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Paleogeography, paleoecology and eustacy: Miocene 3rd order cycles of relative sea-level changes in the Western Carpathian – North Pannonian basins

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Interaction of sea-level changes and tectonics had an important influence on the paleogeography and thereby also on the paleoenvironment of the Western Carpathian – North Pannonian Basins, which formed the northern bays of the Central Paratethys epicontinental sea in the Miocene. The depth and the shape of the basins were predominantly controlled by the main tectonic events. Relative eustatic changes reflected in coastal onlaps were followed mostly by the rise of water paleodepth in the offshore environment. The correlation of the constructed curves for the coastal onlap and estimated paleodepth with the global reference curves (Haq et al. 1988; Haq 1991) shows some discrepancies, predominantly caused by tectonics during the basin development.

The Late Egerian–Early Eggenburgian isolation or erosion was followed by the Eggenburgian transgression (ca. 21 Ma; NN2 zone) and deepening of the sedimentary environment in the Western Carpathian – North Pannonian Basins. Later on, during the Ottnangian (NN3 zone), a brackish paleoenvironment developed in the Vienna and the Danube Basins (TB 2.1 global cycle;Haq et al. 1988). In the East Slovakian Basin the Ottnangian uplift was associated with hiatus, unlike the Novohrad Basin, where deposition of the continental coal-bearing formations passed upward into the regional highstand deposition characterized by lacustrine to brackish environment with the Late Ottnangian marine ingressions.

The Early Karpatian (NN4 zone) transgression and highstand in the Western Carpathian – North Pannonian Basins can be correlated with transgression and global sea-level rise of the TB 2.2 global sea-level cycle (Haq et al. 1988). In the East Slovakian Basin the local sea-level fall at the late highstand led to salinity crisis.

The following Late Karpatian–Early Badenian sea-level cycle (NN4 zone) is observable only in the East Slovakian and the Novohrad Basins and it can be correlated with the TB 2.3 global cycle (Haq et al. 1988). In the Vienna and Danube Basins the erosion of uplifted areas or terrestrial deposition in the depressions occurred.

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Pronounced late Early Badenian transgression (NN5 zone) followed by highstand was observed in all Western Carpathian – North Pannonian Basins during the extensional synrift stage of the development. The late Early Badenian to Middle Badenian relative sea-level cycle, which can be correlated with the TB 2.4 global eustatic cycle (Haq et al. 1988) is known only from the East Slovakian Basin, where the late highstand characterized the evaporite sedimentation. The next sea level change, which can be correlated with the TB 2.5 global cycle (Haq et al. 1988) is proved by transgression followed by deepening of the sedimentary environment during the Late Badenian (NN6 zone). In the Vienna and Danube Basins the late Early Badenian transgression was followed by highstand conditions, which lasted until the Late Badenian.

The last well-observed relative sea-level cycle, which can be correlated with the TB 2.6 global eustatic cycle (Haq et al. 1988; Haq 1991) was associated with the Sarmatian transgression (NN7 zone), highstand and gradual shallowing in the Early Pannonian.

The Late Miocene global sea-level changes cannot be satisfyingly interpreted by means of paleoecology in the Western Carpathian basins due to their isolation and lack of relevant chrono- and biostratigraphic data in the Pannonian and Pontian deposits.

Key words: Early and Middle Miocene, Central Paratethys, biostratigraphy, paleoecology, sequence stratigraphy

Introduction

The Neogene paleogeography of the Western Carpathian – North Pannonian region (Fig. 1), as a combination of aquatic and continental environments in the Central Paratethys, was influenced not only by geodynamic factors but also by the regional manifestations of global sea-level changes (sensu Haq et al. 1988; Haq 1991; Hardenbol et al. 1998).

The proper determination of depositional systems tracts was carried out by detection of relative eustatic changes using geophysical methods and sedimentology, but the study of relative sea-level cycles based on paleoecology also played an important role (Kovác and Hudácková 1997; Kovác and Zlinská 1998; Kovác et al. 1999). Along with the estimation of onlap curves, the paleobathymetric reconstruction of sedimentary environment was carried out as well. The correlation of cycles in the respective time intervals was enabled by new bio- and chronostratigraphic data. Simultaneously, the principal rule of sequence stratigraphy was taken into account in that the reflection of a sea-level change can be strengthened, weakened or it may disappear completely, depending on the value of tectonic subsidence and the rate of sediment input (Brown and Fischer 1977; Vail et al. 1984; Posamentier and Vail 1988; Van Steen and Winkler 1988). Therefore the study of the relative sea-level cycles manifestations in the Western Carpathian – North Pannonian Basins (Figs 5, 6, 7, 8) also took into consideration, beside the global sea-level changes (Haq et al. 1988; Haq 1991), data on the tectonic subsidence and detrital input rates proved by analysis of the subsidence history (Lankreijer et al. 1995; Baráth et al. 1997; Lankreijer 1998)

In the paleogeography of the Western Carpathian – North Pannonian Basins representing individual bays of the Paratethys sea during the Miocene,

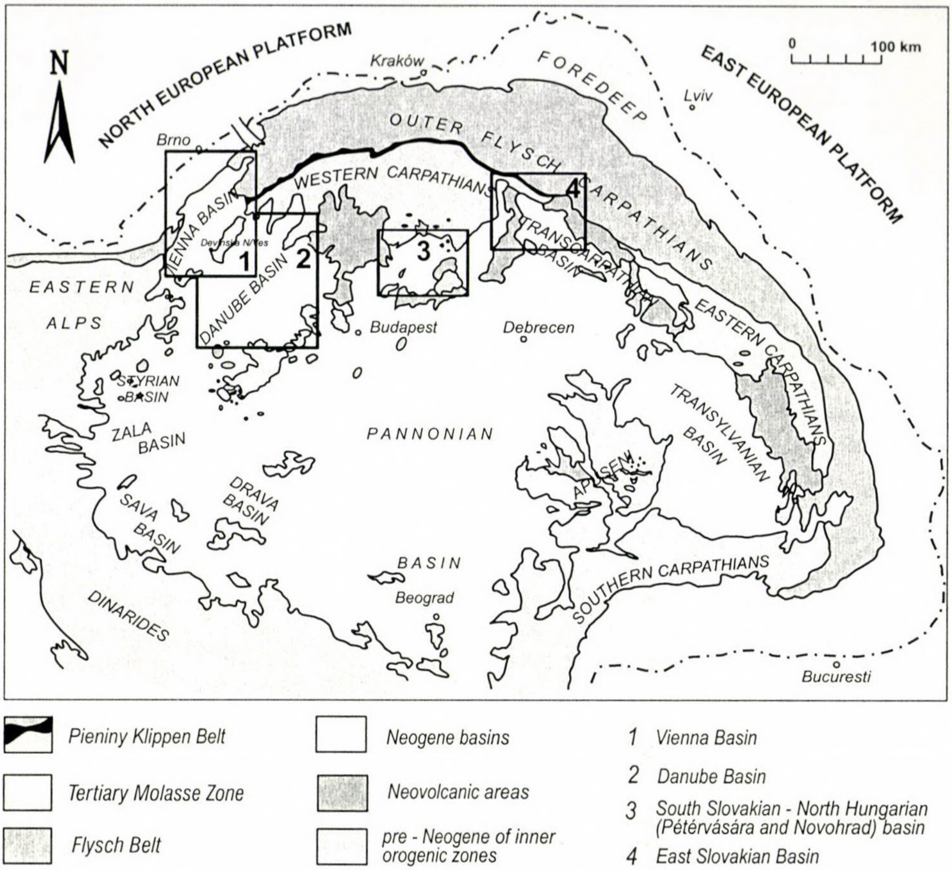


Fig. 1
Study areas – the Vienna, Danube, South Slovakian – North Hungarian and the East Slovakian Basins – are indicated

tectonically-controlled local relative sea-level changes became dominant (Nagymarosy 1980, 1985; Nagymarosy et al. 1995; Kókay 1996). Hence, the correlation of global sea-level changes with regional, relative sea-level cycles in the studied region is very difficult due to its dramatic geodynamic evolution and sometimes also due to lack of relevant chronostratigraphic data.

In the following those relative 3rd order sea-level cycles will be described that can be traced by means of paleoecology in the Western Carpathian – North Pannonian region during the Miocene (Fig. 1). These cycles of the relative sea-level changes are marked by symbols from CPC 1 to CPC 6 (3rd order relative sea-level cycles in the Carpathian-Pannonian region).

Methods

On the basis of changes in the calcareous nannoplankton and foraminiferal assemblages we were able to determine the paleoecological conditions of the sedimentary area in the time and space, as well as establish more precise biostratigraphy. The most important factors, which are reliably interpretable from the presence of foraminifera genera in the assemblages, were the salinity changes in the environment (Elphidium, Prosononion, Conorbella, Ammonia) and decreased oxygen level at the bottom (Bulimina, Bolivina, Uvigerina). In estimating the paleodepth of the sedimentary environment, a large variety of factors can be taken into account (e.g. temperature, hydrostatic pressure, hydrodynamics, substrate, nutrients, etc.). We omitted the detailed bathymetric zonation used in older literature. Instead we use the stenohaline environment depth division: shallow neritic (predominance of epiphytic species), deep neritic (cibicidoid species, lagenides, Cyclammina, Haplophragmoides) and shallow bathyal environments (Hoeglundina, Pullenia, Hansenisca, Oridorsalis). For the purpose of sequence stratigraphy and interpretation of the cyclic events during deposition, some other characteristics of the foraminiferal assemblages have been studied, e.g. presence and type of redeposited forms, immigration of new species, diversity, plankton/benthos ratio of the foraminifera, ratio of agglutinated/calcareous foraminifera and overall number of foraminifera (abundance) in the assemblages. In the following overview we characterize the important terms:

Sea-level lowstand (LST) at its beginning is usually manifested by erosion of older sediments of the emerged basin margins and their rapid redeposition into the lowstand wedges. Later the incised valleys will be filled up by fine or coarse-grained sediments until the next transgression. The lowstand in foraminiferal assemblages is characterized by frequent microfaunal redeposited forms from older strata (Poag and Comeau 1995), rapid changes in composition of the assemblages caused by instability of paleoecological conditions (Gaskell 1991; Rey et al. 1994) and predominance of shallow-water species (Armentrout et al. 1990;). The best verified sign of the sea-level lowstand assemblages is the low abundance (Vail et al. 1984; Schaffer 1990; Vail and Wornandt 1990; Armentrout et al. 1990).

Marine transgression (TST) is characterized by onlap of shallow marine environment on erosional sequence boundaries (SB 1 type), deepening-upward successions, ravinement surfaces, etc. In the offshore environment it is documented by onset of new micro- and macrofauna elements, namely in half-closed epicontinental seas, e.g. the Central Paratethys during the Neogene. In the central part of the basin, where the transgression is manifested only by faunal changes in condensed sections, the sequence boundary become a conformable surface (Vail et al. 1977). Marginal transgressive facies may reflect either normal or decreased salinity; in appropriate paleogeographical conditions formation of coal seams occurs (Kovác et al. 1999). The culmination of the transgression and beginning of sea-level highstand is known as the period of maximum flooding and is marked by a condensed section at the seaward end of the sequence (maximum flooding surface – mfs). It can be characterized by restoration of the marine seaways. In case of the Central Paratethys it is often documented by immigration of new faunal elements from other bioprovinces (e.g. Mediterranean, Eastern Paratethys).

Marine transgression in the foraminiferal assemblages is also characterized by a relatively high content of redeposited forms due to ravinement from older sediments (Fürsich et al. 1991). A typical sign of transgression is the appearance of new species, represented mainly by planktonic foraminifera. Diversity of the assemblages increases with the progressing transgression, together with the abundance and the plankton/benthos ratio. In the paleoecological record the progress of the transgression is usually connected to a deepening of the sedimentary environment and transition from oligohaline to euhaline conditions (Armentrout et al. 1990). The increasing diversity is caused by the introduction of new species and especially by the change of unstable sea-level lowstand to stable highstand conditions. The diversity values of the foraminiferal assemblages from the Karpatian deposits of the Novohrad Basin can be mentioned as an example. The diversity values range between 2 and 5 during the sea-level lowstand, whereas it reaches a value of 15 during the sea-level highstand (according to Simpson´s formula; Holcová-Šutovská 1996).

Sea-level highstand (HST) is often demonstrated by maximal depth of the sedimentary basins during the early HST close to mfs. Stable paleoecological conditions are mirrored in highly diversified foraminiferal assemblages (Gaskell 1991; Rey et al. 1994). Abundance of fossils reaches the maximum values (Vail et al. 1984; Schaffer 1990; Vail and Wormandt 1990). During the sea-level highstand stratification of the water column may occur; water circulation is restricted to just its upper portion and a low oxic environment prevails near the basin bottom. Such phenomena have been observed e.g. during the Eggenburgian, characterized by Bathysiphon–Cyclammina foraminiferal assemblages, further in the Karpatian with deposition of sediments dominated by Uvigerina and Pappina foraminiferal assemblages and later in the Late Badenian Bulimina–Bolivina Biozone (Kovác and Hudácková 1997; Kovác and Zlinská 1998; Kovác et al. 1999)

Late highstand regression is often characterized by seaward migration of the shoreline and shallowing up that may result in an isolation of the basins. From the global point of view, this is a period when immigration bridges of mammal fauna were created. During the late highstand (falling stage), an onset of poor, low-diversified foraminiferal assemblages was observed in the Western Carpathian – North Pannonian Neogene Basins. Assemblages are adapted to shallow-water hypo- or hypersaline stress paleoecological environment and contain euryhaline foraminiferal assemblages due to the restriction of the sedimentary areas and due to the poor communication with the open marine environment. An oxygen insufficiency at the bottom of hypertrophic shallow-water environment led to the evolution of assemblages of tiny foraminifera, ecologically tied to algae (Kovác and Hudácková 1997; Kovác and Zlinská 1998; Kovác et al. 1999).

Late Egerian to Ottnangian 3rd order relative sea-level cycles

The time interval 23.5–20.5 Ma, represented by the Late Aquitanian–Late Egerian to Early Eggenburgian–Late Karadzhalgan stages (Berggren et al. 1995; Rögl 1998) is defined by the upper part of the NN 1 and the lower part of the NN2 nannoplankton zones (sensu Martini 1971). The time interval between 20.5–18.8 Ma, represented by the Early Burdigalian–Eggenburgian – Sakaraulian stages (sensu Berggren et al. 1995; Rögl 1998) is defined by the presence of the upper part of the NN 2 and the lower part of the NN3 nannoplankton zones (sensu Martini 1971). The time interval between 18.8–17.5 Ma, represented by the Middle Burdigalian– Ottnangian – Early Kotsakhurian stages (sensu Berggren et al. 1995; Rögl 1998) is defined by the presence of the upper part of the NN 3 nannoplankton zone in the Western Carpathians (sensu Martini 1971), but also by the lower part of the NN 4 zone in Hungary (Báldi-Beke and Nagymarosy 1979; Bohn-Havas and Nagymarosy 1985) and is correlated with the M3 planktonic foraminifera zone (Berggren et al. 1995). The Eggenburgian/ Ottnangian boundary in the Western Carpathian area was estimated at about 19 Ma by radiometric dating (Vass et al. 1985).

The Late Egerian paleogeography (Fig. 2) of the Central Paratethys was characterized by disintegration of a large communication system between the Mediterranean area, Paratethys and the Indian Ocean (Rögl 1998). The existence of the former communication system is reflected by the same mollusc fauna elements from the Bavarian Molasse Zone to the Transylvanian Basin (Báldi 1983). The Eggenburgian paleogeography (Fig. 3) is overprinted by an extensive transgression connected with immigration of a new marine fauna from the Mediterranean (Rögl and Steininger 1983). Typical littoral marine sediments with the pectinid *Chlamys gigas* (Schlotheim) became widespread from the Alpine foredeep as far as the Transylvanian Basin (Steininger and Senes 1971; Báldi 1979, 1983; Rusu 1996; Báldi 1998).

During the Early Ottnangian (NN 3 nannoplankton zone) a marine connection between the Mediterranean and the Alpine Foredeep was still active (Fig. 3). The foredeep sediments are characterized by mollusc fauna with *Flabellipecten*

Fig. 2 →
Egerian isolation during the relative sea-level fall and Late Egerian-early Eggenburgian faunal migration pathways

Fig. 3 →
Egerian isolation during the relative sea-level fall and faunal migration pathways during the Eggenburgian transgression

Fig. 2

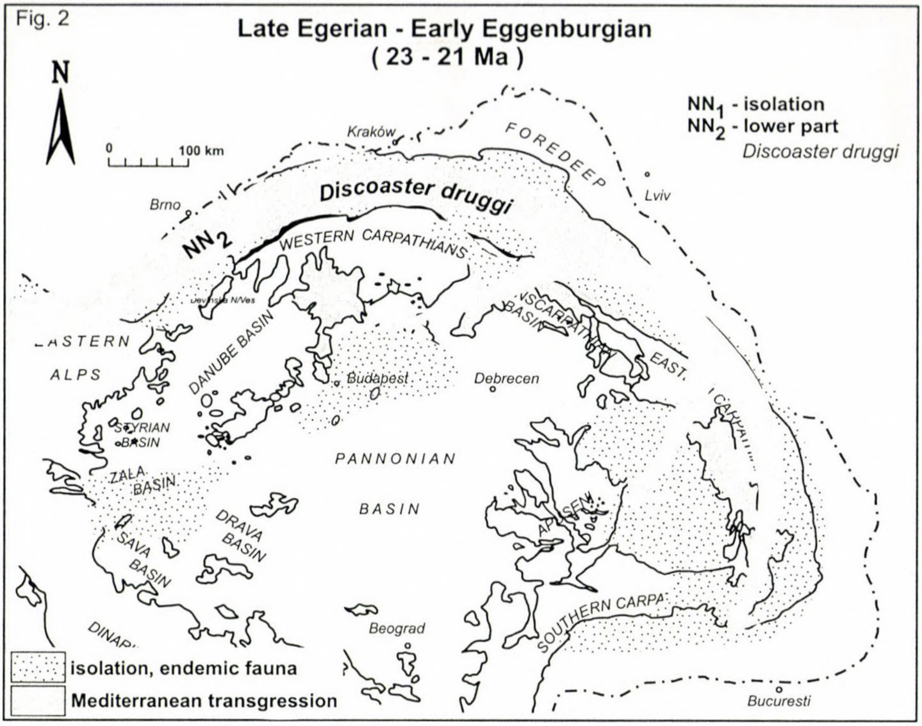
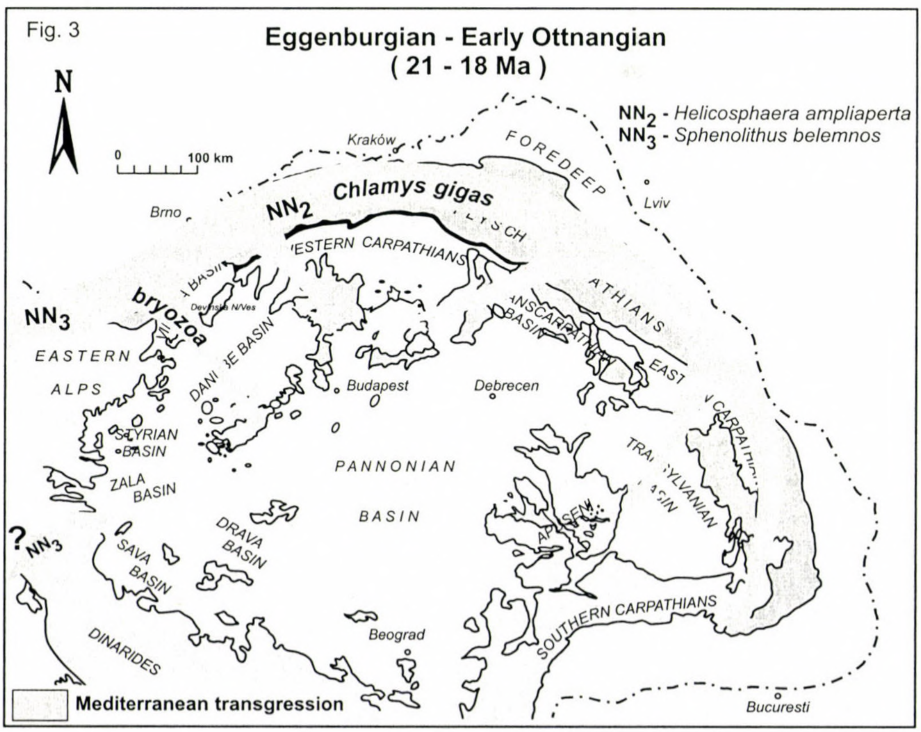

Late Egerian - Early Eggenburgian
(23 - 21 Ma)
NN1 - isolation
NN2 - lower part
Discoaster druggi
100 km
Kraków
FOREDEEP
Lviv
Brno
Discoaster druggi
NN2
WESTERN CARPATHIANS
DANUBE BASIN
EASTERN
ALPS
Budapest
Debrecen
STYRIAN BASIN
ZALA BASIN
PANNONIAN
BASIN
DRAVA BASIN
SAVA BASIN
Beograd
SOUTHERN CARPA
Bucuresti
isolation, endemic fauna
Mediterranean transgression

Fig. 3


Eggenburgian - Early Ottnangian
(21 - 18 Ma)
NN2 - Helicosphaera ampliaperta
NN3 - Sphenolithus belemnos
100 km
Kraków
FOREDEEP
Lviv
Brno
NN2
Chlamys gigas
WESTERN CARPATHIANS
bryozoa
NN3
EASTERN
ALPS
Budapest
Debrecen
STYRIAN BASIN
ZALA BASIN
PANNONIAN
BASIN
DRAVA BASIN
SAVA BASIN
?NN3
DINARIDES
Beograd
SOUTHERN CARPATHIANS
Bucuresti
Mediterranean transgression

hermansenni (Dunker) (Rögl 1998) and bryozoans of Atlantic provenance (Vávra 1981). The littoral Bántapuszta beds in the Várpalota Basin (Transdanubian Central Range; see Kókay 1973) are also considered to be relics of the sedimentary fill of possible prolongation of this Ottnangian marine bay. Their mollusc fauna shows similarities to the coeval faunas in the Alpine region as well as to the Apennines. This opinion is also supported by the presence of similar bryozoan-rich marine deposits in the northern part of the Vienna Basin near Holíc (Zágoršek, pers. comm.) and in the Mediterranean, i.e. the western immigration is also documented by the Ottnangian marine ingressions in the Novohrad Basin situated in the Western Carpathian hinterland (Vass et al. 1987; Šutovská 1993).

The Late Ottnangian paleogeography (Fig. 4) of the Central Paratethys is characterized by an extensive sea-level fall which led to emersion of the continental immigration bridges between Eurasia and Africa during the MN4 mammal zone (Thenius 1979; Barry and Flynn 1989). At the same time it caused the closure of the seaways connecting the Paratethys with the Mediterranean. In the Paratethys endemic mollusc assemblages evolved. This time the shallow-water estuarian facies with *Rzehakia socialis* (Rzehak) developed in the Central Paratethys; in the Eastern Paratethys the shallow-water endemic mollusk assemblages with *Rzehakia dubiosa* (Hoernes) dominated. In the Eastern Carpathian foredeep basin isolation resulted in deposition of evaporites (Kovác et al. 1999).

Shelf margin system tracts (SMST) in the Late Egerian – Early Eggenburgian

In the Western Carpathian – North Pannonian region the Upper Egerian sediments in the present seem tobe isolated from the Outer Carpathian units, mainly from the South Slovakian – North Hungarian and the East Slovakian Basins (Fig. 1). Parts of these deposits were, and still are, assigned to the Eggenburgian, because the Egerian – Eggenburgian boundary can hardly be distinguished in the Western Carpathian basins, since the index microfossils are rare or absent at this time similarly to the previous NN1 nannoplankton zone. One of the possible explanations for this fact is that migration of NN1 zone index species from the open ocean was sporadic, due to isolation of the Central Paratethys during a short time span.

The Late Egerian isolation of the basins is demonstrated by small-size foraminifera fauna reflecting the stressed paleoecological conditions in this time (Kantorová 1977, 1978). The first representations of the Early Miocene immigration were the NN2 nannoplankton zone index fossils during the Late Egerian (Early Eggenburgian resp.). In contrary to the residual flysch basins at the orogenic front (Pouzdrany Unit; Krhovský et al. 1995), in the intramontane basins of the Western Carpathian – North Pannonian region this event is demonstrated by the replacement of tiny and relatively poor foraminiferal assemblages by rich assemblages of normal size specimens.

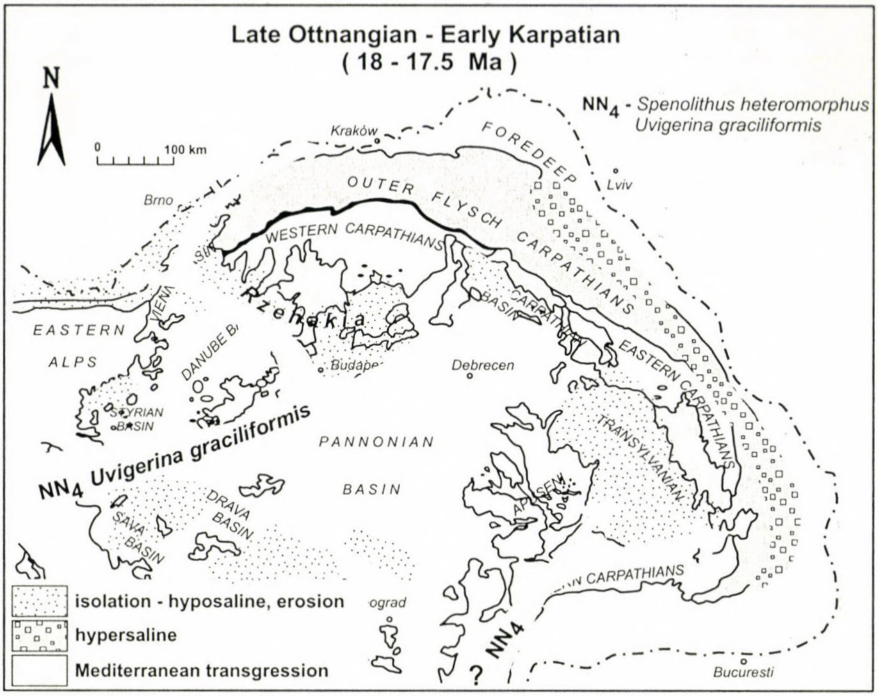


Fig. 4
Late Ottnangian isolation during the relative sea level fall and faunal migration pathways during the Karpatian transgression

The Late Egerian to Eggenburgian lowstand can be biostratigraphically defined on the basis of calcareous nannoplankton and foraminifera as follows: onset (FAD) of *Reticulofenestra pseudoumbilica* (Gartner) Gartner, *Helicosphaera scissura* Miller and *Discoaster druggii* Bramlette et Wilcoxon, later also FAD of *Globigerinoides trilobus* (Reuss). The upper boundary is characterized by transgressive beds with *Helicosphaera ampliaperta* (Bramlette et Wilcoxon).

At the northwestern margin of the East Slovakian Basin (Figs 2, 5) the Egerian to Eggenburgian age of the Biely Potok Formation is proved by the calcareous nannoplankton of the NN2 zone in the upper part of the sedimentary sequence (Soták and Stárek 1999). The formation was deposited in a residual (expiring) fore-arc basin with its center located in the present Levocské Vrchy and the Šarišská Vrchovina Mts and is considered as deposit of lowstand systems tracts (Soták 1998). Gradual transition from the pelagic to low-oxygen environment is documented by diversified foraminiferal assemblages with *Almaena osnabrugensis* (Roemer), *Bolivina fastigia* Cushman, *Globigerina ouachitaensis* Howe et Wallace and *Virgulinella chalkophila* (Hagn) from the Šarišská Vrchovina Mts. (Molnár et al. 1992). On the other hand, the sediments from the HGZ-9 borehole contain sparse pyritized casts of the euryoxibiontic genera Bolivina and Praeglobobulimina.

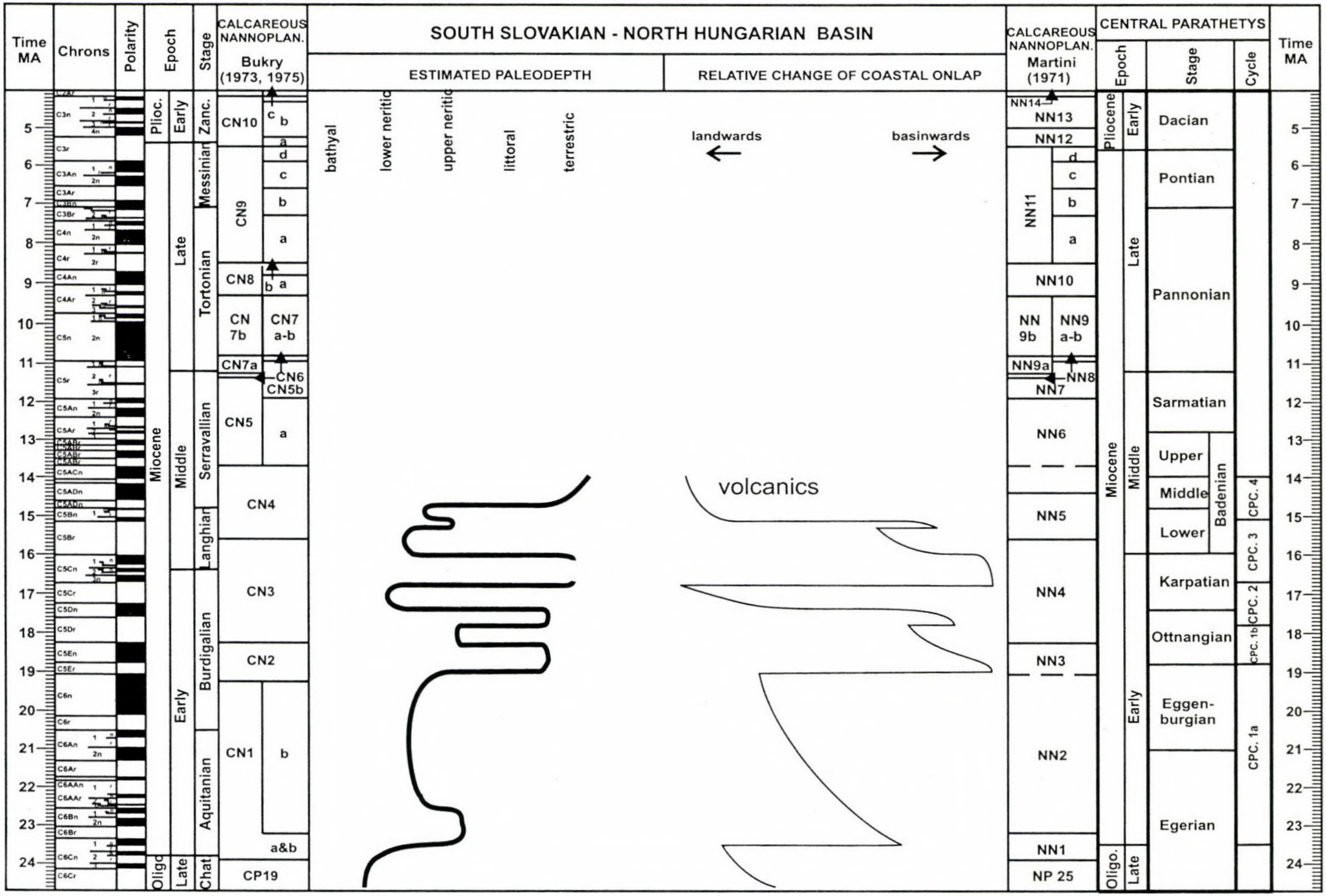


Fig. 5
Estimated paleodepths and coastal onlap in the South Slovakian – North Hungarian Basin

The South Slovakian – North Hungarian Basin (Figs 1, 5) situated in the Western Carpathian hinterland, during the Late Egerian to Early Eggenburgian was a marine embayment (Pétervására Basin; Báldi 1983) connected via the East Slovakian Basin to the residual flysch basins of the Outer Eastern Carpathians (sensu Sztanó 1994). Deltaic sediments of the Opatovce beds with redeposited Early Egerian foraminifera (*Uvigerina hantkeni* Cushman et Edwards, *Globorotalia opima opima* Bolli) are considered to be deposits of the Late Egerian lowstand systems tracts at the basin northern margin (Holcová-Šutovská et al. 1993), similarly to the littoral–brackish sediments near Kovácov and Estergom (Lehotayová in Papp et al. 1975; Sztanó et al. 1998). The Becske and the Andornaktalya beds were deposited at the basin southern margin in Northern Hungary (Hámor and Nagymarosy in Gyalog 1996; Császár 1997).

In the central part of the basin a continuous deposition of the Egerian Szécsény Schlier Formation occurred (= the Egerian Lučenec Formation and the lower part of the Eggenburgian Filakovo Formation, sensu Vass and Elecko 1989). The beginning of the Late Egerian–Eggenburgian period was only demonstrated in offshore facies by the onset of new nannoflora of the NN2 zone. Along with the common species *Coccolithus pelagicus* (Wallich) Schiller and *Cyclicargolithus floridanus* (Roth et Hay) Bukry, *Reticulofenestra pseudoumbilica* (Gartner) Gartner, *Helicosphaera carteri* (Wallich) Kamptner, *H. euphratis* Haq, *H. scissura* Müller and *Discoaster druggii* Bramlette et Wilcoxon also occur (Halásová et al. 1996). The Upper Egerian sediments of the Szécsény Formation contain neritic foraminiferal assemblages similar to those from the older strata, but the Early Eggenburgian new cycle is marked by increased amount of agglutinated species in foraminiferal assemblages and appearing of the *Globigerinoides trilobus* (Reuss) (Halásová et al. 1996). Marginal sandy facies (Tachy, Jalovská Samota and Darmoty Beds) overlying the Széchény Schlier Beds of the Lučenec Formation (Vass and Elecko 1989) is represented by the deposits of shallow-water, hyposaline environment, including tiny individuals of the genera Ammonia, Porosononion and Elphidium. In the Hungarian part of the basin the Szécsény Schlier Formation is partly overlain by the lower part of the Pétervására and the Budafok Formations' sandy marginal facies (Sztanó and Tari 1993).

From the Eggenburgian transgression to the Ottnangian regression

The Eggenburgian to Early Ottnangian relative sea-level rise in the Western Carpathian – North Pannonian region is biostratigraphically dated by the first appearance of the species *Chlamys gigas* (Schlotheim) in the lower part of the NN2 zone and somewhat later by the appearance of the calcareous nannoplankton assemblages with *Helicosphaera ampliaperta* (Bramlette et Wilcoxon) Bukry (FAD 20.5 Ma, sensu Berggren et al. 1995) during the Eggenburgian transgression. The first appearance of *Helicosphaera ampliaperta* marks the beginning of the Burdigalian in the Mediterranean area (Fornaciari and Rio 1996). The end of the

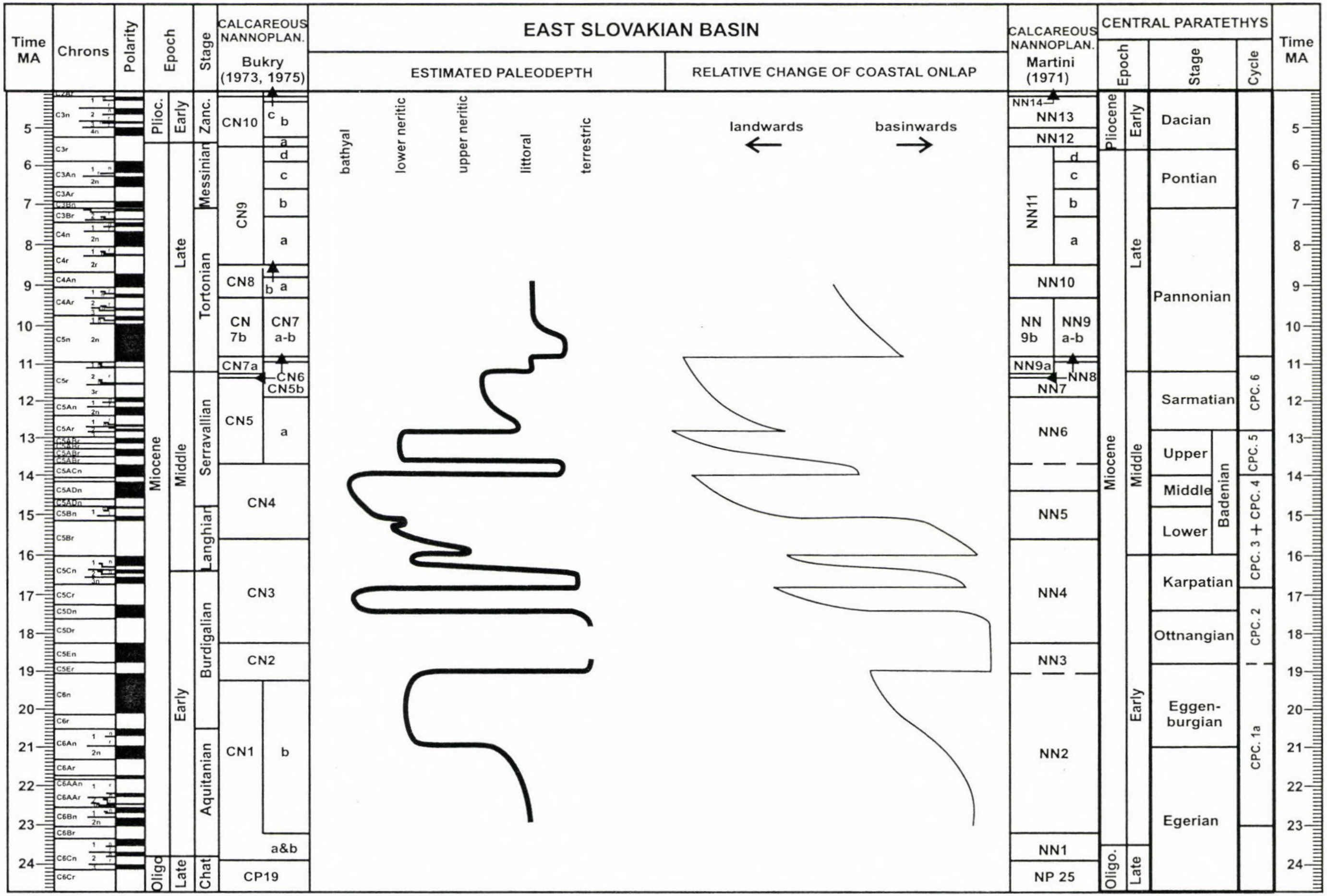


Fig. 6
Estimated paleodepths and coastal onlap in the East Slovakian Basin

cycle is marked by the latest occurrence (LAD) of *Sphenolithus belemnos* Bramlette et Wilcoxon (18.3 Ma, sensu Berggren et al. 1995) in the Upper Ottnangian sediments. The transgression of the next cycle is correlated with the first appearance of the nannoplankton species *Sphenolithus heteromorphus* Deflandre in the Central Paratethys (FAD 17.4 Ma, sensu Rögl 1998), hence in the Mediterranean the first appearance of this species is much earlier (FAD 18.2 Ma, sensu Berggren et al. 1995).

In the South Slovakian – North Hungarian (Pétervására) Basin (Fig. 5) the Eggenburgian transgression is documented by the littoral sandy facies of the Lipovany Beds, with the occurrence of the characteristic mollusc species *Chlamys gigas* (Schlotheim) at the basin's northern margin (Ondrejicková in Steininger and Seneš 1971). In Hungary, the upper part of the Budafok Formation and the Darnó conglomerates as a marginal facies of the Szécsény Schlier Formation also document the Eggenburgian transgression in the south (Báldi 1983, 1997). The character of the Eggenburgian foraminiferal assemblages in the basinal Cakanovce beds of the Filakovo Formation in Slovakia and the upper part of the Szécsény Schlier in Hungary remains stable; nevertheless, the assemblages are more numerous and of normal size. The assemblages in the deepest parts of the basin contain predominantly individuals of the genus Marginulina; in the marginal parts the epiphytic foraminiferal species prevailed. Among the new species the most important is the genus Monspeliensina, documenting the Early Miocene marine connection between the Rhone Basin (where it occurs in the Aquitanian assemblages) and the South Slovakian – North Hungarian Basin. Among the calcareous nannoplankton (in the Filakovo Formation sensu Vass and Elecko 1982), the species *Helicosphaera ampliaperta* (Bramlette et Wilcoxon) Bukry appears and *H. euphratis* Haq fades away. In the Hungarian part of the basin the regressional position of the Pétervására beds, a basinward-retreating sandstone body, already indicates the filling up of the basin during the lower part of the NN 3 nannoplankton zone (Nagymarosy and Báldi-Beke 1988; Sztanó 1994). Before the end of the Eggenburgian a transition from marine to terrestrial environment is documented in the Slovakian part of the basin. The radiometric age of the Bukovinka Formation (Vass et al. 1985) measured on rhyodacite tuffs is above 20.1 ± 0.3 Ma (Repcok 1987). The Zagyvapálfa Formation and the Ottnangian rhyodacite tuffs from Gyulakeszi (19.6 ± 1.4 Ma) are the Hungarian equivalents of the Bukovinka Formation.

Due to tectonic control of the geodynamic evolution of the area at the end of the Eggenburgian and the beginning of the Ottnangian (uplift and subsequent subsidence?) a new relative sea-level cycle appeared in the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5). The Ottnangian Salgótarián Formation contains the lower (coal-bearing) Potor Beds deposited in an alluvial plain environment overlain by the limnic Plachtince beds with marine ingressions in the Slovakian part of the basin (Vass et al. 1987). On the basis of the presence of the calcareous nannoplankton species *Sphenolithus belemnos*

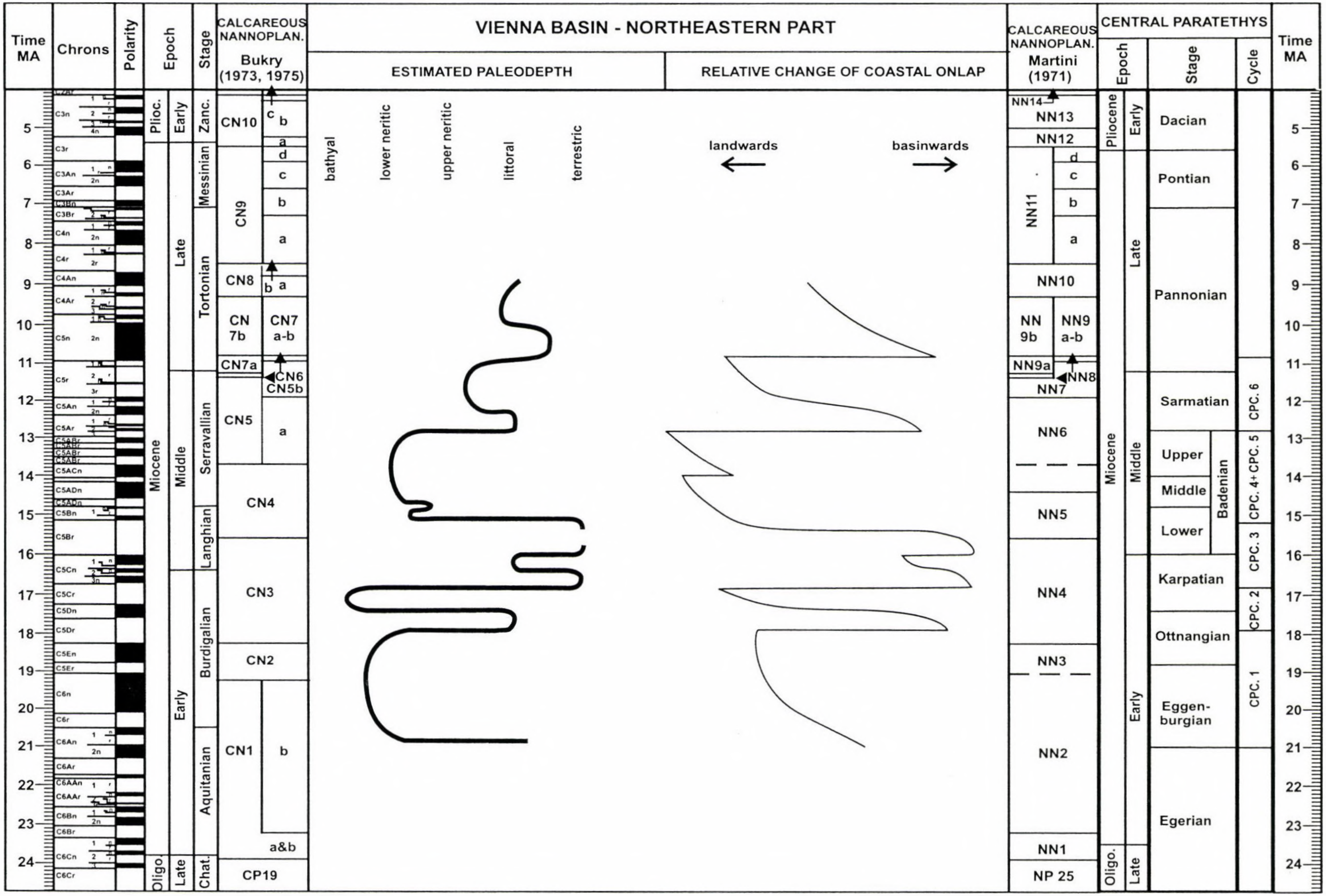


Fig. 7
Estimated paleodepths and coastal onlap in the Vienna Basin

Bramlette et Wilcoxon (Vass et al. 1987), these sediments are attributed to the NN3 zone (sensu Martini 1971). The depocenter of the Ottnangian marine ingressions is more or less identical with that of Eggenburgian time. Foraminiferal assemblages indicate its paleodepth as shallow to deep neritic zone (Vass et al. 1987; Holcová-Sutovšká et al. 1993). In the southern, Hungarian part of the basin the deposition of terrestrial to brackish series of the Kisterényi (limnic to paralic coal seams), Nógrádmegye and Mátranovák Beds occurred simultaneously (Hámor in Gyalog 1996 and in Császár 1997).

In the Bánovce Depression (Fig. 8) the transgressive littoral Eggenburgian sediments contain shallow-water foraminiferal assemblages of Cibicides–Cibicidoides type. The Eggenburgian sea-level rise is characterized by deposits with diversified deep neritic foraminiferal assemblages with prevailing lagenidic taxa (e.g. Lenticulina and Marginulina). They pass, as in the Dobrá Voda Depression, into foraminifera-free dark sediments with abundant fish remnants, which were deposited in an environment with low oxygen content near the bottom of the basin. The considerable content of the thecamoebes in the upper part of the complex documents gradually decreasing salinity (Brestenská 1975, 1977, 1980; Kovác et al. 1991). The determination of the sediment's age as Ottnangian, i.e. the NN3 nannoplankton zone was based on the presence of the species *Sphenolithus belemnos* Deflandre (Lehotayová 1977; Šutovská in Kovác et al. 1991).

The Eggenburgian communication between the South Slovakian – North Hungarian and the Vienna Basins via the Bánovce Depression is documented by foraminiferal assemblages occurring in the sediments below the Badenian volcanics of the Krupinská Vrchovina Upland (Halásová et al. 1996).

In the northern part of the Vienna Basin (Fig. 7) the Eggenburgian transgression is characterized by the occurrence of the species *Chlamys gigas* (Schlotheim) and shallow-water foraminiferal assemblages of Cibicides–Cibicidoides type at the basin margins and in the Váh River Valley (Ctyroký in Buday et al. 1965; Ctyroký in Steininger and Seneš 1971). The open marine conditions during the Eggenburgian to Early Ottnangian are represented by the lower part of the Lužice Formation (Špicka 1969). Stenohaline Batysiphon–Cyclammina foraminiferal assemblages from the western part of the basin dominated by the agglutinated foraminifera *Cyclammina praecancellata* Voloshinova, *C. rotundidorsata* (Hantken), *Haplophragmoides vasiceki vasiceki* Cicha et Zapletalová, *Batysiphon taurinense* Sacco and *Textularia gramen abbreviata* Orbigny indicate a cool, nutrient-rich, upper bathyal marine environment (sensu Murray 1991). The end of this cycle is represented by the deposition of the upper part of the Lužice Formation, including Cibicides–Elphidium-bearing foraminiferal assemblages, indicating further shallowing of the basin during the Ottnangian regression (Kovác and Hudácková 1997).

The erosional remnants of the offshore silts found in a tectonic slice underlying the Manín Unit in the Váh River Valley (Salaj and Zlinská 1991) can be also

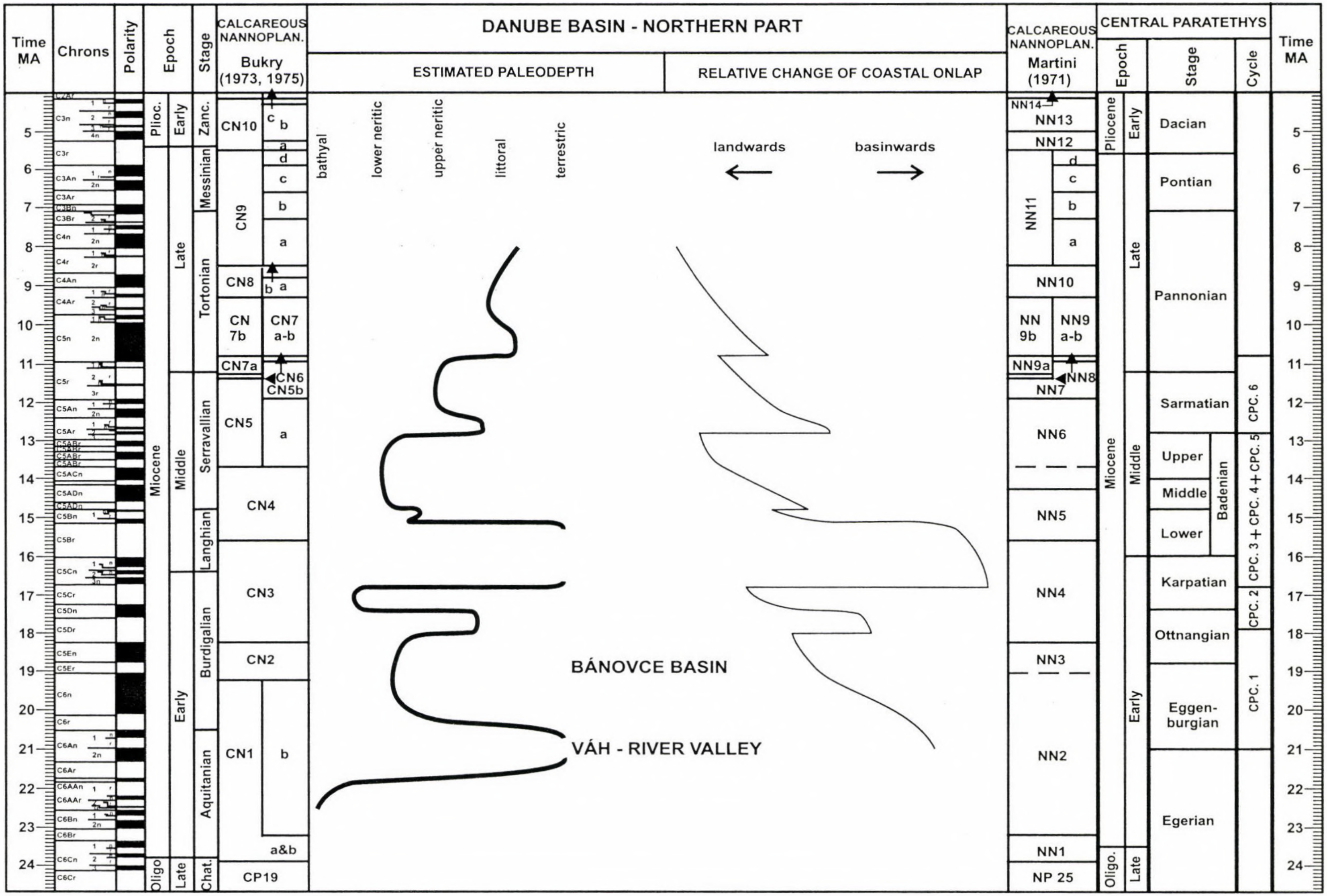


Fig. 8
Estimated paleodepths and coastal onlap in the Danube Basin

attributed to the Eggenburgian highstand period. They contain a foraminiferal assemblage with *Hansenisca soldanii* (Orbigny), *Bulimina elongata* Orbigny, *Oridorsalis umbonatus* (Reuss), *Nodosaria raphanistrum* (Linné), *N. pyrula* Orbigny, *Lagena arcuatostriata* Reuss, *Hanzawaia boueana* (Orbigny), *Lenticulina cultrata* (Montfort), *L. totomiensis* Makiyama, *L.* cf. *inornata* (Orbigny), *Pseudonodosaria* sp., *Heterolepa dutemplei* (Orbigny), *Caucasina schischkinskayae* (Samoilova), *Lagenodosaria* sp. and *Cibicides ungerianus* (Orbigny) reflecting an outer shelf deep neritic environment.

In the East Slovakian Basin (Fig. 6) the Biely Potok Formation is unconformably overlain by the Eggenburgian deposits of the Prešov Formation (Vass and Cvercko 1985). They contain calcareous nannoplankton assemblages with *Helicosphaera scissura* Müller, and *H. carteri* (Wallich) Kamptner also appears (Zlinská 1996a). The absence of the species *Helicosphaera ampliaperta* (Bramlette et Wilcoxon) Bukry (Šutovská in Vass et al. 1993) led us to the conclusion that the beginning of sedimentation can be correlated with the lower part of the NN2 nannoplankton zone, i.e. the sequence (resp. its lower part) might have been deposited during the Late Egerian. The foraminiferal assemblage with *Cyclammina acutidorsata* (Hantken), *Pappina bononiensis primiformis* (Papp et Turnovsky), *Lenticulina arcuatostriata* (Hantken) and *Spirolutilus carinatus* (Orbigny) indicates depositional environment in the outer shelf neritic zone (Zlinská 1992b) and the presence of the species *Pappina bononiensis primiformis* (Papp et Turnovsky) indicates the Eggenburgian age.

The Celovce Formation, a lateral equivalent of the upper part of the Prešov Fm., is characterized by a rapid input of coarse clastics associated with tectonic uplift of the basin margins. Basin isolation led to salinity decrease, which is documented by foraminiferal assemblages dominated by *Ammonia beccarii* (Linné), *Porosononion* aff. *subgranosum* (Egger), *Nonion commune* (Orbigny) and *Elphidium* sp. The presence of the Egerian and the Eggenburgian species inlittoral to neritic zones: *Cibicidoides budayi* (Cicha et Zapletalová), *Lenticulina meznericsae* (Cicha), *Bulimina elongata* Orbigny and *Uvigerina hantkeni* Cushman in thanatocenoses, displays an erosion of older and contemporary sediments; hence a redeposition of the foraminiferal tests during the relative sea-level fall resulted from the tectonic evolution of the area. During the Ottnangian a depositional break is assumed in the East Slovakian Basin (Rudinec 1978, 1989), being interpreted as a consequence of the compressional tectonic regime connected with further uplift of the active margin of the Western Carpathians.

The 3rd order CPC 1a relative sea-level cycle started ca. 23 Ma (23.5–18.8 Ma) ago in the Western Carpathian – North Pannonian region and cannot be correlated with the TB 1.5 cycle of the global sea-level change (22–21 Ma, sensu Haq et al. 1988 and Haq 1991). We can conclude that the Late Egerian–Eggenburgian relative sea-level cycle CPC 1a in the Western Carpathian – North Pannonian region is observable in the South Slovakian – North Hungarian (Pétervására) Basin and the East Slovakian Basin (Figs 5, 6). In the Vienna and the

Bánovce Basins (Figs 7, 8) the Eggenburgian–Ottnangian cycle CPC 1 (21–17,8Ma) is documented (not in the case when the Eggenburgian contains the lower part of the NN 3 zone); moreover, in the South Slovakian – North Hungarian (Novohrad) Basin the next Ottnangian regional sea-level cycle is observed (CPC 1b 18,8–17,8 Ma).

Based on recent knowledge from the Central Paratethys, the Eggenburgian–Ottnangian relative sea-level changes in the Western Carpathian – North Pannonian Basins (CPC 1, CPC 1a and CPC 1b) may be partly correlated with the TB 2.1 cycle of the global changes (21 to 17.5 Ma, sensu Haq et al. 1988; Haq 1991), where the maximum flooding was estimated as 18.5 Ma, and would be reflected in the Ottnangian marine ingressions during the tectonically-controlled development of the Novohrad Basin.

Late Ottnangian to Early Karpatian cycle of relative sea-level changes

The time interval 17.8?–17.5 Ma, represented by the Middle Burdigalian–latest Ottnangian–latest Early Kotsakhurian stages (sensu Rögl 1998) is defined by the lower part of the NN4 zone (sensu Martini 1971). The time section 17.5 to 16.4 Ma represents the Late Burdigalian–Early Karpatian s.s. – Late Kotsakhurian stages (sensu Berggren et al. 1995; Rögl 1998). It is defined inside the NN4 nannoplankton zone (sensu Martini 1971), which can be correlated with the M4 planktonic foraminifera zone (sensu Berggren et al. 1995).

In the Late Ottnangian the shallow-water brackish facies with endemic mollusc fauna, the so-called "Oncophora or Rzehakia Beds" known from Bavaria to Northern Hungary (Rögl 1998) became widespread in the Paratethys at the beginning of the NN4 nannoplankton zones. From the viewpoint of sequence stratigraphy they are considered to be a consequence of the global eustatic sea-level fall. The Early Karpatian transgression (Fig. 4) during the NN4 nannoplankton zone is well documented by immigration of new foraminiferal assemblages and mollusc fauna different from the Ottnangian ones (Rögl 1998). The restored western seaway of the Central Paratethys to the Mediterranean (the "Transdinarid or Slovenian Corridor" Rögl and Steininger 1983) was fully demonstrated by the appearance of the foraminifera genus *Uvigerina graciliformis* Papp et Turnovsky. Migration of foraminiferal assemblages can be traced from Slovenia, via the Styrian Basin, the Central Transdanubian belt and the Novohrad, Vienna and the East Slovakian Basins, as far as the Eastern Alpine and the Western Carpathian Foredeeps (Kovác et al. 1993; Kovác and Hudácková 1997; Kovác and Zlinská 1998). In the Carpathian Foredeep an alternation of marine, brackish and terrestrial sequences is present (marine in the west, than the brackish and terrestrial facies increases eastwards).

The base of the Late Ottnangian–Early Karpatian relative sea-level cycle in the Western Carpathian – North Pannonian region is biostratigraphically defined by the latest occurrence (LAD) of *Sphenolithus belemnos* Bramlette et Wilcoxon (18.3

Ma, sensu Berggren et al. 1995) and the first appearance (FAD) of *Sphenolithus heteromorphus* Deflandre, further by the first appearance of *Uvigerina graciliformis* Papp et Turnovsky within the NN4 nannoplankton zone (17.4 Ma, sensu Rögl 1998). The transgression of the next cycle is marked by the FAD of genus Praeorbulina (16.4 Ma, sensu Berggren et al. 1995).

The LAD of *Sphenolithus belemnos* Bramlette et Wilcoxon defines the beginning of the NN4 zone at the level of 18.3 Ma (Berggren et al. 1995) and the FAD of *Sphenolithus heteromorphus* Deflandre defines the base of the CN 3 zone (sensu Okada and Bukry 1980). These zones were, however, not demonstrated in the Western Carpathian area. Nevertheless, on the basis of the observations of Nagymarosy a common occurrence of both mentioned nannoplankton species has been observed in the Ottnangian Bántapuszta beds (Kókay 1973) in the Transdanubian Central range, as well as the individual occurrence of the species *Sphenolithus heteromorphus* Deflandre in the Late Ottnangian marine sediments of Northern Hungary. This fact suggests that nannoplankton did not reflect the restricted communication between the Mediterranean and the Central Paratethys; hence the sea-level fall at the end of the cycle influenced only migration of foraminifera that reached our territory as late as during the Karpatian transgression.

During the late highstand of the Eggenburgian–Ottnangian cycle Ottnangian anoxic sediments appeared in the Carpathian–Pannonian region, containing the NN3 nannoplankton zone with Sphenolithus belemnos Bramlette et Wilcoxon. These beds, known from the Bánovce Depression (see above), appear in the same stratigraphic level along the eastern border of the Bohemian Massif in Lower Austria (Zellendorf Formation, Upper Eggenburgian–Lower Ottnangian), and are represented by anoxic/carbonate free sediments with many fish scales (Steininger et al. 1991). They are somewhat older than the shallow-water "Oncophora or Rzehakia Beds" of the Alpine–Carpathian Foredeep and the intra-Carpathian area containing nannoplankton of the NN4 zone (Horváth and Nagymarosy 1978; Báldi-Beke and Nagymarosy 1979) and represent sediments of the lowstand depositional system (Ctyroký 1968). Based on their geologic position they are also interpreted as the beginning of the 3rd order eustatic cycle (TB 2.2), the transgressive part of which is characterized by the onset of foraminiferal assemblages with Uvigerina graciliformis Papp et Turnovsky (Horváth and Nagymarosy 1978; Holcová, in prep.) within the "Oncophora or Rzehakia Beds" in the Western Carpathian hinterland.

The "Oncophora or Rzehakia Beds" in the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5) were usually mentioned as Upper Ottnangian in age (Papp et al. 1973). Along with *Rzehakia socialis* (Rzehak), however, they already contain the foraminiferal assemblage with *Uvigerina graciliformis* Papp et Turnovsky, the Karpatian index fossil for the Central Paratethys (Cicha et al. 1983). Moreover, the beds transgressively overlie the underlying Ottnangian deposits. In the autochthonous part of the assemblage the genus *Ammonia* predominates (Holcová 1996a).

From the paleogeographical point of view it is interesting that the compositions of both suspension-transported and autochthonous foraminiferal assemblages are similar to the "Oncophora or Rzehakia Beds" of the Upper Austrian Molasse (Holcová, in prep.). The latter, due to their regressive position, represent the relative sea-level fall at the end of the previous Eggenburgian–Ottnangian cycle, as well as the lowstand deposits at the beginning of the latest Ottnangian–Karpatian cycle in the Western Carpathian – North Pannonian area. The communication between both mentioned areas is also documented by a reappearance of the genus *Monspeliensina* in the "Oncophora or Rzehakia Beds" of the Novohrad Basin (Holcová 1996b), known also from the underlying Treubach Sands of Upper Austria (Rupp, pers. comm. 1996). Foraminiferal assemblages from the overlying Karpatian deposits of the Strháre – Trenc Graben indicate the depositional environment of a shallow bathyal zone (genera *Pullenia, Chillostomella, Hoeglundina, Melonis*) with a great number of planktonic foraminifera (Zlinská and Šutovská 1990).

In the Bánovce Depression as well as in the Danube Basin (Fig. 8) the first sign of the Karpatian transgression or sea-level rise (compared the beginning of the TB 2.2 cycle of the global sea-level changes sensu Haq 1991) was a change of the Ottnangian sedimentary environment with decreased oxygen level into the environment with good circulation and aeration. This event still appeared before the immigration of the foraminiferal assemblages containing *Uvigerina graciliformis* Papp et Turnovsky (Brestenská 1980).

The Karpatian foraminiferal assemblages, in deposits transgressively overlying the marine Ottnangian sediments, are characterized by a dominance of deep neritic rotaliid foraminifera (mainly Lagenida and Cibicidoides). On the other hand the assemblages dominated by deep neritic agglutinated forms (genus Cyclammina) occur in the Karpatian deposits directly overlying the pre-Neogene basement of the Bánovce Depression and the Danube Basin. The majority of the assemblages is rich in redeposited Cretaceous and Paleogene material (Kovác et al. 1999).

In the Vienna Basin (Fig. 7) the Late Ottnangian–Early Karpatian lowstand is represented by the Štefanov Sand, littoral sediments with a predominance of the foraminiferal assemblages containing Ammonia (Kovác and Hudácková 1997). Our opinion is that the beds can be correlated with the "Ammonia Beds" from the Bánovce Depression and the "Upper Ottnangian" sediments from the Novohrad Basin with *Uvigerina graciliformis* Papp et Turnovsky. The overlying Karpatian marine "schlier" sediments (Lakšár Formation; Špicka 1969) contain highly-diversified foraminiferal assemblages rich in *Uvigerina graciliformis* Papp et Turnovsky, *Pappina bononiensis primiformis* (Papp et Turnovsky), *P. parkeri breviformis* (Papp et Turnovsky), *Cyclammina karpatica* Cicha et Zapletalová and *Reticulophragmium karpaticum* Cicha et Zapletalová. They represent deep-water sediments of the shallow bathyal zone (Zlinská 1994; Kovác and Hudácková 1997).

The Karpatian foraminiferal assemblages in the Vienna Basin (Kovác and Hudácková 1997) indicate the same paleoecological conditions as the Eggenburgian ones with the genera Cyclammina and Bathysiphon (i.e. cool, well-oxygenated deep-water environment rich in nutrients, sensu Murray 1991). They are considered to be assemblages living during the sea-level highstand or occurring in a basin with considerable paleodepth. A major southward shift of the Karpatian depocenters of the Vienna Basin (Jirícek and Seifert 1990), as well as the high sedimentation rate, show a dominant tectonic influence on the subsidence and evolution of the depositional systems during this period (Lankreijer et al. 1995; Kovác et al. 1997). It is important to note that the Vienna and Danube Basins and the Bánovce Depression formed a common water-circulation system (basin) during the Karpatian (Brzobohatý 1987; Brestenská 1980; Kovác et al. 1993).

In the East Slovakian Basin (Fig. 6) the (latest Ottnangian?) Karpatian cycle of the relative sea-level changes began with the transgression of the Teriakovce Formation (Vass and Cvercko 1985). The Karpatian marine microfauna documents open-marine conditions of deep neritic to shallow bathyal zone during the transgression and sea level highstand. The predominant species are *Uvigerina graciliformis* Papp et Turnovsky, *Pappina parkeri breviformis* (Papp et Turnovsky), *P. bononiensis primiformis* (Papp et Turnovsky), *Lenticulina calcar* (Linné), *L. inornata* (Orbigny), *L. cultrata* (Montfort) and planktonic forms such as *Globigerina ottnangiensis* Roegl and *G. praebulloides* (Blow) (Zlinská 1992b). The overlying Solná Bana Formation represents an evaporitic event in the period of basin starvation and isolation during the late HST sea level fall (Kovác and Zlinska 1998). Assemblages of foraminifera, dominated by *Ammonia beccarii* (Linné), *Bulimina elongata* Orbigny, *B. striata* Orbigny, *B. pupoides* Orbigny, *Bolivina dilatata* Reuss, *Uvigerina graciliformis* Papp et Turnovsky, *Elphidium macellum* (Fichtell et Moll), *E. fichtelianum* (Orbigny), *Nonion pompilioides* (Fichtell et Moll) and *Globigerina bulloides* Orbigny document low oxygen content at the basin bottom and gradual transition to hypersaline environment (Zlinská 1992b). This opinion is also supported by the study of evaporites (Karoli et al. 1997), where part of the halite crystals originated at the brine/air contact ("hopper crystals") and subsequently sank to the bottom of the basin with its gradually shallowing environment. In the final stage a gradual desiccation of the basin occurred, accompanied by rupture deformations of the unlithified halite crusts during dewatering.

The Karpatian foraminiferal assemblages in the Vienna, Danube, Novohrad and the East Slovakian Basins uniformly document an accelerated deepening of the sedimentary areas (Figs 5, 6, 7, 8). The deepening was induced both tectonically and by transgression related to the sea-level rise at the beginning of a new eustatic cycle. We can also conclude that the Late Ottnangian–Early Karpatian CPC 2 (17.8–16.8 Ma) relative sea-level cycle in the Western Carpathian

– North Pannonian area is partly correlatable with the TB 2.2 cycle of the global sea-level changes (17.5–16.5 Ma, sensu Haq et al. 1988; Haq 1991).

Late Karpatian – Early Badenian cycle of relative sea-level changes

The 16.4–15.1 Ma time interval, represented by the Langhian–Late Karpatian s.l. to Early Badenian – Tarkhanian stages (sensu Berggren et al. 1995; Rögl 1998) is defined by the presence of the upper part of the NN4 and the lower part of the NN5 nannoplankton zones (sensu Martini 1971). Its lower boundary is biostratigraphically defined by the FAD of *Globigerinoides bisphaericus* Todd and somewhat later by that of the genus *Praeorbulina*. The upper boundary of the cycle is marked by the FAD of the genus Orbulina. According to Berggren et al. (1995) it is a period which might be correlatable with the TB 2.3 cycle of the global sea-level changes (sensu Haq 1991).

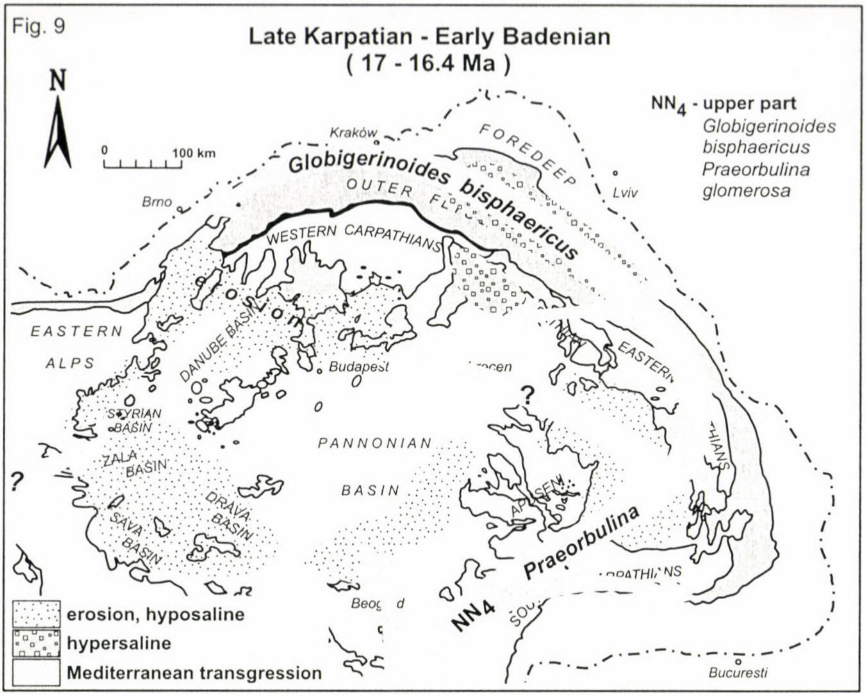
In the Eastern Carpathian Foredeep an onset of the foraminiferal assemblages with *Globigerinoides bisphericus* Todd within the Tarkhanian stage, the time span of which is correlated with the upper part of the NN4 and the lower part of the NN5 nannoplankton zones (Nosovsky et al. 1976; Goncharova 1989; Andreeva-Grigorovich et al. 1997; Rögl 1998), can be interpreted as a new Karpatian (Langhian) transgression documenting the activity of the Central Paratethys connections to the Eastern Mediterranean (Fig. 9). The onset of the genus Praeorbulina within the Karpatian has also been mentioned by Popescu (1998), who found the species *Candorbulina (=Praeorbulina) glomerosa* Blow and *Globigerinoides bisphericus* Todd in the Transylvanian Basin and the Eastern Carpathian Foredeep in Romania. His statement supports the hypothesis on the time span of the Karpatian stage in the Central Paratethys (Figs 5, 6, 7, 8), i.e. it would reach the Mediterranean Langhian stage in its upper part (Cicha et al. 1998).

In the area of the East Alpine – Western Carpathian Foredeep, along with *Praeorbulina sicana* de Stefani (which represents *G. bisphericus* Todd; Cicha et al. 1998; Salaj, pers. comm.), Praeorbulina appears in the foraminiferal assemblages for the first time in the upper part of the NN4 zone. Sediments with Praeorbulina are described as belonging to regressive deposits (lowstand wedge) at the Karpatian/Badenian boundary (Cicha 1995). This fact, together with the occurrences of this species in the Transylvanian Basin and in the Eastern Carpathian Foredeep, indicate a connection between the Central Paratethys and

Fig. 9 →
Late Karpatian isolation and erosion during the relative sea-level fall and faunal migration pathways during the Late Karpatian–Early Badenian (Langhian) transgression

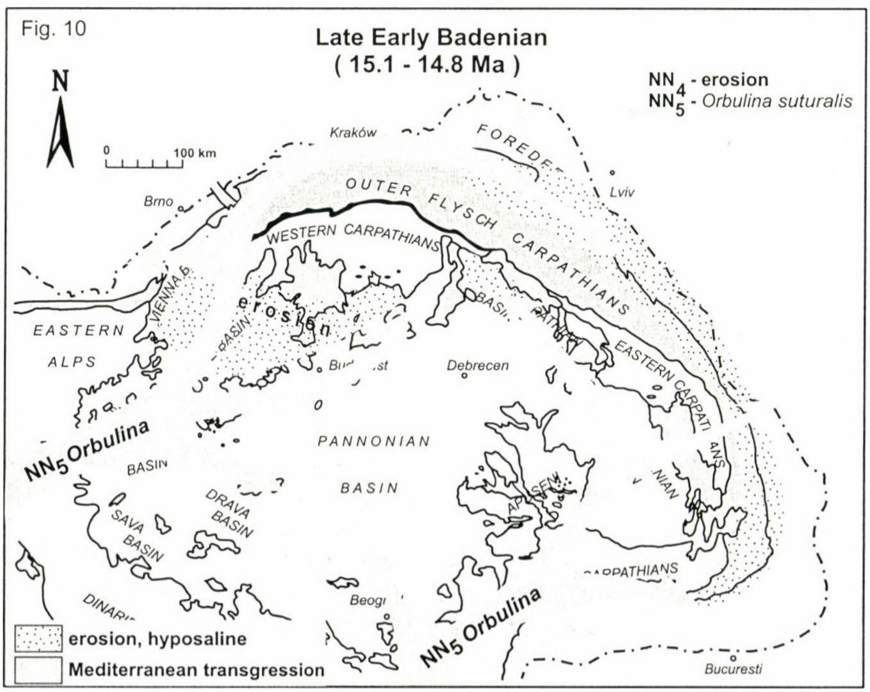
Fig. 10 →
Early Badenian isolation and erosion during the relative sea-level fall and faunal migration pathways during the late Early Badenian transgression

Fig. 9



Late Karpatian - Early Badenian
(17 - 16.4 Ma)
N
0 100 km
NN_4 - upper part
Globigerinoides bisphaericus
Praeorbulina glomerosa
Kraków
FOREDEEP
Globigerinoides bisphaericus
OUTER
Lviv
Brno
WESTERN CARPATHIANS
erosion
EASTERN ALPS
DANUBE BASIN
Budapest
STYRIAN BASIN
ZALA BASIN
PANNONIAN BASIN
DRAVA BASIN
SAVA BASIN
Praeorbulina
NN4
Bucuresti
erosion, hyposaline
hypersaline
Mediterranean transgression

Fig. 10



Late Early Badenian
(15.1 - 14.8 Ma)
N
0 100 km
NN_4 - erosion
NN_5 - Orbulina suturalis
Kraków
Lviv
Brno
OUTER FLYSCH CARPATHIANS
WESTERN CARPATHIANS
EASTERN CARPATHIANS
erosion
EASTERN ALPS
Debrecen
NN_5 Orbulina
PANNONIAN BASIN
DRAVA BASIN
SAVA BASIN
NN_5 Orbulina
Bucuresti
erosion, hyposaline
Mediterranean transgression

the Mediterranean, still during the Karpatian and gradually also during the Early Badenian transgression (Fig. 10).

In the East Slovakian Basin (Fig. 6) the evaporites of the Solná Bana Formation are overlain by the Kladzany Formation, containing mixed foraminiferal associations (Kovác and Zlinská 1998). They are predominantly represented by shallow-water, inner shelf species tolerant to salinity changes, e.g. *Elphidium macellum* (Fichtell et Moll), *E. fichtelianum* (Orbigny), *Ammonia beccarii* (Linné), but stenohaline, outer shelf forms like *Uvigerina graciliformis* Papp et Turnovsky, *Cyclammina karpatica* Cicha et Zapletalová and *Globigerinoides sicanus* de Stefani are also present (Zlinská 1992b). The mixed foraminiferal assemblages, together with the facies of the sediments (extensive sediment input from the coastal area), indicate deposition during the beginning of the sea-level rise, which can be already correlated with the beginning of the Langhian TB 2.3 global sea-level cycle (sensu Haq et al. 1988; Haq 1991).

In the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5) the sediments containing *Praeorbulina* sp., together with the redeposited Karpatian foraminifera species, are considered by Holcová et al. (1996a) to be an indicator of a transgression, correlatable with the transgressive systems tracts of the TB 2.3 global cycle (sensu Haq et al. 1988; Haq 1991).

The Late Karpatian horizon with *Globigerinoides bisphericus* Todd or Early Badenian *O. sicana* de Stefani (Rögl 1986; Cicha et al. 1998) is missing in the Slovakian Vienna and the Danube Basins (Figs 7, 8). An extensive erosion of the area is supposed here due to the tectonic activity at the end of the Karpatian (uplift?), synchronous with an important global sea-level fall at the base of the TB 2.3 global sea-level cycle (sensu Haq et al. 1988; Haq 1991). Then the Jablonica Conglomerate represents a regressive delta deposited in the Late Karpatian – Early Badenian lowstand wedge (BFF), comparable with depositional systems tracts of the TB 2.3 cycle (sensu Haq et al. 1988; Haq 1991).

The hypothesis of the widespread erosion in the Western Carpathians is also supported by some other facts, such as clay mineral analysis of the Karpatian fill in the South Slovakian – North Hungarian (Novohrad) Basin which indicates large-scale erosion of the Karpatian sediments (Vass and Šucha 1994) before deposition of the Lower Badenian sequences, which are compared with deposits of the TB 2.3 and the TB 2.4 cycles of the global sea-level changes (Upper Lagenide Zone). The reliability of the mentioned assumption is also increased by the residual presence of the deposits that are correlatable with the TB 2.3 cycle in this area (sediments with the genus Praeorbulina). Similar, though unpublished, are the data also coming also from the Vienna Basin (Francu, pers. comm.), which suggest that the Early Karpatian sediments of the Dobrá Voda Depression had to be buried at least under a 600 m-thick sedimentary cover.

As it has been mentioned before, in most of the Western Carpathian basins (except for the Alpine–Carpathian Foredeep) the Late Karpatian horizon with *Globigerinoides bishaericus* Todd or the Early Badenian horizon biostratigraphically defined by the presence of the planktonic foraminifera Praeorbulina and the absence of the genus Orbulina (16.4–15.1 Ma, sensu Berggren et al. 1995) is missing. This fact displays a gap between the latest Karpatian and the beginning of the late Early Badenian, probably caused by uplift and erosion due to local tectonic events.

The CPC 3 cycle of the relative sea-level changes, which can be correlated approximately with the TB 2. 3 global cycle (16.5–15.5 Ma, mfs 16 Ma, sensu Haq et al. 1988; Haq 1991), is traced only in the Alpine–Carpathian Foredeep and the Novohrad Basin. In the Vienna, Danube and the East Slovakian Basins it is not yet proved; hence in the Vienna and Danube Basins there are regressive deposits of this age flooded by transgression of the CPC 4 cycle; in the East Slovakian Basin a similar cycle occurs, but somewhat earlier due to increased tectonic activity and subsidence.

Late Early Badenian to Middle Badenian and Late Badenian to Early Sarmatian relative sea-level cycles

The Late Langhian and the Serravalian–late Early Badenian (Moravian), Middle Badenian (Wieliczkian) and the Late Badenian (Kosovian) – Tschokrakian, Karaganian and the Konkian stages represent the time interval 15.1– 12.8? Ma (sensu Berggren et al. 1995; Rögl 1998). Its lower boundary is biostratigraphically defined by the FAD of the genus Orbulina 15.1 Ma ago (Berggren et al. 1995; Fornaciari and Rio 1996). The Middle Badenian is further defined by the presence of the NN5 and the lowermost part of the NN6 nannoplankton zones; the Late Badenian is defined by the presence of the NN6 zone (sensu Martini 1971; Berggren et al. 1995). Some sporadic data in the Mecsek Mts, Hungary (Nagymarosy 1985) and in the Transylvanian basin (Mészáros, Marunteanu) suggest, that the top beds of the Badenian comprise also the lowermost part of NN7 nannoplakton zone. The upper chronostratigraphic boundary of the Badenian (Vass et al. 1985; Steininger et al. 1996) was estimated at 13.6 ± 0.2 Ma for the Central Paratethys, based upon radiometric data.

The time span of the NN5 zone is defined as 15.6–13.6 Ma according to Berggren et al. (1995), and is defined by the LAD of the species *Sphenolithus heteromorphus* Deflandre. This would give an age too young for the Central Paratethys. However, Aubry (in Berggren et al. 1995) admitted the possibility of older age of LAD for the species *Sphenolithus heteromorphus* Deflandre in the Paratethys; he placed the NN5/NN6 zone boundary at 14.4 Ma. Rögl (1998) even shifted this boundary somewhat "deeper", i.e. he placed the Early/Middle Badenian boundary to ca. 15 Ma, the Middle/Late Badenian boundary at ca. 14 Ma and the Badenian/Sarmatian boundary at 13 Ma, which is more convenient from the point of view of our study.

The late Early Badenian (Fig. 10) marine transgression was characterized by a uniform tropical-subtropical mollusc and foraminifera fauna in the Carpathian

Foredeep and in the Inner Carpathian area as well, from the Vienna Basin as far as the Transylvanian Basin. The Central Paratethys had a sea connection with the Mediterranean (Studencká et al. 1998). The western connection is represented by a restored seaway (after the Karpatian) via the "Transdinarid-Slovenian Corridor" (Rögl and Steininger 1983). There are several hypotheses about the eastern connection (Rögl 1998).

The closure of the eastern seaway during the "Middle Badenian" led to the "salinity crisis". Thick evaporitic deposits, namely halite and gypsum, covered the foredeep of the Western and Eastern Carpathians, the Transylvanian and the Transcarpathian Basins (Ney et al. 1974; Sandulescu 1988). This regional sea-level fall of the Paratethys might be correlated with the end of the TB 2.4 global sea-level cycle (sensu Haq et al. 1988; Haq 1991). In this case, we should take the Middle/Late Badenian boundary 14 Ma (sensu Rögl 1998) into consideration. It is necessary to emphasize here that some of the evaporites in the Western Carpathian Foredeep (Poland) are probably younger (Gazdzicka 1994) and may also partially represent a sea-level fall at the end of the TB 2.5 cycle (sensu Haq et al. 1988; Haq 1991).

Though the TB 2.5 represents just a short time section, it is important from the viewpoint of the Central Paratethys evolution. At this time, the latest marine flooding event took place, covering almost the entire Carpathian–Pannonian area (Fig. 11). Since the "Transdinarid–Slovenian Corridor" was definitively closed (Rögl and Steininger 1983), an extensive reopening of the marine connection with the Eastern Mediterranean is supposed (Nevesskaya et al. 1984, 1987; Studencká et al. 1998). At the same time it is necessary to point out that for most of this time the Eastern Paratethys basins were covered by a brackish water mass; thus marine euhaline faunas may not be derived from this region. Therefore the connection between the Eastern and Central Paratethys was interpreted in the opposite manner by Kókay, i.e. the Central Paratethys would be the source of the marine transgression and faunal migration toward the Eastern Paratethys (Kókay 1984). The new Late Badenian transgression documents the onset of a new fauna in the Eastern Carpathian Foredeep and the Transylvanian Basin, where radiolaria-bearing shale and Spirialis Marl (pteropoda marl) were deposited upon the Middle Badenian evaporites. It is important to point out that the calcareous nannoplankton, Diatomaceae and pteropods are related to the Indo-Pacific province; such radiolaria- and pteropod-rich sediments are not known from the Western Mediterranean area (Dumitrica et al. 1975; Rögl and Müller 1976; Popescu 1979).

Since the Early Badenian fairly uniform paleogeographic conditions occurred in the West Carpathian–North Pannonian region. It led to the possibility that ecostratigraphy, mainly the application of Grill's (1941, 1943) biozones, might become important in biostratigraphy.

The late Early Badenian deposits of the intramontane basins of the Western Carpathian – North Pannonian region are characterized by maximum species diversity and evolution of foraminiferal assemblages of the Upper Lagenide Zone (Grill 1941, 1943), reflecting convenient subtropical climatic conditions and stability of the marine environment. The overlying Middle Badenian sediments are characterized by assemblages of the *Spiroplectammina (=Spirolutilus) carinata* zone (Grill 1941; Zlinská and Ctyroká 1993; Zlinská 1993). The existence of the marine seaway via the "Transdinarid Corridor", connecting the Pannonian and the Venetian Basins, is assumed to have survived until the latest Middle Badenian; it is also considered to be a reason why occurrences of the evaporites are situated only in the northern and eastern portions of the Central Paratethys (Rögl and Steininger 1983). The "Middle Badenian sea-level fall" can be also correlated with erosion and hiatus, which is observed in many seismic sections in the Pannonian Basin as a major regional unconformity (Horváth 1995).

Above the *Spiroplectammina carinata* zone (Grill 1941), the Upper Badenian sediments of the Bulimina–Bolivina Zone were deposited (Grill 1941). Whereas in the Vienna and the Danube Basins the transition from the Spiroplectammina to the Bulimina–Bolivina zone is gradual with an almost indeterminable boundary and unobservable shallowing in the basin, in the East Slovakian Basin the Middle and the Late Badenian biozones are separated by shallow-water, lagoonal evaporites of the Zbudza Formation

In the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5) the sequences dominated by Praeorbulina are overlain by marine deposits of the Vinica Formation with *Orbulina suturalis* Brönnimann (onset at 15.1 Ma). On the basis of the basin evolution, where the Early Badenian depocenters were shifted westward with respect to the Karpatian ones, together with the transgressive character of the deposits we can conclude that they represent depositional systems which can be correlated with the TB 2.4 global sea-level cycle (sensu Haq et al. 1988; Haq 1991). Unlike the underlying beds with Praeorbulina, the Vinica Formation marine deposits contain rich foraminiferal assemblages. In the western, deeper part of the basin, highly diversified assemblages occur, dominated by the epiphytic species *Lobatula lobatula* Walker et Jacobs, *Pararotalia stellata* (Reuss), and *Asterigerinata planorbis* Orbigny. Marine sedimentation in the Novohrad Basin ended after the Early Badenian; later only an accumulation of products of the Central Slovakian neovolcanics continued at its northern margin.

In the Danube Basin (Fig. 8) the sediments of the Upper Lagenide Zone are restricted to its southeastern and southwestern parts. Index species the NN5 nannoplankton zone, *Uvigerina macrocarinata* Papp et Turnovsky and the planktonic *Orbulina suturalis* Brönimann are the age indicators. The assemblages are deep neritic with predominance of the genus Lenticulina. In the Bajtava Formation of the Želiezovce Depression, similarly as in the Novohrad Basin, the genus Praeorbulina occurs along with Orbulina (Zlinská 1996b). The foraminiferal assemblage reflects the deep neritic to shallow bathyal environment (Zlinská et al. 1997). Based on the aforementioned facts and on the high diversity of the assemblages with high plankton/benthos ratio as well, we correlate the sediments of this area with the transgressive systems tract of the TB 2.4 cycle of the global sea-level changes (sensu Haq et al. 1988; Haq 1991). The younger deposits with *Spiroplectammina carinata* Zone (Grill 1941) play a major role in the Danube Basin, with their areal extent widely exceeding that of the older, Lower Badenian sediments. The diversity of the assemblages is similar to that in the Vienna Basin. The assemblages gradually pass into the Bulimina–Bolivina Biozone (Kovác et al. 1999).

In the northeastern part of the Vienna Basin (Fig. 7), similarly as in the Danube Basin, Early Badenian sedimentation starts within the Upper Lagenide Zone (Grill 1943; Cicha et al. 1975). This period can be correlated with the first appearance of the genus Orbulina (15.1 Ma). The prevailing sedimentary environment of the Lanžhot Formation (Špicka 1969) represented the neritic zone characterized by the foraminiferal assemblage with *Lenticulina echinata* (Orbigny), *L. cultrata* (Montfort), *Planularia antillea ostraviensis* Vašícek, *P. dentata* Karrer, *Vaginulina legumen* (Linné) and *Uvigerina macrocarinata* Papp et Turnovsky. The geophysical study of the Vienna Basin documents the Early Badenian lowstand systems tract transiting into the transgressive system (Hudácková et al. 1997). Agglutinated foraminifera of the overlying Middle Badenian *Spiroplectammina carinata* biozone (Grill 1941) in the Vienna Basin display the euryhaline neritic environment of the Jakubov Formation (Špicka 1969). They contain mainly the species *Cyclammina pleschakowi* Pishwanova, *Spirolutilus carinatus* (Orbigny), *Martinotiella communis* (Orbigny), *Textularia gramen* Orbigny, and *Haplophragmoides vasiceki vasiceki* Cicha et Zapletalová (Hudácková and Kovác 1993). Geophysical indications of the Upper Badenian depositional systems tracts unambiguously document the sea-level highstand deposits (Hudácková et al. 1997); however, in the Vienna Basin margin (Sandberg Formation) they are transgressive in the lower part (Švagroský 1978, 1981; Baráth et al. 1997).

In the East Slovakian Basin (Fig. 6) the Lower Badenian deposits (Nižný Hrabovec Formation, Vass and Cvercko 1985) contain shallow-water foraminiferal assemblages with *Praeorbulina glomerosa* (Blow), *Orbulina suturalis* Brönniman, *Globigerinoides quadrilobatus* (Orbigny), and *G. trilobus* (Reuss), documenting the neritic, open-marine environment at the time of transgression in the central and the eastern parts of the basin. The Middle Badenian Vranov Formation (Vass and Cvercko 1985) were deposited in the neritic to shallow bathyal zone, as documented by the assemblages of agglutinated foraminifera with *Spirolutilus carinatus* (Orbigny), *Cyclammina vulchoviensis* Venglinsky, *C. complanata* Chapman and planktonic species *Globigerina praebulloides* (Blow) and *Globorotalia mayeri* (Cushman et Ellisor) (Zlinská 1992b, 1996a, 1998). The Middle Badenian late highstand sedimentation is represented by the deposition of shallow-water, lagoonal evaporites of the Zbudza Formation (Vass and Cvercko 1985). Based on the analogs in the Transcarpathian and the Transylvanian Basins, as well as in the Carpathian Foredeep (Rögl 1998), the Middle Badenian evaporites in the East Slovakian Basin (Kovác and Zlinská 1998) may be correlated with the end of the TB 2.4 cycle of the global sea-level changes (sensu Haq et al. 1988; Haq 1991).

The Late Badenian sedimentary environment of the Bulimina–Bolivina Zone (Grill 1941) in the Vienna, Danube and the East Slovakian Basins (Figs 6, 7, 8) characterizes water-column stratification and decreased oxygen content at the basin bottom, similar to the highstand conditions of the Eggenburgian–Ottnangian cycle (Kovác and Hudácková 1997; Kovác and Zlinská 1998; Kovác et al. 1999). In the Western Carpathian – North Pannonian region this zone is characterized by the foraminiferal assemblages of *Bolivina dilatata maxima* Cicha et Zapletalová, *Bulimina striata striata* Orbigny, *Praeglobobulimina pyrula* (Orbigny), *Uvigerina venusta liesingensis* Toula, *Pappina neudorfensis* (Toula), *Globigerina nepenthes* Todd, *G. druryi* Akers, and *Globigerinoides quadrilobatus* (Orbigny), displaying a deep neritic environment. In the Late Badenian Bulimina–Bolivina Zone (Grill 1941; Zlinská 1992a; Hudácková and Kovác 1993), an increased content of pteropods was registered in the Vienna Basin (Zorn 1991).

The Late Badenian sedimentation in the Vienna and the Danube Basins ended with hyposaline deposits with foraminiferal assemblages dominated by Ammonia (Kovác et al. 1999). In the East Slovakian Basin, instead of Ammonia, the brackish environment is characterized by a presence of the genus Egerella. These assemblages evolved gradually from the aforementioned Bolivina–Bulimina assemblages. The Ammonia-rich strata (ranging as far as the Early Sarmatian) frequently also contains the redeposited foraminiferal taxa from the underlying Late Badenian Bulimina–Bolivina Zone (Hudácková and Kovác 1993; Hudácková 1995; Kovác and Zlinská 1998), manifesting a regression at the marginal parts of the basins during the end of the Badenian.

On the basis of the mentioned Badenian bio-events and paleogeographic changes in the Carpathian–Pannonian region, the late Early and the Middle Badenian period (CPC 4, 15.1–14 Ma) can be approximately correlated with the TB 2.4 cycle of the global sea-level changes, (15.5–13.8 Ma, mfs 15 Ma; Haq et al. 1988; Haq 1991); the Late Badenian to Early Sarmatian relative sea-level cycle (CPC 5, 14–12.8 Ma; Haq et al. 1988; Haq 1991) is correlated with the TB 2.5 global sea-level cycle, (13.8–12.6 Ma, mfs 13.4 Ma; Haq et al. 1988; Haq 1991). It is documented by a transgression, which can be coeval with that of the TB 2.4 cycle (after the corresponding SB-s in all basins), but a late HST can be found only in the East Slovakian Basin. The transgression coeval with the TB 2.5 cycle can be recognized only in the East Slovakian Basin; in the Vienna and the Danube Basins a long-duration relative highstand developed and most likely basin subsidence overcame the eustatic sea-level fall at the base of the TB 2.5.

Sarmatian to Early Pannonian cycle of relative sea-level changes

The 12.8?–10.8? Ma time interval is represented by the Late Serravalian to Early Tortonian – Sarmatian to Early Pannonian stages (Berggren et al. 1995) and covers the time of the (upper part of the NN 6?) NN 7 and the total time span of the NN8 nannoplankton zone (sensu Martini 1971).

For the entire Paratethys (Fig. 12) quite uniform conditions are characteristic at this time (Rögl 1998). From the Vienna Basin as far as the Caspian Basin, very similar biofacies evolved, characteristic for reduced salinity and a change in chemical composition of the sea water, which became oversaturated with respect to carbonates and possessed high alkalinity at the same time (Pisera 1996). Such conditions were convenient for the growth of red algae, vermitides, and "nubecularian" bioherms.

A sea-level fall near the Late Badenian/Early Sarmatian boundary was accompanied by the deposition of a brackish sequence with the prevailing Ammonia genus (Zlinská 1997) and higher up with the *Anomalinoides badenensis* (Orbigny) species in the Western Carpathian – North Pannonian Basins. On the basis of the newest data from the Carpathian Foredeep (Poland), the late highstand of the relative CPC 5 cycle, which can be correlated with the end of the TB 2.5 global cycle (sensu Haq et al. 1988; Haq 1991) was also demonstrated by salinity crisis and evaporitic sedimentation. This fact is also documented by means of calcareous nannoplankton studies in the sediments underlying and overlying the evaporites with the ages of the NN6, NN7 and NN8 zones (Andreeva-Grigorovich 1994; Peryt 1991; Peryt et al. 1997, 1998). The radiometric age of the underlying beds was determined as 12.5 ± 0.9 Ma; the age of the overlying strata is 12.0 ± 0.8, 11.9± 0.8 Ma (Oszczypko 1997; Couvering et al. 1981). This fact would shift the beginning of deposition of the Sarmatian–Early Pannonian regional cycle of the sea-level changes to 12.0 ± 0.8, which can be correlated with the beginning of the TB 2.6 global cycle (sensu Haq et al. 1988; Haq 1991).

