–Neocomian flysch. The pelitic beds of this series contain some tintinnid, calcispherullide and radiolarian remnants. The sporadically occurring detritic limestone and calcarenite beds abound with redeposited detritus of shallow-water organisms (*Clypeina jurassica*, etc.).

The Upper Cretaceous is represented in the Fruška Gora by younger Senonian carbonate-clastic, pelagic and flysch deposits.

The carbonate-clastic series is well developed at the northwestern flanks of the Fruška Gora. The Maastrichtian clastic developments, described as Hypersenonian by Pethö (1892), contain a rich fauna of corals, brachiopods, gastropods, lamellibranchs, cephalopods and larger foraminifera. At this northwestern flank of the mountain Late Senonian pelagic globotruncana limestone was encountered in outcrop (Danilova 1960).

The flysch series are well developed at the northeastern flanks of the Fruška Gora. These deposits are characteristic of a rather rhythmic sequencing of graded bedding and laminations. The flysch series were studied by Radoševic and Marković (1967) and Djurdjanović (1971). The estimated thickness of the turbidite series is 960 m.

Paleozoic evidenced by well data from the southern Vojvodina

Carboniferous

The presence of Bashkirian deposits was proved by Canovic (1991) based on the findings from a well near Platičevo (No. 20; Fig. 4.) in the southern Srem.

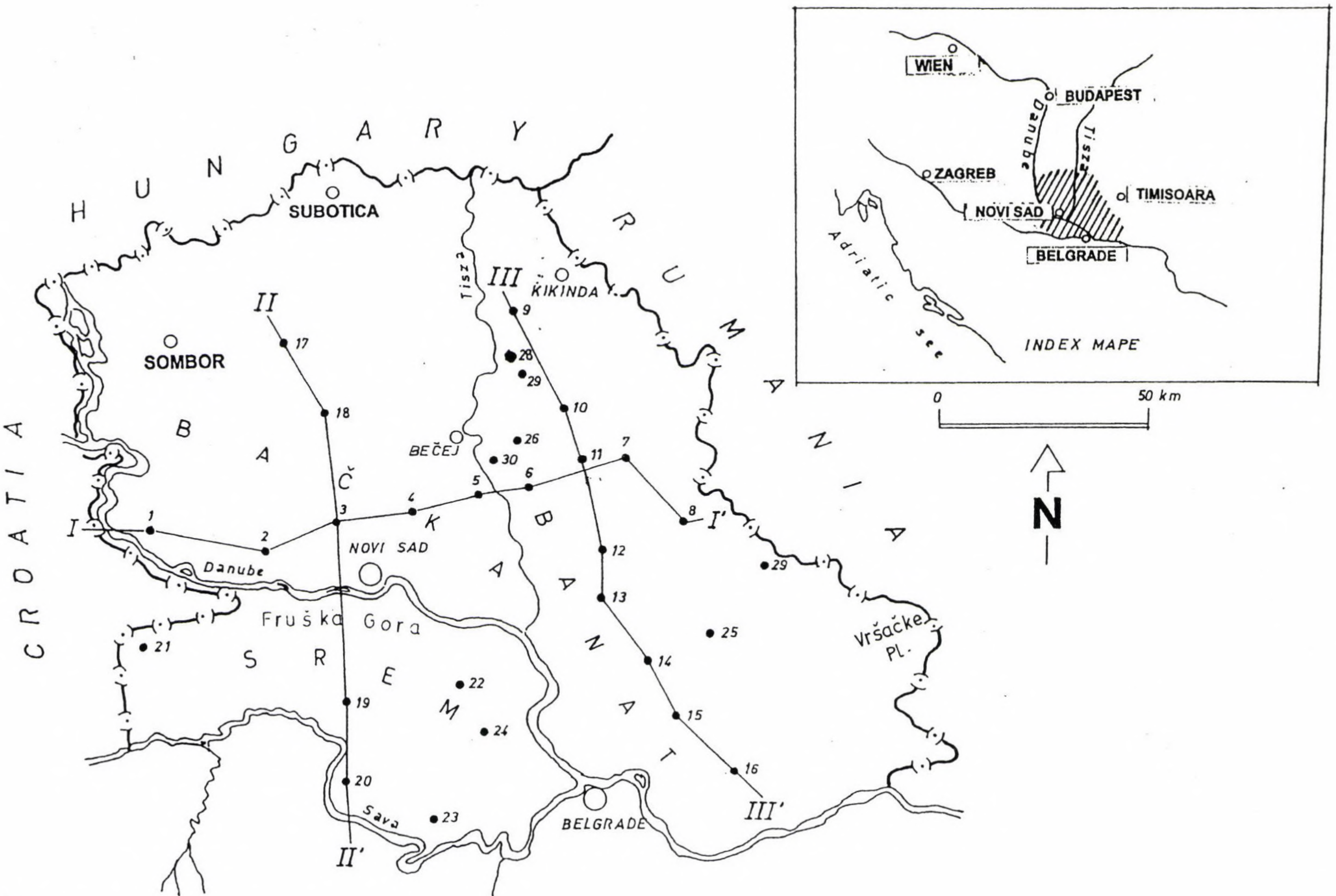

Fig. 4
Index map of well positions and cross-section lines

No other paleontological proof for the presence of the Paleozoic in the Fruška Gora, nor in the entire territory of the southern Vojvodina, has yet been found.

The deposits are characteristic for shallow-water environment: carbonaceous sandstone, limestone, and siltstone containing benthonic foraminifera (*Millerella marblensis* Thorm., *Lasiodiscus* sp., *Archeodiscus* sp.), Amodiscidae, Erlandinae and brachiopod debris. These Platicevo deposits are identical to the Bashkirian sediments in the Jadar block in West Serbia. Capped by Tertiary deposits the Carboniferous at Platičevo overlies phyllite beds.

Mesozoic evidenced by well data from the southern Vojvodina

Triassic

Čanović (1970–1990) established the presence of the Triassic by evidence from three wells in the southern Srem.

The Lower Triassic (Seissian) was identified below the Tertiary in wells at Kupinovo (No. 23; Fig. 4.) and Vojka (No. 24; Fig. 4). These beds include marly and fine-grained limestone with annelids, ostracods and foraminifera (*Meandrospira pussila* (Ho), *Glomospira sinensis* (Ho) *Gl. sigmoidalis* (Ho)).

The well at Golubinci (No. 22; Fig. 4) provided evidence of Anisian shallow-water deposition in the form of organogenic limestone containing associations of foraminifera and algae: *Pilammina densa* Pant., *Meandrospira dinarica* Koch-Dev., and *Physoporella pauciforata pauciforata* Bystr.

Some occurrences of shallow water limestones with the algae *Cayeuxia* sp. encountered in the well at Bački Petrovac (No. 3; Fig. 4.) also belong to the Anisian. Above them are found low-grade metamorphites analogous to the Ladinian–Upper Triassic rocks in the Fruška Gora (Figs 3, 5). Several wells at Ilinci in the western Srem (No. 21; Fig. 4) drilled into calcschist identical to that in outcrops in the Fruška Gora in the vicinity of the well site that contains Triassic conodonts.

Jurassic

On the basis of the palynomorph find in the well at Padina (No. 25; Fig. 4) in the southern Banat by Baltes (1959) suggested the existence of Liassic sediments. These formations consist of black anchimetamorphosed shale, siltstone and sandstone that contain *Ophioglosmudelectus* sp., *Osmudites plicata* Kara-Mursa, *Ginco praecuta* Balt. and *Benettites dilicidus* Balt.

Below the Lower Cretaceous formations in the well near Banatsko Novo Selo (No. 14; Fig. 4), Pantic and Dulic (1990) found identical deposits with Lower Liassic palynomorphs: *Concavisporites* sp., *Deltoidospora* sp., *Cyathidites* sp., etc.

The metasediments drilled in the central Banat in the area of Elemir (No. 6;. Fig. 4), generally referred to as the Elemir Formation, were distinguished as Dogger–Malm basinal developments that evolved in a deep-water environment. The dating was based on pollen and spore analyses: *Contignisporittes durbonensis*

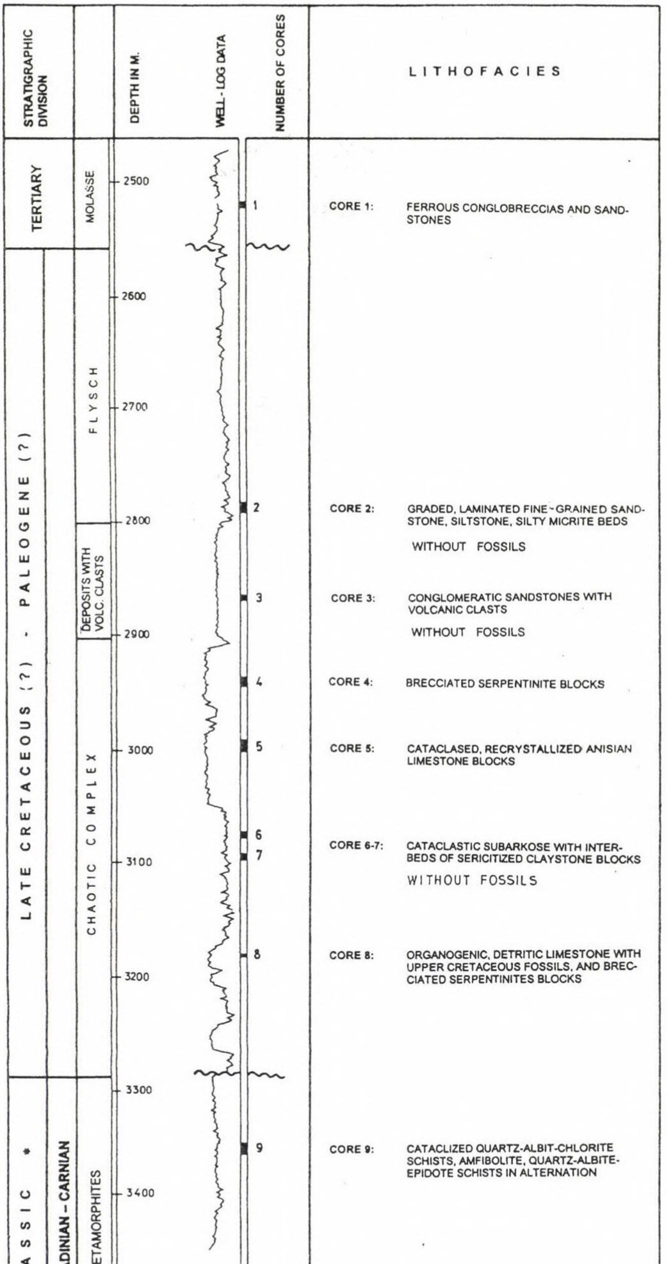

Fig. 5
Lower section of the stratigraphic column in Well 3, * possibly also part of chaotic complex

Cup., *Classopollis meyeriana* (Klaus) de Jersey, *Deltoidospora* sp. and others (Pantic 1982). These deposits commonly consist of fine-grained clastic material, less frequently of chert with rare pelagic microfossils: radiolarians, globochaets and spirillinas. These deposits were partially subjected to low-grade metamorphism.

Canovic and Kemenci (1989) reported that the well near Rusanda (No. 26;. Fig. 4) drilled below the Senonian through some 400 m of the metasediment package, including sericite phyllite, metasandstone, thin carbonatized sericite schist and calcschist. From the last cored interval biotite and amphibolite-riebeckite schist samples were recovered. The preliminary palynologic investigations by Dulic (1988) indicated a Jurassic age of the rocks.

Most wells penetrated several tens of meters into the Dogger–Malm formations where the drilling was completed. The Dogger–Malm formations generally underlie the Tertiary except in well 26 (Fig. 4), where they are below the Senonian.

Ophiolite Complex

The Mesozoic basinal series, particularly the Late Jurassic and Neocomian ones, are commonly tectonically associated with the rocks of the Ophiolite Complex (serpentinite, gabbro, diabase, spilite, radiolarite, etc.). The peridotite–pyroxenite rocks show the largest extent, mostly occurring beneath the Upper Cretaceous in the Banat, the Bačka and the Srem (Fig. 6). The actual thickness of the ophiolite members in the wells has not been assessed. The wells generally drilled only ten to several tens of meters into the ophiolite assemblage. According to the interpretation of geomagnetic data (Starčević 1991) the ophiolite is particularly thick along the Danube in the southern Bačka. Among the samples from the Banat serpentinized harzburgite was observed; only in the eastern Banat was serpentinized lerzolite also found. The wells in the Srem and the Bačka generally penetrated into dunite or homogenous serpentinite (Fig. 6)

Almost all samples of the peridotite rocks encountered in the wells in the basement beneath the Tertiary of the Vojvodina are highly brecciated. Their alternation along the cataclastic fissures was due to hydrothermal and metasomatic processes reflected in the replacement of serpentinite by carbonate. The carbonatization of serpentinite is often complete and the type of original peridotite cannot be distinguished. In addition to carbonatization the original rocks were subjected to silicification and argillization.

Gabbroid rocks were found in the Neogene base in the southeastern Banat along a major tectonic dislocation, extending southward approximately from the Romanian border in the east to the Danube (Fig. 6). Several wells in that area drilled through coarse gabbro. Gabbro occurs in the association with serpentinized lherzolite and amphibolite or shows transition to diorite.

These rocks were subjected to subsequent dynamic-metamorphic changes. The recovered samples are cataclastic, some mylonitic. Diabase and spilite, occasionally amygdaloid, were also observed in the southeastern Banat together

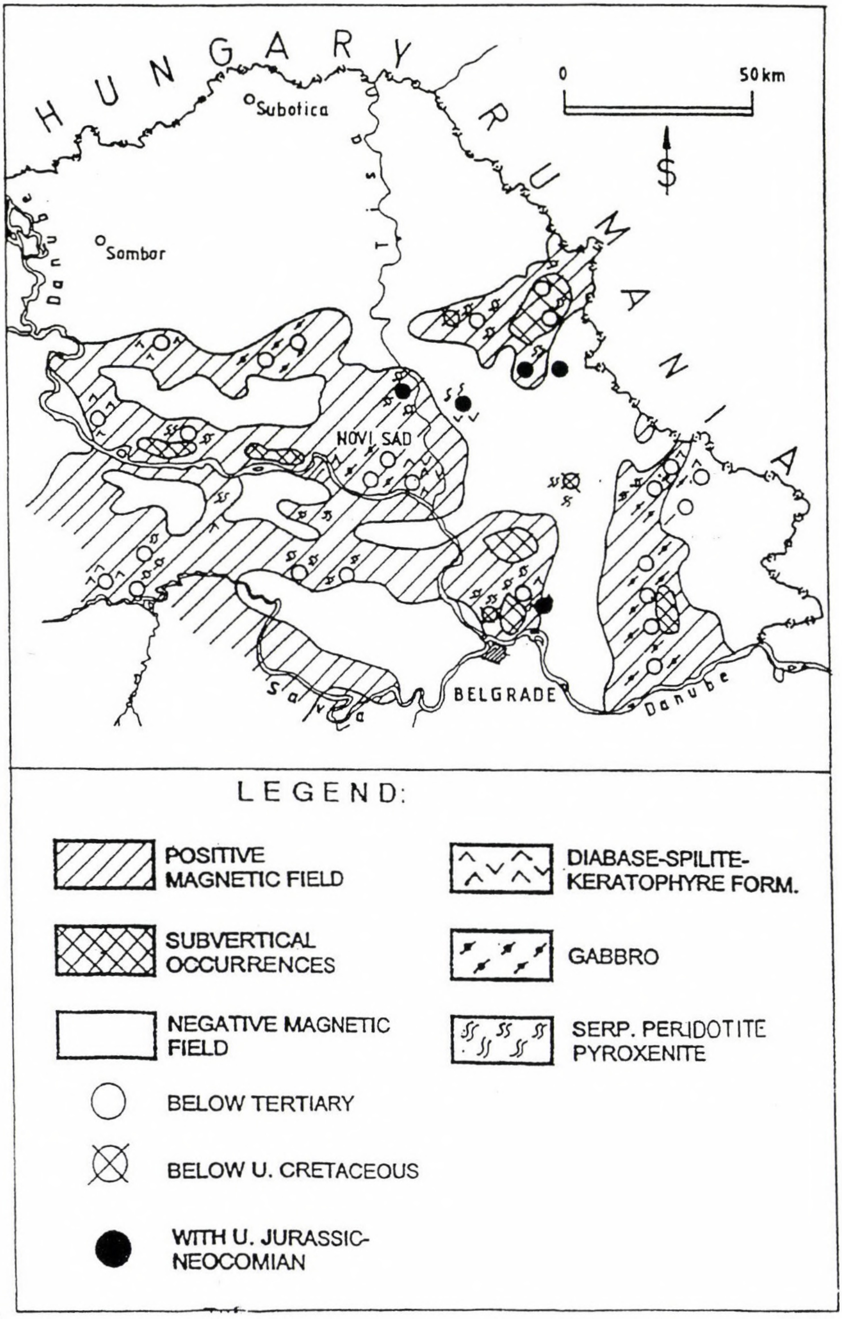

Fig. 6
Distribution of ophiolite rocks in the basement of the Pannonian Basin in the Vojvodina, (according to geophysical and well data). From Starčević (1991). Modified by Kemenci

with gabbro. Reworked, often well-rounded fragments of these rocks abound in the Upper Cretaceous clastites, indirectly indicating their earlier origin.

Evaporite formation

Beneath the Upper Jurassic and Neocomian pelagic developments of the basinal facies 75 m of evaporites were drilled in a well at Kraišnik (No. 8 at Fig. 4), Central Banat, near the Romanian border, before drilling was discontinued at 4575 m. (Fig. 7). The age of the evaporite formation has not yet been established. It is described here according to its position in the column of well 8 (Fig. 7). The evaporite formation contains breccia composed of one to ten cm-sized fragments of sericitized claystone, fine-grained sandstone and dolomite cemented by anhydrite, and pure, crystalline anhydrite with dolomite laminae.

Upper Jurassic – Neocomian

Well data from the central Banat and the southern Backa indicate the presence of Tithonian, Neocomian–Berrisian, Valangianian and Hauterivian beds (Čanović 1970–1991). The formations deposited in a basinal environment consist of marly limestone, marlstone, siltstone and chert. The microfossil assemblage includes *Calpionella alpina* Lor., radiolarian, aptichus and micromollusc remnants.

Relying on palynomorph evidence Pantić (1982) placed the age of the garnet-bearing sericite phyllite sampled from a well at Mramorak (No. 16; Fig. 4) in the southern Banat into the Jurassic to Neocomian range

The Upper Jurassic – Neocomian deposits in the central and southern Banat and the southern Bačka wells were found to underlie different stratigraphic units of the Tertiary, Upper Cretaceous and late Lower Cretaceous. The greatest thickness of approximately 960 m was drilled at Kraišnik (No. 8) in the central Banat (Fig. 7). The wells generally reach their total depths in these highly tectonized formations.

Late Neocomian (Valanginian, Hauterivian) and Barremian pelagic and deep-water series were established in several wells in the southern Banat. These series are either very poor in pelagic fossils (stomiosphera, cadosina, tintinnid, radiolarian, globochaeta, and hedbergella remnants) or completely sterile. The lithologic composition includes pelites, fine-grained clastites, chert, radiolarite but also some calcarenite and interserial breccia. Apparently redeposited, shallow-water fossil material was observed in some samples.

Late Lower Cretaceous

Čanović (1970–1990) identified Barremian and Aptian, shallow-water deposits in several central Banat and southern Bačka wells (Fig. 1).

These deposits of Urgonian-type facies contain coral-algal-bryozoan limestone and siltstone rich in different fossil associations (corallinaceans,

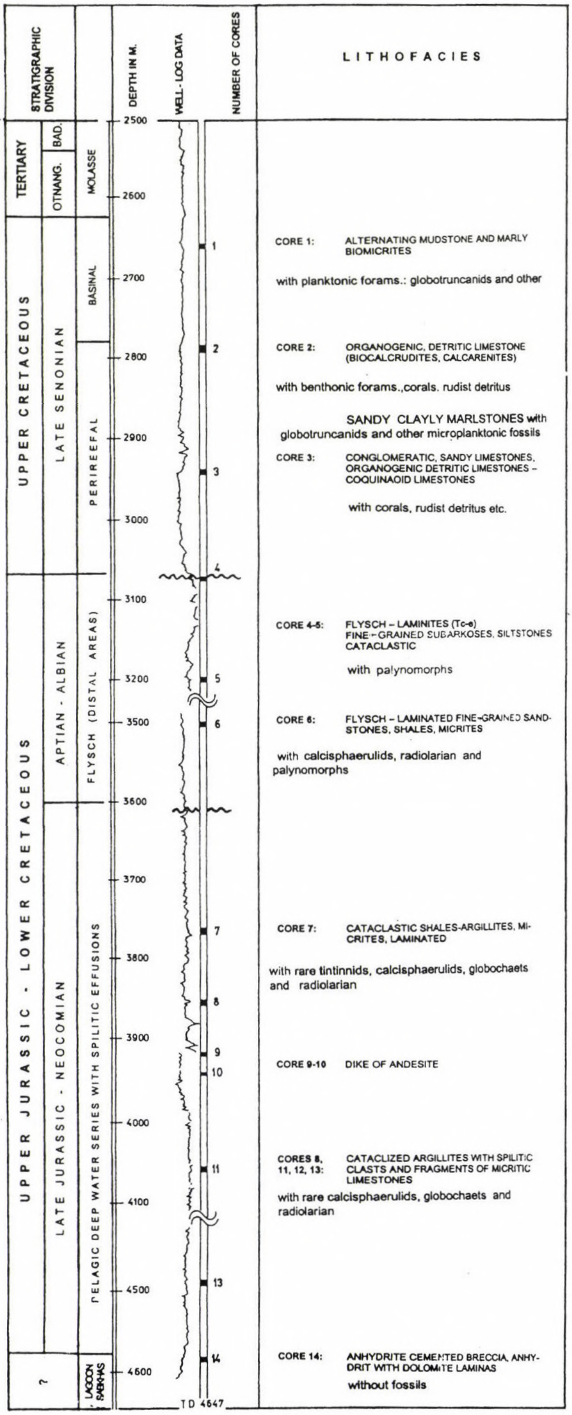

Fig. 7
Lower section of the stratigraphic column in Well 8

udatoceans, dasycladaceans, bryozoans and mollusks, Orbitolinida and other foraminifera).

The Aptian–Albian age of the beds developed in flysch facies was proved by Dulić (1989) who distinguished *Apendicisporites* sp., *Cicatricosporites* sp., *Corniculatisporites* sp. and *Tricolpites* sp. in the palynomorph association of the samples recovered from Well 8 at Krajišnik in the central Banat (Fig. 7). The lithologic composition of these beds includes fine-grained sandstone and laminated siltstone.