In the Vienna Basin (Fig. 7) the regressive sediments with fresh-water Sarmatian mollusc fauna and redeposited foraminifera of the Bulimina–Bolivina zone (only the relative age determination is available, showing that they are younger than 13 Ma) may be considered as the Early Sarmatian lowstand systems tracts, together with prograding sands from the paleo-Danube delta at the western margin of the basin (Jirícek and Seifert 1990). Another good example of the lowstand systems tracts (over the SB1) are the continental-brackish sediments

Fig. 11 →
Middle Badenian isolation during the relative sea-level fall and faunal migration pathways during the Late Badenian transgression

Fig. 12 →
Late Badenian isolation during the relative sea level fall and faunal migration pathways during the Sarmatian transgression

Fig. 11

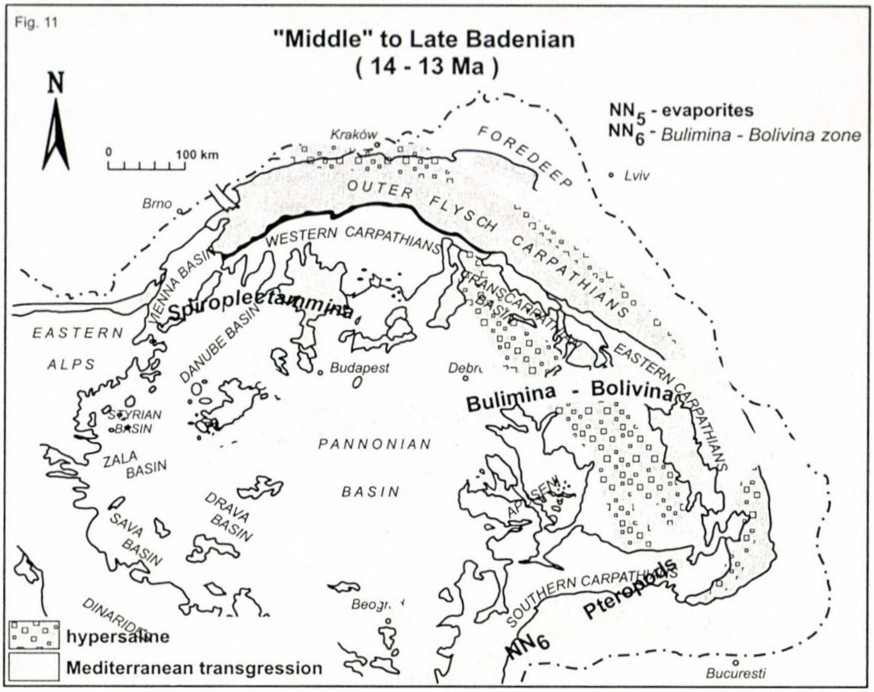
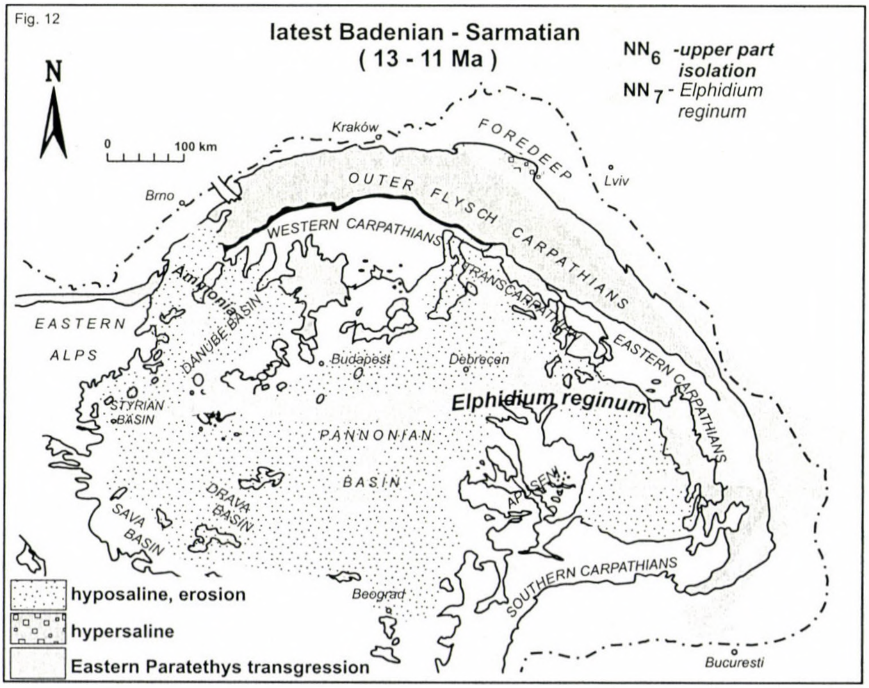

"Middle" to Late Badenian
(14 - 13 Ma)
N
0 100 km
NN5 - evaporites
NN6 - Bulimina - Bolivina zone
Kraków
Lviv
Brno
FOREDEEP
OUTER FLYSCH CARPATHIANS
WESTERN CARPATHIANS
VIENNA BASIN
TRANSCARPATHIAN BASIN
EASTERN CARPATHIANS
EASTERN
ALPS
DANUBE BASIN
Spiroplectammina
Budapest
Bulimina - Bolivina
STYRIAN BASIN
ZALA BASIN
PANNONIAN
BASIN
APUSENI
DRAVA BASIN
SAVA BASIN
DINARIDES
SOUTHERN CARPATHIANS
Pteropods
NN6
Bucuresti
hypersaline
Mediterranean transgression

Fig. 12


latest Badenian - Sarmatian
(13 - 11 Ma)
N
0 100 km
NN6 -upper part isolation
NN7 - Elphidium reginum
Kraków
Lviv
Brno
FOREDEEP
OUTER FLYSCH CARPATHIANS
WESTERN CARPATHIANS
TRANSCARPATHIAN
EASTERN CARPATHIANS
Ammonia
EASTERN
ALPS
DANUBE BASIN
Budapest
Debrecen
Elphidium reginum
STYRIAN BASIN
PANNONIAN
BASIN
APUSENI
DRAVA BASIN
SAVA BASIN
SOUTHERN CARPATHIANS
Beograd
Bucuresti
hyposaline, erosion
hypersaline
Eastern Paratethys transgression

with redeposited forms on Devínska Kobyla Hill and in the Leitha Hills – base of the Karlova Ves Formation (Nagy et al. 1993).

In the East Slovakian Basin (Fig. 6) the Klcovo Formation (Vass and Cvercko 1985), deposited in the environment of a prograding deltaic front to deltaic plain, can be mentioned as an example of the continuous sedimentary transition through the Late Badenian/Early Sarmatian boundary. The Badenian portion of the formation contains mixed fauna of the shallow-water (littoral) and neritic foraminifera species, documenting gradual shallowing and coastal erosion. The Lower Sarmatian transgressive part of the formation already involves the shallow-water species of the genus Elphidium (Zlinská 1997).

The "Sarmatian transgression" in the Vienna, Danube and the East Slovakian Basins also began with shallow-water, brackish sedimentation, with characteristic occurrence of large elphidians of the *Elphidium reginum* Biozone (Grill 1941). The assemblages involve mainly *Elphidium reginum* (Orbigny), *E. aculeatum* (Orbigny), *E. macellum* (Fichtel et Moll), *E. samueli* Zlinská, *Articulina articulinoides* Gerke-Issaeva, *A. problema* Bogdanowicz, and tiny miliolids (Zlinská 1997). For the Middle Sarmatian the presence of the *Elphidium hauerinum* Biozone (Grill 1941) is characteristic, including the euryhaline species *Elphidium aculeatum* (Orbigny), *E. reginum* (Orbigny) and *E. hauerinum* (Orbigny). The sedimentary environment of the Western Carpathian basins is deepened slightly. The Late Sarmatian was characterized by a further decrease of salinity, shallowing and growth of deltaic systems. Foraminiferal assemblages of the *Porosononion granosum* Biozone (Grill 1941) contain individuals of *Porosononion* ex gr. *granosum* (Orbigny), rarely also *Ammonia parkinsoniana tepida* (Cushman), *Elphidium hauerinum* (Orbigny) and *Miliolina* sp. The Late Sarmatian to Early Pannonian isolation led to a complete change to a fresh-water environment in several places in the Western Carpathian basins.

The CPC 6 (12.8–10.8Ma) relative sea-level cycle, which can be approximately correlated with the TB 2.6 global cycle (12.5–10.5 Ma, mfs 11.6 Ma; sensu Haq et al. 1988; Haq 1991) is well manifested in the Vienna, Danube and the East Slovakian Basins. It began after the Late Badenian–Early Sarmatian late highstand Ammonia Beds, by deposition of the lowstand sediments, with the occurrence of the species *Anomalinoides badenensis* (Orbigny). The subsequent transgression (ca. 12.5 Ma) is characterized by the *Elphidium reginum* Biozone (Grill 1941), passing upwards into the *Elphidium hauerinum* one (Grill 1941). Late highstand represents beds of the *Porosononion granosum* Biozone (Grill 1941) passing upwards into the Pannonian Zones A and B (sensu Papp 1951).

Late Miocene cycles of relative sea-level changes

The 10.8?–7.1 Ma time interval is represented by the Tortonian–Pannonian–Late Bessarabian to Meotian stages. Vass et al. (1985) proposed for the Western Carpathian area the Sarmatian/ Pannonian boundary at 11.5 ± 0.5 Ma, the

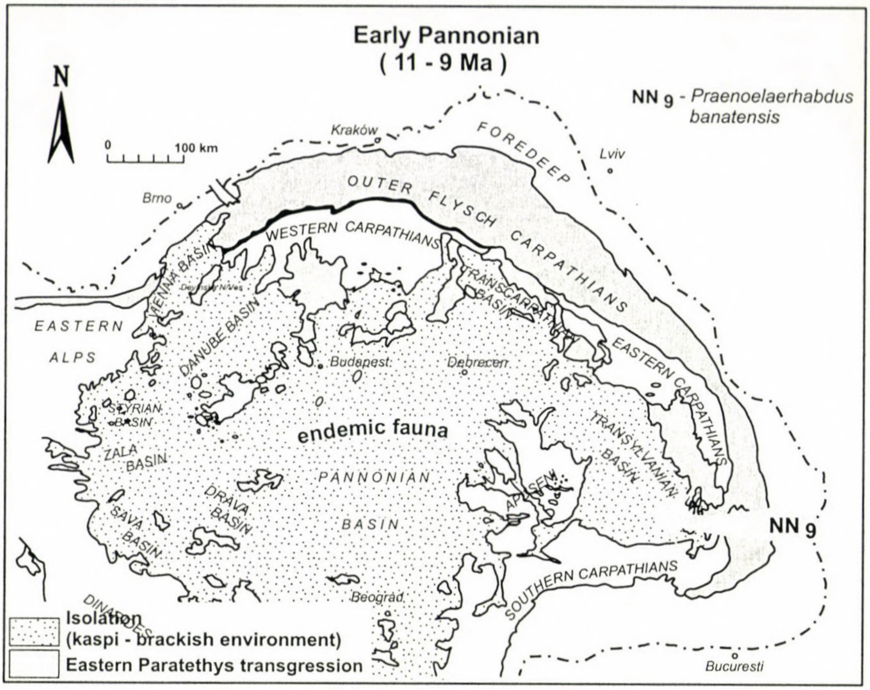


Fig. 13
Early Pannonian isolation during the relative sea-level fall and faunal migration pathways during the late Early Pannonian transgression

Pannonian/Pontian boundary at 8.5 ± 0.5 Ma and the Pontian/Dacian boundary at 5.6 ± 0.2 Ma.

In the Late Miocene, the water-covered area of the Central Paratethys was restricted only to the intra-Carpathian region (Fig. 13). The isolated Pannonian lake system developed in a common water regime and considerably decreased salinity, similar to that of the recent Caspian Sea. The faunal correlation between the Central and the Eastern Paratethys is very difficult due to higher salinity in the Eastern Paratethys, where Sarmatian s.l. - type faunas still survived (Late Bessarabian, Khersonian).

The isolation of the Western Carpathian – Pannonian Basins, manifested by local biostratigraphy, practically disables the use of paleoecology for correlation with the TB 3 cycles of the global sea-level changes (sensu Haq et al. 1988; Haq 1991). It can be said in general that the highstand falling stage depositional system represents the Pannonian Zones A-B (Papp 1951). The appearance of a three-fingered ancestor of the horse (Hipparion) at the margin of the Vienna Basin during the B–C zone (Papp 1951) may be considered as one of the signs of the sea-level lowstand before the transgression in Zone C (Kovác et al. 1998). On the other hand, the highstand of the TB 3.1 global sea-level cycle is documented by a boom of dinoflagellates in the Pannonian E zone (Papp 1951), known also from the Vienna Basin (Kovác and Hudácková 1997; Kovác et al. 1998).

Late Miocene sequence stratigraphy uses sedimentology and seismic data to document the basin evolution (see Vakarcs et al. 1994, Vakarcs 1997; Pogácsás et al. 1993).

Discussion and conclusions

The relative sea-level changes in the Western Carpathian – North Pannonian Basins, as can be deduced from the previous text, have also been considerably shaped, apart from eustacy, by local tectonic events; thus, they are neither always identical with the global cycles defined by Haq et al. (1988) and Haq (1991), nor with others found in different basins of the region. On the other hand, we were able to date many biostratigraphically and paleoecologically important paleogeographic events in the Western Carpathian – North Pannonian region, as well as to correlate them with the events in the Central Paratethys and the Mediterranean area.

The Late Egerian – Early Eggenburgian SMST or LST in the Western Carpathian – North Pannonian area (23.5 – 21? Ma) have been caused by terminal deposition of the Biely Potok Formation in the fore-arc basin at the active margin of the Western Carpathians (Fig. 6), and in the South Slovakian – North Hungarian Basin (Fig. 5) in the Western Carpathians hinterland by deposition of the Opatová Beds deltaic sequences overlying the "Szécsényi Schlier beds" of the Lucenec Formation.

The Eggenburgian transgression in the Western Carpathian – North Pannonian area is manifested by sequence boundary of the SB1 type in the Vienna Basin, the Bánovce Depression, in the Váh River Valley, the South Slovakian – North Hungarian (Pétervására) and the East Slovakian Basins (Figs 5, 6, 7, 8). Littoral sediments are characterized by the occurrence of large pectinids and calcareous nannoplankton of the NN2 zone (sensu Martini 1971). The sediments deposited in neritic zone (somewhat later) also contain *Helicosphaera ampliaperta* (Bramlette et Wilcoxon) Bukry in nannoflora assemblages (FAD 20.5 Ma, sensu Berggren et al. 1995).

In the South Slovakian – North Hungarian (Pétervására) Basin (Fig. 5) the Eggenburgian transgression is followed by HST deposits of the Cakanovce beds of the Filakovo Formation (upper part of the Szécsény Schlier Formation in Hungary) covered by the terrestrial Bukovinka Formation, Eggenburgian in age (Zagyvapálfa Formation in Hungary). The above-mentioned deposits together with the Late Egerian sediments (SMST) form a relative sea-level cycle (CPC 1a) similar to the cycle of the East Slovakian Basin (Fig. 5), where the Egerian–Early Eggenburgian LST represents the Biely Potok Formation and the Eggenburgian TST and HST is represented by the Prešov Formation and the regressive Celovce beds in the terminal part.

In the southern part of the Danube Basin (Transdanubian Range) the Ottnangian relative sea-level cycle (CPC 1b) began by marine transgression in the Bántapuszta region; in the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5) it is represented by lower coal-bearing beds deposited in the alluvial plain environment and upper limnic to marine beds (TST, HST).

In the East Slovakian Basin the Ottnangian deposits are missing (uplift due to tectonic activity).

In the Vienna Basin, the Váh river valley and the Bánovce Depression (Figs 7, 8) large-scale transgression was followed by a deepening of the sedimentary environment and the end of the cycle is represented by shallow-water, brackish sediments of the Ottnangian age. The Eggenburgian to Ottnangian relative sea-level cycle (CPC 1) in the Vienna Basin can be compared to the TB 2.1 global cycle (21–17.5 Ma, mfs 18.5 Ma; sensu Haq et al. 1988; Haq 1991).

The Latest Ottnangian–Karpatian relative sea-level cycle in the Western Carpathian – North Pannonian area – CPC 2, which is correlated with the TB 2.2 cycle of the global sea-level changes (17.5–16.5 Ma, mfs 17 Ma; sensu Haq et al. 1988; Haq 1991), is demonstrated by increased intensity due to the synrift stage of the intra-Carpathian basin evolution, with a high rate of tectonic subsidence in the Vienna, Danube, South Slovakian – North Hungarian (Novohrad) and the East Slovakian Basins (Lankreijer et al. 1995; Baráth et al. 1997; Lankreijer 1998). The sequence boundary of SB 1 type is covered by transgressive deposits of the latest Ottnangian or the Early Karpatian. The end of the cycle is represented by erosional surfaces or by the Karpatian evaporites in the East Slovakian Basin, both transgressively overlain by the latest Karpatian to Early Badenian, mostly deltaic, strata.

The latest Karpatian–Early Badenian CPC 3 relative sea-level cycle in the Western Carpathian – North Pannonian area, which is correlated with the TB 2.3 cycle of the relative sea-level changes (16.5–15.5 Ma, mfs 16 Ma; sensu Haq et al. 1988; Haq 1991) started with a sharp boundary of SB 1 type, documented by an angular unconformity between the Lower and the Middle Miocene sediments in most of the Western Carpathian basins. The CPC 3 cycle is, however, hardly determinable as an interference between the sea-level fall at the Early/Middle Miocene boundary and the considerable slowing down of the subsidence (or uplift) which occurred at that time. The cycle was obviously completely controlled by tectonics acting throughout its duration, which disables an accurate correlation of the evolution of individual basins (heterochronous deposits of lowstand and transgression between the Late Karpatian and the Early Badenian).

In the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5) this period is represented by regressive-transgressive Lower Badenian deposits with the Karpatian redeposited foraminifera and Praeorbulina sp., whereas in the East Slovakian Basin (Fig. 6) it is represented by the terrestrial interbeddings in the Kladzany Formation deposits with *Globigerinoides bisphericus* Todd (its upper part).

In the Vienna Basin (Fig. 7) the upper part of the Karpatian/Early Badenian conglomerates and sandstones of the Aderklaa and the Jablonica Beds and the conglomerate at the base of the "Lanzendorf Series" may be attributed to a new – "Langhian" cycle of the relative sea-level changes (TB 2.3). The upper boundary of the SB 2 type was identified in the Austrian part of the basin within the Upper Lagenide Zone (Weissenbäck 1996).

The late Early Badenian "tectonically-controlled" transgression started during the deposition of the Upper Lagenide Zone in all Western Carpathian basins (Figs 5, 6, 7, 8) and is characterized by the onset of the genus Orbulina (15.1 Ma ago, sensu Berggren et al. 1995).

The late Early to Middle Badenian relative sea-level cycle CPC 4 in the Western Carpathian – North Pannonian region can be more or less correlated with the TB 2.4 global sea-level cycle (15.5–13.8 Ma, mfs 15 Ma; sensu Haq et al. 1988; Haq 1991), but its termination in the Western Carpathian – North Pannonian area is heterochronous except for the East Slovakian Basin, where it was manifested by the evaporitic sedimentation at the NN5/NN6 nannoplankton zone boundary, similarly to the Carpathian Foredeep, the Transcarpathian and the Transylvanian Basins (Buday et al. 1965; Kovác et al. 1989; Rögl 1998). This period in the Central Paratethys area represents the chronostratigraphic boundary at 14.4 to 14 Ma (sensu Berggren et al.; 1995, Rögl 1998).

The Late Badenian to Early Sarmathian relative sea-level cycle CPC 5 in the Western Carpathian – North Pannonian area can be approximately correlated with the TB 2.5 global cycle of the sea-level changes (13.8 – 12.5 Ma, mfs 13.4 Ma; sensu Haq et al. 1988; Haq 1991). In the Western Carpathian area the lower sequence boundary of SB 1 type is known only from the East Slovakian Basin, followed by transgression of the Upper Badenian sediments.

The Late Badenian pelites of the Bulimina–Bolivina Zone (Grill 1941) are considered to be the deposits of the sea-level highstand during the NN6 nannoplankton zone in the East Slovakian Basin (Fig. 6), similarly as in the Danube and the Vienna Basins (Figs 7, 8). The end of sedimentation represents a transition into the brackish Ammonia-bearing beds during the sea-level fall in the basal part of the NN7 nannoplankton zone. In the mentioned cycle no influence of increased tectonic activity on the paleogeographical evolution of the intra-Carpathian basins was registered.

In the South Slovakian – North Hungarian (Novohrad) Basin (Fig. 5) the Early Badenian deposits with *Praeorbulina* sp. are transgressively overlain by deposits of the Upper Lagenide Zone (Grill 1941 1943) and at the same time the marine sedimentation terminated (CPC 4) and it was replaced by deposition of the Middle to Late Badenian volcano-sedimentary complexes at the Novohrad Basin northern margin.

In the Vienna and the Danube Basins (Figs 7, 8) a slight decrease in salinity and shallowing occurred (Cicha in Buday et al. 1965; Papp in Papp et al. 1978) at the end of the Upper Lagenide Zone (Grill 1941, 1943). The boundaries between sedimentary bodies, displaying the relative sea-level oscillation in the Vienna and the Danube Basins at the Lower/Middle Badenian boundary, are interpreted to be the result of a decrease in tectonic activity and increased sediment input. In the Vienna Basin this event is represented by the transgressive-regressive body of the Matzen Sandstone overlying the Upper Lagenide Zone (Kreutzer and Hlavatý 1990) and it reflects development of a new drainage pattern and delta progradation at the western margin of the basin (Paleo-Danube Delta). The Middle Badenian deposits of the *Spiroplectammina (=Spirolutilus) carinata* Zone (Grill 1941) possess a transgressive character in the Vienna and the Danube Basins (Buday et al. 1965; Hudácková et al. 1998), passing into the highstand systems tracts. They are separated from the overlying sediments of the Bulimina–Bolivina Zone (Grill 1941) by a conformable surface (CPC 4+5), representing in the East Slovakian Basin the boundary of the next relative sea-level cycle. These facts are consistent with the synrift stage of the basins evolution at that time (Lankreijer et al. 1995; Kovác et al. 1997; Baráth et al. 1997).

In the East Slovakian Basin (Fig. 6) a continuous transition from the Upper Lagenide Zone to the overlying *Spiroplectammina (Spirolutilus) carinata* Zone (Grill 1941) was documented. The late highstand of the CPC 4 cycle represents the transition of the Vranov Formation to the evaporite-bearing Zbudza Formation at the end of the Middle Badenian (Baráth et al. 1997). The transgression of the Bulimina–Bolivina Zone deposits represents the next CPC 5 cycle relative sea-level cycle in the East Slovakian Basin. These facts are consistent with the synrift stage of the basin evolution at that time (Baráth et al. 1997).

The Sarmatian to Early Pannonian relative sea-level cycle CPC 6 in the Western Carpathian – North Pannonian area can be approximately correlated with the TB 2.6 global cycle of the relative sea-level changes (12.5–10.5 Ma, mfs 11.5 ma; sensu Haq et al. 1988; Haq 1991), if correlatable at all, due to the isolation of the intra-Carpathian region from the Mediterranean.

The Early Sarmatian transgression in the Vienna, Danube and the East Slovakian Basins (Figs 5, 7, 8) is characterized by the onset of the large Elphidium Zone (Grill 1941). The transgression was enhanced by the tectonic subsidence in the northern part of the Vienna and the Danube Basins, as well as in the East Slovakian Basin (Lankreijer et al. 1995; Baráth et al. 1997). The overlying sediments of the *Elphidium hauerinum* Zone (Grill 1941) can be considered to be the highstand systems tract. During the global sea-level fall between 11.5–10.5 Ma ago the late highstand – *Porosononion granosum* Zone (Grill 1941) was formed in the Western Carpathian area through decreasing salinity and fresh-water sedimentation (the Late Sarmatian coal-bearing Kochanovce Formation of the East Slovakian Basin and the Late Sarmatian deposits of the *Porosononion granosum* Zone overlain by relics of the terrestrial-lacustrine deposits at the base of the Pannonian Zone C (sensu Papp 1951) in the Vienna and the Danube Basins).

The tectonically controlled subsidence in the Late Miocene was a reflection of the second rifting phase of the Pannonian back-arc basin (Lankreijer 1998) and began ca. 10.5 Ma ago, similarly as the TB 3 cycles of the relative sea-level changes (sensu Haq et al. 1988; Haq 1991).

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Chloritoid schist from the Uppony Mts (NE Hungary): mineralogical, petrological and structural data from a new occurrence

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Chloritoid from the Lower Paleozoic metasandstone (Tapolcsány Fm., Rágyincsvölgy Sandstone Mb.) in the Uppony Mts. (NE Hungary) has been known for a long time, although there were some debates on its origin. A previously not described chloritoid schist occurrence was found in the type locality (Rágyincs valley) of the metasandstone. This rock type consists of idiomorphic, lath-shaped chloritoid porphyroblasts of 0.5–2 mm in a very fine-grained quartz, white K-mica and subordinately albite-bearing matrix. Textural relationships as well as the high modal content of chloritoid (c. 25–30 vol.%) clearly prove the metamorphic origin of this index mineral.

Outcrop-scale structural observations on the metasandstone and surrounding Tapolcsány slate reveal the presence of two foliation generations: the first one is sub-parallel to bedding (S_{0-1}); the second foliation cuts it at high angle and forms the axial plane foliation (S_2) of gently northeastward or southwestward-plunging folds. Post-tectonic formation of chloritoid with respect to S_2 foliation indicates that the major phase of Alpine deformation (e.g. folding of the Uppony Paleozoic sequence) predates Cretaceous peak metamorphic conditions.

XRD investigations show that chloritoid belongs to the triclinic structure polytype. According to microprobe investigations chloritoid has a relatively Fe-rich composition. It shows a slight zoning from the core toward the rims, with increasing Mg/(Mg+Fe) ratio (core: 0.11, rim: 0.16), suggesting prograde metamorphic conditions during its growth.

The doubtlessly metamorphic origin of the chloritoid from this rock type contradicts the previous idea of Noske-Fazekas (1973) arguing for its detrital origin and unambiguously confirms the conclusions of Árkai et al. (1981), deduced from the chloritoid-bearing metasandstone.

Key words: Chloritoid, metasandstone, low-grade Alpine metamorphism and ductile deformation, Uppony Paleozoic, NE Hungary

Introduction

Chloritoid is one of the most important metamorphic index minerals, occurring predominantly in greenschist facies metamorphic rocks. It is mostly formed by the metamorphic transformation of Al-rich metapelites with generally relatively high Fe/Mg ratio (Spear 1995, p. 349). Its stability field also extends into the lower amphibolite facies (Bucher and Frey 1994; p. 200–201). In case of Mg-rich bulk chemistry, chloritoid – in association with talc – can be stable also at high pressure

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values (over 10 kbar) in the blueschist and/or eclogite facies (see e.g. Spear 1995; p. 363).

Chloritoid as a newly-formed, metamorphic mineral was described only from very few localities in Hungary: it occurs in the medium-grade metapelites of the Sopron Mts (Kisházi 1977; Török, unpubl. data), in the metabasites of the Bódva Valley (Horváth 1997) and in the Lower Paleozoic (Ordovician–Silurian?) metasandstone (Tapolcsány Fm., Rágyincsvölgy Sandstone Mb.) of the Uppony Mts (Noske-Fazekas 1973; Árkai et al. 1981; Ivancsics and Kisházi 1983).

A new chloritoid schist occurrence was found in the classical outcrop (Rágyincs valley) of the chloritoid-bearing metasandstone in the Uppony Mts. A chloritoid schist fragment from this type locality was also mentioned by Noske-Fazekas (1973), who first described chloritoid from the Rágyincsvölgy Sandstone. However, the exact locality, extent and the relationship of the chloritoid schist to the surrounding metasandstone was not discussed by her. The published photos (Noske-Fazekas 1973) suggest that chloritoid schist fragments form only very thin (mm-scale) lenses in the metasandstone. Furthermore a detailed mineralogical and petrological description of this rock type was not available until now.

Therefore, the aim of the present study is to characterize this rock type in detail from a petrographic–petrological point of view and describe its structural–lithological relationships to the neighboring rocks. Considering these results we will discuss previous ideas about the origin of chloritoid and structural features of the Rágyincsvölgy Sandstone; furthermore, combining the obtained data we can reconstruct the temporal relationship between metamorphism and deformation.

Geologic setting

The Uppony Mts in NE Hungary form a smaller pre-Tertiary basement exposure built up predominantly by low-grade, Lower Paleozoic meta-sedimentary sequences. They are located in the direct north-northwestern neighborhood of the Bükk Mts (Fig. 1) and belong to the so-called Gemer-Bükk region, which comprises the innermost tectonic units of the Inner Western Carpathians.

The Uppony Mts are bounded entirely by the approximately NE–SW-trending Darnó fault zone: to the NW, Mesozoic rocks of the Aggtelek–Rudabánya Mts are separated by the Uppony fault, while the Nekézseny fault to the SE separates the Paleozoic and Mesozoic sequences of the Bükk Mts from the Uppony Paleozoic in the pre-Tertiary basement (Fig. 2). The present structure of the Uppony Mts was interpreted as the result of polyphase Tertiary activity along the Darnó zone: sinistral transpressional and pure sinistral regimes were active in several phases during Paleogene–Middle Miocene times (Fodor et al. 1992, 1999). The last phase produced transtensional (releasing bend: Csernely trough) and transpressional areas (restraining bend: Uppony Mts).

The rather complicated internal structure of the Uppony block is characterized by upright, open to tight folds of probable Middle Cretaceous age (Csontos 1989). Fold axes are predominantly NE–SW-trending, parallel to the main strike of the exposed stratigraphic units. Steeply SE-dipping axial plane foliation of these folds is very frequently observed in the outcrops, often representing the only visible

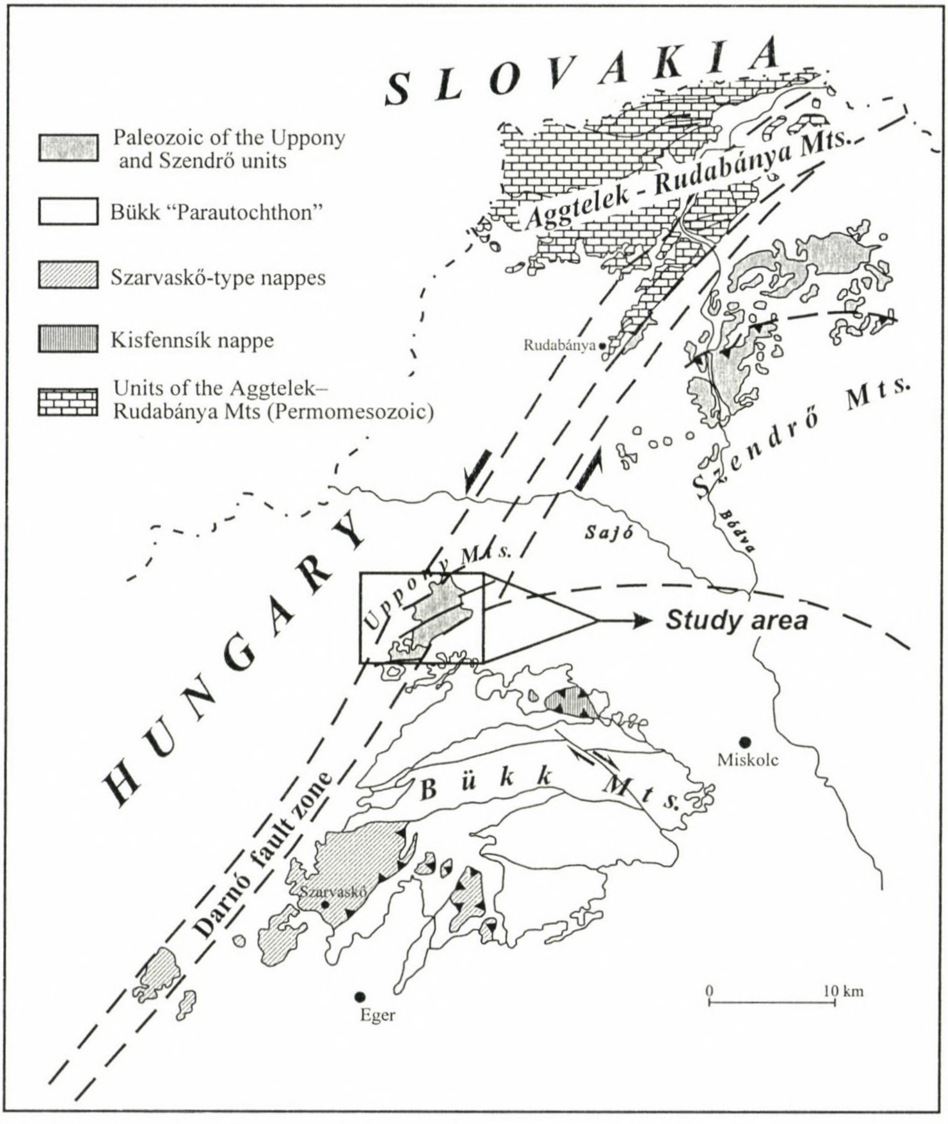


Fig. 1
General overview map of the pre-Tertiary basement in NE Hungary with the study area (Modified after Balogh 1964)

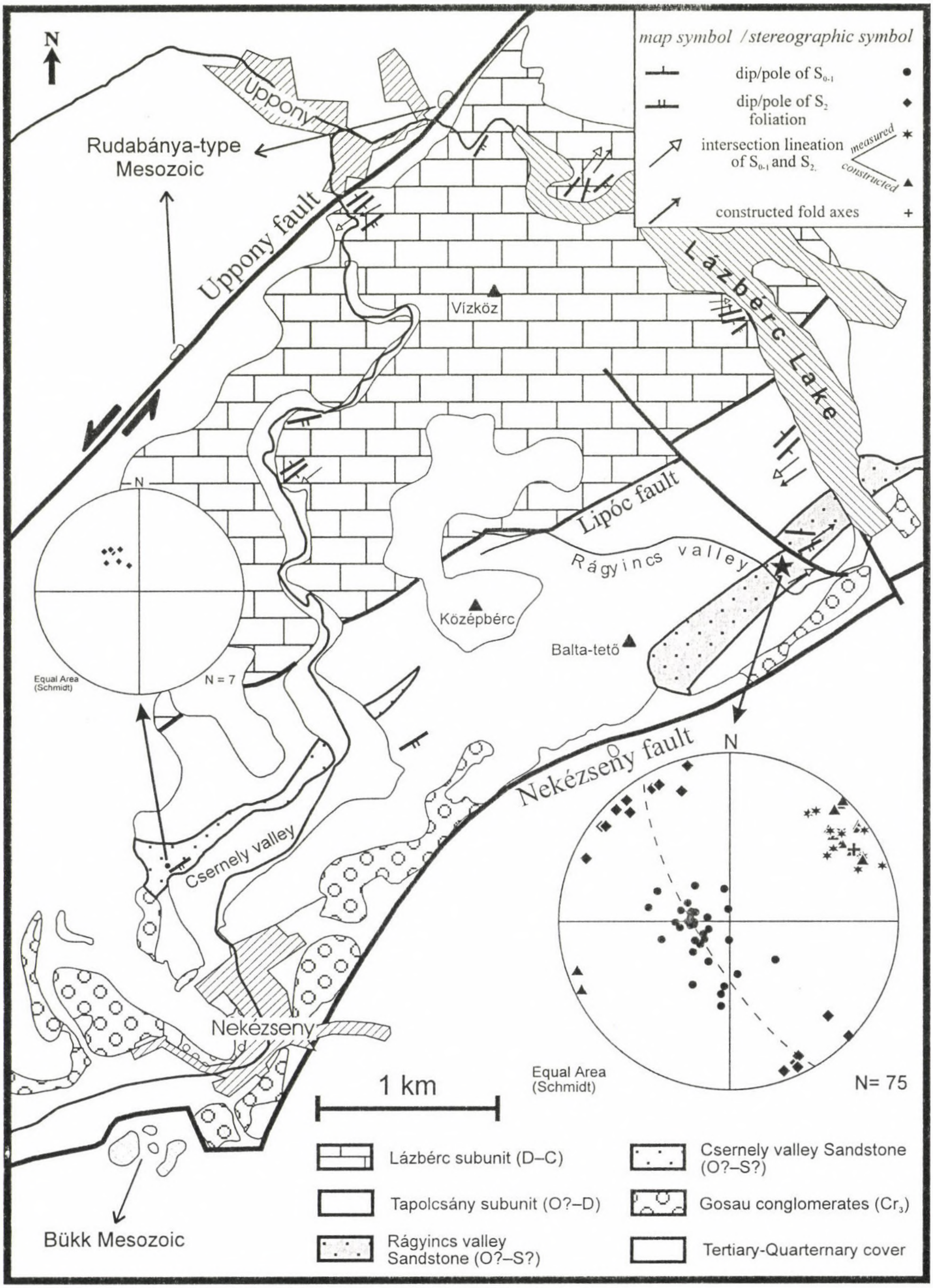


Fig. 2

Simplified geologic map of the Uppony Mts. with the main bounding faults and subunits (modified after Kovács 1983). The two metasandstone intercalations (Csernelyvölgy Sandstone and Rágyincsvölgy Sandstone Mb.) in the southern Tapolcsány subunit are indicated. The symbol "★" shows the locality of the chloritoid schist. Stereograms show the orientation of the different foliation generations, measured and constructed intersection lineations, and fold axes structural feature. Internal thrusts also occur; however, the imbricated structure may be connected – at least partly – to Early Miocene tectonics, as was e.g. demonstrated at the northern Uppony boundary fault (Pantó 1954). Both brittle and ductile structures show generally N–NW vergency (Schréter 1945; Koroknai and Frisch 1998). The grade of metamorphic alteration, style of ductile deformation and also the orientation of the main structures show great similarity to the so-called Martonyi Unit (Fodor and Koroknai 2000) also bounded by the Darnó fault zone in the Rudabánya Mts further to the NE.

In the Uppony Mts two tectonic subunits were distinguished, which are separated by the Lipóc fault (Kovács 1992, see Fig. 2):

1. The northern *Lázbérc subunit* consists of mainly Middle Devonian to Middle Carboniferous platform and pelagic carbonates with minor siliciclastic intercalations. Middle–Upper Devonian (dated: Frasnian) volcanic activity is present in this subunit (Kovács and Vető-Ákos 1983).

2. The southern *Tapolcsány subunit* is built up by predominantly siliciclastic rocks (siliceous slate, metasandstone) of unknown age (probably Ordovician and/or Silurian) and a presumably Middle Devonian volcano-sedimentary formation containing Silurian and Lower Devonian olistoliths in a strongly altered basic volcanic matrix (Kovács and Vető-Ákos 1983).

Both subunits suffered Alpine regional metamorphism that reached approximately the transitional zone between the very low and low-grade (lower greenschist facies) conditions (Árkai et al. 1981). K-Ar ages on K-white mica ($<2\,\mu$m grain-size) are in the range of 115–130 Ma (Árkai et al. 1995). There is no direct evidence for a pre-Alpine tectonometamorphic event, but this could not be excluded either (Árkai 1983). However, if it was present it surely did not exceed the Alpine metamorphic conditions, since there is no trace of any previous, higher-grade, retrogressed mineral assemblage; furthermore, there is no sign of any Variscan event in the geochronological data set (Árkai et al. 1995).

Based on lithostratigraphic and facies characteristics the Paleozoic of the Uppony Mts was correlated with the Paleozoic of the Carnic Alps and the Graz Paleozoic, suggesting their original proximity on the southern shelf of the Prototethys (Ebner et al. 1998).

The Lower Paleozoic sequence is covered by Upper Cretaceous, Gosau-type conglomerates at the southern part of the Uppony Mts. (Brezsnyánszky and Haas 1984; Clifton et al. 1985). This clastic formation contains low-grade metamorphic pebbles of the Uppony Paleozoic, proving its uplift and exhumation during the Late Cretaceous. The surroundings of the Uppony Mts are covered by Miocene sediments and volcanics.

Analytical methods

The SPHERISTAT program was used for evaluation and graphical presentation of field tectonic data. Foliation data were plotted by their pole in the Schmidt net (equal area, lower hemisphere projection). Intersection lineation data were either measured in the field or partly calculated and constructed from the S_{0-1} resp. S_2 dataset (if it was not detected in the outcrops). Beside the usual macro- and microscopic investigations the following methods were applied for checking the mineralogical and mineral chemical composition of the studied samples.

X-ray diffraction analysis on whole rock sample was performed with a Philips PW-1730 diffractometer equipped with a graphite monochromator using Cu-K_α radiation at 45 kV and 35 mA with 1° divergence slit and 1° receiving slit samples in the Laboratory for Geochemical Research, Hungarian Academy of Sciences, Budapest. Scanning rate was 0.05° 2Θ per second from 3° to 70°. Modal composition was determined by semi-quantitative phase analysis using random (disorientated) powder mounts.

Chemical analyses of minerals were carried out with a JEOL JXCA-733 electron microprobe equipped with 3 WDS in the Laboratory for Geochemical Research, Hungarian Academy of Sciences, Budapest. The measuring conditions were: 15 kV acceleration voltage; 40 nA sample current; defocused electron beam with a diameter of 5–10 μm; 5 s counting time. Matrix effects were corrected by using the ZAF method. The following standards were used for quantitative analysis: orthoclase (K, Al, Si), synthetic glass (Fe, Mg, Ca), spessartine (Mn), rutile (Ti) and albite (Na).

Structural observations

The Tapolcsány subunit, which is built up predominantly by dark, siliceous slate and lydite, contains two metasandstone intercalations (Fig. 2):

a) in the southwestern part of the mountains the *Csernelyvölgy Sandstone Member* (exposed mainly near Nekézseny);

b) in the southeastern part the *Rágyincsvölgy Sandstone Member* (exposed mostly in the Rágyincs valley).

It must be emphasized that these two metasandstones show very different mineralogy and deformation style, which was already outlined by Ivancsics and Kisházi (1983). The Csernelyvölgy Sandstone is a very massive sandstone body without any visible sedimentary structure (e.g. bedding). Ductile deformation resulted in the formation of a very weakly developed, gently SE-dipping foliation (fracture cleavage). From a mineralogical point of view it is a poorly sorted, lithic greywacke (Ivancsics and Kisházi 1983) containing a considerable amount of lithoclasts, feldspars and micas, but no chloritoid.

The previously not-described chloritoid schist occurrence was found on the northern side of the Rágyincs valley, at the type locality of the Rágyincsvölgy Sandstone Member (Fig. 2). This chloritoid-bearing metasandstone is macroscopically a light-gray, fine-grained rock with two well-observable foliations: the first one (S_1) is a sub-horizontal, widely spaced (in the range of c. 2–6 cm) foliation, dipping mostly to (E)SE (Fig. 2). At first glance this foliation is

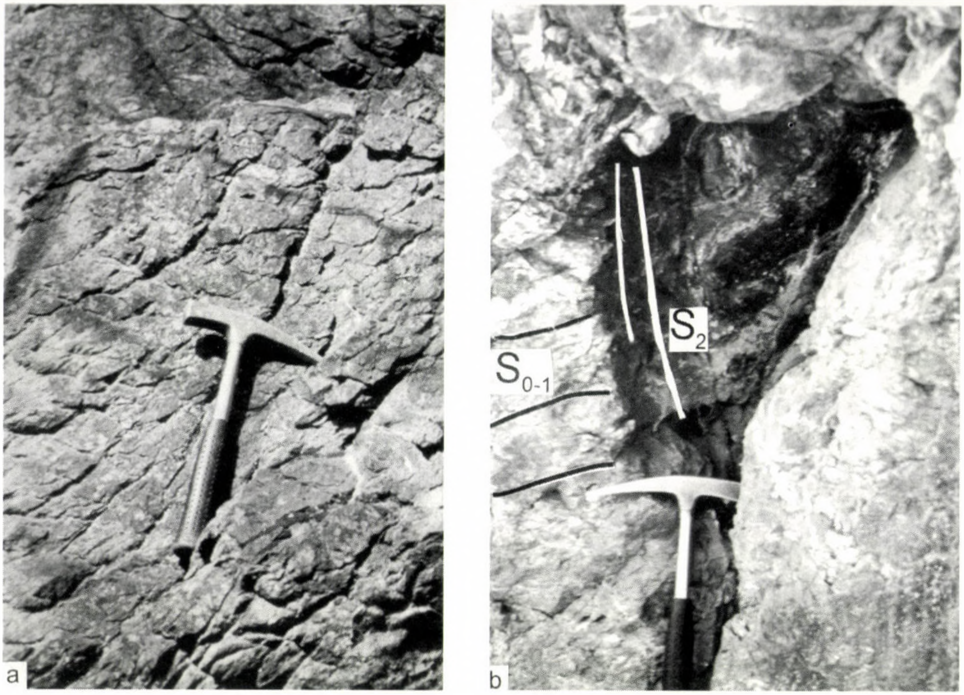


Fig. 3
a) Outcrop view of the Rágyincsvölgy Sandstone. Note the widely spaced, gently dipping (from left to right), bedding-parallel first foliation (S_{0-1}) and steeply dipping, penetrative S_2 (from upper right to lower left). Observable wedge-shaped splitting of the rock originates from the intersection of S_{0-1} and S_2. b) Photo of the newly observed chloritoid schist block in the Rágyincsvölgy Sandstone. Traces of S_{0-1} resp. S2 are indicated by white and black lines, respectively

the most characteristic structure in outcrop scale (Fig. 3a) to be described as a metamorphic foliation (Fülöp 1994) not related to the original bedding of the rock. The second foliation (S_2) is a steeply SE– (resp. NW) dipping, closely spaced (mm-scale) foliation crenulating S_1 that can be observed both in macroscopic (Fig. 3a) and microscopic scale (Fig. 5a). The intersection of S_2 on S_1 surfaces resulted in a sub-horizontal (or gently) NE (resp. SW) plunging intersection lineation (Fig. 2) that is parallel to the majority of fold axes measured in the outcrops of the Uppony Mts. The wedge-shaped splitting of the rock (Fig. 3a) is also the consequence of the intersection of these two foliation planes.

In the vicinity of this classical metasandstone outcrop very fine-grained, banded metasiltstone and slates with post-tectonic, skeletal chloritoid crystals are exposed (Ivancsics and Kisházi 1983). Two foliations were developed in these rocks as well, showing the same above-mentioned wedge-shaped structure. The first foliation in this metasiltstone is clearly parallel to the original bedding (S_{0-1}), defined by the alternation of compositionally different (pelitic and silty) layers.

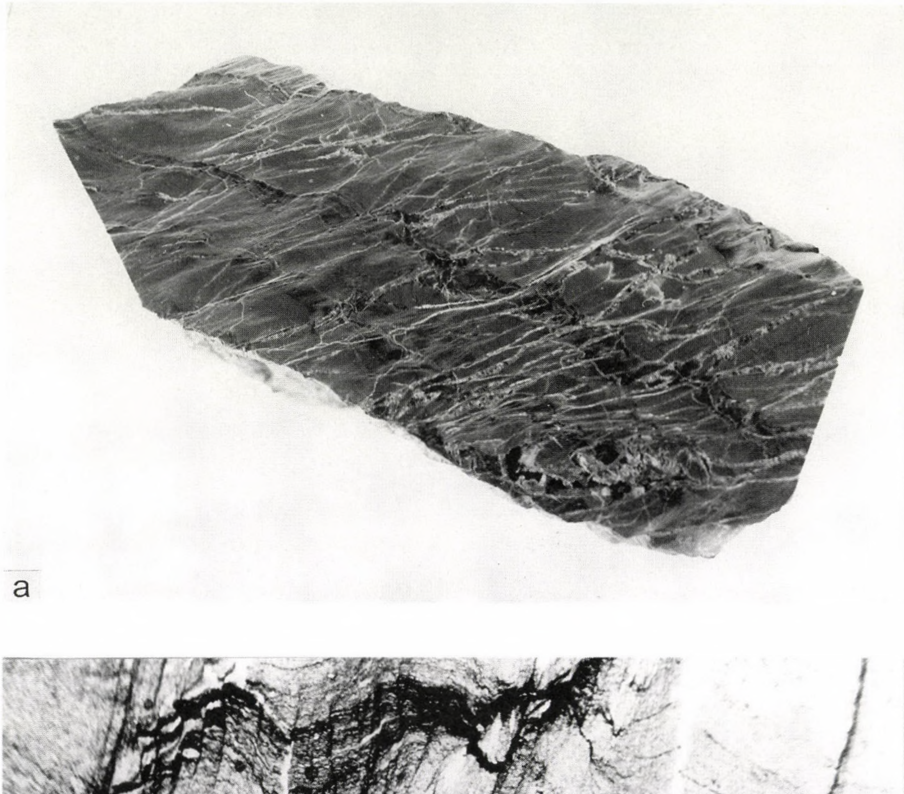


Fig. 4
a) Polished surface of the surrounding metasiltstone (Tapolcsány Formation). Note the two foliations: S_{0-1} (from upper left to lower right) parallel to the original sedimentary bedding (marked by the alternation of darker and lighter layers). Anastomosing S_2 foliation surfaces cut S_{0-1} at high angle. Length of picture: 10 cm. b) Thin section from the same rock with folded S_{0-1} and S_2 (nearly vertical). Length of picture: 3 mm, 1N

This feature can be studied very well both on polished surfaces and in thin sections (Figs 4a, b). Crenulation of S_{0-1}, caused by the development of S_2, is also very characteristic here (Fig. 4b).

Petrography and mineral chemistry

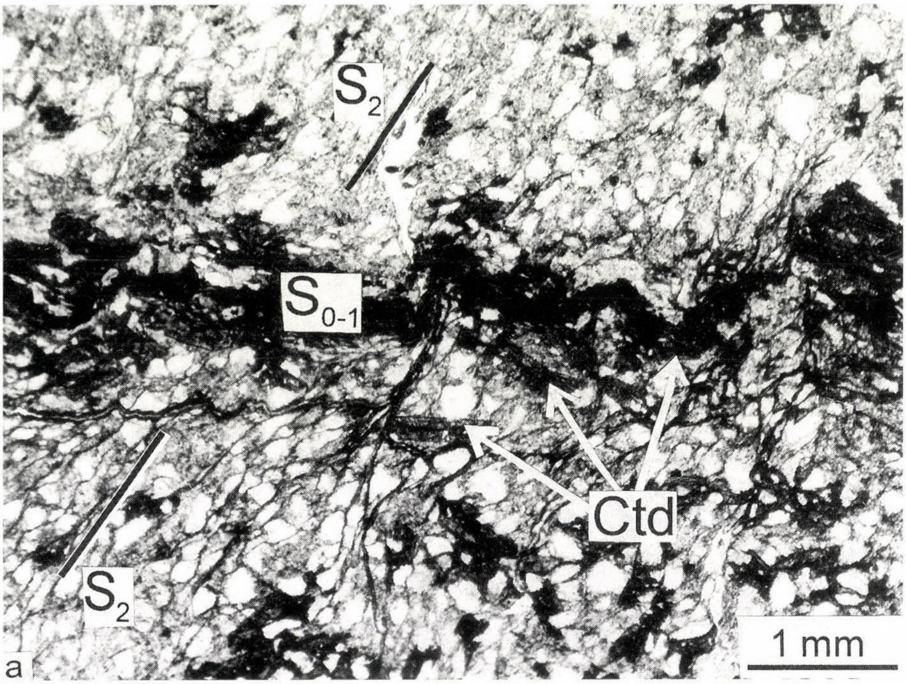
The new locality of chloritoid schist was found in the easternmost cliffs of the metasandstone type locality (Rágyincs Valley). It forms blackish, dm-scale blocks (Fig. 3b) or lenses that are strongly transposed into the anastomosing S_2 foliation (see previous section) of the light-gray metasandstone. In the well-foliated hand specimens dark chloritoid grains, 1–2 mm in size, are visible.

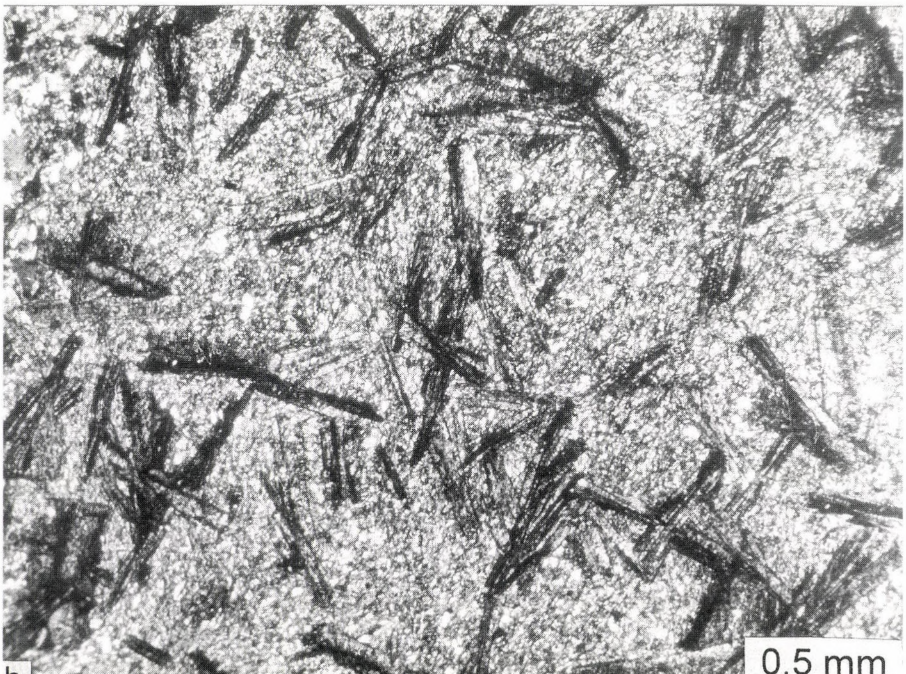
Microscopic and XRD investigations show that this rock has a relatively simple mineralogy. Chloritoid forms relatively large (0.2–2 mm), randomly oriented porhyroblasts overgrowing the S_2 foliation. This records its post-tectonic character in accordance with previous observations of Árkai (1982) on the host metasandstone and those of Ivancsics and Kisházi (1983) on the neighboring metasiltstone. It occurs both as single idioblastic, lath-shaped crystals (often forming penetrating twins) or as rosettes (Figs 5b, 7a–b) in a very fine-grained matrix. Polysynthetic twinning and hourglass structure is a common feature. It reaches surprisingly high modal content (approx. 25–30 vol.%). The fine-grained matrix contains white K-mica, quartz and opaque minerals (pyrite, hematite); furthermore accessory zircon and tourmaline could be optically determined.

The rock exhibits porphyroblastic texture. At microscopic scale foliation is defined by oriented white micas. Although this rock is of metamorphic origin its texture also shows – at least at first sight – great similarity to basalt with intersertal texture because of the high proportion of the lath-shaped, polysynthetically twinned chloritoid crystals occurring in a fine-grained matrix (Fig. 5b).

According to the semi-quantitative phase analysis of the XRD investigations (Fig. 6) the sample consists of quartz (c. 40–45 wt%), chloritoid (c. 20–35 wt%), 10 Å phyllosilicate (sericite–muscovite, ca. 10–15 wt%) and a minor quantity of plagioclase (about 5 wt%). The plagioclase shows albitic composition. The subordinate presence of rutile is probable as well.

Two structure polymorphs of chloritoid are known to exist in nature: a triclinic and a monoclinic one. In case of the triclinic polytype the highest intensity peak is located at d(002)=4.45Å, while the monoclinic type shows its highest intensity at the d(111)=4.47Å peak. Based on these data the chloritoid schist from the Uppony Mts is made up of the triclinic polytype. This is also supported by the presence of the strong 3.25Å, 2.69Å, 2.40Å and 1.83Å reflections. However, because of the appearance of the peaks at 2.59Å and 2.49Å, the monoclinic polytype can be present in subordinate quantity as well. The close association of these two polytypes in the same sample is not rare; several authors described samples containing both structure types (e.g. Halferdahl 1961 and Hanscom 1980). It is generally accepted that the triclinic polytype is more stable in rocks


S2
S0-1
Ctd
S2
1 mm
a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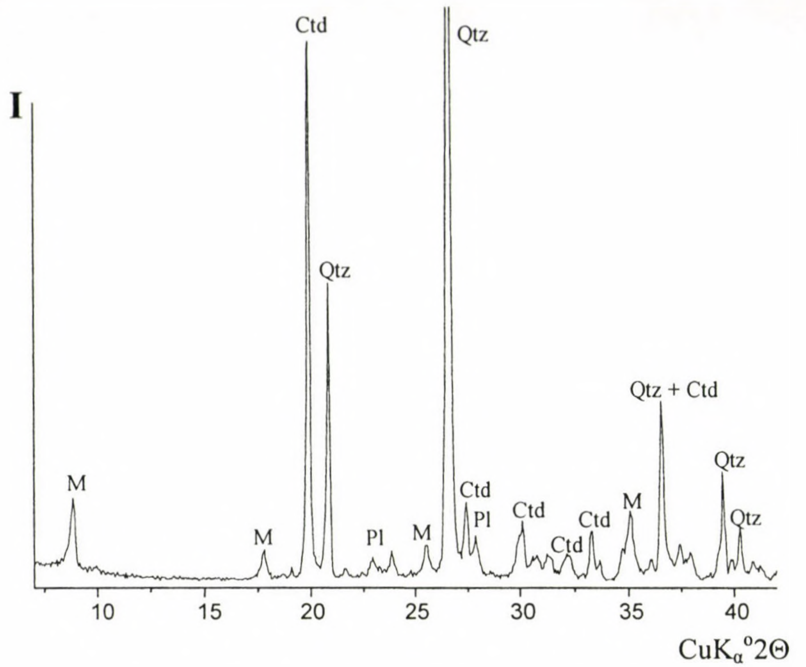

Fig. 6
Results of the XRD investigations. Ctd=chloritoid, Qtz=quartz, M=white K-mica (sericite), Pl=plagioclase

representing (very) low-grade metamorphic conditions and disappears in higher-grade rocks. Chloritoid with high Mg content (so-called magnesiochloritoid) belongs mainly to the monoclinic polytype. With increasing Mg content, the cell parameters also increase (Chopin et al. 1992).

The polytype of the 10Å layer-silicate cannot be unambiguously identified because its reflections are present only in small number. Based on the 2.99Å reflection, it probably belongs to the 2M1 polytype which is a very characteristic and frequent structure type of the higher temperature K-white micas (Velde 1965).

It is noteworthy that chlorite is not present in the assemblage although it appears both in the metasandstone and the neighboring metasiltstone (Árkai et al. 1981; Ivancsics and Kisházi 1983).

← Fig. 5
Microfabrics of the chloritoid schist and the surrounding Rágyincsvölgy Sandstone. a) Crenulation of the first foliation (S_{0-1}) in the Rágyincsvölgy Sandstone: small, lath-shaped chloritoid grains (with high refractory) are enriched in the crenulated, more pelitic, S_{0-1}-parallel layers (dark, ca. sub-horizontal in the middle). Steeply dipping S_2 is also indicated by white, resp. black lines. 1N. b) Microfabrics of the chloritoid schist showing post-tectonic chloritoid both as rosettes and single, lath-shaped grains in the fine-grained matrix. +N

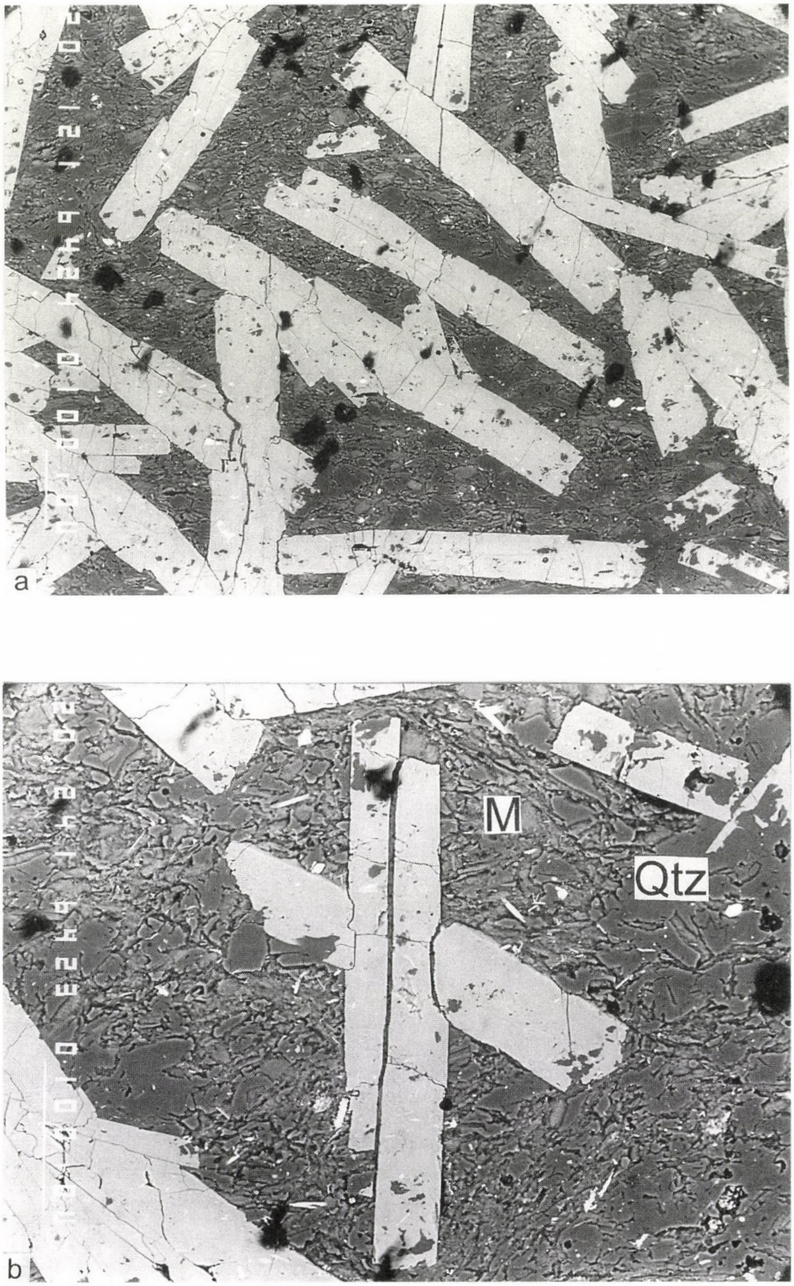

Fig. 7
BSE images of the chloritoid schist. a) Single, idiomorphic, lath-shaped chloritoid grains. b) Penetrating chloritoid twins in the fine-grained matrix consisting of white K-mica (M), quartz (Qtz), feldspar and opaque minerals (tiny, bright grains)

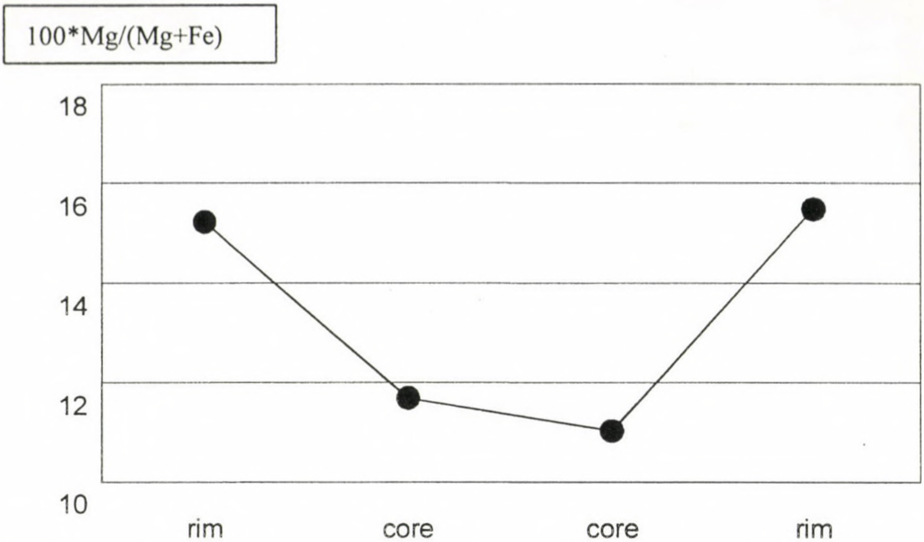

Fig. 8
Zoning profile across a chloritoid grain. Note the increasing X_{Mg} trend from the core toward the rims

Microprobe analyses were performed on chloritoid grains (Figs 7a–b) in order to determine their chemical compositional and zoning characteristics. Representative core and rim compositions are given in Table 1. Structural formulae were calculated on the basis of 12 oxygens. The Si content is close to the stoichiometric values of 2 atoms per formula unit (a.p.f.u.), ranging between 1.918 and 2.013. The Al content is always above 4 a.p.f.u (4.01 to 4.11). The analyzed chloritoids are Fe-rich, with slightly variable Mg content: X_{Mg} (X_{Mg}=Mg/($Mg^{+}Fe^{2+}$) ranges from 0.11 to 0.16. This variation is systematic: it is connected to an observable compositional zoning, with increasing X_{Mg} from core to rim (Fig. 8). The MnO content is between 0.15 and 0.35 wt%, and there is no correlation between MnO and MgO or FeO.

Our microprobe results are in agreement with the data of Árkai et al. (1981). However, they did not detect any compositional zoning in the chloritoid of the metasandstone. This is probably related to the formation of relatively badly developed, often skeletal chloritoid grains in that rock type.

Discussion

Structural features

As was described in detail (see section "Structural observations"), two foliation generations were observed in outcrop scale both on the Rágyincsvölgy metasandstone and the surrounding Tapolcsány metasiltstone and slate: the first

Table 1

Representative microprobe analyses for the core and rim of 2 chloritoid grains from the chloritoid schist

	Grain 1				Grain 2			
	rim	core	core	rim	rim	core	core	rim
SiO_2	23.86	23.67	24.44	23.33	23.10	22.77	22.55	22.51
TiO_2	0.13	0.08	0.15	0.08	0.43	0.70	0.04	0.91
Al_2O_3	41.46	41.56	41.37	41.61	42.04	40.83	40.79	40.50
FeO^*	24.07	24.70	23.40	23.24	22.93	23.40	24.11	22.85
MnO	0.16	0.30	0.34	0.29	0.19	0.17	0.20	0.28
MgO	2.22	1.75	2.42	2.49	2.31	1.74	1.68	2.35
CaO	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00
Na_2O	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00
K_2O	0.01	0.05	0.00	0.00	0.00	0.00	0.00	0.00
Total	91.91	92.14	92.13	91.04	91.00	89.61	89.37	89.40
	Cation numbers on the basis of 12 oxygens							
Si	1.977	1.965	2.013	1.948	1.926	1.937	1.931	1.918
Ti	0.008	0.005	0.009	0.005	0.027	0.045	0.002	0.058
Al	4.050	4.067	4.015	4.094	4.130	4.093	4.117	4.067
Fe^{2+}	1.668	1.715	1.612	1.623	1.599	1.664	1.727	1.628
Mg	0.011	0.021	0.024	0.021	0.013	0.012	0.014	0.020
Mn	0.274	0.216	0.297	0.310	0.287	0.221	0.214	0.298
Ca	-	0.002	-	-	-	-	-	-
Na	-	0.002	0.002	-	-	-	-	-
K	0.001	0.005	-	-	-	-	-	-
Total	7.990	7.999	7.971	8.000	7.982	7.972	8.007	7.990
$Mg/(Mg+Fe^{2+})$	0.14	0.11	0.16	0.16	0.15	0.12	0.11	0.15

is clearly parallel to bedding (S_{0-1}) in the metasiltstone developed due to deep burial (sedimentary and/or tectonic). The second foliation (S_2) cuts it at high angle, resulting in characteristic wedge-shaped splitting. This is a very characteristic structural feature of the investigated rocks, suggesting their common deformation history. Therefore, contrary to Fülöp (1994), we interpret the relatively widely spaced first foliation in the Rágyincsvölgy metasandstone as bedding-parallel first foliation (S_{0-1}). The common tectonometamorphic evolution is also definitely proved by the presence of chloritoid in both rock types.

The steeply SE-dipping S_2 foliation is considered as axial plane foliation of gently NE (resp. SW) plunging folds. These folds are often not directly observable in the outcrops of the Uppony Mts., since they form larger scale structures. Fold axis orientation is parallel to the observed and calculated intersection lineations (Fig. 2) suggesting cylindrical fold geometry. The NE–SW-trending axis direction is very characteristic in the entire Uppony Mts. The post-tectonic character of chloritoid with respect to S_2 foliation shows that its formation occurred after the major Alpine (Cretaceous) folding event of the Uppony Paleozoic sequence.

The described structural features also suggest that the original contact between the Rágyincsvölgy Sandstone and the Tapolcsány Slate may have been sedimentary. It is of course possible that this lithological contact was later re-activated as a smaller, (Tertiary?) brittle fault (as it was indicated on the map of Kovács (1983); see in Fülöp 1994), but this phenomenon often occurs at the contact of lithologies with very different rheological properties. However, the major ductile deformation phase occurred in the same way in both rocks, indicating their close spatial relationship already before the major Cretaceous folding event.

Origin and formation of chloritoid

Noske-Fazekas (1973) argued for the detrital origin of chloritoid in the meta-sandstone. Our data from the chloritoid schist leave no doubt about the metamorphic origin, since the formation of chloritoid of such an idiomorphic appearance and high modal content is only possible during metamorphic processes. This result confirms the previous statements of Árkai et al. (1981) and Ivancsics and Kisházi (1983). Furthermore, it is worth mentioning that a new chloritoid schist occurrence was recently described (Koroknai et al. 2000) from a water-prospecting borehole at Kazinbarcika (Szendrő unit). The encountered dark phyllite with light, thin metasandstone intercalations (Szendrő Phyllite Fm., or possibly the Irota Fm. ?) shows very similar microstructural and mineralogical-petrological features to the chloritoid schist from the Uppony unit. This suggests that the major phase of folding occurred prior to peak metamorphic conditions in both Paleozoic sequences.

In the metamorphic literature there are slightly different opinions on the lower temperature boundary of chloritoid-in reaction. Hoschek (1969) has given the following reaction during progressive metamorphism:

$$\text{chlorite+Al-silicate (kaolinite or pyrophyllite) = chloritoid+quartz+}H_2O \qquad (1)$$

If this reaction was responsible for the formation of chloritoid in our case, it would mean that both chlorite and Al-silicate were totally consumed during reaction (1), since neither microscopic nor microprobe and XRD investigations could detect any of them. This reaction has a very steep, negative slope of about 400 °C in the petrogenetic grid of Hoschek (1969), which is somewhat higher than the estimated temperature values (300–350 °C) of Árkai (1983), obtained by the IC method for the investigated rock type. Microtectonic observations in the surrounding metasandstone and slate (dynamic recrystallization of quartz) also indicate temperatures at least in the range of 270–300 °C (Koroknai and Frisch 1998).

Chloritoid appears in metapelites of the KFMASH system (K_2O–FeO–MgO–Al_2O_3–SiO_2-H_2O) at about 300 °C according to Bucher and Frey (1994, p. 206), and at about 350 °C according to Spear (1995, p. 374). Frey (1987) regarded chloritoid as a typical index mineral of the epizone (over ca. 300 °C) as well. However, it was also described from high-pressure, anchimetamorphic (T<300 °C) terranes (Seidel 1978).

Further theoretical possibilities for the Ctd-in reaction were described by Miyashiro (1973):

$$\text{hematite}+\text{chlorite} = \text{magnetite}+\text{chloritoid} +\text{quartz}+H_2O \quad (2)$$

$$\text{paragonite}+\text{chlorite}+\text{quartz} = \text{chloritoid}+\text{albite}+H_2O \quad (3)$$

Since microprobe and XRD investigations show the presence of both Fe-oxide(s) and albite, it cannot be excluded that reactions (2) and (3) were also active during chloritoid formation. However, the subordinate quantity of albite and Fe-oxides suggests that presumably reaction (1) was the major chloritoid-producing process. Considering the large amount of idioblastic, progressively zoned chloritoid grains, the results of detailed IC study (Árkai et al. 1981) and microtectonic investigations, it seems to be very probable that temperature was over the boundary between anchi- and epizone (c. 300 °C), presumably even reaching 350 °C during he Alpine metamorphism.

Chloritoids with different zoning types were reported by Franchescelli et al. (1997) from the low-grade, chlorite-bearing chloritoid schist in the Apuan Alps, NW Italy ("Breccia di Seravezza Formation"). They interpreted rimward-increasing X_{Mg} ratio in chloritoid as evidence of prograde metamorphic conditions. Although chlorite is missing in our assemblage it seems reasonable that a similar zoning trend indicates prograde metamorphic conditions during the growth of chloritoid.

It is an interesting question what kind of rock could be the precursor of the chloritoid schist with such a high modal portion of chloritoid. Petrographic and field observations indicate that chloritoid schist blocks could formerly have been fine-grained, pelitic sediments forming either ripped-up intraclasts or smaller lenses in the sandstone. Giglia and Trevisan (1967) interpreted chloritoid schist of the Apuan Alps (see previous paragraph) as the equivalents of modern laterites. Those rocks form a meter-thick layer above a Triassic carbonate sequence ("Marni a Megalodonti Fm."). However, this interpretation does not apply to our case. According to Kovács (1992) the metasandstone containing the studied chloritoid schist forms the stratigraphic base of the pelagic, euxinic slate of Tapolcsány Formation. Nevertheless the clear nature of this relationship requires further detailed stratigraphical and sedimentological investigations.

Conclusions

1. Our detailed petrographic, microprobe and XRD investigations on the chloritoid schist from the Uppony Mts. clearly prove the metamorphic origin of chloritoid from this newly described rock type in the Rágyincsvölgy Sandstone.

This is in agreement with the conclusions of Árkai et al. (1981) and Ivancsics and Kisházi (1983) who also regarded the chloritoid from the metasandstone and the neighboring metasiltstone as newly-formed metamorphic mineral, but contradicts the previous idea of Noske-Fazekas (1973) who argued for its detrital origin.

2. Structural observations in the outcrops of the Rágyincsvölgy Sandstone and on the neighboring rocks record that they have experienced the same ductile deformation processes, which are indicated by the consequent relationship of the S_{0-1} and S_2 foliations in the different rock types. The second (S_2) foliation forms axial plane surface of larger-scale fold structures and is related to the Cretaceous folding of the sequence.

3. Post-tectonic growth of chloritoid with respect to S_2 foliation indicates that the major phase of Alpine deformation (e.g. folding of the Uppony Paleozoic sequence) predates peak metamorphic conditions at c. 110–130 Ma (Árkai et al. 1995).

4. XRD investigations show that the chloritoid belongs to the triclinic polytype. This agrees well with the general observation that this structure polytype occurs predominantly in (very) low-grade metamorphic rocks. Zoning profiles (with increasing XMg from core to rims) of chloritoid crystals suggest prograde metamorphic conditions during its growth.

5. Macroscopic and microscopic observations suggest that the precursor of the chloritoid schist must have been a fine-grained, Al-rich pelitic sediment forming smaller (dm-scale, eventually meter-sized) clasts or lenses in the metasandstone. Therefore, our study also demonstrates how important the effect of bulk chemistry on the formation of metamorphic minerals may be. The generally badly-developed, skeletal chloritoid crystals in the metasandstone reflect its original relatively Al-poor bulk chemistry, while the Al-rich chemistry of the pelitic precursor made possible the formation of idiomorphic grains in a large quantity.

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Relationships of thrust-fold and horizontal mechanisms of the Mt. Žumberak part of the Sava nappe in the northwestern Dinarides, West Croatia

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On the basis of structural indicators Mt. Žumberak was emplaced in the tectonostratigraphic unit of the Sava nappe, which is of key importance for the explanation of thrust structures of the northwestern Dinarides and transitional Alpine–Dinaridic region (Zagorje–Mid-Transdanubian Zone). In this region the External Dinaridic units (e.g. Mesozoic carbonate platform) confront Cretaceous flysch formations of the Apulian passive continental margin and allochthonous Triassic–Paleozoic formations. Dominant morphostructure and orographic features are controlled by kinematic processes, which took place along two major fault systems. The fault system striking northwest–southeast facilitated thrusting of Triassic carbonates southeastward over carbonate–clastic Cretaceous formations. The southeastern margin of Mt. Žumberak is limited by the system of sinistral strike-slip faults separating it from Neogene carbonate–clastic deposits. It is assumed that the thrusting preceded the faulting, since the remains of km-sized overturned folds were noted at the southwestern margin of Mt. Žumberak. Progressive thrusting reduced most of the fold, and Cretaceous clastics were overthrust onto part of the overturned fold limb.

Key words: Southern Alps, Dinarides, Sava nappe, fault kinematics, stress orientation

Introduction

Mt. Žumberak (Žumberačka Gora) with neighboring Mt. Samobor (Samoborska Gora) forms a unique orographic unit with a significant geologic setting in the marginal part of the northwestern Dinarides and Southern Alps; in this area the tectonostratigraphic units of the Internal and External Dinarides and Southern Alps come together. Therefore this area was grouped into different geotectonic units, depending on the approaches of different authors (Petkovic 1958; Dimitrijević 1982; Herak 1986, 1991; Pamić and Tomljenović 1998).

The aim of this paper is to present new geologic data for the area of Mt. Žumberak, with an emphasis on structural relationships. The description of the main structural elements is supported by the schematic presentation of fold-overthrust relationship development of the southwestern margin of this allochthonous terrane. On the other hand, along its southeastern margin, notable morphological and structural characteristics are defined by a system of left lateral

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strike-slip faults. The study is based on data collected during work on the Geologic Map of the Republic of Croatia at 1:50000 scale.

Overview of previous investigations

First observations on specific differences in the orientation of the morphostructural elements in Mt. Žumberak were made by Gorjanović-Kramberger (1894). He distinguished the SW–NE and NW-SE striking structures and interpreted them as a consequence of polyphase tectonics.

Prelogović (1970), Šikić and Basch (1975) contributed to the understanding of neotectonic movements. The stratigraphy of Hauterivian–Cenomanian deposits and position of the boundary between carbonate platform and adjacent basinal areas were studied by Babić (1974). Later, Pleničar and Premu (1977) classified the northwestern part of the region, which belongs to the border area of Slovenia, into the Balaton Range.

Šikić et al. (1979) delimit the southeastern Croatian margin of this area with the subsided Tertiary block of the northwestern Sava Depression. Within the area of the southern Žumberak and Pokuplje Bukovac (1988) distinguished between structures belonging to the External Dinarides and those of the Internal Dinarides. Contemporaneously, in the area of Duralija (SW Žumberak), Herak and Bukovac (1988) indicated the relationship between the Vivodina flysch and Triassic dolomites as significant evidence for the overthrusting relationship of the External and Internal Dinarides.

Bukovac and Sokać (1989) explained the contact of Upper Triassic and Lower Liassic limestone units as the consequence of overthrusting. Bukovac et al. (1995) proposed the first detailed lithostratigraphic subdivision of Mesozoic units of Mt. Žumberak and the southwestern part of Mt. Samobor, which will be partly used in this paper.

Geologic setting

The investigated area, covering 180 km^2 (Fig. 1), has considered a part of different tectonic units: 1. Inner Dinaridic Zone (Petković 1958), 2. High Karst Zone (Grubić 1980), 3. Julian Alps–Sava folds–Zagreb Zone (Dimitrijević 1982), 4. Sava nappe (Mioč 1982; Pamić 1993), 5. Supradinaricum (Herak 1986, 1991), 6. Middle Transdanubian Zone (Fülöp et al. 1987; Haas et al. 1995) and 7. Zagorje–Mid-Transdanubian Zone (Pamić and Tomljenović 1998).

Fig. 1 →
Tectonic map of Mt. Žumberak modified after Bukovac et al. (1983, 1985). 1. Quaternary; 2. Neogene; 3. Upper Cretaceous–Maastrichtian; 4. Lower-Upper Cretaceous (Hauterivian-Cenomanian); 5. Jurassic; 6. overturned Upper Triassic near Duralije; 7. Upper Triassic; 8. Middle Triassic; 9. Lower Triassic; 10. nappe front; 11. sinistral strike-slip fault; 12. reverse fault; 13. normal fault; 14. fault with marked dip of the fault plain; 15. supposed fault; 16. normal and overturned b-axis; 17. normal and transgressive boundary; 18. normal and overturned bed. Legend for the inserted map: Ž – Žumberak; ID - Internal Dinarides; JSN – Julian–Savinja nappe; SN – Sava nappe; ADCP – Adriatic-Dinaridic carbonate platform; ZZ – Zagreb-Zemplen Fault; PL – Periadriatic lineament; TS – Tisia


0 1 2 km
Carbonate platform
RADOVICA
VIVODINA
Sošice
Drašći vrh
Tihočaj
Durdjići
Goljak
Puškarevi
Konjarić vrh
Rodina Gorica
Hutin
N
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
AUSTRIA
SLOVENIA
ITALY
JSN
SN
PL
ZZ
ZAGREB
ADCP
TS
ID
BOSNIA AND HERZEGOVINA
ADRIATIC SEA
0 50 100 km

On the basis of field observations it seems logical to include Żumberak into the Sava nappe and in the southwesternmost part of the Zagorje–Mid-Transdanubian Zone, respectively.

In this part of the Sava nappe the older part of Lower Triassic formations (usually referred to as "Seiser Schichten") are represented by varieties of dolomite-mica siltite and silty shale, approximately 300 m thick (Fig. 1). The younger part of the Lower Triassic deposits (usually referred to as "Campier Schichten"), of approximately the same thickness, are represented by tabular, marly biomicritic limestone and calcareous marly shale, as well as micrite with laminae of silty marl and marly shale (Pleničar and Premu 1977).

Younger Triassic sequences begin with the Anisian–Ladinian Ruškovlje Formation, about 590 m thick, which is composed mostly of early diagenetic dolomite (Bukovac et al. 1995) (Fig. 1). However, periodical interruptions of carbonate sedimentation were characterized by influx of clastic–pyroclastic material in their lower and upper parts. The origin of these sediments in the wider Dinarides area is ascribed to rift processes which were most intensive during the Middle Triassic (Pamić 1984). The upper part of this formation is represented by thin-bedded limestone and chert. The Upper Triassic begins with the fossiliferous Slapnica Formation (Bukovac et al. 1995), which is 360 m thick and characterized by rhythmical alternation of 20–65 cm-thick beds of dolomicrite, fenestral dolomicrite and dolomite stromatolite (Fig. 1). Two stratigraphic members can be distinguished within this formation: the Lower Vranjak unit is characterized by locally up to 50 cm-thick intercalations of yellow-green shale, while in the overlying Drenovac unit black dolomicrite and thin coal intercalations occur.

The uppermost part of the Triassic is represented by the Norian–Rhaetian Hauptdolomit Formation, 1200 m thick. It is characterized by an irregular lateral and vertical alternation of early diagenetic dolomicrite, fenestral dolomicrite and dolomitic stromatolite. Some layers are up to 40 cm thick.

The Triassic formations of Mt. Žumberak, except for minor differences in the Anisian, show great similarity to Middle Triassic sediments of the neighboring, northwestern extension of the Sava nappe in Slovenia (Mioč 1982; Pamić et al. 1998). Identical spilite-keratophyre volcanics overlain by Upper Triassic Hauptdolomit, were also noticed in the southeastern extension of the Sava nappe toward Bosanski Novi (Šikić 2000), as well as in the hanging wall of the Sana–Una nappe (Maksimčev and Jurić 1964).

The Mesozoic carbonate succession of Mt. Žumberak ends with the Rajići unit, which is represented by thin-layered to platy recrystallized micrite and/or intrasparite ranging in age from Liassic to Tithonian (Gušić and Babić 1970), which have all the characteristics of condensed sedimentation. In the southwestern part of the study area they have been tectonically reduced, while in the northeastern part they are overthrust by Albian–Cenomanian clastics; thus, their total thickness cannot be determined. On the basis of the correlation with similar formations in the northwestern extension toward Slovenia their thickness can be estimated at 100 m.

The Triassic and Jurassic formations were thrust over Cretaceous flysch composed of breccia, calcarenite and shale sporadically grading into siltite, marly shale and subgraywacke with intercalations of pelagic micrite and chert (Scaglia) of Albian–Cenomanian age (Babić 1974) (Fig. 1). In the area of Vivodina an Early Maastrichtian age was determined for these deposits along the southwestern wedge of Mt. Žumberak (Devidé-Nedêla et al. 1982) (Fig. 1).

These Cretaceous clastic formations correspond to the deposits of the southeastern part of Slovenian Trough (Cousin 1972) which could be correlated to the "flysch bosniaque" (Aubouin et al. 1970; Blanchet 1975) or "Sarajevo–Banja Luka flysch" (Pamić et al. 1975; Mojičević et al. 1979) of the Central Dinarides.

The youngest sediments are Lower Miocene clastic–carbonate formations, which occur, in the spacious depression along the southeastern margin of Mt. Žumberak. In the central part of Mt. Žumberak several small, isolated erosional remaining occurrences, 1–2 m thick, can also be found.

Structural data

The investigated area obtained its main structural features during Middle/Late Eocene compression, which comprised the entire area of the Dinarides characterized by NW–SE-striking fold-thrust-imbricate structures. Mesozoic platform carbonates along their southwestern margin are in fold-thrust relationship with Upper Cretaceous clastics along the older fault system, which is parallel to the front of the Sava nappe. However, recent morphostructural and orographic characteristics originated after the Late Miocene, when old folds and reverse faults were affected by NE–SW-directed strike-slip faulting. Mt. Žumberak is separated from Neogene carbonate–clastic sediments by a major subvertical fault, which is included in the NE–SW-striking Zagreb–Zemplen fault system.

The NE–SW-striking fault system is characterized by gently curved bedding planes and/or an anastomozing pattern. Sinistral movements are observed on the majority of fault planes and the biggest throw is noticed along the marginal fault (Fig. 2). On the fault paraclasis 6 meter-long striations with left lateral shift and 250 dip angle were found. From Goljak (Fig. 2) westward the fault zone is marked with more or less cataclased dolomite. Near Puškarev Jarak and Hutina a dense fracture system with 5–20 cm fractures, each with millimetric striations within this dolomite, occurs.

Along the fault plane the total sinistral movement is not visible because Middle Triassic sediments are in contact with Neogene deposits. In this zone Neogene deposits are locally steepened to 600 and slightly folded.

A more important, though indirect, indication of complete kinematic deformations in this fault zone is visible in its northwestern block. In this segment

Fig. 2
Fault plain of the marginal Mt. Žumberak fault with diagonal striations (near Goljak)

there occurs on the surface a 1–1.5 km wide zone of Middle Triassic dolomite which is in contact with Neogene sediments. In its southeastern part and toward the northwest this zone is bounded by the fault striking from Tihočaj to Konjarić vrh, that is, subparallel to the previously mentioned main strike-slip fault. The northwest block of this fault was shifted southwestward by 1.5 to 2 km. As in the previous case the fault plane is subvertical to vertical, with small inclinations in northeast or southwest directions.

On the surface it is marked by 10 cm-thick lenses composed of pulverized rock material and/or slickenslides, which are often subparallel to the fracture system, with about one millimeter wide striations. This fault separates Middle Triassic from Upper Triassic dolomite of the northwestern fault block.

The dip of the fault plane ranges between 10–40 degrees which, together with its normal movement character, indicates a regional compression axis σ_1=354 to 5 degrees and a regional extension axis σ_3= 98 to 90 degrees as determined by the Angelier and Mechler (1977) method. Within the block delimited by these two faults a horizontal decrease of width and vertical uplift of Middle Triassic dolomite was determined. An identical geometrical relationship of faults, together with a record of lineations on the fault plane, can be followed in the inner part of Żumberak, indicating widely present east-west extension and north-south compression.

The southeastern margin of Mt. Žumberak represents a morphological step block (e.g. marginal fault zone) formed due to the uplift of the northwestern block and the subsidence of the southeastern block. The diagonal shift of the blocks parallel to the Goljak–Hutin marginal fault resulted in overturning of Miocene beds and reduction of their proximal parts.

The structural relationships of the southwestern edge of the Mt. Žumberak Triassic carbonate succession are more complex, since in this part faults with horizontal movements cut across the Sava thrust front. Interactions of these two systems caused disharmonic movements and the rotation of blocks in the hanging wall of the previously formed thrust. The rate of movement of the blocks in the south and southwest directions is not equal; a gradual increase of the thrust index from the northwestern to the southeastern margin is obvious.

In this area NW–SE-striking faults with reverse character predominate. The reverse faults were reactivated during younger periods and thus changed their primary reverse character, becoming a conjugate system for the NE–SW-oriented sinistral strike-slip faults. Faults with southeastern strike are mainly marked by cataclased Triassic dolomite, typically grouped in zones as follows:

- cataclasites, fault breccia (belonging to the distal parts of the fault zone),
- completely disintegrated dolomite (transitional zone), and
- pulverized rock and/or slickenslides (sliding surfaces).

This zonality is not always completely developed along the strike, since some segments are missing at certain localities. However, the occurrence of any of these zones indicates proximity of sliding surfaces or of the fault contact itself (Fig. 3).

The front of the Sava Nappe is marked by a wide zone of cataclased Triassic carbonates, which can be traced from Sošica in the northwest to Radina Gorica in the southeast. The main fault plane is commonly not visible at the surface since the rigid dolomitic sediments were pushed over the more plastic Cretaceous clastics during thrusting. Only locally, as in the neighborhood of the village of Konjarić Vrh, faulted folds were noticed in underlying Cretaceous calcarenite and shale. The folds have rounded hinges, an irregular geometry and decimeter to meter amplitudes. The axes of these folds plunge northwestward with an average B= 320/20 degrees and axial planes have a distinct southern vengeance.

The frontal part of the thrust is divided by the afore-mentioned sinistral, NE–SW-directed strike-slip faults, which shifted parts of the thrust front with different intensity, subsequently reflecting the varying thrust indices (Fig. 4).

Especially complex relationships can be followed in the wider neighborhood of the Duralija area, where overturned Upper Triassic dolomite thrust by Cretaceous dolomites has been registered for the first time during this research.

A virtually normal succession is, in this case, a consequence of complex kinematic processes. This example is specific, since overturned dolomite

Fig. 3
Fault plain of the reverse fault with lenses of silt-sized crushed (pulverized) rock and cataclased dolomite (near Sošica)

represents the remnants of the southwestern limb of an overturned and later eroded kilometer-size fold (Fig. 5).

Accordingly, Cretaceous clastics of the Veliki Oštrc Peak area near the village of Duralije lie on the overturned Hauptdolomit unit, and in their basal part contain tectonic breccia and/or recrystallized limestone and dolomite. In the apical part of these clastics meter and decameter-size faulted folds are found. A thrust block in front of which clastics were squeezed out has a flexure-shape structure, with the core composed of Middle Triassic dolomite. Such structural relationships are the consequence of progressive stress action that formed shallow sliding surfaces, over which pushing of the back parts of the hanging wall forward took place. The main thrust plane was differentiated into smaller parts and, consequently, in this part has a duplex structure. Accordingly, Cretaceous clastics of the Duralija area are klippen on the overturned fold, composed of Triassic carbonates of kilometric amplitude, remaining after erosion and/or consummation of its northeastern

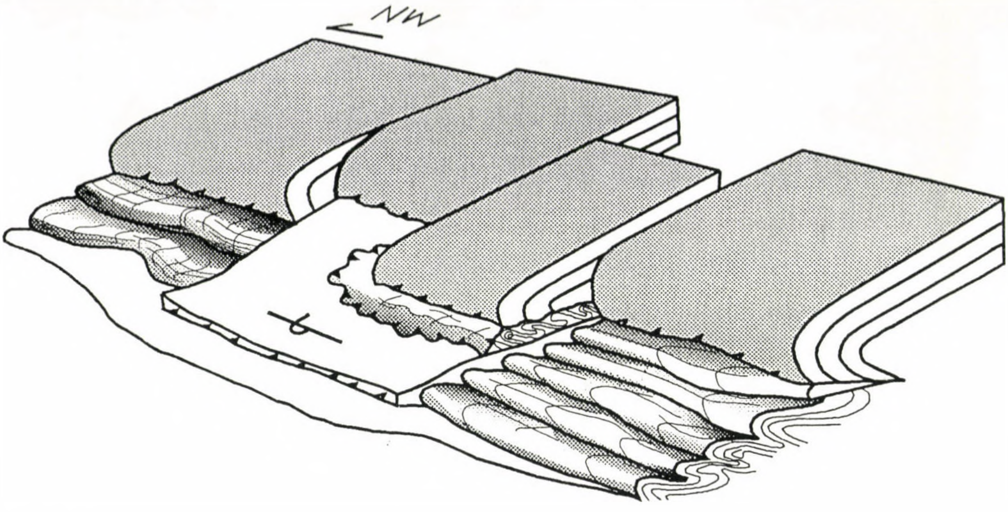

Fig. 4
Schematic block diagram of the southwestern margin of Mt. Žumberak

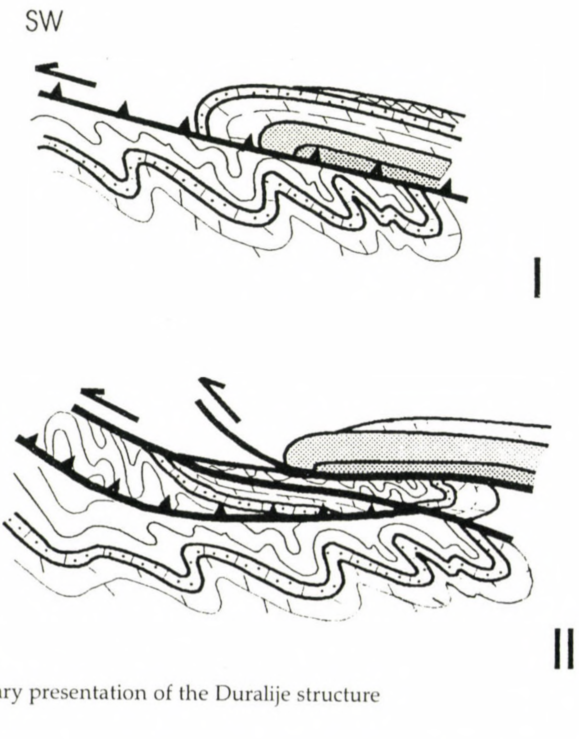

Fig. 5
Schematic evolutionary presentation of the Duralije structure

limb. Extension of the overturned part of the structure southeastward was tectonically completely reduced, implying more than ten kilometers of thrusting of the carbonate part of the structure over Upper Cretaceous clastics.

A subparallel reverse fault extends about 1–2 km northeastward of the frontal part of Sava nappe along the River Kupčina. This shear zone is a cataclased and recrystallized dolomite zone, locally up to hundred meters wide. Along its strike southeastward drag folds are developed as the result of the differential sliding velocities of the reverse-fault hanging wall (Fig. 6). They have rounded hinges, flexure shape, and the axes are oriented NW–SE with southern vergencies. Shale intercalations of the Slapnica Formation also bear drag folds with centimetric amplitudes, with southern and northwestern vergencies. Similarly, they indicate shear between different layers in the thrust front. This indicates that larger amounts of thrusting took place along the fault plane of the first (outer) fault in this system, similar to "the leading imbricate fan" (Ramsay and Huber 1987).

N–S oriented faults opened in the youngest kinematic period and thus contributed to the complexity of this part of the studied area. They are associated with the hanging wall of the Sava nappe and can be followed within smaller blocks without cutting the older faults. Vertical movements took place along these faults and they mostly show normal character. Fault planes are subvertical to vertical (75–90°), and can be traced along strike up to one kilometer. Throw of the blocks along this system have decametric to hectometric dimensions and do not significantly disturb structural relationships. However, interactions of this system and the NE–SW strike-slip fault system gave rise to irregular displacement of the blocks in the hanging wall of the thrust, and their crashing produced folds with gently dipping (5–10°) limbs of hectometric dimensions. Since axes of these structures show divergent strike it may be concluded that some blocks were probably exposed to the influence of a dispersed (local) stress field, also indicating recent allochtony of Mt. Žumberak.