The Albian and Cenomanian deposits were proved by paleontological evidence (microfossils and palynomorphs) recovered from a well at Šurjan (No. 29; Fig. 4), eastern Banat. The lithologic sequence observed in these formations from the base upward contains black, silty claystone, micrite and conglomerate. The clay segment was found to contain pelagic microfossils (*Planomalina* sp., *Calcisphaerulla innominata* (Bon.). Pantić and Dulić (Pantić et al. 1991) associate the identified palynomorphs (*Apendicisporites matesovae* Bolk., *A. tricispidatus* Wey. et. Gr., *Atlantispollis* sp., *Alisporites grandis* Cook, *Liliacidites* sp., Tenerina sp.) with the Cenomanian. In the matrix of the conglomerate sequence, red algae (*Archaeolithothamnium rude* Lem., *Agardhiellopsis cretacea* Lem., *Paraphyllum amphiroeforme* Roth.), pachyodon and echinoderm skeleton detritus was found. Čanović (Pantić et al. 1991) determines these conglomerates as Albian or Cenomanian. The geologic cross-section of the Šurjan well shows Senonian formations both in the base and on top of the Albian–Cenomanian sequences.

Upper Cretaceous

The Upper Cretaceous is extensively developed at the base of the Tertiary. Upper Cretaceous deposits were encountered in over a hundred wells in central and southern Banat and southern Bačka (Fig. 1). Approximately 90% of these wells never cut across more than several tens of meters of Upper Cretaceous deposits. Some wells drilled a few hundred meters of Upper Cretaceous, commonly Senonian, formations. Six wells penetrated the base of the Upper Cretaceous. In the central Banat the subcrop of the Upper Cretaceous was found to be Lower Cretaceous; however, some wells drilled into crystalline schist below the Upper Cretaceous, near the limit of the Tisza Mega-unit.

Microfossil evidence (planktonic foraminifera, calcispherulids, rare benthonic foraminifera, palynomorphs and nannofossils) helped identify the Upper Turonian – Lower Senonian, Santonian, Campanian – Maastrichtian and Maastrichtian deposits. The paleontological studies were performed by Wicher and Obradovic (1952) and Čanović (1970–1990). The palynologic analyses of the Upper Cretaceous were carried out by Šećerov and Pantić (1975) and Dulić (1988, 1989); the nannofossils were determined by Mihajlović (1980–1991). The sedimentological and petrographic data were investigated and interpreted by Kemenci (1961–1994).

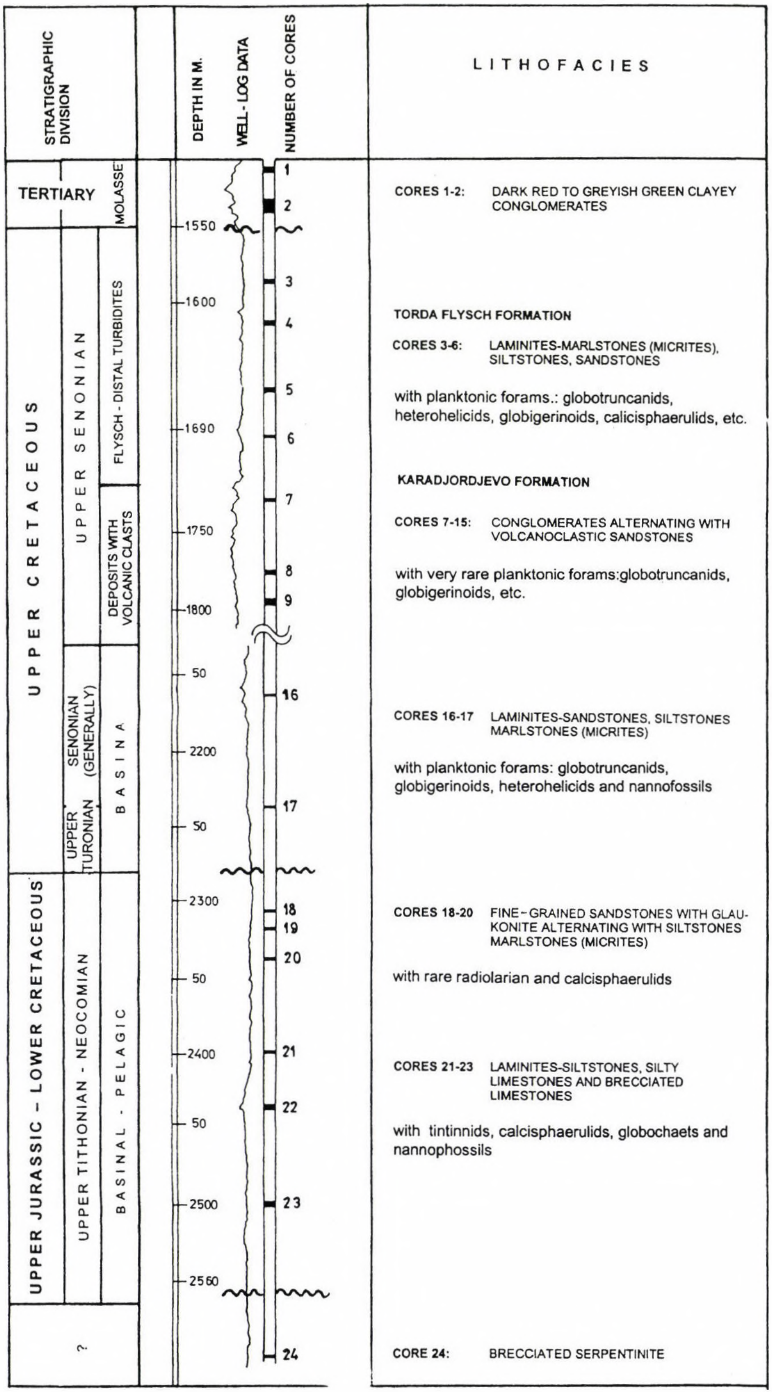

Fig. 8
Lower section of the stratigraphic column in Well 7

From the point of view of sedimentology the Upper Cretaceous can be subdivided into the following informal lithostratigraphic units (Fig. 8):

Clastic carbonate deposits

Organogenic, detritic limestone rich in rudist detritus ranging from the Upper Turonian to the Lower Senonian, and pelites with Senonian planktonic foraminifera;

The Karadjordjevo formation

Made up of Senonian basinal deposits bearing andesite clasts.

The Torda flysch formation

Built up of Senonian turbidity current deposits.

The Upper Cretaceous clastic rock samples from the wells drilled into the Karadjordjevo Formation in the southern Backa and the central and southern Banat contain a particularly large volume of andesite, but also trachyte and dacite clasts. Other than these volcanic fragments, the sandstone and conglomerate beds contain variable volumes of granitoid, crystalline schist, Lower Triassic sandstone and limestone, Upper Jurassic and Lower Cretaceous limestone, chert, diabase, spilite and argillaceous schist fragments. Obviously, the site was severely affected by volcanic activity in the Upper Cretaceous; however, there are no reliable data on its outset and duration. The available well data indicate that the Karadjordjevo Formation underlies the turbidites of the Torda Formation (Fig. 8).

The studies of well data show the existence of an Upper Cretaceous (Senonian) trough, the upper levels of which were filled up by turbidity currents (Kemenci and Čanović 1987). This turbidite belt, 15–20 km wide on average, trends from the west southward in southern Bačka and from southwest northeastward across the central Banat. The northern boundary of this unit is made up of crystalline schist, granitoid and Triassic deposits of the Tisza Mega-unit. Ophiolites of the southern Bačka and central Banat underlie the trough in the south (Fig. 1).

Turbidites developed in the proximal domain were encountered only in the central Banat but those of the distal zone both in the central Banat and the southern Backa. Laminites frequently interbedded with thin pelitic beds that contain pelagic microfossils were also observed.

Upper Cretaceous – Paleogene (?)

Clastic series with turbidity features, 350 to 600 m thick, were encountered in the well at Kumane (No. 30) in the central Banat, at Temerin (No. 4) and Backi Petrovac (No. 3) in the southern Bačka (Fig. 4). These series underlie Tertiary (Miocene) formations. As the lithofacies of these series differ in some respects from the Senonian flysch in the Torda Formation they are distinguished as the Temerin Formation. It is possible that they were formed in the Late Upper Cretaceous depositional cycle as the uppermost levels of the Torda

Formation, but also that they were deposited during the post-Maastrichtian, in the Paleogene.

The cross-section of well 3 (Fig. 5) shows that this series is underlain by deposits similar to the Karadjordjevo Formation. These deposits with volcanoclasts are in turn underlain by a chaotic complex of blocks (Fig. 5). The origin and age of the blocks are different (cataclized serpentine, Anisian limestone, Triassic low-grade metamorphites, Senonian organogenic reef limestone with dark gray subarkosic matrix. Their size is expressed in terms of tens of meters.

Litho-structural units in the southern Vojvodina and correlation with the Vardar Zone in Serbia

The subjects of this article are the Ophiolitic and Flysch belts in the southern Vojvodina. These litho-structural units, according to their stratigraphic and facial characteristics and magmatism, can be correlated with similar units in the Vardar Zone to the south in Serbia, according to the Basic Geologic Maps of the SFRJ at 1:500 000 scale, from the Federal Geologic Institute (1985).

In the most recent papers by Dimitrijević (1995), Karamata and Krstić (1996), and Karamata et al (1998), the Vardar Zone is defined as a Composite Terrane, with separate zones and the Jadar Block.

Prior to analyzing the presented data three representative cross-sections of southern Vojvodina are given in Fig. 9. On the cross-sections three morphological regions of the Pre-Tertiary basement can be distinguished. The eastern units stretch northeastward and consist of a Neogene depression deepening northward in the Banat and a ridge along the Danube and Tisza Rivers. The western part embraces the W–E -extending Neogene depression in the southern Bačka and Srem, and the ridges of the Fruška Gora and south of the Sava River. This morphology at the basement is connected to the geology of the basement.

The basement morphology, as well as geologic units encountered in the wells, represent the continuation of the geologic framework south of the Danube and the Sava Rivers (Fig. 2): The Serbo-Macedonian Massif, the eastern, main ophiolitic belt of the Vardar Zone, the continuation up to Belgrade of the Kopaonik Block as a ridge unit, the western ophiolite belt of the Vardar Zone and the Jadar Block.

In the east, along the margin (with metamorphites) of Vršac, analogous to the geology in the Central Serbia, exists a zone of gabbro and diabase. To the west is a belt composed of serpentinite, rare Liassic and Upper Jurassic–Neocomian deposits, and widespread Lower Cretaceous basinal sediments. Low-grade metamorphosed Dogger–Malm sediments occur with the serpentinite in the west (Elemir Formation). It should be noted that similar, low-grade metamorphosed Jurassic–Neocomian deposits were found in Central Serbia. In the main, central parts of the Banat they are covered by

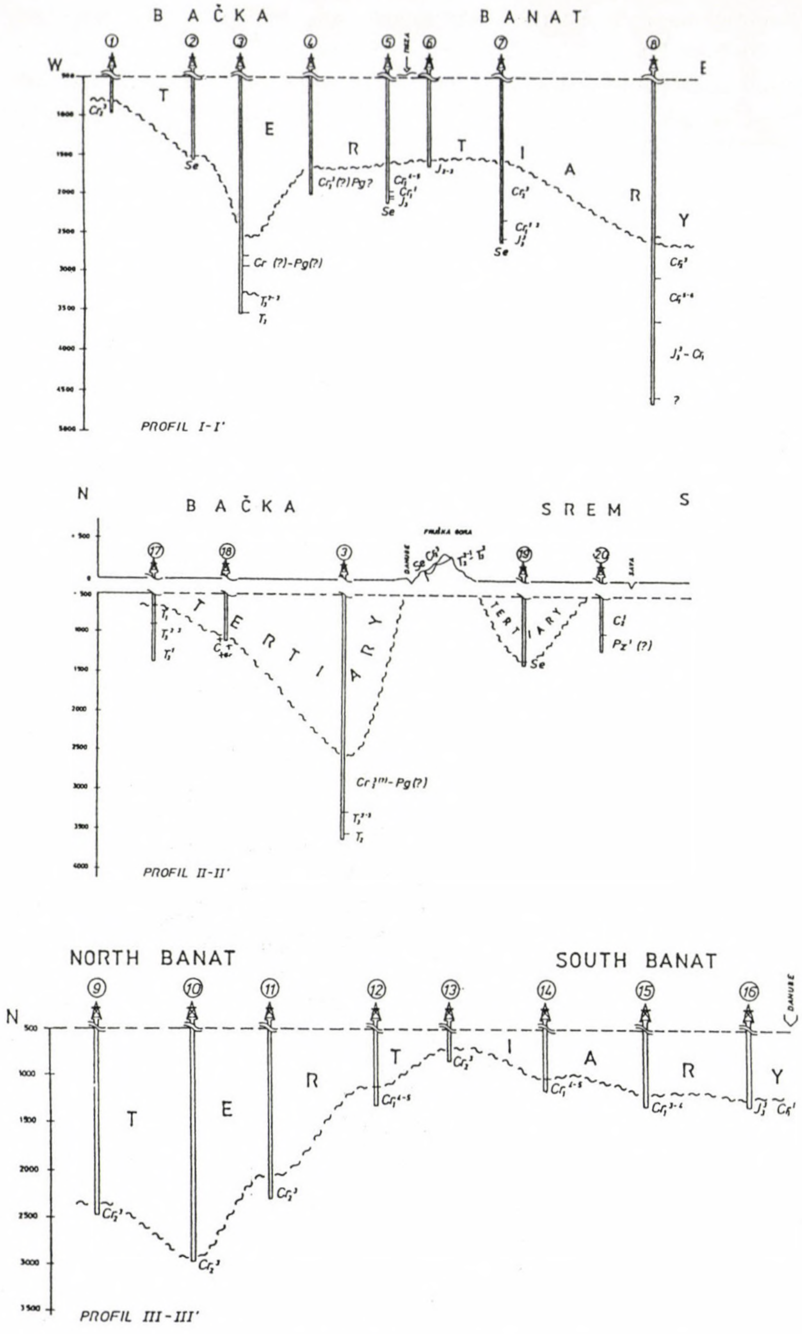

Fig. 9
Burial depths of the Mesozoic in the Pre-Tertiary basement in the Vojvodina

Tithonian–Neocomian basinal sediments as the first overstep sequence, followed by Late Neocomian–Barremian pelagic sediments, themselves covered by Aptian and Albian (partly flysch) deposits. In this area shallow-water deposits of Urgonian type occur. This part of the Banat is identical in geology and distribution of units to the main belt of the Vardar Zone.

Along the Danube and Tisza Rivers, i.e. in the middle of the southern Vojvodina, to the west of the previously described units, a ridge (Fig. 9) stretches as a continuation of the Kosmaj–Avala Ridge to the south. There are no data for the Pre-Tertiary evolution of the southern part of this unit; only its margin at the east is defined by serpentinite NE of Belgrade, and the western one by diabase and gabbro in the area of the southeastern Bačka (Fig. 1). Only northeastward from the outcrop at Krčedin were Tithonian–Neocomian basinal deposits found. The basement, proved only in the Banat, is the previously-mentioned serpentinite. In southeastern area of these deposits they are overlain by Lower Cretaceous sediments, similar to ones of the same age near Belgrade.

To the west in the southern Bačka, and in the Srem with the Fruška Gora, the geology of the Pre-Tertiary basement is different and very complex. This area consists of unmetamorphosed and metamorphosed Triassic sediments in greenschist facies, Jurassic (?) anchimetamorphosed basinal facies (DCF), serpentinite bodies, diabase and gabbro blocks as well as Senonian shallow-water limestone. The complexity of the geologic setting of this region confirms the data from the Fruška Gora, and from the column of Well 3 near Bački Petrovac (Fig. 5). The geologic setting of this formation, according Karamata (1999, oral comm.), points to a trench.

The area of the Fruška Gora, according to Grubić et al. (1998), judging by the type of deposits and the orientation of folds as well as the mutual relations, form a packet of obduction thrusts of northern vergency. These overthrusts cover the old basement and Triassic deposits, which are uncovered west of the monastery of Jazak.

The first unit covering all three previously distinct units is the Karadjordjevo Formation of Senonian age. This unit was encountered in the southern Bačka and the central Banat. The Karadjordjevo Formation is rich in andesitic and associated volcanics, similar to subduction-related ones. This unit shows an analogy with the upper levels of the Turonian (?) – Senonian andesite-bearing sediments south of Belgrade, and with identical formations in the western parts of the south Apusenides and the Mures Trough described by Lupu (1991).

Finally, the Maastrichtian Torda Formation overlies the former units in the form of a belt 15 to 25 km wide, extending from the Danube in the west to the Romanian border in the east. To the north its boundary is the transcurrent fault system to the south of the Tisza Mega-unit. The Torda Formation is a flysch, deposited in a trough of unknown primary extent, since it is cut off to the north and the west by Tertiary transcurrent movements.

Final Remarks

In spite of limited drilling through the relevant series, the scarcity of core material and their poor paleontological content, the analyses of the core samples recovered from the wells which penetrated the Paleozoic and Mesozoic formations in the Pre-Tertiary basement in the Vojvodina did reveal some very important evidence for a more positive comprehension of the geologic evolution (primarily stratigraphic, but also paleogeographic and to some extent geodynamic) of the study area during the Paleozoic and Mesozoic, and its relation to adjacent areas in Yugoslavia and neighboring countries.

Acknowledgements

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Radiometric dating of Tertiary glauconite-bearing formations in the northeastern part of the Lublin Upland, Poland

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This paper presents new stratigraphic data concerning glauconite-bearing Tertiary formations in the northeastern part of the Lublin Upland (Chełm Hills area) based on radiometric dating of glauconitic minerals. Until now the analyzed glauconitic sands were regarded to be of Lower Sarmatian (samples G1, G2 and G3) and Lower Oligocene (sample G4) age. Because of the absence of fossils it was not possible to date them paleontologically. Therefore, hitherto the correct stratigraphic position of the sand deposits in that area was not established. The tested glauconites, due to a relatively high potassium content, showed a very high degree of maturation and could be dated with the potassium-argon method.

For estimating the quality of glauconites as geochronometers their physical features (color, morphology, grain size, and specific density) as well as the chemical-structural ones (chemical composition, structure type, and polytypic features) have been determined.

Radiometric dating showed an Eocene age for the glauconites and therefore the same age of three sampled deposits (samples G2, G3 and G4). On the other hand, the radiometric age of glauconite from one sample (G1) proved its allogenic character in the Sarmatian sediments, redeposited from older (Eocene) deposits.

Key words: glauconitic minerals, radiometric dating, Tertiary

Introduction

Tertiary deposits of the Lublin region, especially in the Chełm Hills area, are rich in glauconitic minerals; some of them are even called glauconitites (glauconitic sandstones). They show a great variety in facies and are poor in fossils. For this reason it was a problem to identify their stratigraphic levels. Moreover, the occurrence of glauconitic deposits in the form of isolated layers leads to further difficulties in their correlation. Until now their dating was mainly based on geologic-morphological circumstances, petrographic likeness, and on the analogy of deposits in comparison with respective layers located in other regions. As there are no fossils, adequate criteria to establish their stratigraphic attachment could not be found.

Until now sands of the Czułczyce, Chełm, and Janów localities were regarded to be of Lower Sarmatian age (Kowalewski 1924; Turnau-Morawska 1949; Prószyski 1952; Morawski 1957; Harasimiuk and Rutkowski 1972; Henkiel 1983)

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and the Lechówka locality of Lower Oligocene age (Szelągowska-Skrzypczak 1971), but the correct stratigraphic position of these deposits has not yet been resolved.

Recently, more and more attention has been paid to the problem of the geologic significance of data obtained by means of isotopic geochronology. Among the minerals of sedimentary rocks that can be radiometrically dated, glauconite seems to be one of the greatest significance because it contains potassium in an amount that permits isotopic age determination.

Isotopic chronometry is a powerful tool in geologic investigations but until now it was not sufficiently used in Poland. The author assumed that the Tertiary, paleontologically barren, clastic sediments of the Chełm Hills area could be dated by means of radiometric determination of the absolute age of the glauconite occurring in those deposits.

The application of radiometric dating to glauconite, despite its great diagnostic advantages, should be treated cautiously, because the glauconite grains evolved, matured, and then may have come under the influence of various post-depositional processes that can cause under – or overestimation of its isotopic age.

The specific methods for estimating the quality of glauconite as a good chronometer were defined by Odin (1975). The mineralogical quality of glauconite as a chronometer can be estimated by means of X-ray diffraction and thermal analysis. The chemical quality of glauconite depends on the amount of potassium. According to Odin (1982) the K_2O content must be higher than 6.5%.

The measured age of glauconite is often different from the age generally admitted for the formation studied. Geochemistry and mineralogy of glauconite help in understanding the isotopic dating and how this chronometer works in sedimentary rocks. Only glauconites in which the expandable layers of smectite-type are absent are suitable for radiometric dating (Odin 1982). The presence of smectite material in the glauconite structure is responsible for the lower K-content. The loss of radiogenic argon from post-diagenetic glauconites and the ratio of radiometric to geologic ages are more or less constant. For this reason it is necessary to know the genetic character of the glauconite in the deposits (authigenic, allogenic), its alteration, chemical composition and internal structure. Therefore, the tested glauconites were characterized in respect to their physical and chemical-structural features. To accomplish this thermal, X-ray, and chemical analyses were performed. Scanning electron micrographs of selected grains of glauconite were also produced.

General conditions of glauconite formation

In the Tertiary period favorable conditions for the formation of glauconite prevailed (McRae 1972; Logvinenko 1980; Nikolaeva 1980; Odin and Matter 1981). Such conditions as transgressive cycles of sedimentary basins and slow rates of sedimentation promoted glauconitization processes. Among Tertiary glauconite-bearing formations Paleogene and especially Eocene formations are globally the most extensive (Hein et al. 1974; Logvinenko 1976; Nikolaeva 1981). Paleogene glauconite occurs practically all over the world on platforms as well as in geosynclines.

In the Tertiary period favorable paleogeographic–environmental conditions for glauconite formation also occurred within the territory of Poland. In the Lublin region, glauconite-bearing sediments of that period are known from the Lower Paleocene, Upper Eocene, and Lower Oligocene. In the author's opinion (Krzowski 1998) the glauconite of the Miocene period is of an allochthonous character and was redeposited from older sediments, especially from the Eocene ones.

In the Upper Eocene particularly favorable conditions for glauconite formation occurred, that is: transgressive stratigraphic position, low rate of sedimentation, supply of terrigenous material from the land and a warm climate. During that period a climatic optimum of global character occurred (Ahmetev 1976; Dyjor and Sadowska 1986). In the Eocene period the climate in Europe was warm and wet, of tropical type, with a mean annual temperature of about +22 °C (Dyjor and Sadowska 1986; Głazek and Szynkiewicz 1987). On the territory of Poland the climate was tropical, as well. The studies of paleotemperatures based on the analysis of stable isotopes of oxygen and carbon of fossils (Krzowski 1995) confirmed the Upper Eocene climatic optimum (+23.8 °C) as well as the open marine character of the sedimentary basin.

The sandy facies of sediments and diagnostic features of glauconite included in them show that the Eocene deposits in the Lublin region consist of shallow-water, littoral marine sediments (Krzowski 1998). These sediments are fine-grained, gray-green in color, frequently muddy, and weakly marly quartzose-glauconitic sands with well-rounded grains of gravels and locally of lydites, silty-sandy muds and claystones, and sometimes gray-green and olive-colored clays. The sediments frequently contain numerous, small-sized phosphorite concretions and amber fragments.

Poor sorting of the glauconitic sands and the lack of a distinct density selection of glauconite indicate a low-energy sedimentary environment and weak bottom currents. The variable heavy mineral composition (Roztocze Area) in the sediments suggests different source areas and different sedimentological conditions within the Eocene basin (Buraczynski and Krzowski 1994).