Discussion

The investigated area, though relatively small, is of key importance in the explanation of thrust structures of the northwestern Dinarides and transitional Alpine–Dinaridic area (e.g. Zagorje–Mid-Transdanubian Zone). In this area the units of the External Dinarides (Mesozoic carbonate platform), Cretaceous flysch of the passive continental margin of Apulia and allochthonous, mainly Triassic, formations of the Sava nappe are included.

The lithostratigraphic characteristics of Triassic units of the Sava nappe show identical development to those of the Outer Dinaridic Mesozoic carbonate platform. By structural processes these formations were detached from the basement, i.e. the distal Adriatic–Dinaridic carbonate platform (ADCP). Regionally the beginning of the disintegration could be linked with the Lower Jurassic opening of the Tethys when clastic–carbonate formations of the Apulian

Fig. 6
Drag folds in dolomite in the Slapnica Formation near the Kupčina river

shelf subsided. Later, during the Jurassic and Cretaceous the Triassic formations of the present Sava nappe represented the basement of the southwestern parts of the Dinaridic Tethys on which the Mesozoic oceanic crust was generated. The occurrence of Jurassic condensed carbonates in the apical part of the Sava thrust of Mt. Žumberak, as well as in the allochthonous Triassic units of the southwestern Dinarides, suggest that the process of disintegration of the northeastern margin of the ADCP and its subsidence could be of regional importance (Pamić 1993).

In the area of Mt. Žumberak the Cretaceous period was characterized by deposition of flysch-turbiditic sediments, which commenced in the Albian and lasted until the Senonian, gradually expanding southward and southwestward within the Dinaridic Tethys. Continuous tectonic changes, most probably radial in character, are indicated by facies variability (basal clastics, delta and prodelta sediments, biolitites, pelagic sediments, flysch-turbidites, debrites). These movements culminated at the end of the Eocene when the area of Žumberak and its wider surroundings were affected by the main Alpine compressional deformation.

The end of the Eocene (45–40 Ma) and the beginning of the Oligocene also represented a period of major uplift, followed by the creation of major structural units. Structures formed in this kinematic period are characterized by km-sized folds, which are partly overturned with southwest vergency and NW–SE strike. Along the southwest flank of the Žumberak area, from Sošica to Konjarić Vrh and Radina Gorica, in the present structure there are remnants of overturned folds of kilometric amplitude, which were detached progressively and displaced southwestward and southward for more than 10 km over the Cretaceous clastics of the Apulian passive continental margin.

The recent morphostructural and orographic features of the investigated area were acquired by the end of the Miocene (10–6 Ma). In this period the older structures of Dinaridic strike were influenced by reorientation of the stress field (north–south). The hinge of the Sava nappe and its parallel system of faults with NW–SE strike was reactivated. Therefore the previous reverse faults became dextral strike-slip faults. Simultaneously, along the Goljak–Hutin fault and the NE–SW fault systems parallel to it, sinistral movement was taking place. Through interaction of these two systems lateral shift and extrusion occurred.

In the evolution of the study area, as well as in the neighboring southern parts of the Pannonian Basin, the northeast-striking fault system played an important role in the youngest kinematic periods (Prelogović et al. 1995). during the Late Pliocene sinistral to diagonal movements of blocks took place along it, giving rise to the creation of local zones of transpression/transtension. In the part of the structure where the system of subparallel faults cuts dolomite it is reflected in the horizontal reduction of width and uplift of dolomite blocks. However, in the zone of Goljak–Puškarov Jarak–Hutin fault, lateral (east–west) movement took place on the account of more plastic Miocene clastic-carbonate sediments of the southeastern limb. Therefore, in the marginal zone of Mts. Žumberak and Samobor, 60° bedding dip is accompanied by gentle folding and reduction of proximal parts of Miocene sediments.

Based on all aforementioned data it can be concluded that the main role in the formation of the recent structure of Mt. Žumberak can be attributed to north–south compression and east–west extension. Moreover the system of NE–SW sinistral faults played a major role at regional scale as well, since it represented the most noticeable western part of the subvertical NW–SE-striking Zagreb–Zemplen fault system (Pamić and Tomljenović 1998). This fault system bounded Mts Žumberak and Samobor to the southeast, and locked the southwesternmost part of the Zagorje–Mid-Transdanubian zone between the Dinarides and the Alps.

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Two new rare earth minerals in an unusual mineralization of the Nissi bauxite deposit, Greece

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The Nissi bauxite deposit in the Lokris area of Greece represents a transitional deposit between the karst bauxites and karstic nickel deposits. It is characterized by high enrichment of the rare earth elements and Ni at the base of the deposits where authigenic rare earth minerals and Ni clay-like mineral have been formed. XRD and microprobe study of these materials revealed an unusual association of takovite, halloysite, gibbsite, lithiophorite, birnessite, bastnäsite and two new rare earth minerals: hydroxylcarbonate-(Nd) and hydroxylcarbonate-(La). According to the XRD and microprobe analyses the two new rare earth minerals are the end members of the series (La, Nd)CO_3F – (La, Nd)CO_3OH, belonging to the bastnäsite group.

Key words: new rare earth minerals, bastnäsite, karst bauxite and nickel deposits, Greece, Lokris area, XRD and EPMA study

Introduction

The study of trace elements in karstic bauxites in Greece on a large scale was begun in 1973 by Maksimović and Papastamatiou. They found different trace element content in bauxites of different ages and also an enrichment of some trace elements at the base of the deposit, above the carbonate footwall. This enrichment was particularly marked for Ni and rare earth elements (REE) (Maksimović 1976; Maksimović and Roaldset 1976). As a result of this discovery the authigenic minerals of Ni and REE were soon found, including some new minerals: brindleyite, synchysite-(Nd), monazite-(Nd), hydroxylbastnäsite-(Nd) (Maksimović and Bish 1978; Maksimović and Pantó 1978, 1980, 1985). Further studies, which included nickel laterites and bauxites (Rosenberg 1984; Valeton et al. 1987), gave more data on trace elements in these deposits and explained their relation to ultramafic parent rocks. More detailed study comprised geochemistry of the REE in karst-bauxites and karstic nickel deposits (Maksimović and Pantó 1991; Maksimović et al. 1993), as well as mineralogy of the authigenic RE minerals in these deposits (Maksimović and Pantó 1996). The latest study presented geochemical and mineralogical characteristics of Fe–Ti- and bauxitic-lateritic deposits in Greece (Eliopoulos and Economou-Eliopoulos 2000), based on large

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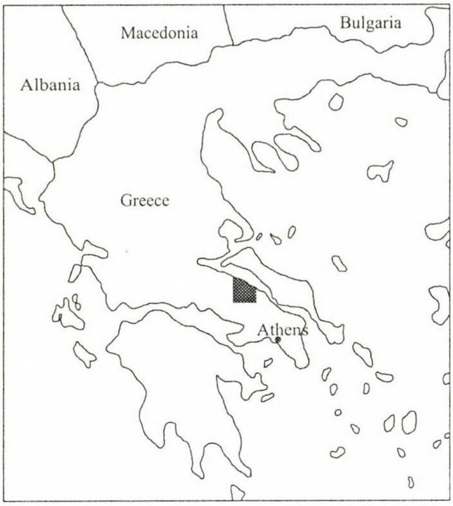

Fig. 1
Location of the Lokris area in Greece

number of trace element data, including REE.

The Nissi bauxite deposit occurs in the Lokris area of continental Greece (Fig. 1), relatively close to other bauxite and karstic nickel deposits in this area. The closest (3–4 km), Aghios Ioannis and Marmeiko, are karstic nickel deposits and their mineralogy and geochemistry was presented elsewhere (Maksimović et al. 1993). The Nissi deposit is a ferruginous bauxite deposit, transitional between karstic bauxites and karstic nickel deposits. The deposit is excavated and only the lower part of it remained, where authigenic takovite and bastnäsite were discovered (Rosenberg 1984).

All these deposits belong to the same stratigraphic horizon. The footwall limestone is of Middle to Late Jurassic age. The hanging wall, transgressively overlying the Ni–Fe ores and bauxites, is made up of Early Cretaceous limestone. From the stratigraphic point of view these deposits are equivalent to the B2 bauxite horizon within the Parnassos Ghiona geotectonic unit (Combes 1977, 1978; Biermann 1983; Rosenberg 1984).

Materials and methods

The samples of Ni–Fe clays and ferruginous bauxitic clays were collected along a vertical profile in the lower, accessible part of the Nissi deposit. The position of samples is given in Fig. 2. Mineralogical composition was studied by means of XRD, chemical and thermal analyses. Authigenic RE minerals were analyzed with a JEOL Superprobe JXA-733 in the Laboratory for Geochemical Research of the Hungarian Academy of Sciences, Budapest. An operating voltage of 20 kV was used and a beam current of 0.05 A. The data were corrected using the ZAF method adapted to the instrument by JEOL Ltd. The standards used were Y-Al garnet for Y; apatite for Ca and F; synthetic glass

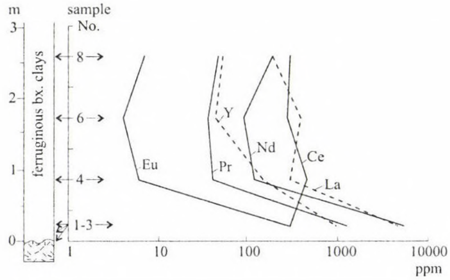

Fig. 2
Distribution of the REE in a profile of the lower part of the Nissi bauxite deposit, Greece. Footwall is made of karstified Jurassic limestone

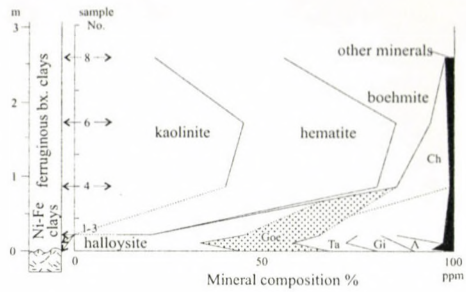

Fig. 3
Mineralogical composition of the lower part of the Nissi bauxite deposit. Ch = chlorite; Goe = goethite; Ta = takovite; Gi = gibbsite; A = asbolite; Other minerals: hydroxylcarbonate-(Nd, La), hydroxylbastnäsites, calcite and anatase

standards of Drake and Weill (1972) for rare earth elements. Rare earth elements (REE) in the Ni–Fe clays and bauxitic clays were determined by ICP-AES (Maksimović et al. 1993; Skarpelis et al. 1995).

Mineralogy of the deposit

Mineralogical composition of the lower part of the Nissi deposit, derived from all available data, is presented in Table 1 and Fig. 3. Three samples, N1, N2 and N3 were collected at the base of the deposit, 0–30 cm from the footwall limestone. They have similar mineralogy consisting mainly of hydrated halloysite, goethite,

Table 1
Mineralogical composition of samples in the lower part of Nissi bauxite deposit, Greece (in %)

Minerals	N1	N2	N3	N4	N6	N8
Kaolinite	-	-	-	40.0	45.0	21.4
Chlorite	-	-	-	14.0	5.0	Tr
Boehmite	-	-	tr	6.0	8.0	40.0
Hematite	-	-	10.0	39.4	41.0	37.2
10 A Halloysite	43.4	33.0	45.0	-	-	-
Gibbsite	9.4	13.3	6.0	-	-	-
Goethite	24.0	25.2	20.0	-	-	-
Takovite	12.0	14.0	10.0	-	-	-
Asbolite	5.0	12.0	3.5	-	-	-
Calcite	4.0	tr	3.3	-	-	-
REE-minerals	2.0	2.3	2.0	tr	-	-
Anatase	0.2	0.2	0.2	0.6	1.0	1.4
	100.0	100.0	100.0	100.0	100.0	100.0

N1–N3 = samples of Ni-Fe clays 0–30 cm from the footwall; N4 = sample 75 cm from the footwall; N6, N8 = samples of ferruginous bauxitic clays; for position of samples see Fig. 2 and Fig. 3

gibbsite, takovite and asbolite. Takovite, Ni_6Al_2 $(OH)_{16}$ (CO_3). $_4H_2O$ (Bish and Brindley 1977), is the major nickeliferous clay-like mineral, indicating an intermediary character of this deposit, between bauxites and karstic nickel deposits. This mineral, as well as bastnäsite and asbolite were described in this deposit by Rosenberg (1984). RE minerals, including hydroxylbastnäsite-(Nd, La) and hydroxylcarbonate-(Nd, La), were also determined in these three samples (Maksimović and Pantó 1996), in amounts of 2.0–2.5%. XRD study of black asbolite revealed the presence of manganese minerals: lithiophorite, rancieite and birnessite.

The upper three samples, N4, N6 and N8, have similar mineralogical composition, representing the ferruginous bauxitic clays: the major phases are kaolinite and hematite, with variable amounts of chlorite and boehmite. The content of boehmite is increasing upwards, representing probably the major mineral phase in the excavated part of the bauxite deposit.

The abrupt change in the mineralogy between two set of samples (N1–N3 and N4–N8) is the result of high enrichment of Ni, Mn, and REE, and formation of authigenic minerals of these elements above the footwall limestone (Fig. 2, Table 1).

Chemical properties of the RE minerals

The RE minerals occur in the black Mn–oxide concentrations, and were not found in the pale, bluish-green Ni–rich material (Fig. 4). They form irregular segregations, micropore and space fillings and microveins (Fig. 5). Larger grains up to 70x100 μm in diameter are rare. The microprobe analyses of the RE minerals in samples collected (N1–N3), as well as in the sample F. Rosenberg kindly sent to Gy. Pantó, have shown variable composition (Table 2 and Table 3). Each analysis represents an average of 3–5 analyses of a grain.

Variations in the composition of the RE minerals are mainly expressed in terms of La and Nd content, two leading elements in these minerals. In addition to substitution in the REE sites, there is also substitution in the F site by OH groups (Kirillov 1964; Maksimović and Pantó 1985, 1996). In case of analyzed RE minerals in Table 2 and Table 3, this substitution is partial to complete. Therefore, the following RE minerals could be defined: a) hydroxylcarbonate-(Nd), b) hydroxylcarbonate-(La), and c) hydroxylbastnäsite-(La). The first two hydroxylcarbonates represent two new rare earth minerals found only in this locality. These minerals are enriched in the light rare earth, generally in order Nd>La, or La>Nd, with Ce being totally absent. Although endmember hydroxylbastnäsite has been synthesized (Christensen 1973), naturally occurring hydroxylbastnäsite usually contains an appreciable amount of F (Maksimović and Pantó 1996). Only in the Nissi deposit were two hydroxyl analogues of bastnäsite found, representing the endmembers of the series (La, Nd)CO_3F–(La, Nd)$(CO_3)OH$, which belong to the bastnäsite group.

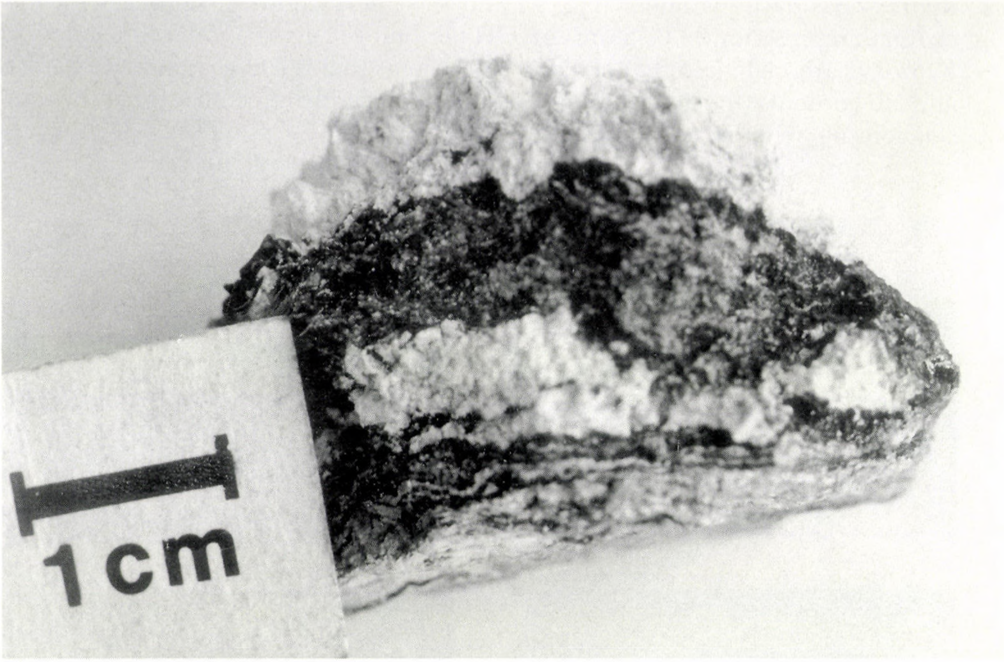

Fig. 4
Black Mn–concentrations and pale, bluish-green Ni-rich clay-like material at the contact with the footwall limestone

X-ray powder diffraction study

The XRPD patterns of the black Mn–concentrations ("asbolite") and pale, bluish-green Ni-rich clay-like material revealed the presence of an unusual association of Mn, Ni, Al and RE minerals. In the pale, bluish-green material the major constituents are hydrated halloysite, takovite and gibbsite (Fig. 6). In the black "asbolite" the major phases are: lithiophorite, rancieite and birnessite, as manganese minerals, with some reflections of calcite and RE minerals: hydroxylbastnäsite-(Nd, La) and hydroxylcarbonate-(Nd, La) (Fig. 7).

An XRPD study of the black samples from the Nissi deposit has shown reflections due to minerals from the bastnäsite group. In most of the samples, however, only the strongest reflection d112 was observed, which varied also in its d spacings from d = 2.875 Å to d = 2.930 Å (Fig. 8). The reflection with the highest d112 value appeared in the sample from Nissi deposit where fluorine was absent in the analyses of hydroxylbastnäsites. The substitution of OH for F in the structure of bastnäsite is likely to lead to an increase in the unit cell volume due to the different sizes of these ions, and in the value of the strongest d112 reflection (Table 4). The increase of the unit cell volume from bastnäsite to hydroxylbastnäsite is linear (Fig. 9), and the intersection of the line with the ordinate representing 1.00 atoms of OH per unit cell gives a unit cell volume of hydroxylcarbonate-(Nd, La) of 455.5 Å3. It seems that relative changes to La, Ce and Nd contents (the major cations) do not significantly affect the size of the unit cell volume, whereas substitution of OH for F does.

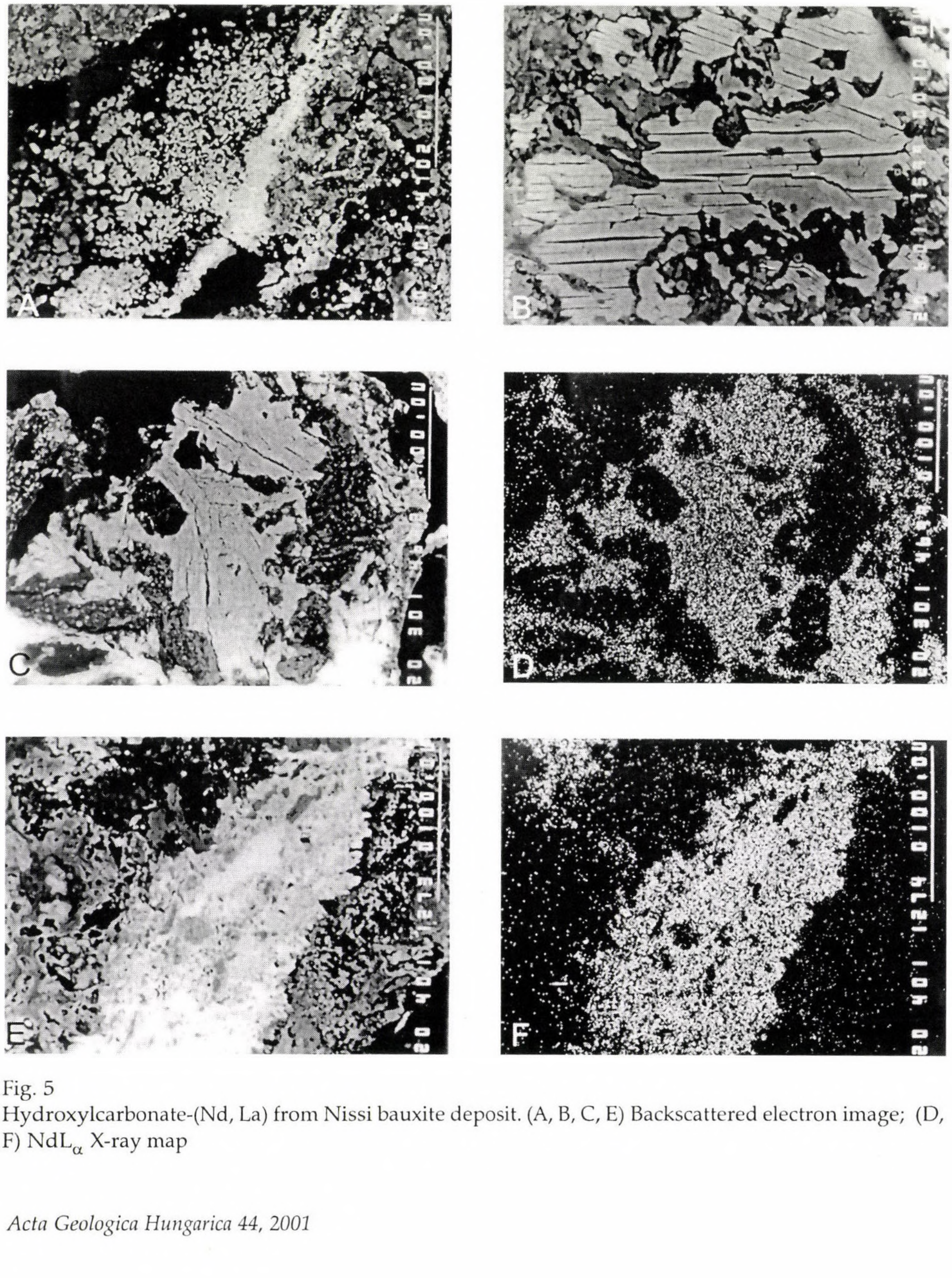

Fig. 5
Hydroxylcarbonate-(Nd, La) from Nissi bauxite deposit. (A, B, C, E) Backscattered electron image; (D, F) NdL_{α} X-ray map

Table 2
Microprobe analyses (wt%) of authigenic hydroxylcarbonate-(Nd) from Nissi bauxite deposit, Greece

	1	2	3	4	5
La_2O_3	17.9	17.3	27.5	27.7	26.0
Ce_2O_3	–	–	–	–	–
Pr_2O_3	7.2	8.0	8.5	7.3	9.1
Nd_2O_3	32.5	34.5	32.3	29.9	33.1
Sm_2O_3	6.1	5.3	3.7	4.7	3.9
Eu_2O_3	1.4	1.4	0.9	1.5	0.9
Gd_2O_3	2.6	2.1	1.5	2.6	1.5
Tb_2O_3	–	–	–	–	–
Dy_2O_3	0.3	0.3	0.2	0.7	–
Er_2O_3	–	–	–	0.1	–
Yb_2O_3	–	–		0.2	–
Y_2O_3	0.3	0.6	0.1	0.1	0.2
CaO	5.5	4.5	0.1	–	0.1
F	0.4	0.2	0.1	–	–
H_2O^*	3.8	3.8	4.1	4.2	4.2
CO_2*	22.1	22.1	21.0	21.0	21.0
	100.1	100.1	100.0	100.0	100.0
–O=F	0.1	0.1	0.04	–	–
Total	100.0	100.0	99.96	100.0	100.0
	Atomic ratios calculated to (O, OH, F) = 4				
La	0.22	0.22	0.36	0.37	0.34
Pr	0.09	0.10	0.11	0.10	0.12
Nd	0.39	0.42	0.41	0.38	0.42
Sm	0.07	0.06	0.05	0.08	0.05
Eu	0.02	0.02	0.01	0.02	0.01
Gd	0.03	0.02	0.02	0.03	0.02
Tb	–	–	–	–	–
Dy	<0.01	<0.01	<0.01	0.01	–
Er	–	–	–	<0.01	–
Yb	–	–	–	<0.01	–
Y	<0.01	<0.01	<0.01	<0.01	<0.01
Ca	0.20	0.17	<0.01	–	<0.01
F	0.04	0.02	0.01	–	–
OH	0.86	0.87	0.98	1.00	1.00
C	1.03	1.03	1.02	1.02	1.02

* The presence of carbon was detected by electron probe. H_2O is assumed to make up the remainder

Genetic consideration

Distribution of the REE along a vertical profile in the lower part of the Nissi deposit shows a high concentration of all REE, except Ce, above the footwall limestone (Fig. 2; Table 5). This trend of concentration was found in all karstic bauxites and karstic nickel deposits formed in situ (Maksimović and Pantó 1996), and represents the primary distribution pattern formed during the bauxitization process. The enrichment of the REE resulted in the formation of RE minerals

Table 3
Microprobe analyses (wt%) of authigenic hydroxylbastnasite-(La) and hydroxylcarbonate-(La) from Nissi bauxite deposit

	1	2	3
La_2O_3	28.4	31.6	35.3
Ce_2O_3	–	–	–
Pr_2O_3	6.1	7.0	7.8
Nd_2O_3	27.9	29.0	28.0
Sm_2O_3	3.9	3.9	2.8
Eu_2O_3	1.1	0.9	0.7
Gd_2O_3	2.8	2.0	1.2
Tb_2O_3	–	–	–
Dy_2O_3	1.3	0.2	–
Er_2O_3	0.7	–	–
Yb_2O_3	–	–	–
Y_2O_3	2.0	0.3	0.1
CaO	1.6	0.1	–
F	4.0	0.1	–
H_2O*	1.3	4.0	4.0
CO_2*	20.5	20.94	20.1
	101.7	100.04	100.0
–O=F	1.7	0.04	–
Total	100.0	100.0	100.0
Atomic ratios calculated to (O, OH, F) = 4			
La	0.38	0.42	0.48
Pr	0.08	0.09	0.10
Nd	0.36	0.37	0.36
Sm	0.05	0.05	0.04
Eu	0.01	0.01	0.01
Gd	0.03	0.02	0.01
Tb	–	–	–
Dy	0.01	<0.01	–
Er	0.01	<0.01	–
Yb	–	–	–
Y	0.04	<0.01	–
Ca	0.06	<0.01	–
F	0.46	0.01	–
OH	0.32	0.96	0.98
C	1.02	1.02	1.00

* The presence of carbon was detected by electron probe. H_2O is assumed to make up the remainder.
1. Hydroxylbastnäsite-(La); 2–3. Hydroxylcarbonate-(La)

above the footwall limestone, which acted as an effective alkaline barrier for descending solutions. The same distribution trend occurred for other "mobile" trace elements, including Ni and Mn.

Naturally occurring hydroxylbastnäsites usually contain an appreciable amount of fluorine. In the Nissi deposit, however, the lack of this element was evident, promoting the formation of fluorine-free hydroxylcarbonate-(Nd, La).

Both light REE (LREE) and heavy REE (HREE) show the same concentration trend in the Nissi deposit (Table 5). There is, however, a general decrease of the ratios LREE/HREE and La/Y in samples with depth, with a sudden increase of these ratios in the lowermost sample N1, and especially in the hydroxylcarbonate–(Nd, La). This extremely high enrichment of the LREE relative to HREE was found in all authigenic REE minerals encountered in karstic bauxites and karstic nickel deposits (Maksimović and Pantó 1991, 1996).

Figure 10 shows the chondrite-normalized REE distribution patterns in four samples from Nissi deposit, including the average N1 to N3 sample. The negative Ce anomaly is very pronounced in the three bottom samples (Av. N1–N3) due to the total absence of cerium in the authigenic hydroxylcarbonate-(Nd, La). This sample differs in absolute REE contents from three other ones. Negative Y anomaly is evident in all four samples. Three samples of the ferruginous bauxitic clays (N4, N6, N8) have similar distribution patterns except for slight differences in the absolute REE content.

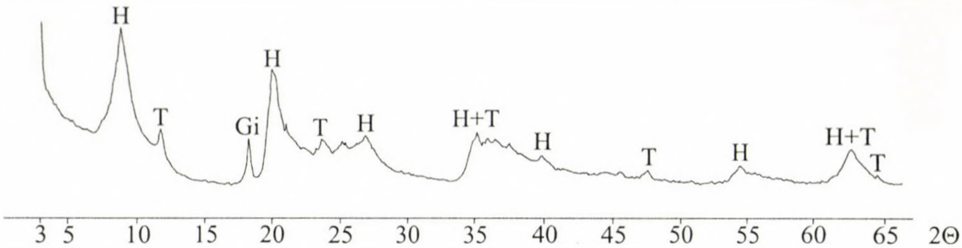

Fig. 6
XRD pattern of the bluish-green Ni-rich clay-like material. H = hydrated halloysite; T = takovite; Gi = gibbsite

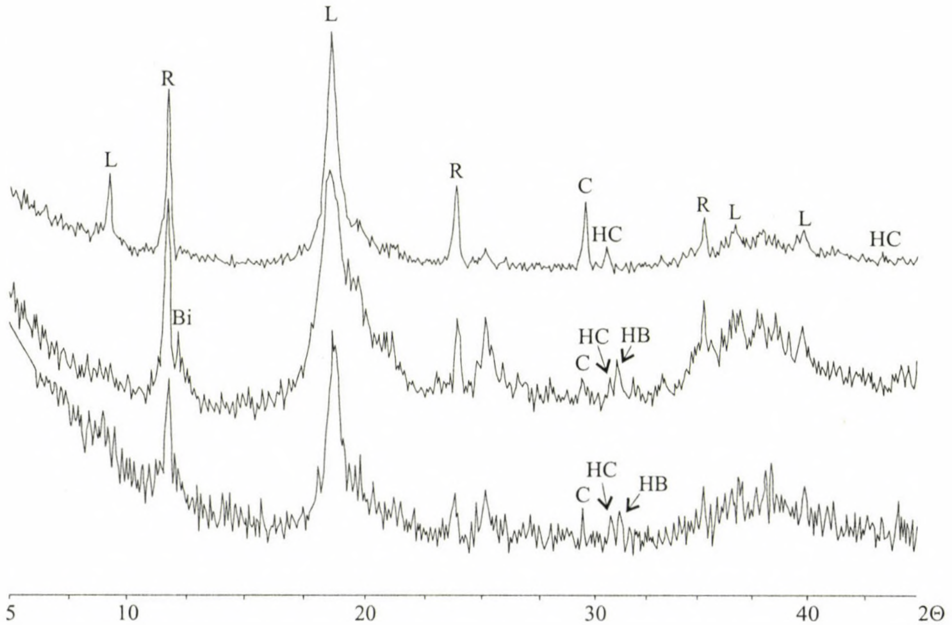

Fig. 7
XRD patterns of black concentrations with Mn and RE minerals. R = rancieite; L = lithiophorite; Bi = birnessite; C = calcite; HB = hydroxylbastnäsite-(Nd, La); HC = hydroxylcarbonate-(Nd, La)

Considering the behavior of the REE in the Nissi deposit it may be concluded that concentration per descensum of these elements at the base of the deposit is the same as in all karstic deposits formed in situ. When this deposit was exposed on the surface some alteration processes occurred, and the formation of Al-bearing minerals: takovite, hydrated halloysite and gibbsite above the footwall limestone, was probably the result of the activity of some acid solutions. It is not clear, however, whether some small scale re-distribution of the REE also took place in the framework of the primary distribution pattern, with the formation of hydroxylcarbonates-(Nd, La) in the epigenetic stage.

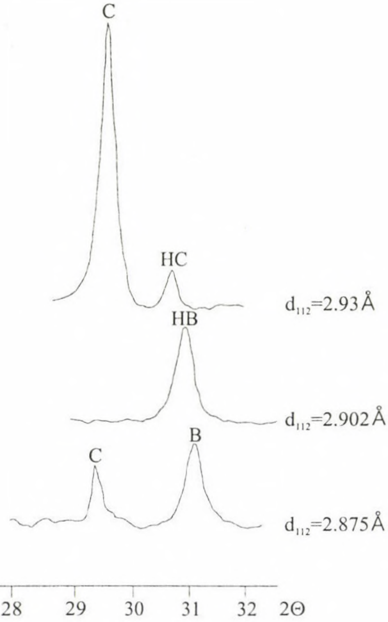

Fig. 8
A part of the XRD patterns of bastnäsite (B), from Marmeiko deposit, hydroxylbastnäsite-(Nd, La) (HB) and hydroxylcarbonate-(Nd, La) (HC) from Nissi deposit, with corresponding d112 values; C = calcite

Conclusions

The behavior of the REE during the formation of the Nissi deposit was the same as in all karstic deposits formed in situ. As a result of high concentration of the REE, especially of Nd and La, at the bottom of the deposit, the authigenic RE minerals were formed, including hydroxylcarbonate-(Nd) and hydroxylcarbonate-(La). According to the chemical composition and the XRD data, these are new minerals, the endmembers of the series (La, Nd) CO_3 F- (La, Nd) CO_3 OH, belonging to the bastnäsite group.

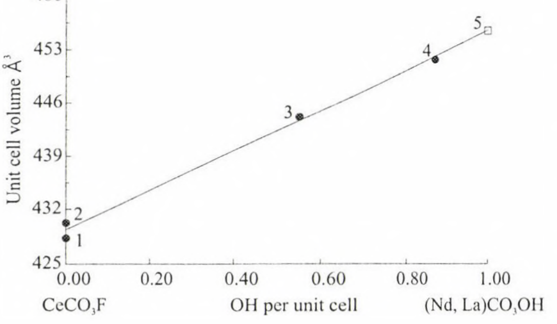

Fig. 9
Increase of the unit cell volume of bastnäsite by substitution of OH for F. 1. bastnäsite-(Ce), Ni et al. (1993); 2. bastnäsite-(Ce), Glass et al. (1958); 3. hydroxylbastnäsite-(Nd), Maksimović and Pantó (1985); 4. hydroxyl-bastnäsite-(Ce), Kirillov (1964); 5. hydroxylcarbonate-(Nd), Nissi deposit, Greece, Maksimović and Pantó (1996)

Table 4
Cell parameters, volume and the strongest reflections of bastäsite, hydroxylbastnäsite and hydroxylcarbonate-(Nd, La) (Å)

Mineral	Unit cell dimensions Å	Volume ($Å^3$)	Strongest reflection (I=100) d_{112} (Å)
Bastnäsite-(Ce)			
(Ni et al. 1993)	$a_0 = 7.1175$	428.3	–
	$c_0 = 9.7619$		
Bastnäsite-(Ce)	$a_0 = 7.129$	430.2	2.879
(Glass et al.1958)	$c_0 = 9.774$		
Hydroxylbastnäsite-(Nd)	$a_0 = 7.191$	444.3	2.911
(Maksimović and Pantó 1985)	$c_0 = 9.921$		
Hydroxylbastnäsite-(Ce)	$a_0 = 7.23(2)$	451.8	2.92
(Kirillov 1964)	$c_0 = 9.98(5)$		
Hydroxylcarbonate-(Nd, La)		455.5[a]	2.93
Hydroxylcarbonate-(La, Nd)			
Nissi deposit, Greece			

a Estimated in Fig. 1.

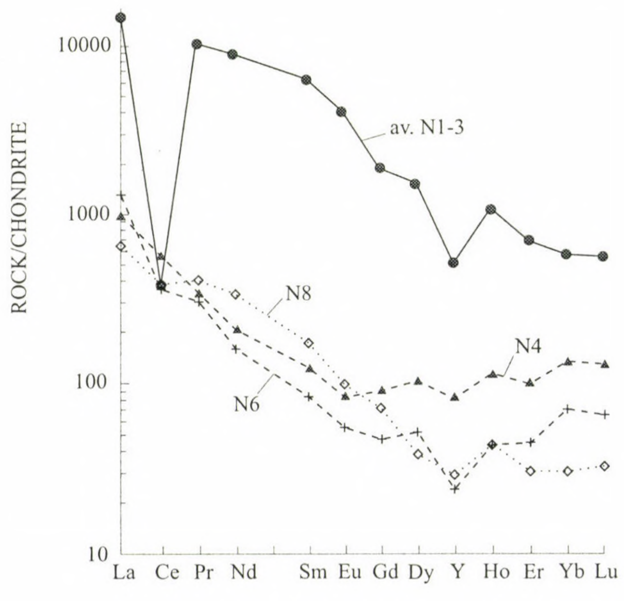

Fig. 10
Chondrite-normalized REE patterns of samples from the lower part of the Nissi deposit

Table 5
REE distribution along a vertical profile in the lower part of the Nissi deposit, including hydroxylcarbonate-(Nd, La)

Samples	ΣREE[a] (ppm)	ΣLREE[b] (ppm)	ΣHREE[c] (ppm)	ΣLREE / ΣHREE	La/ Y
N8 Ferrugin. bx clay	901	800	101	7.9	3.6
N6 Ferrugin. bx clay	943	842	101	8.3	8.9
N4 Ferrugin. bx clay	1240	970	270	3.6	2.0
N3 Ni–Fe clays	14565	12302	2263	5.4	4.3
N2 Ni–Fe clays	16832	13387	3445	3.9	3.2
N1 Ni–Fe clays	14259	13180	1079	12.2	9.7
Hydroxylcarbonate-(Nd, La)	636500	605000	31500	19.2	236

a ΣREE = La - Lu, Y, bΣLREE = La - Eu, cΣHREE = Gd - Lu, Y,

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Middle Triassic dasycladales in Sicily: Evidence of an Anisian?–Ladinian carbonate platform

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The Upper Triassic (Carnian) Mufara Formation outcrops in several localities in Western Sicily. It consists of deep-water marl-limestone alternations with intercalations of calcareous turbidites and debrites. Shallow water carbonates of both reef and lagoonal facies are abundant constituents of the clastic intercalations. In channel-filling carbonate debrites outcropping at Cozzo Paparina, near Altofonte, SW of Palermo, the elements exhibit different facies types as well as different age. The paleontology and facies of carbonates belonging to the reef facies of Carnian age, were described by Senowbari-Daryan and Abate (1986).

The lagoonal facies is represented by two distinct facies types: "Cayeuxia" bindstone and dasycladacean grainstone. *Diplopora annulatissima* Pia is the most abundant species in the dasycladacean association of this grainstone facies, followed by *Teutloporella peniculiformis* Ott and a further dasycladacean species tentatively assigned to *Physoporella lotharingica* (Benecke). The problematic organism (possibly a dasycladacean alga) *Zornia obscura* (= "Problematikum 1" Zorn 1971), is newly described. The stratigraphic age of the dasycladacean-bearing boulders is discussed and an Anisian?–Ladinian age is suggested. The source area of dasycladacean grainstone was an unidentified Middle Triassic carbonate platform. This represents the first evidence for an Anisian?–Ladinian carbonate platform in Sicily.

Key words: Dasycladaceans, Systematic Paleontology, Problematica, *Zornia obscura* n. gen., n. sp., Anisian–Ladinian, Mufara Formation, Sicily, Italy

Introduction

Carnian deep-water deposits known as the Mufara Formation outcrop in several localities in Western Sicily. They consist of brown marl and calcilutite containing halobians, radiolarians, conodonts, ammonoids, ostracods and foraminifers (Schmidt Di Friedberg 1965; Lentini 1974; Montanari 1989; Di Stefano 1990; Martini et al. 1991). Calcareous turbidites are commonly interbedded in these deposits, as well as thin levels of basalt. Owing to Neogene compressive tectonics the true thickness of this formation is very difficult to estimate but probably does not exceed a few hundred meters.

Channel-filling carbonates consisting of coarse-grained conglomerate and breccia are frequently interbedded in these deposits, as can be observed in the

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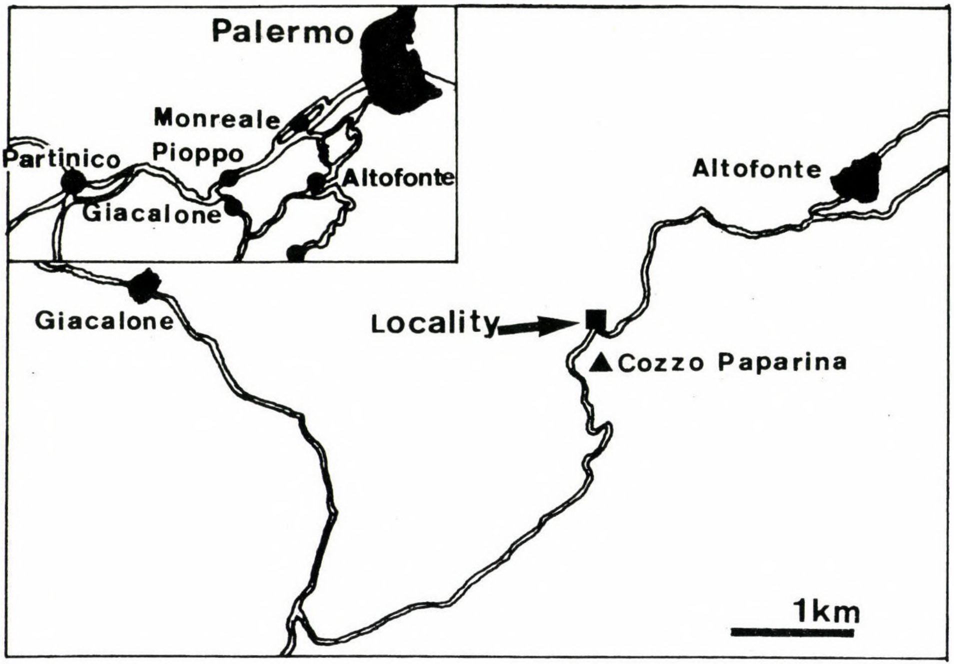

Fig. 1
Geographic position of the Cozzo Paparina locality where the Ladinian boulders were found

Palermo and Madonie Mountains. These carbonates are represented by platform/slope-derived elements of Anisian?–Ladinian and Carnian age (see Senowbari-Daryan and Abate 1986; Di Stefano 1990).

Among the platform-derived elements we can differentiate three main facies:

1) reef facies
2) lagoonal facies
    a. dasycladacean grainstone facies
    b. "Cayeuxia" boundstone facies
3) pisolite facies.

The reef facies is characterized by boundstone to bindstone dominated by sponges and "*Tubiphytes*" ssp. The paleontological contents of the reef facies has been described by Senowbari-Daryan and Abate (1986) and Senowbari-Daryan and Abate (in preparation). According to the association of reef-building organisms, foraminifers (especially the spiriamphorellid and pseudocucurbitid types) and microproblematic organisms the boulders of this reef facies are Carnian in age (Martini et al. 1991).

The lagoonal facies is characterized by "Cayeuxia" bindstone and dasycladacean grainstone. The dasycladacean grainstone contains the abundant species *Diplopora annulatissima* Pia, followed by *Teutloporella peniculiformis* Ott and ?*Physoporella lotharingica* (Benecke). This dasycladacean association is typical of lagoonal facies of Late Anisian to Early Ladinian and probably also Late Ladinian age. The age of the "Cayeuxia" boundstone as well as of the pisolithic facies is not clear. The source area of these reworked carbonates is not known; however, they indicate the existence of carbonate platform in Sicily during Middle Triassic and Carnian times. According to Di Stefano (1990) and Di Stefano et al. (1996) the reworked carbonates in the Mufara Formation were derived from a platform-upper slope area, adjacent to the Mufara basin, in which Middle Triassic to Carnian carbonates were exposed. This sector acted as the sedimentary substratum of the well-known upper Triassic-Eocene platform limestone pertaining to the Panormide domain. Due to the compressive tectonics related to the formation of the Apenninic–Maghrebian chain this (unknown) substratum was detached from the overlying carbonate succession during the Neogene.

We describe here the dasycladacean flora of the boulders found in Cozzo Paparina outcrops near Palermo (Fig. 1). The investigated material is kept in the Geologic Museum in the Department of Geology of the University of Palermo.

Paleontology

Class Chlorophyt Papenfuss, 1946

Order Dasycladales Kützing, 1843

Family Diploporaceae Deloffre, 1988

Tribe Diploporeae Pia, 1920

Genus *Diplopora* Schafhäutl, 1863

Synonym: Favoporella Sokac 1968 (non Favoporella Wu 1991).

Diplopora annulatissima Pia 1920

(pl. I, figs 1–6)

1920	*Diplopora annulatissima* n. sp.	– Pia, p. 67–68, pl. 4, figs 11–16
1931	*Diplopora annulatissima* Pia	– Pia, p. 273, pl. 21, figs 1–2
1964	*Diplopora annulatissima* Pia	– Bystricky, p. 137, pl. 24, figs 4–5, pl. 25, figs 1–4 pl . 26, figs 1–4, pl. 27, figs 1–4, pl. 28, figs 1–4
1965	*Diplopora annulatissima* Pia	– Herak, p. 23, pl. 5, figs 3–4, pl. 7, figs 1–2, pl. 12, fig. 7
1966	*Diplopora annulatissima* Pia	– Bystricky, p. 254, pl. 8, fig. 4
1966	*Diplopora annulatissima* Pia	– Ott, p. 5, fig. 4
1968	*Favoporella annulata* n. g. n. sp	– Sokac, p. 208–211, pl. 1–4
1979	*Diplopora annulatissima* Pia	– Emberger, p. 22 (Synonymy-List!)
1982	*Diplopora annulatissima* Pia	– Bystricky, pl. 2, fig. 1, pl. 4, fig. 3
?1982	*Diplopora* cf. *D. annulatissima* Pia	– Flügel and Mörtl, p. 99, figs 3–5
1983	*Diplopora annulatissima* Pia	– Fois and Jadoul, pl. 1, fig. b
non1986	*Diplopora annulatissima* Pia	– Sudar, pl. 25, fig. 4
non 1986	*Diplopora annulatissima* Pia	– Flügel, pl. 4, fig. 8
1986	*Diplopora annulatissima* Pia	– Bystricky, p. 304, pl. 3, figs 8–9
1986	*Diplopora annulatissima* Pia	– Senowbari-Daryan and Abate, pl. 5, fig. 5, pl. 12, fig. 1
1993	*Diplopora annulatissima* Pia	– Senowbari-Daryan et al., pl. 55, figs 1, 6, 8

Description

The cylindrical thallus of this alga shows a distinct annulation with low "segments". The annulation runs deeply into the internal cavity which is surrounded by a very thin wall (Plate I, figs. 3–6). Each "segment" contains only one whorl. The diameter of the pores increases a little closer to the periphery of the thallus. The majority of the specimens are recrystallized, only in some specimens the pores and the structure of the calcareous skeleton can be recognized. Table 1 shows some dimensions of *Diplopora annulatissima* Pia.

Stratigraphic range: Middle? Upper Anisian (Pelsonian?, Illyrian) – Upper Ladinian (Langobardian).

Table 1

Dimension of the skeleton of *Diplopora annulatissima* Pia from the Anisian?–Ladinian boulders within the Mufara Foramation

Thin section	D	d	d/D in %
S/6/120	4.0625	1.8750	46.15
"	4.7500	2.1875	46.05
"	3.1250	1.2500	40.00
"	3.7500	1.8570	49.52
"	4.3750	2.4375	55.71
"	5.4375	2.4375	44.82
"	3.3500	1.2500	33.33
S/6/a/a	4.3750	2.3750	54.28
"	4.0625	1.8750	46.15
"	5.3125	2.8125	52.94
"	5.1250	3.1250	60.97
"	3.4375	1.5625	45.45
"	5.1852	2.1875	42.18
S/6/a/b	3.3500	1.6250	43.33
"	3.4375	1.5625	45.45
"	4.1875	2.8125	67.16
"	4.6875	2.8750	61.33

D = outer diameter, d = inner diameter (all dimensions in mm)

Tribe Teutroporelleae Bossoullet et al., 1979

Genus *Teutloporella* Pia, 1912

Teutloporella peniculiformis Ott, 1995 (in Granier and Deloffre, 1994) (non Ott, 1963)

(pl. II, figs 1–6)

1963	*Teutloporella peniculiformis* n. sp.	–	Ott, figs 20–24
1964	"Bryozoa"	–	Bystricky, pl. 32, figs 2–4
1966	*Teutloporella peniculiformis* Ott	–	Ott, figs 3–4
1971	*Teutloporella peniculiformis* Ott	–	Zorn, pl. 4, fig. 2
1974	"Bioclastic dasycladaceous limestone"	–	Mello, pl. 1, fig. 3 (*Physoporella* and *Teutloporella peniculiformis*)
1978	"Fragment of a ?bryozoan zoarium".	–	Mello, Taf. 28, fig. 4
1976	*Teutloporella peniculiformis* Ott	–	Tollmann, p. 84, figs 29–30
1987	"Dasycladacea biosparite"	–	Oravecz-Scheffer, pl. 25, fig. 3
1993	*Teutloporella peniculiformis* Ott	–	Senowbari-Daryan et al., pl. 56, fig. 13
1994	*Teutloporella peniculiformis* Ott	–	Piros et al., p. 351, pl. 1, fig. 6
1994	*Teutloporella peniculiformis* Ott	–	Bucur, et al., p. 90, pl. 10, fig. 8, pl. 11, fig. 1 (Synonymy list!)
1994	*Teutloporella peniculiformis* Ott	–	Granier and Deloffre (no illustration)
1997	*Teutloporella peniculiformis* Ott	–	Rüffer and Zamparelli, pl. 26, figs 3–4

Description

Having a "fox tail" appearance of the calcareous skeleton, this alga is characterized by a small diameter of the central stem (cavity). Some specimens, however, exhibit a wider internal cavity on their upper part because of the V-shaped uncalcified segments (Plate II, fig. 3). The relatively thick wall of the skeleton is pierced by numerous circular and weakly recognizable pores (Plate II, fig. 1). The skeleton is composed of funnel-shaped segments giving it a cone-in-cone structure.

In the original description by Ott (1963) the species was incorrectly typified (see Bucur et al. 1994). Recently a new type species (fig. 3 in Ott, 1966 = fig. 23 in Ott, 1963) has been designated by Ott (in Granier and Deloffre, 1994).

Stratigraphic range: Middle Anisian (Pelsonian) – Lower Ladinian (Fassanian).

Tirbe Salpingoporelleae Bossoullet et al., 1979
Genus *Physoporella* Steinmann, 1903
?Physoporella lotharingica (Benecke, 1892)
(pl. III, figs 1–5)

1920 *Physoporella pauciporata* var.? *lotharingica* (Benecke) – Pia, p. 52, pl. 3, figs 11, 13
1931 *Physoporella lotharingica* (Benecke) – Pia, p. 266, pl. 21, fig. 8
1965 *Physoporella lotharingica* (Benecke) – Herak, p. 21, pl. 14, figs 4–6

Description

The cylindrical skeleton of this species is characterized by a thin wall and a wide inner space. The inner boundary of the wall is smooth, the outer surface is uneven (Plate IV, figs 3, 5). Almost all specimens of this alga are recrystallized and show a micritic rim. The interior of the skeleton is filled with sediment or sparry calcite cement. Table 2 gives the skeletal dimensions of *?Physoporella lotharingica*.

?Physoporella lotharingica represents a rare dasycladacean alga in this limestone. Thin section S/6/130, which bears a few specimens of *?Ph. lotharingica* lacks other species of dasycladacean algae. It is rarely associated with *Diplopora annulatissima* and *Teutloporella peniculiformis*. Some porostromate

Table 2
Dimensions of the skeleton of *?Physoporella lotharingica* (Benecke)

Thin section	D	d	d/D in %
S/6/130	1.8750	1.5625	83.33
"	2.0000	1.5625	63.12
"	1.8750	1.2750	68.00
"	1.3750	1.1250	81.81
"	1.1250	0.8125	72.23
"	2.5000	2.1250	85.00
S/6/120	2.5000	2.3125	92.50
S/6/?	2.6250	2.2500	97.29
"	2.5000	1.7500	70.00

D) Outer diameter, d) inner diameter (all dimensions in mm) algae, *Zornia obscura* n. gen., n. sp., gastropods, foraminifers (*Diplotremina* sp., *Trocholina* sp., *Endteba/ Endotebanella*; see Vachard et al. 1994), and ostracods occur in association with *Ph. lotharingica*.

Family incertae sedis

Zornia n. gen.

Zornia obscura n. sp.

(pl. IV, fig. 1–5, text-fig. 2)

1971 "Problematikum 1". – Zorn, pl. 17, fig. 7.
1972 "Problematikum 1". – Zorn, pl. 3, fig. 1

Derivatio nominis: In honor of Dr. H. Zorn, who first described this microfossil from the Ladinian carbonates as "Problematikum 1".

Holotype: We designate as holotype the specimen illustrated in Plate IV, fig. 4.

Paratypes: All specimens illustrated in Plate IV, figs. 1–3, 5.

Locus typicus: Cozzo Paparina, SW Palermo, at the road from Giacalone to Altofonte (see Fig. 1).

Stratum typicum: Ladinian (Anisian?) boulders within the Carnian Mufara Formation.

Diagnosis: Calcareous aggregates of horseshoe or crescent-shaped, 3/4-circular, circular or irregular skeletons. Relatively thick skeletal wall with an inner and outer micritic rim (diagenesis?). The wall may be pierced by pores.

Type species: Zornia obscura n. gen., n. sp.

Description

The skeleton of this microfossil consists of horseshoe or crescent-shaped, 3/4-circular to circular or sometimes irregular walls surrounding a wide cavity of approximately 1 mm in diameter. The ends of the crescent-like walls are turned into the internal cavity which is usually infilled with sparry calcite cement and rarely with micritic sediment (see also Zorn 1971, fig. 7). The common feature of all specimens are the 3/4-circular elements which are arranged in curved or circular lines. The space between these elements appear to be pierced by pores (Plate IV, figs 3–5: arrows). Apparently these elements were together during the life of the organism and fell apart after death.

The skeletal walls appear white in transmitted light and show an inner and outer micritic rim. The walls are 0.1–0.22 mm thick and are composed of calcite (primary aragonite?).

Zornia occurs in clusters, usually associated with cavities filled with sparry calcite cement as shown in text Fig. 2. Some dasycladacean algae described in this paper may occur in association with *Zornia obscura*. In some samples the organism is embedded in a grapestone facies associated only with some agglutinated foraminifers and no other fossils.

Discussion

Zornia obscura was first described by Zorn (1971, 1972) as "Problematikum 1" from the lagoonal grapestone facies of the Ladinian Salvatore dolomite of St. Salvatore in Ticino (Switzerland). In Sicily *Zornia obscura* occurs in reworked calcareous boulders representing a lagoonal grapestone or grainstone facies. In both localities the organism may be associated with dasycladacean algae. Zorn (1972) mentions the occurrence of oncolites, stromatolites, codiacean algae (*Orthonella*) and dasycladacean algae (*Diplopora annulata* Schafhäutl and *Physoporella lotharingica* (Benecke). In both localities the small cavities between the *Zornia* fragments may reach more than 50% of the total rock volume.

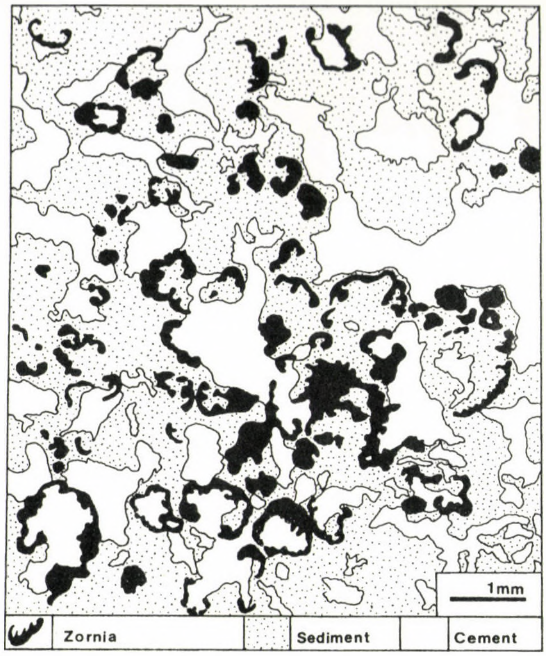

Fig. 2
Zornia obscura n. gen., n. sp. and the associated sediments or cavities

The different shape of the organism (fragments?) make interpretation of the overall form problematic. Two reasons suggest a plant affinity for this organism: in the first place the calcareous skeleton is pierced, possibly by pores (Plate IV, figs 1, 3–5: arrows). This could indicate a dasyladacean alga affinity with an umbrella-like construction whose upper part is calcified. Secondly, *Zornia obscura* is found associated with other dasycladacean green algae.

Stratigraphic range of algae-bearing boulders

The dasycladacean green algae (*Diplopora annulatissima, Teutloporella peniculiformis, ?Physoporella lotharingica*) and the enigmatic fossil, *Zornia obscura* n. gen., n. sp., described here, occur in reworked calcareous boulders of a lagoonal facies (dasycladacean grainstone or algal grapestone) exposed in Cozzo Paparina near Palermo. In addition to algae, some foraminifera and ostracods are present.

The dominant dasycladacean species is represented by *Diplopora annulatissima* Pia, followed by *Teutloporella peniculiformis* Ott, and *?Physoporella lotharingica* (Benecke). *Diplopora annulatissima* is restricted to the time-interval Upper Anisian (Illyrian) to Upper Ladinian (Longobardian) (Herak 1965, Ott 1972a, Bystricky 1986). According to Flügel and Mörtl (1982) *D. annulatissima* may also occur in Middle Anisian (Pelsonian). In some regions (e.g. Romania, Hungary), however, *D. annulatissima* seems to be restricted only to the Anisian (Dragastan 1981; Piros et al. 1994).

Teutloporella peniculiformis Ott occurs also in Anisian and Lower Ladinian (Ott 1972b; Bystricki 1986; Bucur et al. 1994).

There are few reports concerning the occurrence and stratigraphic range of *Physoporella lotharingica*, but according to Herak (1965) this alga also characterized Anisian–Upper Ladinian. Zorn (1971, 1972) has described this species from the Ladinian of the St. Salvator dolomite (Ticino, Switzerland).

The association of this dasycladacean algae, which also occurs in the Ladinian, with *Zornia obscura*, an organism not known from the Anisian, suggests that these dasycladacean-bearing boulders from Cozzo Paparina are most probably Early Ladinian in age. An Anisian or Late Ladinian age of these boulders, although not excluded, appears unlikely.

Acknowledgements

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Plate I

1–6. *Diplopora annulatissima* Pia from the Ladinian calcareous boulders within the "Mufara Formation" (Carnian–Norian) of Cozzo Paparina near Palermo

1. Tangential to longitudinal section. S/6/?, ×10
2. Tangential section. S/6/6/1, ×10
3. Similar to fig. 1. S/6/?, ×8
4. Tangential to longitudinal section. S/6/6/1, ×6
5. Tangential section S/6/120, ×8
6. Tangential sections through two specimens. S/6/6/1, ×10

Plate II

1–6. *Teutloporella peniculiformis* Ott from the Ladinian calcareous boulders within the "Mufara Formation" (Carnian–Norian) of Cozzo Paparina near Palermo.

1 Tangential section. S/6/120, ×24
2 Tangential section. S/6/?, ×15
3 Longitudinal section with a wide axial canal. S/6/6/1, ×15
4 Tangential section. S6/6/1, ×12
5 Fragment. S/6/?, ×12
6 Tangential section. S/6/?, ×12

Plate III

1–5. ?*Physoporella lotheringica* (Benecke) from the Ladinian calcareous boulders within the "Mufara Formation" (Carnian–Norian) of Cozzo Paparina near Palermo.
1. Oblique to cross section through two specimens. The alga is characterized by very thin wall of the calcareous skeleton. S/6/124, ×30
2. Tangential section. S/6/6/1, ×30
3. Cross-section. S/6/120, ×30
4. Cross-section through two specimens. S/6/61, ×30
5. Longitudinal section. S/6/6/1, ×8

6–7. Foraminifera from the Ladinian calcareous boulders within the "Mufara Formation" (Carnian–Norian) of Cozzo Paparina near Palermo associated with the algae described in this paper.
6. Agglutinated test of the ex. gr. *Endotebanella* sp. S/6/124, ×30
7. Duostominidae and Endotebidae. S/6/130, ×20

Plate IV

1–5. *Zornia obscura* n. gen., n. sp. from the Ladinian calcareous boulders within the "Mufara Formation" (Carnian-Norian) of Cozzo Paparina near Palermo.
1. Sections through several specimens. The arrows indicate pores in the skeletal wall. S/6/144, ×30
2. Sections through three specimens. S/6/148, ×30
3. Section through numerous specimens and some cavities filled with cement. S/6/148, ×7.5
4. Holotype. Section through a circular specimen composed of several crescent-like elements. S/6/148, ×3.
5. Section through several specimens. S/6/148, ×18

Plate I

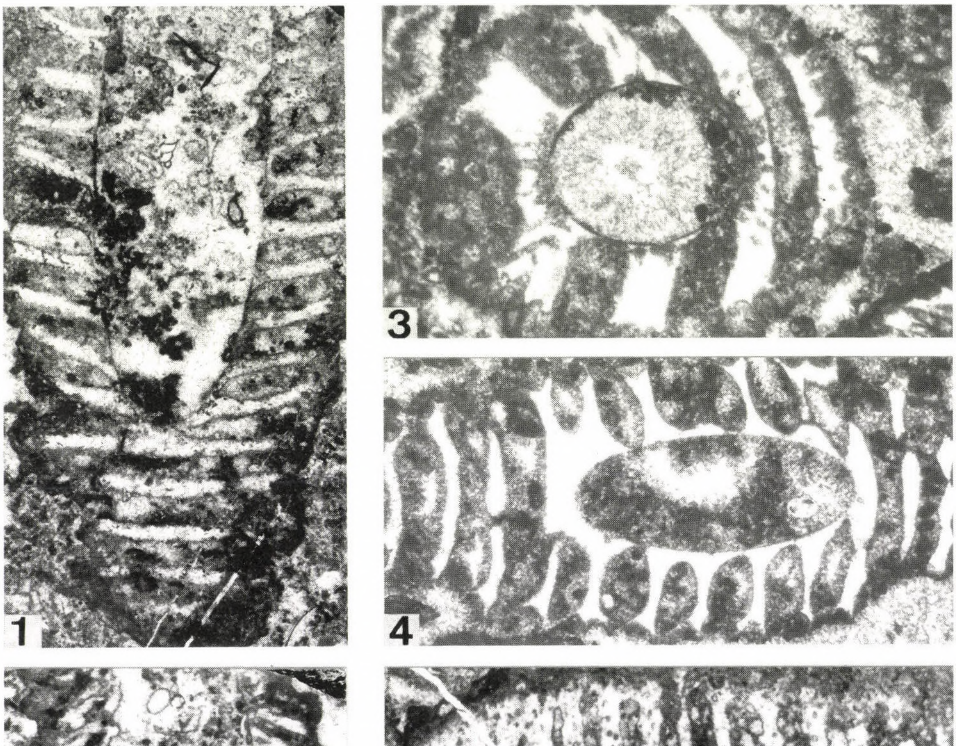


Plate II

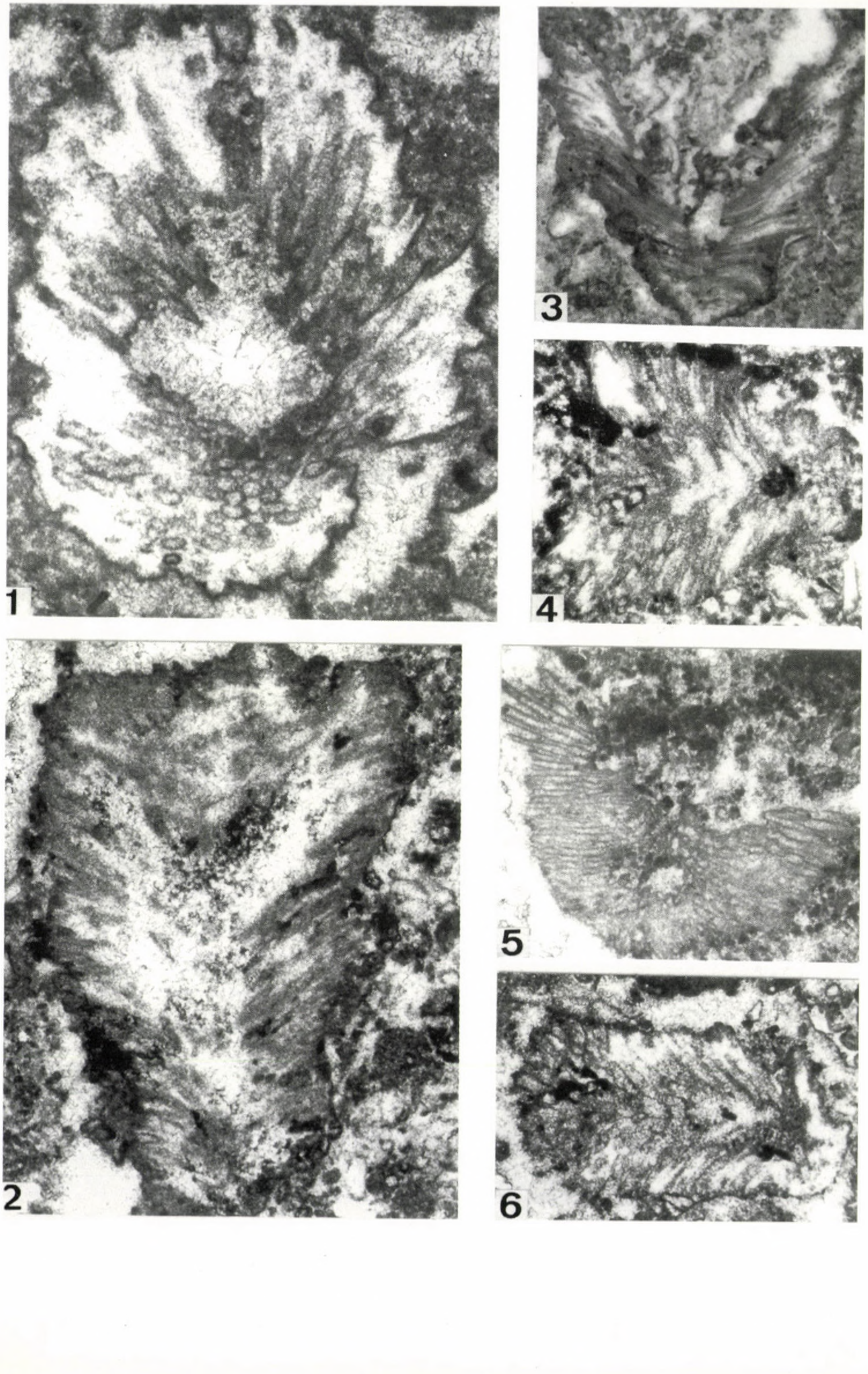


Plate III

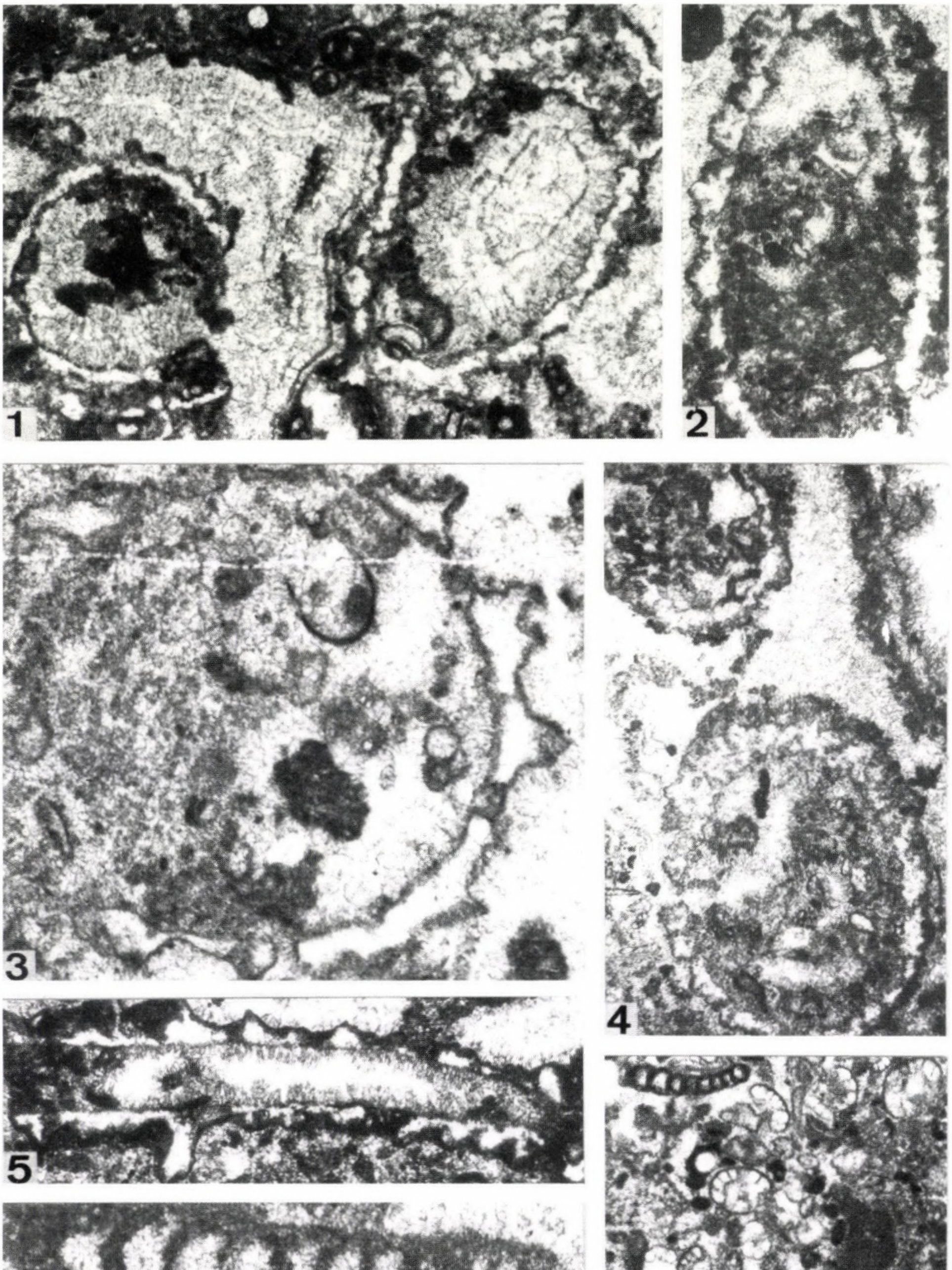
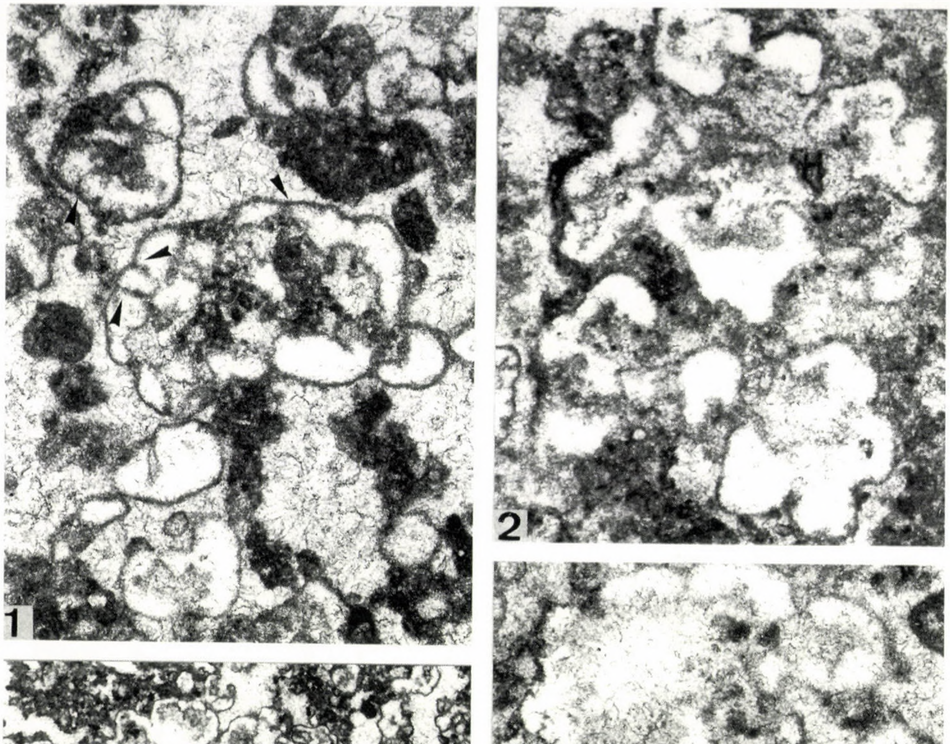


Plate IV



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Preface

to special issue devoted
PANCARDI 2001 Meeting
Sopron, Hungary

The present special issue of Acta Geologica Hungarica is devoted to the Hungarian results obtained within

- the framework of the PANCARDI (Pannonian Basin, Carpathian Arc, Dinaride System) key project, which focuses on the dynamics of ongoing orogeny;
- the EUROPROBE (Origin and Evolution of the Continents) lithosphere dynamics program, and which were
- supported by the ESF (European Science Foundation).

Of course, some of these results have already been published in detail in Hungarian and international journals (most recently, the European Geophysical Society (EGS) has also prepared some "Special Publications", including several papers on the results of the PANCARDI project). The edition of the present number of Acta Geologica Hungarica is connected to the PANCARDI closing conference to be organized this autumn in Sopron, W Hungary.

The organizers have requested the leaders of the Hungarian working groups to write together with co-authors (generally Hungarian experts but in a few cases foreign colleagues, as well) a review paper on the main results of their activities. The annotated themes correspond to those summarized in the EUROPROBE Program booklets for the "PANCARDI research".

The editors wish to thank the authors for their contributions, and also thank the chairman of the PANCARDI key project, Prof. C. Tomek, for overseeing of the project during the last ten years.

A. Ádám, J. Haas

Acta Geologica Hungarica, Vol. 44/2–3, pp. 113–134 (2001)

Metamorphic and rheologic effects shown by seismic data in the Carpathian Basin

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Antal Pápa
Hungarian Oil and Gas Company LTD (MOL RT), Budapest

Tamás Fancsik
Hungarian Eötvös Loránd Geophysical Institute, Budapest

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The authors have attempted to interpret the origin of subhorizontal reflections often observed in crustal seismic sections. They have compared the approximate locations of metamorphic facies and the probable strength of the rocks depending on depth to a few seismic sections of eastern Hungary. Estimation of depth of metamorphic mineral stability zones was carried out by the data of Fyfe et al. (1978) and geothermal data of the area. The strength profile was computed by the diagram of Meissner et al. (1991) and the seismic P-wave interval velocities determined in the region.

The authors conclude that the series of subhorizontal reflections are connected to metamorphism and strength envelopes, and may even be of importance in interpreting industrial seismic sections. The examples presented in their paper suggest that the retrograde metamorphism of the pre-Tertiary basement was made possible by the softening effect of shear zones and their water-conducting capacity.

It was found that the subhorizontal reflections of highest energy in the consolidated crust are to be found in the depths of greenschist, amphibolite and granulite metamorphic mineral facies originating in geothermal and pressure conditions similar to the present ones. Thus the overprint of earlier (Variscan) metamorphic structure by a retrograde metamorphism can be supposed.

Key words: subhorizontal reflections, metamorphism, overprint, rheology, strength, deep seismic survey, P-wave velocity

Introduction

The task of geophysical exploration has gone beyond the investigation of sedimentary structure and aims at determining the quality of rocks, their fluid content and physical properties. In this paper – through some examples – while looking for the conditions producing reflections in the basement and deeper in the crust, we try to determine the factors influencing the state of rocks.

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In addition, it often occurs in industrial seismic sections (also in Hungary) that series of subhorizontal reflections appear in the consolidated basement, sometimes even in the depth interval of exploration boreholes. These reflections must be interpreted in a way not usually carried out in industrial seismic exploration; therefore, our results may command interest among their interpreters as well. Some examples:

– in the structural study of the Mecsek Mountains (Ráner et al. 1979) this phenomenon appeared in the northern end of seismic section Gö-5 from the uppermost part of the crust to its bottom.

– in several localities of the Transdanubian Central Range high-amplitude subhorizontal reflections could be observed at relatively shallow depths (4–5 km) and in the same localities in similar depths magnetotelluric measurements showed a conducting layer (Pápa et al. 1990; Ádám 2001; Redler-Tátrai and Varga 2001).

– in deep seismic surveys high-energy subhorizontal reflections appeared at several depth intervals of the lithosphere and asthenosphere (Posgay et al. 1995). Similar reflections could be observed in the basement as well (Posgay et al. 1996).

In the last decades a large number of papers has been published in this field. Just to illustrate the development of the ideas we will try to review some of them.

Dohr and Meissner (1975) interpreted the lamellae of the lower crust as the result of either intrusions or crystallization seams or peeling of mantle material.

According to Klemperer (1987) in the consolidated crust below the sedimentary layers, generally very few reflections can be found. Below this transparent zone reflections appear where the temperature becomes higher than 300–400 °C. As the upper boundary of the zone of rock ductility coincides with that of increasing reflectivity, he considered it probable that ductility may be the factor promoting reflectivity. He had problems, however, in explaining the origin of layered patterns. Matthews (1986) supposed that beside temperature, other factors are also needed for the increase of reflectivity, e.g. difference in mineral composition or the presence of high-pressure fluids.

Christensen (1989) described subhorizontal reflections in Inner Piedmont (Southern Appalach Mts. of South Carolina), where a deep borehole penetrated rocks of middle and lower crustal origin (upper amphibolite facies). He studied the reflections using synthetic seismograms, computed using density and velocity data obtained from drill cores. He concluded that the silicic and mafic layers of thickness varying between 0.3 and 13.7 m, corresponding to biotite-bearing quartz-feldspathic gneiss, granitic gneiss and amphibolite, respectively, originated in the lower crust by metamorphism, most probably by ductile flow. From the layers of sharp physical contrasts the biotite quartz-feldspathic gneiss may have originated from metasediments, the granitic gneiss from low potassium-content granite and the amphibolite from basalt. The foliation is the result of preferred mineral orientation of quartz, amphibole and mica.

Holbrook et al. (1992) suggest the following causes of reflectivity in the lower crust:

1) layering of magmatic or metamorphic rocks,
2) ductile shear zones,
3) magmatic intrusions,
4) lenses of partial melting, and
5) fluid-containing fissures.

Referring to wide-ranging literature Mooney and Meissner (1992) reviewed the results of deep seismic reflection profiling in the continental lower crust and associated the lower-crust reflectivity with the physical properties and evolution of its rocks.

According to Pavlenkova (1996) the low-velocity and low-resistivity zone of high reflectivity described by Western European authors as part of the lower crust, in areas of inland thick continental crust, can be found in the middle crust. There the steeply dipping reflections of the upper crust are replaced by subhorizontal ones. Below 10 km the fluid-saturated fine fissuring causes low velocity and high conductivity as well as ductility, all contributing to intensive metamorphism.

Borradaile et al. (1999) studying the Kapuskasing Structural Zone (KSZ) found that the gneissic layering of a 30 km-thick Archean granoblastic gneiss block of granulite and upper amphibolite facies metamorphic rocks – overthrusting from the depths of the lower crust to the surface – cannot be recognized in seismic sections. On the other hand the observed anisotropy of low-field magnetic susceptibility (AMS) and anisotropy of anhysteritic remanence (AARM) showed subhorizontal schistosity and extension lineation in the mineral patterns, parallel to scattered but concordant reflections. In thin sections perpendicular to magnetic schistosity of the layering is conspicuous. According to the authors the schistosity appearing both in magnetic and seismic data – originating before the emergence of the KSZ block – has overprinted the gneissic layering.

Subhorizontal reflections of the pre-Tertiary basement

In Fig. 1 a portion of an oil exploration seismic depth section from eastern Hungary is presented. Its location can be found in Fig. 2, marked by A. In its central part a steeply dipping shear zone can be observed, which is disturbed in the depth range of 4.5–5.5 km by a series of subhorizontal reflections. In the seismic section of Fig. 3 (marked B in the location map of Fig. 2) one can also recognize subhorizontal reflections at similar depths. In the following, these subhorizontal reflections will be compared to the metamorphic and rheologic conditions presumed for the given depth range. A similar study will be carried out along the PGT-1 lithosphere and asthenosphere investigating deep seismic section.

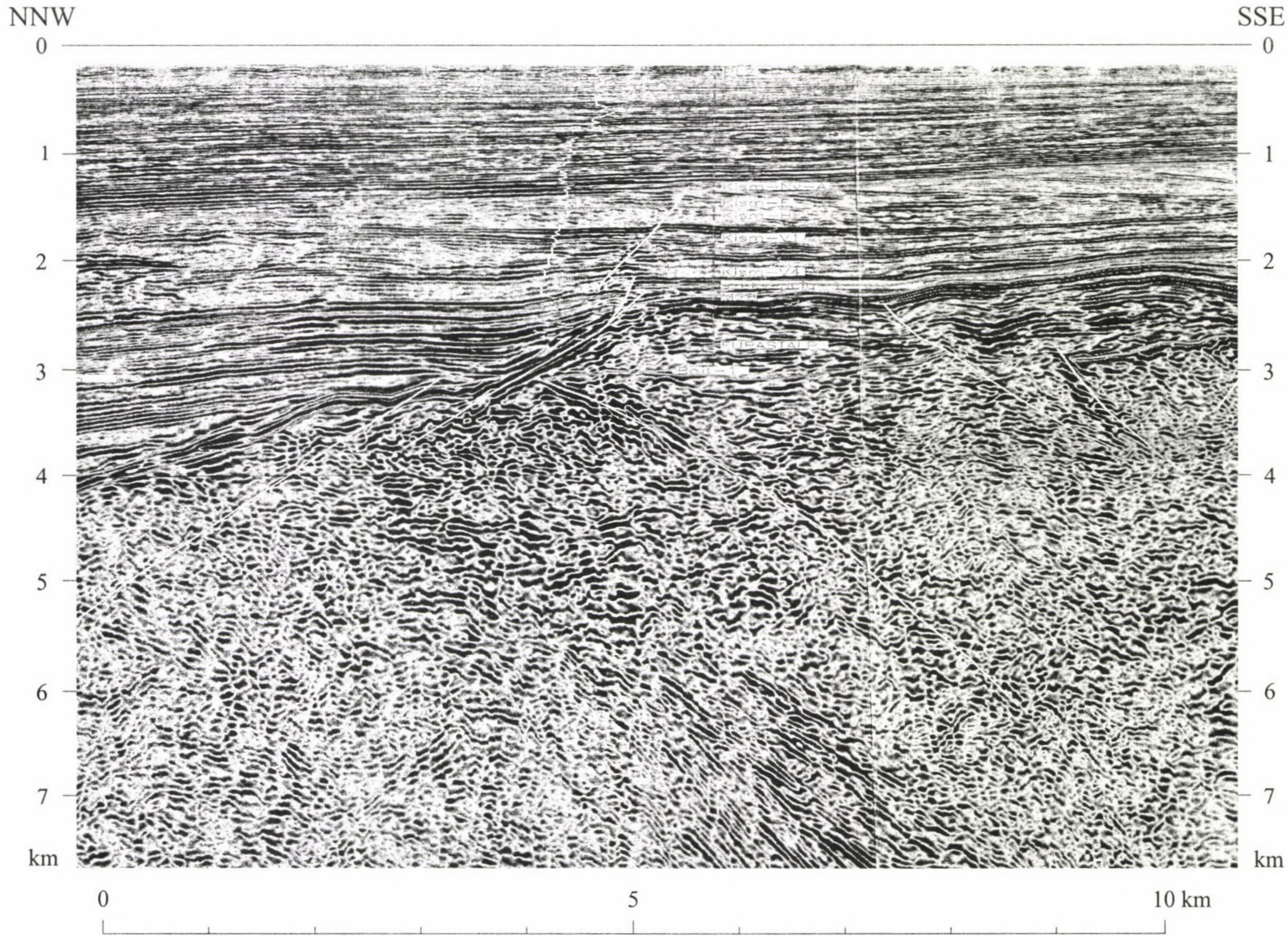


Fig. 1
Part of an oil-exploration seismic depth section from eastern Hungary. Its location is marked by A in Fig. 2. In the depth range of 4–5 km, subhorizontal reflections overprint the shear zone in the basement. We suppose that the water content of the shear zone made the retrograde metamorphism of the metamorphic complex possible

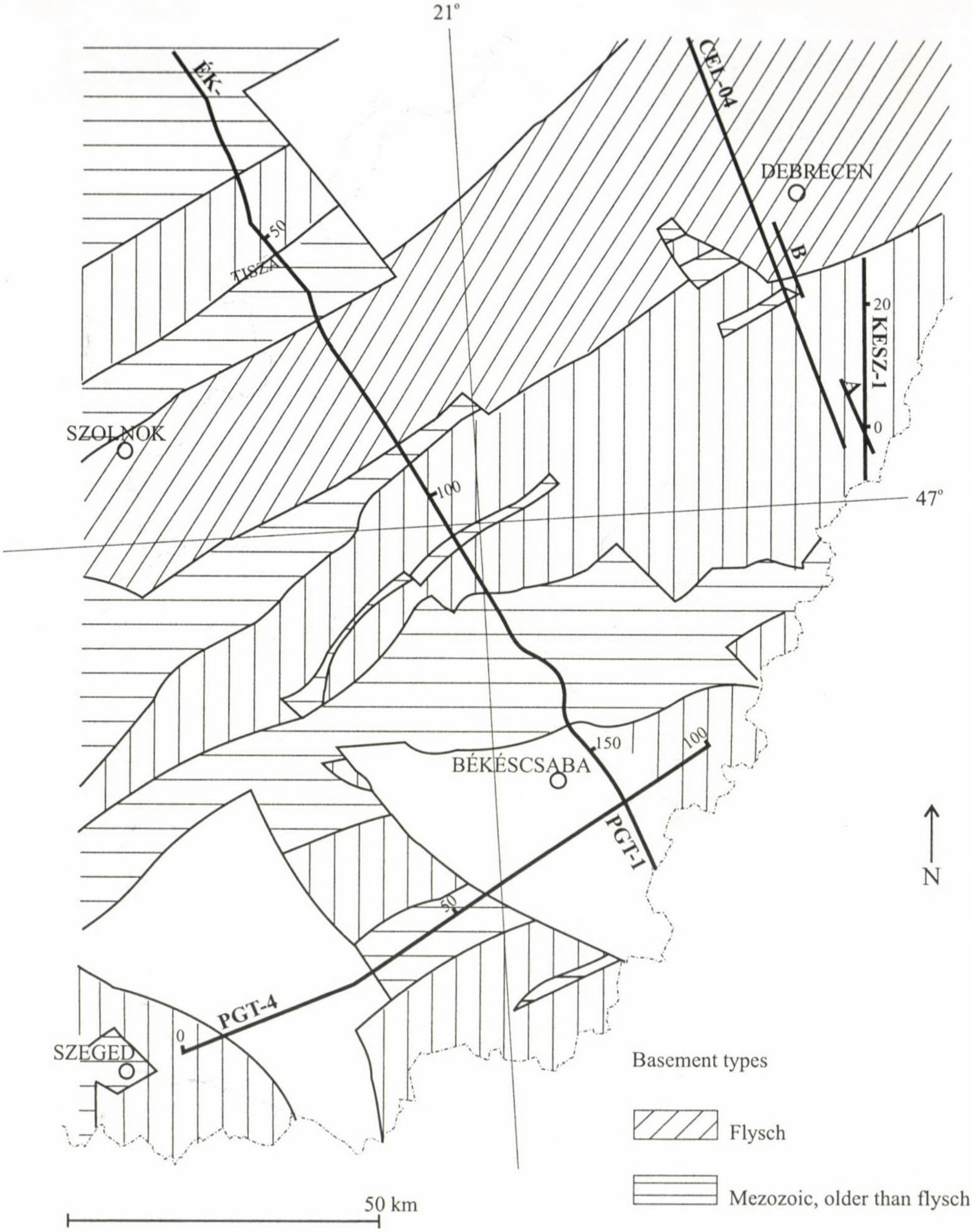


Fig. 2
Location map of seismic sections. The sketch of geology of the pre-Neogene basement is taken from the structural map of Hungary (Dank et al. 1990)

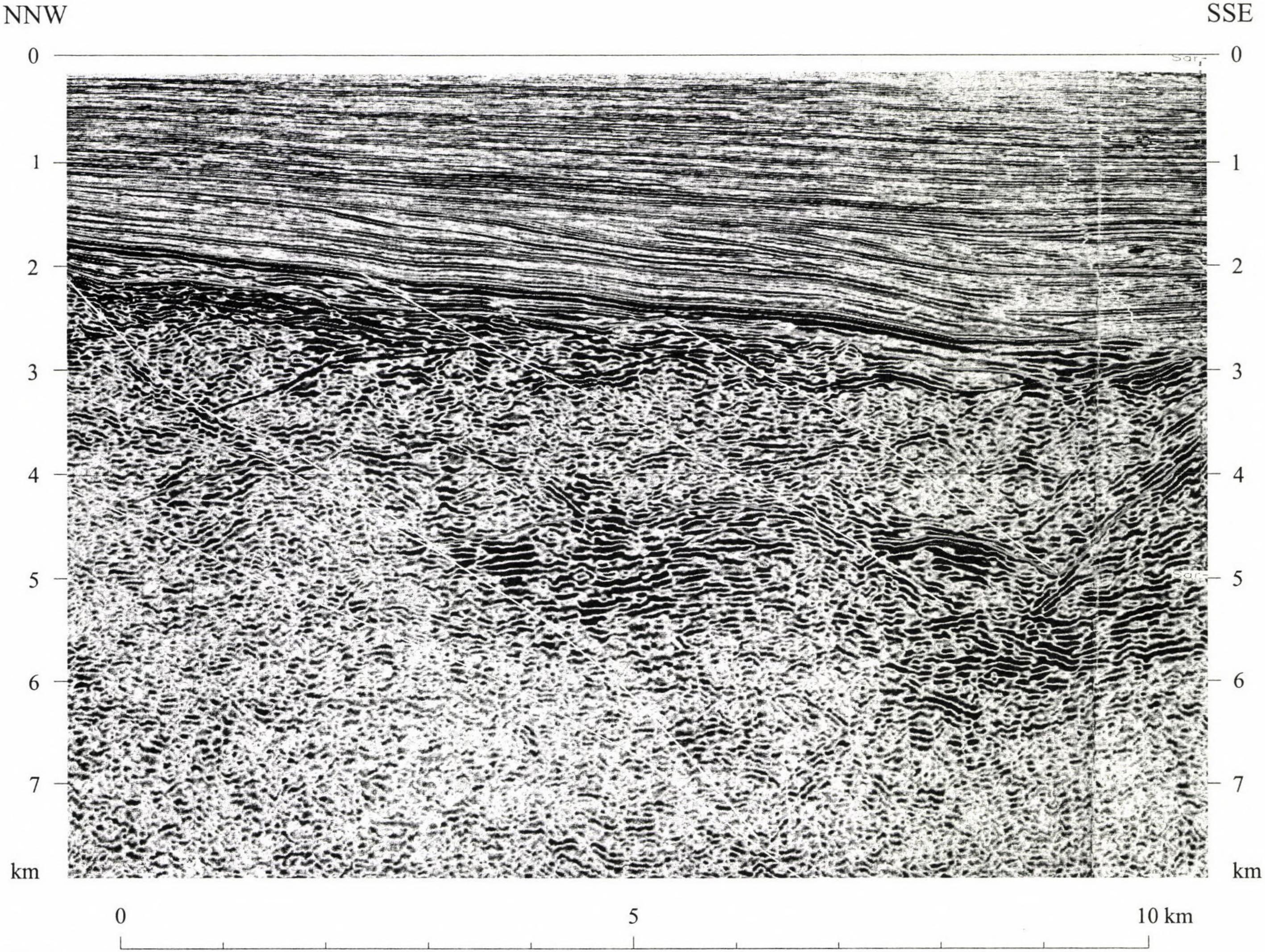


Fig. 3
Part of an oil exploration seismic depth section, marked by B in Fig. 2. Subhorizontal reflections can be observed in the same depth range as in Fig. 1

Figure 4 shows the approximate locations of mineral facies in terms of experimentally determined mineral stabilities (modified after Fyfe et al. 1978). The metamorphic facies are normally named after one of the characteristic assemblages found in metabasalt. We have plotted in the same figure the temperature versus depth curve valid for the area (Dövényi et al. 1983; Cermák and Bodri-Cvetkova 1987; Lenkey 1999) and the interval velocities determined in the KESZ-1 deep seismic section (Posgay et al. 1981). Using this chart, we can estimate the probable stability zones of mineral facies in the area within presently existing conditions. As the subhorizontal reflections of Figs 1 and 3 appear in the depth range of 4.5–5.5 km, according to Fig. 4 in our study area we are dealing with the zeolite metamorphic zone.

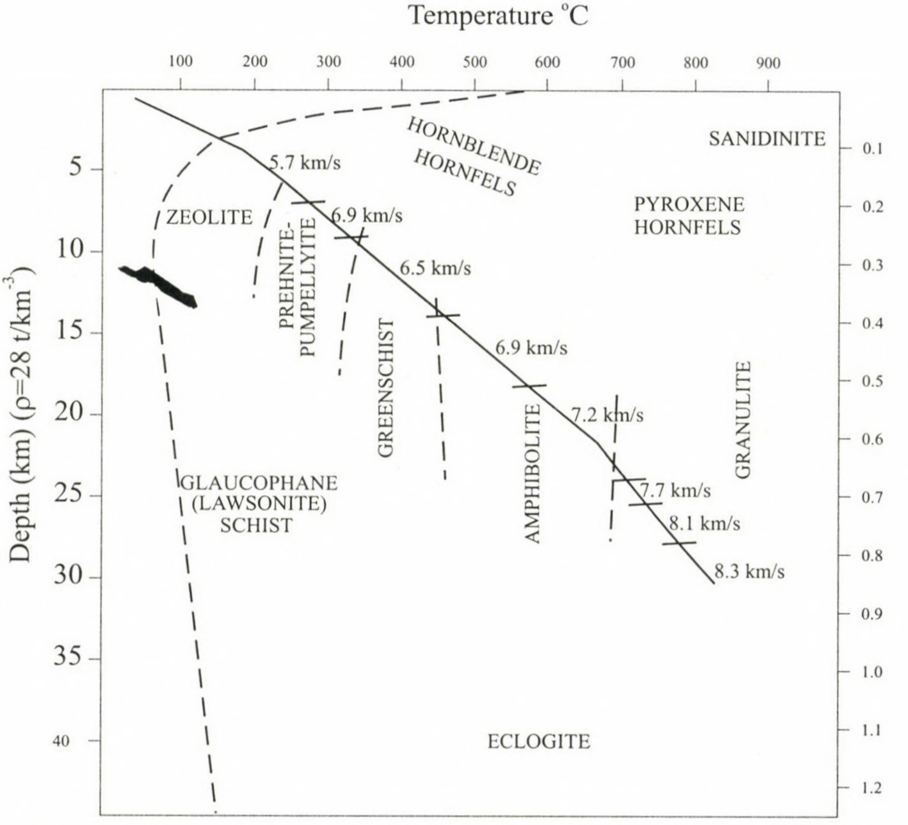


Fig. 4
Approximate locations of mineral facies in terms of experimentally-determined mineral stabilities (modified after Fyfe et al. 1978). Facies are normally named after one of the characteristic assemblages found in metabasalt. The temperature versus depth diagram (Dövényi et al. 1983; Cermák and B-Cvetkova 1987; Lenkey 1999) and the interval velocities characteristic of the area, are also plotted (Posgay et al. 1981)

It seems, however, expedient to study these results from the point of view of rheology as well, and to interpret the data according to both fields.