Epeirogenetic movements of the Laramide phase (Alpine Orogeny) had a decisive effect on the type and kind of sedimentation and consequently on glauconite formation during the Eocene (Henkiel 1984). Secondary alterations of glauconite confirm the existence of a weathering zone in the Eocene deposits which developed after their exposure in later geologic periods, especially in the Miocene.

Geologic setting

The study area comprises the NE part of the Lublin Upland called the Chełm Hills (Fig. 1). Tertiary deposits of this region are very diversified with regard to facies. They begin with Paleocene (Danian-Montian) deposits that lie on the corroded, hardened surface of Upper Maastrichtian age (hard-ground). Paleogene and Neogene deposits occur on the surface or beneath a very thin cover of Quaternary deposits. They consist of sands, sandstones, chalky marls and gaizes[1]. The sands are well-sorted, fine and medium grained, composed chiefly of quartz and glauconite. Feldspar, muscovite, iron oxides, and heavy minerals are present in minor amounts.

The Tertiary glauconite-bearing deposits of the Chełm Hills area have interested geologists for many years but only in recent times have provided increased development of knowledge about glauconite and its applications to the interpretation of the sedimentary environment (Gazda et al. 1992; Krzowski 1998), to the stratigraphic correlation of sediments (Hałas et al. 1990; Krzowski 1980, 1991, 1993), and to the analysis of stratigraphic-paleogeographic problems (Buraczynski and Krzowski 1994).

Glauconite in the western part of the Lublin Upland, with regard to its mineralogy, was studied by Wiewióra and Łącka (1985).

Samples were collected from the quartzose-glauconitic sands and clayey sands with glauconite exposed in the outcrops of Czułczyce (G1), Chełm (G2), and Lechówka (G4) localities and of the Janów stratigraphic borehole (G3). Three of the above-mentioned samples (G2=CH1, G3=JA2, G4=LE1) were described earlier from the aspect of paleoenvironmental studies (Krzowski 1998).

Experimental results

Glauconite pellets were electromagnetically separated from the glauconitic sand samples by means of a Frantz electromagnetic separator and, when necessary, were further purified by hand-picking under binocular microscope. In order to minimize mineralogical and chemical heterogeneity of glauconite it was subjected to granulometric analysis and to density separation in a solution of bromoform and ethyl alcohol.

In order to determine the mineralogy of the glauconite the following investigations were made: X-ray diffraction (XRD), differential thermal analysis (DTA), chemical analysis and scanning electron microscopy (SEM). The highly evolved glauconites sometimes show a thin external film covering the grains. This film is composed of less evolved glauconitic minerals, which according to Lamboy and Odin (1975) is easily removed by ultrasonic treatment. For this

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

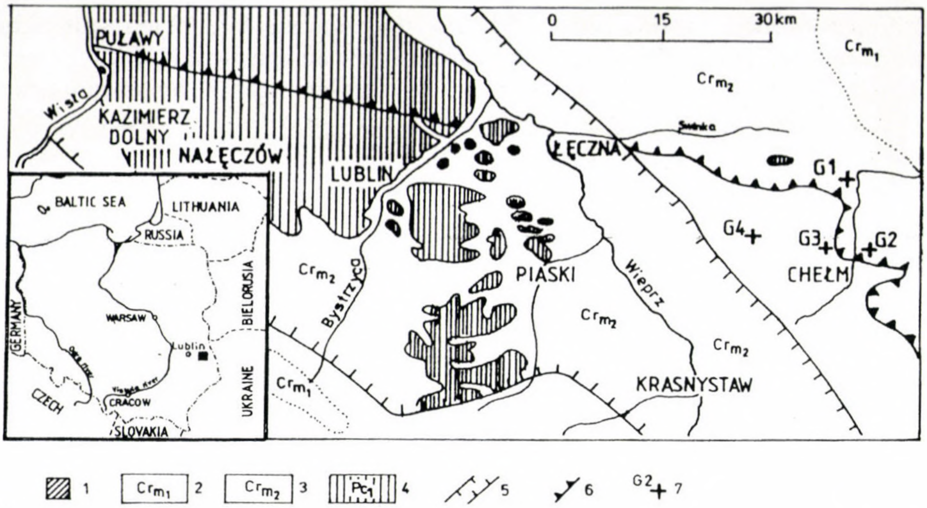

Fig. 1
Geologic sketch map (without the post-Paleocene cover) of the northern Lublin Upland (after M. Harasimiuk 1984) with locations of tested glauconites. 1. study area; 2–3. Upper Cretaceous; 4. Paleocene (undivided); 5. marginal dislocation of the Paleozoic Mazovian–Lublin Graben; 6. northern edge of the Lublin Upland; 7. sampling sites

reason the samples of glauconite for X-ray diffraction analysis were disaggregated by means of an ultrasonic disintegrator. The fraction size smaller than 2 μm resulting from this procedure was suspended in distilled water and separated through sedimentation, according to Stokes' Law.

Analysis of physical features of glauconite

The physical features of glauconite such as color and morphology of grains, their size and specific density have much geologic and mineralogical significance and are useful for the interpretation of various geologic problems, e.g. sedimentological environment of glauconite-containing deposits, genesis of glauconite, post-depositional processes, and others. The analysis of abundance, color and morphological variability of glauconite pellets provides valuable information about the general conditions of the sedimentological environment (Krzowski 1998). With respect to grain size Odin (1982) stated that coarser glauconite (0.15–0.50 mm in diameter) is generally more evolved than finer one. The glauconite occurring in the form of internal molds of foraminifera would indicate that these glauconite pellets might have been formed in an open marine environment.

Previous studies concerning the radiometric ages of different density fractions of glauconite (Halas et al. 1990) showed that not all density fractions of glauconite yield a reliable radiometric age. According to the above-mentioned authors only medium (the most abundant) fractions yield reliable results. The light density fractions (2.3–2.4 g/cm^3) yield too low an age; heavier (>2.6 g/cm^3) fractions behave in a similar fashion, but to a lesser extent. This is connected to their mineralogical (expanding layers content and type of ordering) and chemical (K_2O content) composition.

Color and morphology tests of glauconite grains were performed under binocular microscope. Grain size was determined by means of sieve analysis; specific density tests were performed using a bromoform and ethyl alcohol solution.

Color and grain morphology were variable (Table 1). Glauconite grains were green, light green, dark green, black, brown, honey yellow, rusty, beige, and celadon. Black-colored glauconite is considered to be mature and well crystallized (Ehlmann et al. 1963); on the other hand, beige and brown ones represent evidence of weathering processes.

Table 1
Physical features of glauconite

Sample No.	Percent of glauconite in the sample	Colour	Pellets morphology*	Density (g/cm^3)
G1	6.3	light green green brown honey yellow black	spheroidal ovoidal mammillated tabular discoidal ellipsoidal	2.2 - 2.6
G2	10.5	light green green dark green black rusty	tabular lobate spheroidal ovoidal	2.3 - 2.6
G3	17.8	light green green black beige celadon	spheroidal ovoidal mammillated tabular capsule internal moulds	2.2 - 2.6
G4	14.0	light green green black light brown	spheroidal ovoidal tabular capsule mammillated internal moulds	2.5 - 2.6

* after Triplehorn's classification (1966)

Glauconite grain morphology was differentiated as well. A large variety of shapes could be found: spheroidal, ellipsoidal, ovoidal, mammillated, tabular, discoidal, lobate, capsule and as internal molds of foraminifera. Variation in grain size could also be observed: from Ø 0.50 mm to Ø < 0.071 mm (Table 2). None of the grain size fractions dominated in any of the tested samples. A different grain size fraction was dominant in every sample. In the tested samples two to five density fractions of glauconite occur; values between 2.5–2.6 g/cm^3 were dominant.

Table 2
Grain size of glauconite

Sample No.	Grain size (mm) (percent)						Total
	0.50-0.25	0.25-0.20	0.20-0.16	0.16-0.10	0.10-0.071	< 0.071	
G1	37.36	17.63	22.63	18.94	3.94	0.52	101.02
G2	8.20	14.50	41.91	23.52	8.56	3.28	99.97
G3	36.60	-	-	58.20	5.40	-	100.20
G4	-	0.78	6.61	41.43	42.65	8.40	99.87

Grain size of glauconite subjected to analytical (mineralogical, chemical and radiometric) studies

X-ray diffraction analysis (XRD)

The most useful and necessary study for aiding in the interpretation of glauconite ages is diffraction analysis. The quickest and most informative technique is the use of non-oriented powder.

In this work X-ray diffraction analysis was performed on mounts with non-oriented powder and oriented-untreated and glycol-solvated specimens. The oriented samples were realized by pipetting clay suspension on glass slides (10 mg/cm^2) and were glycolated by storing in glycol vapor overnight at the temperature of 60°C.

Glauconite was examined by X-ray diffraction using Cu-K_α radiation, on HZG-4 and DRON diffractometers and a goniometer scan speed of 1^0 2Θ/min. Powder diffractograms of the glauconite samples are presented in Fig. 2. The type of ordering of glauconites studied and the content of smectite type layers were defined using the Ir index (representing the intensity ratios of the 001 and 003 reflections from the air-dried and the glycolated samples) and the computer plots proposed by Środoń (1984). On the basis of X-ray diffraction patterns, three types of glauconitic minerals have been distinguished: 1.) ordered 1M polytype, 2.) disordered 1Md polytype and 3.) glauconitic mixture (Table 3). Glauconitic mixture may also be regarded as disordered glauconite (Odom 1984).

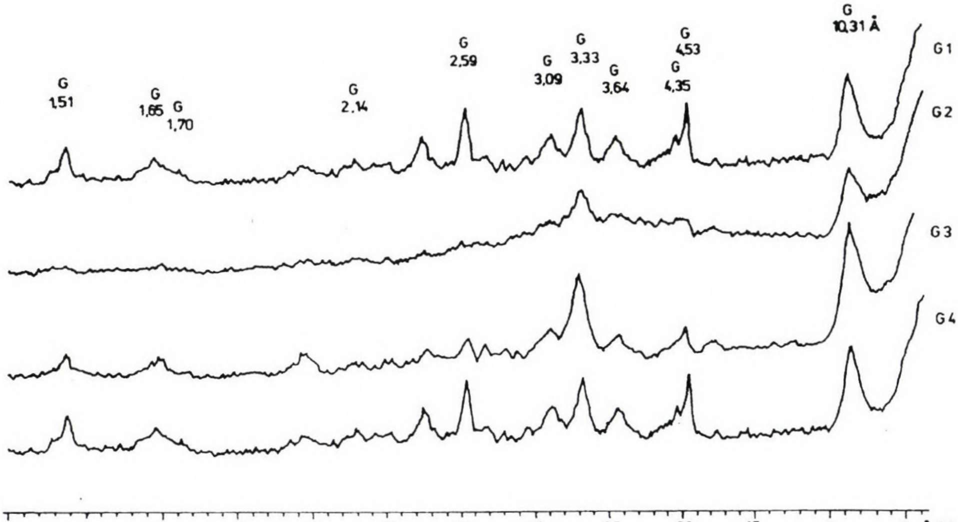

Fig. 2
X-ray diffractograms of unoriented powder specimens of glauconites

Generally, XRD patterns from samples studied reveal narrow 001 reflections of glauconite at 10Å. The intensity of this peak is weakly increased after ethylene glycol solvation and heating, which suggests a very low presence of expandable layers. The sharp and symmetric 020 and 130 and the well-developed 112 and $11\bar{2}$ reflections in samples G1 and G3 indicate an ordered structure. Sample G2 shows XRD patterns having slightly more asymmetric 10 Å peaks in which ethylene glycol solvation produced little change, but heat treatment increased the intensity of reflections and diminishes broadening. Furthermore, weak and broad peaks between 12 and 14 Å were observed, which disappear after heating and are found at 17 Å after ethylene glycol solvation. These data indicate the presence of material having layer spacings >10 Å. In this sample the small

Table 3
Mineralogical-structural features of glauconite

Sample No	Percent of smectite-type layers	Type of "illitic material"	Type of ordering	Polytype
G1	6	ISII	ordered	1 M
G2	8 - 15	IS + I	disordered	1 Md
G3	< 5	I/S	disordered	1 M
G4	> 10	I/S	ordered	1 Md

number of diffraction observed as well as the low intensity of 112 and $\bar{1}1\bar{2}$ peaks can be indicative of structural disorder.