Rheology studies

Duration of stress originating in the Earth may vary between a few seconds to several hundred million years. Rheologic response of rocks on stresses of varying magnitude and duration can strongly differ, from elastic to ductile (Ranalli 1995). Unless the oscillations of atoms surpass the potential threshold of the cohesive force, the response of the rock will be elastic (generally in the case of the effect being shorter than 10 million years). In elastic materials deformation is proportional to stress. Magnitude of tectonic force depends on direction. If the difference (differential stress) between the maximum and minimum stresses – called principal stresses – surpasses the yield stress, the rock fails. Beyond the range of elastic behavior the material either fractures or plastic flow takes place.

If high temperature induces atomic vibration above the elastic threshold, the rocks become viscous and their deformation takes the form of creep. This threshold is to be found at 1/3–1/2 of the melting point. Lithostatic pressure decreases the mobility of atoms: creep takes place at a few hundred MPa pressure in the temperature range of 300–400 °C. In the case of ductile flow, in ideal conditions, constant stress induces increasing deformation. It can be supposed, however, that if the strain rate is higher than 5×10^{-15}/s strain softening occurs, while for low strain rate strain hardening take place (Kusznir and Park 1987). Scholz (1988) describes a brittle-plastic transition zone between the elastic and the ductile zones. In the case of rocks of quartz-feldspar composition this transition zone is determined by the initial temperature of their plastic behavior. According to his model earthquakes generally originate in the brittle crust (and partly in the upper mantle) but large earthquakes may originate even in the transition zone by localized shear. At lower stresses the response of rocks in the transition zone is plastic flow.

Rheology of the lithosphere can be characterized by a strength versus depth diagram. Strength, in this diagram, is described by the differential stress (or by its logarithm=$\log(\sigma_1-\sigma_3)$) at which the cohesion of the rock ceases to exist. Constructing this diagram, the authors extrapolate the laboratory steady-state data by several orders of magnitude (Chen and Molnar 1983; Strehlau and Meissner 1987).

In the elastic lithosphere, rheology is described by the value of frictional failure (Sibson 1974; Byerlee 1978) taking into account the pressure of the pore fluid (Ranalli and Murphy 1987). For the ductile part of the lithosphere, where deformation takes place in the form of steady-state creep, the stability computed by a power-law empirical function (Kirby 1985) is lower than the one computed by friction, therefore the power-law function is used. Through these studies it has been pointed out that in the case of heat flow $q\geq55$–70 mW/m^2, at the bottom of the crust and within the crust as well, detachment zones of low stress and stability can be found. If the heat flow is higher, these detachment zones are to be found at shallower depths.

For the conditions of Hungary calculations were carried out by Bodri (1994). In his model the upper crust consists of silicic rocks of low metamorphism, while below 10–15 km, they give place to rocks of greenschist-grade metamorphism. The middle crust consists of silicic and intermediate rocks of amphibolite facies metamorphism. In the lower crust intermediate and mafic rocks of granulite facies metamorphism, and in the upper mantle ultramafic rocks dominate. Creating his model, he used the stability values published for water-saturated rock samples for both the crust and mantle lithosphere. He presents depth maps for the brittle and ductile zones of the crust. The lower crust and the mantle is supposed to be ductile. He compares the distribution of earthquake focal depths to his strength profile (Bodri 1995) for several areas of Hungary. In most cases earthquake foci coincide with the elastic (brittle) zones of the upper and middle crust.

In the above-described models the crust is built up from one or two isotropic layers. Estimating their physical parameters, their seismic velocities were compared to those obtained by laboratory measurements using high pressure and temperature (e.g. Christensen 1979). In this method the possibility of intra-layer parameter variations were not taken into consideration. To reduce the errors arising from neglecting these variations we have used the empirical relationship between the longitudinal interval velocities and the activation energy of rocks determined by laboratory measurements (V_p–E diagram; Meissner 1989; Meissner et al. 1991). Determining the interval velocities, we have divided the lithosphere into more intervals than the authors of rheology models did. Thus the empirical relation of Meissner et al. (1991) made it possible to estimate the rock parameters from the actual velocity curve determined in the study area and not from some general laboratory and field data. The physical basis of Meissner et al's diagram is the assumption that increasing packing of rock texture increases both activation energy and shock-wave velocity. Up till now we have not found any reference in the literature that in creating a rheologic model this possibility was considered.

For the above-mentioned purpose we have used the interval velocities determined in the KESZ-1 deep seismic section (Posgay et al. 1981, 1986). In Fig. 5 both the interval velocity curve, V_{int} of KESZ-1 and the tomographic velocity curve, V_{tom} of CEL-04 (for location see Figs 2 and 6) are plotted. This latter section is part of the international CELEBRATION 2000 project (Guterch et al. 2000; CEL Org. Com. & Exp. Team 2001). As the two curves match well it seemed acceptable to use these interval velocities for the computation of strength in the ductile zones. The temperature versus depth curve characteristic for the area is also plotted.

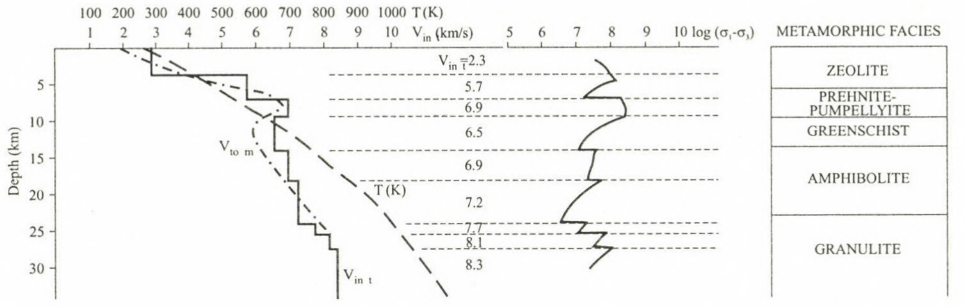


Fig. 5
On the left side the interval velocities (V_{int}) are plotted as determined in deep seismic section KESZ-1 (Posgay et al. 1981), together with the velocities (V_{tom}) determined by ray-tracing tomography (Guterch et al. 2000), as well as the temperature versus depth curve (Dövényi et al. 1983; Cermák and B-Cvetkova 1987; Lenkey 1999). On the right side the strength profile computed from the interval velocities and the respective metamorphic facies versus depth from Fig. 4 are plotted

The experiment called CELEBRATION 2000 (Central European Lithospheric Experiment Based on Refraction 2000) targeted the structure and evolution of the complex collage of major tectonic features in the Trans-European suture zone (TESZ) region, as well as the southwestern portion of the East European craton, the Carpathian Mountains, the Pannonian basin, and the Bohemian massif (Fig. 6).

For the upper crust, we have computed the frictional stress by Sibson's (1974) relationship:

$$\sigma_1 - \sigma_3 = \beta \rho g z (1 - \lambda)$$

where

σ_1 is the largest and σ_3 the smallest principal stress

β is a numerical parameter depending on the type of faulting (our procedure was carried out using $\beta=3$, which means thrusting)

ρ average density

g gravity acceleration

z depth

λ is the pore fluid factor (in the upper crust, supposing the pore pressure to be hydrostatic, l=0.36)

In the ductile zone we have used the so-called Dorn relation (Ranalli 1995) with the necessary modification:

$$\sigma_1 - \sigma_3 = (\varepsilon'/A_D)^{1/n} \exp(E/nRT)$$

where

ε' creep rate

A_D material constant (Dorn parameter)

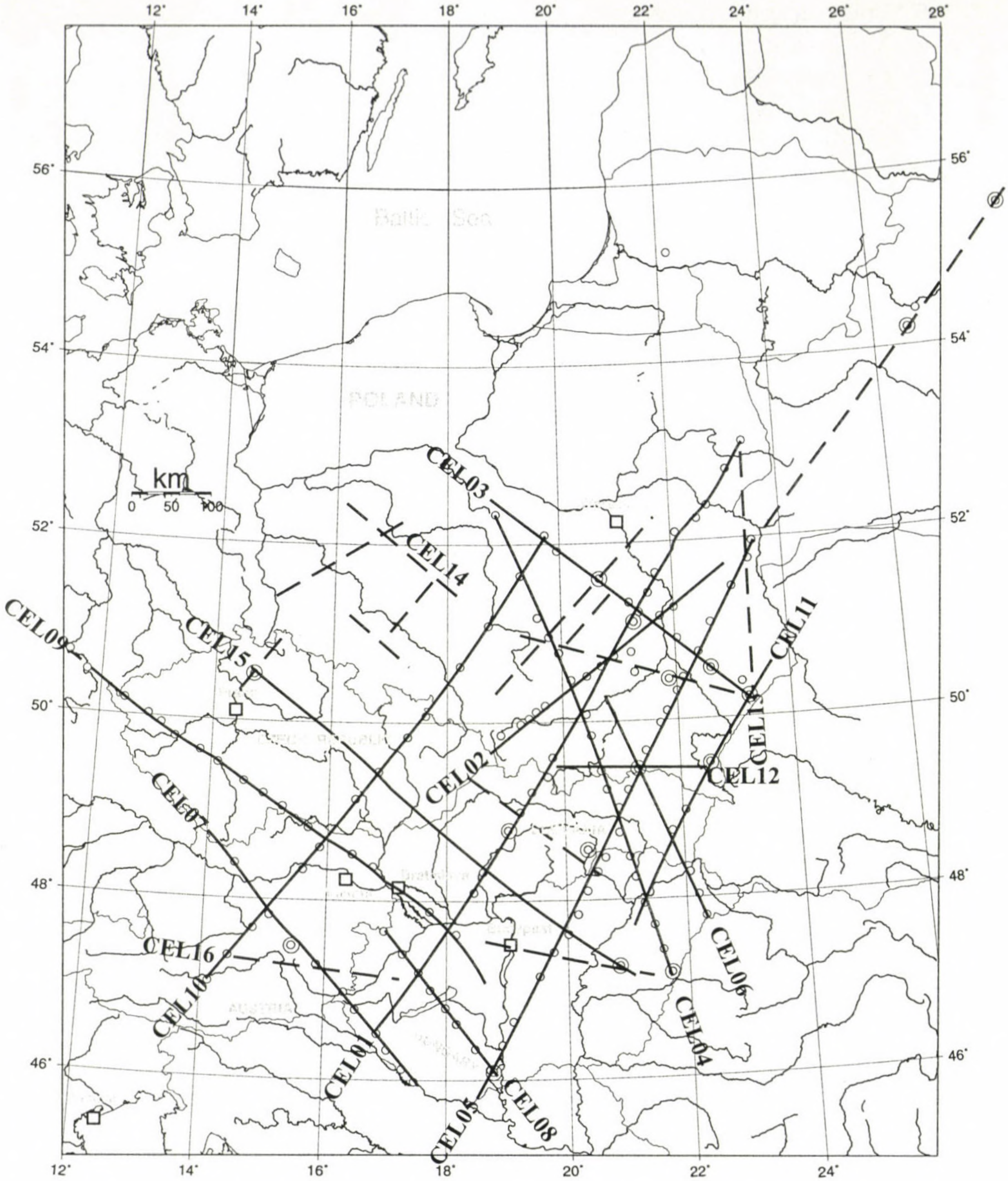


Fig. 6
Layout of the CELEBRATION seismic survey (Guterch et al. 2000)

E creep activation energy
n stress exponent
R gas constant
T temperature

E activation energy was estimated for the consecutive depth intervals from the interval velocities with the help of the diagram of Meissner et al. (1991). Using the E values from the V_p-E diagram we looked for the material constants A_D and n among the published laboratory data of Chopra and Paterson (1984), Kirby and Kronenberg (1987), Ranally and Murphy (1987), Strehlau and Meissner (1987) and Ranalli (1995). A further aspect in selecting a parameter from the wide range of laboratory data was to find those rocks which provided similar velocity values at high pressure and temperature in laboratory as those determined in section KESZ-1, and whose occurrence in the given interval is probable (e.g. Christensen and Mooney 1995).

In the center of Fig. 5 the strength profile computed in this fashion is presented, while the right side shows the respective metamorphic facies sequence according to Fig. 4. It should be noted that AVO inversion for the vicinity of the Mohorovicic discontinuity carried out on the PGT-4 deep seismic data (Takács and Hajnal 2000) resulted the same trend, as can be seen in Fig. 5 strength profile at the 21–26 km depth range.

Interpretation of subhorizontal reflections in the pre-Tertiary basement

Interpreting the high-amplitude subhorizontal reflections of Figs 1 and 3, one must take the following aspects into consideration:

1) According to Figs 4 and 5 these rocks are in the lower range of the zeolite metamorphic facies. In the strength profile a brittle-plastic transition zone can be found there, which is called semi-brittle by Kohlstedt et al. (1995). In this range the flow stress is equal to the effective stress (the effective stress is the difference between lithostatic and pore pressures).

2) In borehole Sár-1, near the southern end of seismic profile *B*, Árkai et al. (1998) determined that the Variscan prograde metamorphism of the 924 m-thick polymetamorphic overthrust block belongs to the amphibolite facies. The retrograde metamorphism of this block is connected to a mylonite formation and should be older than the Cretaceous overthrusting, which is rather widespread in the region. The underlying Mesozoic – probably Triassic – rocks (penetrated between 3941 and 4799.2 m) are in a para-autochthonous position (Árkai et al. 2000). Their upper part suffered a low-temperature anchizonal prograde metamorphism, while the transformation of their middle and lower parts belongs to the transitional anchizonal and epizonal metamorphism. The event most probably happened during the Austrian orogeny (100–105 Ma) but before Austrian overthrusting. However, a 30 million year-old overprint of Mesozoic

rocks suffering prograde metamorphism in the Cretaceous is known from the Dráva Basin in the same Tisia structural unit (Balogh et al. 1990).

3) According to Fyfe et al. (1978) prograde metamorphism is accompanied by loss of water and compaction. If the rock gets into low *P–T* conditions after suffering high-grade metamorphism, it can absorb a small amount of water only and keeps its former structure. If one can find the signature of retrograde metamorphism in metamorphic rocks, Fyfe et al. conclude that it is the result of some tectonic events.

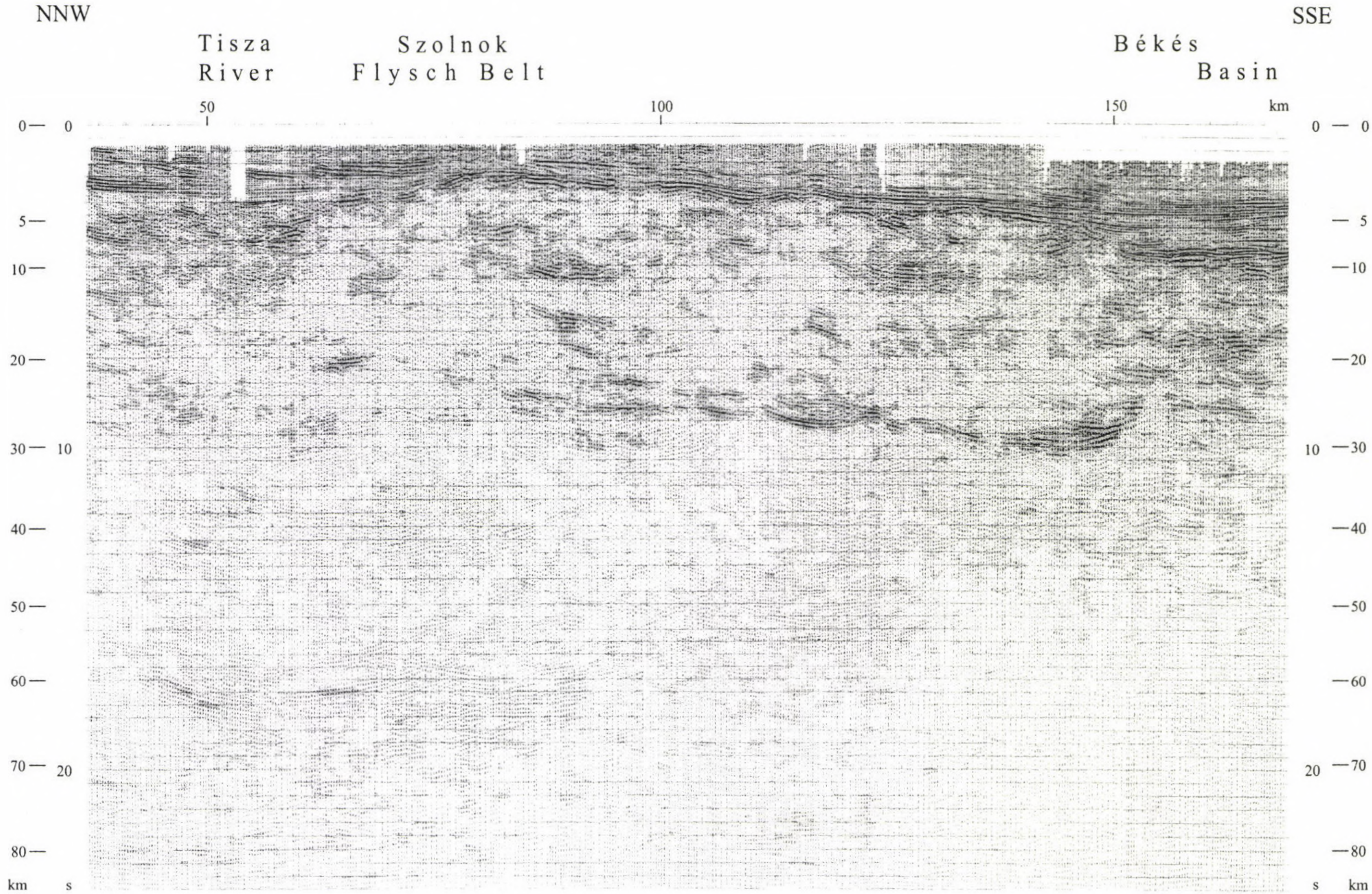
In the lower central part of Fig. 1 the steeply dipping reflections suggest the presence of a shear zone up to the surface of the basement. The bundle of high-energy subhorizontal reflections barely extend this zone in both directions. This seismic section suggests that in the extensional era of the Upper Tertiary (Horváth and Royden 1981) the basement rocks formerly suffering metamorphism, softened and received the water necessary for the retrograde metamorphism from the shear zone. At recrystallization, as the effect of the near-vertical principal maximum stress and the existing geothermal gradient, subhorizontal foliation, overprinting the shear zone, came into existence.

This interpretation is in harmony with the findings of Kozlovsky et al. (1987) in the ultra-deep borehole of the Kola peninsula where retrograde metamorphism was found in fault zones combined with disintegrated amphibolite and granitic gneiss.

In the central part of Fig. 3 – in similar depths to our former example – one can see two shear zones, both dipping to the south (both much narrower than the one in Fig. 1) close to the ends of the series of subhorizontal reflections. In addition, on the left side of the section, a north-dipping shear zone can be recognized. The subhorizontal reflections crossing it show a much more disturbed pattern than the other ones, and the reflections belonging to the shear zone are still recognizable. In trying to interpret the different patterns we suggest that this north-dipping shear zone was revived after the retrograde metamorphism producing the subhorizontal reflections. The similar interpretation of the two profile sections is supported by their proximity (see Fig. 2) and the similar depth range of the subhorizontal reflections.

Subhorizontal, high-amplitude reflections can also be found in deep seismic sections investigating the lithosphere and the asthenosphere, but at greater depths than in our previous examples. Figure 7 presents the reflection strength (instantaneous amplitude) section of profile PGT-1. This (originally colored) time section (Posgay et al. 2000) illustrates the amplitudes by the shades of grey (the darker the higher). The high amplitude patches are located approximately at the same depth range.

Figure 8 shows a part of the migrated depth section of the same profile (Posgay et al. 1995). On the left side of the figure the interval velocities, the stability profile and the estimated metamorphic facies of Fig. 5 are plotted. As all three profiles (A and B discussed above and this part of PGT-1) are in the same Kunságia terrane


NNW
SSE
Tisza River
Szolnok Flysch Belt
Békés Basin
50
100
150
km
0 — 0
5 —
10 —
20 —
30 — 10
40 —
50 —
60 —
70 — 20
80 —
km s
0 — 0
— 5
— 10
— 20
10 — 30
— 40
— 50
— 60
20 — 70
— 80
s km

of the Tisia composite terrane (Kovács et al. 2000) the rheology- and velocity data are assumed to be applicable for this part of PGT-1 as well. Also plotted are the 300–400 °C temperature zone, which according to Klemperer (1987) is the beginning of the reflecting lower crust, and the regionally determined surface of the conducting layer according to Ádám (1987). This latter was calculated by the empirical relation of Ádám (1983):

$$H=1718.7\ q^{-1.09}$$

where
H is depth and
q is heat flow.

The reflections dipping to the SSE starting at shotpoints 85 and 90, are interpreted as overthrusts of the Austrian orogeny in the basement. Below 4 km the dipping reflections are disturbed by subhorizontal ones but later they appear quite clearly again. This pattern is interpreted in the same way as the former two examples.

Interpretation of subhorizontal reflections in the middle and lower crust

In Fig. 8, at the beginning of the section up to shotpoint 90 km, at approximately 7 km depth, a reflecting horizon of medium strength can be seen. This may be assumed to belong to the prehnite-pumpellyite grade metamorphic zone.

Dominating subhorizontal reflections appear in the depth range of 8.5–11 km, coinciding with Klemperer's zone beginning with strong reflections, determined by a wide range of investigations. It is worth mentioning that this depth range is also referred to by Kohlstedt et al. (1995) as a brittle-ductile transition called semi-brittle. The subhorizontal reflections in the depth range of 8.5–11 km coincide with the upper part of the greenschist metamorphic facies. Among the subhorizontal reflections one can also see dipping reflections (at maximum 17°) as well, whose amplitudes are similar to the subhorizontal ones. We may assume tectonic effects following the metamorphism creating the subhorizontal reflections.

The regional conducting layer of Ádám (1987) seems to coincide with a ductile zone. According to the Bostick transformation data of their MT profile, close to PGT-1, a conducting layer was found at the depth of 17.4 ± 5.1 km. They interpret the conducting layers (referring to Klemperer) as the result of free fluids due to metamorphism.

← Fig. 7
Reflection strength (instantaneous amplitudes) of seismic time section PGT-1 (Posgay et al. 2000). The darker the shading the higher the amplitudes

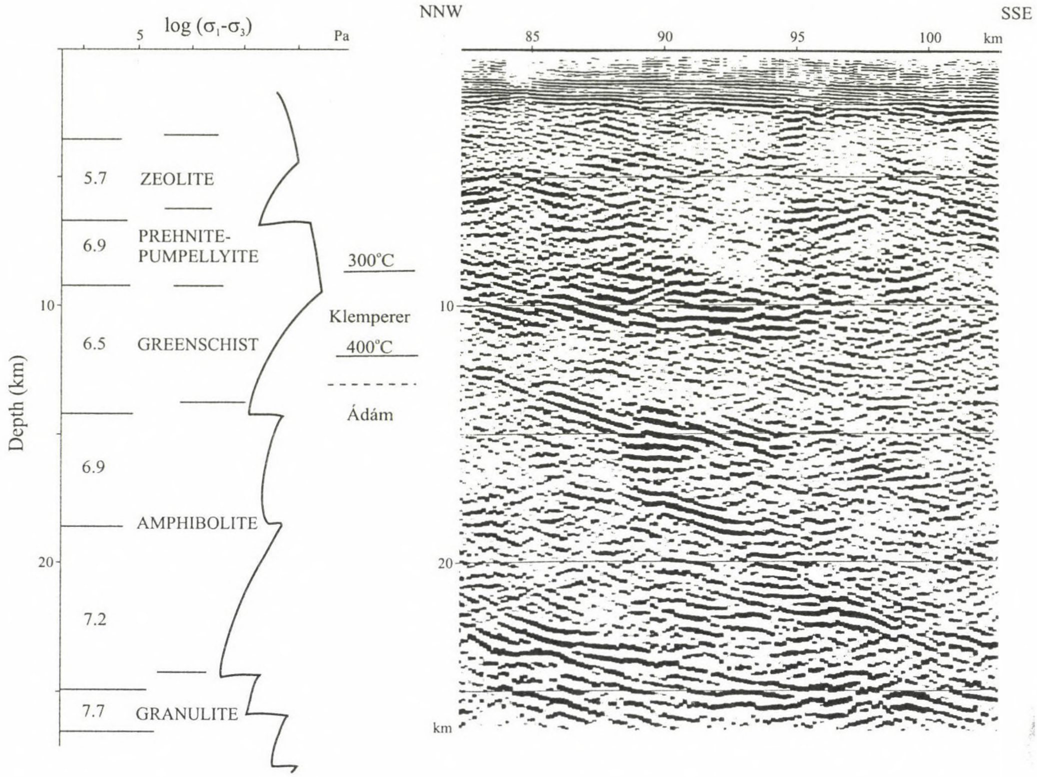


Fig. 8
The upper part of a portion of migrated depth section of PGT-1 (Posgay et al. 1995). On the left side besides the interval velocities, the strength profile and the estimated metamorphic facies, the 300–400 °C temperature range interpreted by Klemperer as the beginning of reflecting lower crust are plotted, as well as the regionally-determined depth of the conducting layer (Ádám 1987)

Starting at about 14 km depth a bundle of subhorizontal reflections of prominent energy can be seen (Fig. 8). In Fig. 7, roughly at the same depth range, similar reflections can be seen in several locations. We believe that these reflections mark the upper part of amphibolite grade metamorphism. This is in good agreement with the results of Mueller et al. (1987) and Mueller (1991), derived by complex interpretation of deep reflection and refraction data. They believe that most frequently reflections in the crust belong to positive velocity steps (increasing interval velocities).

The reflection pattern expected for the granulite-grade metamorphic facies, at around 24 km depth, intermixes with that of the crust/mantle boundary, which is close by. In a previous paper (Posgay 1993) the subhorizontal reflections of this depth range were interpreted as the surviving remnants of a former crust/mantle boundary. In Fig. 7, above the crust/mantle boundary, at the top of the granulite grade metamorphic zone, conspicuous reflections can be seen in many parts of the section, whose interpretation is possible from the point of view of metamorphism and rheology as well.

From the above-discussed examples we may conclude that the high-amplitude subhorizontal reflections of the middle and lower crust are to be found in the depth ranges (within the presently existing pressure and temperature conditions) of metamorphic mineral stabilities. This statement is worth noting because the pre-Mesozoic deformation and metamorphism of the Tisia composite terrane can be characterized by the following grades of development (Kovács et al. 2000):

- between 440–400 Ma: P=0.95–1.2 GPa and T=600–560 °C
- between 350–330 Ma: P=0.4–0.65 GPa and T=640–650 °C
- between 330–270 Ma: P=0.2–0.3 GPa and T=680–685 °C.

At the depths below 7–10 km, mainly pre-Mesozoic, crystalline rocks can be assumed in the region. Comparing their metamorphic grade with those of the metamorphic facies of Fig. 4 it seems probable that they have suffered retrograde metamorphism as well. For this process, beside the transition of solid-solid, the solid-liquid phase reaction was also necessary.

If we want to estimate the porosity and fluid content of these rocks we must take the findings of the following authors into consideration:

i) The high velocities found in the depth range of 5–10 km (Fig. 5) suggest the microfissures of the rocks to be closed as determined by Holbrook et al. (1992).

ii) Kozlovsky and his research team have stated that the water content (and its pore pressure) below the depth of closing microfissures (4.5 km) originated during the prograde metamorphism of the rocks. They assume that the rock can preserve this water for more than 1Ga. They found foliation and zonality, the latter being of complex hydrophysical character, depending on the rock type and the discharged fluids.

iii) According to Fyfe et al. (1978), during prograde metamorphism the water content and pore pressure rise suddenly and later decrease gradually. This process may vary both in time and its course depending on the metamorphic facies. Both the water content and the permeability of rocks will decrease significantly within 10^8–10^9 years.

Applying all these statements to our case, we assume the following process to be probable. Before the formation of the Carpathian Basin (Horváth and Royden 1981) a significant part, if not the whole, of the crust entered into a lower temperature range than it had been subjected to during Variscan metamorphism. The possibility of retrograde metamorphism between the Variscan Orogeny and the formation of the Carpathian Basin cannot be excluded. During the formation of the Carpathian Basin the thinning of the lithosphere (and the increase of heat flow) probably changed the former depth ranges of the various mineral stabilities. The decreasing rock volume above these stability zones caused the decrease of lithostatic pressure (and the effective pressure as well), while the pore pressure showed relative increase. Thus the pore volume and permeability were able to increase, promoting the formation of a new metamorphic facies. Because of the vertical temperature and pressure gradient the layers and their gneissic layering formed horizontally (gneissic layering is a complex metamorphic differentiation process, resulting in non-continual layers of different percentage of silicic and mafic metamorphic minerals. This expression is also used in a generalized form describing the layering of metamorphic rocks).

The retrograde metamorphism in the new mineral stability zones may have limited the upward migration of residual water, thus promoting the forming of a water-containing zone of relatively high pore pressure (Hyndman 1988). High pore pressure decreases both velocity and stability of rocks, in accordance with Christensen (1989) and Mueller (1991), who interpret the low-velocity layers in the continental middle crust as the result of pore water (resulting from metamorphism) and relatively high pore pressure compared to lithostatic pressure. The 'water film' decreases electric resistivity as well (Hyndman 1988). Gneissic layering and the underlying water-containing low-velocity zone may contribute to the formation of the high-energy reflections of crustal seismic sections and, at the same time, to that of the semi-brittle rheology zone and the zonality described by Kozlovsky et al.

It is worth mentioning that the high-amplitude subhorizontal reflections do not exceed 10–20 km horizontally. Looking for the reasons of this interruption of continuity we must suppose that the circumstances contributing to their formation or survival have been changed. Studying the flysch zone and the deep basins of southeastern Hungary in seismic section PGT-1 we have concluded that the NNW-dipping zones cutting off the extension of subhorizontal reflections indicate shear zones, which have broken the reflecting horizons (Posgay and Szentgyörgyi 1991; Posgay et al. 1996, 1997). It is probable that velocity changes caused by tectonic effects were neglected in data processing, thus contributing to the discontinuity of reflections in shear zones (Posgay et al. 2000).

For the time being, we also suppose that the permeability of these shear zones are much higher than in the surrounding rocks; thus the water originating from metamorphism was not trapped below the newly formed metamorphic zones. The repeatedly shearing prevented the cracks and fissures from clogging. It may also be supposed that the effective pressure in these shear zones are different from that existing in the intact parts of the surrounding rocks; thus the depth ranges of stability of metamorphic facies differ in the two media. Furthermore, it can also be supposed that the pressure conditions necessary for the forming of a new metamorphic zone were not stable in such shear zones for the necessary time.

We can observe the discontinuity of subhorizontal reflections not only in shear zones, as can be seen in the right side of Fig. 8. Between shotpoints 93 and 103, at the depth of 7 km, one can observe a reflection pattern characteristic of intrusions. It is possible that in the left side of the section at the depth of 7 km, and also between 8.5 and 11 km, the discontinuity of reflections is also connected with this intrusion, which is most probably the root of the magnetic body causing the Túrkeve magnetic anomaly of (Posgay 1967).

Conclusions

The studies described above led us to the conclusion that in the consolidated crust the reflecting horizons also reflect the temperature and pressure conditions (similar to the present ones), which were altered during the formation of the Carpathian Basin. The depth coincidence of subhorizontal reflections with the present stability zones of metamorphic facies, and with the brittle-ductile transition zones of the strength profile, suggest that the former structure of the crust has been overprinted following the formation of the Carpathian Basin.

In the pre-Neogene basement, at about 5 km depth, the subhorizontal reflections overprint a steeply-dipping shear zone. It can be supposed that the water necessary for retrograde metamorphism was provided by the shear zone.

Our study area lies on the Kunságia terrane of the Tisia structural unit. Considerations deduced from deep borehole data suggest that, below 7–10 km, pre-Mesozoic crystalline rocks are to be found in the area. These rocks were already metamorphosed during the Variscan Orogeny above 650 °C. The temperature and pressure conditions similar to present ones could have initiated retrograde processes in which the water (released by Variscan metamorphism) played an important role in the parts of the crust marked by relatively low permeability. The subhorizontal gneissic layering is the result of the vertical gravity and geothermal gradient. The discontinuity of subhorizontal reflections along shear zones, cutting through a great part of the crust, may suggest the absence of retrograde transitions.

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Tectonic implications of Tertiary paleomagnetic results from the PANCARDI area (Hungarian contribution)

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This paper deals with the tectonic aspects of the Tertiary paleomagnetic results obtained by the Paleomagnetic Laboratory of the Eötvös Loránd Geophysical Institute in bilateral cooperation with several research centres during the lifetime of PANCARDI. It summarizes the tectonic implications of the very recently obtained paleomagnetic data over the foredeep of the West Carpathians, the Inner West Carpathians (mainly the flysch basins), the North Hungarian–South Slovakian Paleogene Basin, the Tokaj Mts and the East Slovak Basin, the Transdanubian Central Range, the intramontane basins of the Eastern Alps and the Mura Depression, the Croatian part of the Pannonian Basin, all of which exhibit counterclockwise rotated declinations. According to the paleomagnetic observations both the angle and timing of the movements is variable though fairly consistent within each of the above areas. Clockwise rotations were observed for the Tertiary of the main Paleozoic-Mesozoic body of the Mecsek Mts., in the Mürztal and in the area between the Periadriatic and the Sava fault systems in Slovenia.

The Tertiary paleomagnetic observations of the last decade require modification and/or refining of the existing reconstruction models of the PANCARDI region.

Key words: PANCARDI area, Tertiary, paleomagnetism, tectonic implications

Introduction

In a review of paleomagnetic results from the Alpine–Mediterranean belt, Márton et al. (1987) suggested that the rotations observed on Paleozoic rocks must be due to post-Paleozoic movements. Later, Márton and Mauritsch (1990) concluded that the most important rotations in the Central Mediterranean, which also includes the PANCARDI region, are of post-Mesozoic age. Thus, paleomagnetists shifted their interests to the Tertiary of the Central Mediterranean. Incidentally one of the projects of EUROPROBE, PANCARDI also needed Tertiary paleomagnetic results in order to better constrain the Tertiary Carpathian evolutionary model (project 9).

Most of the paleomagnetic results reviewed in this paper were obtained during the lifetime of PANCARDI by several teams sharing a common resource, which is the Paleomagnetic Laboratory of the Eötvös Loránd Geophysical Institute of Hungary. This laboratory processed the overwhelming majority of the samples, while field work and tectonic interpretation was performed by the respective

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teams (the names of team members with academic qualifications are found in the reference list).

The age and lithology of the studied rocks range from the K/T boundary (Dinarides) to Upper Pontian (Zagorje–Mid-Transdanubian Zone). Concerning lithology, sediments were preferred to igneous rocks (metamorphic units were avoided, no matter how low the degree of metamorphism was). The reasons for this are that in the first case the control on local tectonics is better, and the paleomagnetic direction is practically not influenced by the secular variation of the Earth's magnetic field. Nevertheless, the tectonic significance of paleomagnetic data from igneous rocks will also be discussed, either in combination with data from sediments or independently, when a large number of igneous sites were studied.

Paleomagnetic vectors from the PANCARDI area will be compared to reference directions. One of them is the expected declination-inclination at the sampling area in the central axial dipole field of the Earth, where the declination is zero and the inclination depends on the present latitude (e.g. Irving 1964). Further reference directions (Table I) will be those calculated from the paleomagnetic poles defined for Africa and stable Europe (data compiled by Besse and Courtillot 1991). Occasionally, paleomagnetic vectors will be compared between different parts of the PANCARDI area.

For the discussion of the tectonic implications of the paleomagnetic results, the PANCARDI region will be subdivided into the following areas: A) West Carpathians including NE Hungary and the Polish segment of the Foredeep, B)

Table I

Expected declination, inclination and paleolatitude for Budapest and Makarska–Rijeka–Kraków–Jaroslaw, respectively for the last 70Ma, assuming direct connection with Europe or Africa

Ma	D°	I°	lat.° N	D°	I°	lat.° N
		Budapest		Makarska – Rijeka – Kraków – Jaroslaw maximum variation in the study area		
			European framework			
10	6	62	44	5 – 7	59 – 64	39 – 46
20	8	61	42	7 – 9	57 – 64	38 – 45
30	11	62	43	10 – 12	58 – 64	39 – 46
40	10	60	41	9 – 12	56 – 63	37 – 44
50	12	58	39	11 – 13	54 – 61	35 – 42
60	5	56	37	4 – 6	52 – 59	31 – 39
70	2	54	35	1 – 3	50 – 57	31 – 37
			African framework			
10	5	62	43	4 – 6	58 – 64	39 – 46
20	6	60	41	5 – 6	56 – 62	37 – 44
30	7	60	42	6 – 8	57 – 63	37 – 44
40	3	57	38	2 – 4	53 – 60	32 – 40
50	2	53	35	1 – 3	49 – 57	30 – 37
60	354	50	31	353 – 354	46 – 53	27 – 34
70	351	48	29	349 – 352	43 – 50	25 – 31

the area south of the Northern Calcareous Alps, north of the Periadriatic line plus the Transdanubian Central Range, C) southwestern part of the Pannonian Basin, D) Adriatic Foreland and Dinarides.

A) West Carpathians and the Polish segment of the Foredeep

This area is characterized by large to moderate CCW rotated paleomagnetic declinations with an interesting time-space distribution.

In the best studied part, which is the area of the North Hungarian–South Slovak Paleogene Basin (Fig. 1: 1), the declination is 280° on average for sediments

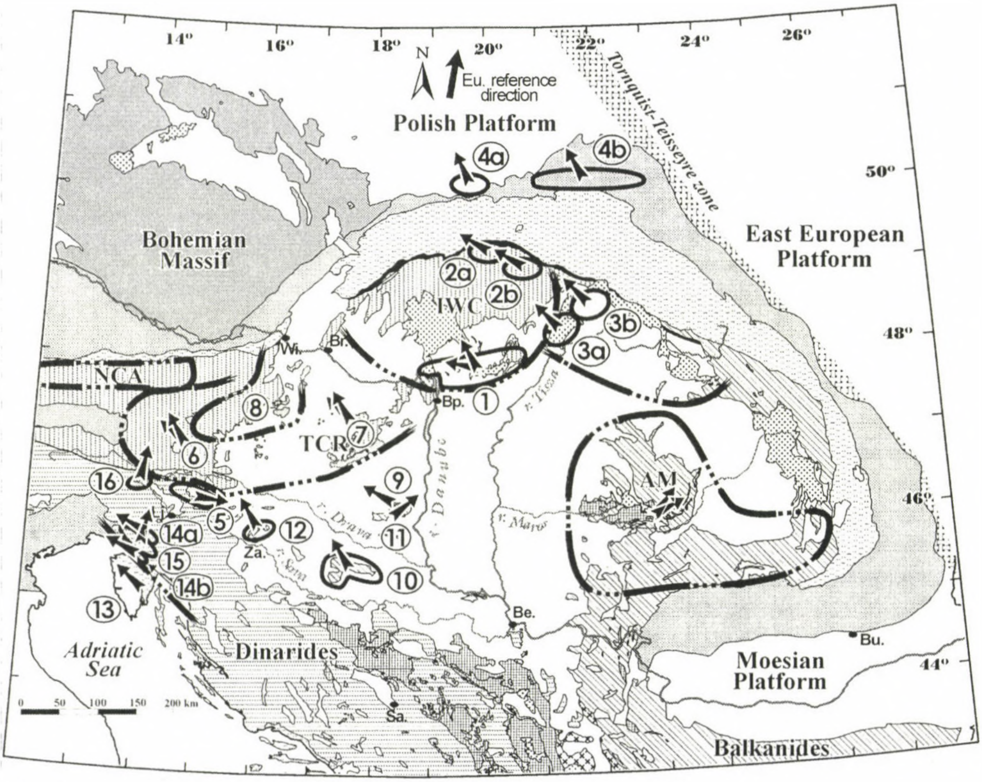

Fig. 1
The PANCARDI area with paleomagnetic declinations (overall paleomagnetic mean declinations, each based on several geographically distributed site and locality paleomagnetic mean directions from the numbered area represented by the arrow). The arrows were plotted relative to the 20° meridian. Note that area 8 has no arrow, since Tertiary rotations of different direction occur there. The numbers are indicated in the text. The area containing the Apuseni Mts. is known to have rotated CW in the time interval 12–14Ma and also earlier in the Tertiary in the same sense (Panaiotu 1998). Abbreviations: OWC – Outer West Carpathians; IWC – Inner West Carpathians; NCA – Northern Calcareous Alps; TCR – Transdanubian Central Range; AM – Apuseni Mts

and ignimbrites of Late Eocene (the oldest rock studied) through Ottnangian age. The declination for Karpatian–Lower Badenian rocks is around 330°, while Upper Badenian and younger igneous rocks are characterized by 0–10° (Márton and Márton 1996; Márton et al. 1996; Karátson et al. 2000). The K/Ar ages for the volcanics of the respective groups are 21.0–18.5Ma, 17.5–16.0Ma and younger than 14.5Ma (Márton and Pécskay 1998). In terms of tectonic rotations these observations mean 50 ° CCW, followed by 40° CCW rotation of the area relative to stable Europe (Table I) in the time intervals of 18.5–17.5Ma and 16.0–14.5Ma, respectively.

In the Podhale and Levoca basins (Fig. 1: 2a and 2b) of the Inner West Carpathians (IWC) the average declination measured for the Paleogene is around 300° (Márton et al. 1999a). Outside of these basins, sporadic data from the Paleogene of the Inner West Carpathians (Túnyi and Márton 1996; Túnyi and Kovác 1991) exhibit similar declinations. Though paleomagnetic constraints younger than Oligocene outside of the Paleogene Basin are lacking, it is reasonable to think of the Inner West Carpathians (as we shall see, east of the Hernád tectonic line) as connected to the North Hungarian-South Slovak Paleogene Basin during Miocene tectonic events.

East of the Hernád tectonic line the oldest Tertiary rocks croping out are of Badenian age. Thus we have no paleomagnetic constraints for pre-Badenian times. However, a large number of volcanic sites and a few sediments, the first from both the Tokaj Mts and from the East Slovak Basin, the second only from the latter (Márton et al. 2000a; Márton and Pécskay 1995) exhibit westerly declinations, while paleomagnetic vectors obtained from igneous rocks, mostly of Pannonian K/Ar age, are aligned with the stable European reference direction of corresponding age (Table I). The angle of the rotation for the first group is about 40–50°; the estimated age of rotation is the Sarmatian–Pannonian boundary. The research in progress east of the Hernád line, both in Hungary and Slovakia, will hopefully supply more precise values for both the angle and timing of the rotation. Nevertheless, we can safely conclude at the present stage that the youngest rotation in the Inner Carpathians, east of the Hernád Line, is a few million years younger than West of it.

A pilot study in the Outer West Carpathians (OWC) revealed CCW rotations of variable angle (Márton et al. 2000b). The positive results are few, represent rocks of different ages, and are distributed over a large area and in different nappe units. Clearly, further research is needed before any conclusion can be reached about possible differential movements within the Outer West Carpathians.

The Molasse of the Foredeep in front of the West Carpathians is subdivided into a folded and unfolded subunit. Badenian in the west (Fig. 1: 4a) and Sarmatian-Pannonian sediments in the east (Fig. 1: 4b) exhibit moderate CCW rotations, except for one Badenian locality in the west (Márton et al. 1999b; Márton et al. 2000b; Márton et al. 2000c). So far, these rotations have been tentatively connected to the accretion of the Inner, later the Outer West

Carpathians to the European margin, which process led to the incorporation of the Molasse into the accretionary wedge (Márton et al. 2000b).

B) The area south of the Northern Calcareous Alps, north of the Periadriatic Line plus the Transdanubian Central Range

This area is bordered by the Northern Calcareous Alps (NCA) and paleomagnetic megaunit A in the north. The southern boundary is the Periadriatic–Sava fault system, and further to the east, the Balaton Line. Concerning the northern boundary, the few Tertiary basins of the Northern Calcareous Alps were not studied in the framework of PANCARDI, because the youngest data presently available (Late Cretaceous–Paleocene) from the Gosau Basins (Mauritsch and Becke 1987) had implied that the Northern Calcareous Alps was practically emplaced prior to the Late Cretaceous. The situation in the Inner West Carpathians was summarized in the previous chapter. The Slovenian part of the Periadriatic-Sava fault system (Fig. 1: 5) exhibits a complex rotation pattern with the dominance of large to moderate CW rotations (Fodor et al. 1998), a paleomagnetic picture which is unique in the PANCARDI area. Right lateral shear is known to be active in this fault system, and the paleomagnetic results are in line with the tectonic observations. The Balaton Line and its neighborhood are paleomagnetically unexplored.

In the area itself CCW rotated declinations prevail. The basic difference, compared to the North Hungarian-South Slovakian Paleogene Basin, is the paleomagnetic evidence for post-Badenian rotations (Mura Depression, Klagenfurt and Lavanttal Basins, Transdanubian Central Range).

The observed declinations between the Northern Calcareous Alps and the Periadriatic Lineament (Fig. 1: 6) are interpreted in the following way (Márton et al. 2000d). During the lateral extrusion of the Eastern Alps (East of the Tauern Window) a large sinistral wrench corridor was created by faster movement along the Periadriatic Lineament than along the Ennstal Fault. In this corridor, domino block rotations occurred and the rotations were accommodated by NNW–SSE-trending dextral strike-slip faults (e.g. Lavanttal Fault). Some of the rotation, however, took place after extrusion: CCW rotation in the order of 30° (with respect to stable Europe) must have been connected to the overall rotation of the area comprising the Tertiary intramontane basins. This angle is similar to the one observed in Lower Pannonian sediments in the Transdanubian Central Range (Márton 2000, Fig. 1: 7). However, there is a transition zone (Fig. 1: 8) with a complex rotation pattern between the Alps and the Transdanubian Central Range (Márton et al. 2000d). Therefore a rigid connection between the two during the last CCW rotation is unlikely. One may be tempted to think of the last rotation of the Transdanubian Central Range as due to domino block rotations as Tari (1991) suggested. At the time of his publication the available paleomagnetic data showing 45° post-Oligocene CCW rotation with respect to stable Europe were not yet resolved into two components. The dilemma, however, is that the tectonic event he held responsible for the domino block rotations is of Sarmatian age, while the recently obtained paleomagnetic data point to post-Lower Pannonian age of the paleomagnetic rotation. This problem calls for a reappraisal of the age of the movements or of the sampled rocks of the Lower Pannonian paleomagnetic directions.

Post-mid Eocene through Oligocene sediments and igneous rocks in the Transdanubian Central Range exhibit 15° larger CCW rotation than the Lower Pannonian sediments. Rocks of similar age from the West Carpathians, including the North Hungarian–South Slovakian Paleogene Basin, are characterized by much larger rotation in the same sense. However, the magnetization of the post-mid Eocene sediments from the Transdanubian Central Range seems to be younger than the stratigraphic age. Thus, paleomagnetic data from the Transdanubian Central Range may be interpreted either as evidence for more or less independent movement during the Miocene of the Transdanubian Central Range and the Inner West Carpathians sensu lato, or as evidence for regional remagnetization of the mid-late Eocene and Oligocene sediments of the Transdanubian Central Range following an Ottnangian rotation.

C) Southwestern part of the Pannonian Basin

The paleomagnetically studied Neogene of the Mecsek Mts (Márton and Márton 1999) and that of the Slavonian Inselbergs (Márton et al. 1999c) represent the young cover of the Tisza (Tisia) magatectonic unit. There are also paleomagnetic results from the Croatian part of the Zagorje–Mid-Transdanubian Zone (Márton et al. 2000e).

The Neogene cover of the Tisza megatectonic unit was studied with the aim of obtaining paleomagnetic constraints for the timing of the CW rotation of the megatectonic unit, defined as a microplate during its Tertiary tectonic history (e.g. Balla 1986; Csontos 1995). Contrary to expectations, CCW rotations of probably Karpatian age were observed for ignimbrites at the northern margin of the Mecsek Mts. (Fig.1: 9) and on all the studied sediments and igneous rocks from the Slavonian Inselbergs (Fig.1: 10), where the rotation is of post-Pannonian age. CW rotations are confined to the main Paleozoic–Mesosoic body of the Mecsek (Fig.1: 11). The timing of the CW rotations is not constrained paleomagnetically. These paleomagnetic results imply that the Tisza megatectonic unit broke apart in the Neogene and differential rotations of probably different age must have occurred in the area, affecting not only the inselbergs but also the buried basement.

The Zagorje–Mid-Transdanubian Zone, which is assumed to be the southern margin of the North Pannonian megatectonic unit (Balla 1984; Csontos 1995; Fodor et al. 1998) was studied in Croatia. Sediments of Egerian through Upper Pontian age (Fig.1: 12) were found to exhibit CCW rotations with a somewhat larger angle for pre-Badenian than post-Karpatian age. In the context of Neogene CCW rotations occurring north of the Periadriatic Line as well as in the Slavonian Inselbergs, it seems plausible that a large area, north of the "rigid" Adriatic microplate, was caught up in the CCW rotation of the microplate (Márton et al. in prep.).

D) Adriatic Foreland and Dinarides

The most important result from this area is the paleomagnetic proof for the significant post-mid-Eocene CCW (about 45°) rotation of the "rigid" Adriatic foreland (Fig. 1: 13) with respect to Africa (Márton et al. 1995). In other words, we now have solid paleomagnetic evidence for the separation of the Adriatic promontory from Africa in the Tertiary.

The Dinarides were studied in Istria (Croatia) and in Slovenia. The results obtained so far suggest larger-angle CCW rotations (Fig. 1: 14) than in the Adriatic foreland, except in the Komen Plateau (Fig. 1: 15), where positive tilt test supports the lack of significant rotation with respect to stable Europe since the mid-Eocene. North, south and east of the Komen Plateau there are observations for CCW rotations, which makes the tectonic interpretation of the paleomagnetic results from the plateau very difficult. It is, however, interesting to note that the Julian Alps (Fodor et al. 1998, Fig. 1: 16) and the Komen Plateau are paleomagnetically similar. This situation differs from the usual picture of tectonically complex areas with highly variable declinations. Future research should focus on the northwestern and southeastern connections of the slivers or nappes in the Dinarides, with declinations consistent within a certain area, yet with profound differences perpendicular to the general strike between neighboring areas.

Tectonic implications of the observed paleoinclinations

As Table I shows, expected coeval declinations vary very little within the study area both in the European and the African reference systems (see max. variation in the area defined by the corner points Makarska–Rijeka–Kraków–Jaroslaw). Consequently, a measured paleodeclination for any point within the PANCARDI area means practically the same rotation with respect to the same reference system. The situation differs for paleoinclinations, since the variation is considerable even within our relatively small study area. Therefore, measured paleoinclinations must be compared with expected inclinations for the source area of the respective paleomagnetic result.

Such a comparison tells us that paleoinclinations characterizing rocks younger than Karpatian of the PANCARDI area agree quite well with the inclination expected in a stable European framework. In other words all units of the area were more or less at the present latitude. The inclinations of pre-Karpatian rocks (Fig. 1: 1, 2a and b, 7, 8 and 12) suggest an important and rapid northward shift following the Ottnangian (several hundreds of kilometres for 7 and 12, less for 6, and even less for 2a and b). The inclinations from paleomagnetic megaunit D are not very useful for estimating the shift in north-south direction, since the remanences may be of post tilting age, except for 15 and 16 (Fig. 1). For them the inclinations are somewhat shallower than expected, suggesting moderate northward shift following the Middle Eocene (Komen Plateau) and the Oligocene (Julian Alps), respectively.

General conclusions

Most of the models published earlier (beginning with the eighties) attempted to satisfy the that time existing paleomagnetic constraints. Moreover, certain elements of the reconstructions, like the Miocene CCW rotation of the Transdanubian Central Range were entirely based on the suggestion of paleomagnetists (Márton and Márton 1983; Márton 1986).

However, Tertiary paleomagnetic results relevant to the paleomagnetic subproject of PANCARDI and obtained through bilateral projects of the Paleomagnetic Laboratory of ELGI with several research groups imply that the tectonic history was more complicated than depicted by the existing simple tectonic reconstruction models of the PANCARDI region. Clearly, paleomagnetic observations on Tertiary rocks call for modifying and/or refining the models.

At certain points there are obviously great difficulties in reconciling the Tertiary paleomagnetic data with a plausible model (e.g. Komen Plateau and its surroundings, West Carpathian Foredeep): there are other results, which permit alternative models (e.g. Croatian part of the Pannonian Basin). Nevertheless, high-quality paleomagnetic data should not be omitted when the tectonic history of a region is evaluated, for they may trigger a new generation of tectonic models.

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Paleobiogeographical analysis: a tool for the reconstruction of Mesozoic Tethyan and Penninic basins

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The paleobiogeographical relationships of the major terranes of the Alpine–Carpathian–Dinaridic region changed several times during the Mesozoic. In the Middle Triassic the "South Alpine" fauna was overwhelming but the influence of the "Dinaric" (in Aggtelekia, Austroalpinia and the eastern part of Tisia) and the "German" (in the western part of Tisia) faunas can also be demonstrated. In the first half of the Jurassic (Sinemurian to Bathonian?) Adria, Austroalpinia and Pelsonia belonged to the Mediterranean province, whereas Tisia and Dacia were inhabited by NW-European faunas. In the later part of the Jurassic and in the Early Cretaceous (up to the Aptian) all terranes of the Alpine–Carpathian region came under the dominance of the Mediterranean province. From the Late Aptian onwards, the influence of the NW-European province rapidly increased throughout the area concerned. These changes are interpreted in terms of plate-tectonic changes in the western Tethys. In the Middle Triassic, the western end of the Tethys (e.g. the Meliata Ocean) was surrounded by a more or less continuous shelf margin, where the differentiation of true faunal provinces was not possible. The Early Jurassic opening of the Ligurian–Penninic ocean formed a barrier between the faunal provinces of stable Europe (NW-European province) and the drifted microcontinent (Mediterranean Province). In the second half of the Jurassic and in the Early Cretaceous, due to the spreading of the Valais–Magura oceanic belt, further microcontinents (including Tisia and Pieninia) drifted apart form the European shelf; these came under the influence of the Mediterranean province. In the Late Aptian and Albian, the progressive consumption of the oceanic Tethys led to continental collisions in the Alpine–Carpathian region and, through these corridors, the NW-European faunal elements migrated to the impoverished areas of the Mediterranean province.

Key words: Mesozoic, Alpine–Carpathian–Dinaridic region, terranes, western Tethys, paleobiogeography

Introduction

The study of distribution of fossil plants and animals is a very important and useful tool in paleogeographic and plate-tectonic analyses. Even E. Suess' ingenious idea, the Tethys, was induced by paleontological observations well before the plate-tectonic formulation of the Tethys Ocean.

Biogeographic provinces (distinctive groups of species confined to certain regions) have been present on the Earth throughout the Phanerozoic. The primary factors causing provincialism have been debated at length, with environmental factors (e.g. temperature, salinity, food) on one hand and

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geographic factors (distance, physical barriers) on the other. Both groups of factors are important in controlling the distribution of plants and animals. However, it is to be noted that, although the environment may be very similar or almost the same at very distant points of the Earth, the biota differs in composition from place to place. Therefore it seems to be justified to use the barrier/distance concept (i.e. the deep-sea/oceanic arms of the Tethys as barriers) for the explanation of the faunal differences in the Alpine-Carpathian region.

Based on paleobiogeographic work in the northern and southern mountains of Hungary (Transdanubian Central Range and Mecsek–Villány Mountains, respectively) Géczy (1973) was the first to state that there was a major difference in the Mesozoic fossil assemblages of the Intra-Carpathian area. The northern mountains and southern mountains belonged to two different faunal provinces during the Jurassic. These Early Jurassic faunal provinces were named by Géczy (1973, 1984a) and Vörös (1977, 1984) as NW-European province, or Neumayria, which corresponds to the European continental margin, and the Mediterranean province, which is found in the Italian peninsula, Southern Alps, and Austroalpine units.

Since Géczy's pioneering work a great amount of Mesozoic paleobiogeographic data have been collected for the Alpine–Carpathian region by working groups of national (OTKA, Hungary) and international (IGCP) projects lead by the present author. These results are summarized and their interpretation is given here.

Methods

The most important task was to produce a more or less uniform database, including the occurrences of the most diverse fossil groups in the greatest number of stratigraphic intervals possible over the entire Alpine–Carpathian region, with special attention to the territory of Hungary. Comparisons with the more distant areas of the respective faunal provinces were based on personal studies or literature data, e.g. classical paleontological monographs.

A proper comparative paleobiogeographic study was not possible because similarity coefficients would have been strongly biased. Poorly known (poorly collected) faunas with low number of taxa should not be compared with rich faunas by means of numerical methods. On the other hand, both poor and rich faunas may furnish valuable and comparable data if we study the distribution of taxa distinctive of separate faunal provinces. The designation of "distinctive taxa" may involve a kind of subjectivity; nevertheless, the method was successfully used by Middlemiss and Smirnova (1988), Prosorovskaya and Vörös (1988), Sandy (1988), Vörös (1993) and many others.

The distribution of "distinctive taxa" of several fossil groups has been recorded on basemaps showing the boundaries of the main tectono-stratigraphic terranes of the study area (after Vörös 1993 and Kovács et al. 2000). Paleobiogeographic maps have been synthesized from these distributions. These paleobiogeographic maps for separate stratigraphic intervals have been plotted on the same terrane basemaps. The terrane names used in the present paper are the following (according to the above-mentioned authors): Adria, Austroalpinia, Penninia, Pieninia, Pelsonia (including Bakonyia, Bükkia, Aggtelekia), Zemplenia, Tisia, Dacia.

Results

The detailed paleobiogeographic studies made by several contributing authors have been focused on the territory of Hungary and on the most important marine fossil groups. The studies have been limited to the Middle Triassic (Anisian) to the mid-Cretaceous (Albian) stratigraphic interval, because this has been thought to offer a nearly continuous data set.

Triassic

From the Triassic, only the Anisian distributions have been analyzed. The Early Triassic faunas are poor and not well studied. The Late Triassic faunas of Pelsonia are very rich and well known; their Alpine affinities have been evident for a long time. However, in the Hungarian part of Tisia, this time interval is represented by non-marine sediments or by stratigraphic gaps.

Anisian

Ammonoids provided data from Bakonyia, Aggtelekia and the Mecsek zone of Tisia (Vörös 1992). The Late Pelsonian paleobiogeographic picture is rather homogeneous: the fauna of Bakonyia and the Mecsek Zone shows South Alpine affinity. It must be remarked, however, that at this time a definite Alpine influence appears in the Germanic Muschelkalk basin as well (Urlichs and Mundlos 1985). In the framework of the Late Illyrian faunal belts, Aggtelekia belonged to the "Dinaric" and Bakonyia to the "South Alpine" belt, with a slight "Dinaric" influence.

The distribution of brachiopods shows a similar pattern (Pálfy 1992). The fauna of Aggtelekia has a "South Alpine" affinity with slight "Dinaric" influence. The brachiopods of the Balaton Highland (Bakonyia) are definitely "South Alpine" in character. Within Tisia, the Mecsek and Villány faunas also show "Germanic" composition, but in the Transylvanian part (Bihor Autochton) some "South Alpine" and even "Dinaric" species were found. This may indicate a paleoenvironmental differentiation within Tisia, involving more open marine facies in the present-day eastern part of this terrane.

Jurassic

Within the Jurassic, the Sinemurian, Pliensbachian, Toarcian, Bajocian, Bathonian, Callovian, Oxfordian, Kimmeridgian and Tithonian stages provided faunistic data for comparison and evaluation. The paleobiogeographical affinities of the Jurassic (and also the Early Cretaceous) faunas may be given in terms of the "NW-European" and "Mediterranean" faunal provinces, respectively.

Sinemurian

According to the studies on Gastropoda, the faunas of Bakonyia are definitely Mediterranean. The scarce data available from the Mecsek zone of Tisia points to NW-European affinity (Szabó 1990). Brachiopod studies, based on a much larger material, underscore the above result (Dulai 1990; Vörös 1993).

Pliensbachian

The detailed comparative studies on Bivalvia (Szente 1990), Gastropoda (Szabó 1990, 1994), Ammonoidea (Dommergues and Géczy 1989; Géczy and Meister 1994) and Brachiopoda (Vörös 1993) led to a concordant conclusion: Bakonyia belonged to the Mediterranean province, while all zones of Tisia showed definite NW-European character. Brachiopods testify to the Mediterranean affinity of Aggtelekia (Vörös 1993).

Toarcian

This stage is extremely poor in benthonic fossils; therefore the evaluation was restricted to ammonoids. On this basis Bakonyia can be qualified as strictly Mediterranean, while the fauna of the Mecsek Zone of Tisia is dominantly NW-European with some Mediterranean elements (Géczy 1990).

Bajocian

Studies on Bivalvia and Brachiopoda reached at the same conclusion: the faunas of Bakonyia are definitely Mediterranean, while the Mecsek Zone of Tisia shows clear NW-European faunal affinity (Szente 1992; Vörös 1993).

On the basis of its rich Gastropoda fauna, Bakonyia belonged to the Mediterranean province (Szabó 1990).

Bathonian

Bivalves seem to prove the NW-European affinity of the Mecsek Zone of Tisia (Szente 1992). This was partly endorsed by brachiopod studies: the fauna of the Villány Zone of Tisia is markedly NW-European, while the Mecsek fauna shows strong Mediterranean influence (Vörös 1995, 2001) (see evaluation below).

On the basis of ammonoids, Pelsonia can be regarded as definitely Mediterranean. The rich fauna of the Mecsek Zone of Tisia does not show any clear character: beside the occurrence of some diagnostic NW-European species, the average of the fauna shows "Sub-Mediterranean" features (Galácz 1990). This would point to the southern, transitional zone of the NW-European province, with strong Mediterranean influence.

Callovian

From this stage, we have faunistic data only from the Villány Zone of Tisia.

Bivalves show a mixed, NW-European + Mediterranean faunal character (Szente 1992). Brachiopods showed a similar picture, with a slight dominance of the Mediterranean faunal elements (Vörös, 1993). According to Géczy (1984b) the extremely rich ammonoid fauna represents the transitional zone of the two faunal provinces.

Oxfordian and Kimmeridgian

Based on rather scarce data on Ammonoidea, the faunas of Bakonyia (Bakony, Gerecse) show definite Mediterranean character (Főzy 1990).

Tithonian

On the basis of ammonoids Bakonyia can be included in the Mediterranean province (Főzy 1990; Cecca et al. 1993), whereas weak NW-European influence was recognized in the substantially Mediterranean fauna of the Mecsek Zone of Tisia (Főzy 1990).

Brachiopods clearly prove the Mediterranean character of Bakonyia (Kázmér 1990, 1993); the scarce fauna of the Mecsek Zone of Tisia also consists of Mediterranean species.

Cretaceous

Detailed comparative studies were restricted to the Lower Cretaceous Valanginian, Barremian, Aptian and Albian stages.

Valanginian

The evaluation of ammonoid faunas proved that both Bakonyia and the Mecsek Zone of Tisia showed Mediterranean faunal character. Surprisingly, NW-European elements were not found in the Mecsek but, occasionally, only in the Bakonyia fauna (Bujtor 1992). This may be explained by the much more voluminous collecting works carried out in the Bakony and Gerecse.

Brachiopods were evaluated only in Bakonyia; here the character of the fauna is definitely Mediterranean (Somody 1992).

Barremian

On the basis of brachiopod studies, the faunas of Bakonyia show clear Mediterranean character (Somody 1992).

Aptian

Brachiopod studies resulted in a remarkable conclusion: the fauna of Bakonyia was definitely Mediterranean in the Early Aptian, whereas the Late Aptian fauna (of the Tata Limestone) is dominated by NW-European elements ("Jura fauna") (Somody 1992).

Albian

The evaluation of ammonoids has shown that the fauna of Bakonyia had a southern (Tethyan) affinity. On the other hand, the fauna of the Villány Zone of

Tisia was dominated by NW-European faunal elements, though a few Tethyan taxa also occurred (Bujtor 1990).

Paleobiogeographical evaluation

The above summary of results concerns mainly the Hungarian or intra-Carpathian paleobiogeographical data but the scope of the cited works covered much wider areas of Europe and the Peri-Mediterranean region. Therefore the evaluation will also be performed on a larger scale but always focused on the Alpine–Carpathian terranes.

On the basis of the paleobiogeographic changes, the Middle Triassic to mid-Cretaceous time interval can be divided into four major periods or stages: (1) Triassic, (2) Sinemurian to Bathonian (?), (3) Callovian to Early Aptian and (4) Late Aptian to Albian. Individual paleobiogeographic maps have been constructed for the first three stages.

Triassic (Fig. 1)

In the present case this stage is represented only by the Anisian; however, its upper boundary may probably be placed at the end of the Triassic. The stable part of the European craton is dominated by the "Germanic" fauna. Most terranes belong to the "South Alpine" faunal belt but in Tisia (especially in its western part) some kind of "Germanic" influence was also recorded. This can be explained by the contemporaneous geographical vicinity to the Germanic Basin or, at least partly, by environmental effects. Various parts of the composite terranes (e.g. the Upper Codru nappes of Tisia, Aggtelekia, Bakonyia, the higher nappes of Austroalpinia with Hallstatt facies) represent the "Dinaric" faunal belt, what means that they belonged to the outermost, pelagic shelf margins of the Tethys.

Sinemurian to Bathonian (?) (Fig. 2)

The degree of NW-European/Mediterranean provincialism reached a maximum in this time interval (Vörös 1993). Accordingly, a clear paleobiogeographical differentiation between the major terranes of the Alpine-Carpathian region can be recorded. Tisia (including at least the Mecsek, Villány-Bihor and Lower Codru Zones) and Dacia definitely belonged to the NW-European province. On the other hand, all parts of Alcapa, together with Adria, clearly represent the Mediterranean province. This strikingly anomalous present-day arrangement requires a plate-tectonic explanation.

The upper limit of this paleobiogeographic period is uncertain. In the Bajocian, the NW-European/Mediterranean provincial boundaries are still clear. A gradual change can be recorded during the Bathonian (e.g. in Tisia); therefore, this may be taken as a transitional age.

Callovian to Early Aptian (Fig. 3)

In this period NW-European/Mediterranean provincialism still existed but the paleobiogeographical affiliation of the Alpine–Carpathian terranes changed

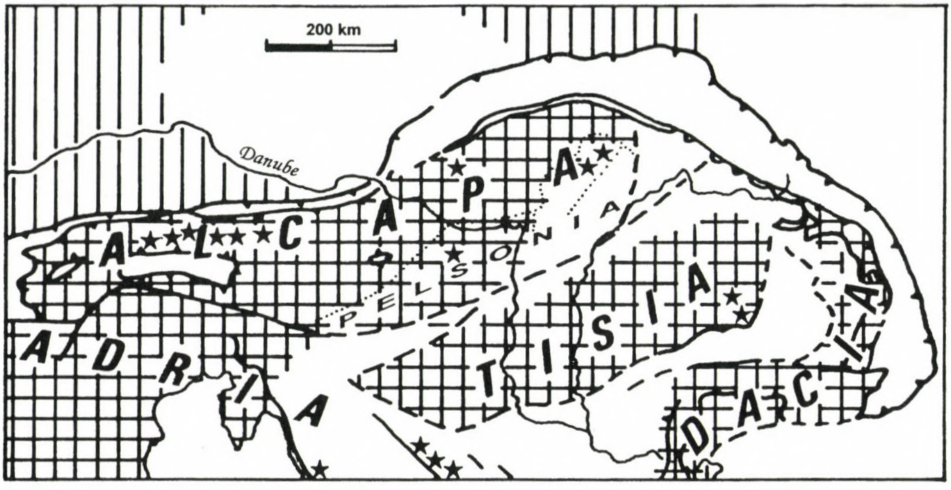

Fig. 1
Paleobiogeographical map for the first stage (Triassic, Anisian) in the Alpine–Carpathian region. Asterisks: faunas of "Dinaric" type; cross-hatching: faunas of "South Alpine" type; vertical hatching: faunas of "Germanic" type

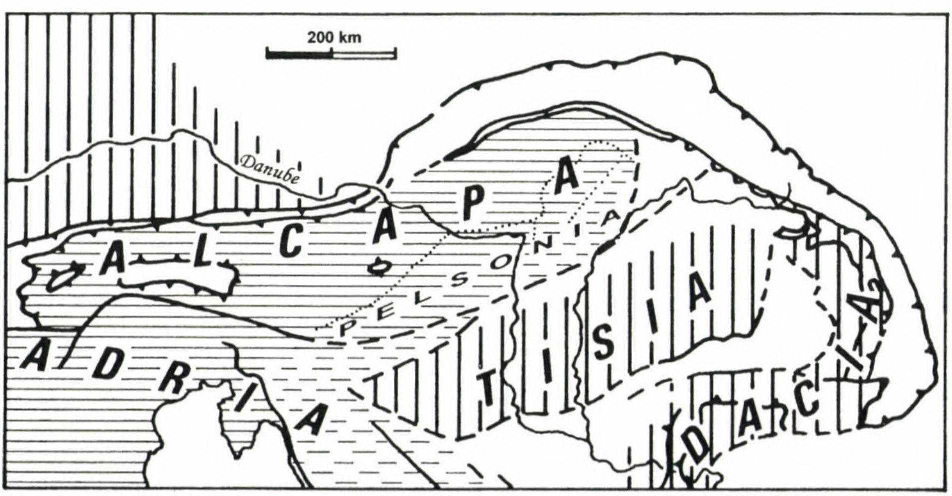

Fig. 2
Paleobiogeographical map for the second stage (Sinemurian to Bathonian ?). Horizontal hatching: Mediterranean province; vertical hatching: NW-European province; dashed: inferred affinity

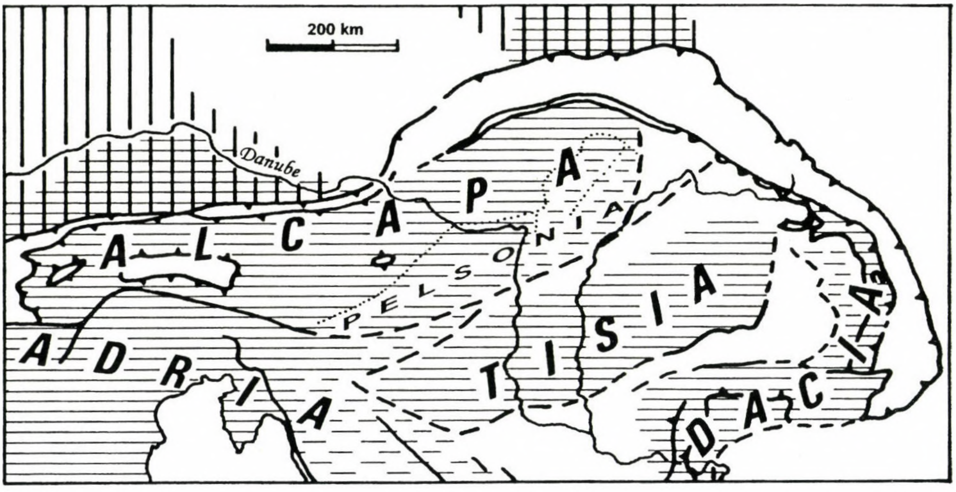

Fig. 3
Paleobiogeographical map for the third stage (Callovian to Lower Aptian). Horizontal hatching: Mediterranean province; vertical hatching: NW-European province; cross-hatching: transitional zones; dashed: inferred affinity

considerably (except for Alcapa and Adria, which remained in the Mediterranean province). Tisia and Dacia gradually came under Mediterranean influence. This change began in the Bathonian, as is well demonstrated in Tisia. Here the appearance of definitely Mediterranean brachiopods is recorded in the deep-water Bathonian of the Mecsek, while the shallow marine Villány region remained under the influence of the NW-European fauna. In this case we must count on the interplay between the general pattern of brachiopod distribution and the environmental control (Vörös 2001). The change continued through time and was apparently accomplished only by the Tithonian. Simultaneously, an increased Mediterranean faunal influence manifested itself in the southern zones of the European shelf (Southern Poland, Southern Germany, Southern France) (Vörös 1993; Cecca et al. 1993).

Thereafter, until the Aptian, the general paleobiogeographic pattern did not show any appreciable change.

Late Aptian to Albian

At the lower boundary of this period, at the middle of the Albian, a significant change in the brachiopod faunas was recorded in Bakonyia, characterized by the disappearance of the Mediterranean taxa and the massive occurrence of NW-European faunal elements. The southward expansion of the NW-European faunal province is confirmed by the character of the Albian ammonoid fauna in the Villány Zone of Tisia.

Discussion: microplates and oceanic basins

The four major paleobiogeographic periods or stages outlined above can be well correlated with the plate tectonic changes in the western part of the Mesozoic Tethys [the definitions of the terms and discussion on the concepts "microplate" "microcontinent" and "terrane" are given in length in Vörös (1993, 2000)].

The Triassic (and later Mesozoic) paleogeographic positions of the Alpine–Carpathian–Dinaridic terranes and their components have been widely and strongly debated. It was the subject of several international paleogeographical projects (Rakús et al. 1990; Dercourt et al. 1993), research papers (Kozur 1991; Kovács 1992; Haas et al. 1995) and paleobiogeographical studies (Vörös 1993). Some of these plate tectonic syntheses suggested that in the Early Mesozoic these microcontinents were closely linked to the European and/or the African craton. Others favored the concept of intra-Tethyan microcontinents isolated from the European and African shelves, named "Kreios" (Tollmann 1976), "Mediterranean Microcontinent" (Vörös 1977, 1980), "Adriatic Microplate" (Ager 1980), "Apulia-Anatolide" Block of the Cimmerian Continent (Sengör 1984). Instead of discussing these opposing views at length, compilations based on recently published reconstructions by Stampfli and Marchant (1997), Stampfli and Mosar (1999) and Vörös (2000) are adopted here to illustrate the supposed paleogeographic positions of the microplates carrying the terranes of the Alpine-Carpathian region. The extensions of the paleobiogeographic provinces are also presented on these paleogeographic sketch maps [these maps are admittedly oversimplified, and in the Dinaridic parts reflect Stampfli's opinions, which may conflict with the results of the local experts, e.g. Dimitrijevic (1982), Karamata and Krstic (1996), and Papanikolau (1997). Nevertheless, they can be used for illustrating the main paleobiogeographic relationships.]

In the first stage (Triassic, Anisian) the western end of the Tethys was surrounded by a more or less continuous shelf margin, where the differentiation of true faunal provinces was not possible (Vörös 1988). We may envisage two oceanic arms penetrating westward into Pangaea with the continental block of Adria between them (Fig. 4). The southern arm was suggested by Brandner (1984) and Sengör (1984); this seaway might have served as a corridor for the very early (Permian) invasion of oceanic plankton into Sicily (Catalano et al. 1996) and its remnant forms the basement of the present-day East Mediterranean Basin. The northern arm may correspond to the Meliata Ocean of Kovács (1982, 1992), known in the form of fragments in the Inner Dinarides, Inner West Carpathians and Northern Calcareous Alps. The subsided, open shelf margin of this Meliata Basin might have been the homeland of the "Dinaric" fauna in the Anisian. The next, concentric faunal belt, here named "South Alpine", was established on the wide, shallower, partly restricted shelves around the Meliata Basin. Toward the European epicontinental areas followed the third, "Germanic" faunal belt. The

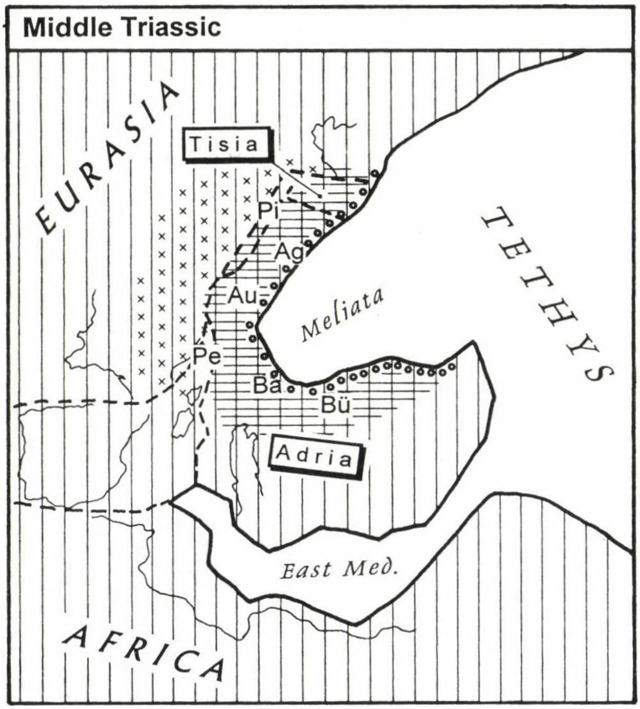

Fig. 4
Reconstruction of the western Tethyan area for Middle Triassic times, showing the supposed positions of the terranes and partial terranes discussed in the paper in relation to the oceanic basins. Compiled after Stampfli and Marchant (1997), Stampfli and Mosar (1999) and Vörös (2000). Blank: oceans; Vertical hatching: main continents and shelves; small circles: Dinaric fauna; horizontal hatching: South Alpine fauna; crosses: Germanic fauna. Elements of the Alcapa composite terrane: Ag: Aggtelekia; Au: Austroalpinia (composite); Ba: Bakonyia; Bü: Bükkia; Pi: Pieninia (Czorsztyn); Pe: Penninia (Briançonnais–Central Penninic)

Germanic fauna represents an impoverished variant of the "South Alpine" fauna; it became an endemic faunal province only after the Anisian.

The second stage (Sinemurian to Bathonian?) was the time of opening of the Ligurian–Penninic oceanic belt (Fig. 5). Due to eastward movement of the African Plate, an arm of the Central Atlantic rift system affected the region and ran into the Meliata Ocean. By this rifting the Austroalpine-Adriatic block (Mediterranean Microcontinent of Vörös 1993), moving as part of the African Plate, was detached from Europe. The newly formed oceanic belt, the Penninic (or Váhic, as it is called in the West Carpathian system) formed a true barrier between the faunal provinces of stable Europe (Neumayria) and the drifted microcontinent (Mediterranean province) (Géczy 1973, 1984a; Vörös 1977, 1993). According to the paleobiogeographic studies, Adria and all partial terranes of Pelsonia and Austroalpinia belonged to the Mediterranean Microcontinent, whereas Tisia, Pieninia and Penninia remained attached to the European margin.

The beginning of the third stage (Callovian to Early Aptian) was marked by several important paleogeographical, paleoclimatological and paleoecological changes in the western Tethys. The most remarkable plate tectonic change was connected to further rotation of the African Plate: the northern arm of the Central Atlantic rift system jumped to a new, northwestern position. The spreading of a new Valais–Magura oceanic belt had been started. By this movement, the next

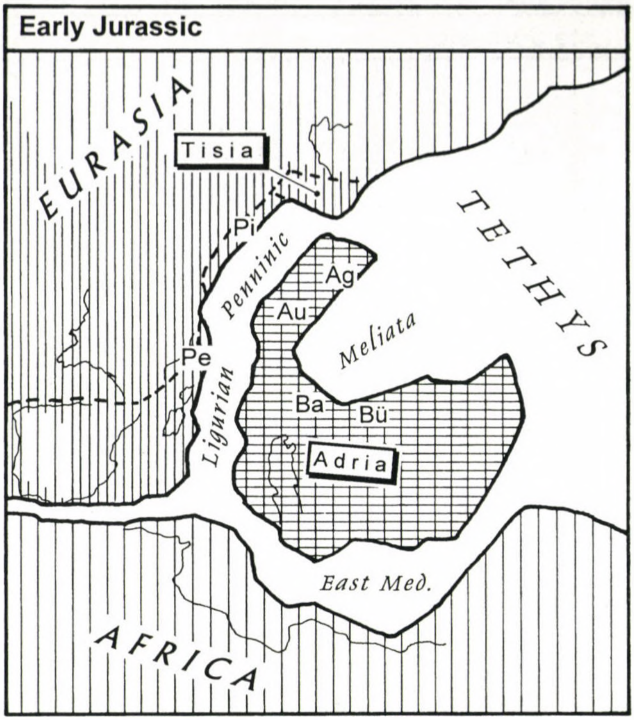

Fig. 5
Reconstruction of the western Tethyan area for Early Jurassic times, showing the supposed positions of the terranes and partial terranes discussed in the paper in relation to the oceanic basins. Compiled after Stampfli and Marchant (1997), Stampfli and Mosar (1999) and Vörös (2000). Blank: oceans; vertical hatching: main continents and shelves; tight vertical hatching: NW-European province; horizontal hatching: Mediterranean province; Elements of the Alcapa composite terrane: Ag: Aggtelekia; Au: Austroalpinia (composite); Ba: Bakonyia; Bü: Bükkia; Pi: Pieninia (Czorsztyn); Pe: Penninia (Briançon-nais–Central Penninic)

segment of the European shelf (including the Tisia and the central Pieninic, or Czorsztyn Microcontinents) drifted apart. The central Penninic (Hochstegen) Microcontinent is envisaged as the northern spur of the Iberian Microcontinent (Fig. 6). The previous, definitely NW-European paleobiogeographic character of these isolated microcontinents gradually changed: they came under the influence of the Mediterranean province. At the same time, they may have served as "stepping stones" promoting the migration of Mediterranean faunal elements, which invaded the southern belt of the European shelf in the Late Jurassic (Vörös 1993). The consumption of the Meliata Ocean apparently left no signs in the paleobiogeographic record.

The fourth stage (Late Aptian to Albian) is the time of the first significant continental collisions in the Alpine–Carpathian segment of the Tethys. The oceanic Tethys became much narrower; some oceanic basins, previously barriers (e.g. Penninic), would have closed. Continuous shelf areas had formed, connecting the microcontinents to the European shelf. Through these shallow marine corridors the NW-European faunal elements migrated to the impoverished areas of the Mediterranean Microcontinent.

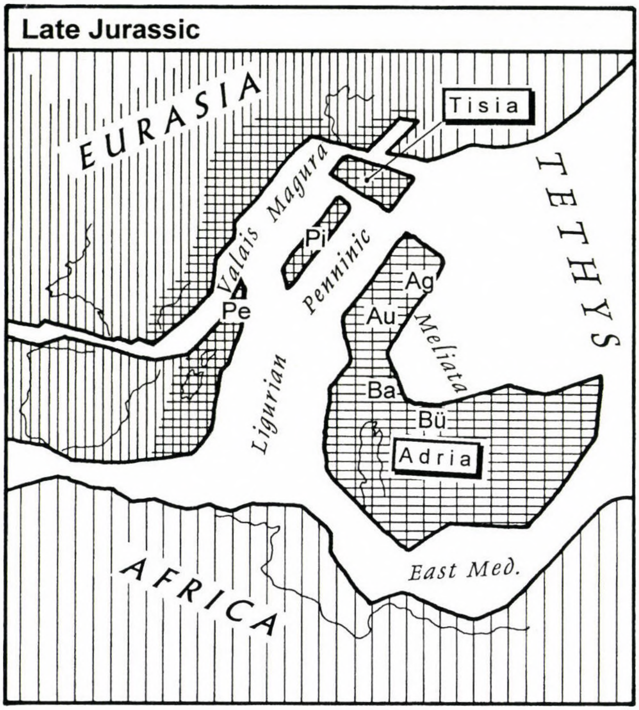

Fig. 6
Reconstruction of the western Tethyan area for Late Jurassic times, showing the supposed positions of the terranes and partial terranes discussed in the paper in relation to the oceanic basins. Compiled after Stampfli and Marchant (1997), Stampfli and Mosar (1999) and Vörös (2000). Blank: oceans; vertical hatching: main continents and shelves; tight vertical hatching: NW-European province; horizontal hatching: Mediterranean province; Elements of the Alcapa composite terrane: Ag: Aggtelekia; Au: Austroalpinia (composite); Ba: Bakonyia; Bü: Bükkia; Pi: Pieninia (Czorsztyn); Pe: Penninia (Briançon-nais–Central Penninic)

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Deep tectonics under the thick limestone in NW Transdanubia by means of magnetotellurics

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The paper describes the main steps of magnetotelluric (MT) data processing and interpretation in the area of the Transdanubian Conductivity Anomaly (TCA). The regional strike of the TCA was found in a NE–SW direction on the basis of the MT phase maps and from the rose diagram of the extreme resistivity sounding curves. Inverting the MT data along ELGI's NW–SE profiles by different 2D programs, strong longitudinal strike slip zones unknown until now could be determined between the Rába and Balaton Lines, which might be in close connection with earthquake activity of the area.

Keywords: magnetotellurics, conducting dike, Paleozoic graphitic layer, seismicity, strike slips, deep fractures

Introduction

During the last 30 years about 300 deep magnetotelluric (MT) measurements have been carried out by the Geodetic and Geophysical Research Institute (GGRI) and the Eötvös Geophysical Institute (ELGI) in the area of the Transdanubian Conductivity Anomaly (TCA) between two great tectonic lines, i.e. the Balaton and Rába Lines (NW Hungary). The results of these measurements – including the methodological problems to be solved – have recently been summarized in a review paper (Ádám 2001). This short paper focuses on the tectonic conclusions of the MT soundings.

A successive approximation toward the tectonics represented by the TCA

After the discovery of the Transdanubian Conductivity Anomaly (TCA) in the sixties (Ádám and Verő 1964; Takács 1968), a regular measuring campaign began in the TCA area at the end of the same decade, both by GGRI and ELGI. The work was supported by the Central Geological Office (KFH). Parallel to the deep magnetotelluric measurements, GGRI carried out precise magnetovariational measurements as well, using an Askania variograph. While the inversion of the magnetotelluric measurement supplies the geoelectric layer sequence with resistivity and thickness of the (quasi) layers, from the relation between the time variation of the three geomagnetic components, primarily the lateral extent of

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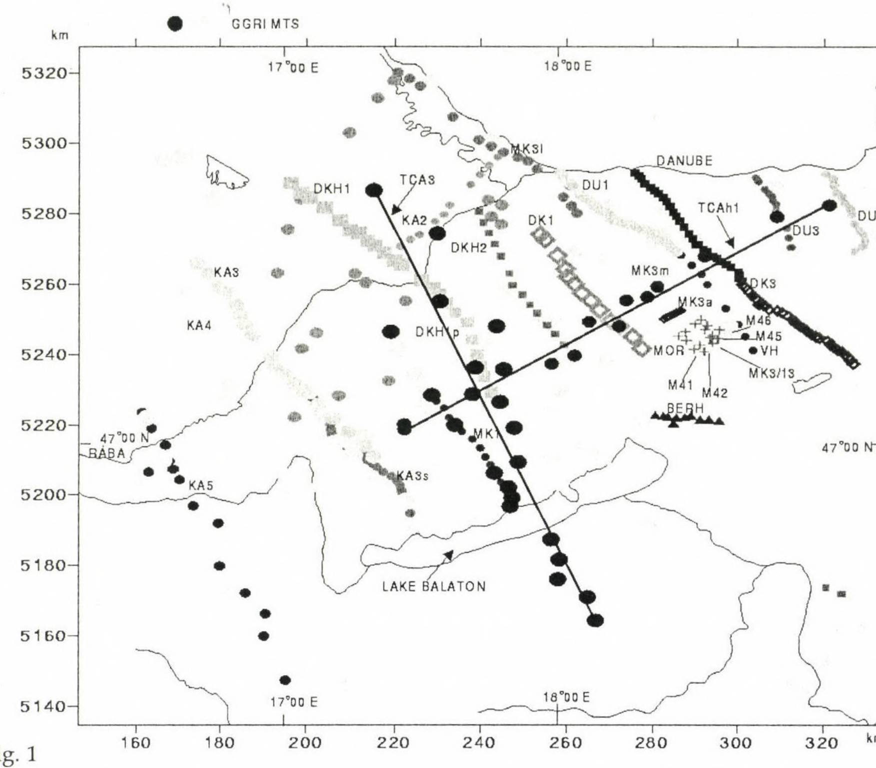

Fig. 1
Magnetotelluric measuring sites in the area of the Transdanubian Conductivity Anomaly (TCA) with profiles TCAh1 and TCA3 (GGRI and ELGI measurements)

conductors can be distinguished. Nevertheless, both techniques play their own role: they complete each other and control the results of the other.

In the present synthesis the data of both institutes were used; as a consequence of this close co-operation, a large data set was at our disposal (see the map of the measurement sites in Fig. 1).

During the interpretation of the magnetotelluric measurement curves, the first step is a rigorous study of the EM distortions. In our case the galvanic distortion was studied after separation of the extreme measurement curves (Rhomin and Rhomax), on the basis of the so-called S-effect, i.e. the change of the apparent depth of a conductor due to the near-surface inhomogeneities, such as changes of the resistivity of the surface rocks. A statistical treatment (Ádám 1981) hinted at a curious situation: to begin with the Rhomax curves are distorted by S-effect. This distortion practically does not appear in the Rhomin curves (Ádám 1981). It was therefore concluded that the Rhomax curves are of B polarization (TM mode) (in this case the magnetic component (B) lies in the strike direction).

On the basis of the impedance phase maps from Rhomax and Rhomin curves it was concluded that the regional geoelectric strike direction corresponds to that of the longitudinal fractures/strike slips in the Transdanubian Central Range, i.e.

NE-SW. As the peak in the rose diagram of the Rhomax directions (Fig. 2) is perpendicular to this regional strike, the B polarization character of the Rhomax curve is also confirmed by this fact.

Geoelectric models have been examined for this special distortion and found by forward modeling in a 2D conducting dike system, where the dikes are isolated from each other. In the case of the TCA the strike of these 2D dikes should be directed NE-SW.

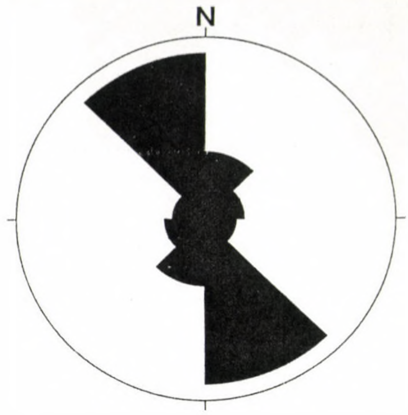

Fig. 2
Rose diagram of Rmax directions (GGRI data)

Further proof for the NE-SW-directed 2D dike system was given by the peculiarity of the Wiese arrows. They show a regional N-S direction in the Pannonian Basin; nevertheless, in the TCA area, in two NE-SW zones, the arrows rotate toward a southwesterly direction (Fig. 3). Similar phenomena have been described by Arora and Ádám (1992) above other conductors in the world (see e.g. the Carpathian Anomaly by Jankowski et al. 1984).

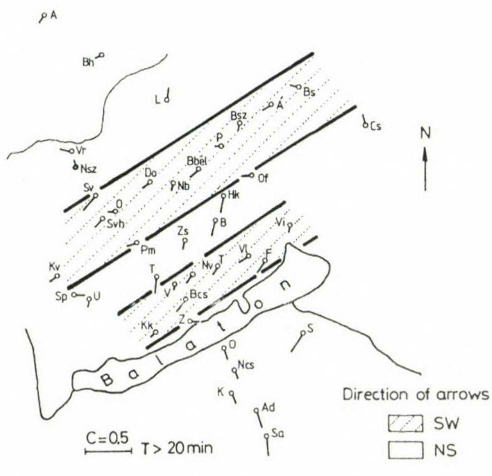

Fig. 3
Long-period Wiese-arrows in the area of the TCA, measured by GGRI (Wallner 1977)

The apparent resistivity pseudo-section along the TCA3 section (Fig. 1) – perpendicular to the conductive zones as indicated by Wiese arrows – clearly shows 3 conductors among them. Two correspond to the Wiese arrow anomalies: the third-one appears above the Rába tectonic line (Fig. 4). This latter one is shown by Madarasi's (2000) map as well, which was computed from the telluric conductance map stripped from the conductance of the sediment.

2D inversion was calculated along ELGI's very dense NW-SE directed MT sections using RRI inversion (Smith and Booker 1991), but part of them were controlled by Occam (Constable et al. 1987) and Win Glink 2D inversion methods as well. All measurement curves involved in the inversion were computed in the

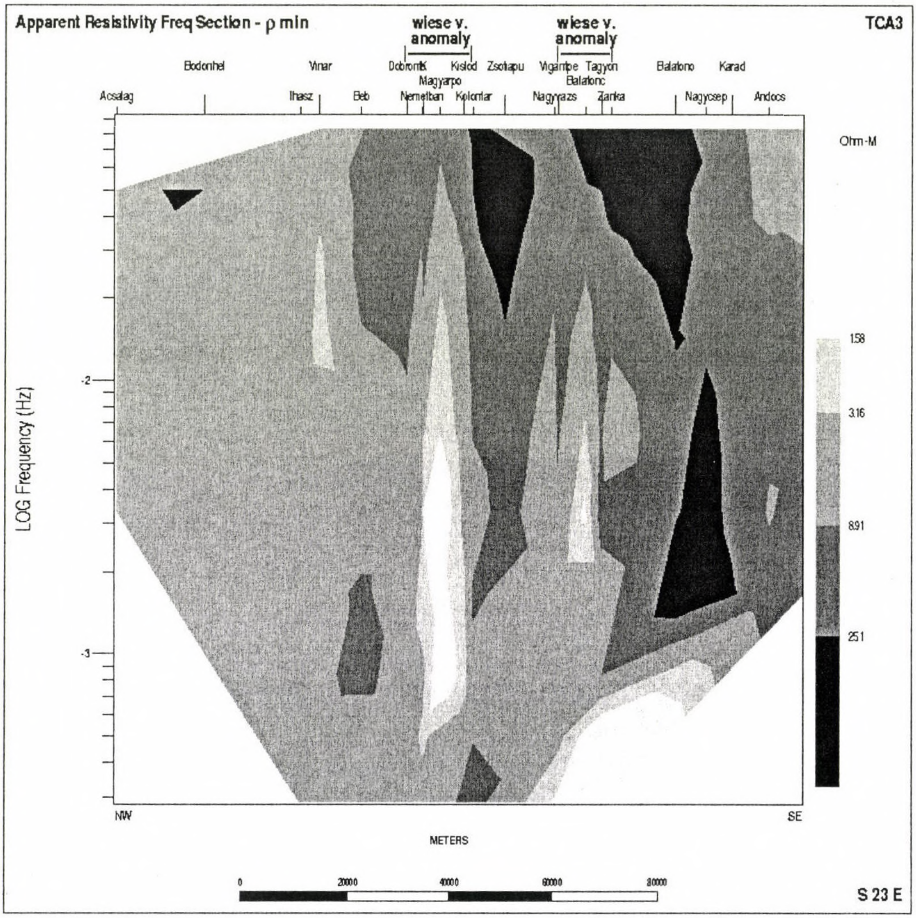

Fig. 4
R_{min} pseudosection perpendicular to the conductive zones (TCA3) indicated by Wiese arrows in Fig. 3

regional strike (NE–SW) direction (TE mode, E polarization) and perpendicularly to it in a NW–SE direction (TM mode, B polarization).

The measurement curves in the regional strike direction generally correspond to the Rhomin curves and are therefore the least distorted.

As result of the inversions three significant NE-SW directed 2D (quasi linear) conducting zones (strike slip/fractures) (Fig. 5) could be distinguished in close connection with the Wiese arrow anomalies (Fig. 3) and the pseudo-section in Fig. 4. The average depth of the conductor is about 4km.