Disordered glauconite was clearly evident in scanning electron microscopy (SEM) in all grains (pls I–IV). It was predominantly of the boxwork and rosette forms with clusters of larger, parallel to subparallel blades of lamellar structure. According to Odin and Matter (1981) the boxwork and lamellar structures of disordered glauconite are typical for contents of 6.5% up to 8% K_2O, respectively; K_2O values in the glauconites studied were, generally, greater than 6%.

Differential thermal analysis (DTA)

The differential thermal analysis of glauconite samples was performed on a Paulik–Paulik instrument.

Micas contain water of two different kinds: molecular and hydroxyl. In the differential thermal curves each kind of water is distinguished by independent endothermic effects. However, the differential thermal analysis of glauconite was mostly characterized by three major endothermic effects.

In the tested glauconites molecular water was eliminated in the temperature range of 100 to 200 °C and hydroxyl water in the temperature range between 545 and 560 °C (Fig. 3). The third endothermic effect occurred at a temperature of 980 °C and was caused by the destruction of the crystal lattice of the mineral and by the crystallization of a new mineral phase (spinel). This effect was not accompanied by weight loss. Total weight loss of the heated samples was up to 11%.

Thermal studies confirmed the mineralogical heterogeneity of glauconite. With the increase of expandable layers content the hydroxylation processes intensified.

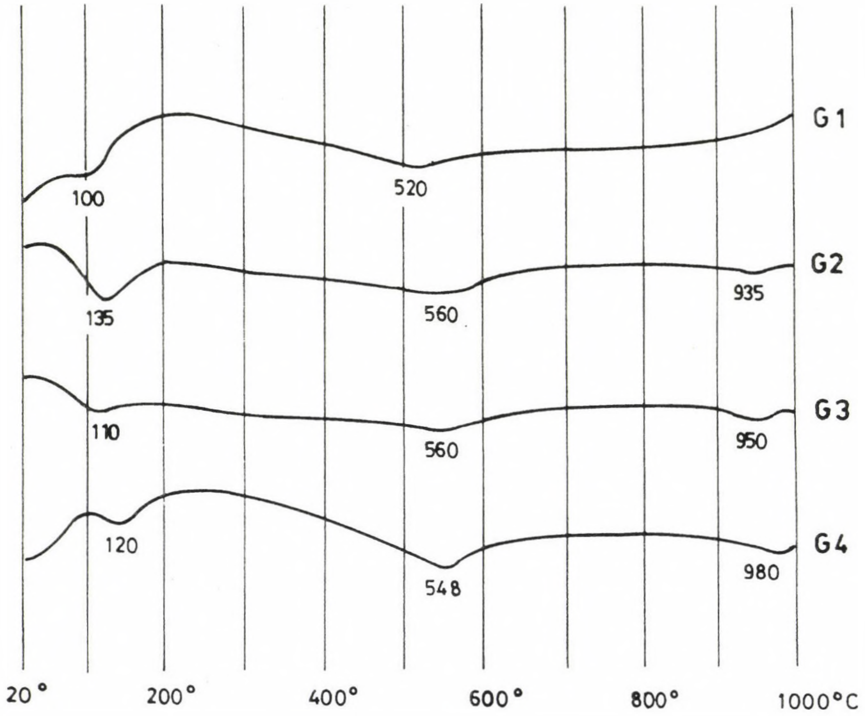

Fig. 3
Differential thermal analysis (DTA) curves of glauconites

Chemical analysis

The determination of chemical elements was performed on dried samples of glauconite at a temperature of 105 °C. The samples were dissolved in HF, HNO_3, and $HClO_4$. The ASA flame method was used; air-acetylene and nitrogen suboxide-acetylene types of flame were applied. Each sample was homogenized by means of rubbing in ethylene alcohol. Loss on ignition (LOI), S, SiO_2 were determined by weight, however TiO_2 with the DAM reagent, and FeO with ofenatroline. Remaining components were determined by the flame method with the application of a SP-9 spectrometer (Pye Unicam production).

The full chemical composition of glauconite is given in Table 4. The chemical composition of glauconite is variable and depends on its mineralogical composition i.e. on the percent contents of expandable layers.

Chemical analysis of glauconites made possible the calculation of the following crystallographic formulae:

Table 4
Chemical composition of glauconite

Component	Sample number			
	G1	G2	G3	G4
	weight percent	weight percent	weight percent	weight percent
SiO_2	50.05	55.50	40.90	46.60
Al_2O_3	9.58	9.43	7.76	14.40
Fe_2O_3 total	19.40	17.80	20.70	16.00
Fe_2O_3	18.20	16.50	19.60	15.10
FeO	1.15	1.17	1.02	0.80
MgO	3.77	3.41	3.00	3.11
K_2O	8.41	6.22	6.49	8.99
Na_2O	0.02	0.02	0.04	0.03
CaO	0.45	0.32	1.68	0.43
H_2O^-	4.56	0.34	1.28	5.23
H_2O^+	6.42	5.70	6.60	8.30
TiO_2	0.08	0.11	0.09	0.16
P_2O_5	0.09	0.08	1.20	0.10
S total	0.03	0.06	0.05	0.03
	0.021	0.031	0.032	0.031
Ba	0.002	0.003	0.004	0.005
LOI (600°C)	6.97	6.04	6.88	8.42
LOI (1000°C)	8.12	7.03	8.05	13.39
TOTAL	99.97	99.88	99.72	99.83
Total of tetrahedral cations	4.000	4.000	4.000	4.000
Total of octahedral cations	1.971	1.960	1.968	1.982
Total of interlayer cations	0.824	0.583	0,865	0.894

Glauconite from Czułczyce locality (G1)

$(K_{0.789}Na_{0.003}Ca_{0.0027})(Fe^{3+}_{1.005}Fe^{2+}_{0.071}Mg_{0.414}Al^{VI}_{0.502})[Si_{3.673}Al^{IV}_{0.327}O_{10}(OH)_2$

Glauconite from Chełm locality (G2)

$(K_{0.577}Na_{0.003}Ca_{0.17})(Fe^{3+}_{0.874}Fe^{2+}_{0.068}Mg_{0.359}Al^{VI}_{0.677})[Si_{3.896}Al^{IV}_{0.104}O_{10}(OH)_2]$

Glauconite from Janów locality (G3)

$(K_{0.625}Na_{0.083}Ca_{0.403})(Fe^{3+}_{0.665}Fe^{2+}_{0.121}Mg_{0.388}A^{lVI}_{0.599})[Si_{3.670}Al^{IV}_{0.330}O_{10}(OH)_2]$ [1]

Glauconite from Lechówka locality (G4)

$(K_{0.547}Na_{0.004}Ca_{0.025})(Fe^{3+}_{0.862}Fe^{2+}_{0.050}Mg_{0.351}Al^{VI}_{0.820})[Si_{3.534}Al^{IV}_{0.466}O_{10}(OH)_2]$

1 without apatite correction

Table 5
Radiometric age of glauconite

Sample No.	Density (g/cm^3)	Weight of the sample (mg)	K_2O content (%)	K content (%)	$^{40}Ar^*$ (nl/g)	$^{40}Ar^*$ (%)	Radiometric age and error (1σ) (Ma)	Geochronological Laboratory
G1	2.5 -2.6	50.25	8.41	6.980 0.059	10.2304 0.2937	72.42 2.53	37.3 ± 1.1	Heidelberg
G2	2.5 -2.6	58.65	6.22	5.163 0.052	9.5334 0.6084	52.43 3.44	46.9 ± 3.0	Heidelberg
G3	2.5 -2.6	30.00	6.49	5.39	9.491	58.49	41.1 ± 4.0	Lublin
G4	2.5 -2.6	40.56	8.99	7.463 0.075	14.9767 0.8500	55.50 3.26	50.9 ± 3.0	Heidelberg

The octahedral cation composition and their number within the limits of 2.0 justifies classifying glauconites as dioctahedral mica with small defects in the octahedra. It should be noted that octahedral Al plays an important role in samples G2 and G4. This can indicate that these samples represent intermediate members of the illite-glauconite series.

Radiometric dating of glauconite

Direct dating of sediments has been the subject of numerous studies since the beginning of geochronology, especially in papers by Hurley et al. (1960), Evernden et al. (1961), Polevaya et al. (1961), Hurley et al. (1963), Hurley (1966) and Amaral and Kawashita (1967). Determination of the numerical chronostratigraphic ages was also carried out by many authors (Odin 1982; Harland et al. 1982; Palmer 1983; Vass and Balogh 1986; Harland et al. 1989).

Since the beginning of the application of radiometric methods for dating sedimentary rocks, much progress has been achieved from an analytical point of view.

The most widely used for glauconite dating is the potassium– argon method. Due to its range of applicability, the method is considered to be the most efficient in determining isotopic geochronology of sedimentary rocks.

In Poland the application of mass spectrometry to the isotopic method of geochronology was initiated in

1991. The first samples of glauconite from the Lublin Upland were dated in the Mass Spectrometry Laboratory of the Physics Institute, Maria Curie-Skłodowska University at Lublin. They were dated by the external standard technique and since 1995 by the isotopic dilution technique with the application of the ^{38}Ar isotope as spike.

The estimation of the radiometric age of glauconite discussed in this paper was performed by the same K-Ar method with the application of isotopic dilution technique in two geochronological laboratories: three glauconite samples (G1, G2 and G4) were dated in the Geochronological Laboratory of Heidelberg University, Germany, and one sample (G3) – in the Physics Institute of Maria Curie-Skłodowska University at Lublin, Poland (Table 5).

The K–Ar age determination of individual samples of glauconite was calculated from the following formula (Dalrymple and Lanphere 1969)

$$t = -\frac{1}{\lambda}\ln\left(1 + \frac{\lambda}{\lambda_e}\frac{^{40}Ar*}{^{40}K}\right)$$

where $^{40}Ar^*$ is the concentration of radiogenic argon, i.e. the quantity of argon generated in one gram of the mineral by the potassium decay (the mineral is considered to be a closed system from its formation to present), ^{40}K is the quantity of radioactive potassium in one gram of the sample (which was estimated from the potassium concentration determined by the ASA flame method and from its natural abundance = 0.0167%, $\lambda=5.543 \times 10^{-10}y^{-1}$ and $\lambda_e=\lambda/9.540$ are decay constants of ^{40}K, (Steiger and Jäger 1977).

$^{40}Ar^*$ was determined mass-spectrometrically by the isotope dilution technique in which an unknown amount of ^{40}Ar was mixed with a known amount of radioactive ^{39}Ar ($T_{1/2}$=269 y) and the resultant isotope ratio $^{40}Ar/^{39}Ar$ was measured. Both argon species were released simultaneously from the solid materials sample and the spike (a granulate of Quaternary biotite in which ^{39}Ar was formed by neutron irradiation in a nuclear reactor). One gram of this spike yields 1.972 ± 0.005 nl of ^{40}Ar at STP with a very low ratio of $^{40}Ar/^{39}Ar$ (Furhman 1983).

A portion of the 50–100 mg glauconite fraction was wrapped in aluminum foil. In similar manner a portion of 5–10 mg of the spike was prepared and wrapped together in common Al-foil. The set of samples was put into a loss extraction line which was connected to the mass spectrometer. Argon was released from each sample by a radio frequency oven. The argon gas extracted was purified in the usual way and analyzed mass-spectrometrically. A MAT GD-150 mass spectrometer with 180^0 magnet sector, 5 cm curvature and 0.51T magnet was used in a static vacuum mode.

The total amount of ^{40}Ar extracted was calculated from the following formula:

$$^{40}Ar = {}^{39}Ar \frac{1-f}{f}\left[\left(\frac{^{40}Ar}{^{39}Ar}\right)_{mixture} - \left(\frac{^{40}Ar}{^{39}Ar}\right)_{spike}\right]$$

where f is the weight fraction of sample, $1-f$ is that of the spike, and ^{39}Ar is the amount of the radioactive argon determined from the weight of the spike; the isotope ratio $^{40}Ar/^{39}Ar$ of the mixture was determined mass- spectrometrically. In the above-mentioned case the ratio ranged from about 10 to 100 while that for the spike was nearly zero, a value of 0.0645 ± 0.077 was assumed on the basis of previous calibration. Note that ^{40}Ar calculated from the above formula required correction for atmospheric argon by taking into account the peak of ^{36}Ar, which appeared in the mass spectra. The radiogenic component of ^{40}Ar is

$$^{40}Ar^{*} = (^{40}Ar)_{\text{measured}} - 295.5(^{36}Ar)_{\text{measured}},$$

where 295.5 is the $^{40}Ar/^{36}Ar$ ratio for atmospheric argon. Finally, from total radiogenic argon a small fraction of this isotope which came from spike was subtracted.

The standard age error was estimated from the error propagation formula assuming relative error of 1% for potassium determination and from the estimated variance of $^{40}Ar^{*}$ on the basis of five repetitive argon spectra.

Results

The radiometric dates of the tested glauconite showed an Eocene age and not a Sarmatian or an Oligocene one as it was mentioned in the literature. The genetic character of the investigated glauconite was estimated as allogenic in the Czułczyce deposit (G1) (Photo 1) – redeposited from older (Eocene) deposits (Krzowski 1998) and authigenic in the Chełm (G2) (Photo 2), Janów (G3) (Photo 3) and Lechówka (G4) (Photo 4) deposits – syngenetic with the host deposits (Table 6).

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

Discussion and Conclusions

The application of glauconite to the geochronological interpretation requires complex analytical investigations and careful estimation of their results. Many factors can cause incorrect glauconite dating. The major ones are factors of geochemical type, that is disturbances of the internal structure of glauconite resulting from advanced alterations that disturbed the K-Ar isotopic equilibrium. However, due to the mica-type layer structure during the alteration processes, K and Ar were lost proportionally and for this reason slightly weathered glauconites can still show a reliable radiometric age.

The degree of glauconitic material maturation influences its compositional and textural features. From a chemical point of view the most discriminating parameter is the K content of the glauconitic material, showing its degree of maturation.

Table 6
Stratigraphic reinterpretation of Tertiary glauconite-bearing formations on the basis of radiometric dating of glauconite

System	Epoch	Stage	Radiometric ages (Ma)*	Glauconite studied: Sample No	Glauconite studied: Location	Glauconite studied: Lithology	Glauconite studied: Apparent glauconite age (Ma)	Stratigraphy: hitherto	Stratigraphy: present	Interpretation
TERTIARY	OLI.	RUPELIAN								
			35.4							
	EOCENE	PRIABONIAN		G1	Czułczyce	quartzose-glautonitic sands	37.3 ± 1.1	Middle Miocene (Lower Sarmatian)	Middle Miocene (Lower Sarmatian)**	glauconite reworked from Eocene deposits**
			38.6							
		BARTONIAN		G3	Janów	quartzose-glautonitic sands	41.4 ± 4.0	Middle Miocene (Lower Sarmatian)	Middle Eocene (Bartonian)	authigenic glauconite - syngenetic with matrix deposits
			42.1							
		LUTETIAN		G2	Chełm	clayey sand with glauconite	46.9 ± 3.0	Middle Miocene (Lower Sarmatian)	Middle Eocene (Lutetian)	authigenic glauconite - syngenetic with matrix deposits
			50.0							
		IPRESIAN		G4	Lechówka	quartzose-glautonitic sands	50.9 ± 3.0	Middle Miocene (Lower Sarmatian)	Lower Eocene (Ipresian)	authigenic glauconite - syngenetic with matrix deposits
			56.5							
	PAL.	THANETIAN								

* - after Harland et al. (1989)

** - after Krzowski (1998)

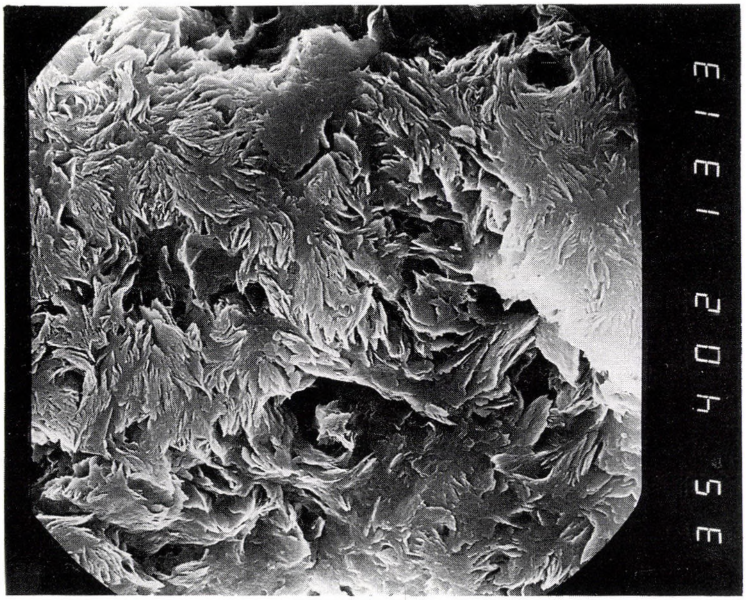

Photo 1
Scanning electron micrograph of glauconite grain from Czulczyce (G1). Magn. ×4000

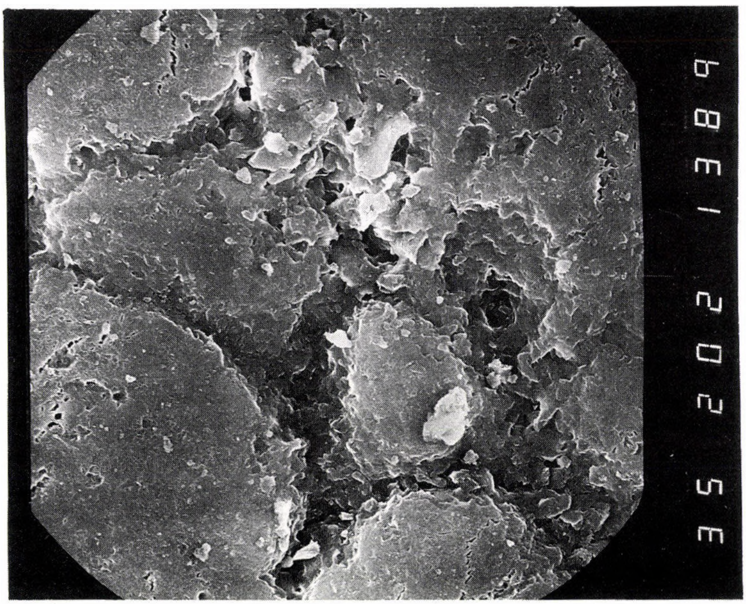

Photo 2
Scanning electron micrograph of glauconite grain from Chelm (G2). Magn. ×2000

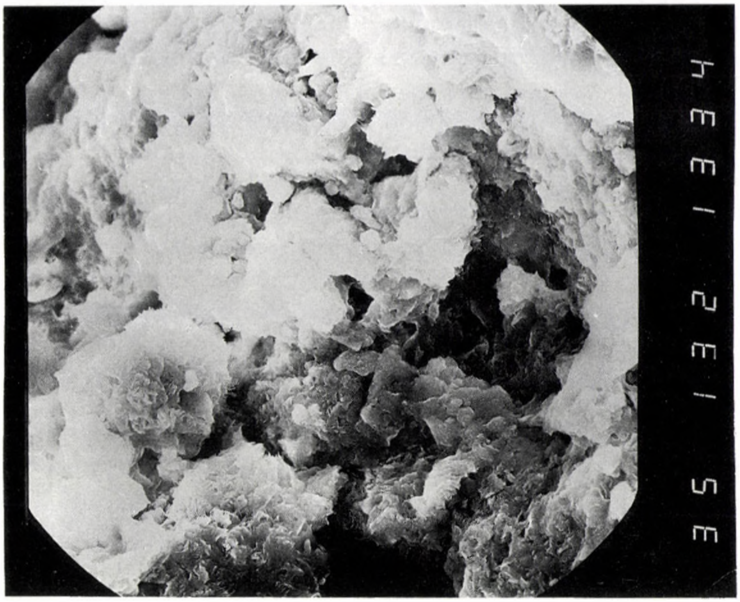

Photo 3
Scanning electron micrograph of glauconite grain from Janów (G3). Magn. ×1300

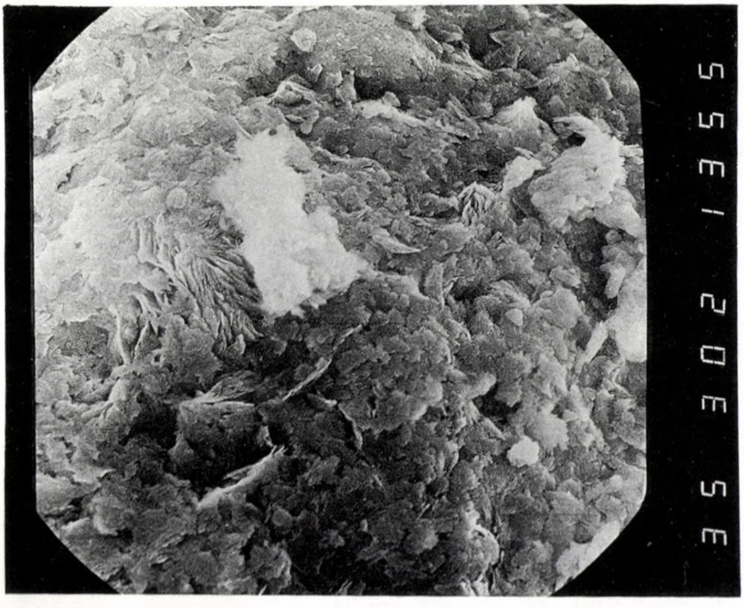

Photo 4
Scanning electron micrograph of glauconite grain from Lechówka (G4). Magn. ×3000

Glauconites from the G1 and G4 samples contain a high K_2O content (8.4 and 8.99%, respectively) and show a high degree of maturation. Their high maturation is proved by the XRD data (narrow XRD peaks). Glauconites from the G2 and G3 samples contain lower values of K_2O (6.22 and 6.49%, respectively) and show a greater presence of smectite areas and broader XRD peaks.

The large grain size of glauconite ranges (from 0.10 mm up to 0.50 mm in diameter) and density selection as well as high K_2O content (higher than 6%) in the tested glauconites had a major influence on the positive result of their dating.

The occurrence of large glauconite grains can be explained on the basis of the size of original pellets and of the more evolved stage of the glauconitization processes. The glauconite occurring as internal molds of foraminifera would indicate that those glauconite pellets might have been formed in an open marine environment (sublittoral or littoral).

Tertiary glauconites of the Chełm Hills area were originally a mature species with relatively high potassium content, although not very much developed. Subsequently they were altered during Miocene weathering, when the Eocene deposits were exposed. It is worth mentioning that the investigated glauconites were altered both in the outcrops and in the borehole and only the degree of their alteration was different.