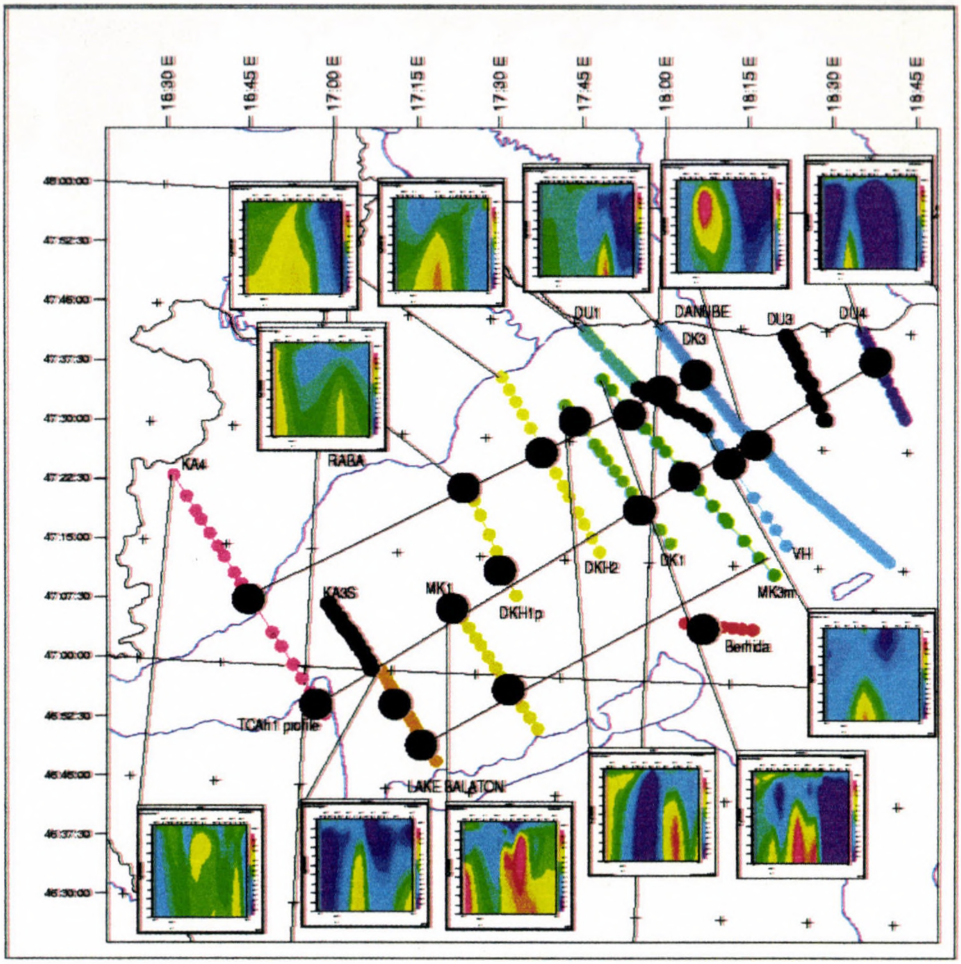

Fig. 5

A map on the conducting dikes (large dots) on the ELGI profiles and the vertical resistivity sections obtained by RRI inversions

In some cases the conductance in the fracture reaches more than 10000 Siemens. Figure 5 shows the position of the conductor with large dots and the vertical resistivity sections obtained by the inversions.

These tectonic zones play an important role in the seismic activity of the area because most of the epicenters lie in their neighborhood (about the supposed relation between the conductors and the earthquake foci see Ádám (1994) and Ádám and Zalai (2000)).

A map was constructed on the depth of an apparent conducting layer by 1D inversion of the least distorted, supposedly E-polarized Rhomin curves. We are

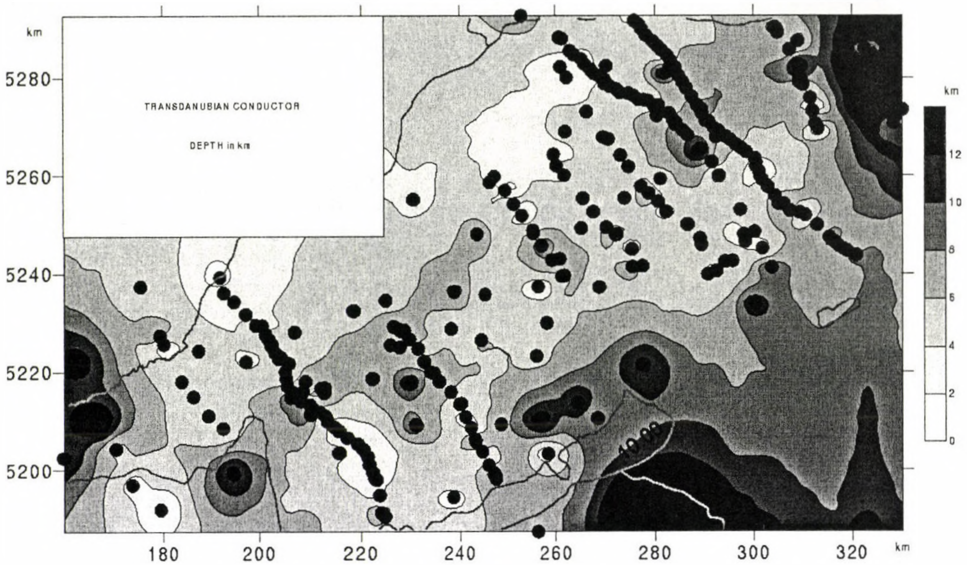

Fig. 6
The depth of the virtual conductive layer computed on the basis of the 1D inversion of Rmin curves (quasi E-pol.)

aware that this map is a strong approximation; nevertheless, it has some common features with the known (or supposed) geology, e.g. Balla and Dudko's (1989) syncline axis coincides with the deepest part of the conducting layer (Fig. 6).

A map could also be constructed by the inversion of the quasi B polarized Rhomax curves on the thickness of the sediment, which has many realistic features as compared to the distorted Rhomax curves (Fig. 7).

Conclusions

The main results of this investigation are the demonstration of new, up to now unknown, important tectonic zones under the thick Mesozoic layers, which can be connected with seismic activity.

Nevertheless, a question arises when comparing the tectonics determined by the three two-dimensional dike systems in longitudinal fractures/strike slips (Fig. 5) and the map of a conducting – supposedly Paleozoic – layer (syncline) (Fig. 6), with peaks on it corresponding to the dikes (Fig. 5). One must take a similar "blending process" between the anomalies into account, due to the induction (side) effect of the E polarized MT curves, which is, for instance, demonstrated in Ádám and Zalai's (2000) paper in their Fig. 7, for the case of four two-dimensional dikes. The question – at the present stage of the investigation – cannot be decided. We are in favor of the existence of a Paleozoic graphitic layer

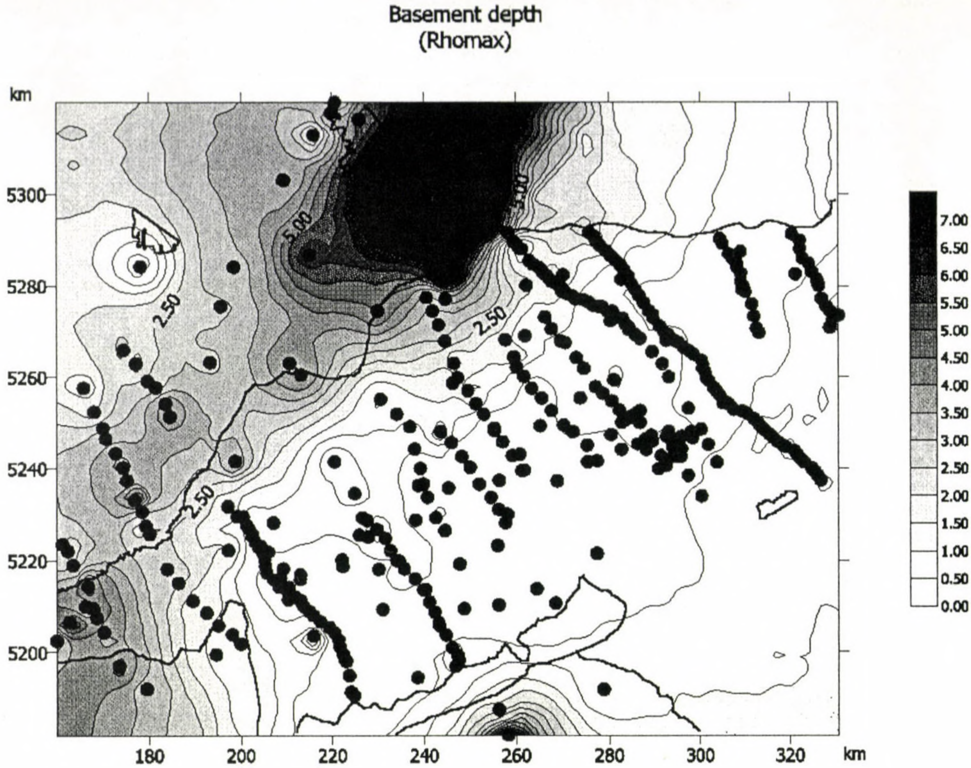

Fig. 7
The thickness of the sediment in the measuring area computed from the inversion of rmax curves (quasi B-pol.) in km

(syncline) from which fractures emerge, containing graphite (+ fluid) which was injected into them.

The problem will only be solved by drilling.

Acknowledgements

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An attempt to map the depth of the electrical asthenosphere by deep magnetotelluric measurements in the Pannonian Basin (Hungary)

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It was stated in the early sixties, on the basis of deep magnetotelluric measurements, that the conductive asthenosphere wells up to a depth of 60 km in the central part of the Pannonian Basin (Hungary), in close connection with the high heat flow of the area (~100 mW/m^2), that it deepens toward the rim of the basin and that its depth reaches 100–200 km beneath the area of the East European Platform (EEP). Beneath deep extensional basins (e.g. the 7 km-deep Békés Graben) the asthenosphere can even reach depths of 40–50 km. This must certainly have been a plume during the Miocene synrift era, causing a strong ultramafic intrusion in the crust.

The authors have attempted to construct a map of the depth of the conductive asthenosphere in the Pannonian Basin with correction of the EM distortions (static shift, 3D inhomogeneity effect by decomposition). They point out the difficulty of the corrections and summarize the main characteristic features of the lithosphere-asthenosphere boundary in the Pannonian Basin.

Key words: conducting asthenosphere, magnetotelluric decomposition, static shift

Introduction

Historic background

The first indication of the relatively shallow depth of the electrically conductive asthenosphere in the Pannonian Basin (its central part belongs to Hungary) was obtained by magnetotellurics in the early sixties (Ádám 1963, 1965). In the electromagnetic observatory near Nagycenk, named by I. Széchenyi, the depth of the asthenosphere was estimated to be about 60 km (later, using a more exact numerical inversion technique, this value was modified to about 80 km). Comparing this shallow depth with asthenospheric depths of 100–200 km measured in the East European Platform (EEP) (Ádám 1970), it became evident that the asthenosphere is an isothermal surface (~1300C°), as Gutenberg initially supposed, and is in close connection with the regional surface heat flow in a "steady state" condition (Ádám 1978) (Fig. 1). In the seventies both seismology (Bisztricsány 1974) and active (reflection) seismic measurements (Posgay 1975), and later seismic tomography (Spackman 1990) confirmed the magnetotelluric results. They found that the LVL layer corresponding to the conductive

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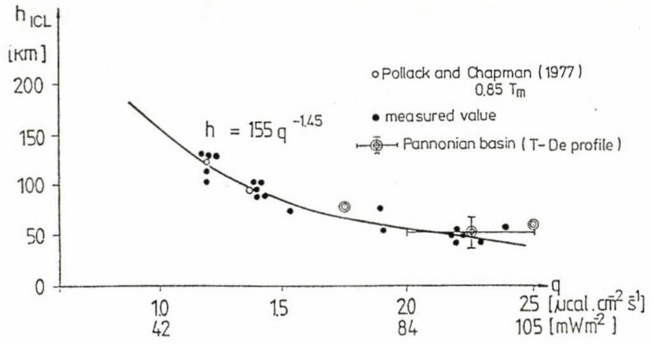


Fig. 1
Empirical relationship between regional surface heat flow and depth of the asthenosphere (in the formula the old HFU has been used) (Ádám 1978)

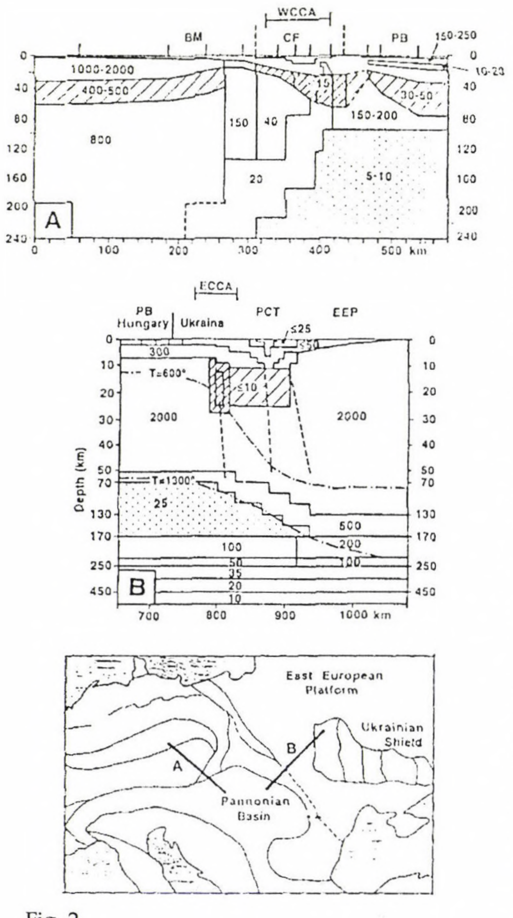


Fig. 2
Electrical resistivity models through the Carpathians from the Carpatho-Pannonian Basin (PB) toward the Bohemian Massif (BM) (A. Cerv et al. 1987) and East European Platform (EEP) (B. Zhdanov et al. 1986) with Carpathian conductivity anomalies (WCCA, ECCA)

asthenosphere lies at a depth of 60–70 km and that it deepens toward Paleozoic (Variscan) Europe and the EEP. The same tendency is shown in Fig. 2 on the basis of 2D magnetotelluric inversion.

Difficulties in the construction of an asthenospheric map in the Pannonian Basin

The historic data mentioned in the previous section are not sufficient to map the depth of the asthenosphere in the Pannonian Basin. For this purpose a network of systematic magnetotelluric measurements is required. Since the basement of the Pannonian Basin is strongly fractured and folded (Fig. 3) it is economically impossible to build up a sufficiently dense measuring network to cover every (mainly 3D) basement element to be taken into account in a distortion study and accounting for it in the data correction. Although there is some general 2D tendency in the deep structure of the Pannonian Basin (Ádám 1969) the distortion of the data is determined by 3D inhomogeneities. Therefore, for satisfactory data processing 3D inversion technique is needed, which is only now appearing, i.e. it is not available at present for a large regional

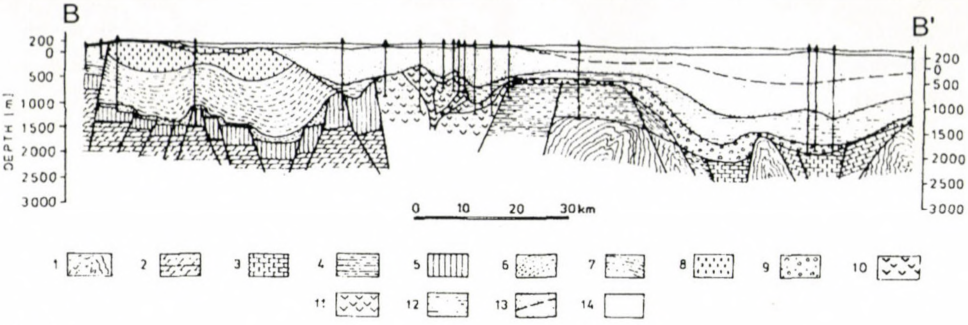


Fig. 3
Geologic section through the Great Hungarian Plain (after Horváth and Berckhemer, 1982). 1. Metamorphosed Paleozoic schist; 2. Upper Triassic carbonate and sandstone; 3. Lower Jurassic sandstone and marl; 4. Cretaceous complex; 5, 6, 7, 8. Paleogene basinal complex (Upper Eocene, Lattorfian, Rupelian, Egerian–Eggenburgian, resp.); 9. Ottnangian gravel and sand passing upwards into Carpathian–Badenian clay, clay marl and limestone; 10. Sarmatian andesite flows; 11. Badenian(?) and Sarmatian rhyolitic tuffs and intercalated Sarmatian sediments; 12. Pannonian calcareous marl, clay and sand; 13. Lower and Upper Pliocene clay and sand complex; 14. Quaternary deposits

data system such as that of the Pannonian Basin (the Hungarian part of which is 96000 km^2). At present we see only the following way to obtain a more or less acceptable data set for mapping:

a) the measurement curves should be freed from distortions, primarily from static shift (frequency-independent galvanic effect and the so-called S effect due to basement relief (Berdichevsky and Dmitriev 1976) and from the near-surface small 3D effects by decomposition,

b) looking for support to separate the E and B polarized curves, if a 2D character of the structure is suspected,

c) calculation of the 1D layer sequence from the (E polarized?) curves (resistivity (Rho) and phase). The fit should be best in the case of the less distorted phase curves.

If there are a few measurements on a quasi-2D structure, a 2D joint inversion of these data should be applied (e.g. Pannonian Geotraverse, PGT1).

A severe problem emerges when, after all corrections, the 1D or 2D inversion gives two quite different layer sequences and as the asthenospheric depth values for the extreme Rho and phase curves. In this case the only apparent solution is to calculate the determinant of the impedance elements and invert it as proposed by Berdichevsky and Dmitriev (1976) for 3D structures, or very simply to take the geometric mean of the two asthenospheric depth values. All these simplifications weaken the resolving power of the MT method and its applicability to deep investigations, including the study of the peculiarities of the conductive asthenosphere.

We only succeeded in determining the conductance of the asthenosphere in a very few cases due to the lack of sufficiently long variations. Due to the equivalence of the conductance one can only obtain a rough approximation of the resistivity of the asthenosphere.

Data base

In the last 30–35 years, since we have been studying the asthenosphere in the Pannonian Basin (Ádám 1965), MT techniques have gone through a very extensive development, from analog to digital data acquisition, from scalar to tensorial data processing, and toward more sophisticated interpretation tools in noise suppression and in the decrease of distortion effects of inhomogeneities. We also continuously improved our technique. At present we use – for long period MT measurement – Polish-made digital MT equipment with very stable quartz variometers, the shortest period of which is 10 s. On our map we distinguish MT sites where this instrument has been applied (Fig. 4). The name of the measuring sites are abbreviated (the full names are given in Table I). The data have been processed by the program elaborated in its original form by Verő (1972). This is a robust processing program, which takes into account the parameters influencing the data quality by different weights. The extreme curves or those in any direction are computed by rotation of the coordinate system. In Fig. 5a–g a series of measurement curves are shown from the different major tectonic–geographic units (see Roman numbers I–VII) with their 1D layer sequence. 1D inversions are made with the GEOTOOL-system containing Fischer/Occam/Marquardt algorithms.

These curves are the raw, i.e. not yet corrected ones (see below). The uncertainty of the asthenospheric depth determination is illustrated by the 1D inversion of two MT measurements carried out in the area of the Békés Sub-basin near the town of Gyula, at a distance of a few km (Fig. 5b). The measurements were carried out with different instruments. Here the difference in the asthenospheric depth is about 10–15 km in the case of the Rhomax curves. Therefore the step in the iso-areas of the depth on the map has been chosen to 10 km.

The fundamental question is the horizontal spacing of sites: how much should the minimum distance between two measuring sites be in order to be able to follow the relief of the asthenosphere? In this respect we can start from our previous depth estimations. The average depth was calculated to be 60 km, and it changes by about 20 km in any direction from the center of the Basin at a distance of about 200–250 km, except southward toward the national frontier. This change is assumed to be continuous (the local surface heat flow does not express exactly the depth variation of the asthenosphere. One must suppose a regional adjustment in the corresponding deep temperature. Ádám (1978) refers to regional heat flow and not to the local one when looking for the relation shown in Fig. 1).

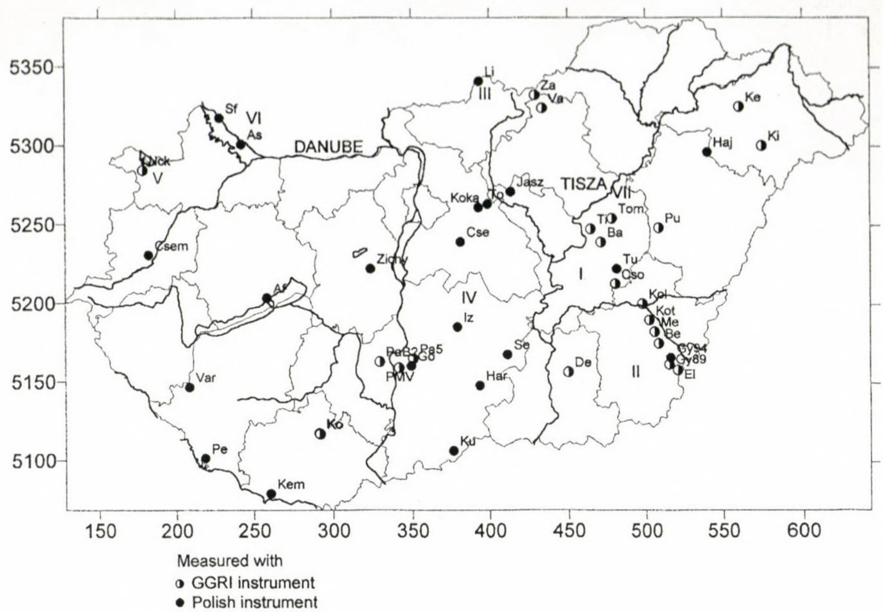


Fig. 4
MT sounding network in the Pannonian Basin. The station names are abbreviated (See Table I). The major geologic/geographical units are identified by Roman numbers

Table 1
MT stations and their abbreviations

Hungary				Transdanubia	
Name	Abbreviation	Name	Abbreviation	Name	Abbreviation
Bánhalma	Bá	Körösladány	Köl	Ásványráró	Ás
Békés	Bé	Köröstarcsa	Köt	Aszófõ*	Af
Csévharaszt	Csé	Kúnbaja	Kú	Csempeszkopács	Csem
Csodaballa	Cso	Litke	Li	Kemse	Kem
Derekegyháza	De	Mezõberény	Me	Komló	Ko14
Elek	El	Paks MV	PaMV	Komló	Ko57
Gombolyag	Go	Paks 5	Pa5	Nagycenk	Nck
Gyula 89	Gy89	Püspökladány	Pü	Paks B2	PaB2
Gyula 94	Gy94	Selymes	Se	Péterhida	Pé
Hajdúdorog	Haj	Tiszagyenda	Ti	Serfenyõ	Sef
Harkakötöny	Har	Tóalmási	Tó	Varászló	Var
Izsák	Iz	Tomajmonostora	Tom	Zichyújfalu**	Zichy
Jászágó	Jász	Túrkeve	Tú		
Kemecse	Ke	Váraszó	Vá		
Kisléta	Ki	Zabar	Za		
Kóka	Kóka				

*, ** not used due to unknown high static shift

1

2

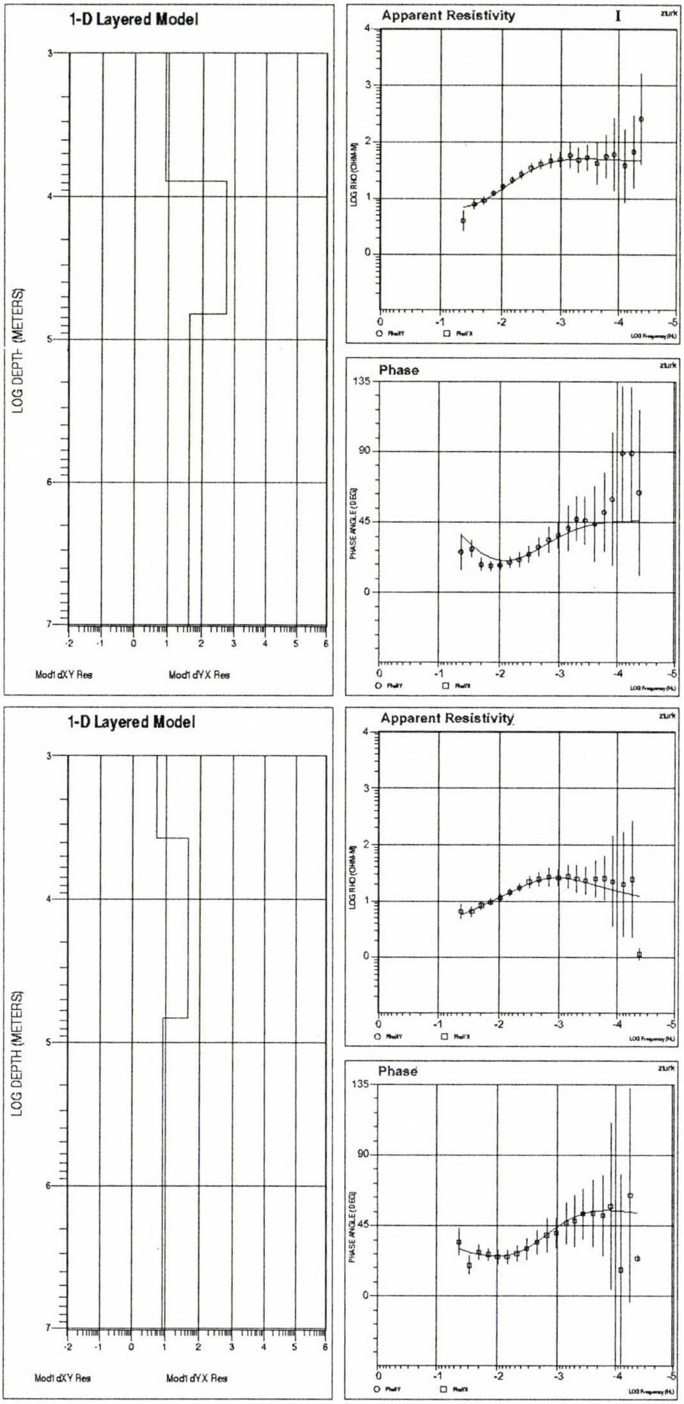


Fig. 5a–g

Rho and phase curves with their 1D inversion from selected regions of the Pannonian Basin (Rhomax (xy) 5a...g/1, Rhomin (yx) 5a...g/2) a) I: Great Hungarian Plain: Túrkeve (Polish instrument (i))

1

2

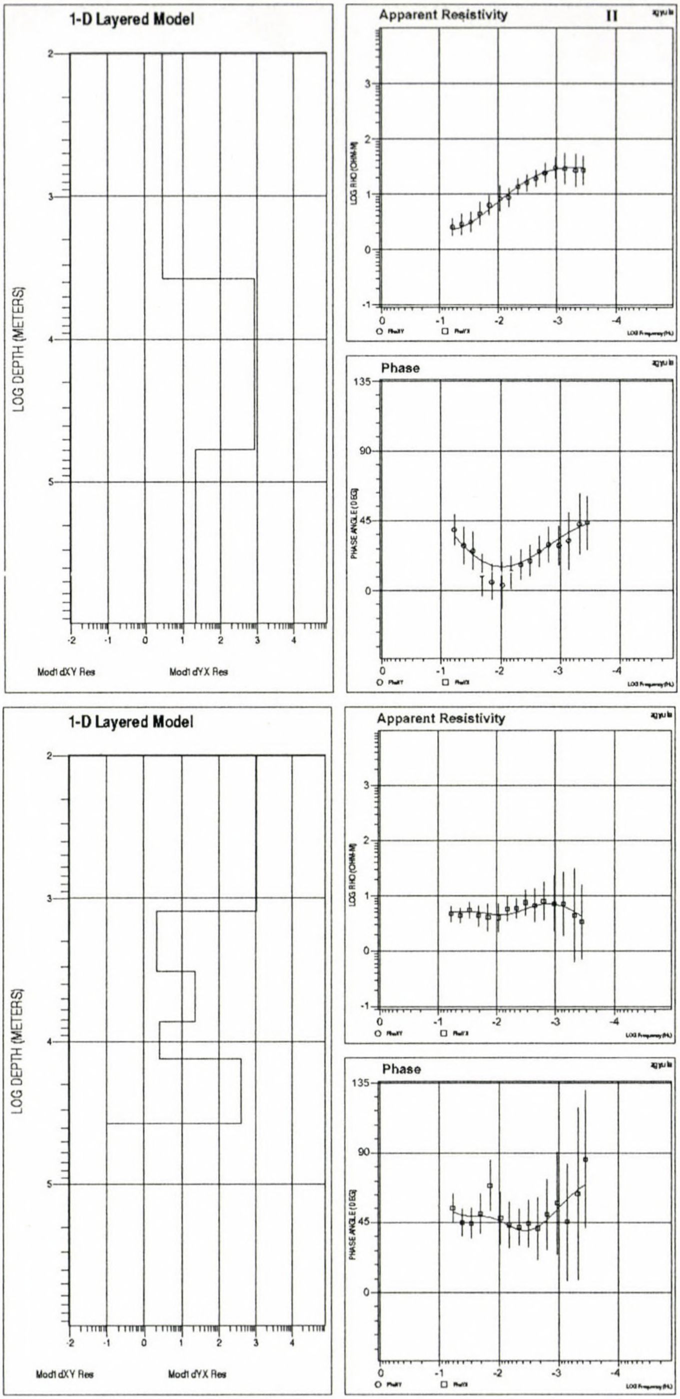


Fig. 5b1
b) II: Békés Sub-basin: Gyula (GGRI i for comparison)

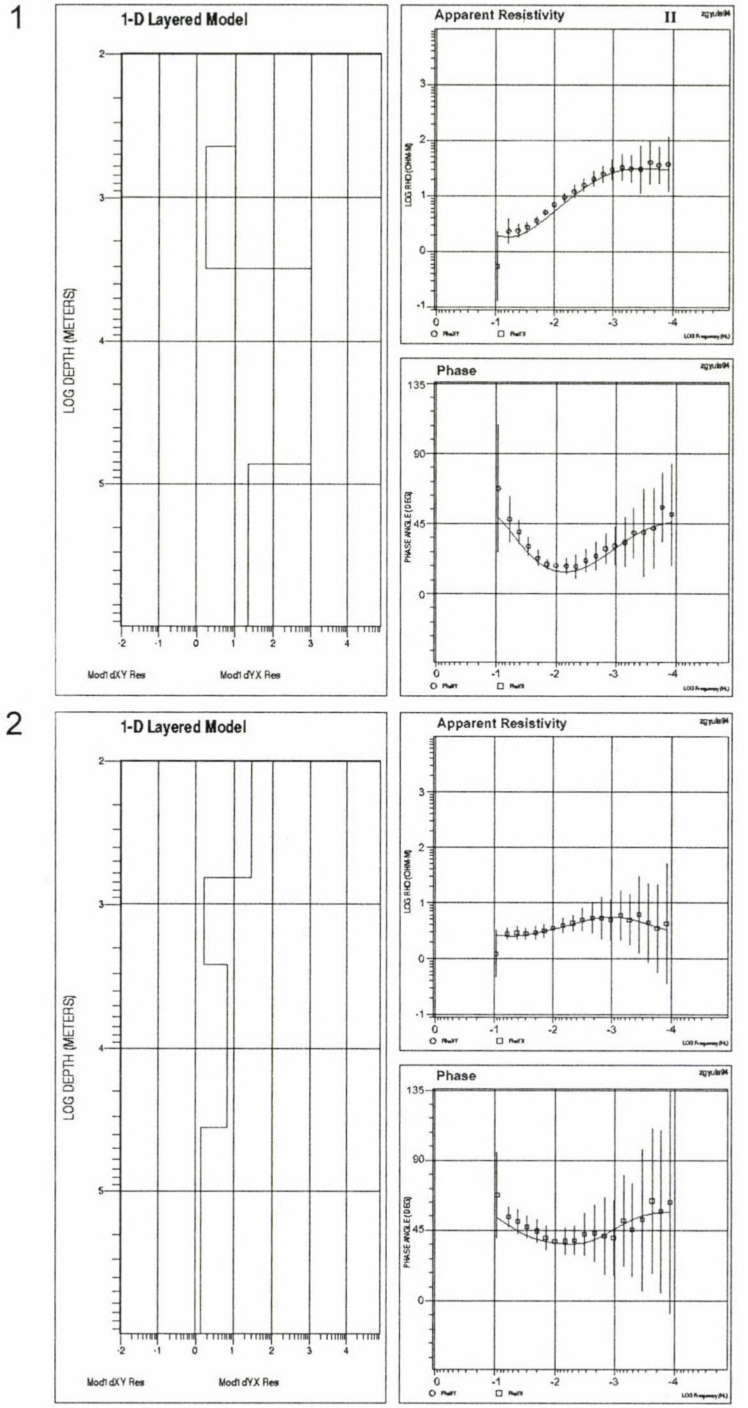


Fig. 5b2
II: Békés Sub-basin: Gyula (Polish i for comparison)

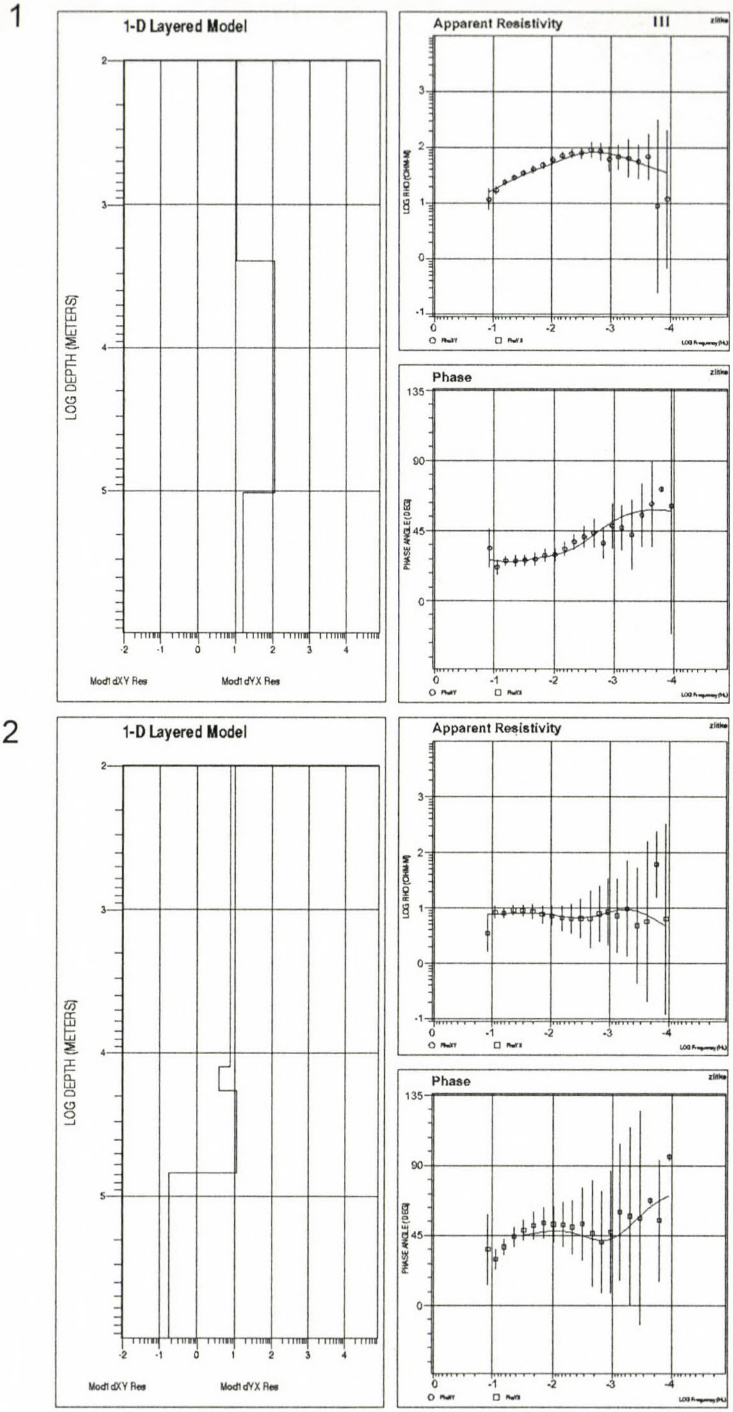


Fig. 5c
III: Gemerids: (?) Litke (Polish i)

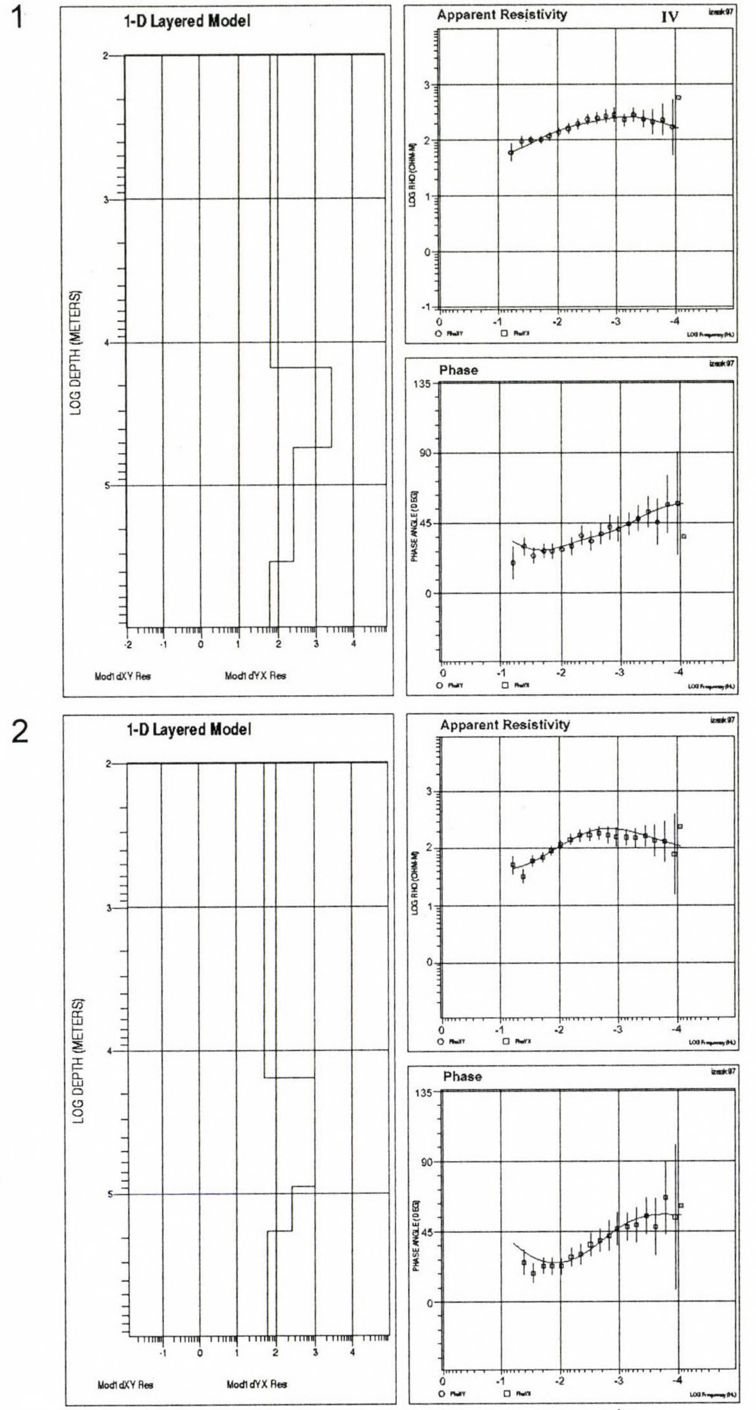


Fig. 5d
IV: Area between rivers Danube and Tisza: Izsák (Polish i)

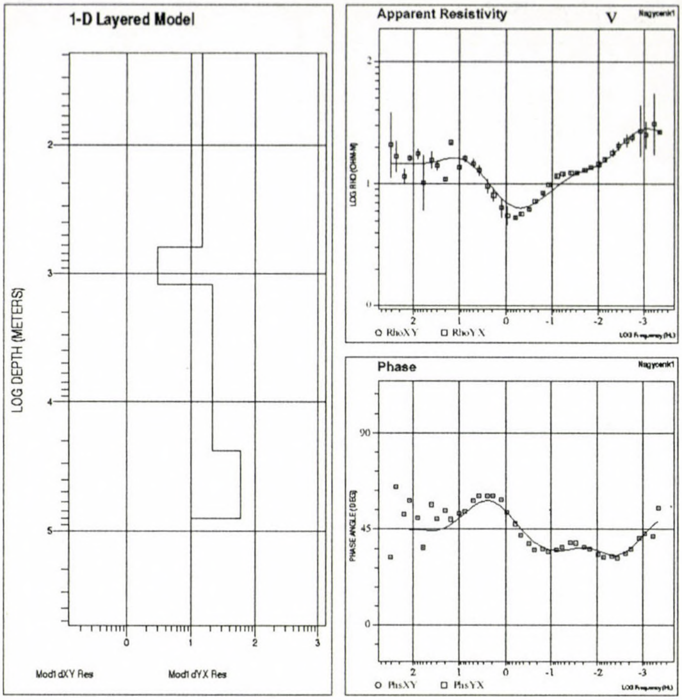


Fig. 5e
V: West Hungary: Nagycenk E–W (yx) curves (Phoenix i)

Horizontal spacing with a few km distance between MT sites would not be needed if the subsurface would be homogeneous or the relief of the asthenospheric layer would continuously (systematically) change, because some tenth-km change in the depth cannot be reliably followed by the 1D inversion, as has been demonstrated.

In the Pannonian Basin, nevertheless, the measurement curves are strongly distorted. The distortion effect can be determined, or at least estimated, mainly by statistical methods. Therefore, despite the above argumentation, the distance between the stations should be as small as economically supportable.

On the map of the asthenosphere there are lone outliers with higher or lower values than the ambient ones. These hint at unsuccessful correction, maybe due to the unknown effect of the 3D inhomogeneities in the sediment or in the basement. More measurements are needed in these areas to permit statistical treatment of the data.

Special cases are the extensional deep subbasins (grabens), e.g. the best-studied Békés Sub-basin (Fig. 6), where deep seismic measurements by Posgay et al. (1995) detected the seismic asthenosphere at a depth as shallow as 40 km. This

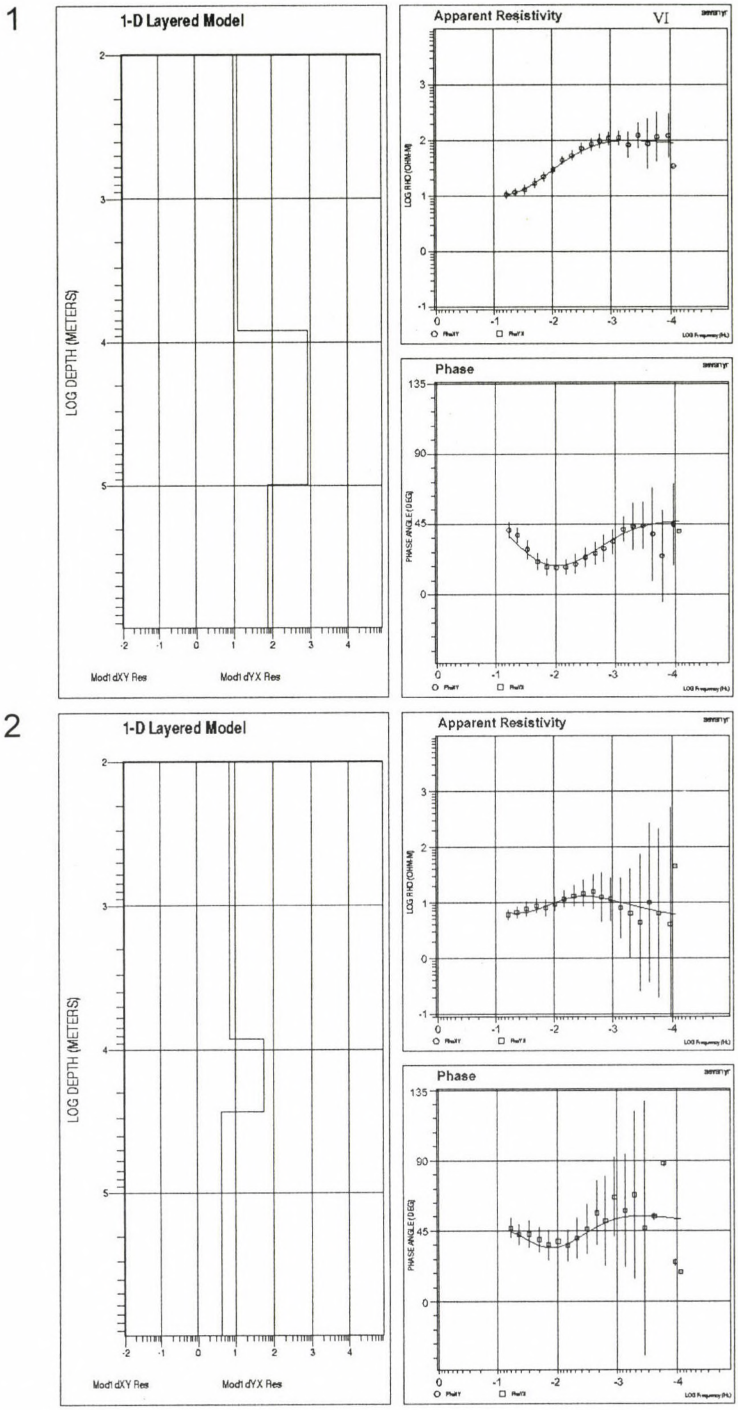
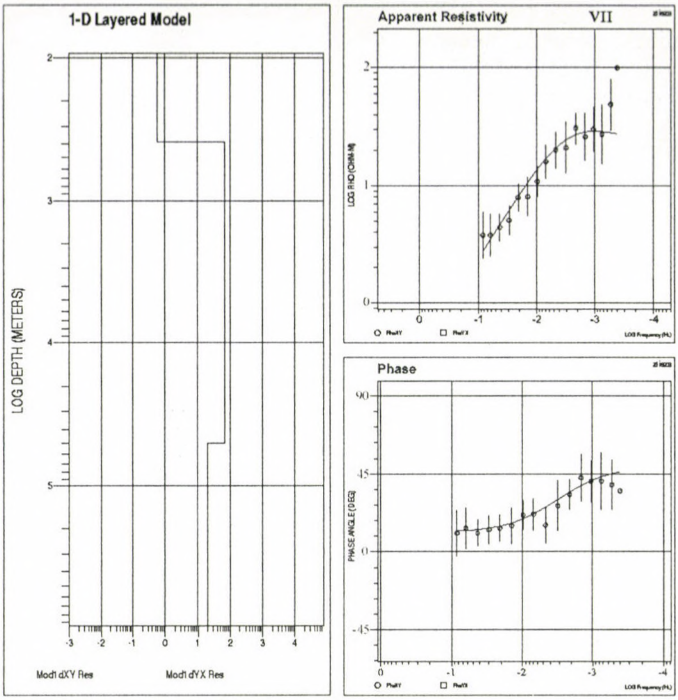


Fig. 5f
VI: Little Hungarian Plain (Danube Basin): Ásványráró (Polish i)

1



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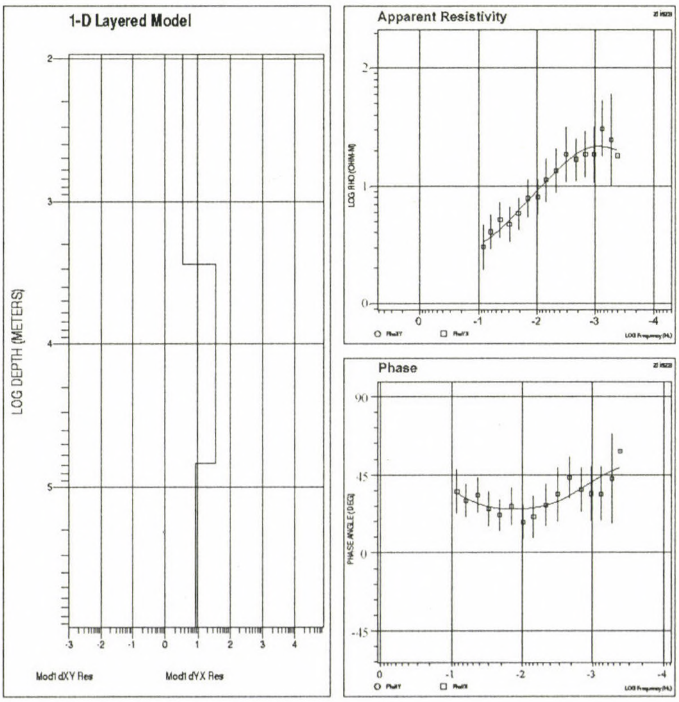
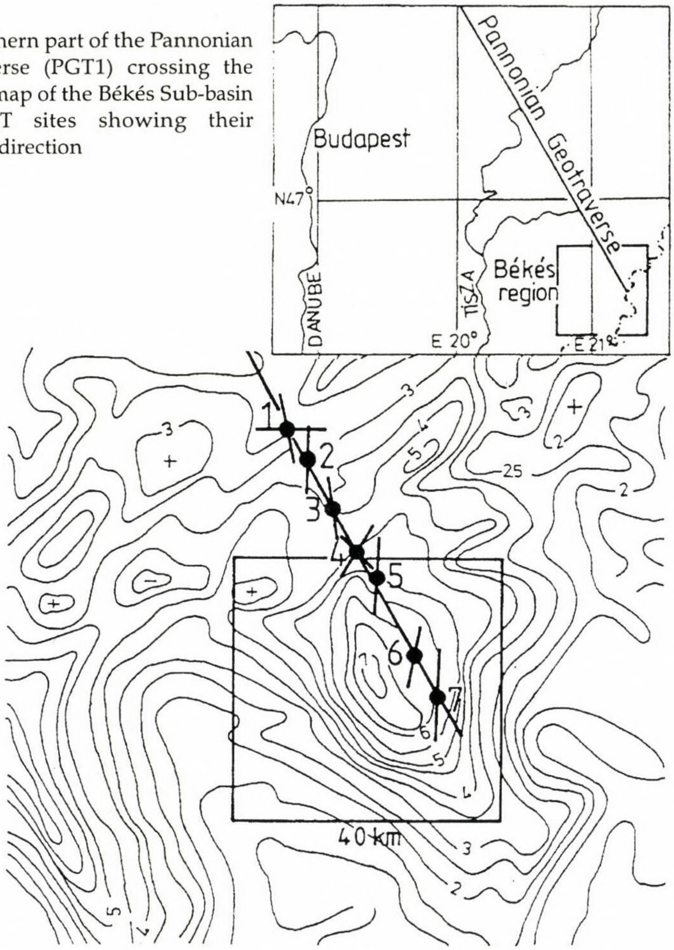


Fig 5g
VII: jászság depression (in the Great Hungarian Plain): Tiszagyenda (GGRi)

Fig. 6
The southern part of the Pannonian Geotraverse (PGT1) crossing the isopach map of the Békés Sub-basin with MT sites showing their Rhomax direction



shallow depth caused a great dilemma for our MT group because its determination is strongly influenced by the distortion effect of the thick sedimentary cover. The shallow depth cannot be accepted without reservations (we shall return to this question when correcting the data).

Corrections used in the determination of the asthenospheric depth

Before constructing a map of the asthenosphere the MT data have been very critically studied from the point of view of different kinds of distortions (e.g. 3D effects, static shift, etc.) and then corrected. In the following the different correction procedures will be described and illustrated by examples.

Decomposition

From the beginning of the use of the magnetotelluric method in Hungary, our attention has been attracted by the difference between the extreme values of the MT apparent resistivity, i.e. the MT anisotropy in the Pannonian Basin.

Very early (Ádám 1969) we found a systematic difference in these data oriented in two characteristic directions of the Pannonian Basin. These directions can be seen in the map of the crystalline basement structure, with parallel NE–SW directed strips of Paleozoic and Mesozoic rocks separated by longitudinal fractures. It has been supposed that this quasi 2D "basin and range" structure of the basin can be cleaned of 3D near-surface effects using decomposition (Groom and Bailey 1989; Bahr 1988) on the basis of the formula

$$Z(\omega) = R(\Theta)C(\Theta)\begin{bmatrix} 0 & Z_{\|}(\omega) \\ -Z_{\perp}(\omega) & 0 \end{bmatrix} R^T(\Theta)$$

where C (Θ) is a real matrix independent of frequency, R (Θ) a rotation matrix R^T (Θ) is its transpose, and $Z_{||}$ and $Z_{\perp}$ are the antidiagonal elements of the 2D impedance tensor.

Using Groom-Bailey's and Bahr's decomposition techniques, which can be transformed to each other, the strike direction has been calculated to select the E and B polarizations. We were disappointed by the results: the average strike direction appears near the N–S direction (Fig. 7) with an angle about 45° to the characteristic strips of the basin. As an example of the results of decomposition, we present the Groom-Bailey parameters (strike, twist, shear) of the MT site Túrkeve from region I (Fig. 8). The original Rho and phase curves are given in

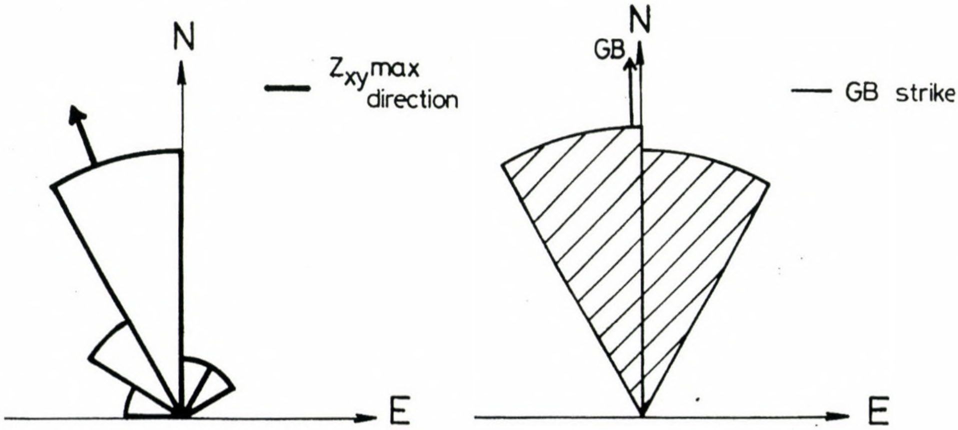


Fig. 7
Rose diagrams for the original Z_{xy}max direction and Groom-Bailey strikes for MT sites in the Great Hungarian Plain (Ádám 1998)

Fig. 5a and the Z_{xx}, Z_{xy} polar diagrams in Fig. 9. The distortion is low. The strike direction changes between about +20 and –20° from North, i.e. near the Rhomax direction. Bahr's strike direction is similar to the one given in Ádám (1998) to Bahr's formulas for other parameters. This result is unusual because induction vectors, which generally indicate the dip direction in a sedimentary basin, also point to the Rhomax direction in the Pannonian Basin. This means that the original idea of decomposition does not work here. We do not know why decomposition fails for the Pannonian Basin. There is a suggestion that full 3D interpretation could explain it. We can also cite Berdichevsky's (1999) opinion

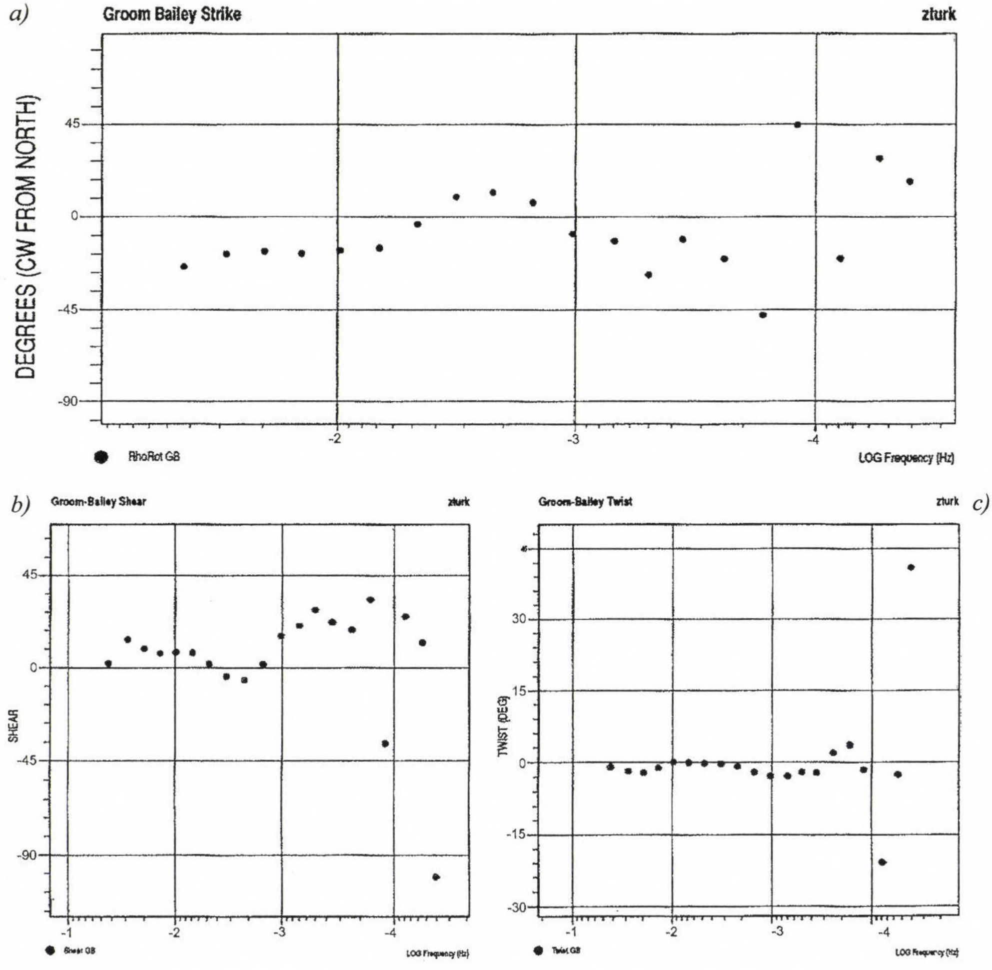


Fig. 8
a) Groom-Bailey (GB) strike vs. log frequency, b) GB shear vs. log frequency, c) GB twist vs. log frequency at MT site Túrkeve

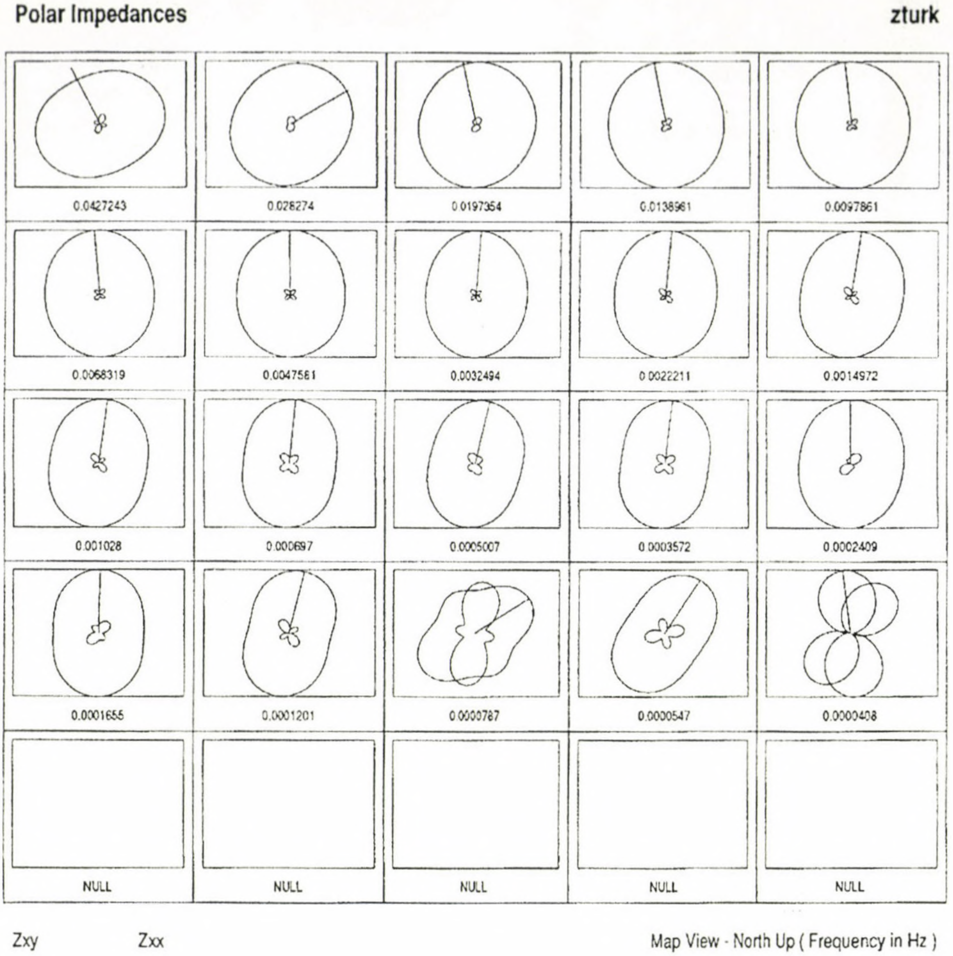


Fig. 9
Z_{xx} and Z_{xy} frequency-dependent polar diagrams for Túrkeve

here: Bahr and Groom-Bailey decomposition give stable results if transverse and longitudinal components of the 2D regional impedance have significantly different phases. In the case of the thick sediment of the Great Hungarian Plain the phase curves do not differ significantly.

Decomposition decreased the deviation of the extreme Rho values. Their ratio decreased in average from 1.5 to 1.1. The Groom-Bailey parameters (twist and shear) are generally low especially in the Great Hungarian Plain (between 7° and 11° in average, taking only their absolute value). For comparison the average asthenospheric depth values are shown in Table 2 with and without decomposition for the Great Hungarian Plain, excluding a few very disturbed data. Hence it can be stated that the difference among them is small.

Table 2
Astenospheric average depth in the Great Hungarian Plain

$h_{1\,\mathrm{Rhomax}}$	= 65.2 ± 14 km (18 values)	
$h^{\mathrm{GB}}_{1\,\mathrm{Rhomax}}$	= 62.1 ± 17.5 km (19 values)	
$h_{2\,\mathrm{Rhomin}}$	= 52 ± 18 km (25 values)	
$h^{\mathrm{GB}}_{2\,\mathrm{Rhomin}}$	= 59.2 ± 17.6 km (21 values)	
$\sqrt{h_1 \cdot h_2}$	= 58.1 km	difference is 1.9 km
$\sqrt{h_1^{\mathrm{GB}} \cdot h_2^{\mathrm{GB}}}$	= 60.0 km	

Static shifts (effect of electric charges)

Static shift is a typical B polarization effect, which is due to the electric charges generated at the surface of the electric inhomogeneities. There are two typical forms.

a) The most common one is a frequency-independent shift of the resistivity sounding curves along the Rho axis. This does not affect the phase curves, which are in differential relation with Rho values.

b) The second one is, as described by Berdichevsky and Dmitriev (1976), connected to the morphology of the highly resistive basement – e.g. horst and graben – covered by sediments of varying thickness. This is the so-called S-effect because it shifts the decreasing branch of the Rho sounding curves which indicates a conductive layer.

Both types of static shifts distort MT curves of the Pannonian Basin, mainly in connection with the morphology of the basement of the sedimentary basins and of the conductive and resistive layers in the upper crust (including the basement).

Some examples of the causes of distortions in the Pannonian Basin:

a) extensional deep basins covered by thick sediment (e.g. Békés (Sub)basin in Fig. 6)

b) Transdanubian Conductivity Anomaly (Fig. 10) with its highly conductive graphitic formations, or as a counterexample, the very resistive ultra-metamorphic rocks in the basement of the area between the rivers Danube and Tisza (region IV) and their morphology. This latter example is shown by a resistivity pseudosection (Fig. 11).

Both effects cause static shift of the MT curves and influence the depth of the asthenosphere as well. Therefore, one must find empirical solutions to free these curves from distortions. Decomposition does not provide a solution for static shift.

The thick, low resistivity sediment in the Békés Sub-basin causes an apparent decrease in the depth of the asthenosphere (Ádám et al. 1996). To eliminate this distortion a network of magnetotelluric measurements have been carried out by the Eötvös Geophysical Institute (ELGI) in the area of the sub-basin, and the relationship between the conductance of sediments (S) and the apparent depth of the asthenosphere was computed (Fig. 12). The upper period limit of these ELGI data is about 1000 s, which is generally not enough for an exact determination of the asthenospheric depth. This explains the large scatter of the data. Extrapolating to zero sediment thickness (S = 0) the asthenospheric depth appears at 44 km in the case of Rhomin curves and at ~60 km at Rhomax ones.

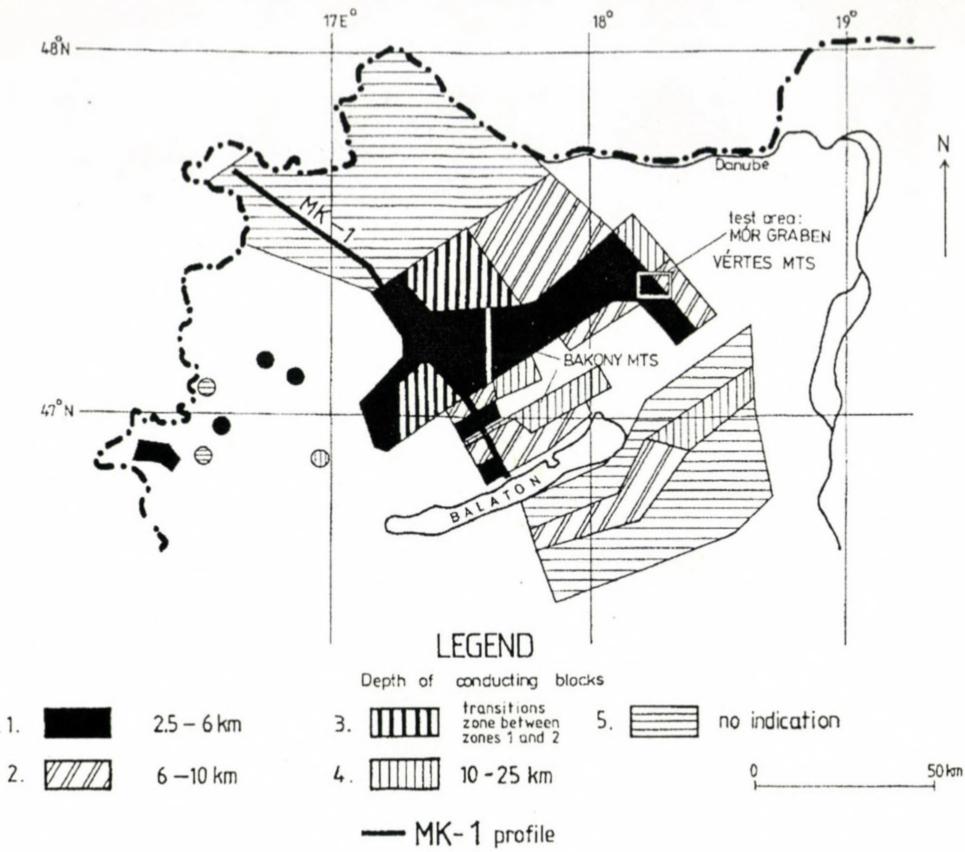


Fig. 10
The lateral extension of the Transdanubian CA (Ádám and Varga 1990)

The Békés Sub-basin being a 3D structure, its asthenospheric depth can be best approximated by the geometric mean of the mentioned two extreme values: it results in ~51 km, approximating the one obtained by the deep seismic soundings (Posgay et al. 1995).

The 2D inversion of another magnetotelluric data set measured by GGRI – having much longer upper periods than ELGI's ones – has given a similarly low asthenospheric depth for the Békés Sub-basin along the PGT1 geotraverse (Ádám et al. 1996; Ádám and Bielik 1998). It is worth noting that the geometric means of the asthenospheric depth obtained by 1D inversion of the Rhomax and Rhomin curves at these sites resulted in about the same values as derived from ELGI's data.

In the area between the rivers Danube and Tisza the highly resistive ultra-metamorphic blocks in the basement distort the MT curves, as mentioned, and

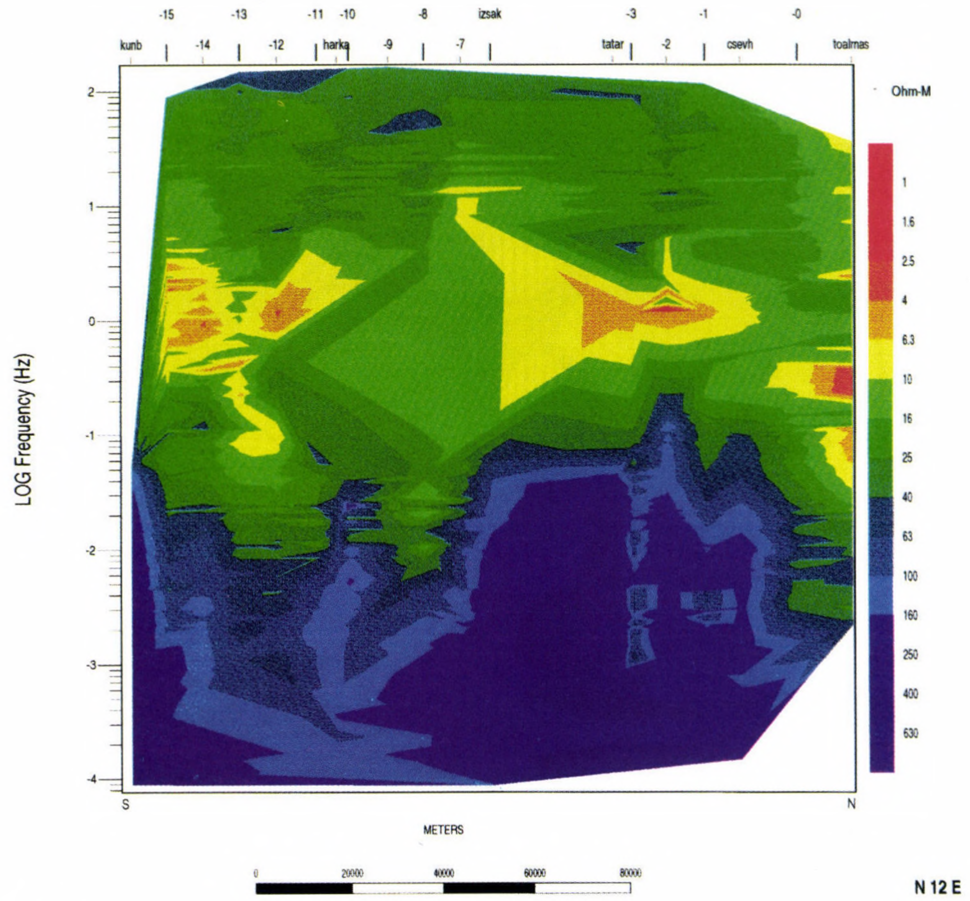


Fig. 11
Apparent resistivity pseudosection across the MT sites between the rivers Danube and Tisza

shift these curves toward greater resistivity values. Here also statistical treatment helped us to find the most probable asthenospheric depth in the case of the scattered data of a N–S profile. Figure 13 shows the close relationship between asthenospheric depth and resistivity of the basement; therefore, correction is necessary in the case of the measuring site Tatárszentgyörgy. In the case of Izsák the depth is acceptable, but the resistivity of the asthenosphere is exceptionally high (instead of about 10–20 Ωm it is about 250 Ωm). The scatter of the depth values measured between the Danube and the Tisza is shown in Fig. 14, with the adjustment line and with the strong outlier of Tatárszentgyörgy.

Against these points (distorted toward higher resistivity) there are resistivities shifted toward lower values in the case of the measuring sites in the Transdanubian CA. Here the shallow apparent depth of the asthenosphere is due to this type of static distortion, and should therefore also be corrected.

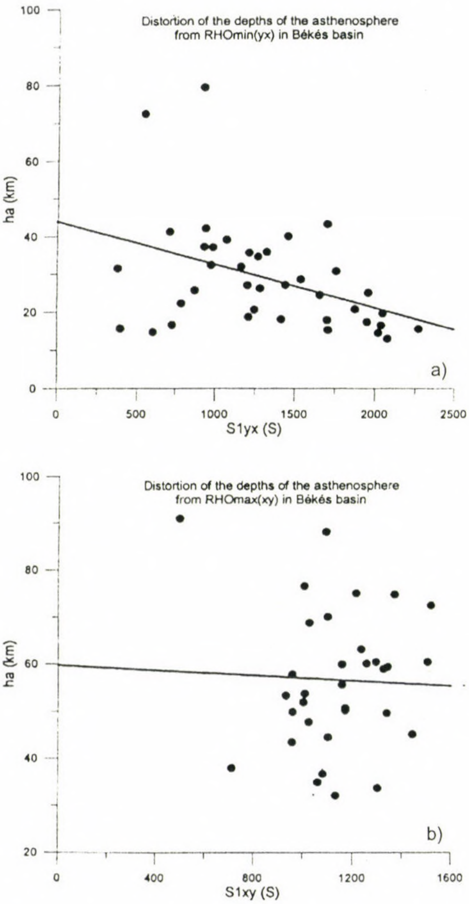


Fig. 12
Depth of the asthenosphere (h) in the Békés Sub-basin in function of sediment conductance (S). Both values have been calculated a) from Rhomin curves and b) Rhomax curves

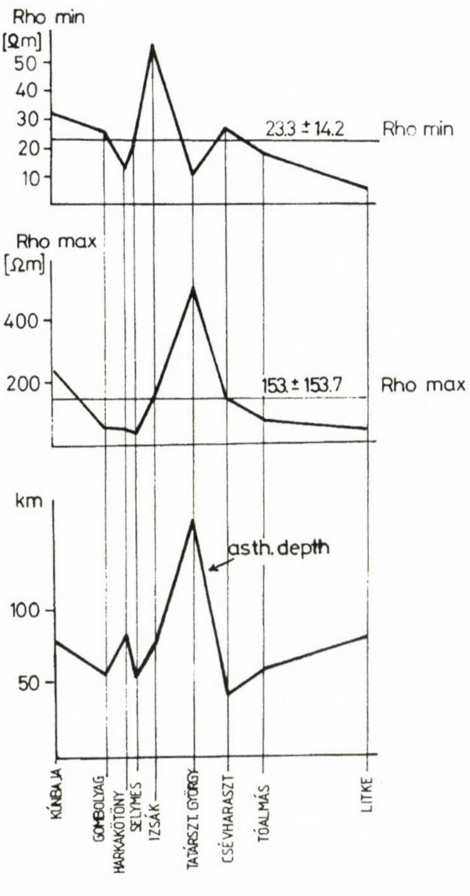


Fig. 13
Apparent Rho values at periods 50–100 s and depth of the asthenosphere in the area between the Danube and the Tisza

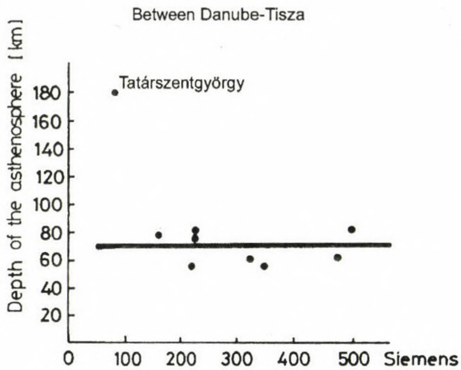


Fig. 14
Asthenospheric depth in the area between the rivers Danube and Tisza, depending on the sediment conductance S and an adjustment line

Map of the depth of the asthenosphere in the Pannonian Basin (Fig. 15)

Only those data have been taken into account where reliable asthenospheric depth could be calculated by 1D inversion from both extreme curves (i.e. Rhomax and Rhomin). Since decomposition did not bring any essential change in the parameters of the asthenosphere, primarily the static shift effect was corrected on the basis of the mentioned aspects. As a next step the geometric mean of the asthenospheric depths derived from Rhomax and Rhomin curves were calculated, approximating Berdichevsky's determinant as proposed for complicated (3D) structures, and used for the map. The depth step was chosen as 10 km, which can correspond to the uncertainty of the determination. This map shows an average depth of 55–65 km in the center of the basin. This large area is embraced by a belt of 65–75 km depth range, which contains sites with asthenospheric depths of 75–85 km as well, i.e. the asthenosphere continuously deepens when moving away from the center of the Pannonian Basin. Inside the central part of the basin there are extensional subbasins, especially the Békés one, where it seems that the asthenosphere is somewhat shallower than 60 km, if the effect of the thick sediments has been correctly eliminated. Reliable and sufficient data are available only for the Békés Sub-basin, which is characterized by a high gravity anomaly, indicating a high-density intrusion in the crust in connection with its extension. Posgay et al. (1995) corrected the heat flow values for the steady-state condition and so obtained high values, which also hint, together with seismic results (Posgay et al. 1995) at the upwelling of the asthenosphere, according to Ádám's (1978) empirical relationship between heat flow and asthenospheric depth (Fig. 1).

To increase data quantity a few other data can be included into the asthenospheric map. Nevertheless, severe problems arise in connection with these data. Generally only one of the extreme measurement curves seemed to be reliable, giving the depth of the asthenosphere in that range, which corresponds to the Pannonian Basin. Such a curious measuring site is "Aszófő"' near Lake Balaton, which lies at the southern boundary of the Transdanubian CA (TCA). Here the 1D inversion of the Rhomax curve yielded 2000 km for the asthenospheric depth, certainly a strongly distorted value, which should be dropped. The depth value derived from the Rhomin curve is also distorted, but downward, by static shift due to the TCA, similarly to other sites in the area of the TCA.

Beside the extensional sub-basins in the Great Hungarian Plain, an upwelling of the asthenosphere appears in the southwestern part of the country, in the so-called Dráva Sub-basin, as well, where the effect of thick sediment cannot be taken into account since data are not sufficient for a statistical distortion analysis similar as shown in Fig. 12 (see MT station "Pe" in Fig. 4). In these uncertain cases, new, more precise, undistorted measurements need to be carried out. Therefore it should be emphasized that, due to the relatively small number of measuring sites, our asthenospheric map is only a draft, and aims to hint at the difficulties in the construction of such a map by magnetotellurics.

Apparent resistivity of the asthenosphere

As a result of the inversion, only a very approximate resistivity value of the asthenosphere can be obtained. As in the case of a conductive layer, the law of equivalence is in effect, and this resistivity is nothing else than a rough, informative value. In addition, all of these Rho values are influenced by the geologic formations through static shift, because its effect cannot be totally eliminated by corrections. This effect is reflected, for instance, in the low resistivity of 8–9 Ωm in the Békés Sub-basin, and tens of Ωm appear above the resistive ultrabasic magmatites between the Danube and Tisza rivers, when derived from Rhomin curves. By averaging all the Rho values for the entire area, the obtained large RMS also makes their value questionable. These approximate Rho values for the conductive asthenosphere are as follows:

on the basis of Rhomax curves:	85 Ωm
on the basis of Rhomin curves:	16 Ωm

The RMS error of these Rho values is unreasonably high.

Since these resistivity values show negative correlation with the asthenospheric depth farther away of the center of the basin, the deepening of the asthenosphere toward the Carpathian rim cannot be attributed to static shift.

In a few cases the conductance of the asthenosphere could only be derived from the deep magnetotelluric sounding curves. These values are much greater than 4 kSiemens calculated by Semenov et al. (1997) by the combination of MT and MV techniques.

Conclusions

It is obvious that the uncertainty of the magnetotelluric parameters increases with the depth of the target (inhomogeneity) to be studied. This is due primarily to the near-surface inhomogeneities, which distort the asthenospheric depth in very different manner, as was shown. Corrections used in our case (decomposition, static shift) could only roughly improve the data. Nevertheless, the

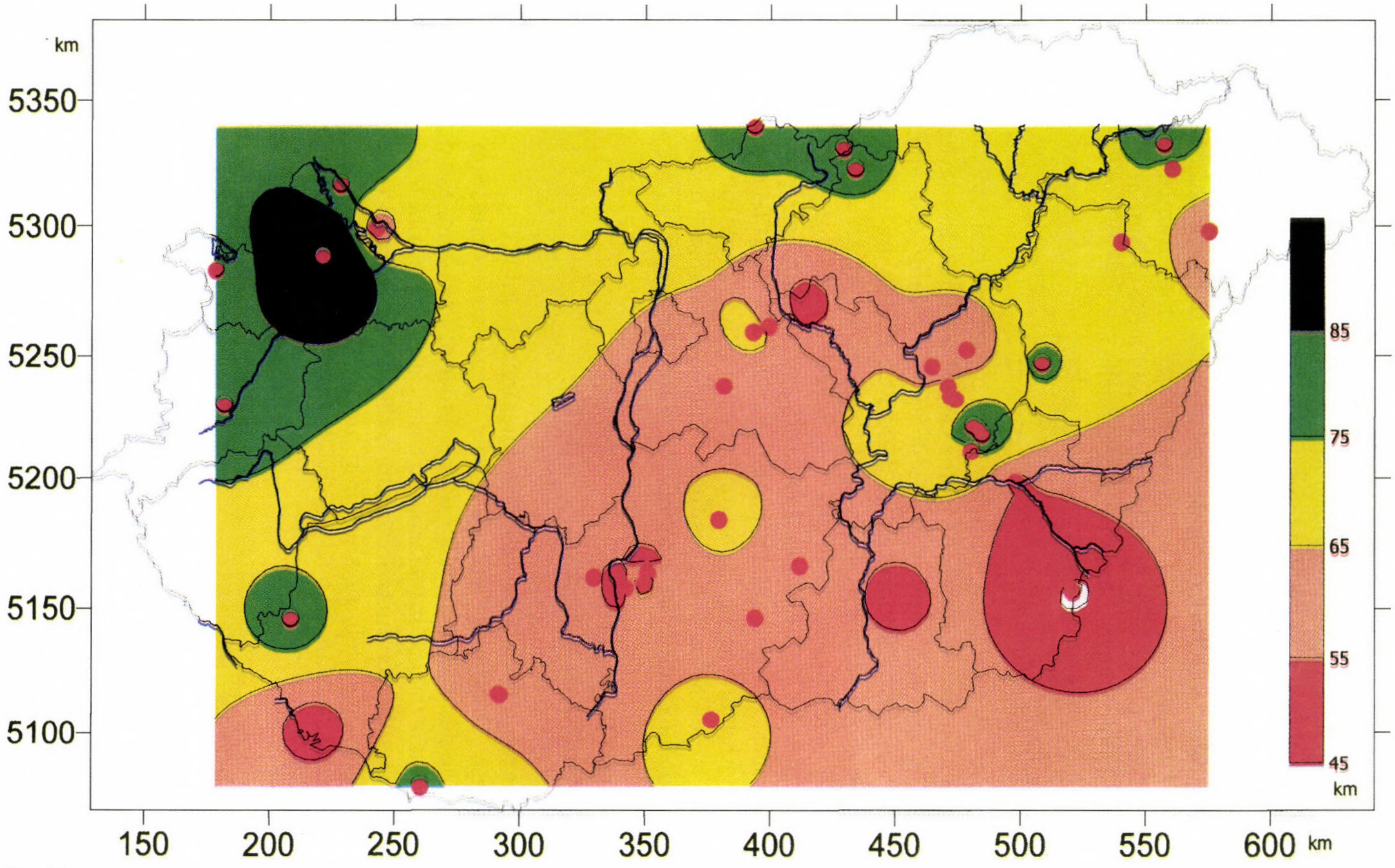


Fig. 15
Map on the asthenospheric depth based on the geometric mean of the depth values derived from Rhomax and Rhomin curve

following earlier statements concerning the conductive asthenosphere are confirmed:

– The depth of the asthenosphere within the Pannonian Basin is at an average of 60–65 km, and is in agreement with the regional heat flow values ($\sim$100mW/m^2).

– Departing from the center of the basin toward the Carpathians asthenospheric depth increases. It reaches about 80 km at the western border of the country toward the Eastern Alps as well (Ádám et al. 1997; Ernst et al. 1997)

– In the extensional deep subbasins, especially in the Békés Sub-basin, there is an upwelling of the asthenosphere to about 50 km. There is a similar upward deformation in the MOHO and the lower crust, as found by the Posgay et al. (1995) deep seismic measurements.

– The lowest average resistivity of the asthenosphere is, in the case of Rhomin curves, $\sim$16 Ωm. This value accounts for a partial melting of only a few percent at the base of the lithosphere (Shankland and Waff 1977).

– The most difficult problem to be overcome is static shift. For its elimination there is no general rule. Since decomposition does not help, uncertain empirical solutions need to be found, which degrade the value of the map. Nevertheless, its regional features can be stressed.

Acknowledgements

The authors express their gratitude for the grant of the OTKA (T 029443), which helped to carry out new long-period magnetotelluric measurements in Hungary, and they especially thank their colleagues for taking part in these measurements and data processing. The publication of these results is supported by one of the referees' remark: The Pannonian Basin might be an attractive target for future 3D modeling. But at this stage, a 1D-approach with empirical kind of 3D corrections might be worth publishing.

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Paleogeographic/structural evolutionary stages and the related volcanism of the Carpathian–Pannonian Region

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Tertiary subsidence of the Pannonian Basin, interrupted locally by inversions, generated huge depocenters, which were filled by detritus eroded from the surrounding orogenic belts, i.e. the Eastern Alps, Carpathians and Dinarids. Based on revision and updating of data sets, a combination of local and regional geologic maps, geophysical, volcanological, tectonic and sedimentological data, paleogeographic/structural, sediment thickness and volcanic maps have been constructed for the evolutionary stages of the Pannonian–Carpathian region as follows.

1. Early Miocene paleogeographic/structural evolutionary stage (Upper Egerian–Eggenburgian–Ottnangian; 24–18 Ma), early synrift phase, NW–SE striking rifts, Mediterranean/Atlantic/Indo-Pacific connections;

2. Middle Miocene paleogeographic/structural evolutionary stage (Karpatian–Middle Badenian; 18–14 Ma), major synrift phase, connection of NE–SW striking rifts with the Mediterranean Sea;

3. Late Miocene paleogeographic/structural evolutionary stage (Upper Badenian–Sarmatian–Pannonian–Pontian; 14–4 Ma), thermal subsidence phase, emerging Carpathian orogenic belt, NW–SE and NE–SW-oriented depocenters, southeastern connections with the Black Sea/Aral–Caspian Basin;

4. Pliocene–Quaternary (4–0 Ma); later stage basin inversion, uplift and erosion in Transdanubia and the Danube-Tisza interfluve, continuous subsidence in the Danube, Drava, and Tisza Basins.

Volcanism of the Carpathian–Pannonian region was related to the thermo-mechanic evolution of the crust/mantle system (the driving mechanism of basin subsidence). The distribution of volcanic activity in space and time were controlled by factors (subduction, astenospheric dome, stress field orientation, terrain displacement), strongly affecting the paleogeographic evolution of the region as well.

Maps showing volcanism versus time and space have been constructed as follows:

1. Crustal origin, areal acidic volcanism
2. Mantle derived, areal andesitic volcanism
3. Subduction slab break off-related (basaltic andesitic to andesitic) volcanism
4. Alkali basalt (not effected by subduction) volcanism

Sediment thickness maps for the above-mentioned paleogeographic/structural evolutionary stages have been constructed for the central part of the Carpathian–Pannonian Basin system.

Key words:paleogeography, sterructural evolution, volcanism, Neogene Carpathian–Pannonian Basin subduction, stress field, astenosphere dome, Pleistocene

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Introduction

The paleogeographic evolution of a sedimentary basin is a result of the interaction of initial physiography, tectonics, sediment supply and eustacy. Physiography, tectonics and eustacy control the amount of space available for sediment to accumulate and sediment supply governs how much of the accommodation space is filled. The multi-phase Tertiary subsidence of the Pannonian Basin, interrupted by inversions, generated huge depocenters, which were filled with detritus eroded from the surrounding orogenic belts, i.e. the Eastern Alps, Carpathians and Dinarids (Fig. 1). Distribution of volcanism in space and time were controlled by subduction, the ascending of an astenospheric dome, stress orientation and tectonostratigraphic terrain displacement.

Sediments transported by rivers were sorted and deposited by the waves and currents of the Central Paratethys Sea, which was from time to time connected to the Eastern Paratethys and to the Mediterranean Sea. During Late Miocene time the Pannonian Basin system was gradually isolated from the oceans and became an inland sea. Its water became brackish, and the basin system was gradually filled up. The total thickness of the Late Paleogene–Quaternary sediments of the Carpathian–Pannonian basin systems amounts to 6–8 km (Fig. 1) in the deepest sub-basins (Carpathian foredeeps, Vienna, Danube, Mura-Dráva, Hódmező-vásárhely, Békés, and Derecske Basins).

The purpose of this paper is to present the results of a multidisciplinary study of the Late Oligocene to Quaternary paleogeographic evolution within the Carpathian–Pannonian region (CPR). In the attempt to derive a set of coherent paleogeographic/structural and volcanic maps we either compiled or reviewed previously compiled data sets, which had been produced by different groups of researchers working for research institutes of several countries of the Carpathian–Pannonian region. These data sets include local and regional geologic maps, geophysical, volcanic, tectonic and sedimentological data. Based on revision and updating of those data sets, paleogeographic/structural and volcanic maps have been compiled for the Pannonian–Carpathian region (CPR). Sediment thickness maps for paleogeographic/structural evolutionary stages have been constructed for the central part of the Carpathian–Pannonian basin system. Each paleogeographic, volcanic and thickness map was prepared on the same geographic base. Stages and their absolute ages used for the Neogene are taken from the time-scale for the Central Paratethys (Hámor and Halmai 1995).

Paleogeographic/structural evolutionary stages

The Pannonian basin is underlain by a large orogenic collage which is built up by several tectonostratigraphic terranes (Brezsnyánszky et al. 2000). The crustal terranes of different origin reached their present positions through strike-slip movements and/or by rotation during step-by-step closure of the Tethys Ocean. The existence of the Tethyan system ended at the Eocene/Oligocene boundary, and a series of basins belonging to the Paratethys (remnant basins of the former Tethys) came into existence in the Early Oligocene (Nagymarosy 1990a, 1990b). The collision effects of the African (Adriatic) and European plates controlled the tectogenesis of the sub-basins of the Pannonian Basin system, and most of them were developed approximately above the "collision-sutures" (Hámor 1983; Haas et al. 1999).

The convergence of the African and European Plates resulted in subduction roll back, subduction slab break-off, terrane displacement, and variations in stress

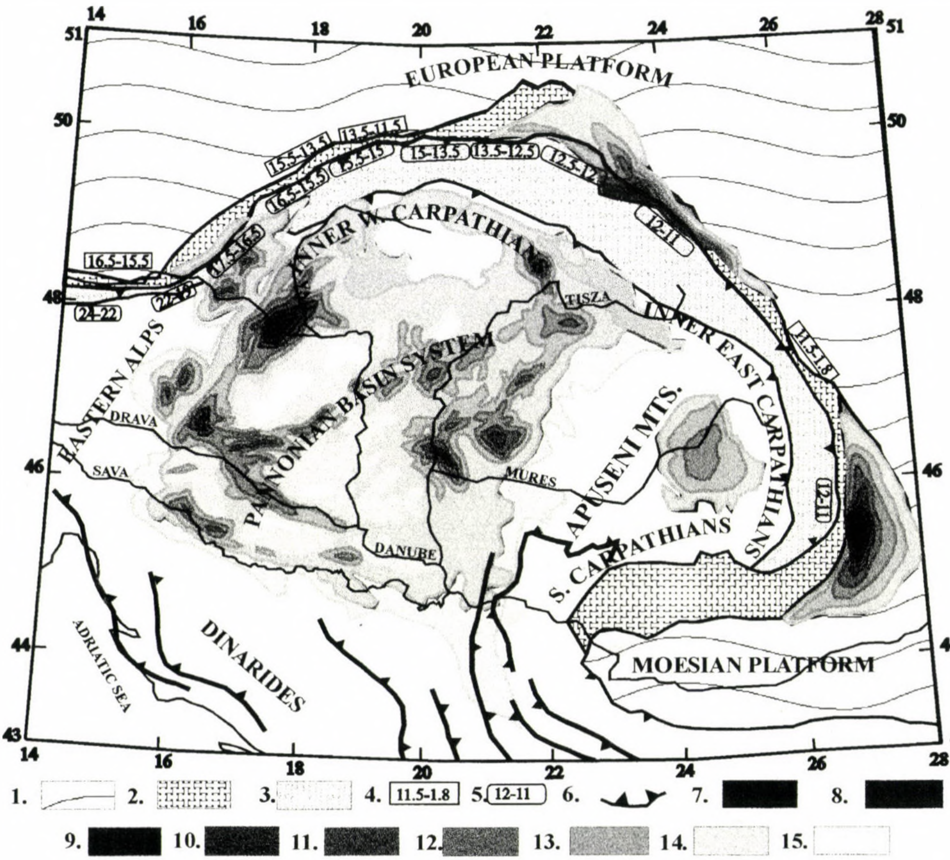

Fig. 1
Late Paleogene–Quaternary sediment thickness map of the Carpathian–Pannonian Region, indicating time periods of last thrusting in the Outer Carpathians and last periods of foredeep molasses sedimentation (modified after Jiricek 1979; Pogácsás 1980; Kilényi and Rumpler 1984; Steininger et al. 1984; Horváth 1988; Royden 1988; Kókai and Pogácsás 1991; Csontos 1995; Meulenkamp et al. 1996; Nemcok et al. 1998). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. Last period of sedimentation in foredeeps; 5. Last thrusting of the Outer Carpathians; 6. Major thrusts; 7–15. Sediment thickness 8, 7, 6, 5, 4, 3, 2, 1, 0.1 km

field orientation. These caused: a) wrench-type rejuvenation of the earlier "sutures", b) formation of a network of synrift transtensional troughs, c) postrift subsidence of interior sags. Latest plate convergence generated late structural inversion in the CPR (Horváth and Cloetingh 1996; Horváth and Tari 1999; Gerner et al. 1999).

The paleogeographic/structural evolution of the CPR consisted of four stages, according to biostratigraphic data, sedimentary cycle data (Hámor 1984, 1995), sequence data (Pogácsás and Seifert 1991; Sztanó and Tari 1993; Vakarcs et al. 1994, 1998; Juhász et al. 1999; Sacchi et al. 1999), reconstruction of paleogeographic and facies relations (Hámor 1995; Hámor et al. 1988, in press; Mindszenty et al. 1988; Kovac et al. 1993; Szentgyörgyi and Teleki 1994; Haas et al. 1999), revision of the Savic, Styrian, Leithaian, Rhodanic orogenic cycles (Hámor 1983), volcanic data (Balkay 1962; Kubovics and Pantó 1970; Embey-Isztin 1976; Póka 1988; Szabó et al. 1992; Simunić and Pamić 1993; Embey-Isztin et al. 1993; Lexa et al. 1995; Pamić and Pécskay 1996; Nemcok et al. 1998) and K/Ar radiometric age determinations (Széky-Fux and Pécskay 1991; Árva-Sós et al.1983; Balogh et al. 1994; Pécskay et al. 1995a, b; Pécskay and Balogh, 2000).

Paleogeographic maps have been constructed for four stages as follows.

1. Early Miocene (Upper Egerian–Eggenburgian–Ottnangian; 24–18 Ma), Figs 2 and 3.
2. Middle Miocene (Karpatian–Middle Badenian; 18–14 Ma), Figs 4 and 5.
3. Late Miocene (Upper Badenian–Sarmatian–Pannonian (14–4 Ma), Figs 6, 7, 8.
4. Pliocene–Quaternary (4–0 Ma); Fig. 9.