The dated glauconites showed that they could be utilized as a geochronometers in the chronostratigraphic studies of the glauconite-bearing Lublin Cenozoic formations.

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Anniversaries

In Love with the Earth
Lajos Lóczy sen. was born 150 years ago

From Transylvania through Switzerland to Budapest

After the disastrous end of the Hungarian War of Liberation (1848–49) fought against the Habsburg dynasty of Austria, and defeat by the Tsarist Russian Army, the middle-class Lóczy family had to flee from their Transylvanian home to Pozsony (now Bratislava) in NW Hungary. It was there that Lajos Lóczy was born on 4 November 1849.

His father died early. The loving mother, recognizing the remarkable talents of the son, sacrificed her modest fortune to send him to study at the Institute of Technology in Zurich, Switzerland (Eidgenössische Technische Hochschule). He graduated as a Civil Engineer, but was so fascinated and attracted by the secrets of the Earth that, instead of beginning a promising career in engineering (under the rather favorable circumstances of the so-called Post-compromise Period) he accepted a more than modestly paid post of junior curator in the Natural History Department of the Hungarian National Museum in Pest. His capabilities were soon also highly appreciated there.

The Great and Fruitful Adventure in Asia

When Count Béla Széchenyi began organizing an expedition to Eastern Asia the young Lóczy was warmly recommended to him as an earth scientist for the team. The journey lasted three years (1877–1880). The (not easily accessible) regions explored were the southern margin of the Himalayas (where Lóczy was the first to recognize the overthrust (nappe) structure), several parts of China (producing, among other things, the first scientific description of the Kun-lun Mountains), Burma and Java.

Back in Hungary

While editing the reports of the expedition (even including a thick volume in which he deals with the History of China!) Lóczy continued working, first in the Museum, later in the Royal Hungarian Geological Institute as a geologist, and became Associate Professor of Geology at the Budapest Technical University.

New Ways of Teaching Geography

In 1889, much to his own great surprise, he was invited to become Full Professor for the Chair of Geography of the Budapest University of Sciences. There he introduced modern Physical Geography, including geology-based genetic geomorphology, organized almost every year far-reaching field trips abroad with his students, and was eventually elected president of the Hungarian Geographic Society as well as Ordinary Member of the Hungarian Academy of Sciences.

Third Director of the Royal Hungarian Geological Institute

Lóczy was already aged 59 when he was appointed Director of the Royal Hungarian Geological Institute (the Geological Survey of Hungary, founded 1869). Accepting this new challenge he reorganized and enlarged the staff. He established instructions and norms for geologic mapping, moving the focus of these operations to Croatia and Dalmatia. During World War I he personally took part in the geologic surveying of Serbia.

All-round Investigation of the Lake Balaton Area

Lóczy initiated the creation of a Commission on Lake Balaton. With an incredible effort and incomparable perseverance he succeeded in implementing a fantastic achievement, which was both interdisciplinary (ranging from geology through hydrology and meteorology to archeology and ethnography) and international (with contributors from Austria and Italy). The results were published in 32 (no mistake: thirty-two) thick A4-format volumes in Hungarian and German. This is an achievement of Hungarian science which has remained unique and exemplary up to the present day.

Edelweiss on a Red Tombstone

At the time of the political troubles of 1919 Lóczy retired from the post of Director to live in a small village on the Lake Balaton Riviera (Balatonarács). When his heart stopped beating in 1920 he was buried there. A Permian New Red Sandstone block has been erected on his tomb. It is decorated with an edelweiss (placed under glass) which was sent by another great Hungarian-born explorer of Asia, Sir Aurel Stein.

L. Lóczy Sen. was undoubtedly one of the greatest earth scientists in Hungary. The use of the term "earth scientist" is intended to indicate that he has acquired immortal merits both in geology and geography. His statue stands in the lounge of the Geological Institute of Hungary; the L. Lóczy Medal is the highest distinction that is awarded by the Hungarian Geological Society. Fossils, streets, grammar schools all bear his name – even a peak in the Himalayas.

On 4 November 1999, the 150th anniversary of his birth, a full-day commemorative session took place at the Hungarian Academy of Sciences. The seven lectures presented on this occasion will soon be published.

Earth Trek – the Next Generation

One of his sons, Lajos Lóczy Jun., was also an outstanding geologist and became Director of the Hungarian Geological Institute. In 1947 he left Hungary and went overseas. He successfully explored the petroleum potential of Brazil and was elected Member of the Brazilian Academy of Sciences. But this is another story.

Endre Dudich

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Books, Maps, Software

New geologic map of the Balaton Highland (1982–1999)

The year 1999 is a notable anniversary for geologic research of the Balaton Highland. This year, the Hungarian earth science community celebrated the 150th birthday of Lajos Lóczy sen., author of the world-famous Balaton monograph. The Geological Institute of Hungary paid its respects to the memory of its former director by publishing, in the same year, the geologic map of the Balaton Highland at 1:50 000 scale (Budai et al. 1999) and the related explanatory notes (Eds: Budai and Csillag 1999), which summarize the results of the geologic survey completed a few years before.

The geological survey of the Balaton Highland, which constitutes the southeastern flank of the synclinal structure of the Transdanubian Range, began in 1982 in the Keszthely Mts and – following the SW–NE strike – ended in 1988 in the area of Balatonfő. The basis of the survey were topographic maps at a scale of 1:10 000 as well as color and black-and-white aerial photographs. Editing and compilation of the basic data collections were carried out according to the map sections at a scale of 1:25 000, partly during the survey, partly thereafter. Unification of the map sheets and editing of the covered version of the regional map at a scale of 1:50 000 took place during 1989–1990, while the uncovered version and the explanatory notes were completed as a manuscript during 1991. The scientific results of the mapping were summarized by several printed publications, manuscript reports, dissertations and maps.

Later – following an involuntary pause of some years – the idea of publication of the regional geologic map at 1:50 000 scale and of the explanatory notes was raised at the beginning of 1999. However, for several reasons the original manuscript versions were not suitable for preparing a published map, nor were the explanatory notes for printing. Basically, the total re-editing of the map was justified by the fact that instead of the previously planned covered and uncovered maps only the publication of a single version was possible. The newly edited map (Budai, T., G. Csillag, L. Koloszár, A. Dudko (Eds): Geologic map of the Balaton Highland, 1:50 000. – Geol. Inst. Hung.) attempts to mix the good points of the covered and uncovered versions in a form which – apart from the detailed, genetic subdivision of the Quaternary formations – over considerable areas leaves out the deluvial detritus in order to reveal the underlying basement formations and the structural forms. Use of the map is made easier for non-Hungarians in that all the texts (titles, legends, etc.) are also given in English. The geologic map was prepared by digital editing (GIS), in a unified national projection, making use of the Digital Topographic Model created by the Cartographic Office of the Hungarian Army.

Akadémiai Kiadó, Budapest

In editing the explanatory notes (Budai, T., G. Csillag (Eds): "Geology of the Balaton Highland. Explanation of the Geologic Map of the Balaton Highland, 1:50 000". – Geol. Inst. Hung., 257 p.), the original manuscripts could also only be used as starting points, since certain chapters practically summarize the scientific results of the period after the survey (the most striking example of this is the representation of the Pannonian basalt volcanism and of environmental geology). In terms of its style the volume attached to the map represents a transition between the traditional explanatory notes published formerly by the Geological Institute and the monographic dissertations, both in form and in content. Based upon the description of certain formations as well as the extent of the chapters the volume is slightly eclectic, since the quantity and depth of the published information attempt to harmonize with the proportion of the assumed significance of geologic build-up, of structure or of evolution. The closing chapters are followed by detailed stratigraphical representation; however, structural geology and hydrogeology (including this time limnogeology, as well) play a more important role than in the former explanatory notes. Similarly, there is a significant difference compared to the traditional explanatory notes in that the volume is bilingual and illustrated with text figures (71), photo plates (47) and color map supplements. The date of manuscript completion and that of its appearance almost coincide with each other, which guarantees the up-to-date character of the contents.

Tamás Budai, Gábor Csillag
Geological Institute of Hungary

Acta Geologica Hungarica, Vol. 42/4, pp. 477–478 (1999)

Commission News

In May 1999 the General Assembly of the Hungarian Academy of Sciences elected a new Chairman and a new Vice-Chairman of the Department of Earth Sciences for the next academic cycle (1999–2002).

Chairman: György Pantó, Ordinary Member of the Hungarian Academy of Sciences Vice Chairman: József Ádám, Corresponding Member of the Hungarian Academy of Sciences

In October 1999, during a series of sessions, the scientific commissions were reorganized and their new office-holders were elected. The commissions in the field of geology, geochemistry and paleontology elected the following leaders for the next academic cycle:

Geological Commission: Chairman: János Haas, Secretary: Géza Császár
Geochemical Commission: Chairman: Péter Árkai, Secretary: Mária Földvári
Paleontological Commission: Chairman: Attila Vörös, Secretary: József Pálfy

On October 26, 1999, during the session of the Department of Earth Sciences, the lists of names for the Editorial Boards of the foreign language journals published under the auspices of the Department were approved. The Department of Earth Sciences extended the assignment of György Bárdossy as the Chairman of the Editorial Board. Simultaneously, János Haas was called upon to continue serving in the capacity of Editor-in-Chief of Acta Geologica Hungarica for the next academic cycle. György Bárdossy invited the former members of the Editorial Board to continue their activity in the next cycle. Thus, there is no change in the composition of the Editorial Board and the members of the International Advisory Board also continue in their office.

János Haas

IUGS Hungarian National Committee renewed

The Hungarian Academy of Science (HAS) has traditionally coordinated the connection between Hungary and the IUGS. The IUGS Hungarian National Committee (hereafter "Committee") works within the framework of the Department of Earth Sciences of the HAS.

In agreement with the general rules of HAS for its commissions, officers and members of the Committee are elected and invited to serve for a three (academic) year-long term. As the terms of the members of all commissions

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expired in 1999, a new IUGS HNC had to be formed for the 1999/2000 - 2001/2002 academic years.

The former officers of the outgoing Commission (Chairman: Prof. Dr. György (George) Bárdossy; Secretary: Dr. Endre Dudich CSc) performed excellently, but had already served for two terms, and were thus not eligible for re-election. The Department of Earth Sciences expressed its gratitude for their six year-long service.

For the new term Prof. Dr. Péter rkai, corresponding member of the HAS, was elected Chairman and Dr. Tamás G. Weiszburg was elected Secretary of the Committee.

The 14 members invited to serve on the Committee for the coming term are Prof. Dr. Gy. (George) Bárdossy, Mr. Károly Brezsnyánszky, Dr. Géza Császár, Prof. Dr. László Cserepes, Dr. László Csontos, Dr. Endre Dudich, Dr. Elizabeth Erdélyi, Prof. Dr. András Galácz, Prof. Dr. Gyula K. Greschik, Prof. Dr. János Haas, Ms. Eszter Havas-Szilágyi, Prof. Dr. Magdolna Hetényi, Prof. Dr. László Kordos, and Prof. Dr. Attila Somfai.

In order to support better understanding and cooperation between the different branches of earth sciences the secretaries of the Hungarian national committees for the IUGG and the IGU will also be invited, as liaison officers, to future meetings of the Committee.

Tamás Weiszburg

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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, 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
- the name of the institution and city where the work was prepared
- an abstract of not more than 200 words
- a list of five to ten key words
- a footnote with the postal address of the author

The SI (System International) should be used for all units of measurements.

References

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.

Figures and tables

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. 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.

Proof and reprints

The authors will be sent a set of proofs. Each of the pages should be proofread, signed, and returned to the Editorial Office within a week of receipt. Twenty-five reprints are supplied free of charge, additional reprints may be ordered.

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