1. Early Miocene paleogeographic/structural evolutionary stage (Upper Egerian–Eggenburgian–Ottnangian; 24–18 Ma, Early Synrift Phase). It is represented by one large sedimentary cycle, which consists of three medium cycles: a) Upper Egerian; regressive half cycle, b) Eggenburgian; an entire transgression–regression cycle, c) Ottnangian; an another entire transgression–regression cycle. Lower Miocene layers were deposited in depocenters whose subsidence was initiated by the Savian tectonic cycle, characterized by escape tectonics (Kázmér and Kovács 1985; Csontos et al. 1991; Fodor et al. 1999), northeastward displacement of the ALCAPA (Pelso) terrane, uplifting of the NW–SE-oriented Neo-Vardar zone (Haas et al. 1999; Hámor, in press).

The recent distribution of Lower Miocene facies points to two marine sedimentary systems (Fig. 2). The Carpathian Flysch Basin (an oceanic crust-floored basin no longer in existence) was also present in the Early Miocene; step by step this basin was subducted, and finally vanished. The large eastern/northeastern marine system was connected to a branch of the Atlantic–Indopacific Ocean, while the southwestern one was a part of the eastern Mediterranean (Adriatic) sea (Transtethyan Corridor). The two marine systems were isolated from each other by a relatively emerged (subaerially exposed) ridge of the Neo-Vardar Zone (Hámor 1995, in press), represented by a wide area of continental and alluvial depositional systems (Fig. 2). Sediments were transported in opposing directions (to the NE and SW) off the ridge. In some cases paleo-valleys of the Neo-Vardar Zone (Vah Valley, Hron Valley, Trnava Basin, Sava Valley, Drava Valley, Bacska, Banat, Mures Valley, and Rába Valley) transported and/or accommodated the erosional debris of the sediment provenance areas (Fig. 2). Locally and periodically continental-marine transitional and marine sediments were also deposited (e.g. between Brno and Ostrava, northern side of the Gerecse and Buda Mts.). Structural (strike-slip?) zones are represented by fluvial channel and floodplain facies (Fig. 2), e.g. Backa,

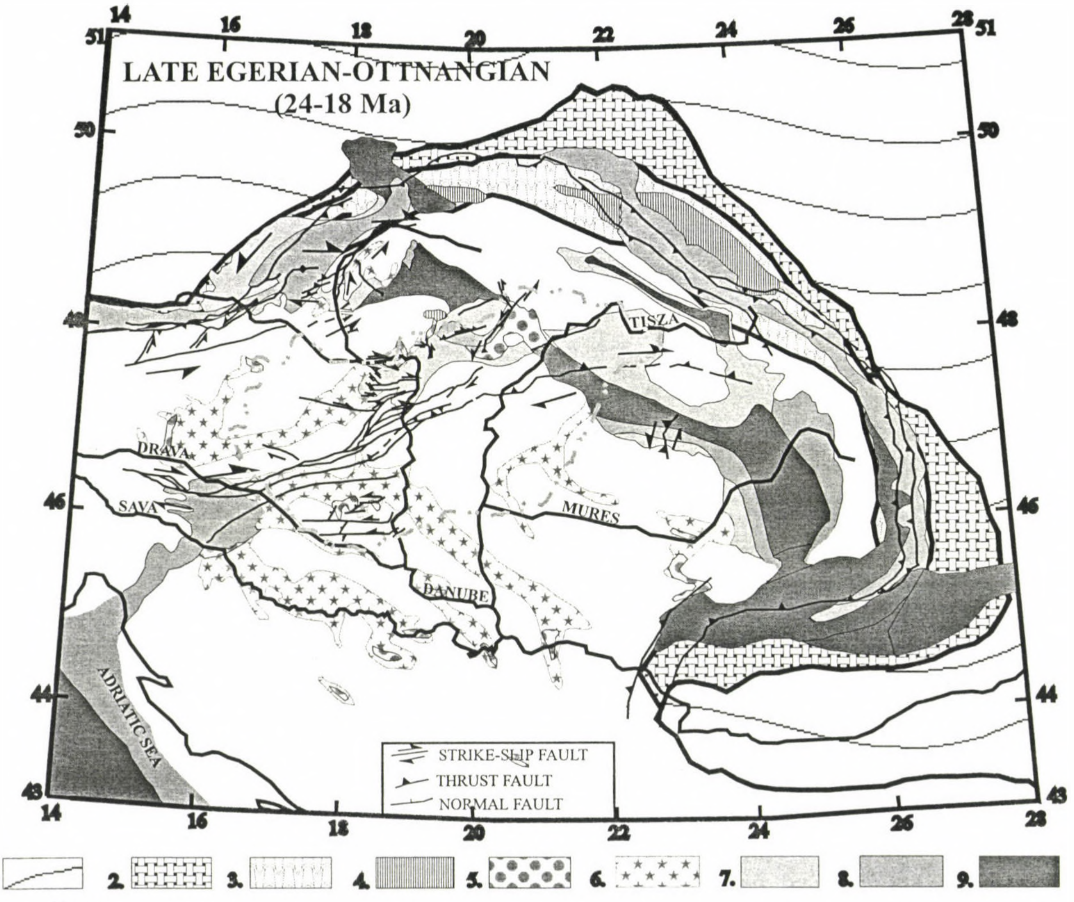

Fig. 2
Late Egerian–Ottnangian (24–18 Ma) paleogeographic/structural evolutionary stage. Paleogeographic sketch has been modified after Hámor (1983, 1984, 1995); Hámor et al. (1988); Báldi (1983). Structural sketch modified after Royden (1988), Kókai and Pogácsás (1991), Pogácsás et al. (1991), Ratschbacher et al. (1993), Csontos (1995), Fodor et al. (1999), and Kiss et al. (in press). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. evaporitic lagoons; 5. coal marsh; 6. fluviatile (channel and/or floodplain) facies; 7. nearshore facies; 8. epicontinental schlier facies; 9. deep water facies

Kolubara-Drava, Sava Troughs in the SW, and the Etes, Sajó–Hajdú, Northern Transylvania and Eastern Slovakian–Maramures Trough in the NE.

During the Early Miocene rotational contact of the ALCAPA and Tisza tectonostratigraphic terranes took place (Balla 1984; Horváth 1993; Csontos 1995; Fodor et al. 1999). Their juxtaposition was produced by the convergence and collision of the European and African plate. Kázmér and Kovács (1985) inferred that the Pelso (ALPACA) terrane, originally lying between the Central Alpine and Southern Alpine Units, reached its recent position by strike-slip movement, resulting in shifting of depocenters from the SW to NE (Báldi and Báldi Beke 1985; Nagymarossy 1990). Microcontinent displacement/rotation was contemporaneous with basin formation. As a consequence a palinspastically restored Early Miocene paleogeography was presumably different from the recent position of Lower Miocene facies belts, as indicated in Fig. 2.

According to Csontos (1995) the observed fault pattern indicates radial shortening at the external part of both the Tisza and ALPACA unit, and shortening along their internal contact (Figs 2 and 3). Fodor et al. (1999) supposed repeated dextral transpression within the ALCAPA, and compression within the Tisza terrane. Thrusts and imbricated structures, associated with shortening, were identified in the Pannonian Basin by seismic (Pogácsás et al. 1991), beneath Middle Miocene and younger sediments. Detailed maps of prerift and early synrift thrusts were presented by Kókai and Pogácsás (1991). Shortening events inverted the former basins (e.g. Szolnok Flysch Trough). Right lateral faults served as accommodation faults between major thrust faults (Royden 1988). The large Petrosani basin (Southern Carpathians) was generated by dextral wrenching along the Moesian corner (Ratschbacher et al. 1993; Csontos 1995). Depocenters with the thickest (more than 800 m) Lower Miocene sediments are related to the Zala–Mura, Danube, and Sajó Basins (Fig. 3).

2. *Middle Miocene paleogeographic/structural evolutionary stage* (Karpatian-Middle Badenian; 18–14 Ma, Major Synrift Phase). It is represented by one large sedimentary cycle, consisting of two sub-cycles: a) Karpatian; an entire transgression–regression cycle, b) Lower Badenian; an entire transgression–regression cycle.

The Middle Miocene paleogeography was dominated by the Styrian tectonic cycle; uplift of the Alps and Carpathians, subsidence of foreland basins in front of the overthrusted Carpathian nappes, collapse of the Dinarids. Fodor et al. (1999) assumed two distinct events within the Carpathian–Pannonian basin: a) the first event is characterized by NE–SW to ENE–NSW tensional or strike-slip-type stress field, b) the second event with E–W to SE–NW tensional or strike-slip-type stress field. During and after the main stages of thrusting in the Outer Carpathians, a series of discrete basins (Figs 4 and 5) opened up behind the Carpathian arc. Development of these basins might have even begun in the Early Miocene. Areas of extension and normal faulting were associated with discontinuous pull-apart basins, push-up swells or divergent strike-slip faults and with fragmentation in zones bounded by two major strike-slip faults (Horváth 1984, 1993; Royden 1988; Kókai and Pogácsás 1991; Tari et al. 1992). The extension resulted from lithospheric thinning, caused by astenospheric flow induced by a steeping (in dip) of the subducting slab.

Royden and Burchfiel (1989) and Royden (1993) proposed that systematic variations in the style of thrust belts resulted from different ratios between the rate of subduction and the rate of plate convergence. Contemporaneous subduction and back-arc extension occurred when the rate of subduction of one plate was greater than the rate of overall plate convergence. This resulted in

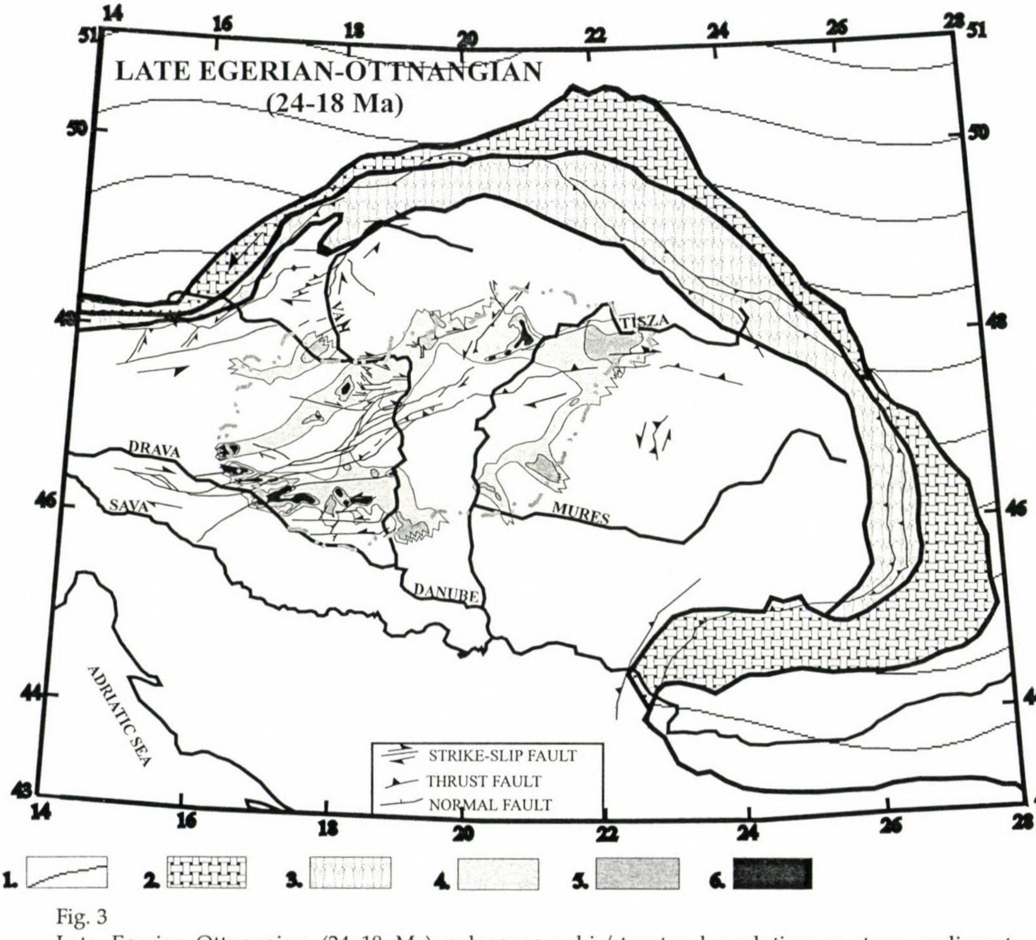

Fig. 3
Late Egerian–Ottnangian (24–18 Ma) paleogeographic/structural evolutionary stage, sediment thickness within the central part of the Carpathian–Pannonian Region. Modified after Báldi (1983), Royden (1988), Bérczi et al. (1988), and Nagymarossy and Müller (1988). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. sediment thickness less than 500 m; 5. sediment thickness between 500 and 800 m; 6. sediments thicker than 800 m

retreat of the subduction boundary toward the oceanic crust-floored flysch basin. Royden (1993) assumed that this process could affect limited segments of the subduction boundary, while other segments of the same boundary remained stationary (with respect to a point in the interior of the upper plate), which led to the formation of strongly arcuate segments within the Carpathian arc, bounded by strike-slip zones on both sides. The driving force of the rollback was gravity acting on a dense subducting slab.

The Middle Miocene stage was characterized by synrift sediments deposited in half-grabens bounded by listric faults, in crestal collapse grabens related to (flat-ramp) listric faults (Kókai and Pogácsás 1991), and in rifts between (smaller size) rotating blocks. Migration of volcanic activity and facies belts took place during relatively short periods of time. Large displacements along listric faults have resulted in tilting of originally horizontal strata, and the formation of a regional unconformity between the synrift Middle Miocene and the postrift Upper Miocene sediments. The wrench fault related synrift basins are filled by terrestrial to marine sediments (Fig. 4). Royden (1988) supposed that sedimentation within each basin was influenced by the proximity of individual basins to the thrust front of the Carpathian. Basins located close to the thrust front contain a thick, fault-bounded synrift sedimentary section, overlain by thin postrift sediments. Basins located in a more internal position within the Carpathian-Pannonian region were starved during the synrift phase. These differences can be related to the differences in thermal subsidence rate of the basement following extension and to the proximity of each basin to the sediment sources in the rapidly uplifted surrounding Carpathians.

The Neo-Vardar Zone was less emerged than previously and acted as a relative levee, but still affected the sedimentary conditions of the Karpatian–Lower Badenian period. In the southwestern part of the Pannonian Basin the thickness of the Karpatian schlier complexes are reduced. In the NE the situation is reversed; the ingressive schlier complex is thick (over 1000 m) but its basal gravel layer is thinner, and in some places is absent. Lower Leitha Limestone reefs and other coastal and nearshore formations are widespread in the southwestern part, and practically missing or substituted by coarse gravel/conglomerate in the northeastern part (Fig. 4). In the NE Pannonian Basin greater water depth is indicated by greater thickness and more significant extent of the Badenian clay. The most important eruption fissure systems of the "Middle Rhyolite Tuff" are NW–SE-striking (Fig. 10). The physiography conditions are reflected in the thickness map of Middle Miocene sediments: depocenters with thicker sediments are related to the southern margin of the Alcapa terrane (Fig. 5). The opening of these basins resulted from wrench movements along a set of roughly NE-SW-trending left lateral and NW-trending right lateral conjugate shears.

In the western part of the CPR, occasionally and locally fluvial sedimentation, regression and swamp formation (Fig. 5) occurred (in the southern part of the Vienna Basin, in the Southern Bakony Mts. and in the area of the Mura–Baranya

Trough). In the eastern part of the CPA, along the inner foreland of the Carpathians, extensive evaporite deposits were simultaneously formed in the East Slovakia–Transcarpathian Depression–Maramures–Transylvanian Basin Belt. The western and eastern areas of the CPR were partly separated from each other by andesitic volcanic ranges (Selmec-Körmöc-Körösvölgy Volcanic Belt, Fig. 11).

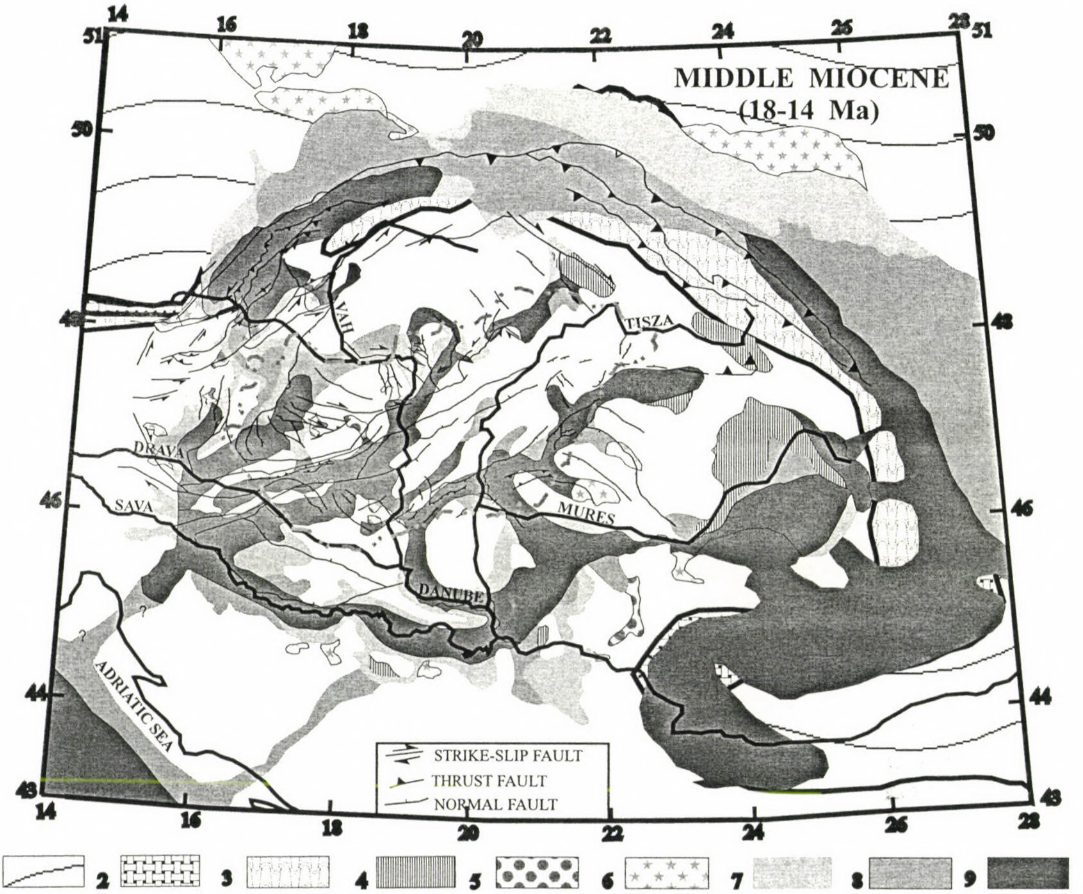

Fig. 4

Karpatian–Middle Badenian (18–14 Ma) paleogeographic/structural evolutionary stage, (see Fig. 2, for references). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. evaporitic lagoons; 5. coal marsh; 6. fluviatile (channel and/or floodplain) facies; 7. nearshore facies; 8. shallow marine facies; 9. deep marine facies

3. *Late Miocene paleogeographic/structural evolutionary stage* (Upper Badenian–Sarmatian–Pannonian–Pontian; 14–4 Ma, Postrift Phase). It is represented by one large sedimentary cycle, consisting of three medium cycles: the Upper Badenian, the Sarmatian, and the Pannonian.

Paleogeographic/structural conditions were controlled by the Leithaian tectonic cycle (Hámor 1978); from west to east step-by-step termination of thrusting in the Carpathians, uplifting of the Alpine–Carpathian–Dinaridic realm, early stage basin inversion (Fodor et al. 1999), ENE–WSW or E–W compression at about 10Ma, followed by renewed extension, slip faulting and postrift thermal subsidence (Figs 6, 7, 8) were characteristic.

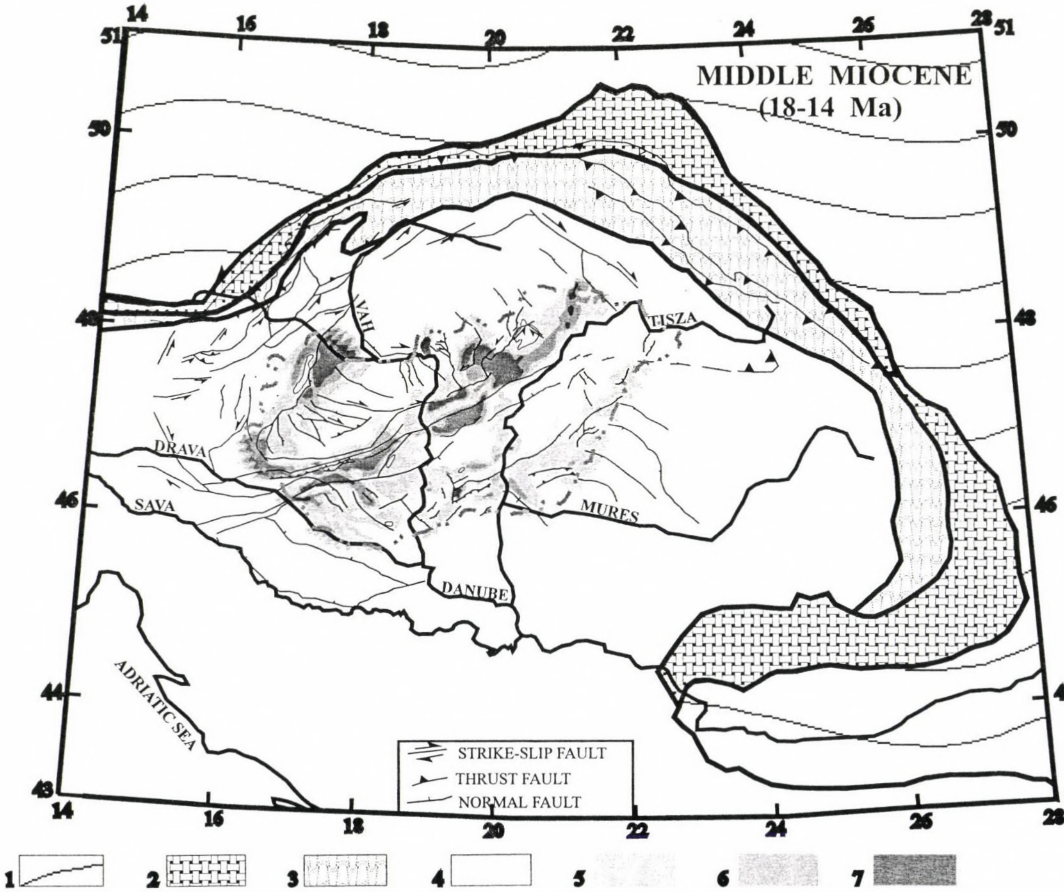

Fig. 5

Karpatian–Middle Badenian (18–14 Ma) paleogeographic/structural evolutionary stage, sediment thickness within the central part of the Carpathian-Pannonian Region (see Fig. 3, for references). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. sediment thickness less than 200 m; 5. sediment thickness between 200 and 500 mé 6. sediment thickness between 500 and 800 m. 7. sediments thicker than 800 m

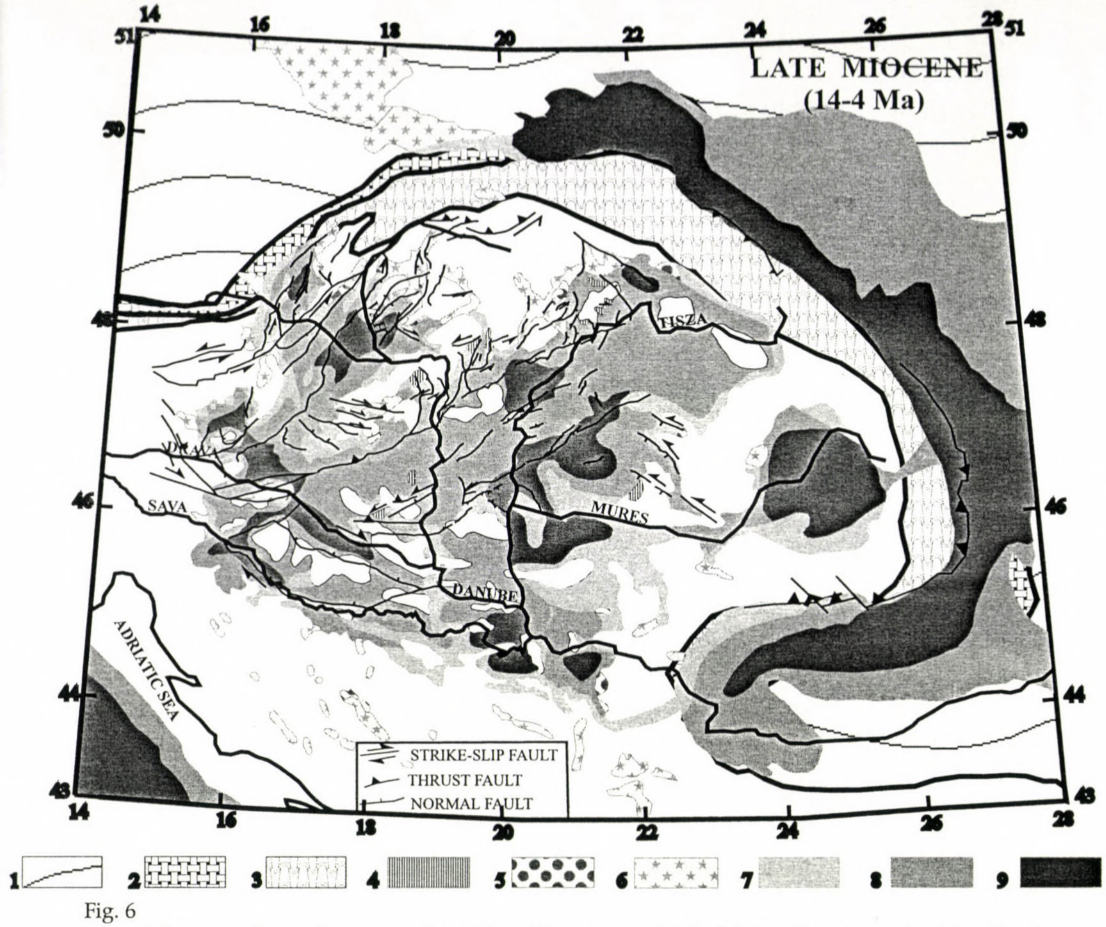

Fig. 6
Late Miocene (Late Badenian–Sarmatian–Pannonian 14–5 Ma) paleogeographic/structural evolutionary stage (see Fig. 2, for references). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. evaporitic lagoons; 5. coal marsh; 6. fluviatile (channel and/or floodplain) facies; 7. nearshore facies; 8. shallow water facies; 9. deep water facies

At the beginning of Late Badenian uplifting of the "inselbergs" (including the newly formed Middle Badenian volcanic mountains) of the Alpine–Carpathian–Dinaridic system and the Carpathian Basin accelerated. Simultaneously, a new southeastward oriented basin system was formed through the Timok Trough (and possibly the Lower Danube Trough) toward the Getic Depression–Black Sea and further eastward. Reefal sedimentation, transgression from the Aralo-Caspian areas toward NW can be demonstrated within the entire CPR with the exception of its northeastern part. Upper Badenian–Sarmatian reef series were deposited at the margins of the emerged areas and in the basins respectively, according to the NE–SW-trending Paleozoic–Mesozoic basement morphology.

The reef series, showing a similar strike, follow the NE–SW-oriented main wrench faults.

During the Sarmatian local deltas were formed at the margins of CPR (northeastern parts of the Vienna Basin–Vah Valley–Nitra Valley, Northern Hungary and Eastern Slovakia, the western margin of the Transylvanian Basin, etc.). The evaporitic diatomaceous lagoon belt of Sarmatian age was developed.

During Pannonian times in the Styria, Mura, Drava, Sava, Vienna, Little Hungarian Plain–Danube Basin, Jászság, Great Hungarian Plain, and Transylvanian Basin, sedimentary sequences of thousands of meters thickness were deposited (Fig. 7). Outside of these areas, in the regions of the former Neo-Vardar ridges, sediments of only some hundreds of meters thickness can be generally found (Fig. 7). According to Royden (1988) the subsidence was

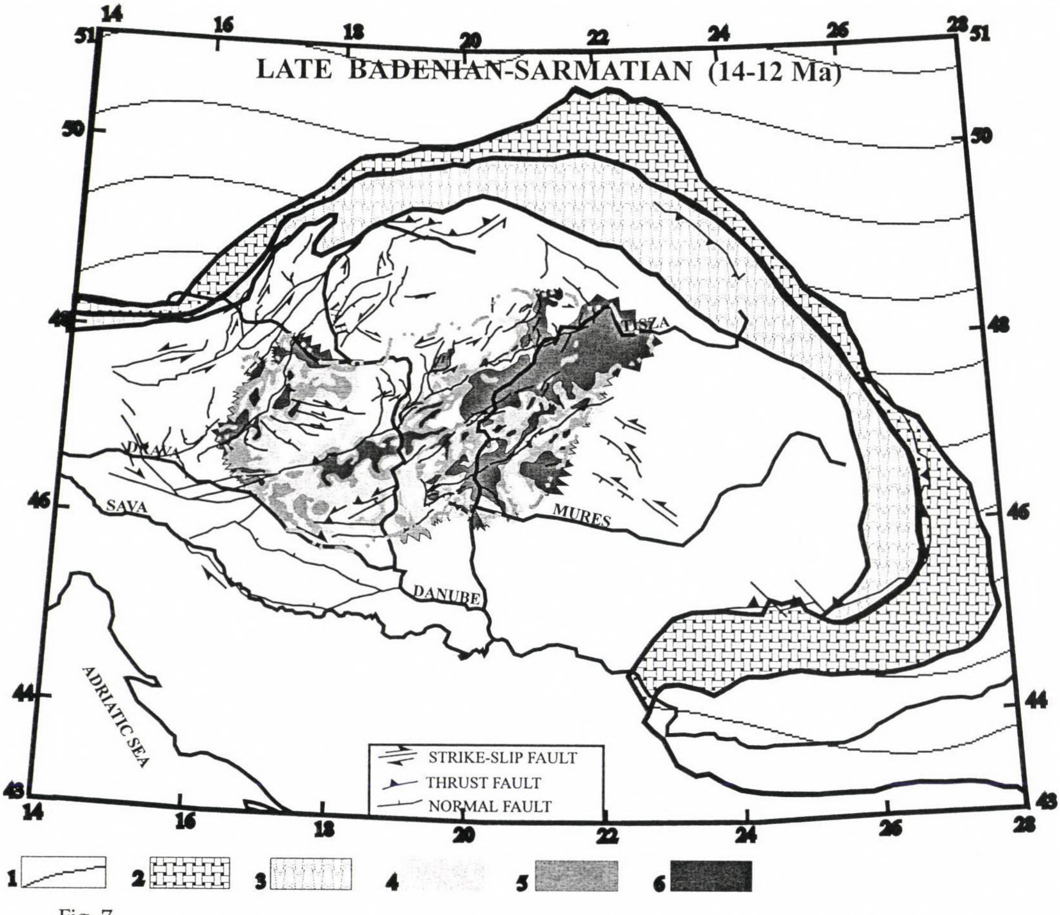

Fig. 7
Late Badenian–Sarmatian (14–12 Ma) thickness within the central part of the Carpathian-Pannonian Region (see Fig. 3, for references). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. total thickness less than 100 m; 5. total thickness between 100 and 500 m; 6. total thickness greater than 500 m

thermally controlled by a decaying thermal anomaly created by extension of the crust and upper mantle. Accompanying the thermal cooling of the lithosphere, interconnected interior sags developed. Sediments filling the interior sags overlie

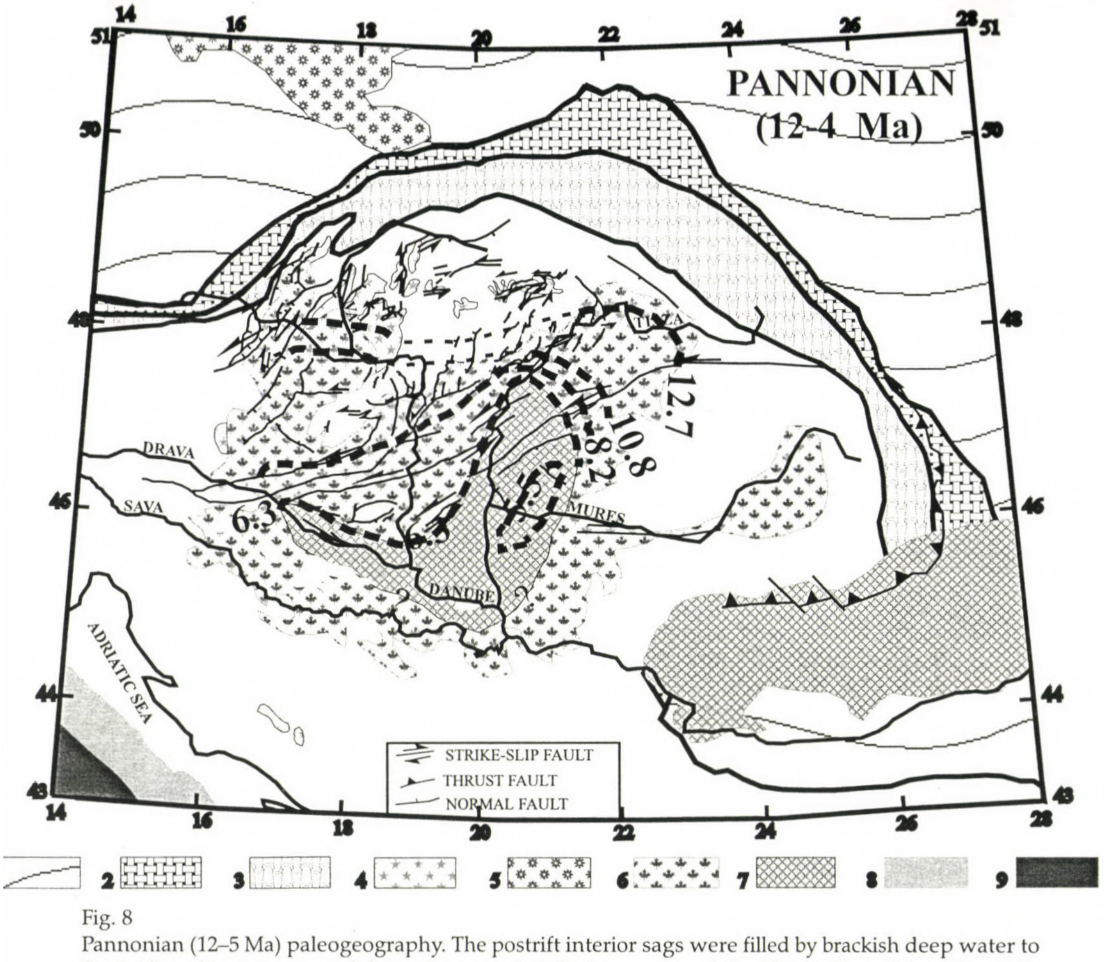

Fig. 8
Pannonian (12–5 Ma) paleogeography. The postrift interior sags were filled by brackish deep water to fluviatile sediments, sourced from the emerging drainage area of the Alpine–Carpathian–Dinaridic realm. The great amount of erosional debris was transported to the Pannonian Basin system by rivers. The southward-prograding, large river-dominated delta systems are represented by deep water shale, prodelta turbidites, slope sediments, delta front mouth bar and foreshore sediments, delta and coastal plain sediments. Position of prograding shelf break line is shown at 12.7; 10.8; 8.2; 6.3; 5.5 Ma. Modified after Pogácsás et al. (1988), Kókai and Pogácsás (1991), Magyar (1991), Ujszászi and Vakarcs (1993), Csató (1993), Juhász (1994), Vakarcs et al. (1994), Juhász et al. (1999), and Sacchi et al. (1999). Paleogeography at 6.3 Ma (time of Messinian salinity crisis in the Mediterranean) is highlighted. 6.3 Ma sequence boundary is marked by a well-developed coastal onlap pattern, referring to a large-scale lake level fall and rise. Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. Fluviatile and alluvial facies at 6.3 Ma; 5. Continental facies of the Polish Lowland Basin; 6. Brackish delta plain and coastal plain facies at 6.3 Ma; 7. Brackish basinal, foreslope and slope facies at 6.3 Ma; 8. Shallow marine facies of the Adriatic region; 9. Deep marine facies of the Adriatic region

the synrift sediments, usually at an erosional unconformity or conformably in the areas inundated at the Middle/Late Miocene boundary. The depocenters of the synrift and postrift phase do not usually coincide. The interior sags are filled by brackish deep water to fluviatile sediments (Jámbor 1989) sourced from the emerging drainage area of the Alpine–Carpathian–Dinaridic realm. A great amount of siliciclastic sediments was transported to the Pannonian basin system and to the Carpathian foredeep system by rivers. Filling up of the interior sag system was controlled by southward prograding, large, river-dominated delta systems. Well-developed third order sequences (consisting of lowstand prograding wedges, backstepping transgressive systems tracts, highstand prograding system tracts) are the building block of the thick postrift unit. Based on high-resolution sedimentological and magnetostratigraphic data of wells with continuous coring, Juhász et al. (1999) observed three major unconformities within the postrift basin fill. The oldest one is between the Pannonian and the older Miocene (or even pre-Miocene strata). The age of basal strata ranges from 9.7 to 11.2 Ma. The second significant unconformity was identified between the Miocene and Pliocene rocks in wells of the Hungarian Geological Institute using continuous coring. The duration of the gap represented by the second unconformity is about 2.0 to 2.5 Ma.

4. *Pliocene–Quaternary paleogeographic/structural evolutionary stage* (4–0 Ma, Basin Inversion Phase). The paleogeographic/structural evolution was controlled by the Rhodanian tectonic cycle. Uplifting of the Alpine–Carpathian arc continued, whereas compression characterized the emerging inner mountainous ranges (inselbergs) as well. Later stage basin inversion took place, characterized by NW–SE to N–S maximal stress axes, strike-slip and compression in the Alpine/Dinaric junction, tensional stress field and subsidence in the central part of the Pannonian Basin (Fodor et al. 1999). Quaternary strike-slip faults have been identified on seismic profiles in several areas within the Pannonian Basin (Pogácsás et al. 1989; Detzky et al. 2000; Bada et al. 2000). A possible marine connection to SE existed through the lower Danube and/or Timok Trough during the Early Pliocene. Progradational delta filling of the Pannonian Basin ended during the Late Pliocene (Fig. 8), and was followed by continental and fluvial depositional systems (Fig. 9). Juhász et al. (1999) observed (the third) unconformity between the Pliocene and the Pleistocene strata. They suppose that in most of the western part of the Pannonian Basin (Transdanubia) unconformities separating Miocene, Pliocene and Pleistocene strata appear together, composing one "super" unconformity separating Quaternary and Miocene strata. This "super" unconformity is the result of regional uplifting caused by the Late Pliocene to recent inversion of the Pannonian Basin. A significant stratigraphic gap associated with the 4 Ma sequence boundary was observed by Sacchi et al. (1999) in SW Hungary. The Late Pliocene/Early Pleistocene tectonic activity was followed by a second phase of uplifting and

erosion at the Early/Middle Pleistocene boundary (1.2 Ma) between the fluviatile Tengelic Formation and the overlaying loess strata. The magnitude of uplift according to the investigation of travertine horizons and Danube River terraces was as high as 80 m in the Middle Pliocene 100 m in the Late Pliocene and 140 m in the Quaternary (Pécsi 1991). Based on seismic sequence analysis Vakarcs et al. (1994) identified 600 m-thick erosion in the Derecske Basin and 850 m thick deposition in the Békés Basin during the last 3.8 Ma.

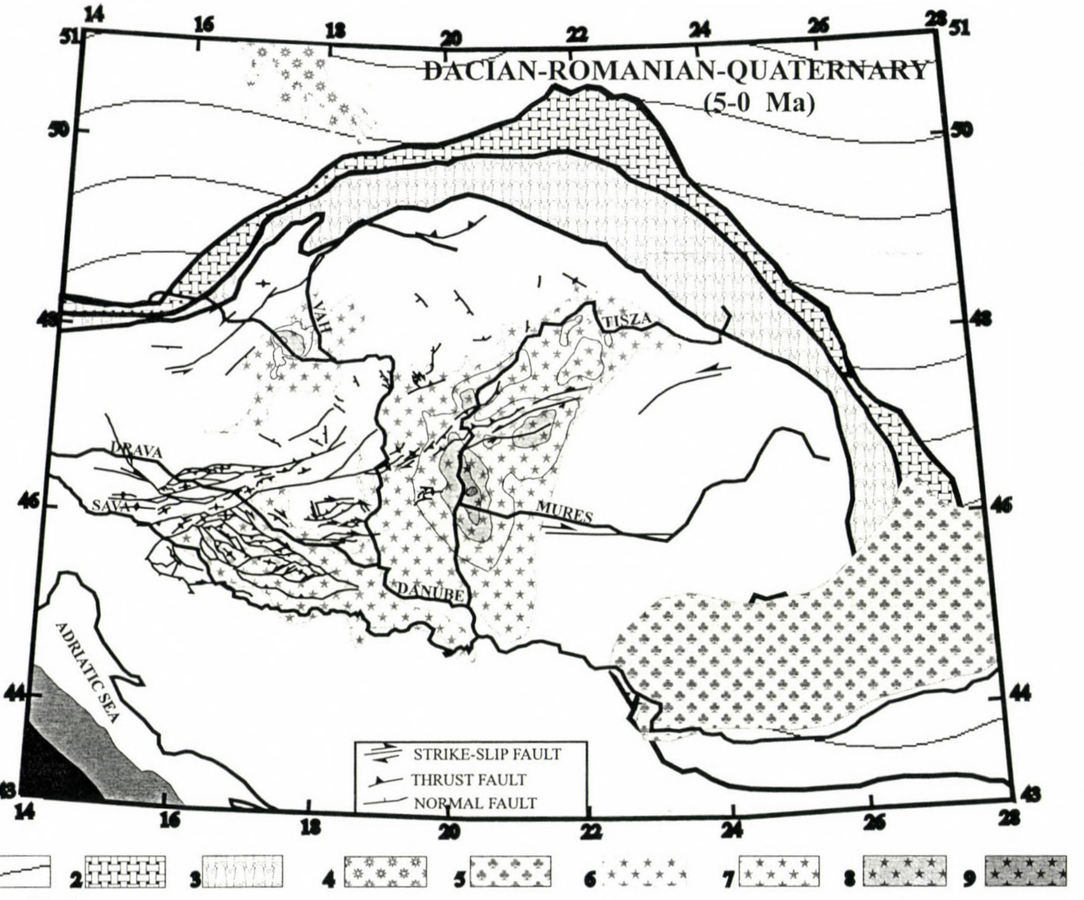

Fig. 9

Pliocene–Quaternary paleogeographic/structural evolutionary stage, sediment thickness and structural elements. Modified after Kretzoi and Krolopp (1972), Molnár (1980), Rónai (1985), Hámor et al. (1988), Pogácsás et al. (1988), Kokai and Pogácsás (1991), Borsy (1992), Franyó (1992), Csontos (1995), Jámbor (1998), Fodor et al. (1999), Nádor et al. (2000), Thamó-Bozsó and Kercsmár (2000), and Bada et al. (2000). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. continental facies of the Polish Lowland Basin; 5. lacustrine, paludal facies of Moesia; 6. sediment (fluvial/alluvial) thickness less than 100 m; 7. sediment (fluvial/alluvial) thickness between 100–300 m; 8. sediment (fluvial/alluvial) thickness between 500–700 m; 9. sediment (fluvial/alluvial) thickness over 700 m; for symbols of shallow and deep marine facies of the Adriatic region see Fig. 8

Pleistocene paleogeography was controlled by climatic factors. Ephemeral lakes/ponds sediments contain thin interbedding dolomite layers. Similar dolomite layers also occur in the delta/coastal plain sediments of the Late Miocene evolutionary stage (Hámor and Hertelendy 1996). The foothills of the earlier uplifted mountainous area are characterized by thick alluvial fans. During the 2.6 Ma-long Pleistocene seven cold climatic intervals (glacial events) were identified, which were separated by climatic periods showing similarities to the recent climate (interglacials). The glacial intervals were characterized by strong winds resulting in deflation and deposition of 20 to 100 m thick loess. The loess strata are divided by interglacial paleosoils. Nádor et al. (2000) suggest that the Pannonian Basin contains the thickest continental Pleistocene sedimentary record in Europe. Based on high-resolution sedimentological data for the 400-800 m thick fluvial section of the central part of the Pannonian Basin, they concluded that sandy layers characterize warm climatic intervals while shaly sediments represent cold climatic periods. Sporadic basalt volcanism resulted in lava flows, basalt tuffs, and filled by alginite and bentonite.

Volcanic stages versus time and space

Based on borehole and outcrop data, the Late Oligocene to Quaternary volcanism of the Carpathian–Pannonian region can be subdivided into four stages, which are related to the thermal mechanic evolution of the crust/mantle systems. The volcanic rocks and the volcanoclastic sediments in several parts of the CPR are covered by thick units of postvolcanic sediments. Limited number of data have been published concerning the petrography, chemistry and age of volcanic rocks known from boreholes. As a consequence the genetics, spatial and temporal distributions of buried volcanic rocks are more uncertain than those of the outcropping volcanic rocks. Figures 10–12 show the generalized distribution of acidic, intermediate and basaltic volcanic rocks in space and time.

1. *Crustal origin, areal acidic volcanism* is represented by tuffs, rhyolites, rhyodacites, ignimbrites, dome/flow complexes. It is the first volcanic activity in most interior areas of the Carpathian Pannonian Region (CPR), with increased content in incompatible elements and relatively high light/heavy rare element ratio (Pantó 1981; Lexa et al. 1995). The total volume of the acidic volcanic rocks is five times greater than that of the intermediate and basaltic material (Póka 1988). It has variable K_2O content and K_2O/SiO_2 ratio. They contain three large and a fourth thinner tuff horizons (Fig. 10): "lower rhyolite tuff" (Hámor et al. 1976; Árva-Sós et al. 1983; Póka 1988) at about 21 Ma (Eggenburgian–Ottnangian boundary), "middle rhyolite tuff" at about 17 Ma (Karpatian–Badenian boundary), "upper rhyolite tuff" at about 14 Ma (Badenian–Sarmatian boundary) and the (fourth) thin Lower Pannonian (11 Ma) rhyolite tuff (Radócz and Jámbor 1969). Eruption of rhyolites was associated in space and time with major phases

of crustal stretching. Magmatic centers were located along major tensional features, (opening ruptures) in the upper brittle crust (Fig. 10). Major changes in stress field orientation led to generation of new set of ruptures and resulted new rhyolite horizon. The appearance of the lower rhyolite tuff indicates the beginning of the synrift phase of Pannonian Basin evolution. The pyroclastic material of the lower rhyolite horizon flowed out through ruptures while middle

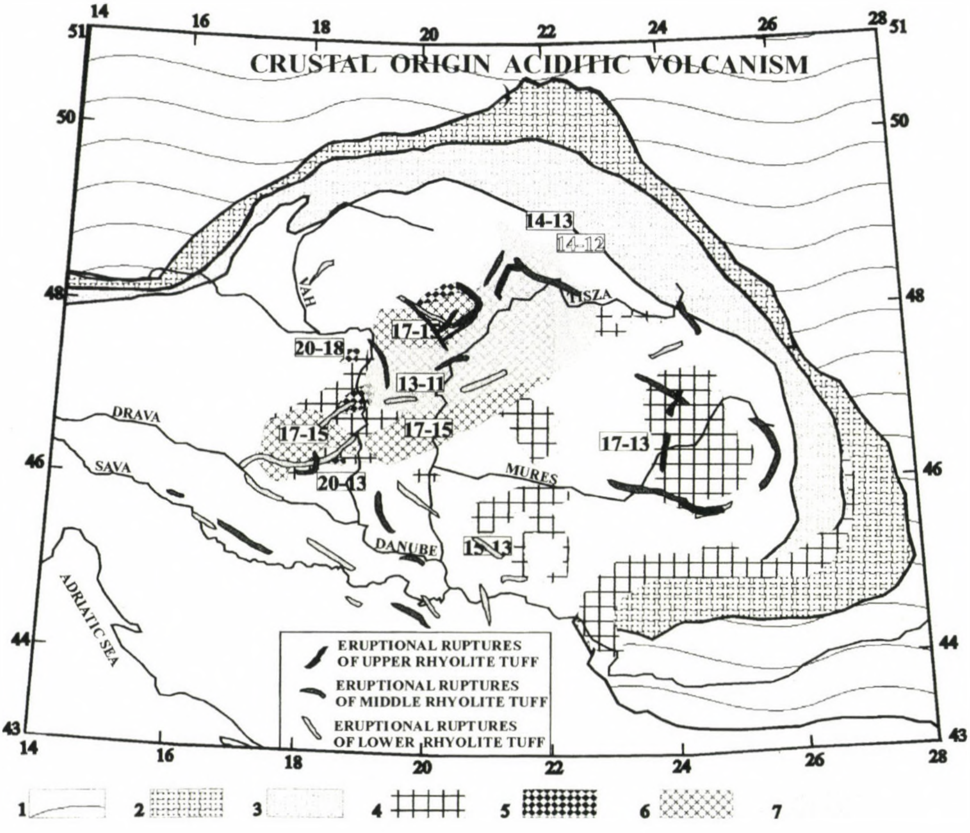

Fig. 10

Space and time distribution of crustal origin, areal acidic volcanism within the Carpathian-Pannonian Region. Collected from Balkay (1962), Radócz and Jámbor (1969), Kubovics and Pantó (1970), Embey-Isztin (1976), Hámor et al. (1976 1988), Árva-Sós et al. (1983), Balogh et al. (1983 1994), Póka (1988), Nemcok and Lexa (1990), Pantó (1991), Széky-Fux and Pécskay (1991), Szabó et al. (1992), Embey-Isztin et al. (1993), Simunić and Pamić (1993), Lexa et al. (1993 1995), Puskás et al. (1993 1998), Pécskay et al. (1995a, b), Downes et al. (1995), von Blanckenburg and Davis (1995, 1996), Pamić and Pécskay (1996), Rosu et al. (1997), Nemcok et al. (1998), Konecny et al. (1999), Pécskay and Balogh (2000), and Rasimic-Saric et al. (2000). Size and location of eruptional fissures based on Hámor (1995). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. undifferentiated acidic volcanism; 5. Lower rhyolite tuffs; 6. Middle rhyolite tuffs; 7. Upper rhyolite tuffs rhyolites were rather blown out. The third (Sarmatian) rhyolite eruption event covered the entire CPR with fine-grained (dust-like) tuff. The thickness of the Lower Pannonian rhyolite tuff is as high as 5 m in the Tokaj Mountains and is only some millimeters in the Mecsek Mountains. The presence of thin Lower Pannonian rhyolite tuffs was identified in several wells in the Danube Basin, Transdanubia, Great Hungarian Plain, and the Mecsek Mts. Petrologic and chemistry data indicate that the Eggenburgian-Early Pannonian dacitic to rhyolitic volcanism was of crustal origin with subordinate mantle component (Póka 1988; Salters et al. 1988; Széky-Fux and Pécskay 1991; Nemcok et al. 1998).

The older part of the volcanic series is characterized by a different paleomagnetic declination from the present North, indicating that the Neogene

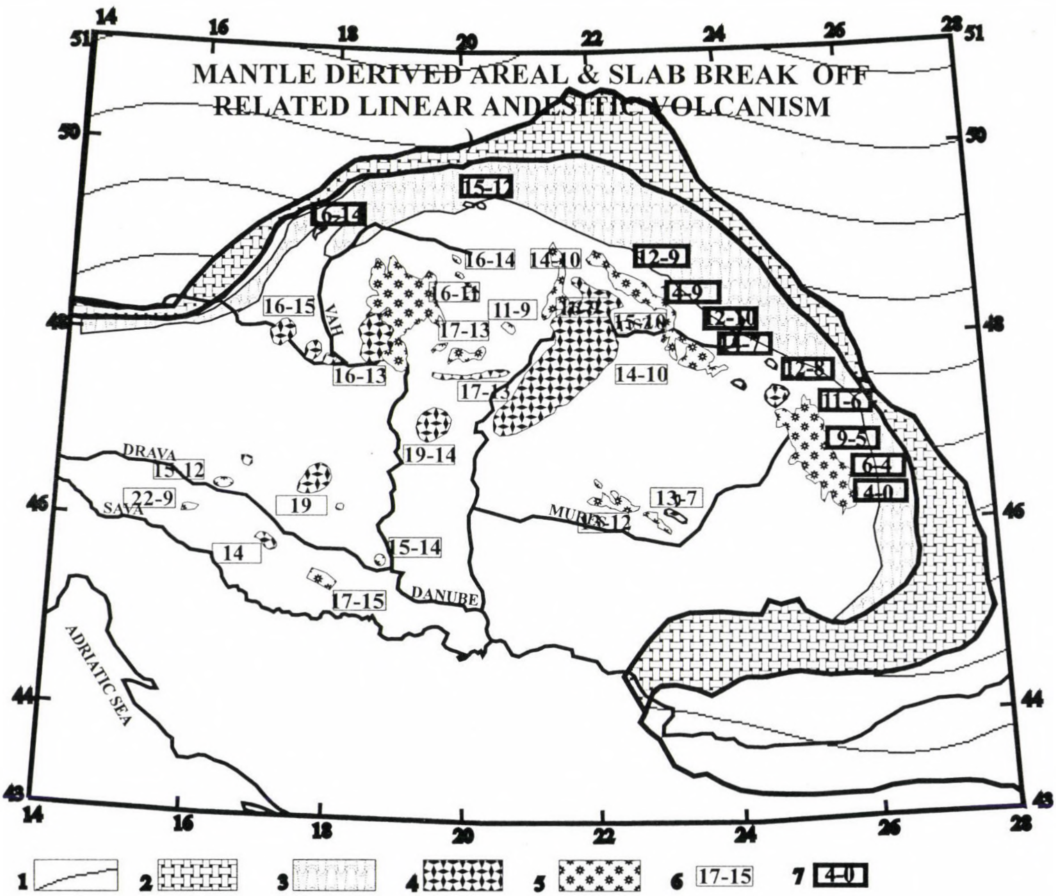

Fig. 11
Space and time distribution of mantle-derived, areal and linear intermediary volcanism within the Carpathian-Pannonian Region (for data sources see references in Fig. 10). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. intermediate calcalkaline magmatism; 5. intermediate stratovolcanoes; 6. timing of areal andesitic volcanism; 7. timing of linear andesitic volcanism

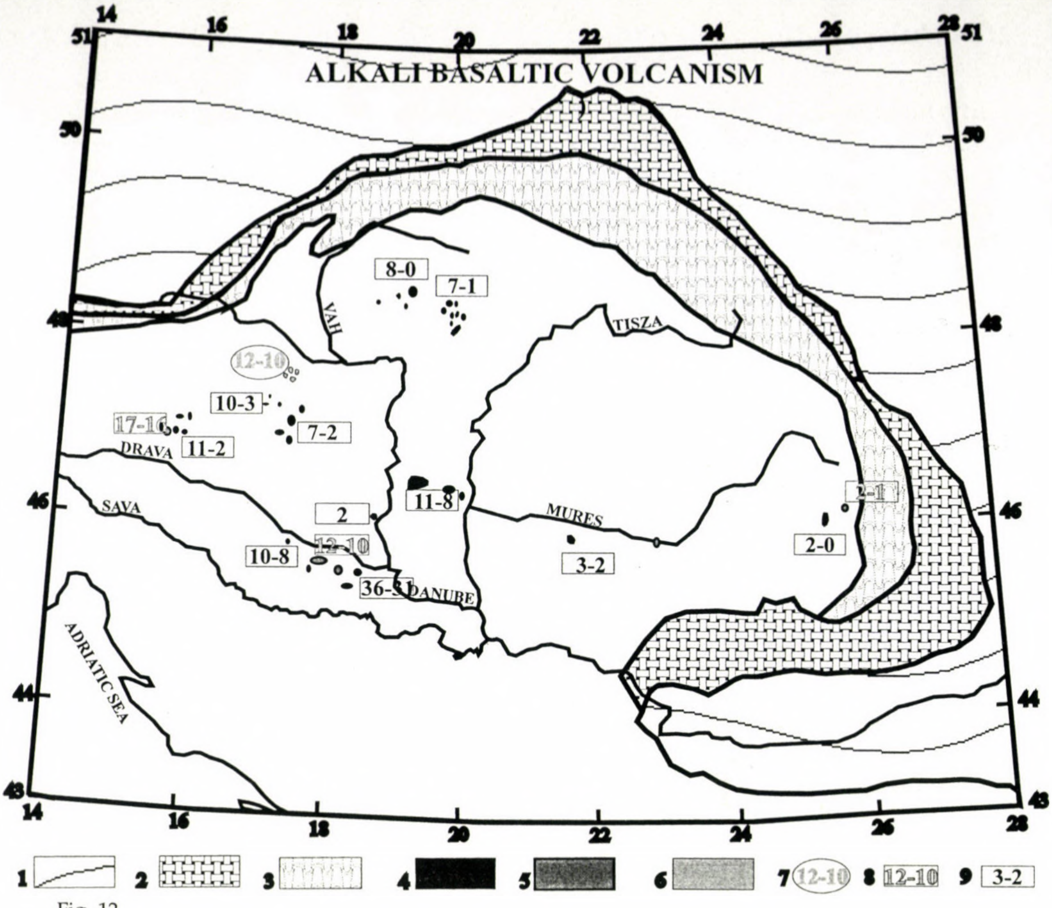

Fig. 12

Space and time distribution of alkali basalt volcanism (not effected by subduction) within the Carpathian-Pannonian Region (for data sources see references on Fig. 10). Legend: 1. East European and Moesian Platform; 2. Alpine and Carpathian foredeep molasse basins; 3. Carpathian flysch nappes; 4. alkali basalt; 5. shoshonites; 6. potassium-rich trachites; 7. age of potassium-rich trachites; 8. age of shoshonites; 9. timing of basaltic volcanic activity

volcanism began before the final emplacement of the tectonostratigraphic terranes of the Carpatho-Pannonian area (Márton and Márton 1989, 1996). Rotation seems to have ended with the onset of the extensive andesitic volcanism, with the exception of the northeastern segment of the Inner Carpathian volcanic belt, where the last significant movement is as young as Early Pannonian.

There is a strong difference between the volcanism of the Pannonian Basin and the East Carpathian arc. In the Pannonian Basin eruption of acidic magmas took place at the same time as the eruption of intermediate magmas. Coulon et al. (1980) have suggested that in Sardinia the Neogene acidic magmas are derived from intracrustal melting and are of different origin from the intermediate ones. They suggest that a secondary magma chamber containing the andesitic magma raised temperatures within the crust to cause partial melting and formation of acidic magma. A similar scenario had been suggested for the Pannonian Basin by Póka (1988): intense intermediate volcanism may have heated the crust beneath the basin enough to cause crustal melting. This acidic melt might have interacted with much less acidic "parent magma". This idea is consistent with the increase in acidity and in potassium content in the intermediate magmas from the beginning of magmatic activity until Late Sarmatian (Early Pannonian) time. After the termination of acid volcanism the magmatism became abruptly more basic and weaker in potassium (Póka 1988). This change in chemistry of magmatism was synchronous with the initial rapid subsidence within the Pannonian Basin (Royden 1993).

2. *Mantle-derived areal andesitic volcanism* (Fig. 11) began in most areas later (Ottnangian-Pannonian; 19–7 Ma) than the acidic volcanic activity (Balogh et al. 1994; Pécskay et al. 1995a, b; Pamić and Pécskay 1996; Pécskay and Balogh 2000). It usually begins with a shallow marine volcano sedimentary unit which is followed by the development of terrestrial stratovolcanoes (Balkay 1962). Stratovolcanoes represent the most intense phase of volcanic activity, which was followed by caldera collapsing and less intense subvolcanic intrusions. Areal andesitic volcanism is characterized by high to medium K-calcalkaline content with high light/heavy REE ratio (Pantó 1981; Lexa et al. 1993; Puskás et al. 1993 1998), which is similar to the character of continental margin calcalkaline volcanics. The geochemical character of Ottnangian–Pannonian (19–7 Ma) andesitic volcanism points to subduction-related mantle-derived magmas (Nemcok et al. 1998). Salters et al. (1988) have suggested that the calcalkaline magmas of the Pannonian realm are mantle-derived by contamination with crustal material (mantle melt and crustal contaminants). Based on petrographic and isotopic data Salters et al. (1988) concluded that the most acidic members of the calcalkaline series (rhyolite) contains up to 67% of crustal material. The least contaminated end member contains only 10% of crustal material. Location of mantle derived areal andesitic magmatism in the Pannonian Basin system shows slight SW to NE and W to E migration through time it might be related to retreat and rollback of the subducted slab. Although Tamburelly et al. (2000) suggested that the presence of potassium-rich magmas of the Apennine should only be taken as evidence that subduction of the lithosphere has occurred beneath that zone, and the uprise of magmas through the brittle upper crust could only occur when extensional tectonic forces significantly attenuated horizontal stresses in the crust. Nemcok et al. (1998) suggest that the spatial distribution of calcalkaline volcanism overlaps with the region of the asthenosphere bulge and extension in the Pannonian Basin. According to Nemcok and Lexa (1990) and Bielik (1991) there is a spatial and temporal relationship between the volcanic activity and the "basin and range"-type extensional structures of the Pannonian Basin, and the most mafic rocks accompany sites with more rapid basin subsidence.

3. Linear zone of basaltic andesite to andesite volcanism (evolving island arc volcanism?) is aligned to the contact of the Outer and Inner Carpathians (Lexa et al. 1993; Nemcok et al. 1998). It began in Eastern Moravia during the Early Sarmatian (13.5–12 Ma). According to detailed studies of Pécskay et al. (1995a, b) in the Vihorlat–Gutin–Gutii mountain ranges this activity lasted throughout the Late Sarmatian and Early Pannonian (13–9 Ma). This period of activity resulted in subvolcanic bodies in the Tibles–Radna Mts. during Early Pannonian (11.9–8.3 Ma). Volcanoes of the Calimani–Harghita area evolved during the Late Pannonian to Quaternary (9–0.1 Ma). Among the well-defined segments of migrating activity along the arc there is only little overlap. The individual segments are characterized by short period of activity.

Linear volcanism is represented by stratovolcanoes with common undifferentiated and rare differentiated rocks. This series has the geochemical character of an evolving island arc volcanism (Lexa et al. 1993); however, it varies in detail according to the crustal composition. Nemcok et al. (1998) interpreted the arc-type andesite volcanism as subduction slab break-off-related volcanism. The volcanic activity is contemporaneous or slightly younger than the final thrusting in the given segment (compare Figs 1 and 12). A similar mechanism was proposed for the Periadriatic slab detachment. Where the slab detachment occurred, rifting resulted in asthenospheric upwelling into the rift, and the resulting thermal perturbation led to melting of the metasomatized overriding mantle lithosphere, producing nearly linear magmatism (von Blanckenburg and Davies 1995).

The Insubric to Periadriatic calcalkali andesites and basaltic andesites are much older than the linear volcanic arc of the Carpathians. The age of this magmatism ranges from 43 to 24 Ma, displaying a pronounced maximum between 33 and 29 Ma (von Blanckenburg and Davies 1995, 1996; Nemcok et al. 1998). This magmatic ranges disappear beneath the Pannonian Basin fill and may be identical with intermediate calcalkaline volcanics of late Eocene to Oligocene age present along the Central Hungarian Fault Zone (Báldi and Báldi Beke 1985; Csontos et al. 1991, 1992).

The Egerian–Eggenburgian age (22.8–19.7 Ma) andesite and basaltic andesite form a 75 km-long linear zone in the easternmost part of the Periadriatic Line (Simunić and Pamić 1993). These andesitic rocks contain xenoliths of basalt, and their $^{87}Sr/^{86}Sr$ ratio indicates a mixing of mantle and crustal isotopic values. These andesites may be considered as the continuation of calcalkaline magmatics, located along the Insubric, Pustertal and Gailtal fault zones (von Blanckenburg and Davies 1995).

4. Alkali basalt (not effected by subduction) volcanism. Sodic-type alkali basaltic volcanism (including shoshonitic, K-trachytic and ultrapotassic rocks) represents the final, Late Sarmatian–Quaternary (11–0 Ma) stage of magmatism in the Pannonian Carpathian Realm (Fig. 12). Their total volume is small. Low-viscosity lavas erupted along faults and formed maars, diatremes, flood basalt and cinder/spatter basaltic cones, necks, and dykes. The duration of alkali basaltic volcanism was relatively long (5–10 Ma). Alkali basalt activity started in different places during Pannonian times (11.5–3.5 Ma) in the Carpathian–Pannonian area and ended in the Quaternary, also in different locations. Lower Pannonian basalt was encountered in several wells (Üllés, Kecel, Ruzsa, Besenyszög, and Nagykörü). Upper Pannonian (7–3 Ma) basalt forms scenic basalt cones, hills in the SW Balaton Highland, S Danube basin, Salgótarján–Filakovo area etc. Pleistocene basalt of the Kemeneshát and Salgótarján–Filakovo area is sodium alkali basalt, while the Pleistocene Bar ones are ultra potassium-rich.

Lherzolite and dunite inclusions within alkali basalt indicate a mantle origin (Embey-Isztin 1976). Phase equilibrium of minerals in inclusions indicate that the basaltic magma originated at a depth of about 70–80 km (Póka 1988). The basaltic magmas were not effected by subduction (Salters et al. 1988; Embey-Isztin et al. 1993; Lexa et al. 1993).

According to Szabó et al. (1992) the alkali basalt of the Carpathian–Pannonian area is not strongly related to the thrusting events of the Carpathian loop and is probably independent of the development of the Pannonian Basin. They suppose that alkali basalt was formed along fault zones cutting through the entire thickening, cooling and increasingly rigid crust, and was related to a kind of passive rifting processes. Variation in composition (becoming more silica-undersaturated and relative alkali-rich) could reflect the level to which the cooling asthenospheric dome (Dövényi and Horváth 1988) has ascended. These faults were reactivated, previously active tectonic lines or ones related to the beginning young structural inversion (Horváth and Gerner 1993; Lenkey 1999; Gerner et al. 1999).

Discussion and conclusions

The paleogeography of the Carpathian Pannonian Region was controlled by the interaction of extension and subsidence within the Pannonian basin systems, compression and uplifting in the orogenic Carpathian thrust belt, and lithospheric bending in response to loading of thrust sheets within the Carpathian foreland. Subsidence and eustacy determined the accommodation space for sediment accumulation. Uplifting, eustacy and climate controlled the erosion of the drainage area, the rate of sediment supply, and how much of the accommodation was filled. In the early phase of Pannonian Basin evolution the overall rate of creation of accommodation exceeded the rate of sediment supply, so water depth increased. The basin-wide average accumulation rate was 3.3 cm/100 years during the Early Miocene and it increased to 7.8 cm/100 years during the Late Miocene. The time of maximum water depth occurred during the Early Pannonian (10 Ma). Prograding seismic reflection patterns indicate more than 1000 m water depth in Early Pannonian time. In a later stage of basin evolution (Late Pannonian) the overall rate of (postrift thermal) subsidence was less than the rate of sediment supply and water depth decreased.

Extension inside the Carpathian loop and compression (thrusting and folding) of the orogenic belt were related to three factors: collision of the Adriatic Plate, the European Platform and several microplates (tectonostratigraphic terranes); evolution of subduction (rollback, hinge retreat, migration of slab detachment) along the Carpathian Arc; mantle upwelling (asthenospheric doming).

There was a causal (but not necessarily plausible) connection between paleogeographic evolutionary stages and subduction and/or mantle upwelling-related volcanic activity. The eruption of crustal-origin acidic rhyolites was associated in space and time with major phases of crustal stretching. Acidic magmatic centers were located along major tensional ruptures (Fig. 10). Major changes in stress field led to the generation of a new set of ruptures and resulted in new rhyolite horizons. The appearance of the lower rhyolite tuffs indicates the onset of the synrift tectonic phase. Mantle-derived areal andesitic magmatism in the Pannonian basin system overlaps with the region of the asthenospheric bulge and extension. The linear zone of basaltic andesite to andesite volcanism aligned with the contact of the Outer and Inner Carpathians is interpreted by Nemcok et al. (1998) as subduction slab break-off-related volcanism. The volcanic activity was contemporaneous with or slightly younger than the final thrusting in the given segment (Fig. 1 and Fig. 12). A similar mechanism was proposed for the Periadriatic slab detachment. Where the slab detachment occurred, rifting resulted in asthenospheric upwelling into the rift, and the resulting thermal perturbation led to melting of the overriding mantle lithosphere, producing nearly linear magmatism (von Blanckenburg and Davies 1995). The eruption of alkali basaltic magmas was the youngest volcanic activity in the Pannonian Basin system. Isotope studies indicate that it was not effected by subduction (Salters et al. 1988; Embey-Isztin et al. 1993; Lexa et al. 1993) and according to Szabó et al. (1992) it might have independent of the development of the Pannonian Basin system.

The size and shape of (marine, fluvial, coastal, etc.) facies belts of each paleogeographic evolutionary stage were strongly connected to the dominant structural style of that stage. The Early Miocene evolutionary stage was characterized by a system of long and narrow grabens connected by XXX, and connected to the Carpathian thrust belt by transform faults. Grabens within the area of the Pannonian Basin were characterized by a fluvial depositional environment (Fig. 2). The NW–SE-oriented marine (flysch and molasses) facies belts of the subducting "Magura Ocean" were almost perpendicular to the direction of the narrow fluvial grabens of the southwestern part of the Carpathian Pannonian region.

Rollback and hinge retreat of subducting slab contributed to the rapid consumption (subduction) of these NW–SE-oriented marine (flysch) facies belts (Fig. 4). The Middle Miocene paleogeographic evolutionary stage was characterized by a wide zone of evaporitic lagoons at the inner basinward side of the emerging Carpathian thrust belts. Syn-orogenic subsidence took place within the compressional foreland basins of the Carpathian Arc, while pull-apart along transform faults led to deepening and widening of the formerly narrow fluvial grabens and to the opening of several new rifts and/or pull-apart basins within the Pannonian Basin system. Overall the Middle Miocene paleogeography was dominated by wide marine facies belts separated by areas of erosion and non-deposition. Termination of the synrift phase in Late Middle Badenian time (14 Ma) is indicated by a structural inversion-related regional unconformity.

The Late Miocene paleogeographic stage was dominated by the rate of postrift thermal subsidence in the Pannonian Basin systems and partly by rate of erosion within the thrust belts. Emerging thrust belts provided sediments for the outer Carpathian foreland basins and for the Pannonian Basin as well. The great amount of erosional debris was transported to both basin systems by rivers. During the Late Badenian and Sarmatian the rate of subsidence was greater than the sediment supply rate. Transgression from the Aralo-Caspian area northwestward can be demonstrated within the entire region. Reef limestone was deposited at the margins of the uplifted areas and in the sub-basins as well. During Late Pannonian time the rate of sediment supply exceeded the rate of creation of accommodation space. Southward-prograding, large rivers dominated delta systems, which controlled the paleogeography of the Pannonian Basin system. Well-developed third-order sequences (consisting of lowstand prograding wedges, backstepping transgressive systems tracts, highstand prograding system tracts) are the building blocks of the thick postrift unit. Progradational delta filling of the Pannonian Basin ended during Late Pliocene (Fig. 8), and was followed by continental and fluvial sedimentation (Fig. 9).

Pleistocene paleogeography was controlled by climatic factors. The foothills of the uplifting mountainous area were characterized by thick alluvial fans. Seven cold climatic intervals (glacial events) were identified which were separated by interglacials. During glacial intervals thick loess layers were deposited. The loess strata are separated by interglacial paleosoils.

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Neogene to Quaternary volcanism of the Carpathian–Pannonian Region – a review

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The Neogene to Quaternary evolution of the Carpathian–Pannonian Region (CPR) was accompanied by various volcanic events. The volcanic products can be classified into four groups based on their compositional characteristics: (1) Miocene silicic volcanism; (2) Miocene to Pliocene potassic and ultrapotassic volcanism; (3) Miocene to Quaternary calc-alkaline volcanism and (4) Late Miocene to Quaternary alkaline volcanism.

The Miocene silicic volcanism could be related to complex tectonic processes including major block rotations and accompanying strike-slip tectonic movements (Early Miocene explosive volcanism) and thinning of the continental lithosphere (Middle Miocene explosive volcanism). Melt generation occurred in the upwelling asthenosphere and the mantle-derived magmas mixed with induced lower crustal melts. The potassic and ultrapotassic volcanic rocks were generated as a response of the overall extension of the Pannonian Basin. Contrasting origins are suggested for the Miocene to Quaternary calc-alkaline volcanic suites of the CPR based on the different space and time distribution and geochemistry of the volcanic rocks and the different deep structure beneath the volcanic complexes. The Middle to Late Miocene calc-alkaline volcanic rocks in the Northern Pannonian Basin were formed due to lithospheric extension following the extensive metasomatism of the lithospheric mantle during the Paleogene to Early Miocene flat subduction. The lithospheric mantle-derived magmas were contaminated by metasedimentary lower crust. In contrast, a direct relationship is proposed between the Late Miocene to Quaternary calc-alkaline volcanism and subduction and slab break-off processes along the Eastern Carpathians. The Miocene calc-alkaline volcanic rocks in the southern part of the Pannonian Basin could have been formed by partial melting of the lithospheric mantle due to lithospheric thinning. The Late Miocene to Quaternary post-extensional alkaline volcanism is explained by upwelling of a relatively hot EAR-type asthenospheric material beneath the abnormally thin lithosphere of the Pannonian Basin.

Key words: silicic volcanism, potassic and ultrapotassic volcanism, calc-alkaline volcanism, alkaline basalts, subduction, extension, Carpathian–Pannonian Region

Introduction

During the last 10–15 years, significant progress has been made in understanding magma generation processes in the mantle and in the crust (e.g. Huppert and Sparks 1988; McKenzie and Bickle 1988; Gallagher and Hawkesworth 1992; Leeman and Harry 1993; Wilson 1993; Black et al. 1997). Integrated studies including interpretation of the geochemical data of the volcanic rocks and the geophysical data of the area as well as the theoretical

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models on the magma generation could result in a better understanding of geodynamic evolution of a region (e.g. Fitton et al. 1991; Hawkesworth et al. 1995; Zhang et al. 2001). The Carpathian–Pannonian Region (CPR) is characterized by various Neogene to Quaternary volcanic products from strongly undersaturated nephelinites through ultrapotassic rocks to high-silica rhyolites. These volcanic rocks were generated at different stages during the evolution of the Pannonian Basin; therefore they are of great importance in the reconstruction of the geodynamic processes of this region. During the last decade, a large amount of new data has been published by the PANCARDI Igneous Team on the volcanology and geochemistry of these volcanic rocks. Based on these data, new models have been developed; most of them were discussed in the last PANCARDI Meetings. This paper attempts to summarize our present knowledge on the Neogene to Quaternary volcanism of the CPR, mainly based on the results published during the last decade. It follows the classic reviews published previously by Póka (1988) and Szabó et al. (1992) and is rather geochemistry-oriented compared to that published by Lexa and Konečný (1999). Although this review attempts to show most of the recently published models on the genesis of these volcanic rocks as objectively as possible, it cannot avoid being strongly influenced by the author's ideas on the Neogene to Quaternary volcanism of the CPR.

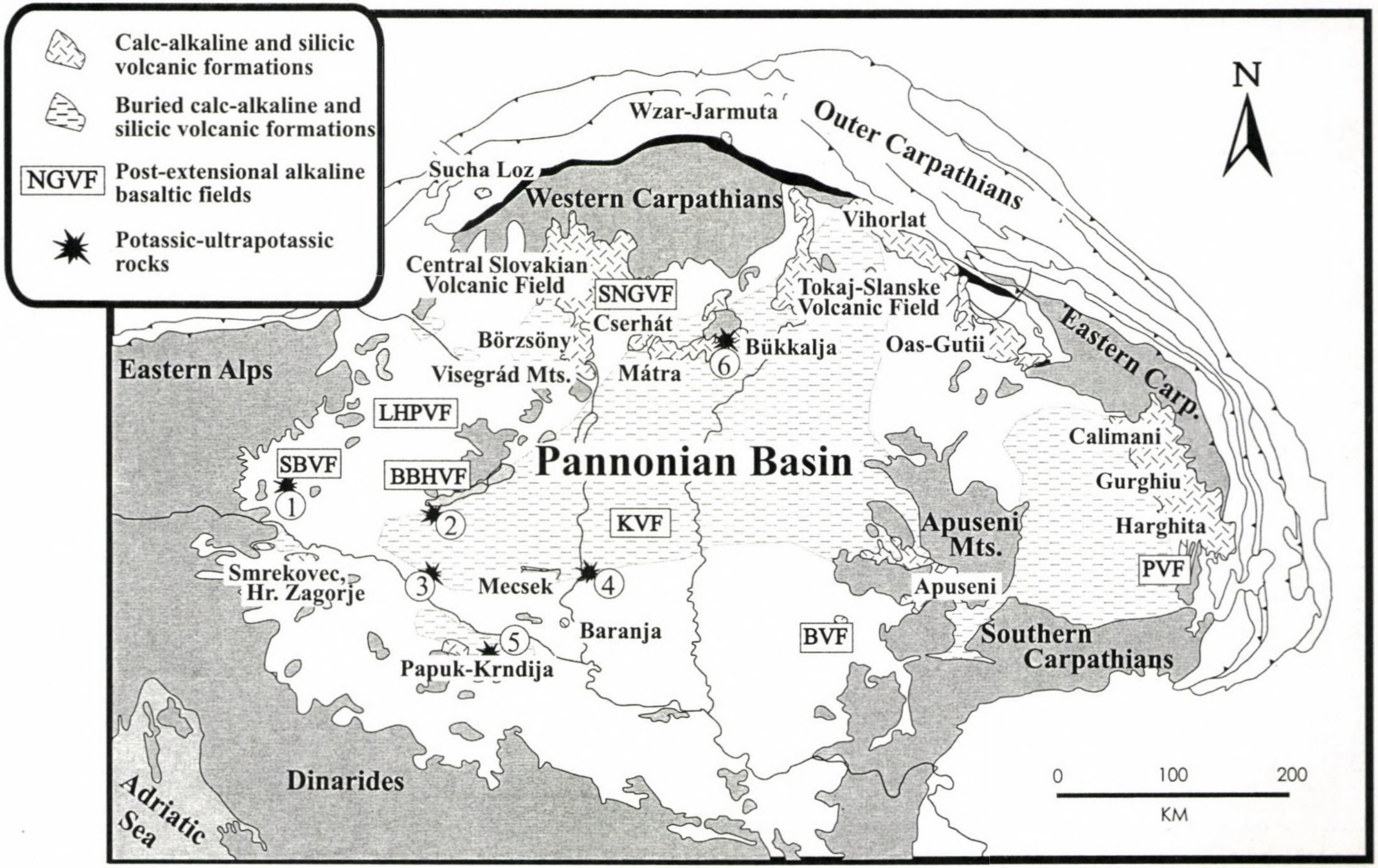
Space and time distribution

Neogene volcanic rocks of the CPR can be classified into four main groups based on compositional characteristics: (1) Miocene silicic volcanism; (2) Miocene to Pliocene potassic and ultrapotassic volcanism; (3) Miocene to Quaternary calc-alkaline volcanism and (4) Late Miocene to Quaternary alkaline volcanism. These volcanic rocks cover large areas of the CPR, although the majority of them are buried by young (mostly Late Miocene to Quaternary) sedimentary formations (Fig. 1).

Miocene silicic volcanism

The first volcanic products during the evolution of the Pannonian Basin were silicic pyroclastic rocks originated by voluminous explosive eruptions. Lexa and Konečný (1999) called these volcanic rocks "areal type silicic volcanics". This volcanic activity occurred repeatedly during the Miocene (Fig. 2) and resulted in a large volume of unwelded and welded pumiceous pyroclastic flow deposits

Fig. 1 →
Distribution of the Neogene to Quaternary volcanic rocks in the Carpathian-Pannonian Region. Post-extensional alkaline basaltic volcanic fields: SBVF – Styrian Basin Volcanic Field; LHPVF – Little Hungarian Plain Volcanic Field; BBHVF – Bakony-Balaton Highland Volcanic Field; SNGVF – Štiavnica-Nógrád-Gömör Volcanic Field; KVF – Kecel Volcanic Field; BVF – Banat Volcanic Field; PVF – Persany Volcanic Field



Potassic-ultrapotassic rocks: 1. Styrian Basin; 2. Balatonmária; 3. Szenta; 4. Bár; 5. Mt. Krndija; 6. Bükkalja

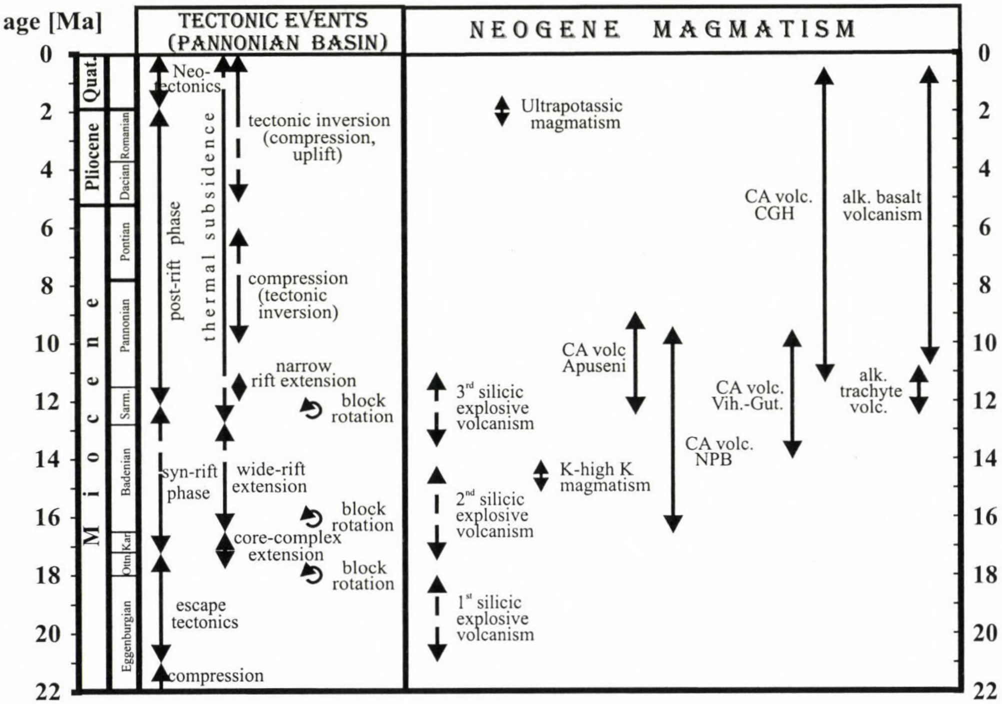


Fig. 2
Relationship between Neogene to Quaternary geodynamic and tectonic events and magmatism in the Pannonian Basin based on the data of Balogh et al. (1986, 1994a), Fodor et al. (1999) and Tari et al. (1999), NPB – Northern Pannonian Basin; CGH – Calimani-Gurghiu-Harghita; CA – calc-alkaline

(ignimbrites) and related pyroclastic fall layers (e.g. Pantó 1962; Capaccioni et al. 1995; Szakács et al. 1998; Harangi et al. 2000). In the Hungarian literature, the Miocene silicic volcanic series have been divided into three main units (e.g. Hámor et al. 1980; Ravasz 1987; Póka 1988), i.e. "Lower Rhyolite Tuff" (LRT) or "Gyulakeszi Rhyolite Tuff Formation" thought to be formed at the Eggenburgian/Ottnangian boundary (19.6±1.4 Ma; Hámor et al. 1980), "Middle Rhyolite Tuff" (MRT) or "Tar Dacite Tuff Formation" that could have been formed at the Karpatian/Badenian boundary (16.4±0.8 Ma; Hámor et al. 1980), and "Upper Rhyolite Tuff" (URT) or "Galgavölgy Rhyolite Tuff Formation" considered to have been formed during the Late Badenian to Sarmatian (13.7±0.8 Ma; Hámor et al. 1980). More recent studies (e.g. Oláh and Harangi 2000), however, questioned this rigid classification and suggested nearly continuous multiple eruptions of silicic magmas from different volcanic centers during the Miocene. The LRT extends along a southwest (north of the Mecsek Mts) to northeast (south of the Bükk Mts.) belt roughly parallel with the Mid-Hungarian Tectonic Zone (Ravasz 1987; Póka 1988; Széky-Fux et al. 1991). The MRT has a much wider areal distribution, extending over the entire Pannonian Basin. In the basement of the Great Hungarian Plain (eastern Pannonian Basin) boreholes penetrated several hundreds of meters thick MRT sequences. Some parts of the MRT are known as "Hrabovec Tuff" in eastern Slovakia and "Novoselica Tuff" in the Ukraine (Lexa and Konečný 1999). They are mostly reworked volcaniclastic rocks formed in a marine environment (Lexa and Konečný 1999). In spite of the large areal distribution of the MRT, we do not know exactly the locations of the eruptive centers, although Hámor (1998) suggested several ones where thick volcaniclastic series are found. The Dej Tuff covering the Transylvanian Basin interbedding within Miocene sedimentary sequences originated between the formation of the MRT and the URT (14.8–15.4 Ma; Popescu 1970; Szakács et al. 2000). It has an extremely variable thickness and contains several independent flow and fall units as well as reworked beds (Szakács et al. 2000). Szakács et al. (2000) suggested that the rhyolitic Dej Tuff volcaniclastic sequence could have been derived either from the Oas–Gutii volcanic complex or from the Ukrainian Carpathians. In addition, Badenian volcaniclastic rocks can be found even at the foredeep of Southern Carpathians. The third silicic volcanic unit (URT) is confined mainly to the eastern-northeastern part of the Pannonian Basin. The possible volcanic centers could be in the Tokaj, Beregovo and Oas–Gutii volcanic fields.

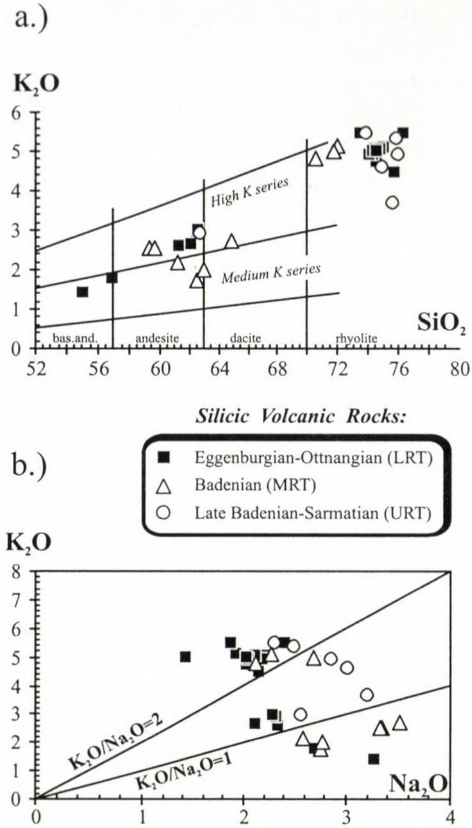


Fig. 3
Classification of the Miocene silicic volcanic rocks of the CPR based on the a) SiO_2 and K_2O and b) K_2O and Na_2O contents

These volcaniclastic horizons are separated by fossiliferous sediments in the basement of the basins, but it is very difficult to distinguish them in other areas. In the southern slope of the Bükk Mts. (Bükkalja region), all of these silicic volcanic products covering a formation age between 21 Ma and 13.5 Ma (Márton and Pécskay 1998) crop out. Therefore, this region provides a unique opportunity to study the volcanology, petrology and geochemistry of these rocks (e.g. Capaccioni et al. 1995; Póka et al. 1998; Szakács et al. 1998; Harangi et al. 2000).

The three volcaniclastic horizons can be well distinguished in this area based on paleomagnetic data (Márton and Pécskay 1998), because two major rotations occurred during the period of volcanic activity (Márton and Fodor 1995; Márton and Márton 1996). The LRT is characterized by 80–90° counterclockwise rotations, whereas the MRT rotated about 30° counterclockwise. No rotation has been detected for the URT deposits. Although paleomagnetic data are powerful tools in correlating the volcaniclastic horizons of the Bükkalja region, it appears that they cannot be used effectively at a larger scale (Márton 2000, personal communication). Therefore, one of the major challenges in the study of the silicic volcanic rocks of the CPR is to find tools to correlate the scattered deposits extending over the Pannonian Basin (Lukács 2001). The K/Ar radiometric age data do not appear to have enough resolution to distinguish clearly the main volcaniclastic horizons (e.g. Márton and Pécskay 1998). Combined physical volcanology, zircon morphology, petrographic, mineral and bulk rock geochemical studies, however, can be effectively used to compare volcanic units (e.g. Harangi et al. 2000; Lukács 2001). Zircon is a common accessory mineral in silicic igneous rocks. Its morphology reflects the type of host magma (anatectic, hybrid or mantle-derived) and the crystallization history of the silicic magma (Pupin 1980; 1988). Recently, zircon typology studies have been carried out on the Miocene silicic rocks of the Bükkalja region (Szabó 2000; Szabó and Harangi 2000; Harangi et al. 2000) and on the scattered deposits thought to belong to the LRT (Oláh and Harangi 2000; Oláh 2001). These results suggest that the volcaniclastic rocks of the LRT and MRT can be well distinguished based on the morphology of zircons (Fig. 4).

Miocene to Pliocene potassic and ultrapotassic volcanism

Potassic ($K_2O/Na_2O=1–2$ and $K_2O>3$ wt%) and ultrapotassic rocks ($K_2O/Na_2O>2$ and $K_2O>3$ wt%; Foley et al. 1987) are minor among the Neogene volcanic rocks of the CPR (Figs 1 and 2). This type of volcanic rock reflects the most extreme incompatible element enrichment processes acting in the upper mantle. Formation of potassic and ultrapotassic magmas by partial melting of metasomatized, K-rich lithospheric mantle region could be related either to lithospheric stretching or to upwelling of hot asthenospheric material beneath the continental lithosphere (e.g. Thompson et al. 1989). Therefore, the occurrence of these rocks has a strong geodynamic implication. In the CPR, potassic and ultrapotassic rocks can be found in the Styrian Basin (Heritsch 1967; Scharbert et al. 1981; Krainer 1987; Harangi et al. 1995a; Harangi, in press), in the central Pannonian Basin (Balatonmária borehole; Harangi et al. 1995a), in the southern Pannonian Basin close to the Drava Fault (Loncarski Vis and Mt. Krndija; Pamić et al. 1992; 1995) and near the village of Bár (Szederkényi 1980; Harangi et al. 1995a). Most of these rocks were formed contemporaneously during the Middle Miocene (13–17.5 Ma in the Styrian Basin; Balogh et al. 1994a; 13–16 Ma in the

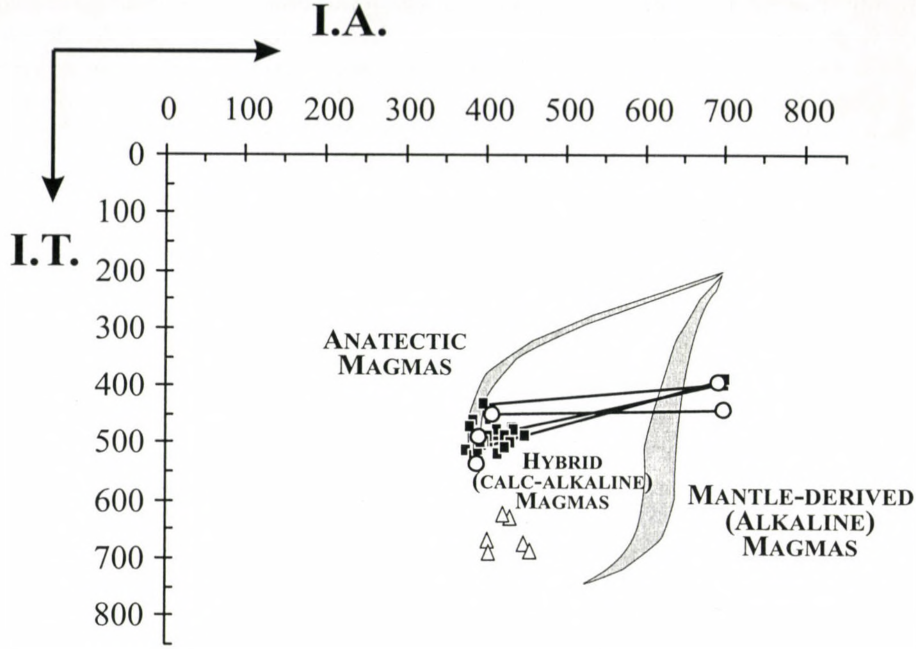


Fig. 4
I.A. (alkalinity index) vs. I.T. (temperature index) diagram (Pupin 1980, 1988) for zircons of the Miocene silicic rocks based on zircon morphology data (Szabó 2000; Oláh 2001). Symbols are as in Fig. 3

central Pannonian Basin; Józsa et al. 1993; Harangi et al. 1995a and 15–17 Ma in the southern Pannonian Basin; Pamić et al. 1995), whereas the olivine leucitite at Bár was formed at 2.1±0.2 Ma (Balogh et al. 1986).

Miocene to Quaternary calc-alkaline volcanism

One of the most striking features of the Neogene volcanic areas of the CPR is the occurrence of large calc-alkaline volcanic complexes along the Carpathian chain. However, they are only the "tip of iceberg", because voluminous calc-alkaline volcanic rocks are buried in the basement of the central and eastern Pannonian Basin. Almost complete, more than a thousand meter-thick stratovolcanic complexes have been recognized by borehole and seismic studies (e.g. Széky-Fux and Kozák 1984; Széky-Fux et al. 1987; Széky-Fux and Pécskay 1991; Zelenka 2001, pers. comm.). Lexa et al. (1993) and Lexa and Konečný (1999) divided the calc-alkaline volcanic rocks of the CPR into two groups: "areal-type" and "arc-type" andesitic groups. In this paper, four main calc-alkaline volcanic regions are distinguished based on areal distribution.

In the northern Pannonian Basin, calc-alkaline volcanic complexes occur roughly perpendicular to the Carpathian Arc, extending as far as 200–300 km from the assumed suture zone (Fig. 1). They are underlain by relatively thin crust (27–30 km) and thin lithosphere (60–80 km; Tari et al. 1999). They include the Visegrád Mts. Börzsöny and the Central Slovakian Volcanic Fields (e.g. Kubovics and Pantó 1970; Konečný et al. 1983; 1995; Harangi et al. 1999; Karátson et al. 2000), Cserhát and the Mátra Mts (e.g. Póka 1968; Kubovics and Pantó 1970; Varga et al. 1975) and the Tokaj–Slanske Volcanic Field (e.g. Gyarmati 1977; Kaliciak and Repcok 1987; Kaliciak and Žec 1995). Lexa et al. (1993) and Lexa and Konečný (1999) called them "areal-type andesite volcanics". In addition, sporadic dyke swarms can be found in the Magura Flysch Zone at Sucha Loz and the Uherské Hradište area (Lexa and Konečný 1999) and in the Mt. Wzar and Mt. Jarmuta area (e.g. Birkenmajer 1984; Youssef 1978; Birkenmajer et al. 1987, 2000; Birkenmajer and Pécskay 1999, 2000). In this region, calc-alkaline volcanic activity was relatively long-lived from about 16.5 to 9 Ma (Pécskay et al. 1995). No temporal variation of volcanism can be observed from west to east and it is remarkable that the volcanic activity started roughly contemporaneously in each volcanic field of this region at 16–16.5 Ma.

Calc-alkaline volcanic complexes in the eastern Pannonian Basin are parallel to the Carpathian orogenic belt and extend 50–100 km from the assumed suture zone (Fig. 1). They are underlain by thin crust and lithosphere (26–28 km and 60–80 km, respectively) in the northern parts, but thick crust and lithosphere (38–42 km and 100–120 km, respectively; Tari et al. 1999) in the southern parts. Lexa et al. (1993) and Lexa and Konečný (1999) called them "arc-type basaltic andesite to andesite volcanics" although they also included the calc-alkaline dyke rocks from the Magura Flysch Zone in this group. These volcanic complexes comprise the Vihorlat-Beregovo volcanic fields (Kaliciak and Žec 1995; Lyashkevich 1995; Pécskay et al. 2000; Seghedi et al., in press), the Oas-Gutii volcanic fields (Kovács et al. 1992; Pécskay et al. 1994; Seghedi et al. 1995) and the Calimani-Gurghiu-Harghita volcanic chain (e.g. Peltz et al. 1974, 1987; Seghedi et al. 1987; Szakács et al. 1993; Szakács and Seghedi 1995; Mason et al. 1996, 1998). The volcanic activity in this region was relatively short-lived (about 4 Myr) at each volcanic complex and a gradual age decrease of the volcanism from 14 Ma to 0.15 Ma can be observed from north to south (Pécskay et al. 1995). This gradual younging is the most remarkable at the southernmost part of the volcanic chain (4.5 Ma to 0.15 Ma; Szakács et al. 1993). The eruption of the Ciomadul volcano in South Harghita (1.0–0.15 Ma; Szakács et al. 1993 or 35–42 Ka; Szakács and Seghedi 2000) was the last volcanic phenomenon in the CPR.

The third group of the calc-alkaline volcanic rocks occurs in the inner part of the Pannonian Basin (Fig. 1). Most of them are buried by thick Late Miocene to Quaternary sedimentary formations in the Great Hungarian Plain, eastern Pannonian Basin (Széky-Fux and Kozák 1984; Széky-Fux et al. 1987; Széky-Fux and Pécskay 1991) and we know little about them in spite of the large volumes.

Sporadic occurrences of basaltic andesite to andesite outcrops were described in the eastern Mecsek Mts. and in the Apuseni Mts (Fig. 1). Both volcanic rocks were termed as "areal type volcanics" by Lexa et al. (1993) and Lexa and Konečný (1999). Andesites in the Mecsek Mts. were formed during the Early Miocene (19–20 Ma; Hámor et al. 1987) and occur as shallow subvolcanic bodies and sills. In the Apuseni Mts., mostly andesitic and subordinately dacitic magmas erupted from 14.5 to 9 Ma, followed by the formation of a basaltic andesite body at Detunata at 7.4 Ma (Lemne et al. 1983; Borcos et al. 1986; Seghedi et al. 1998).

At the southwestern margin of the Pannonian Basin, sporadic Neogene calc-alkaline volcanic rocks occur from the Pohorje to Baranja (Pamić et al. 1995; Fig. 1). Most of these volcanic rocks occur in the Drava Depression along the Drava Fault. Outcrops are very rare but borehole data indicate an up to 1000 meter-thick volcanic succession (Pamić et al. 1995). The volcanic activity in this region commenced in the Hrvatsko Zagorje and the eastern Drava Depression during the Early Miocene (about 18–22 Ma). This was followed by Badenian (13–15 Ma) basaltic to dacitic volcanic activity and eruption of mostly basaltic magmas during the Late Miocene (7–9 Ma) in the Drava Depression and the Baranja area (Pamić and Pikija 1987; Pamić and Sparica 1988; Lugovic et al. 1990; Simunić and Pamić 1993; Pamić et al. 1995).

Late Miocene to Quaternary alkaline volcanism

Late Miocene to Quaternary alkaline volcanism resulted in several isolated volcanic fields in the CPR (Fig. 1) partly overlapping the calc-alkaline volcanic complexes (e.g. Štiavnica–Nógrád–Gömör Volcanic Field and Persany Volcanic Field). This volcanic activity produced almost exclusively basaltic volcanoes (Embey-Isztin et al. 1993; Seghedi and Szakács 1994; Dobosi et al. 1995; Downes et al. 1995b; Embey-Isztin and Dobosi 1995; Harangi et al. 1995b; Konečný et al. 1995b; Németh and Martin 1999; Harangi, in press). The only known exception is the buried trachyandesite to alkaline trachyte stratovolcano at Pásztori in the Little Hungarian Plain (Balázs and Nusszer 1987; Harangi et al. 1995b; Schléder and Harangi 2000; Schléder 2001). The alkaline mafic volcanism began at the western (Burgenland) and central (encountered in boreholes around Kecel) Pannonian Basin in the Late Miocene (10–12 Ma; Balogh et al. 1986, 1994a; Balázs and Nusszer 1987), culminating at 3–5 Ma at each volcanic field (Balogh et al. 1986, 1994a, 1994b; Borsy et al. 198a) and the last eruptions occurred only a few hundreds of thousand years ago at Brehy (Central Slovakia, 0.53 Ma; Balogh et al. 1981) and in the Persany Mts. (0.7–0.8 Ma; Mihaila and Kreutzer 1981). Phreato-magmatic to magmatic eruptions of basaltic magma resulted in various volcanic forms including maars, tuff rings, cinder cones and shield volcanoes (Harangi and Harangi 1995; Konečný et al. 1995b; Németh and Martin 1999; Németh et al. 1999). The trachyandesite-trachyte stratovolcano at Pásztori (Little Hungarian Plain) was penetrated by several boreholes beneath a 2000 meter-thick Late

Miocene to Quaternary sedimentary sequence. Boreholes penetrated a more than 1000 meter-thick volcanic succession, but none of them reached the basement of the volcano. The upper part of the volcanic series is cut by several basaltic dykes and/or lava flows, which formed roughly contemporaneously with the basaltic volcanism in Burgenland (Harangi, in press).

Petrology and geochemistry

Neogene to Quaternary volcanism in the CPR is strongly related to the geodynamic evolution of this region. The geochemistry of magmatic rocks reflects various processes from the nature of the source regions of the primary magmas to deep and high-level magmatic processes (e.g. partial melting, fractional crystallization, contamination, magma mixing, etc.). On the other hand the temporal variation of geochemistry of the volcanic suites could indicate the relationship between the volcanism and the major tectonic processes of the area (e.g. Wilson 1993). During the last decade a considerable amount of new geochemical data obtained by modern analytical techniques has been published for the Neogene to Quaternary volcanic rocks of the CPR. The majority of the silicic and the calc-alkaline volcanic rocks were analyzed at the same laboratory (Royal Holloway University of London; details of the analytical conditions can be found in Mason et al. 1996 and in Harangi et al. 2001); therefore comparisons of these rocks are not affected by interlaboratory bias. In contrast, the alkaline basaltic volcanic rocks were analyzed at different laboratories (University of Edinburgh; Embey-Isztin et al. 1993; University of Florence; Harangi et al. 1995b; University of North Carolina; Dobosi et al. 1995). A comparison of data from different laboratories is discussed by Harangi (in press) suggesting that the interpretations are not influenced by analytical differences.

Among the geochemical data, the First Transition Series Elements (FTSE; e.g. Ni, Cr, Co, V, Sc) are compatible in mafic minerals such as olivine and clinopyroxenes that are early crystallizing phases in basaltic magmas. Therefore, the concentration of these elements drastically drops during the early differentiation of mafic magmas. A high amount of FTSE (e.g. Ni=150–300 ppm, Cr=200–400 ppm), along with a high mg-value ($Mg^{2+}/Mg^{2+}+Fe^{2+}=0.65–0.70$), indicates that the composition of the host rock is close to the primary magma. Incompatible trace elements (e.g. Rb, Ba, Th, Nb, La, Ce, Zr, etc.) combined with radiogenic isotope ratios (e.g. $^{87}Sr/^{86}Sr$ $^{143}Nd/^{144}Nd$ and Pb isotope ratios) are of great importance in igneous geochemistry because they are sensitive indicators of enrichment processes (melting events, mantle metasomatism, crustal contamination, etc.). The Low Field Strength Elements or Large Ion Lithophile Elements (LFSE or LILE; e.g. Rb, Ba, K, Sr, Pb, Cs) are the most incompatible elements under mantle conditions, i.e. they are preferentially concentrated in the melt phase. However, they are soluble in H_2O-rich fluids; therefore LILEs are mobile in secondary processes, such as weathering and also during dehydration of subducting slabs. The High Field Strength Elements (HFSE; e.g. Nb, Ta, Zr, Hf) and the Rare Earth Elements (REE; lanthanides) are variously incompatible and usually immobile elements. Ratios of HFSE/HFSE or REE/HFSE do not change significantly during slight to moderate fractional crystallization; therefore they reflect either the chemistry of the mantle source region or the degree of partial melting. Radiogenic isotope ratios, such as $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ and Pb isotope ratios, are widely used to establish the nature of the source region (i.e. depleted or enriched compared to the Bulk Earth) and to quantify the relative proportions of asthenospheric and lithospheric melts. Crustal contamination of mantle-derived magmas can be revealed by radiogenic and stable isotope ratios (e.g. increase of $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ values) and by studying selected trace elements, which are enriched in the crust (e.g. K, Th).

In the next sections, the main geochemical characteristics of Neogene to Quaternary volcanic rocks of the CPR are described with a special focus on the messages about their origins and the geodynamic implications of the magmagenesis.

3.1 Miocene silicic volcanism

The Miocene silicic volcanic products are represented almost exclusively by volcaniclastic rocks. The primary volcanic products are unwelded and welded pumiceous pyroclastic flow deposits with subordinate pyroclastic fall beds (Capaccioni et al. 1995; Szakács et al. 1998; Harangi et al. 2000). The grain size of the pyroclastic rocks is strongly variable. Block-bearing lapilli tuffs are frequent in the Bükkalja region and northwest of the Mátra Mts. (Simon and Kovács 2001) suggesting proximal facies, whereas ignimbrites in other areas (e.g. Ipolytarnóc, Cserhát, Mecsek Mts.) are usually finer-grained lapilli tuffs and tuffs (Oláh and Harangi 2000). The juvenile components are rhyodacitic to rhyolitic pumices, whereas dacitic scoriae occur in some localities of the MRT in Bükkalja (Póka et al. 1998; Czuppon 2001). Lithic clasts are usually subordinate (<10vol%) and are represented mainly by cognate basaltic andesite and andesite fragments. The geochemical composition of the pumices, fiamme, scoriae and lithic clasts defines a roughly linear trend in the SiO_2 vs. K_2O diagram (Fig. 3a). It is remarkable that a significant compositional gap can be observed between the andesitic lithic clasts and the rhyolitic pumices in each suite. Although the linear trend may suggest cogenetic series derived by a different degree of fractional crystallization, the strong potassium enrichment rather suggests incomplete magma mixing process. The pumices are rich in potassium, whereas the lithic clasts are both sodic and potassic (Fig. 3b). Pumices belonging to the URT are typically less potassic, especially in the eastern Bükkalja region (K_2O/Na_2O=1.2–2.1) than those belonging to the LRT (K_2O/Na_2O=2.1–3.5). The mineral composition of the pumices and the host pyroclastic rocks resembles the I/M-type silicic magmas (i.e. the primary magmas were formed by partial melting either of igneous formations of the crust or the upper mantle) (Harangi et al. 2000). The most dominant phenocrysts in each silicic unit are plagioclases and biotite. Quartz is a frequent mineral in the pyroclastic deposits of the LRT and URT, but is subordinate or even absent in the MRT. One of the most striking features of the MRT silicic volcanic rocks is the occurrence of orthopyroxenes and hornblende. The composition of orthopyroxenes suggests a metaluminuous host magma, whereas the chemistry of the biotites is consistent with calc-alkaline and alkaline host rocks (Harangi et al. 2000). K-feldspar and other Al-rich minerals (e.g. cordierite, garnet) characteristic of peraluminous S-type silicic rocks (i.e. the primary magmas were formed by partial melting of metasedimentary formations of the crust) are absent, although sanidine occurs in minor amounts (<5vol%) in some areas (e.g. Cserhát, Mecsek Mts; Oláh 2001).

The petrogenesis of the Miocene silicic magmas of the CPR is a subject of debate. Most researchers have considered that the primary magmas originated predominantly by partial melting of the continental crust. Póka et al. (1998) suggested that the silicic magmas of the LRT and URT were formed by anatexis of the upper crust, whereas the MRT was generated by mixing of andesitic melt and upper crust-derived silicic magma. Crustal anatexis as a dominant source of the Miocene silicic magmas of the CPR was also proposed by Downes (1996) and Lexa and Konečný, (1999). Downes (1996) invoked delamination of the lower part of the continental lithosphere as a primary cause of the lower crustal anatexis, whereas Lexa and Konečný (1999) suggested that partial melting of the crust was due to heating of the uprising asthenosphere and intrusion (underplating) of mantle-derived mafic magmas into the crust-mantle boundary zone. In contrast, Harangi et al. (2000) emphasized the significance of mantle-derived primary magmas in the genesis of the Miocene silicic volcanism. The morphological characteristics of zircons are consistent with hybrid calc-alkaline host magmas (Fig. 4) derived from the mantle and contaminated by variable crustal material (Szabó 2000; Harangi et al. 2000; Oláh and Harangi 2000; Oláh 2001). Among the different silicic volcanic units, magmas of the MRT have the largest mantle component. The zircon population of some samples from the LRT and the URT shows a bimodal character (Fig. 4) indicating mixing of calc-alkaline and alkaline magmas. An involvement of mantle-derived alkaline magmas in the genesis of the URT is suggested by the composition of plagioclases (mostly oligoclase) and biotites from the eastern Bükkalja region (Harangi et al. 2000). Sr and Nd isotope ratios of pumices and lithic clasts show a large range (initial $^{87}Sr/^{86}Sr$= 0.7067–0.7120; initial $^{143}Nd/^{144}Nd$=0.51220–0.51251), but fall in the narrow variation trend of the Miocene calc-alkaline volcanic rocks from the Northern Pannonian Basin (Fig. 5). This variation in the isotope ratios suggests again mantle-derived primary magmas and contamination by a variable amount of possibly metasedimentary lower crustal material. Magma mixing was a characteristic petrogenetic process throughout the Miocene silicic volcanism. During the Middle Miocene, intrusion of andesitic magma into the shallow-

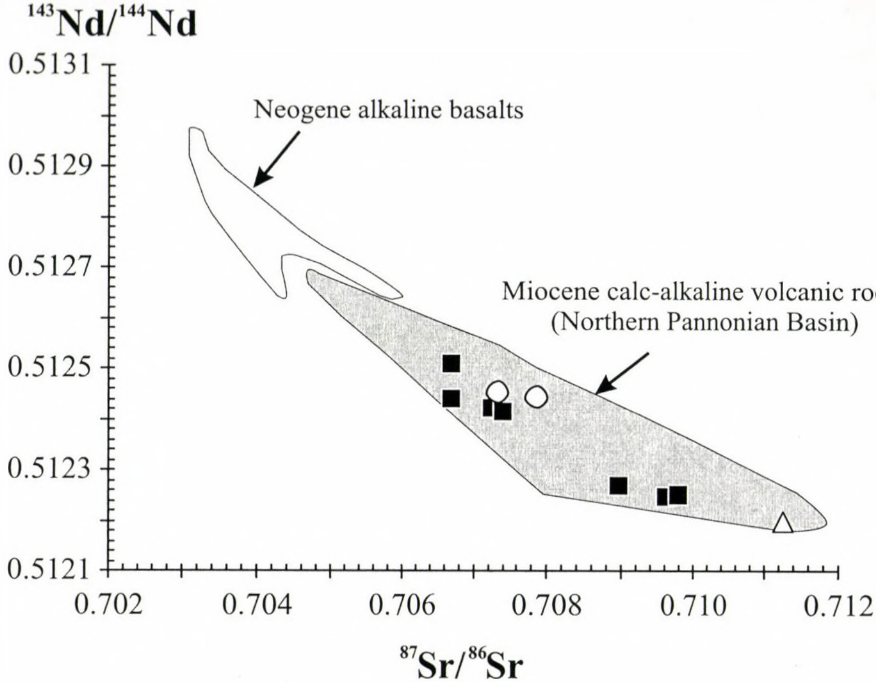


Fig. 5
$^{87}Sr/^{86}Sr$ vs. $^{143}Nd/^{144}Nd$ isotope plot for the Miocene silicic volcanic rocks of the CPR. Neogene alkaline basalts: Embey-Isztin et al. (1993), Harangi et al. (1995b); Miocene calc-alkaline volcanic rocks of the Northern Pannonian Basin: Salters et al. (1988), Harangi et al. (2001), Harangi, unpublished data. Symbols are as in Fig. 3

crustal rhyolitic magma chamber triggered the explosive eruption of the most evolved silicic magma from the upper part of the magma chamber followed by a hybrid dacitic-rhyodacitic one from the deeper parts. Incomplete syn-eruptive mixing resulted in the occurrence of both dacitic scoriae and rhyolitic pumices in the same pyroclastic deposit (Czuppon 2001).

Miocene to Pliocene potassic and ultrapotassic volcanism

The Neogene potassic and ultrapotassic rocks of the CPR mainly form lava flows with minor pyroclastic intercalations. The Middle Miocene rocks are silica-saturated shoshonites and trachyandesites (Krainer 1987; Pamić et al. 1992; Harangi, in press), whereas the Pliocene ultrapotassic rocks at Bár are silica-undersaturated and are classified as olivine leucitite (Harangi et al. 1995a). Based on their K_2O/Na_2O ratios (Fig. 6) the volcanic rocks from Balatonmária and Bár are ultrapotassic, whereas those that occur in the Styrian Basin and in the Drava Depression are potassic. The olivine leucitite from Bár contains olivine and

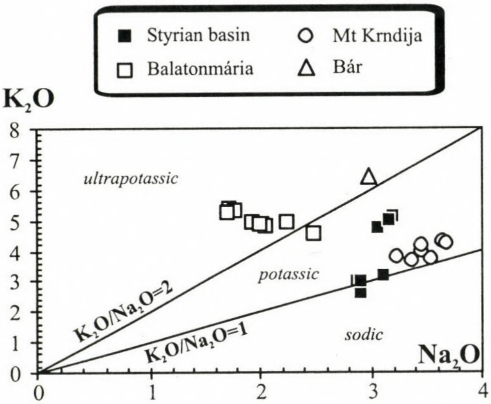


Fig. 6
Na_2O vs. K_2O plot for the K-rich volcanic rocks of the CPR

clinopyroxene phenocrysts with high mg-values (0.8–0.9) in the groundmass, consisting mainly of leucite with a lesser amount of olivine, clinopyroxene, Fe–Ti oxide and apatite. The silica-saturated potassic and ultrapotassic rocks contain phenocrysts of often oscillatory-zoned clinopyroxene, phlogopite, less olivine, ±sanidine, ±plagioclase. In the Balatonmária rocks apatite microphenocrysts are common. The groundmass contains sanidine, clinopyroxene and orthopyroxene, ±plagioclase and ±glass.

The leucitite from Bár has a high mg-value (0.7) and high FTSE concentration (e.g. Ni=150ppm, Cr=240 ppm) suggesting that its composition is close to the primary magma. In contrast, the other potassium-rich rocks (mg-value=0.54–0.62) underwent variable olivine+clinopyroxene fractionation. All of these rocks are enriched in highly incompatible elements (LILE) and show a negative Nb-anomaly in the normalized multi-element diagrams (Harangi et al. 1995a; Pamić et al. 1995). Trace element distribution of the potassic trachyandesites of Balatonmária (central Pannonian Basin) and Gleichenberg (Styrian Basin) is remarkably similar (Harangi et al. 1995a), whereas the trachyandesites from the Drava Depression are strongly enriched in barium (Pamić et al. 1995). The leucitite from Bár shows a distinct trace element distribution (Harangi et al. 1995a), which resembles that of lamproites. The isotopic composition of these rocks is variable. The initial $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios (0.7050 and 0.51259, respectively) of the Bár leucitite overlap the most enriched field of the ocean island basalts (OIB) and are comparable with the ultrapotassic rocks of Muriah (Sunda Arc – Edwards et al. 1991) and SE Australia (Nelson et al. 1986). The ultrapotassic trachyandesite from Balatonmária shows the highest initial $^{87}Sr/^{86}Sr$ (0.70943–0.70945) and the lowest $^{143}Nd/^{144}Nd$ ratios (0.51228–0.51230) among the potassium-rich rocks of the CPR.

Potassic and ultrapotassic rocks worldwide are considered to represent magmas derived from metasomatized continental lithospheric mantle (CLM). Phlogopite-bearing ultramafic xenoliths, which represent such a region occur sporadically in the Cretaceous to Tertiary alkaline igneous rocks of the CPR (lamprophyre and alkali basalts; Szabó 1985; Szabó et al. 1995). The geochemistry of the K-rich volcanic rocks of the CPR indicates no or only minor influence of crustal contamination (Harangi et al. 1995a; Pamić et al. 1995; Harangi, in press); therefore their radiogenic isotope ratios and trace element ratios (e.g. Zr/Nb, La/Nb), thought not to be affected by fractional crystallization, can be used to

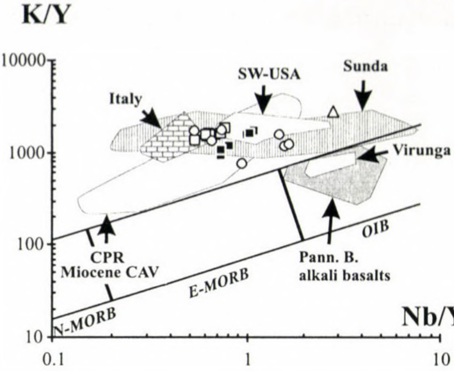


Fig. 7
K/Y vs. Nb/Y plot for the K-rich volcanic rocks of the CPR. Reference data: Italy: Conticelli and Peccerillo (1992), Rogers et al. (1985); Sunda: Edwards et al. (1991), van Bergen et al. (1992); SW USA: Gibson et al. (1992); Virunga: Rogers et al. (1992); CPR Miocene CAV: Downes et al. (1995a), Mason et al. (1996), Harangi et al. (2001), Harangi, unpublished data; Pannonian Basin alkali basalts: Embey-Isztin et al. (1993), Harangi et al. (1995b). Symbols are as in Fig. 6

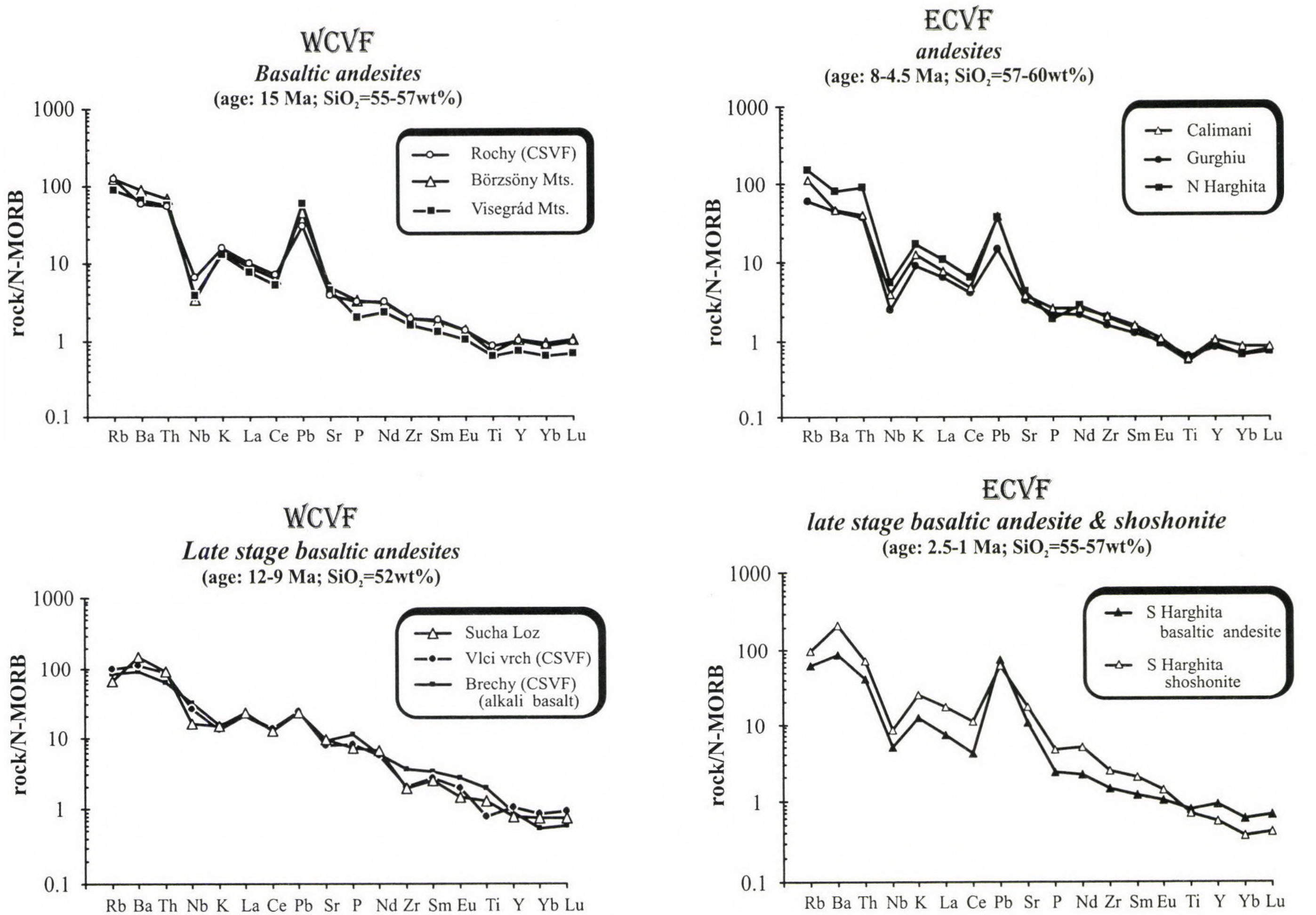
characterize the lithospheric mantle beneath the CPR. The LILE-enriched and Nb-depleted character of these rocks suggests that their source region in the CLM was previously metasomatized by hydrous fluids derived from subducted slab. HFS elements are not incorporated into hydrous fluids rising from the subducting slabs (Tatsumi et al. 1986); therefore HFSE/ HFSE ratios (e.g. Zr/Nb, Nb/Y) can be used to obtain information on the nature of the mantle before the subduction-related metasomatism occurred (e.g. Pearce 1982). On the K/Y vs. Nb/Y diagram (Fig. 7), uncontaminated mantle-derived volcanic rocks fall in a narrow band, whereas potassic rocks from various provinces are displaced vertically upwards. The K/Y ratio of the potassium-rich volcanic rocks of the CPR is consistent with the ultrapotassic rocks worldwide and indicates a K-rich metasomatized source region. The Bár leucitite has a high Nb/Y ratio, similar to the ultrapotassic rocks of Virunga (East Africa Rift), Muriah (Sunda Arc) and the Rio Grande Rift (southwestern USA). This suggests an involvement of OIB-like asthenospheric component in its genesis and/or a very small degree of partial melting leaving residual garnet in the source. Involvement of an enriched asthenospheric mantle is supported also by the Sr and Nd isotope ratios of the Bár leucitite. Thus, mixing between magma derived from the upwelling hot asthenosphere and melts from the lower CLM consisting of K-rich metasomatic layers is suggested for the genesis of the Bár leucitite (Harangi et al. 1995a). The lower Nb/Y ratio of the other K-rich rocks of the CPR implies a moderately enriched (E-MORB-type) mantle source and a higher degree of partial melting. The low Ce/Pb ratios and high $^{87}Sr/^{86}Sr$ ratios of these rocks can be explained by source contamination via melts from subducted sedimentary material (Harangi et al. 1995a). The origin of the Middle Miocene potassic and ultrapotassic rocks of the CPR was coeval with the main rifting period of the Pannonian Basin (Fig. 2). Thinning of the lithosphere reactivated the metasomatic K-rich veins at the base of the CLM and caused decompression partial melting that resulted in strongly potassic magmas.

Miocene to Quaternary calc-alkaline volcanism

Calc-alkaline volcanic rocks of the CPR are usually andesites with subordinate basaltic andesite, dacite and rhyolites. These volcanic rocks usually built up stratovolcanoes consisting of initial small-volume ignimbrites (e.g. Börzsöny; Karátson 1995; Karátson et al. 2000; Visegrád Mts; Gméling et al. 2000) and lava dome complexes followed by extrusive domes with extensive brecciation, block-and-ash flow, debris avalanche, lava flows and epiclastic volcanic deposits (e.g. Konečný et al. 1995a; Kaliciak and Žec 1995; Szakács and Seghedi 1995; Karátson et al. 2000). Subvolcanic intrusive complexes are characteristic of each calc-alkaline volcanic field. One of the striking features of the calc-alkaline volcanic rocks of the CPR is the relative abundance of garnet (almandine) phenocrysts (Embey-Isztin et al. 1985; Harangi et al. 2001). This mineral occurs rarely in volcanic rocks worldwide and its presence contains a strong geodynamic message. Ca-bearing and Mn-poor almandine garnet crystallizes from hydrous magma at high pressure (>7kbar, i.e. >20–25 km) and is not stable at lower pressure (e.g. Green 1992). Therefore, these magmas must ascend rapidly from the lower crustal depths that requirs a tensional tectonic regime. Indeed, garnet-bearing calc-alkaline volcanic rocks occur along major faults in the CPR. These particular volcanic rocks (andesites, dacites and rhyodacites) are the most frequent in the volcanic complexes at the Northern Pannonian Basin (Visegrád Mts, Börzsöny, Central Slovakian Volcanic Field; Harangi et al. 2001), but also occur sporadically in the volcanic complexes along the Eastern Carpathians (e.g. Mason et al. 1996). Most of the garnets are primary in these rocks, i.e. they crystallized directly from silicate melt, although xenocrystic almandine derived from metasedimentary lower crust can be also found (Harangi et al. 2001). The compositions of the primary garnets and the coexisting minerals are consistent with mantle-derived host magmas and crystallization at the mantle-crust boundary zone. It is remarkable that the eruption of garnet-bearing magmas occurred during the initial phase of volcanism at each volcanic field (Konečný et al. 1995a; Harangi et al. 2001). The garnet-free andesites are predominantly plagioclase-phyric with a variable amount of hornblende and/or clinopyroxene and orthopyroxene. Basaltic andesites are more frequent in the East Carpathian calc-alkaline volcanic fields, whereas they occur only during the late stage of the volcanism in the Northern Pannonian Basin volcanic fields (Konečný et al. 1995a; Mason et al. 1996). Late stage rhyolites are characteristic of the volcanic complexes at the Northern Pannonian Basin.

The difference found in the areal and temporal distribution as well as in the deep structure of the calc-alkaline volcanic complexes of the western and eastern segments of the volcanic belt along the Carpathians (hereafter called Western Carpathians Volcanic Field – WCVF, and Eastern Carpathians Volcanic Field – ECVF; Fig. 1) is also reflected by the geochemical composition of these rocks (Lexa and Konečný 1999; Harangi and Downes 2000). The WCVF calc-alkaline volcanic rocks ("areal-type" by Lexa et al. 1993) belong predominantly to the high-K series, whereas those from the ECVF ("arc-type" by Lexa et al. 1993) show greater compositional variation. They are predominantly medium-K rocks, but low-K (tholeiitic) series and shoshonites also occur (Mason et al. 1996). The trace element composition of these rocks is typical of subduction-related magmas, i.e. they are enriched in LIL elements (e.g. Rb, Ba, Pb) and depleted in the HFS elements (e.g. Nb, Ti, Zr; Fig. 8). Nevertheless, some differences can be observed between the early-stage andesites and basaltic andesites and the late stage volcanic rocks. In the WCVF, the late-stage basaltic andesites show only a very weak subduction signature and resemble the trace element distribution of the subsequent alkali basalts (Fig. 8; Harangi et al. 1998). In the ECVF, the late stage basaltic andesites and shoshonites have a distinct trace element distribution with typically higher Ba and Pb concentration, and depletion in Y and heavy rare earth elements (Fig. 8; Mason et al. 1996, 1998). The contrasting geochemical character of the WCVF and ECVF is also reflected by their radiogenic isotope composition. The $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotope ratios of the WCVF rocks define a fairly narrow curvilinear trend with usually higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ values (0.7048–0.7100 and 0.51225–0.51268, respectively; Salters et al. 1988; Harangi et al. 2001) than the ECVF rocks (0.7041–0.7101 and 0.51239–0.51290, respectively; Mason et al. 1996), which show a larger variation in isotope composition.

The genesis of calc-alkaline magmas is traditionally interpreted as the result of the introduction of volatile-rich fluids from the subducted slab and from the subducted sediments (e.g. Hawkesworth et al. 1994; Pearce and Peate 1995). Partial melting usually occurs in the metasomatized mantle wedge above the subducting oceanic lithospheric slab. The results of recent U–Th disequilibria isotope studies suggest that this complex scenario, i.e. dehydration of the slab, metasomatism of the mantle wedge, partial melting and eruption of the magmas, could take place in a time scale of a few tens of thousands of years (e.g. Elliott et al. 1997). Therefore, it can be assumed that calc-alkaline magmatism is associated with contemporaneous subduction. However, recent studies have pointed out that calc-alkaline magmas may also be erupted in extensional areas, where no active subduction occurs (e.g. Basin and Range, western USA; Hawkesworth et al. 1995 and Turkey, Wilson et al. 1997). In these areas melt generation took place as a response of lithospheric stretching due to the decompression melting of metasomatized lithospheric mantle. In the CPR, both petrogenetic models have been suggested. The Slovakian researchers proposed in many papers (e.g. Lexa and Konečný 1974, 1999; Lexa et al. 1993; Konečný et al. 1995a) that the calc-alkaline volcanic rocks in the Northern Pannonian Basin ("areal-type" as defined by them) were generated as a result of diapiric uprise of asthenospheric mantle due to the extension of the lithosphere. This asthenospheric material could have been metasomatized by subduction-related fluids. In contrast the calc-alkaline volcanic rocks at the eastern Carpathians ("arc-type" as defined by them) had a direct relationship with subduction or the subduction break-off process (Nemčok


WCVF
Basaltic andesites
(age: 15 Ma; SiO2=55-57wt%)
Rochy (CSVF)
Börzsöny Mts.
Visegrád Mts.
rock/N-MORB
1000
100
10
1
0.1
Rb Ba Th Nb K La Ce Pb Sr P Nd Zr Sm Eu Ti Y Yb Lu
ECVF
andesites
(age: 8-4.5 Ma; SiO2=57-60wt%)
Calimani
Gurghiu
N Harghita
rock/N-MORB
1000
100
10
1
0.1
Rb Ba Th Nb K La Ce Pb Sr P Nd Zr Sm Eu Ti Y Yb Lu
WCVF
Late stage basaltic andesites
(age: 12-9 Ma; SiO2=52wt%)
Sucha Loz
Vlci vrch (CSVF)
Brechy (CSVF)
(alkali basalt)
rock/N-MORB
1000
100
10
1
0.1
Rb Ba Th Nb K La Ce Pb Sr P Nd Zr Sm Eu Ti Y Yb Lu
ECVF
late stage basaltic andesite & shoshonite
(age: 2.5-1 Ma; SiO2=55-57wt%)
S Harghita
basaltic andesite
S Harghita
shoshonite
rock/N-MORB
1000
100
10
1
0.1
Rb Ba Th Nb K La Ce Pb Sr P Nd Zr Sm Eu Ti Y Yb Lu

et al. 1998). The latter model has been suggested by many other researchers since the 1970s (e.g. Bleahu et al. 1973; Boccaletti et al. 1973; Szabó et al. 1992; Mason et al. 1996; 1998; Seghedi et al. 1998). However, the genesis of the calc-alkaline volcanic rocks of the WCVF is still a subject of debate. Balla (1981), Szabó et al. (1992) and Downes et al. (1995a) among others suggested that these rocks were formed as a direct response to the subduction along the Carpathian Arc. Later, Harangi and Szabó (2000) and Harangi and Downes (2000) proposed a multi-stage model involving subduction-related metasomatism of the CLM and subsequent partial melting in the extensional phase of the Pannonian Basin. The contrasting behavior of the WCVF and ECVF calc-alkaline volcanic rocks (space and time distribution, deep structure, geochemistry of the volcanic rocks) requires different models for their origin. Harangi and Downes (2000) used incompatible trace element, radiogenic and stable isotope data to constrain the genesis of these rocks. The variation trend of the WCVF calc-alkaline volcanic rocks in the $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$ isotope diagram (Fig. 9) can be explained by two-component mixing between enriched asthenosphere-derived magmas and metasedimentary lower crust, whereas that of the ECVF calc-alkaline volcanic series can be modeled by a more complex petrogenetic scenario. In the ECVF most of the calc-alkaline magmas could have been generated from a depleted mantle source metasomatized by fluids derived from subducted flysch sediments. Subsequently, the uprising magmas underwent strong contamination by upper crustal material (Mason et al. 1996). The late stage shoshonites from the South Harghita show a distinct geochemical composition (Szakács et al. 1993; Mason et al. 1996; 1998) that can be explained by derivation from different mantle sources. Harangi and Downes (2000) suggested EMI-type lithospheric mantle source and subsequent contamination by upper crustal material.

Late Miocene to Quaternary alkaline volcanism

The Late Miocene to Quaternary alkaline volcanism resulted in a range of mafic rocks (nephelinite to trachybasalt, but dominantly basanite and alkali basalt) and subordinate basaltic trachyandesite and alkali trachyte (Embey-Isztin et al. 1993; Embey-Isztin and Dobosi 1995; Harangi et al. 1995b; Harangi, in press). The mafic rocks are porphyritic and contain predominantly magnesium-rich olivine and less clinopyroxene phenocrysts. The groundmass is composed of plagioclase, clinopyroxene, olivine, Fe-Ti oxides, apatite and occasionally glass, nepheline or leucite. Olivine and amphibole megacrysts and ultramafic and granulite xenoliths are common in the alkaline basalts (e.g. Embey-Isztin et al. 1989, 1990 2000; Kurat et al. 1991; Downes et al. 1992; Szabó and Taylor 1994;

← Fig. 8
N-MORB (Pearce and Parkinson 1993) normalized trace element patterns of the calc-alkaline volcanic rocks of the Western Carpathians Volcanic Field (WCVF) and Eastern Carpathians Volcanic Field (ECVF). Data are from Mason et al. (1996), Dobosi et al. (1995) and Harangi, unpublished data

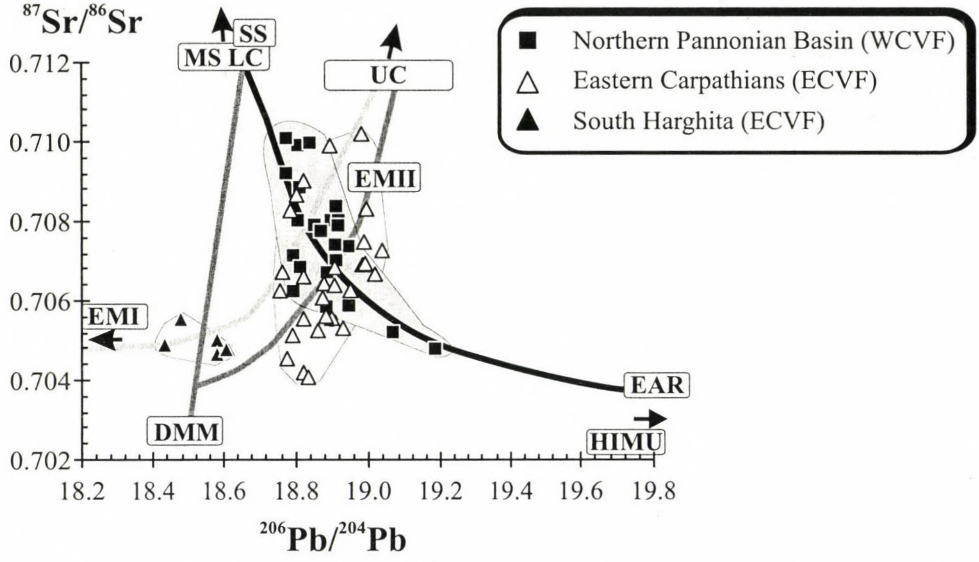


Fig. 9
$^{206}Pb/^{204}Pb$ vs. $^{87}Sr/^{86}Sr$ plot for the calc-alkaline volcanic rocks of the CPR and the petrogenetic interpretation of the variation trends. DMM, EMI, EMII and HIMU are mantle components (Zindler and Hart 1986); EAR is European Asthenospheric Reservoir (Cebriá and Wilson 1995); SS is subducted Cretaceous flysch sediments (Mason et al. 1996); MS LC is metasedimentary lower crust (Harangi et al. 2001); UC is upper crust (Mason et al. 1996)

Szabó et al. 1995; Vaselli et al. 1995, 1996; Kempton et al. 1997). These rock fragments provide direct information on the nature of the lithospheric mantle and the lower crust beneath the CPR (Downes and Vaselli 1995; Embey-Isztin et al. 2001). The trachyandesites and alkali trachytes in the Little Hungarian Plain are variably altered and contain antiperthitic sanidine and plagioclase phenocrysts in a groundmass composed of plagioclase, sanidine, quartz, clinopyroxene (aegirine-augite to aegirine), ±amphibole and biotite (Schléder 2001; Harangi, in press).

The mafic rocks show a wide compositional range from strongly silica-undersaturated to silica-saturated varieties (S.I.=–60 to +5, where S.I. is the silica-saturation index; Embey-Isztin and Dobosi 1995) and have variable K_2O/Na_2O ratio (0.2–0.9; Embey-Isztin et al. 1993; Harangi et al. 1995b). The mg-value varies between 0.56 and 0.76 but is mostly between 0.63 and 0.68. This suggests that most of the mafic magmas underwent only little olivine and ±clinopyroxene fractionation and that the compositions of the basaltic rocks are close to the primary magmas. Among the basaltic rocks of the CPR, the basanites from the Nógrád–Gömör Volcanic Field are the most differentiated (Dobosi et al. 1995). Trace element abundances of the alkaline basalts are typical of intra-plate mafic rocks, i.e. they show incompatible element enrichment relative to the primitive mantle. Comparing them to the average ocean island basalt (OIB) composition that is typical of enriched asthenosphere-derived magmas (Fitton et al. 1991), the mafic rocks from the CPR show an elevated LILE (e.g. Ba, K, Pb) concentration (Fig. 10; Embey-Isztin and Dobosi 1995; Harangi, in press). The relative LILE-enrichment is most characteristic in the alkaline basalts from the central Pannonian Basin and from the Persany Mts. In the basanite of the Persany Mts a small negative Nb anomaly can be also observed (Downes et al. 1995b). All of these features indicate the presence of a subduction-related component in the source region of the magmas. In contrast, this subduction-related signature is missing in the alkaline basalts of the Styrian Basin and from the Štiavnica-

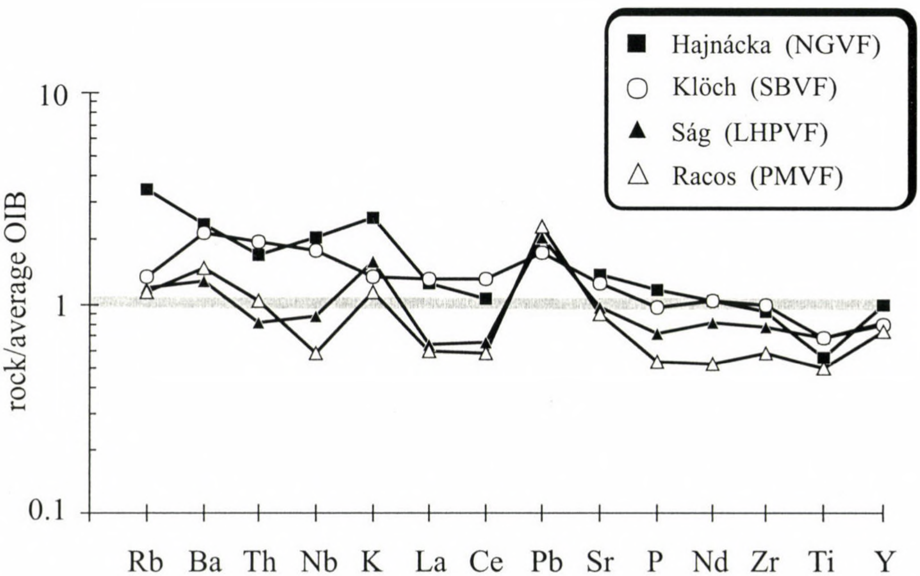


Fig. 10
Average OIB (Fitton et al. 1991) normalized trace element patterns for representative Late Miocene to Quaternary alkaline basalts of the CPR (data are from Embey-Isztin et al. 1993; Dobosi et al. 1995; Harangi et al. 1995b)

Nógrád–Gömör Volcanic Field (Fig. 10). The Sr and Nd isotope ratios in the latter volcanic rocks are also fairly homogeneous (except for the shoshonitic basalts from the Styrian Basin; Harangi, in press) and show the lowest $^{87}Sr/^{86}Sr$ and highest $^{143}Nd/^{144}Nd$ values. In contrast, the alkali basalts from the central part of the Pannonian Basin and from the Persany Mts. have more variable isotope ratios (Fig. 11). The relative LILE enrichment and high $^{87}Sr/^{86}Sr$ isotope ratios of these mafic rocks accompany with relatively high $^{207}Pb/^{204}Pb$ values at given $^{206}Pb/^{204}Pb$ ratios (Salters et al. 1988). It is remarkable that the alkali trachyte from

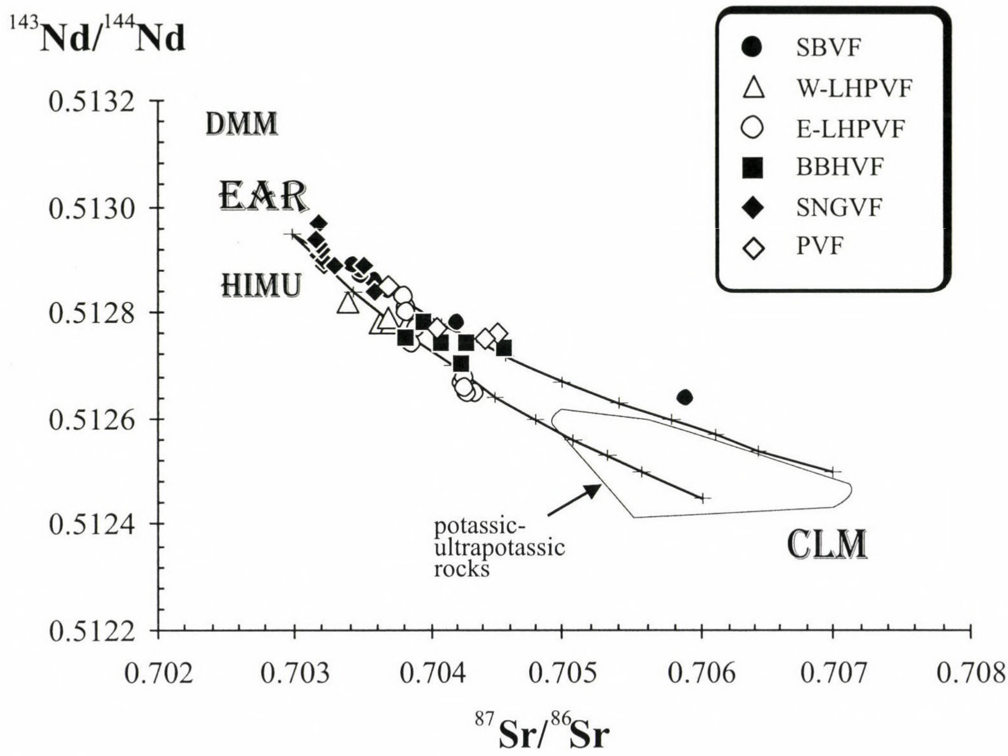


Fig. 11
$^{87}Sr/^{86}Sr$ vs. $^{143}Nd/^{144}Nd$ isotope plot for the Late Miocene to Quaternary alkaline basalts of the CPR and petrogenetic interpretation of the variation trend. Abbreviations are as in Fig. 9. CLM – continental lithospheric mantle

the Little Hungarian Plain has fairly low $^{87}Sr/^{86}Sr$ and high $^{143}Nd/^{144}Nd$ isotope values (Harangi et al. 1995b). This suggests that the trachyte is a differentiation product of asthenosphere-derived mafic magma. In contrast to the wide variation of the radiogenic isotope ratios of the basalts, the oxygen isotope ratios measured on olivine and clinopyroxene phenocrysts and megacrysts show a rather restricted range ($\delta^{18}O=5.0$–5.3‰; Dobosi et al. 1998). The uniform oxygen isotope ratio suggests that the mantle source of the alkali basalts was fairly homogeneous with respect to its oxygen isotope composition.

The relatively large geochemical variation in the alkaline basalts with fairly primitive nature suggests different source regions and different degrees of partial melting. Incompatible trace element ratios (e.g. La/Nb, Zr/Nb) and radiogenic and stable isotope ratios indicate an origin from an enriched asthenospheric source region. Trace element ratios which are sensitive to the mineralogy of the mantle source and therefore to the depth of the melting (e.g. Nb/Y) imply that partial melting occurred predominantly in the garnet stability zone, i.e. at about 80–100 km depth (Embey-Isztin and Dobosi 1995; Harangi et al. 1995b; Harangi, in press). The estimated degree of partial melting (0.5–2%; Harangi, in press) is consistent with the variation of the silica-saturation of the mafic volcanic rocks. On the other hand, the variation of radiogenic isotope ratios and some incompatible trace element ratios (e.g. La/Nb, La/Ba) can only be explained by mixing of asthenospheric magmas and lithospheric melts (Figs 11 and 12). The asthenospheric mantle could have been identical with the mantle source of Neogene alkaline basalts in Europe (Embey-Isztin and Dobosi 1995; Harangi, in press) that was termed as European Asthenospheric Reservoir (EAR; Cebriá and Wilson 1995) or as Low Velocity Component (LVC; Hoernle et al. 1995). The enriched lithospheric component could be partly an EMI-type lithosphere and

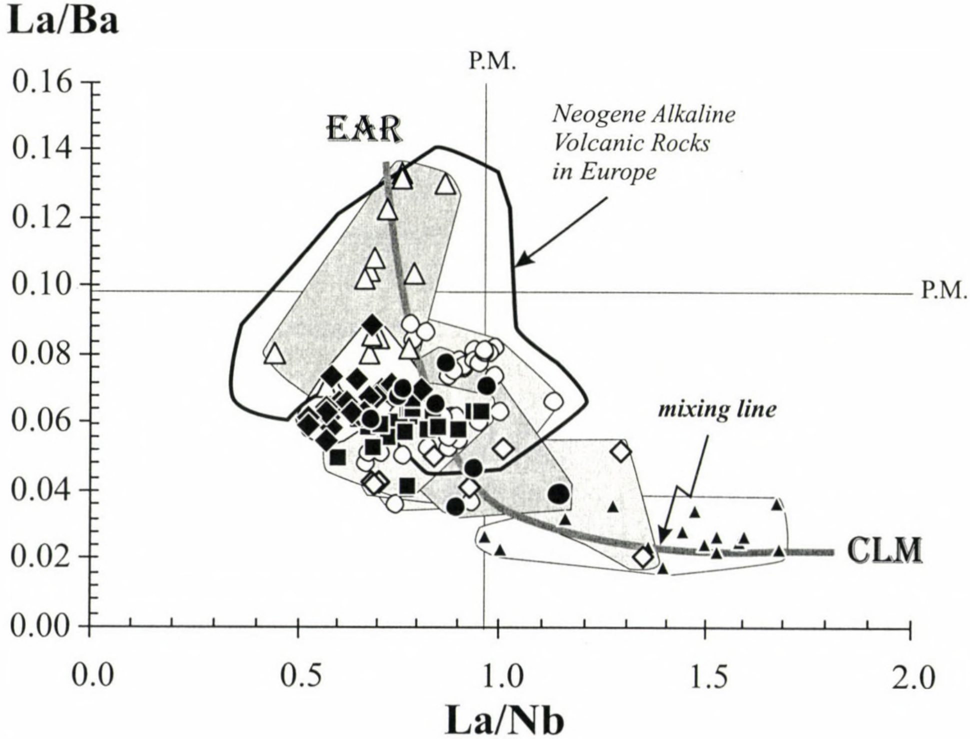


Fig. 12
La/Nb vs. La/Ba diagram for the Late Miocene to Quaternary alkaline basalts of the CPR and the petrogenetic interpretation of the variation trend. Symbols are as in Fig. 11. Filled triangles are potassic and ultrapotassic rocks of the CPR. Field of the Neogene alkaline volcanic rocks in Europe is after Harangi (in press). P.M. – primitive mantle

partly be similar to the source regions of the potassic and ultrapotassic magmas of the CPR (Harangi, in press).

Discussion

Melt generation in the mantle and in the lithosphere occurs in particular conditions, when the geotherm intersects the solidus characterizing the given region (Wilson 1993). This usually requires uprising of the asthenospheric mantle that, in turn, occurs in a particular geodynamic situation such as active or passive thinning of the lithosphere or pervasive metasomatism of the asthenosphere above the subducting slab. Therefore, there is a strong relationship between magmatism and the geodynamic evolution of a given area. During the last 20 Ma various volcanic events occurred in the CPR. What is the geodynamic implication of these volcanic activities? The geochemical composition of the volcanic products could tell us where and how the primary magmas were generated.

The CPR underwent a complex tectonic evolution from the Early Miocene (Fig. 2) that was discussed in several papers during the last decade (e.g. Csontos et al. 1992; Csontos 1995; Nemčok et al. 1998; Fodor et al. 1999; Tari et al. 1999). The main elements of this history were the southwestward and westward subduction of the European Plate and the reorganization of microplates (ALCAPA and Tisza-Dacide) behind the subduction zone. The Pannonian Basin was formed by major lithospheric extension during the Middle Miocene followed by thermal subsidence during the postrift phase. Subduction of the possibly oceanic lithosphere of the Magura Ocean began in the Paleogene and was terminated gradually from west (Middle Miocene) to east (Quaternary) when continental lithosphere entered the subduction zone. Recently a slab break-off mechanism has been proposed by many authors to explain the termination of subduction in the Carpathian–Pannonian Region (Nemčok et al. 1998; Seghedi et al. 1998; Wortel and Spakman 2000).

During the Early Miocene, the ALCAPA and Tisza–Dacide microplates are assumed to have been far away from one another (Fodor et al. 1999); therefore silicic volcanic rocks of the LRT occurring now in both microplates cannot belong to one cogenetic suite, as previously suggested by Szabó et al. (1992). Nevertheless it is remarkable that most of these volcanic rocks are situated along the Mid-Hungarian Tectonic Zone, which separates the two microplates. Here, it is tentatively suggested that the Early Miocene explosive volcanism of the silicic magmas (LRT) could have been the result of strike-slip tectonic movements along the southern margins of the ALCAPA Microplate and the northern margin of the Tisza Microplate during major block rotations. Mafic melts from the asthenosphere intruded into the weakening and possibly warm continental lithosphere and mixed with the induced partial melts of the lower crust. Lexa and Konečný (1999) suggested a close spatial and temporal relationship between the back-arc extension of the Pannonian Basin and the silicic magmatism. Their model involves generation of hydrous basaltic magmas from the uprising asthenosphere, the emplacement of these mafic magmas at the base of the thick crust and an associated anatectic melting of the crust. Downes (1996) proposed an alternative model invoking lithospheric delamination and crustal anatexis due to the heating of the uprising asthenosphere.

The Paleogene–Early Miocene times were characterized by advancing subduction, when the rate of overall plate convergence was greater than the rate of subduction (Royden 1993). The flat subduction and the compressional tectonic regime were able to prevent subduction-related calc-alkaline volcanism. However, dehydration of the subducted slab resulted in extensive metasomatism of the lithospheric mantle leading to the formation of hydrous minerals (amphibole and phlogopite; e.g. Downes and Vaselli 1995) and glass veins in the mantle peridotite (e.g. Szabó et al. 1996). The tectonic situation was changed to a retreating style of subduction when the rate of subduction exceeded the rate of convergence after the Early Miocene (Royden 1993). The principal deformation style became horizontal extension. The overall extension of the Pannonian Basin during the Middle Miocene (17–12 Ma) was associated with various volcanic activities. Thinning of the lithosphere could be accompanied with a suction effect in the upper mantle resulting in the upwelling of enriched asthenospheric material identical with the EAR ("thin spot effect"; Thompson and Gibson 1991). The consequence of decompression melting of the asthenosphere beneath the Pannonian Basin was different that might be related to the different structure of the lithosphere. In the central parts mantle-derived magmas underwent variable crustal contamination by the lower crust and subsequent fractionation in shallow crustal magma chambers (dacitic to rhyolitic magmas of the MRT and URT). The degree of contamination decreased with time due to the progressive thinning of the lithosphere. The rhyolitic magma chamber in the shallow crustal level was replenished repeatedly by mantle-derived basaltic to andesitic melts triggering explosive eruptions of rhyodacitic to rhyolitic magmas followed by dacitic ones. In the Northern Pannonian Basin, where the previous flat subduction resulted in extensive metasomatism at the base of the lithosphere, lithospheric thinning and the intrusion of EAR-derived magmas caused partial melting in the lithospheric mantle (Harangi et al. 1998, 1999). Lexa et al. (1993) and Lexa and Konečný (1999) considered that the andesitic magmas ('areal type') were formed in the metasomatized asthenosphere due to its extension-induced diapiric uprise. The basaltic to andesitic magmas were variably contaminated by the metasedimentary lower crust. The increase of alumina content of the hydrous melt led to the high-pressure crystallization of almandine garnet. The tensional stress field could enhance the rapid ascent of the garnet-bearing magmas along reactivated major faults. Subsequently the degree of crustal contamination decreased and calc-alkaline magmas with higher mantle component erupted in a Basin-and-Range-type region (Nemčok and Lexa 1990). From the Late Miocene the depth of melting increased and basaltic andesites with a strong EAR-signature were formed. In the southwestern part of the Pannonian Basin thinning of the lithosphere resulted in decompression melting of K-rich (phlogopite-bearing) zones at the base of the lithosphere, producing potassic and ultrapotassic magmas.

Calc-alkaline volcanic rocks along the eastern Carpathians ("arc type"; Lexa and Konečný 1999) show different space and time distribution and geochemical composition than those in the Northern Pannonian Basin (Harangi and Downes 2000). These differences, along with the different deep structure beneath the volcanic complexes, can be explained by a stronger relationship with subduction and slab break-off processes (Mason et al. 1998; Nemčok et al. 1998; Seghedi et al. 1998; Lexa and Konečný 1999). During the Late Miocene to Early Pliocene (11–6 Ma) calc-alkaline magmas were formed in the metasomatized, depleted mantle wedge above the subducting slab. These magmas subsequently underwent strong contamination by upper crustal material (Mason et al. 1996). As thick continental crust entered the trench, slab break-off began and progressed from north to south (Mason et al. 1998; Wortel and Spakman 2000). The gradual slab break-off could have occurred at relatively shallow depth beneath the southern Harghita. Upwelling of asthenospheric mantle through the slab window resulted in decompression melting of the asthenosphere (alkali basaltic magma; Persany Mts.) and at the base of the lithosphere (shoshonitic melts). The lithospheric mantle beneath the South Harghita appears to have been less metasomatized by subduction-related fluids, because it was closer to the suture zone. This is consistent with the unmetasomatized nature of the ultramafic xenoliths (Vaselli et al. 1995) and the assumed EMI-type mantle source of the shoshonites. In contrast, Mason et al. (1998) suggested that the unusual geochemical character of the shoshonites could be explained by the more efficient dehydration of the subducted lithosphere, due to the upwelling of hot asthenosphere into the gap behind the detached lithosphere and also by melting of the subducted eclogitic oceanic crust. Gîrbacea and Frisch (1998) and Chalot-Prat and Gîrbacea (2000) proposed an alternative model invoking horizontal intra-mantle delamination-induced volcanism in the South Harghita.

The Late Miocene to Quaternary (11–0.5 Ma) alkaline volcanism of the CPR postdates the synrift phase of the Pannonian Basin (Fig. 2) and therefore can be classified as post-extensional volcanic activity (Embey-Isztin and Dobosi 1995; Harangi 1996; in press). The geochemistry of the mafic rocks is consistent with derivation from the asthenosphere. Since lithospheric extension and related passive upwelling of the asthenosphere already ceased by the Late Miocene, melt generation during the Late Miocene to Quaternary could only have occurred assuming upwelling of a relatively hot (about 1400 °C of potential temperature) asthenospheric material. Upwelling of the EAR-type asthenospheric material could be induced by the thin-spot effect (Thompson and Gibson 1991) of the abnormally thin lithospheric structure beneath the Pannonian Basin.

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Geochemical characterization of the Pannonian Basin mantle lithosphere and asthenosphere: an overview

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Petrologic, geochemical and geophysical research revealed that the mantle lithosphere is thin and hot beneath the Pannonian Basin. Deformation, recrystallization and shear zones are characteristic of the uppermost part of the mantle lithosphere and they are intensively developed in the central part of the basin. Ductile shearing is linked to LREE-enrichment in the mantle. Major and trace-element abundance as well as Sr, Nd and Pb isotope studies enabled us to delineate the major geologic events occurring in the mantle lithosphere. The oldest known event was a partial melting and a major extraction of basalt that resulted in LREE-depleted refractory mantle. Later in its history the refractory mantle was re-enriched in highly incompatible elements and components with radiogenic Sr and Pb isotope compositions. The enrichment was a complex process that occurred in several steps. We can identify at least two types of enrichment and metasomatism in the mantle, (1) related to the passage of alkaline magmas similar to the host lavas, and (2) related to the subduction-derived fluids. The latter component is not seen in other regions of Neogene alkaline magmatism of Europe where subduction has not occurred since Hercynian times.

Key words: mantle lithosphere, astenosphere, depletion, enrichment, deformation, peridotite xenolith, alkali basalt

Introduction

Petrologists and geochemists have essentially two possibilities for studying the upper mantle. The first one is a direct approach, i.e. investigation of rock samples coming straight from the mantle. However, the continental crust constitutes a major obstacle for direct sampling. Petrologists and geochemists therefore rely heavily on mantle-derived rock fragments, so-called xenoliths, transported to the surface by magmas, to study the composition and evolution of the mantle.

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Mantle xenoliths are generally fresh and have not been re-equilibrated during their rapid ascent; they therefore provide representative samples of the deep-seated rocks. The rare outcrops of tectonically-emplaced ultramafic massifs also enable us to study the mantle directly.

The second possibility is an indirect one, which is the study of mantle-derived magmas, such as basalt, because these rocks mirror the composition of their mantle source region. However, only those compositional features that are unaffected by magmatic processes are useful as tracers of mantle processes. These include radiogenic isotope ratios such as those of Sr, Nd, Pb, and stable isotope ratios, e.g. O and H, and ratios of highly incompatible elements or elements of similar incompatibility, such as Ba/Nb or Pb/Ce. The term "incompatible" denotes a preference of the element for a melt over mantle minerals. Highly incompatible elements will partition entirely into the melt under most circumstances, so that the ratio of two such elements in a basalt will be virtually identical to that ratio in its source. This is also true to a lesser degree of ratios such as La/Sm and Zr/Nb, as Zr and Sm are not highly incompatible elements. Of these mantle tracers, isotope ratios have proven the most useful because they are insensitive to magmatic processes and because they are functions of time.

In the Pannonian Basin the petrologic and geochemical study of upper mantle rocks started in the 70s following the first finds of upper mantle peridotite xenoliths in the young (2.3–4 my.) basaltic tuffs and lavas at Szentbékkálla, Szigliget, Bondoróhegy and Gérce in Hungary (Fig. 1) (Embey-Isztin 1976a, b, 1978, 1984). This study was rapidly followed by modern petrologic investigations of the alkali basaltic magmas (Embey-Isztin 1980, 1981; Embey-Isztin and Scharbert 1981; Embey-Isztin et al. 1985; Pantó 1981; Poultidis and Scharbert 1986; Salters et al. 1988). In the last 10–12 years, however, considerable progress was made in the study of petrology and geochemistry of the upper mantle below the Pannonian Basin. Both the direct method (i.e. the study of mantle xenoliths, e.g. Embey-Isztin et al. 1989a; Kurat et al. 1991; Downes et al. 1992; Downes and Vaselli 1995; Szabó and Taylor 1994; Szabó et al. 1995a, b; Rosenbaum et al. 1997; Demény and Embey-Isztin 1998; Dobosi et al. 1998a) and the indirect method (the investigation of mantle-derived alkali basalt, e.g. Dobosi 1989; Dobosi et al. 1991; Dobosi and Fodor 1992; Embey-Isztin et al. 1993a, Embey-Isztin and Dobosi 1995; Harangi et al. 1995; Embey-Isztin and Kurat 1996; Embey-Isztin and Dobosi 1998; Dobosi et al. 1998b; Embey-Isztin 1999) have been widely applied. This review has been compiled on the initiative of the Hungarian National Committee of the International Lithosphere Project with the aim of summarizing our present state of knowledge about the Pannonian mantle, which has been greatly improved by the scientific activity promoted by the EUROPROBE PANCARDI Project. As far as mantle xenoliths are concerned, the review concentrates on the Pannonian Basin sensu stricto and especially on the Transdanubian area (Fig. 1) where large and fresh mantle xenoliths are abundant. Very small xenoliths are known from the Nógrád region and from the late Cretaceous–Paleogene alkaline mafic rocks

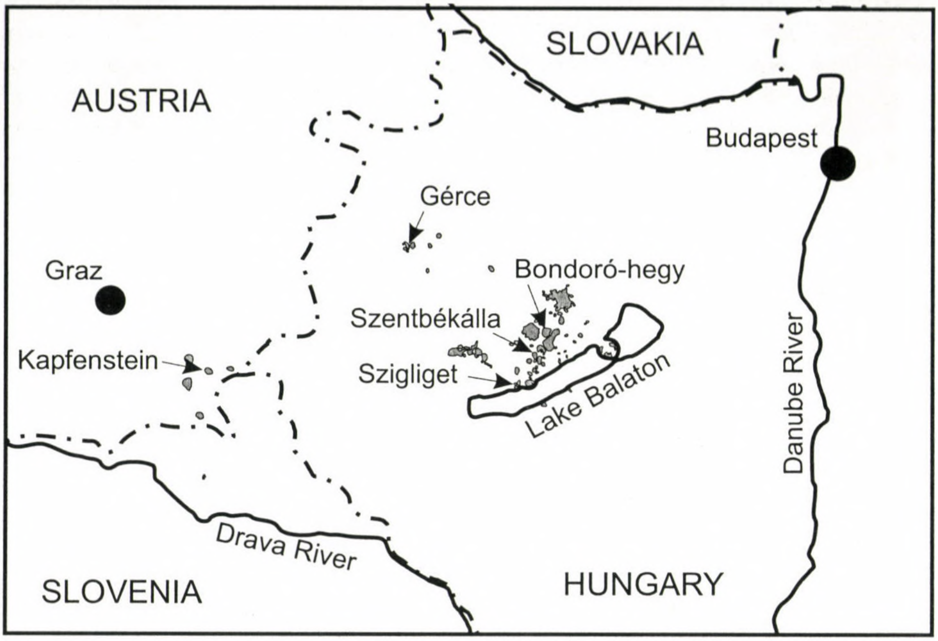


Fig. 1
Geographic sketch-map showing xenolith localities in the western Pannonian Basin

recovered from the Alcsutdoboz borehole (Szabó 1985). The latter xenoliths are, however, severely altered. The main results will be frequently compared to those obtained in the adjacent regions such as the Transylvanian and Graz Basins.

Lithology of the mantle lithosphere beneath the Pannonian Basin

Peridotite xenoliths are fragments from the mantle lithosphere and as such they provide first-order information on the mineralogy, petrology, tectonic history and geochemical evolution of the mantle. The number of xenoliths which have been studied under the microscope exceeds 800 and this sample number is high enough to be representative for the Pannonian upper mantle, at least beneath the Balaton and Kisalföld area. The majority of the peridotite xenoliths are Type I four-phase spinel lherzolites (diopside>5%), indicating that this is the main lithology which forms the shallow lithospheric mantle beneath the area. A smaller number of harzburgite xenoliths (diopside<5%) was also found. No peridotite xenolith contains garnet and/or plagioclase and this strongly constrains the depth interval from which the rock fragments originate, namely to the stability field of spinel peridotites.

Mantle xenoliths other than spinel lherzolites and harzburgites are rare in the region studied. Among these, the most "common" group is the anhydrous Type II pyroxenites ranging from pure spinel pyroxenites to olivine pyroxenites and wehrlites (Embey-Isztin et al. 1990). These lithologies may have formed narrow dykes and veins within the spinel peridotite host rock as attested by the rare composite xenoliths. Thin, cm-wide Type II amphibolite veins in spinel peridotite are extremely rare. Such composite xenoliths were found and studied in detail from the basaltic tuff at Szigliget (Embey-Isztin 1976a) and Kapfenstein (Styria, Austria) (Kurat et al. 1980). In contrast, amphibolite veins are more frequent in the peridotite xenoliths of the Transylvanian Basin, indicating that here the depleted mantle was enriched by vein metasomatism (Vaselli et al. 1995; Chalot-Prat and Boullier 1997). Amphibole selvage has been found in Nógrád-Gömör (Szabó and Taylor 1994).

Mineralogy of peridotite xenoliths

The mineralogy of mantle-derived peridotite xenoliths is dominated by olivine (50–75 per cent by volume), followed by orthopyroxene (15–25%), clinopyroxene (2–15%) and spinel (1–4%). All the silicate phases are rich in Mg, e.g. the composition of olivine is Fo (forsterite) 89–92 and the coexisting orthopyroxene is an enstatite. In addition, the clinopyroxene is rich in Cr_2O_3 (ca. 1%). The composition of the minerals and their modal abundances are interdependent. In olivine-rich and clinopyroxene-poor rocks the silicate phases have high mg-values (Mg/(Mg+Fe)) and low abundances of Al, Ca and Ti. Conversely, olivine-poor and clinopyroxene-rich xenoliths have silicates with lower mg-values and higher concentrations of Al, Ca and Ti. Spinels in olivine-poor and clino-pyroxene-rich xenoliths are rich in Mg and Al, but poor in Fe and Cr, whereas in olivine-rich and clinopyroxene-poor rocks spinels are rich in Fe and Cr and poor in Mg and Al. Though very rare accessory minerals have occasionally been identified, most xenoliths are composed entirely of the four anhydrous phases mentioned above. However, about 7–8% of the peridotite xenoliths is hydrous, containing discrete amphibole grains as a fifth phase and occasionally the quantity of amphibole may be >5% (e.g. in the peridotite xenoliths of the Nógrád region; Szabó and Taylor 1994). In the pyroxenite xenoliths silicate phases have distinctly lower mg-values and the associated spinel is highly aluminous.

Textures

The study of texture is important because it has implications regarding the tectonic evolution (e.g. recrystallization, shear processes) of the mantle lithosphere. Texture types of peridotite xenoliths and their relative abundances at different volcanic vents have been the subject of detailed studies (Embey-Isztin 1984; Kurat et al. 1991). Most volcanic vents have yielded both undeformed (protogranular) and deformed (equigranular and/or porphyroclastic) peridotites; however, the proportion of these types is variable among the various localities.

The volcanic vent of Kapfenstein (Styria) is an exception, where only untectonized protogranular xenoliths have been encountered. Peridotite xenoliths of the Nógrád region show textural bimodality (Szabó and Taylor 1994). Transitional textures between protogranular and porphyroclastic textures, pointing to an incipient stage of deformation, are, however, not rare at all localities. The overall picture indicates that the mantle lithosphere is essentially undeformed in the marginal zone of the basin (Kapfenstein), and at least partially deformed in the central region (Kisalföld and Balaton Highland – Kurat et al. 1991). This may be due to rise of a mantle diapir beneath the central part of the basin and/or more intense stretching of the lithosphere. Rare texturally composite xenoliths (i.e. containing an undeformed and a deformed portion) indicate the presence of shear zones (Downes et al. 1992). By comparison we note that the mantle lithosphere below the Transylvanian Basin is again largely undeformed (Vaselli et al. 1995).

In addition to the three main texture types that constitute a series of increasing tectonization in the order protogranular–porphyroclastic–equigranular, a fourth texture type is also frequent in peridotite xenoliths of the central region. This is called poikilitic by virtue of having small, rounded, sometimes subidiomporphic spinel grains poikilitically enclosed in large olivine, orthopyroxene and clinopyroxene crystals (Embey-Isztin 1984; Embey-Isztin et al. 1989a). This texture type is rare on a global scale and it is virtually absent in the xenoliths of the Graz Basin and the Transylvanian Basin. It seems that these peculiar xenoliths also constitute a deformational series from a coarse-grained, undeformed to a fine-grained recrystallized variety (Embey-Isztin et al. 2001).

Thermal state and evolution of the mantle lithosphere

The thermal state of the mantle can be investigated by geophysical means, such as heat-flow or electric conductivity measurements, and by evaluation of frozen-in equilibrium temperature and pressure values from the phase chemistry of mantle-derived peridotites. The latter approach was already applied by Embey-Isztin et al. (1989a), Downes et al. (1992) and Szabó et al. (1995a); however, recently a larger series of peridotite xenoliths (including the localities of Gérce, Szentbékkálla, Bondoróhegy, Szigliget and Kapfenstein) have been studied than ever before, using a high precision Cameca SX50 electron microprobe equipped with five wave-length dispersive spectrometers (University of Heidelberg). For the first time, detailed concentration profiles were determined for each of the mineral phases in the majority of samples (Embey-Isztin et al. 2001). It has been concluded that most of the intra-grain compositional variations reflect thermal processes characterizing ambient conditions in the upper mantle. By far the largest proportion of the xenoliths show more or less homogeneous mineral compositions and may thus be regarded as equilibrated under the prevailing pressure-temperature conditions. Thus this group of xenoliths is suitable for the

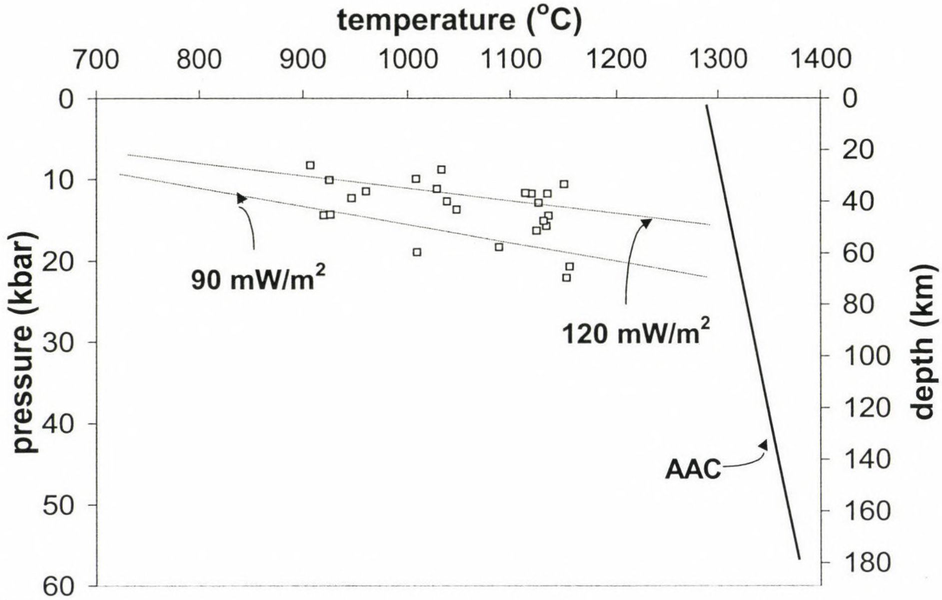


Fig. 2
P–T diagram of selected, well-equilibrated peridotite xenoliths. Data source: Embey-Isztin et al. (2001). Temperature and pressure values were estimated using the two-pyroxene thermometer (Brey and Köhler 1990) and the Ca-in-olivine barometer of Köhler and Brey (1990). Note that the xenoliths are concentrated along and between the geotherms calculated from the 90 and 120 mW/m^2 surface heat flow values by Pollack and Chapman (1977). AAC: adiabatic upwelling curve for normal temperature asthenosphere in the absence of significant amounts of melting with a potential surface temperature of 1280 °C (McKenzie and Bickle, 1988). The intersection of the estimated geotherm with the AAC curve gives an apparent thickness of 60–70 km for the lithosphere

evaluation of equilibrium temperatures and pressures based on the distribution of certain elements between the coexisting mineral phases. The geothermobarometric calculations (based on the two-pyroxene thermometer of Brey and Köhler 1990 and the Ca-in-olivine barometer of Köhler and Brey 1990) on these xenoliths yielded equilibration temperatures ranging from about 850 to 1175 °C and 8–22 kbar, pointing to a provenance depth ranging from the Moho at 25–29 km to 70 km (see explanation in Fig. 2). The latter value should approximate the thickness of the lithosphere beneath the Pannonian Basin. The results of thermobarometry substantiate a hot lithosphere, as high-temperature spinel-bearing lherzolites (>1000 °C) indicate high geothermal gradient and samples equilibrated near 1200 °C should approach the asthenosphere. This is in agreement with the known geophysical properties of the area (e.g. Posgay et al. 1986; Ádám et al. 1989; Praus et al. 1990).

A much smaller group of the studied xenoliths, however, exhibits significant intra-grain compositional variations reflecting partial re-equilibration due to changing thermal conditions characterizing the mantle environment prior to the entrainment in melts. Xenoliths of this group are less suitable for determining equilibrium temperatures and pressures. In contrast they provide information on the thermal evolution of the lithosphere. Among the xenoliths studied by Embey-Isztin et al. (2001), compositional heterogeneity indicating chemical disequilibrium is only important in pyroxenes of some porphyroclastic xenoliths from the Gérce locality (Kisalföld area). The zoning patterns with regularly decreasing Al abundances of both coexisting orthopyroxenes and clinopyroxenes (Fig. 3)

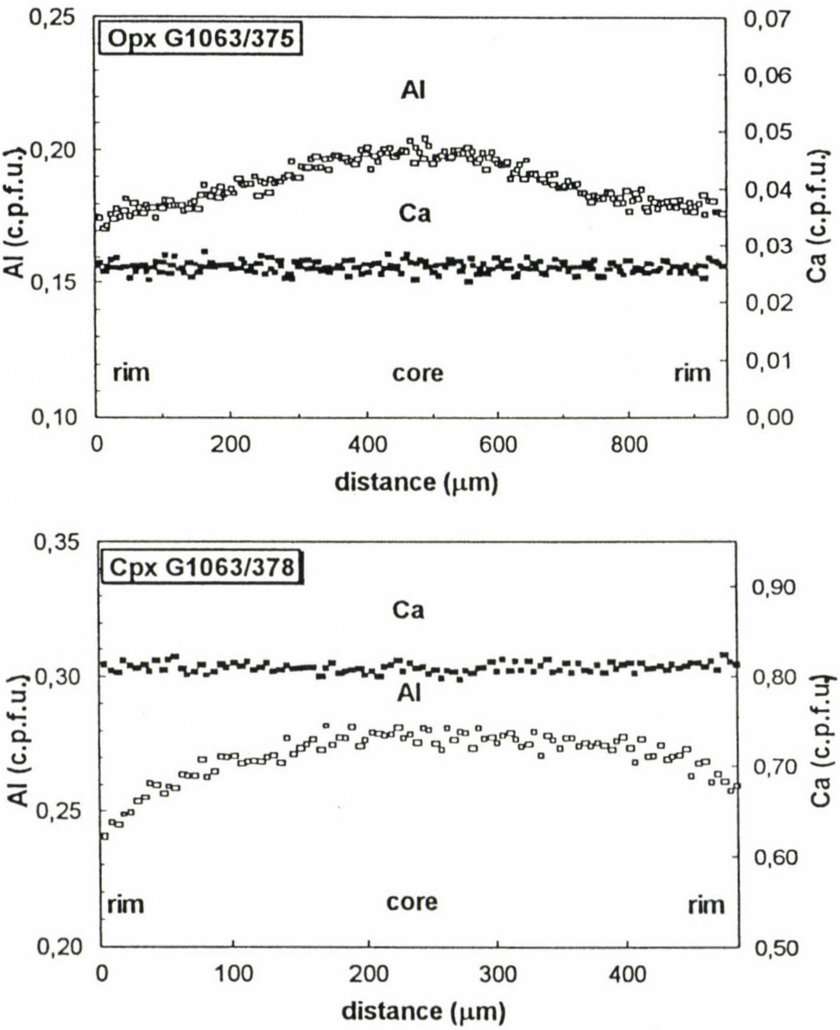


Fig. 3
Complex Al and Ca zoning patterns of orthopyroxene (opx) and clinopyroxene (cpx) of a non-equilibrated mantle xenolith from Gérce (Kisalföld) (after Embey-Isztin et al. 2001). The regularly decreasing Al from the core towards the rims indicates a cooling event in the lithosphere

attest to simple cooling of the lithosphere before entrainment of xenoliths by the host magma.

An important observation is that the equilibrium temperatures and the texture type of the xenoliths are interdependent. Undeformed coarse-grained protogranular and poikilitic xenoliths tend to show high temperatures, whereas deformed, fine-grained equigranular and poikilitic xenoliths yield low temperatures of equilibration (Downes et al. 1992; Embey-Isztin et al. 2001; Fig. 4). We note that coarse-grained poikilitic peridotite xenoliths from Borée, Massif Central (France) also gave high temperatures of equilibration (Xu et al. 1998). The fact that volcanic vents yielded both low-T deformed and high-T undeformed xenoliths suggests that the xenoliths may originate in a diapir, probably due to uplift of hot mantle material into a cooler uppermost mantle. Rare texturally composite peridotite xenoliths (i.e. containing coarse-grained and fine-grained parts in planar contact) corroborate this idea (Downes et al. 1992). It seems that shear processes causing deformation in peridotites were associated with a temperature decrease from about 1150 to ca. 800 °C. Al and Ca zoning patterns such as preserved in large pyroxene grains of some of the porphyroclastic

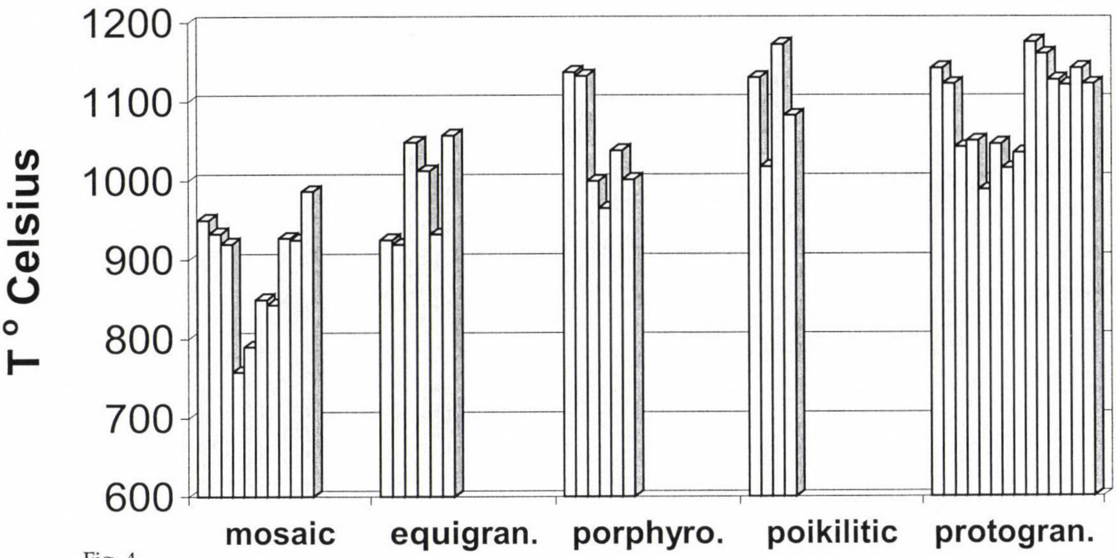


Fig. 4
Frequency distribution of equilibrium temperatures of xenoliths sorted according to their texture (after Embey-Isztin et al. 2001). Mosaic: equigranular mosaic, Equigran.: equigranular tabular, Porphyro.: porphyroclastic, Protogran.: protogranular. Note that xenoliths with undeformed texture tend to have higher equilibrium T than deformed xenoliths

xenoliths from Gérce may be the frozen-in witnesses of this cooling process (Embey-Isztin et al. 2001).

Geochemistry of the mantle lithosphere

Major elements

The bulk rock chemistry of the Pannonian mantle xenoliths forms tight and well correlated arrays ranging from 36 to 46 wt% MgO, 0.5 to 4.0 wt% CaO and 1.0 to 4.0 wt Al_2O_3 (Fig. 5). Al_2O_3, CaO and TiO_2 all correlate negatively with MgO, and the observed trends (that are also mirrored in the mineralogical variations discussed above) reflect a mantle which has been depleted to various extents in "basaltophile" elements. In fact the compositional spread encompasses the whole range from fertile lherzolite to refractory harzburgite compositions. Thus the implication is that variable amounts of basalt must have been extracted from the fertile mantle, presumably through a partial melting event. This basaltic component was probably lost at some time in the geologic past rather than recently, judging from the Nd model ages of depleted peridotite xenoliths (see below).

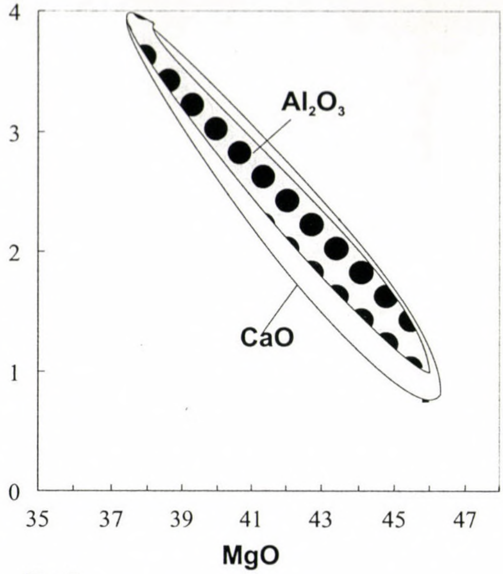


Fig. 5
Al_2O_3, CaO vs. MgO diagram for western Pannonian Basin peridotite xenoliths. Data source: Embey-Isztin et al. (1989a); Downes et al. (1992)

Incompatible trace elements

The lithosphere is a non-convective layer in which chemical heterogeneities may be preserved for billions of years. Therefore, the trace element and isotope signatures of mantle xenoliths can be used to decipher its long-term chemical evolution. The rare earth elements (REE) are incompatible with anhydrous spinel peridotite, and removal of a basaltic component from the mantle, as indicated by the major element chemistry, must strongly affect their relative abundances. As the light REE (LREE) are more incompatible with mantle assemblages than the heavy REE (HREE), they become strongly depleted in the refractory mantle. The resulting REE pattern is therefore fractionated. Thus the LREE-depleted patterns of separated clinopyroxenes in Fig. 6 coupled with the major element abundances of their respective whole rock samples are readily explained as the result of extraction of basalt from the mantle (the great majority of the REE of a peridotite reside in the clinopyroxene, so that the REE pattern of separated

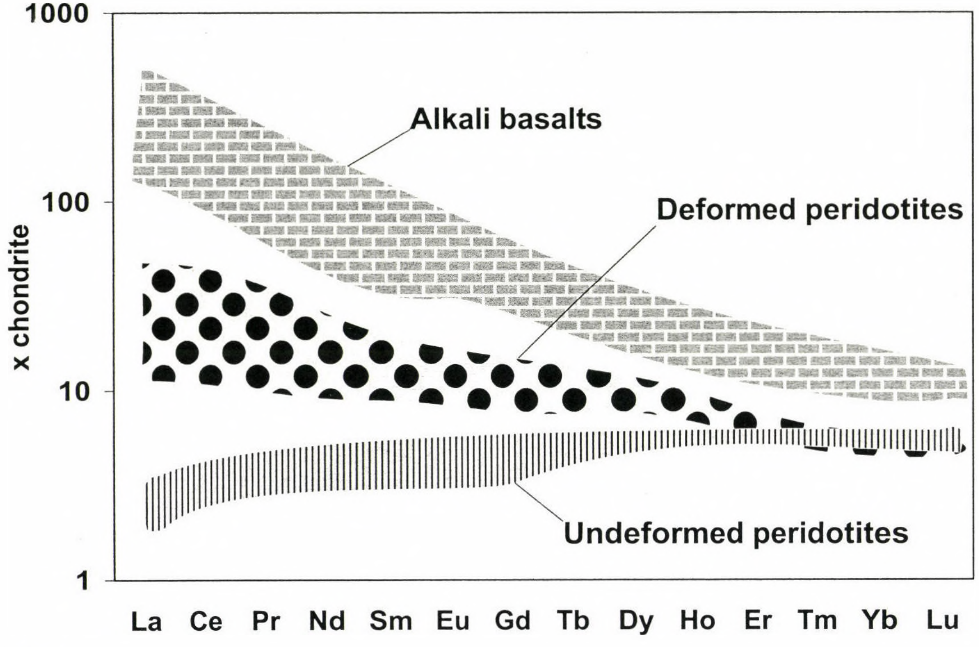


Fig. 6
Ranges of chondrite-normalized REE patterns for deformed and undeformed mantle peridotites of the western Pannonian Basin and alkali basalt of the Carpathian Basin. Data source: Downes et al. (1992); Embey-Isztin et al. (1993a)

clinopyroxene grains should be similar to that of the whole rock sample, but at higher relative values). However, in the Pannonian Basin, just as in the Graz Basin (Kurat et al. 1980; Vaselli et al. 1996), Transylvanian Basin (Vaselli et al. 1995) and the French Massif Central (Zangana et al. 1997), both LREE-enriched and LREE-depleted xenoliths can occur in the same volcanic vent (Fig. 6). One would expect that the LREE-enriched xenoliths would have major element abundances characteristic of fertile mantle. However, this is not the case; on the contrary, such xenoliths are often more depleted in terms of major elements than the LREE-depleted xenoliths. This clearly shows that in these rocks the major and incompatible trace elements are decoupled. As the extraction of basalt must have depleted the LREE, the observed elevated LREE abundances should be the consequence of a later re-enrichment as a result of the passage of LREE-enriched fluids or magmas.

Most of the LREE-enriched xenoliths do not show evidence of later growth of a LREE-rich accessory phase such as phlogopite, amphibole or apatite. Such xenoliths have therefore been referred to as being "cryptically" metasomatized in the sense of Dawson (1984). However, the presence of amphibole-bearing peridotite xenoliths strongly suggests that modal metasomatism also took place in the mantle. Cryptic and modal metasomatism are most probably separate processes both in space and time, though it has been shown that the growth of amphibole can increase the total LREE budget of the host rock, at least locally. For example, amphiboles in veins and in amphibole-rich peridotites have high REE and are usually enriched in LREE in Transylvania and Germany (Vaselli et al. 1995; Witt-Eickschen et al. 1998). The effect of an amphibole-rich vein on the REE pattern of the adjacent mantle is very clear, in that an increase in LREE enrichment can be observed in clinopyroxenes near the vein, compared with those farther away (Vaselli et al. 1995). However, examples from the Pannonian Basin indicate that the effect of modal metasomatism on the LREE-budget is not always evident. Though the Szigliget and Kapfenstein hornblendite veins were considered as LREE-enriched alkali basaltic mobilisates (Embey-Isztin 1976a; Kurat et al. 1980) and most interstitial amphiboles are enriched in LREE, at least one xenolith containing interstitial amphibole from Szigliget and three others from Kapfenstein and two more from Gérce show LREE-depleted patterns (Downes et al. 1992; Kurat et al. 1980; Vaselli et al. 1996; Szabó et al. 1995b). Kurat et al. (1980) interpreted these xenoliths as the products of pure water metasomatism, by which amphibole, characteristically poor in K, has been formed by reaction of spinel with clinopyroxene and water. Also the differences in bulk compositions between vein amphiboles and interstitial amphiboles point to distinct genetic events. Vaselli et al. (1995) however, preferred an alternative explanation according to which the differences would be due to equilibration of interstitial amphiboles with the host mantle.

The trace elements Sr, Sc, V, Y and Zr follow the major element trends, i.e. negative correlation between MgO and these elements are observed. Y behaves geochemically like the HREE; its behavior is largely governed by partial melting, rather than by any subsequent enrichment. From the transition metals only Ni forms positive correlation with MgO, indicating its compatible nature in olivine.

Radiogenic isotope geochemistry

$^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic compositions of separated clinopyroxenes of Pannonian Basin peridotite xenoliths show a much wider variation than the host alkali basalt ranging from ratios similar to those of depleted mantle (DM) to values more enriched than Bulk Earth (Fig. 7). Knowledge of these isotope ratios allow us to estimate the Nd model ages and thus bracket the timing of incompatible element depletion and enrichment events occurring in the mantle lithosphere. The depleted mantle signature is the time-integrated effect of extraction of basalt leaving a LREE-depleted residue, which has evolved to high $^{143}Nd/^{144}Nd$ and low $^{87}Sr/^{86}Sr$. As the half-life of the isotopic systems involved are very high, xenoliths plotting close to DM may have lost their basaltic component in the geologic past, probably more than 1 Ga ago. In contrast, values around Bulk Earth do not necessarily signal a time-integrated response of an old incompatible element enrichment process; they are more probably the result of a

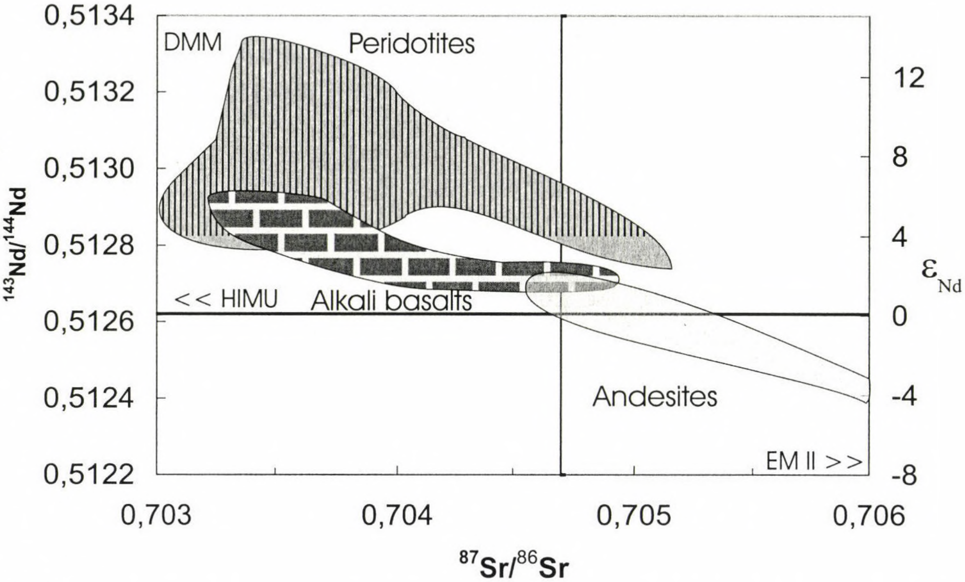


Fig. 7
Sr and Nd isotope variation diagram for peridotite xenoliths, alkali basalts, and the least radiogenic end of the calc-alkaline lavas. Data sources: Salters et al. (1988); Downes et al. (1992); Embey-Isztin et al. (1993a); Downes et al. (1995a). Dividing straight lines show the isotope ratios of the undifferentiated Bulk Earth. DMM, HIMU and EM mantle components are taken from Zindler and Hart (1986)

relatively young event, such as the passage of fluids and melts through the mantle. In fact, xenoliths may have been overprinted by the Neogene alkali basaltic magmatism, which lies at the high $^{87}Sr/^{86}Sr$ and low $^{143}Nd/^{144}Nd$ end of the xenolithic trend (Fig. 7). However, xenoliths approaching the least radiogenic end of the Miocene calc-alkaline magma composition field in Fig. 7 may signal a mixing array towards another enriched component characterized by high $^{87}Sr/^{86}Sr$ ratio but little decrease in the $^{143}Nd/^{144}Nd$ ratio. This was interpreted by Downes et al. (1992) as being the influence of subduction-derived fluids, as the Balaton and Kisalföld area are situated above a Neogene subduction zone. Thus the compositional field of peridotites in Fig. 7 is probably the result of mixing between an ancient depleted mantle with high $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios and two enriched end-members with lower $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios.

The suggestion that the originally depleted mantle lithosphere may have been influenced by both subduction-derived fluids and an asthenospheric component is further corroborated by Pb isotope ratios obtained on carefully leached clinopyroxenes (Rosenbaum et al. 1997). In Fig. 8a, b the mantle xenoliths indicate a mixing between depleted mantle with unradiogenic Pb compositions and a plume-type mantle (HIMU) with more radiogenic Pb. The latter component may relate to the upwelling of the asthenosphere during Tertiary times (Hoernle et al. 1995). However, some xenoliths have high ratios of $^{207}Pb/^{204}Pb$ at a given $^{206}Pb/^{204}Pb$, plotting away from the Northern Hemisphere Reference Line in Fig. 8a, which is a strong evidence of admixing of a component of subducted sediment derived from ancient continental crust. This component is present only in the mantle xenoliths of the western Pannonian Basin and is absent from the xenoliths of the Transylvanian Basin (Rosenbaum et al. 1997). In the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ plot (Fig. 8b) the mantle xenoliths form a subparallel array closer to the NHRL, partially overlapping the compositional field of the alkali basalt. In this diagram the mixing trend between the depleted mantle (DMM) and an asthenospheric component is more evident than toward a subduction-derived component.

Asthenospheric mantle

As we have seen, the mantle lithosphere shows signs of interaction with the asthenosphere due to the passage of asthenospheric melts. Also the lithosphere-asthenosphere boundary is not fixed for ever, but changes during tectonic evolution. In the Pannonian Basin this boundary has been moved to a shallow depth due to the upwelling of hot asthenospheric mantle material during Tertiary times. The most important consequence was the formation of alkali basaltic and nephelinitic magmas due to pressure release in the upwelling mantle. As alkali basalt generally forms below the mantle lithosphere, its composition may thus provide information on the asthenospheric mantle. However, ascending alkali basalt may interact with the lithospheric mantle en route, by which it can contaminate the lithospheric mantle but also become contaminated by any enriched and low melting point component within the lithospheric mantle.

The petrology, major and trace-element, as well as radiogenic and stable isotope geochemistry of the alkali basalt magmas of the Pannonian Basin (Embey-Isztin et al. 1993a; b; 1995; Dobosi et al. 1995; 1998b) have revealed that the magmas are close to the primitive composition and show little evidence of subsequent alteration or contamination by crustal material. Thus they likely preserved the geochemical and especially isotope geochemical characteristics of their mantle sources. Though the whole rock O-isotope ratios of the alkali basalt (Embey-Isztin et al. 1993) are variable and they signal a shift toward higher values, it was later shown that only the $\delta^{18}O$ values measured on separated olivine and clinopyroxene phenocrysts as well as clinopyroxene and amphibole megacrysts can provide authentic information on the O-isotope composition of the mantle (Dobosi et al. 1998b). In Fig. 9a, b $\delta^{18}O$ values measured on phenocrysts form a rather restricted array (approximately at 5.0–5.1‰), which is characteristic of the pristine mantle. It is interesting to note that while the Sr and Pb isotope ratios show important variations due to the mixing of different components, the corresponding O isotopes do not. This is however, under-

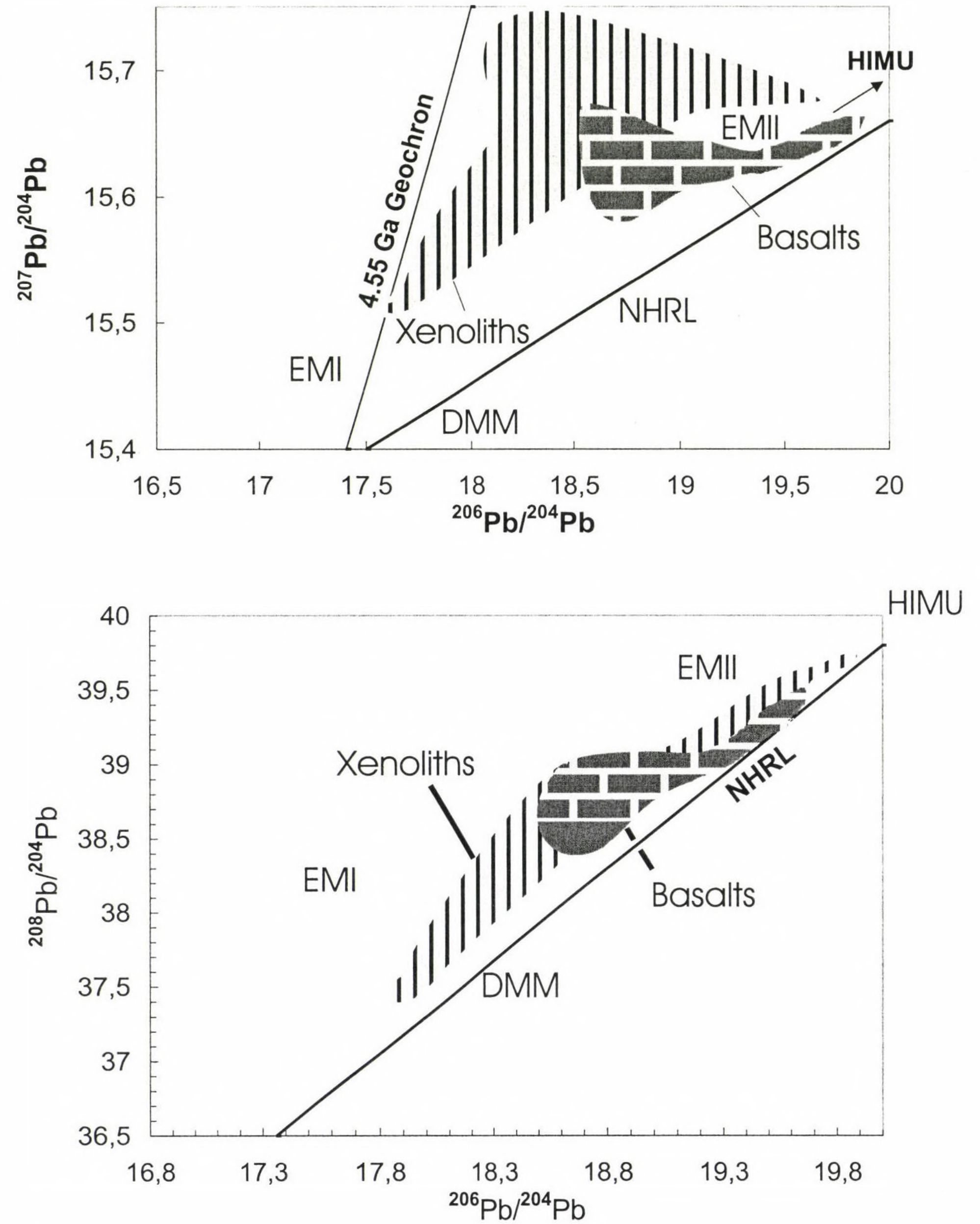
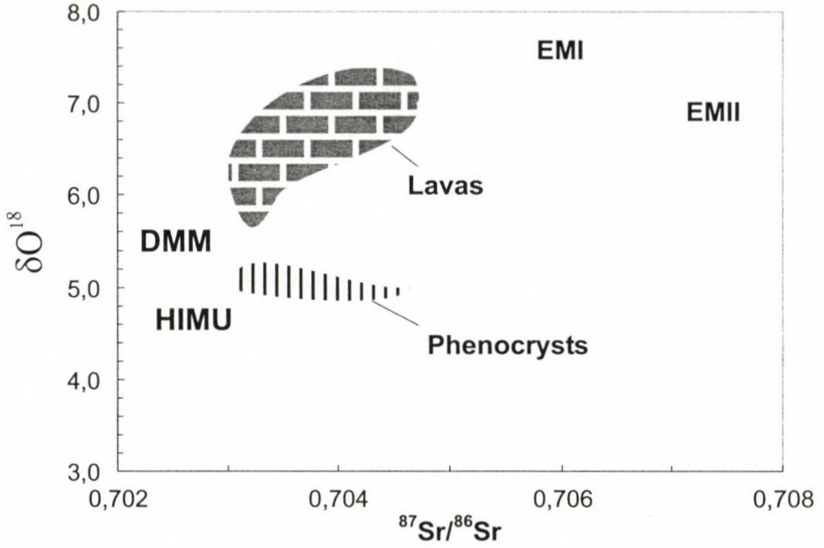


Fig. 8
$^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ (a); $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ (b) isotope diagram for mantle xenoliths and alkali basalts. Data sources: Salters et al. (1988); Rosenbaum et al. (1997); Embey-Isztin et al. (1993a). The NHRL (Northern Hemisphere Reference Line) is from Hart (1984). DMM, HIMU and EM mantle components are taken from Zindler and Hart (1986)

standable as oxygen is the most abundant element in the upper mantle (about 50 wt%); thus the addition of substantial amounts of fluid with distinctly different $\delta^{18}O$ values is required to produce even small shifts in peridotite oxygen isotope



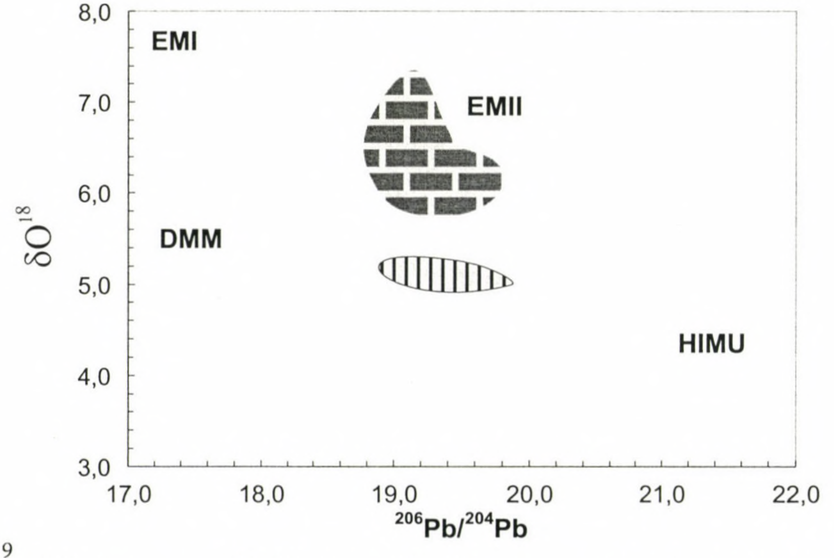


Fig. 9
Plots of $\delta^{18}O$ (in ‰) versus $^{87}Sr/^{86}Sr$ (a) and $^{206}Pb/^{204}Pb$ (b) for the Pannonian basalts and its olivine and clinopyroxene phenocrysts (after Dobosi et al. 1998). Bulk rock data are taken from Embey-Isztin et al. (1993a); Downes et al. (1995a; b). Oxygen isotope values of mantle end-members (DMM, HIMU, EM I and EM II are taken from Harmon and Hoefs (1995))

composition, while a much smaller amount of fluids may cause significant changes in the trace element and Sr and Pb isotopic composition of the same peridotite.

Taking into account both the trace element and isotope geochemical data, Embey-Isztin et al. (1993a) and Embey-Isztin and Dobosi (1995) concluded that the composition of all Pannonian basalt occurrences is dominated by an asthenospheric component. Thus the magmas were formed in the upwelling asthenosphere and not in the mantle lithosphere. The asthenospheric component is characterized by high Nb/La, Ce/Pb, $^{143}Nd/^{144}Nd$, $^{206}Pb/^{204}Pb$, and low $^{87}Sr/^{86}Sr$ ratios. An important observation was that alkali basalt erupted on the northern (Nógrád) and western (Graz Basin) border of the Pannonian Basin is composed exclusively of this component, whereas basalt of the central region shows a more complex composition (see below). In the Pb isotope space diagrams (Fig. 8a, b) the asthenospheric component appears to lie parallel with the NHRL, indicating that it may be in itself a mixture between the depleted mantle (DMM) and HIMU mantle (with extremely high ratios of $^{206}Pb/^{204}Pb$). The HIMU signature (high $^{238}U/^{204}Pb$ mantle end-member) may be indicative of a plume-type, rather than a normal asthenospheric (MORB-type) component. As the influence of this signature is also discernible in the Tertiary alkali basalt of Western Europe, Hoernle et al. (1995) speculated that the Western and Central European alkali basalt originate from a "common European asthenospheric reservoir" (see also Cebriá and Wilson 1995).

Alkali basalt erupted in the central part of the basin (Balaton region, Kisalföld) shows evidence of the influence of an additional, incompatible element-enriched component. The latter component has considerably lower Nb/La, Ce/Pb, $^{143}Nd/^{144}Nd$ and higher $^{87}Sr/^{86}Sr$ and $^{207}Pb/^{204}Pb$ ratios. These characteristics together, but especially the high $^{207}Pb/^{204}Pb$ ratios relative to a given value of $^{206}Pb/^{204}Pb$, strongly suggest the influence of an enriched subduction-derived component (Embey-Isztin et al. 1993a, 1995; Harangi et al. 1995). Figure 8a demonstrates the striking similarity of alkali basalt with the lithospheric mantle xenoliths in this respect. Both rock types have extensions pointing towards the Geochron. The implication of this picture is that the location of the subduction-related enriched component should lie in the mantle lithosphere, and that basaltic magmas of the central region were contaminated by this component during their passage through the lithosphere. In the northern and western border zone of the basin either the mantle lithosphere is devoid of this component, or basalt erupted here failed to tap it.

Discussion

Evidence from the study of mantle peridotite xenoliths indicates that spinel lherzolite is by far the main lithology in the mantle lithosphere beneath the Pannonian Basin. Temperature and pressure estimates of these xenoliths substantiate a hot and thin (~60–70 km) lithosphere. The spatial distribution of different texture types and thermobarometry suggest that the uplift of mantle material culminated below the central part of the basin where lithospheric thinning has resulted in deformation and uplift of the lower portion of the lithosphere upward to shallower depth. The increase in textural complexity, deformation and development of shear zones seems to be connected with metasomatic enrichment of the lithosphere. Shear zones could thus serve as channels for upward transport of fluids and melts during lithospheric thinning.

The xenolith evidence points to a long and complex history of the mantle lithosphere. The dating of the individual events is not yet possible; however a relative chronology can be attempted. The oldest event we can decipher is the depletion in major elements like Al, Ca, Na and Ti, as well as in highly incompatible elements. The depletion is widespread and must be the result of extensive melting and melt extraction. This cannot be correlated with any particular tectonic or magmatic process of the Pannonian Basin, but rather is identical with world-wide trends. The timing of this event is uncertain, but Nd model ages of the highly-depleted peridotites indicate that it can be old, maybe around 1 Ga, but certainly much older than Tertiary.

Compelling evidence exists that the lithosphere had been re-enriched in highly incompatible elements. Sr and Nd isotope compositions indicate that this was a much younger event, certainly younger than a few hundred millions of years. However, several lines of evidence suggest that the enrichment was complex and likely occurred in several steps that were separated from each other in space and time. The observed relationship between metasomatic enrichment and texture complexity suggests that much of the enrichment took place subsequent to the deformation and recrystallization associated with the mantle upwelling in Tertiary times. However, the observation that interstitial amphiboles form texturally-equilibrated grains in equigranular peridotites (Embey-Isztin 1984) indicates that amphiboles were already present in these rocks prior to the recrystallization process. In contrast, the formation of vein hornblendites can be younger on the ground that they cut any previous texture in the host peridotites. This idea is corroborated by the geochemistry and isotope characteristics of the hornblendite veins that are indistinguishable from the alkali basalt. The most probable scenario is that relative to the host lava, the vein hornblendites represent an earlier pulse of alkali basalt that invaded the upper mantle lithosphere somewhat earlier than the final eruption took place. Some of the Type II pyroxenites could have formed in a similar way (Vaselli et al. 1995).

The formation of interstitial amphiboles indicates that enrichment of the lithosphere could have also happened prior to the vein-forming process. Some of the interstitial amphiboles are also LREE-enriched and cryptic metasomatism of peridotites shows trace element and isotope characteristics that may be explained by enrichment due to infiltration of alkaline melts. An even earlier, probably late Mesozoic or early Paleogene carbonatite metasomatism related to an old alkali basalt (Embey-Isztin et al. 1989b) and carbonatite volcanism (Horváth and Ódor

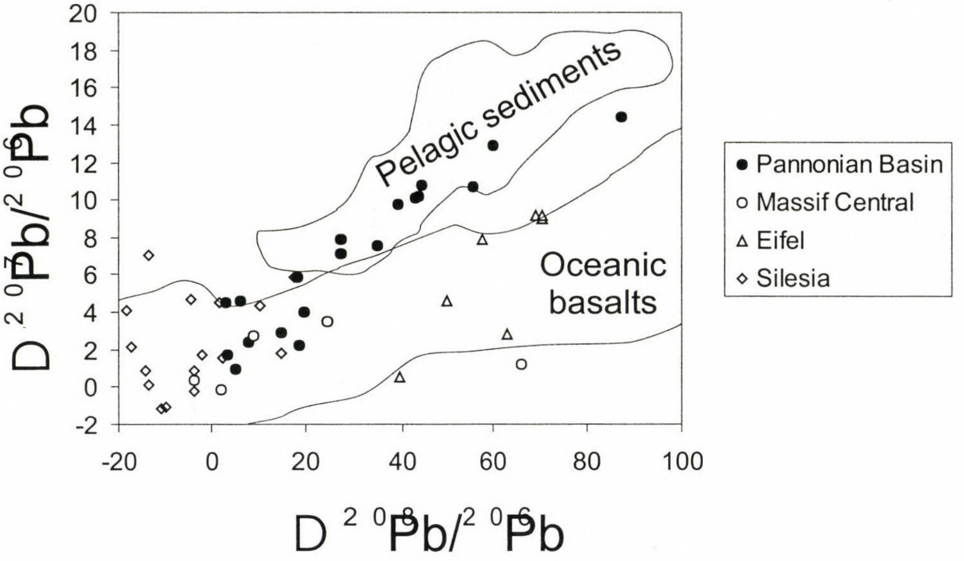


Fig. 10
D(207/204) – D(208/204) Pb isotope diagram for European young alkali basalt (after Embey-Isztin and Dobosi 1998). D-value is after Hart (1988). Full circle: Pannonian Basin (Salters et al. 1988; Embey-Isztin et al. 1993a; Dobosi et al. 1995; Downes et al. 1995), Triangle: Eifel, Open circle: Massif Central (Wilson and Downes 1991), Diamond: Silesia (Blusztajn and Hart 1989). Note that only a part of the Pannonian basalt plots in the field of 'Pelagic Sediments'

1984) was tentatively supposed to have occurred in the mantle (Downes and Vaselli 1995). However, no convincing evidence exists in the mantle xenoliths of the young alkali basalt to support this idea. Carbonates have been observed in glass veins and melt pockets of mantle-derived xenoliths (Szabó et al. 1995b; Bali et al. 1999; Embey-Isztin 2000) and they were interpreted as the products of carbonatite metasomatism (e.g. Bali et al. 1999). However, on the basis of new stable isotope results (Demény et al., in prep.), the carbonates cannot be primary mantle materials as their $\delta^{13}C$ and $\delta^{18}O$ values are far too high and are similar to those of the "secondary carbonates" coating entire peridotite nodules and replacing their olivine grains (Demény and Embey-Isztin 1998). The latter carbonates apparently formed by reaction with the ground water. The presence of this "surface" signature in carbonates hosted by unaltered glass in peridotites can probably be explained by subduction processes (Demény et al., in prep.).

The enrichment due to subduction-related fluids seems to be well documented both in the mantle xenoliths and the basalt. This event can be related to the Tertiary subduction process of the Pannonian-Carpathian region. Mantle xenoliths high in $Sr^{87}Sr/^{86}Sr$ but low in $^{143}Nd/^{144}Nd$ are likely the product of this type of enrichment (Downes et al. 1992). Rosenbaum et al. (1997) arrived at a similar conclusion on the basis of a group of xenoliths, which have higher

$^{207}Pb/^{204}Pb$ at a given $^{206}Pb/^{204}Pb$. This trend can be interpreted as being due to an influx of subduction-zone fluids derived from an ancient subducted sedimentary component. This component is not seen in other regions of Neogene alkaline magmatism of Europe, including the Persányi Mts in the Transylvanian Basin. The alkali basalt lavas show the same picture: only basalt erupted in the central region of the Pannonian Basin exhibits the subduction-related component (Fig. 10; Embey-Isztin et al. 1993a; 1995). However, unlike mantle xenoliths, the alkali basalt of the Transylvanian Basin does show the fingerprint of this component (Downes et al. 1995b).

Acknowledgements

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K/Ar and Ar/Ar geochronological studies in the Pannonian-Carpathians-Dinarides (PANCARDI) region

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Uplift ages of 150.6–127.3 Ma have been detected in the Batocina area, Serbo-Macedonian Massif, and of 160–120 Ma in the Bükkium. Cretaceous metamorphic ages and overthrust ages were measured in the Sopron Mts., in the Hungarian part of the Veporides, in the Bükkium and along tectonic zones of the Tisza Unit. The age of retrogressive metamorphism, associated with ore mineralization in the Central Bosnian Schist Mts. is Eocene–Oligocene.

Triassic gabbro has been dated in the Bükkium, Jurassic magmatic rocks from Jabuka Island in the Adriatic Sea, in the Bükkium and along the Vardar Zone in Serbia. Cretaceous magmatites in and around the Mecsek Mts, S Hungary were partly overprinted during the Austrian and Laramian phases.

Within the Tertiary calc-alkaline volcanism for the entire PANCARDI region, two age groups have been distinguished; an older group, Eocene–Oligocene in age (45–24 Ma), and a younger, Neogene-Quaternary one (ca. 20–0.2 Ma). Radiometric data for hydrothermal minerals and for the unaltered volcanic host rocks suggest close temporal relationships between volcanism and hydrothermal activity.

Paleogene alkaline volcanic rocks are present from Donji Milanovac to Pirot, E Serbia. Alkali basalt in SE Bulgaria is Oligocene and in the Early Miocene of the Moesian Platform. Within the Carpathian Arc alkaline basaltic volcanism began right after the Sarmatian in Austria's Burgenland and about 8 Ma ago in Transdanubia and Central Slovakia. It ended near the Pliocene/Pleistocene boundary in Transdanubia and in the Styrian Basin and in the Quaternary in Central Slovakia and in Transylvania, Romania.

Key words: K/Ar, Ar/Ar, Carpathians, Pannonian Basin, Dinarides, metamorphism, tectonics, igneous rocks, ore mineralization

Introduction

Space and time distribution of orogenic zones, metamorphic rocks, plutonic and volcanic activity, tectonic processes and ore mineralization in the PANCARDI region have been dated by K/Ar and Ar/Ar methods in the Institute of Nuclear Research of the Hungarian Academy of Sciences, Debrecen. The limited space available for this paper restricts us to radiometric chronology, to describing the problems, listing the most important results and drawing attention to the still

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unresolved chronological questions. More detailed geologic background information can be found in the cited original papers and references therein.

Metamorphism, very low-grade metamorphism, tectonic processes and associated ore mineralization

The interpretation of K/Ar ages on metamorphic rocks has been based mostly on the concept of closure temperature introduced by Dodson (1973). According to this concept K/Ar ages give the time when the rock cooled below the "closure temperature" characteristic for the dated mineral. This interpretation was successful in many studies; however, evidence has accumulated against a well-defined closure temperature characterizing the minerals under variable geologic conditions. Villa (1998) argued that in the absence of fluids and stress, closure temperatures are higher than the accepted values, but the presence of fluids and stress greatly reduces the closure temperatures. This does not mean that K/Ar ages are meaningless, but indicates the need for further studies of K-Ar systematics in rocks cooling under different geologic conditions.

A similar observation has been described by Morton (1985), who noted that in a continuously sinking basin where the illite/smectite ratio gradually increased with depth, the age of illite from different depths below about 3200 m was uniformly 23.6 ± 0.8 Ma. This means that illitization was an episodic event at 23.6 Ma before present and, in spite of continuous sinking and temperature increase, the formation of new illite stopped; the age of 23.6 Ma was "frozen in". Morton called this phenomenon "punctuated" diagenesis and explained it by a change in pore-water chemistry.

In the Veporic crystalline basement in Northern Hungary K/Ar and Ar/Ar data on micas and amphibole range from 95 Ma to 87 Ma, i.e. show complete Alpine overprint. Additional fission track ages are only about 10 Ma younger, indicating very rapid cooling during the Late Cretaceous (Koroknai et al. 2000). This rapid cooling is most probably related to the extensional unroofing of the Veporic core complex (Plašienka et al. 1999).

The first K/Ar studies on pelitic sedimentary rocks in Hungary were performed in connection with dating the manganese ore mineralization at Úrkút (Grasselly et al. 1994). Árkai and Balogh (1989) suggested using <2 μm illite-rich fractions and connecting their dating with illite crystallinity (IC) measurement. This method proved to be very successful, opening the way to date orogenic processes in low and very low-grade metamorphic rocks, where other minerals were not available for age determination.

Various lithotypes from the Bükkium (innermost Western Carpathians) were dated in this way. Their metamorphism ranged between the eo-Hellenic (160–120 Ma) and Austrian (100–95 Ma) phases (Árkai et al. 1995). No evidence was found for Hercynian recrystallization.

In the Sopron Mts, NW Hungary, muscovite, biotite and feldspar were dated from leucophyllite and gneiss of the Sopron Gneiss Formation (SGF) and from the andalusite-sillimanite-biotite schist of the Sopron Micaschist Formation (SMF) by Balogh and Dunkl (1998). The youngest ages on fine-grained white mica and biotite, the isochron ages on feldspars and the Ar/Ar age of the lowest temperature step of a fine-grained phengite from the SGF are in the 78.5–71.1 Ma range, which is interpreted as the time of overthrust. However, K/Ar ages on coarse-grained muscovite and Ar/Ar ages at higher temperature can be as old as 160 Ma or >200 Ma, respectively. This confronts thermobarometric results giving >500 °C for the overthrust. On the basis of K/Ar and Ar/Ar measurements a metamorphism of >500 °C temperature cannot be younger than 200 Ma. In the SMF muscovite ages are similar to the ones in the SGF, but biotite ages are older (328.5–235.7 Ma) and Ar/Ar ages are almost plateau-like, suggesting ages from 260 Ma to 240 Ma (Balogh and Dunkl 1998). These older biotite ages are not explained by excess Ar, but by the reduction of muscovite closure temperature. Ar/Ar age spectra point to the possibility of a Permian event.

The metamorphic history and magmatism of the basement of the Great Hungarian Plain and South Transdanubia have been described by Szederkényi et al. (1991) and Lelkes-Felvári et al. (1996). Within the Tisza Unit Szederkényi (1996) demonstrated Alpine overthrust (nappe) structures. Very low and low-grade metamorphism of Mesozoic basement rocks from the eastern part of the Tisza Unit was reported by Árkai et al. (1998). Previous K/Ar age determinations on the metamorphic basement and Variscan granites of the Villány–Bihor and Békés–Codru sub-units have been extended by Ar/Ar dating (Balogh 1999) and, in the frame of a systematic regional study, with K/Ar dating of illite-muscovite-rich <2 μm grain-size fractions from the post-Variscan basement along main structural zones (Árkai et al. 2000). It is observed that along overthrust zones, radiometric ages are mostly overprinted during tectonism, while at a greater distance from the zone the age of an earlier metamorphism is preserved. An explanation can be either that along the overthrust zones rocks from greater depth (higher temperature) approach the surface, or that fluids from the covered rocks facilitate rejuvenation of radiometric ages. Two Ar/Ar spectra recorded on amphiboles (boreholes Biharkeresztes-1 and Szeghalom-4) from the Körös metamorphic complex of the Villány-Bihor Sub-unit suggest Variscan metamorphism with (Late?) Cretaceous overprint. Three muscovites from the Algyő–Ferencszállás metamorphic area (Békés–Codru Sub-unit) resulted in Late Cretaceous (84.36±8.4 Ma) plateau ages. At lower temperature steps the age spectra indicate Oligocene–Early Miocene overprint, which is most likely caused by fluid circulation (Balogh 1999). Dating of <2 μm fractions proved that the age of prograde metamorphism, which affected a considerable part of the post-Variscan basement, is Cretaceous (Árkai et al. 2000).

In the Balkan Peninsula metamorphism and orogeny were dated in the Batocina area, belonging to the northernmost part of the Serbo-Macedonian

Massif (Balogh et al. 1994c). This area is built up of various metamorphic rocks originating from different protoliths. K/Ar dating (i) proved the presence of Variscan metamorphic rocks which were not overprinted during later orogenic phases, and (ii) resulted in an average age of 150.6 ± 5.8 Ma for muscovites and 127.3 ± 4.8 Ma for K-feldspar. These latter ages mark the time when the minerals cooled below their closure temperatures, i. e. they are interpreted as orogenic (uplift) ages. Ar/Ar dating of amphibole with pre-Variscan K/Ar age (951 ± 73 Ma) suggests that the old K/Ar age is caused not only by incorporation of excess Ar but may indicate also a Pan-African event.

The retrogressive metamorphism in the Busovaca region (Mid-Bosnian Schist Mts, central part of the Inner Dinarides, Central Bosnia) has been dated. Ages on whole rock and white mica samples from phyllites and chloritoid schists range from 46.2 Ma to 36.9 Ma. The Ar/Ar plateau age on hyalophane is 31.2 ± 0.3 Ma. This gives the time of cooling below about 250 °C, i.e. the closure temperature for hyalophane determined from the age spectrum. These ages also characterize the stratabound Ba-Cr-Fe-Cu-Zn mineralization of the area. The zoned Cr-spinels, Cr-rich minerals and hyalophane are interpreted as products of retrogressive hydrothermal activity during the final uplift of the Dinarides (Balogh et al. 1999; Bermanec et al. 1999).

Mesozoic magmatism and associated ore mineralization

Most of the Mesozoic magmatic rocks in the studied region are related to the opening of Tethys or to its pre-rifting magmatism. These rocks are frequently affected by the younger phases of Alpine orogeny; therefore, beside dating magmatic activity they can be also used to trace zones of different orogenic phases. Dating was mainly performed on whole rock samples, which are more sensitive to rejuvenating effects (temperature, stress, fluids, etc.). Timing of ore mineralization can be obtained either by dating minerals formed in these processes or by measuring the minerals (or the host rock) that were fully reset at the elevated temperature of mineralization.

In the Bódva valley (N Hungary) previous K/Ar measurements performed on gabbro gave a Triassic age for biotite (233 ± 10 Ma; borehole Bódvarákó-4, 99.3–99.5 m). Feldspar ages were partly Triassic, but partly showed Early Cretaceous rejuvenation. This early Cretaceous overprint was confirmed by K/Ar ages on illite (Árkai et al. 1995). An Ar/Ar plateau age on the same biotite (239.9 ± 1.9 Ma) confirmed the K/Ar age (Balogh 1999) and does not indicate an overprint younger than Triassic.

In the Bükk Mts., at Szarvaskő, muscovite from the contact of the gabbro in the Újhatárvölgy quarry gave an Ar/Ar plateau age of 162.9 ± 0.9 Ma (Balogh 1999), which confirmed previous K/Ar results (muscovite: 165 ± 5 Ma, amphibole: 166 ± 8 Ma). Similar isochron K/Ar ages (160.6 Ma and 159.4 Ma) were obtained on the spilitic diabase from borehole Hejőszalonta-1 (Balogh et al. 1989). On the other hand, K/Ar ages on rhyolite and basalt in the eastern part of the Bükk Mts. (Bagolyhegy, Lillafüred) showed rejuvenation around the middle of Cretaceous. These results are in accordance with ages on white K-mica by Árkai et al. (1995).

The Ditró Syenite Massif is an allochthonous body and forms part of the Bukovina Nappe. It was dated by Pál-Molnár and Árva-Sós (1995). Triassic age is suggested for the intrusion and an overprint during the Alpine orogeny in the middle of the Cretaceous.

Mesozoic alkali basaltic rocks form a zone in the northwestern part of the Tisza Unit. Dating was performed on whole rock and mineral samples from boreholes (Máriakéménd-3, Nagykozár-2, Báta-3, Somberek-1) between the Villány and Mecsek Mts. (Árva-Sós and Ravasz-Baranyai, 1992) and on outcropping rocks and cores (Kurd-2, -3, Döbrököz-1, Dunaszekcső-1, -2) in the Mecsek Mts. and surrounding area (Harangi and Árva-Sós 1993). Ages range from the Jurassic/Cretaceous boundary to the end of the Cretaceous. The great scatter of ages is explained by rejuvenating effects during the Austrian and Laramian orogenic phases. A few Late Jurassic ages need confirmation but are promising for future studies. Cretaceous K/Ar ages were reported on the "bostonites" from Ófalu and Mórágy, Mecsek Mts, by Szabados and Árva-Sós (1996). The authors assume that radiometric ages are rejuvenated, but the extent of rejuvenation and original age of the bostonite dykes remained still uncertain. It is thought that dating of more retentive mineral(s) from the contact could be more successful.

The Ophiolitic Complex of Ždraljica (OCŽ) is situated in Central Serbia around 150 km S of Belgrade. It was emplaced during the Middle Jurassic closure of the Main Vardar Ocean and it now belongs to the eastern branch of the Vardar Zone Composite Terrane, directly juxtaposed to the Serbo-Macedonian Massif. In OCŽ occur also calc-alkaline igneous rocks (pre-collisional granitoids) suggesting the operation of an intraoceanic subduction. K/Ar ages on basaltic rocks were overprinted in the Cretaceous, but an amphibole from a quartzdiorite yielded 168.4 ± 6.7 Ma. This may be the youngest age of the oceanic crust of the OCŽ (Resimic-Šaric et al. 2000a, b).

At the northern margin of the African Plate intensive rifting began in the Middle Triassic; the best-developed rift zone was the Budva-Cukali Zone. A characteristic rock of this zone, the spilite of Becici, was dated. The K/Ar age of 107.3±5.4 Ma indicates rejuvenation in the mid-Cretaceous (Bilik et al. 1993).

The Jabuka Islet (Scoglio del Pomo) is an intrusion of a gabbro crystallized from a slightly differentiated subalkaline magma, located within the Mesozoic succession of the Adriatic Foreland. Ages on fractions of different mineral composition yielded concordant isochron ages (200.3 ± 7.9 Ma and 199.5 ± 11.9 Ma) showing that it was not affected by younger Alpine orogenic phases (Balogh et al. 1994a). Intrusion of the gabbro was probably connected to the extensional tectonic phases (early continental rifting) affecting the Adriatic Plate since the Middle Triassic.

The Mt. Moslavacka Gora crystalline complex is located in the southwestern part of the Tisza Unit. The granitoid pluton (>100 km^2) is bordered by an Abukuma-type metamorphic sequence. Pegmatite is very common in the granite; the one at Srednja Rijeka was dated using muscovite. An Ar/Ar plateau age of 73.2 ± 0.8 Ma was obtained, but further studies are in progress to decide if this age marks the time of magmatic activity or whether it is a rejuvenated one (Palinkaš et al. 2000).

Preliminary radiometric data obtained on core samples of drill holes in the Mura Depression and Drava Depression show that magmatism covered a much longer time span than was previously supposed. Most of them are Late Cretaceous ophiolite and bimodal basalt and rhyolite, Late Oligocene–Early Miocene dacite and trachydacite and Badenian metabasalt and dacitic tuff. The oldest radiometric age of 80.3 Ma was obtained from a gabbro. Younger K-Ar ages between 66.8 to 62.2 Ma were measured on diabase, and 61.1 to 59.9 Ma were determined on quartz-trachyte and alkali-feldspar rhyolite (Pamic and Pécskay, 1994).

Tertiary volcanic rocks of the Slavonija–Srijém Depression are genetically related to the evolution of the Pannonian Basin. The oldest Late Oligocene–Early Miocene trachydacites (31.7±1.2 Ma) are connected to initial extensional tectonism. Penecontemporaneous latite outcrops in the adjacent Mt. Fruska Gora. Radiometric dating on this latite yielded 35 Ma (Knezevic et al. 1991). K-Ar dating indicates that acid and basaltic volcanism also took place during the Paleogene (44 to 32 Ma) in the Slavonija–Srijém Depression and in the Drava Depression.

Within the Banat–Srednagorie Arc, the Poiana Rusca, Ridanj–Krepoljin belt (RKB) and Timok magmatic complex (TMC) have been studied. According to the radiometric data, the Cu- and Pb-Zn mineralization is related to the Late Cretaceous–Paleocene calc-alkaline magmatism of this area.

In the Hauzesti–Drinova region (Poiana Rusca) during the period between about 91 to 71 Ma ago, predominantly granitic rocks were emplaced (Cioflica et al. 1994). Calc-alkaline basaltic andesite and andesite magmas related to subduction erupted 77–66 Ma ago. This andesite was followed at a much later time (48–58 Ma) by small-volume intrusions of mafic alkaline magmas (basanite) (Downes et al. 1995b).

The RKB calc-alkaline volcanic activity most probably took place 74–70 Ma ago. The younger ages can be related to the effects of hydrothermal alteration and/or the thermal influence of the younger intrusions (Karamata et al. 1997a; Pécskay et al. 1998). On the basis of K/Ar data for country rocks of the hydrothermal system that formed the Coka Marin polymetallic ore deposit (TMC), the mineralization occurred approximately 71 Ma ago (Karamata et al. 1997b).

Tertiary-Quaternary magmatism and associated ore mineralization

In order to improve the knowledge of space and time evolution of the Tertiary-Quaternary volcanism in the PANCARDI region, systematic K-Ar age determinations have been undertaken during the last decades. Researchers from all of the PANCARDI countries have been engaged in joint collaboration both among themselves and with colleagues from the UK, Italy and the USA. Numerous projects on individual volcanic regions have been carried out, resulting in numerous publications. Most of the K-Ar age determinations were performed on whole rock samples because of the mineralogy and texture of the investigated magmatic rocks. Mineral separates (biotite, hornblende, feldspar), however, were used for the sake of getting more reliable geochronological information. In the case of dating ignimbrites/tuffs exclusively monomineralic fractions have been measured.

Within the Carpathian–Balkan region of Central and Eastern Europe at least six belts of predominantly andesitic rocks of post-mid-Cretaceous age can be identified (Mitchell 1996)

1. Inner Carpathian Arc
2. Periadriatic–Recsk Arc
3. South Apuseni Mts.
4. Banat–Srednogorije Arc
5. Drina–Rhodope Arc
6. Sumadija–Chalkidiki Arc

The extensive geochronological study which has been carried out on different ore districts of these belts provides radiometric data on the timing of mineralization relative to magmatism. In addition, an estimate of the duration of the lifetime of the hydrothermal system responsible for alteration and mineralization can be made.

Tertiary–Quaternary magmatism can be broadly classified into two types:

1) an earlier phase of volcanism of calc-alkaline affinity,

2) a generally later (though partly overlapping) phase of alkaline volcanism.

Within the subduction-related calc-alkaline volcanism two age groups can be distinguished: an older group, Eocene/Oligocene in age (45–24 Ma), and a younger one, Neogene–Quaternary in age (20.0–0.2 Ma).

Eocene/Oligocene calc-alkaline volcanic rocks

In Hungary, Eocene–Oligocene magmatism is located along, and in the vicinity of, the Periadriatic–Recsk Arc (Balaton Line) from N Hungary to the Zala Basin.

In the Mátra Mts, close to Recsk, the best approach for determining the real age of the magmatism has been amphibole (35.7 Ma). Biotites separated from cores of boreholes Kiscell-1 (Budapest) and Alcsútdoboz-3 yield K-Ar ages of 33.7±1.0 Ma and 31.7±0.8 Ma, respectively. Andesitic samples from the Velence Mts. and its vicinity also show Eocene–Oligocene ages (42–30 Ma). Coupled with the available geologic data on the high sulphidation-type epithermal and Cu-porphyry mineralization of the Velence Mts, the K-Ar ages are used to determine the timing of mineralization. Preliminary data obtained on illite suggest that hydrothermal activity took place between 32.0 Ma–30.0 Ma (Bajnóczi et al. 2000).

Based on the geologic data the andesite and tonalite encountered in boreholes in the Zala Basin are mostly regarded as Eocene volcanic rocks; however, the stratigraphy is uncertain for most of the occurrences. Recent radiometric data obtained on whole rocks and mineral separates (biotite, hornblende and feldspar) also suggest that volcanic activity occurred in the Oligocene; however, in some cases K-Ar ages can be slightly younger than the real geologic age, due to significant tectonic movement and re-heating by Miocene volcanism (Széky-Fux et al. 1991; Pécskay and Balogh 2000).

High-K calc-alkaline and shoshonitic rocks from Srebrenica and Maglaj in the Vardar Zone of the North Dinarides have been dated. According to the K-Ar whole rock ages of 30.4–28.5 Ma these rocks appear to be contemporaneous with shoshonite (latite) from the southeastern Vardar Zone and with Periadriatic tonalite (Pamic et al. 2000).

A detailed geochronological study has been carried out on dacitic and latitic rocks of the Rogozna (Central Serbia). The K-Ar ages range between 31-28 Ma. In accordance with the geologic data the dacitic volcanic rocks are slightly younger than the latite (Karamata et al. 1994).

During the last decades a considerable effort has been devoted to making a correlation of the ages of the various Tertiary volcanic areas in the Rhodopes (Bulgaria and Greece). At present about 300 radiometric data are available to obtain a reasonably complete picture of their space and time evolution. Volcanism started in the Late Priabonian–Early Rupelian with rhyolitic-rhyodacitic explosive events, followed by caldera formation. Bimodal dyke swarms were emplaced between 35–31 Ma. The age of the acid subvolcanic bodies and domes in the Mesta valley and of the voluminous dacitic-rhyolitic high-aspect ratio ignimbrites is 32–28 Ma. The youngest volcanic activity (25–24 Ma) is recorded in the Perelik Center (Harkovska et al. 1998a). The correlation of some Oligocene volcanic complexes along the West-East traverse in Central Balkan Peninsula has been described by Cvetkovic et al. (1995).

K-Ar ages indicate that acidic dykes of Zvezdel were intruded contemporaneously with the paroxysmal Tertiary magmatism of the Rhodopes (31.2–32.2 Ma). Slightly younger ages were determined on the Tsomakovo and Bagrjanka dykes (Harkovska et al. 1998). The oldest radiometric age (39.1 Ma) in the Eastern Rhodopes was measured on a latite and a younger age of 35.8 Ma on a monzonite outcropping in the Yabalkovo-Stalevo region (Yanev et al. 1998). In the Topolovo–Pilashevo dyke swarm acidic and intermediate dykes are exposed. K-Ar ages range between 34.9–28.0 Ma. The older ages reflect the age of the precaldera complex (Harkovska et al. 1998b).

The thickness of Tertiary volcanic sequences in the Briastavo volcano exceeds 1500 m. On the basis of radiometric data two main volcanic phases have been distinguished; (1) 33.4–31.0 Ma and (2) 30.4–29.0 Ma (Yanev and Pécskay, 1997). The age pattern also shows a very short interval for the volcanism of the Perelik volcanic massif. The most probable age of the volcanism is 30.9±0.9 Ma (Pécskay et al. 1991). Close to the Greek–Bulgarian border, in the Kotily–Vitina volcanic massif, an age of 30.3 Ma has been determined by Eleftheriadis and Lippolt (1984). Whole rock samples and biotite, sanidine and plagioclase mineral separates have been measured from the Mesta massif. K-Ar ages indicate that the volcanic activity occurred between 34.3–32.4 Ma (Pécskay et al. 2000a).

Until now the only Miocene age has been determined in the Rhodope Massif, on acidic rocks of Petritsy, SW Bulgaria (12.0±0.5 Ma) (unpublished data).

The low-sulphidation (adularia-sericite type) epithermal gold deposit of Obichnik, Eastern Rhodopes, is associated with mostly andesitic-latitic volcanic rocks of Tertiary age. So far the age of the host rock emplacement (33.5±1.9 Ma) has been determined on a fresh andesitic sample (Kunov et al. 2000).

Miocene-Quaternary calc-alkaline volcanism

Neogene volcanism in the Carpatho-Pannonian Region (CPR) started at about 20 Ma, with high-volume calc-alkaline explosive eruptions. Much of this material is highly siliceous acidic pyroclastic tuffs and ignimbrites, which may have traveled far from their source areas. These rocks are found partly in the Pannonian Basin and partly in the West Carpathians. Recently investigations (including K-Ar geochronology and paleomagnetic studies) were undertaken in the Bükk Foreland area, North Hungary. On the basis of newly obtained data three tuff complexes have been identified, distinguished and characterized. The ignimbritic volcanism occurred between 21–13.5 Ma (Márton and Pécskay 1998; Szakács et al. 1998; Póka et al. 1998).

The most intense activity of Neogene acidic volcanism was the eruption of "Middle Tuff Complex", which has its greatest extent in the Pannonian and Transylvanian Basins (Széky-Fux and Pécskay 1991; Szakács, A., unpublished data).

The main outcropping calc-alkaline volcanoes are predominantly intermediate in composition and form a generally arcuate chain in the Carpathians, with a direct relation to ongoing subduction processes. However, calc-alkaline magmatism in the Danube Basin, Central Slovakia, Northern Hungary and Apuseni Mts. "areal-type volcanism" may be related to the advanced stage of back-arc extension (Lexa et al. 1993).

More detailed reviews of different aspects of Neogene-Quaternary magmatism for the entire CPR and for the Eastern Carpathians have already been published (Pécskay et al. 1995a, b; Downes and Pécskay 1996). Since then a systematic geochronological study has been carried out on different volcanic fields within the CPR. Here we discuss the newly obtained data.

K-Ar dating was performed on whole rock, groundmass and mineral separates (amphibole, feldspar) of andesite intrusions from the Pieniny Mts., Poland. Most of the K-Ar ages range between 13.5 Ma and 11.0 Ma, i.e. Sarmatian (Birkenmajer and Pécskay 1999).

Extrusive domes, lava flows and stratovolcanoes of Vihorlat Mts. (E Slovakia) have been sampled and dated. Against previous geologic expectation radiometric dating proved that andesitic volcanism occurred between 12.6 Ma and 9.4 Ma (Kaliciak et al. 1995).

K-Ar ages were measured on rhyolitic, andesitic and basaltic-andesite samples covering most of the Transcarpathian volcanic centers in the SW Ukraine. New K-Ar data range between 13.4–9.1 Ma, similar to the time interval of the neighboring Carpathian volcanic regions (Pécskay et al. 2000).

Volcanism in the Oas Mts., Romania occurred within a relatively short-time interval (12.9–9.5 Ma) (Kovacs et al. 1997b). The andesitic volcanism in the Gutii Mts. was initiated in the Middle Miocene (ca 13.4 Ma) and ceased at 9.0 Ma. The dominant volcanic phase belongs to the Pannonian (Pécskay et al. 1994). Basaltic intrusive activity followed after a period of about 1 Ma of quiescence (Edelstein et al. 1993).

According to radiometric data an obvious short-distance age progression is seen along the southernmost South Harghita segment. K-Ar ages range between 2.8 to 0.15 Ma. Ciomadul is the youngest volcano of the East Carpathians (Pécskay et al. 1992; Szakács et al. 1993). K-Ar data combined with paleomagnetic and biostratigraphic data from the Apuseni Mts indicate that the Neogene volcanic activity occurred during the Late Badenian–Pannonian (15–7 Ma). The volcanism decreased in the Pannonian (10–7 Ma) and was restricted to the central and northeastern parts of the Apuseni Mts. A much younger age was determined on shoshonitic rock of Uroiu Hill (1.6 Ma), which is coeval with the Southern Harghita shoshonite (Pécskay et al. 1995b; Rosu et al. 1997).

Neogene volcanic rocks crop out in numerous places in the southern and southwestern parts of the Pannonian Basin; however, they are most common in the subsurface, particularly in the Drava Depression, Croatia. These volcanic rocks have been dated and different volcanic formations were distinguished: Egerian–Eggenburgian (28–18 Ma) – andesite with dacite, and andesite with dacite and rhyolite, and post-Badenian (10–8 Ma) – alkali basalt and basalt (Pamic and Pécskay 1996).

So far only very few data are available from the Neogene volcanic rocks buried in the Northern Backa (N Serbia). The only K-Ar age determined on the Cantavir borehole core revealed a Badenian age of 16 Ma (Cvetković et al. 2000a).

The Borac Eruptive Complex (BEC), SW Sumadija in Central Serbia, is built of different volcanic facies of dacites, lamprophyric rocks, quartzlatites and andesites to basaltic andesites. The volcanic activity of the BEC occurred from 23 to 20 Ma. Within this time interval different volcanic events have been distinguished (Cvetkovic and Pécskay 1999). K-Ar ages of phlogopites (22.78 ± 0.88Ma and 22.65 ± 0.89 Ma, respectively) indicate that the hypabyssal lamprophyres of the Borac area formed during the first volcanic phase (Cvetković et al. 2000b).

In the Neogene-Quaternary calc-alkaline volcanic activity of the Inner Carpathian Arc is mostly characterized by low-sulphidation type, epithermal Au-Ag-Pb-Zu-Cu mineralization (Börzsöny, Mátra, Tokaj Mts, Banska Stiavnica, Transcarpathia [Began], Oas-Gutii-Tibles, etc.). However, some volcanic districts also show Cu-porphyry-related high-sulphidation-type mineralization (Banska Stiavnica, Vihorlat, etc.). The interesting features of these Neogene–Quaternary calc-alkaline volcanism are that these mineralizations are related to andesitic-rhyolitic volcanic districts with complex volcanological evolution. Due to the age differences among the volcanic districts along the arc, various erosion levels of the hydrothermal system are exposed. Therefore shallow steam-heated alteration zones are preserved in some parts (e.g. Tokaj Mts); however various deeper adularia-bearing alteration zones with vein-type mineralization are also exposed (e.g. Mátra, Oas-Gutii). The areally distributed rhyolitic tuff units (Pannonian and Transylvanian Basins) and alkaline basaltic volcanic units do not show any significant hydrothermal mineralization.

The main masses of the Börzsöny and Mátra Mts consist of andesitic stratovolcanoes and large caldera (?) structures that host low sulphidation-type epithermal mineralization in siliceous and carbonate veins. Fresh rock and hydrothermal mineral separates (hydromuscovite) of Börzsöny Mts yield K-Ar ages indicating that premineralization volcanism occurred at 16.0–15.0 Ma and that the hydrothermal system was active at 14.6 Ma. (Pécskay and Nagy 1993). A similar conclusion was reached by Korpás and Lang (1993).

Epithermal mineralization of the Neogene Unit is restricted exclusively to the Western Mátra Mts. The low-sulphidation character of hydrothermal alteration resulted in several zones with intense adularia-sericite alteration, outlined by potassium anomalies (Gatter et al. 1999). Well-constrained K-Ar ages for the epithermal deposits were obtained by dating unaltered "host rocks" belonging to the "Middle Andesite Unit" (15.8–15.4Ma) and an intensely altered K-metasomatite (13.7±0.5 Ma and 14.6±0.6 Ma, respectively), adularia monomineralic fractions (13.7±0.5 Ma and 14.0±0.6 Ma, respectively) and also illite mineral separates (13.5±0.6 Ma). Badenian and Sarmatian volcanic cycles of the Tokaj Mts were characterized by intense hydrothermal activity. In general, both volcanic cycles generated low sulphidation-type epithermal deposits with associated alteration. K-Ar ages for adularia from veins, illite from wall rocks of veins as well as alunite from acid steam-heated alteration zones indicate that alteration-mineralization occurred between 10.9 Ma and 12.5 Ma, respectively. Radiometric data for hydrothermal minerals and for unaltered parts of their volcanic host rocks are indistinguishable from each other, suggesting close temporal relationships between volcanism and hydrothermal activity (Molnár et al. 1999).

Recent age determinations and geophysical data from the Baia Mare district suggest that mineralization succeeded volcanism by 1–2 Ma and accompanied crustal extension and cooling of a large pluton. The K-Ar age of the adularia and illite from hydrothermally altered rock indicate that alteration-mineralization occurred during the Pannonian (11.5 Ma–7.8 Ma). On the basis of geologic and radiometric data within this interval two "substages" have been distinguished (11.5–10.0 Ma and 9.4–7.9 Ma, respectively) (Lang et al. 1994; Kovacs et al. 1997).

Basaltic volcanic activity

Basaltic volcanism in the PANCARDI region is mostly alkaline. From among the few exceptions we deal with here are the high alumina basalt and basaltic andesite from the Central Western Carpathians, Slovakia, where the presence of excess Ar necessitated the use of the chronological method elaborated for the Pliocene-Pleistocene alkaline basalt.

Dating of alkaline basalt in Hungary and in Slovakia during the 70s and 80s revealed that K/Ar ages are frequently biased by the presence of excess Ar. Isochron methods, especially when applied to fractions of a single piece of rock differing in their density and magnetic susceptibility, helped to account for the excess Ar and to obtain the real age. When isochron methods also failed, due to the inhomogeneous distribution of excess Ar, the real age could be obtained by selecting for dating suitable fractions on the basis of their atmospheric Ar concentration. This method has been elaborated in the course of a very detailed study of Somoška (Balogh et al. 1994d). A good example for the applicability of this method was obtained when the nepheline basanite of Vel'ké Dravce was dated. Due to the excess Ar, the average age of whole rock samples was 1.90 ± 0.13 Ma; the isochron age on the same samples was 1.61±0.32 Ma. All fractions of a single whole piece of rock did not fit a straight line, but fractions selected according to their atmospheric Ar contents gave the most reliable ages of 1.29 ± 0.34 Ma and 1.27 ± 0.15 Ma.

The alkaline basaltic magmatism (28–26 Ma) of SE Bulgaria follows the most voluminous Paleogene orogenic calc-alkaline and shoshonitic volcanism and is partly coeval with the felsitic rhyolite dykes of the Krumovgrad area. This alkaline magmatism is about 4–5 Ma older than the Moesian Plate alkaline volcanism (Marchev et al. 1997). Basanite samples of the Moesian Platform yielded K-Ar ages 24–16 Ma (Yanev et al. 1993). Alternatively, the Moesian and Rhodopean alkaline basaltic rocks could be related to different structural and magmatic events.

Paleogene mafic alkaline volcanic rocks are distributed along a NNW–SSE belt, stretching from Donji Milanovac to Pirot, Eastern Serbia. K-Ar ages reveal a time span of 62–39.5 Ma, post-dating the Late Cretaceous subduction-related "banatitic" volcanism. This alkaline magmatism preceded the formation of widespread Oligocene-Miocene collision-related igneous provinces of the central axis of Balkan Peninsula (Jovanovic et al. 2000).

Alkaline volcanic activity occurred in the Persani Mts. regions of Romania (eastern Transylvanian Basin) between 2.5 Ma and 0.7 Ma. This volcanism followed an extended period of subduction-related, mostly andesitic and dacitic magmatism in the Eastern Carpathian Arc (Downes et al. 1995a).

As a result of a systematic study reported in a number of publications (Konecný et al. 1995, 1999a, b; Vass et al. 2000) seven volcanic phases of alkaline basaltic volcanism in Central and Southern Slovakia have been distinguished from the Late Pannonian to the Late Quaternary. The relatively low volume of individual eruptions and prolonged volcanic activity indicate a low magma production rate characteristic of this type of intraplate volcanism.

Eruption of alkali basalt was preceded in the Central Western Carpathians by two volcanic formations of high alumina basalt and basaltic andesites. This volcanic phase began at about 12.0 Ma and ended 8.2 ± 0.5 Ma ago, when alkali basaltic volcanism started (Balogh et al. 2000).

During the 90s limited efforts were made in Hungary to improve the chronology of alkali basalt; the new results are mostly unpublished.

The alkali basalt and nepheline basanite in N Hungary form a common basaltic field with the rocks in Southern Slovakia. It has been proved that the smaller necks and lava flows several km west of the plateau basalt of Medves Magosa (Kis-Salgó, Nagy-Salgó, Pécskő, Somlyód) are 2–3 million years older (4–5 Ma) than the plateau basalt.

In Transdanubia the most important result of the last decade is that new data support the reality of the older (7–8 Ma) age of the basaltic tuff at Tihany. This means that further studies are needed to control the age e.g. of Ragonya at Mencshely, where the old K/Ar age of 8 Ma was previously disregarded because it was explained by excess Ar. At present we believe that basaltic volcanism in the Balaton Highland Volcanic Field (BHVC) began about 7–8 Ma ago. On the basis of volcanological-morphological observations Embey-Isztin (personal communication) regarded Bondoró-hill as the youngest basalt in the area. Our measurements of 2.29 ± 0.22 Ma confirmed this opinion: this age marks the end of volcanism in the BHVC.

A highly reliable isochron age of 11.5 ± 0.7 Ma has been obtained on the diabase of the Pauliberg in Burgenland and a similar age (11.1 ± 1.2) was measured on the olivine basalt of Pullendorf. These are the first products of post-Sarmatian alkaline basaltic volcanic activity in Transdanubia (Balogh et al. 1994b).

In the Styrian Basin basaltic volcanic activity began at Neuhaus 3.76 Ma ago and ended at Wilhelmsdorf at 1.72 Ma (Balogh et al. 1994b).

Ore mineralization in sedimentary and metamorphic environment

The timing of oxidation of Mn-carbonate ore (formation of cryptomelane) of Úrkút (Bakony Mts, Hungary) has been dated by Grasselly et al. (1994). The fine grains of cryptomelane did not allow mineral separation; thus the cryptomelane age and its meaning was obtained by dating carefully selected whole rock samples, considering the different effects that might have influenced K/Ar ages and performing several additional experiments. While K/Ar ages on clay mineral-bearing Mn-carbonates scattered from 293 Ma to 110 Ma, most of the ages of Mn-oxide ores were in the 105-91 Ma interval, the rest being older. Cretaceous ages might be the result of an overprint, frequent in the Carpathian Basin in the Cretaceous, but they can indicate the real age of oxidation. A regional overprint in the Cretaceous was ruled out by the older K-Ar ages; the possibility of local, e.g. fluidal effects was checked by dating Mn-oxide nodules and their clayey crust. The crusts of Cretaceous Mn-oxides were older. This does not rule out the possibility of local overprint if the Ar retentivity of cryptomelane is less than that of the clay minerals. Ar-release studies proved that cryptomelane retains Ar better than the clayey crust, so ages on Mn-oxides give the time of oxidation. The older ages on Mn-oxides nodules, as demonstrated by dissolution of cryptomelane and dating the residue, were caused by detrital contamination.

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On the structure and tectonic evolution of the Pannonian Basin and surrounding orogens

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An overview of the Cenozoic tectonic evolution of the PANCARDI (PANnonian–CARpathian–DInaric) region with special emphasis on the Neogene through Quaternary deformation history of the Pannonian Basin system is presented here. First, the tectonic framework is defined by introducing the principal features of the Cenozoic structural evolution of the wider geologic environment, the Alpine–Mediterranean system. This is followed by a historical overview of the kinematic models explaining the deformation history of the PANCARDI system. Differences in plate tectonic models of translation, rotation and internal deformation of the major structural units highlight some of the most peculiar tectonic problems addressed in this study. Several dynamic considerations about the Neogene history of the back-arc-type Pannonian Basin may shed light on the physical problems related to the driving forces behind extensional basin formation in an overall compressional setting, i.e. the Alpine–Carpathian orogenic belt. Thereafter the present-day build-up of the PANCARDI system is further described by a general introduction of the physical properties of the underlying lithosphere.

We conclude with a general discussion on the most recent geodynamic scenarios that attempt to integrate the results of various fields of tectonic studies from field measurements through basin analysis to numerical modeling and some related geophysical techniques. We demonstrate the supreme importance of the rapid changes of boundary conditions, plate boundary forces and internal body forces affecting the PANCARDI system. The formation of the Pannonian Basin (Early Miocene) was initiated and to a great extent controlled by the combination of two main processes. These are the collisional forces exerted by the indentation of the Adriatic plate against the Alpine–Dinaric chain on the one hand and orogen-parallel extension due to the presence of a thickened, asymmetrically elevated, thermally weakened and laterally unconstrained lithospheric segment, the ALCAPA block on the other. During the Middle Miocene to Pliocene period the deformation history was mainly governed by the boundary conditions acting along the Carpathian zone of subduction. The main source of extension in the Pannonian Basin was a set of arc-normal trench suction forces acting at the interface of the overriding plates, i.e. the North Pannonian and Tisza–Dacides units, and the descending lithosphere, i.e. the subducting Magura Ocean. The termination of subduction along the entire arc together with collisional forces due to the further indentation of the Adriatic Microplate has resulted in the change of tectonic regime in the Pannonian basin from general extension to gradual positive inversion during late Pliocene–Quaternary times.

Key words: Pannonian basin, back-arc basins, extension, inversion

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Tectonic framework: Cenozoic evolution of the Alpine–Mediterranean system

The Pannonian Basin is located, at least in a geologic sense, in the northern sector of the Central Mediterranean. Orogenic belts of multiple-compression origin from Late Jurassic to recent times surround this extensional basin, which itself was formed during the Neogene through Quaternary period. To understand its tectonic evolution one should first examine the tectonic development of its broader geologic surroundings, the Mediterranean region (Fig. 1a). This wide zone of convergence between the Eurasian and African plates went through a polyphase deformation history since the opening of the Atlantic Ocean. This latter process represents the ultimate cause for the formation of the Alpine orogenic belt (e.g. Biju-Duval et al. 1977; Dercourt et al. 1986; Dewey et al. 1973, 1989; Şengör 1993; Yilmaz et al. 1996). One of the most striking features in this compressional setting is the abundant occurrence of extensional basins (Horváth and Berckhemer 1982; Cloetingh et al. 1995). From west to east, these are the Alboran, Balearic, Ligurian, Tyrrhenian, Pannonian and Aegean Basins. Although their age, nature and origin are far from being identical, there are several common features in their evolution. The most important appears to be that they are all located in the backyard of a once or still active subduction front. This suggests a casual and genetic relationship of primary significance (e.g. Royden 1993; Giunchi et al. 1996; Meijer and Wortel 1997). Other possible mechanisms, such as the collapse of gravitationally unstable orogenic crust (Dewey 1988), the lateral escape of microplates (Şengör et al. 1985) or their combination (Ratschbacher et al. 1991) may have also significantly contributed to the formation of these basins. However, their relative importance is still a subject of ongoing debate (for a discussion see Şengör 1993).

Using information obtained by the application of seismic tomography techniques, Wortel and Spakman (2000) proposed a model for the Cenozoic dynamic evolution of the Mediterranean system. Although numerous details of the model still require more evidence and improvement, it serves as a working hypothesis for explaining the evolution of Mediterranean back-arc basins, including the Pannonian Basin system. With the aid of the combined process of retreat and lateral detachment of the subducted slabs between the southern edge of the Eurasian Plate and the northern edge of the African Plate, a multi-stage dynamic model was set up (Fig. 1b) that led to the present-day configuration of the system. Slab detachment occurs in the form of the slab's self-perpetuating tear off under its own weight. This can happen when thick and buoyant continental lithosphere enters into the trench zone. More precisely, the process of subduction or underplating is blocked when the positive buoyancy of the arriving continental crust becomes equal in magnitude to the negative buoyancy of the already subducted oceanic lithosphere. Lateral migration of slab detachment results in the concentration of tensional forces on a continuously decreasing portion of the arc which then itself leads to a gradual, sometimes accelerating retreat of the trench system (Dvorkin et al. 1993). In the model, this process is

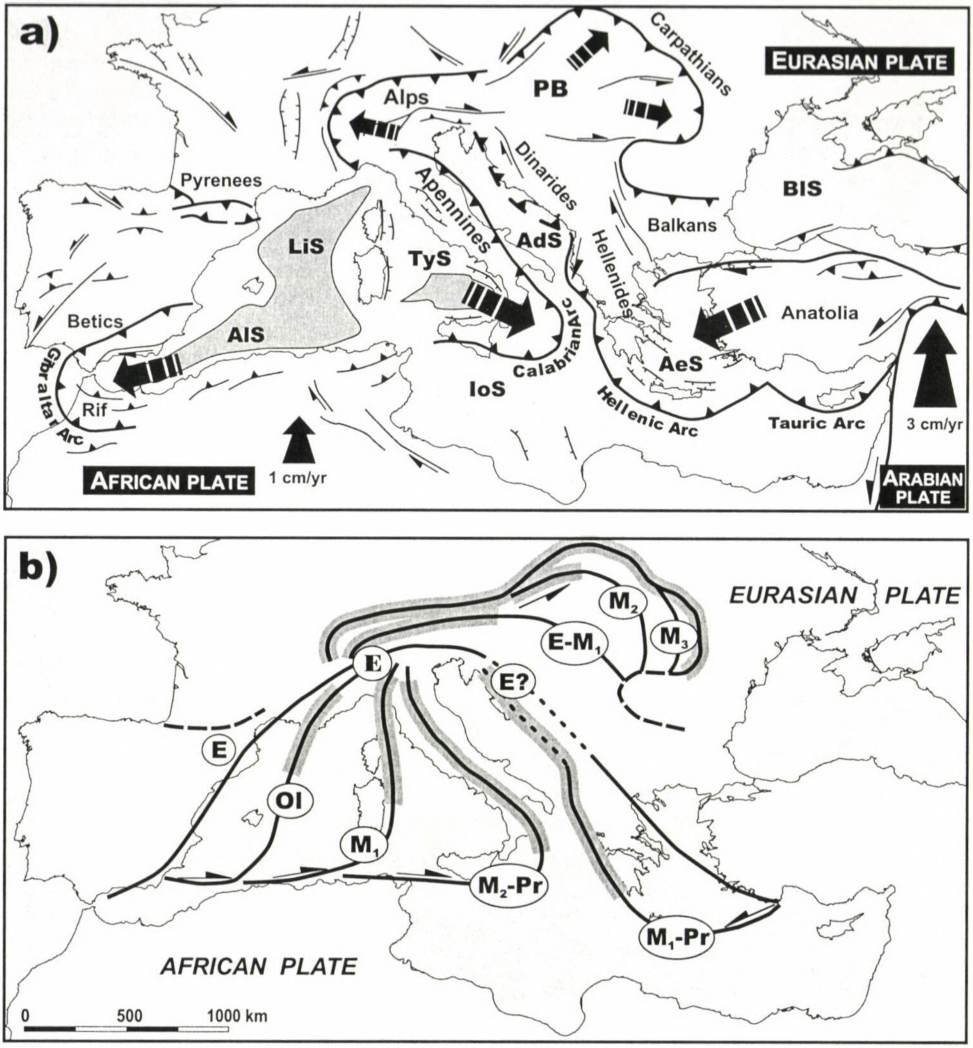

Fig. 1
(a) Sketch map showing the present-day configuration and the late Cenozoic structural features of the Mediterranean and adjacent areas. The overriding plates above subduction zones have suffered rigid body rotation, translation and extension that led to the formation of back-arc basins. Darker shading indicates oceanic crust. PB – Pannonian basin; AdS – Adriatic Sea; AeS – Aegean Sea; AlS – Alboran Sea; BlS – Black Sea; IoS – Ionian Sea; LiS – Ligurian Sea; TyS – Tyrrhenian Sea. (b) Dynamic model for the evolution of the same area during the Cenozoic proposed by Wortel and Spakman (2000). Solid line: area of active subduction, thick grey line: area where slab detachment already occurred at the given time. E – Eocene; Ol – Oligocene; M_1 – Early Miocene; M_2 – Middle Miocene; M_3 – Late Miocene; Pr – present

responsible for arc migration and, as a consequence, back-arc basin formation from the Oligocene until recent times in the Western and Central Mediterranean and the build-up of frontal compression at the orogenic chains of the Betics, Rif and Apennines (e.g. Dercourt et al. 1986; Dewey et al. 1989). The formation of the Pannonian Basin in a similar setting was somewhat delayed, beginning during the Early Miocene (Horváth 1993), while it appears that extension had already come to an end in this region (Horváth and Cloetingh 1996) due to the complete consumption of the subductable lithosphere of the European foreland. In the Eastern Mediterranean the Aegean Sea was probably formed during a first stage of rifting in the Early Miocene (e.g. Robertson and Dixon 1984), and the Hellenic arc with seaward trench retreat represents a presently active region of subduction (cf. Jackson 1994).

Cenozoic evolution of the PANCARDI system: kinematic considerations

The PANCARDI system occupies the northeastern sector of the Alpine-Mediterranean system (Fig. 2). The curving Alpine chain splits into two belts at the western edge of the Pannonian basin. The Dinarides and the Carpathians become again unified in the vicinity of the Moesian Platform. The Cenozoic orogeny of the region is mainly governed by the northward drift and indentation of the Adriatic promontory, producing a total amount of convergence of about 500 km in the Western and Central Alps (Schmid et al. 1996) and 600 km across the Eastern Alps (Roeder and Bachmann 1996). More than two-thirds of this convergence took place during Early Paleogene times. In the proximity of this zone of intense shortening the basement of the Pannonian basin has suffered an overall extension of about 150–200 km in E-W to NE-SW direction during Miocene-Pliocene times (Tari et al. 1995), which is quite similar to the amount of coeval shortening at the Carpathian Arc (Ellouz and Roca 1994; Roca et al. 1995). From the viewpoint of material balance this may shed some light on the relationship between the structural evolution of the basin system and the surrounding Carpathian fold-and-thrust belt. However, as already pointed out by Tari (1994) and Fodor et al. (1998), the importance of the push exerted by the northward drift of the Adriatic microplate should not be underestimated. This process significantly contributed to the geometry and arrangement of the structural elements in the entire area.

Several kinematic models, i.e. palinspastic reconstructions, have been published for the Cenozoic evolution of the PANCARDI region since the general acceptance of the theory of plate tectonics. The Late Jurassic through Cretaceous history is beyond of the scope of the present study, although this period certainly has important bearings on the present-day structure of the region. For instance, the internal architecture of the first-order structural units constituting the basement of the Pannonian Basin (Fig. 3) was already configured before the Tertiary, mostly during Cretaceous times. Considering their easily traceable

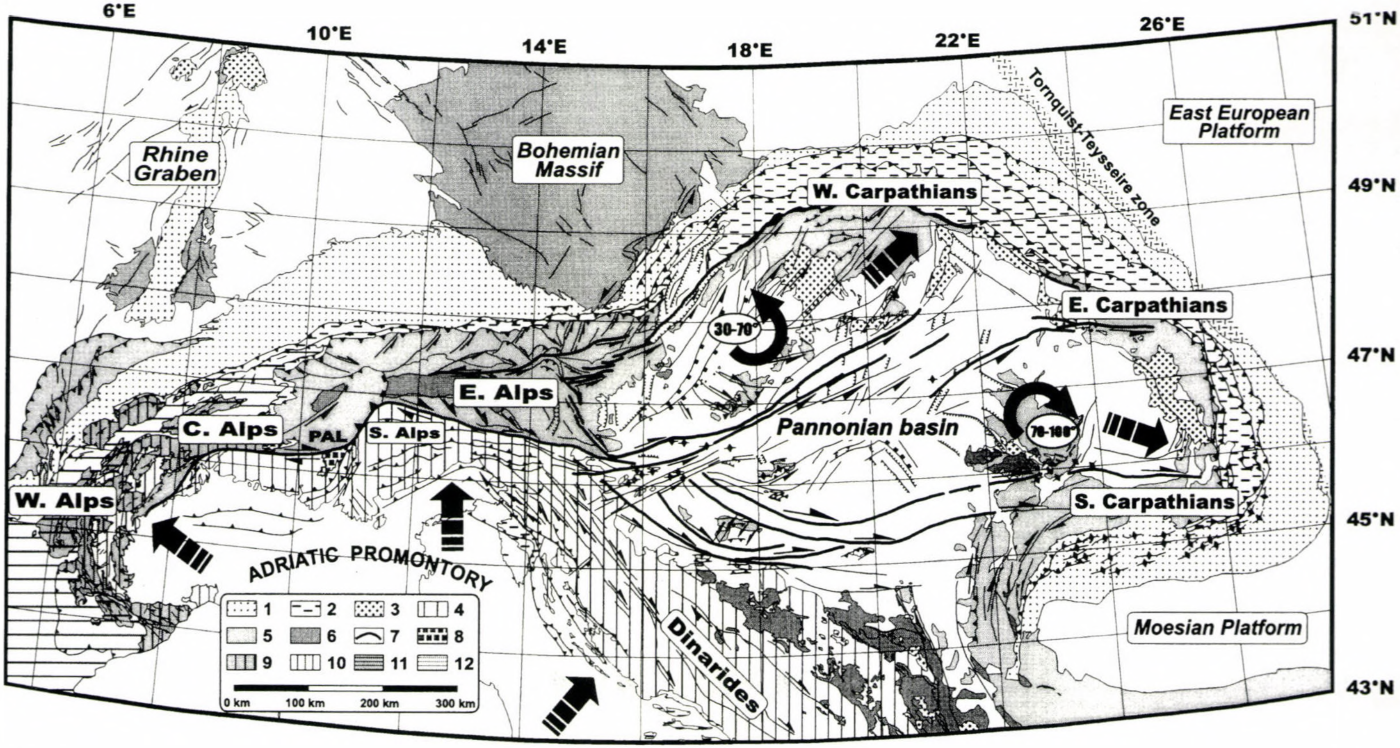

Fig. 2

Simplified Late Cenozoic tectonic map of the Alpine–Carpathian–Pannonian–Dinaric system. The Adriatic promontory or microplate has been indenting and pushing the Alpine–Dinaric belt since the Cretaceous. The northern domain beneath the Pannonian Basin was rotating counterclockwise and was translated to the east-northeast, while the southern unit rotated in an opposite manner and was moving to the east-southeast. Arrows indicate the translation and rotation of various tectonic units. See text for references of the kinematic and paleomagnetic data. 1. foreland (molasse) basins; 2. flysch belts; 3. Neogene volcanites; 4. Southern Alps and Dinarides; 5. pre-Tertiary units of the East Alpine–Carpathian domain and the Jura Mts; 6. Variscan basement of the European Plate; Vardar, Mures ophiolites; 7. Pieniny Klippen Belt; 8. Oligocene tonalites; 9. Penninic basement; 10. Penninic cover; 11. Helvetic basement; 12. Helvetic cover, PAL – Periadriatic Line (PAL)

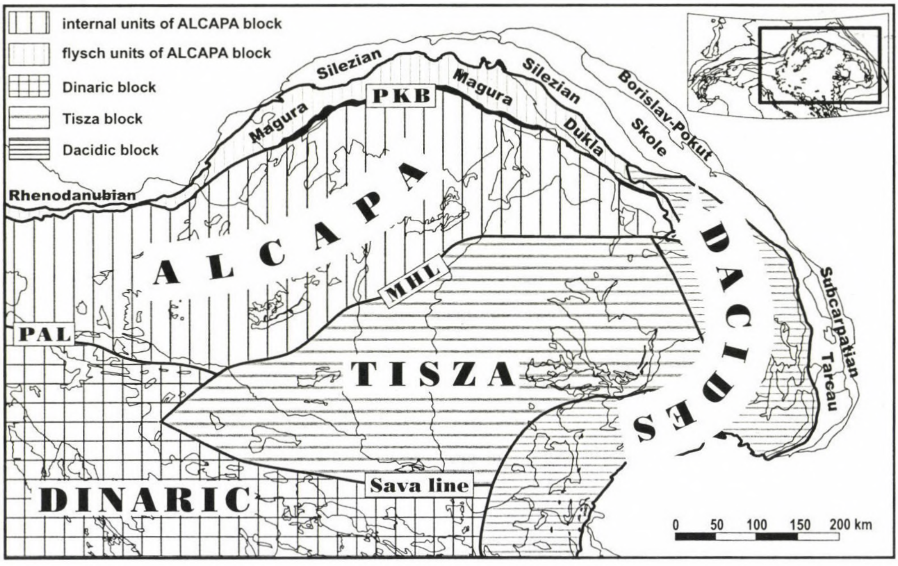

Fig. 3
Present-day configuration of the main structural units in the PANCARDI system (after Balla 1988; Csontos 1995; Fodor et al. 1999). The internal crustal blocks (e.g. ALCAPA, Tisza, Dacides) are surrounded by the external nappe piles of the Carpathian fold-and-thrust belt. MHL – Mid-Hungarian Line; PAL – Periadriatic Line; PKB – Pieniny Klippen Belt. Upper right inset shows the location of the study area as in Fig. 2

boundaries and distinct Mesozoic and Cenozoic tectonic behavior, these continental blocks form key elements in any of the numerous Cenozoic reconstructions. The ALCAPA block of Apulian origin (in Tethyan terms) is situated north of the Mid-Hungarian Line and, at least sensu Csontos (1995) and Fodor et al. (1999), contains two of the most internal flysch nappes of the Western Carpathians, the Magura and Dukla Nappes. The ALCAPA block in a much stricter sense, i.e. without the Eastern Alps and the two above-mentioned flysch nappes of Paleogene tectogenesis, is also known as the North Pannonian Unit (=NPU - Balla 1984) referring already to the evolution of the Pannonian basin. In this sense this unit is limited in the north by the Pieniny Klippen Belt, an intensely deformed, narrow and nearly 600 km-long shear zone made up of a nappe pile of Mesozoic rocks (Birkenmajer 1986). The Tisza and Dacides Units, separated from the southern margin of the European plate during the Middle Jurassic (cf. Géczy 1973), together form the South Pannonian Unit (=SPU). The Intra-Carpathian area is bordered to the southwest by the fold-and-thrust belt of the Dinarides, which itself went through a complex evolution of multistage convergence. These internal masses are surrounded by the continuous loop of an external fold-and-thrust belt made up of the Cretaceous-Lower Miocene flysch sequences of the Carpathians.

The tectonic evolution of the Pannonian Basin and its surroundings represents a polyphase deformation history with separate structural episodes in space and time (Fig. 4). There is overall agreement on the principal kinematic features, i.e. the location and activation of major fault zones and the timing and amount of deformation (for a compilation of kinematic data see Csontos 1995 and Fodor et al. 1999). The rotation history of the main tectonic terranes has been constrained with the aid of paleomagnetic data collected during the last decade (e.g. Márton 1990, 1993, 2001). Data show a characteristic opposite rotation pattern of the basement units in the Pannonian area. The NPU suffered a rapid, ca. 30–70° counterclockwise rotation while the roughly coeval rotation of the SPU was ca. 70–100° in the opposite sense. The bulk of the rotation took place during Early Miocene (mainly Ottnangian) times.

As a very first approach two distinct Cenozoic periods can be defined. The earlier is characterized mainly by compressional deformation in both the internal and external areas during the Paleogene–earliest Miocene (Eggenburgian) (cf. Tari et al. 1993). This is a direct consequence of the Mesoalpine convergence between the Adriatic and European Plates in a dextrally oblique, transpressional manner. At the leading edge of the Adriatic Plate in the Eastern Alps, nappe stacking was dominantly NW-directed (Linzer et al. 1997; Peresson and Decker 1997). In the external zone of the Carpathians, the overthrust of the most internal flysch nappes (i.e. Magura Nappe) already began during the Late Eocene implying a small amount of northward movement of the internal units as well (Nemčok 1993). For the areas characterized by Paleogene sedimentation inside the Carpathian Arc, either flexural or piggyback origin has been put forward by different authors. The Hungarian and Slovenian Paleogene Basins were interpreted as a retroarc flexural basin in the hinterland of the Western Carpathian compressional arc (Tari et al. 1993). In the northern part of the Transylvanian Basin, which was probably in connection with the Intra-Carpathian Paleogene Basin (Tari et al. 1995), basin formation and evolution was controlled by the overthrust of the ALCAPA unit onto the Tisza unit (Csontos and Nagymarosy 1998; Györfi et al. 1999).

The Paleogene–Early Miocene history of the PANCARDI region was followed by a rapid and dramatic change in tectonic style during the late Early Miocene that resulted in the formation of the Pannonian Basin. Consequently, the relatively stable Paleogene to Early Miocene assembly of continental blocks at the axial zone of Adria-Europe convergence was completely disintegrated and the units experienced significant amount of rigid body rotation and translation. This process was partly coeval with the formation and early evolution of the Pannonian Basin and the large-scale tectonic transport of the flysch nappes in the Carpathian Arc. In his plane-view kinematic reconstruction, Balla (1984, 1986,

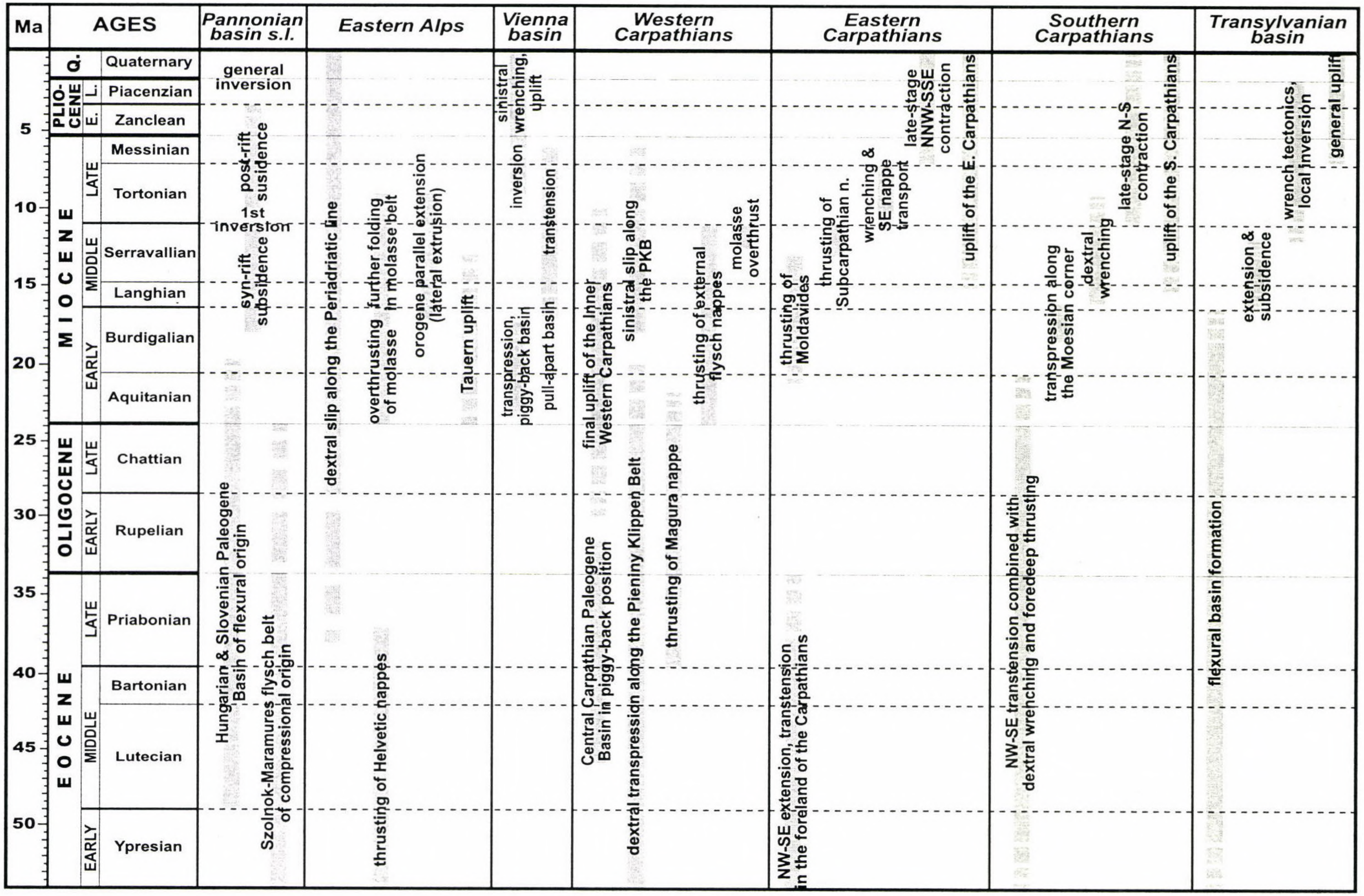

Fig. 4.
Simplified Cenozoic evolutionary chart of the main tectonic domains in the PANCARDI system showing the duration and characteristics of the major structural episodes. The diagram serves as a basis for comparative purposes and to analyze the differences and similarities of tectonic styles in various sectors of the area

1988) was the first to suggest that the main structural domains, i.e. the North Pannonian Unit north of the Mid-Hungarian Line and the Tisza and Dacia (i.e. the South Pannonian Unit) units south of it, were rotating against each other. According to him, this process was basically coeval with the eastward movement of the ALCAPA block, which was expelled from between the converging Bohemian Massif and Adriatic promontory. Beyond this, the main driving force of plate movements in this system was the northward drift and counterclockwise rotation of the SPU. As a consequence, the intense push between the two terranes resulted in active shearing and the development of radial extension along their interface. This deformation was directly responsible for the development of the Intra-Carpathian basins in such a way that a system of narrow grabens connected by transform faults were formed in the vicinity of the contact zone of the NPU and SPU. The main ideas of the pioneering work of Balla (1984), at least in principle, were very often accepted, adopted and developed in later reconstructions. The opposite rotation of the two terranes, however, creates severe space problems, which were unanswered by this author. Furthermore, the general transtensional character of the early (syn-rift) period of the Pannonian Basin can hardly be reconciled with a model of "active push from the south" proposed by Balla (1984). An alternative way of thinking was offered by Csontos et al. (1992). They proposed that the northern and southern units were juxtaposed during the Late Oligocene–Early Miocene period. This model implies that internal units had already nearly reached their present-day position before the onset of overall extension in the Intra-Carpathian realm. Consequently, there was insufficient space left for stretching the lithosphere underlying the Pannonian Basin. In addition, they assume no rigid body rotation for either the NPU or the SPU, which is now difficult to accept from a kinematic point of view. Kováč et al. (1994) combined the two previous scenarios and came up with the conclusion that the large-scale right-lateral slip along the Mid-Hungarian Line, the opposite rotation of the two terranes and, implicitly, the extension in the Pannonian Basin are coeval and genetically related processes.

Accepting and slightly modifying the results of previous reconstructions and, furthermore, employing kinematic indicators, Fodor et al. (1999) suggested a detailed plane-view restoration of the Miocene evolution of the PANCARDI region. The model indicates the most essential boundary conditions affecting the entire system (Fig. 5). The North (NPU) and South Pannonian Units (SPU) occupied the Carpathian embayment in four successive steps. The irregular geometry of the rigid and immovable foreland plays a key role in the reconstruction. The onset of extension and the opposite rotation of the NPU and SPU were coeval during the Late Ottnangian (at ca. 18–20 Ma) and mark a plate configuration where the retreating Alpine-Carpathian subduction front passed through the corners composed by the Bohemian Massif and the Moesian Platform (Fig. 5a). These two buttresses formed a bottleneck-like geometry at the entrance of the Carpathian embayment and to a great extent influenced the

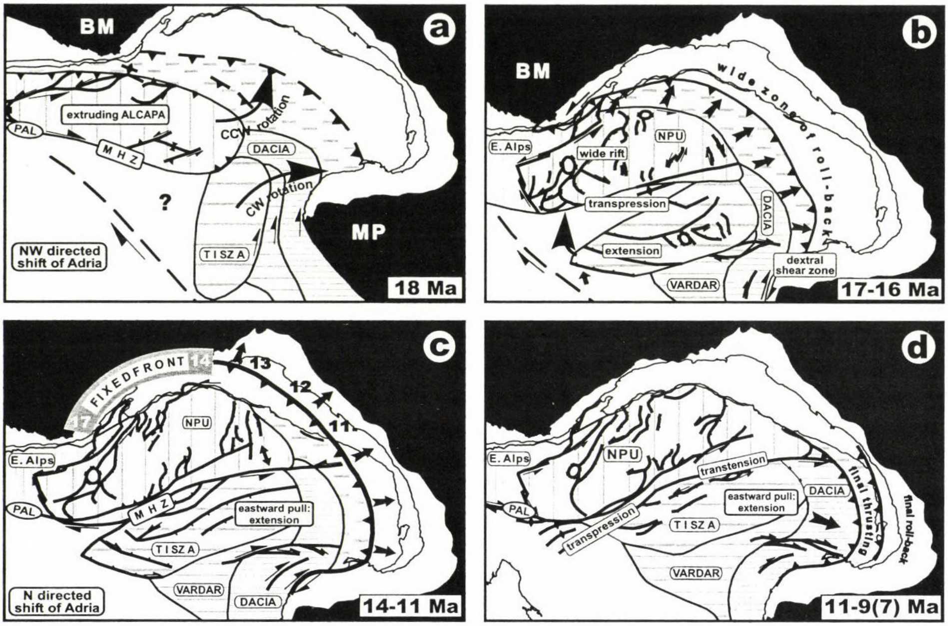

Fig. 5
Late Early through Late Miocene plate tectonic reconstruction of the PANCARDI region proposed by Fodor et al. (1999). (a) Late Ottnangian (18 Ma); (b) Karpatian (17–16 Ma); (c) Middle Miocene (14-11 Ma); (d) Late Miocene (11-9(7) Ma) configuration. BM – Bohemian Massif; MHZ – Mid-Hungarian Fault Zone; MP – Moesian Platform; NPU – North Pannonian Unit; PAL – Periadriatic Lineament. Numbers in panel (c) indicate final thrusting of the Carpathian flysch units after Jiříček (1979). See Fig. 3 for the present-day configuration of the tectonic units

strain and stress pattern. In this geometry the eastern edge of NPU and SPU and the Carpathian subduction front had a continuous and smooth curvature that gradually advanced towards the NE and underwent reduction of radius through time. During Karpatian times (17.2–16.4 Ma), the Intra-Carpathian area was influenced by a rapid hinge retreat of the subducting slab and, as a consequence, both units were pulled and extended along a wide zone of subduction roll-back (Fig. 5b). The Middle Miocene evolution is characterized by the complete consumption of the subductable margin of the European foreland in front of the northern unit (Fig. 5c). As indicated by the subsidence pattern in the Alpine foredeep, a slowdown of continental convergence marked by the transition from flysch to molasse sedimentation occurred during the Middle Oligocene (25 Ma) in the Central Alps and during the Early Miocene in the Eastern Alps (22–18 Ma) (Sinclair and Allen 1992; Sinclair 1997). This temporal pattern is traceable further to the east along the Carpathian arc (Meulenkamp et al. 1996): the final collision between the North Pannonian Unit and its foreland commenced during the Karpatian (17 Ma) and the NPU reached its present-day position at the end of the

Middle Miocene (11.5 Ma). This process proceeded until the Late Miocene when the South Pannonian Unit became docked (Linzer 1996; Maţenco 1997) (Fig. 5d).

The situation is also quite controversial when one starts dealing with neotectonics, i.e. the history of the last few million years of the region. Although the average level of knowledge about the geology of the area in general increases as we approach the present, the neotectonic features are not yet sufficiently well defined to judge their nature and origin unambiguously. Nevertheless, together with the advent of sophisticated techniques of stress measurement methods, seismology, space geodesy and computer sciences, it now appears possible to work out comprehensive models to describe the contemporary tectonic behavior of the Intra-Carpathian area. It has been recently recognized that extension has already terminated and that general inversion of the Pannonian Basin is in progress (Horváth 1995; Gerner et al. 1999; Bada et al. 1999). The latest results of GPS measurements (Grenerczy et al. 2000) suggest that the ALCAPA unit is still moving to the east-northeast. Since no space is available at its leading edge due to the final termination of subduction in the Carpathians, this movement may indeed result in buckling of the entire Pannonian lithosphere as proposed by Horváth and Cloetingh (1996) and increased seismicity within the basin system with respect to the European foreland or the main parts of the Carpathian orogen. As an ultimate consequence, extensional basin evolution has come to an end and positive (compressional) structural inversion of the Pannonian Basin is in progress.

Formation and evolution of the Pannonian Basin: dynamic considerations

The previous section overviewed some kinematic models that deal with the movement (translation, rotation) and internal deformation of the various tectonic units constituting the PANCARDI system. In the following we provide insights into the dynamic behavior and evolution of the system. The emphasis is on the causes of basin formation and evolution and, in addition, the physical processes influencing the present-day architecture of the Pannonian Basin and its surroundings.

The most evident features of the PANCARDI system, i.e. thinned and hot vs. thickened and cold lithosphere in the central and peripheral sectors, respectively, provoked various theoretical solutions even before the general acceptance of plate tectonics. For instance, Szádeczky-Kardoss (1967), Szénás (1968) and, at least in his early works, Stegena (1967) argued for the presence of a mantle diapir beneath the Intra-Carpathian area (Fig. 6a). According to these authors, ascending mantle flow resulted in thinning and subsidence (active rifting) in the central areas, whereas the nappe structure and the root of the surrounding mountain belts were formed above the descending branch of this local convection cell.

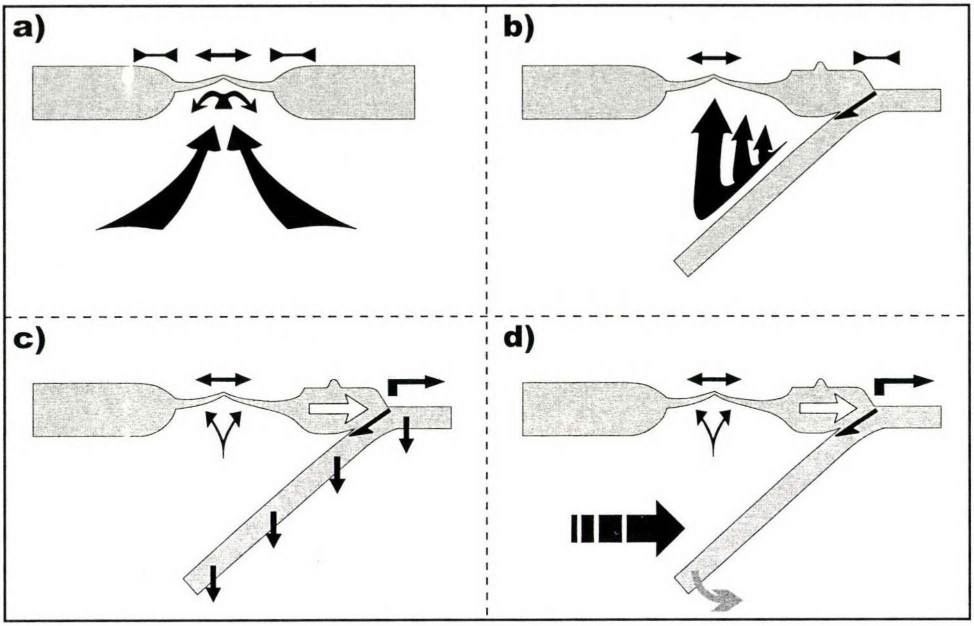

Fig. 6
Dynamic models proposed for the formation and evolution of the Pannonian basin system. (a) Asthenospheric doming can cause active rifting of the lithosphere above the central axis of the dome, whereas shortening is taking place at the peripheral areas (e.g. Stegena 1967). (b) Active rifting may also be caused by a subduction generated mantle diapir (e.g. Horváth et al. 1975; Stegena et al. 1975). (c) Hinge retreat of the subducting European margin driven by the negative buoyancy of the slab that induces trench suction forces and, hence, passive rifting in the overriding plate (cf. Royden and Karner 1984). (d) The same hinge retreat may be sustained by an eastward mantle flow pushing against the descending slab (Doglioni 1993)

Active rifting was also proposed by Stegena (1972), Horváth (1974), Horváth et al. (1975) and Stegena et al. (1975) as an ultimate origin of the Pannonian Basin. However, their models already employed the concept of plate tectonics for the Pannonian region and, hence, reflect a completely different viewpoint. Observing the analogy with other young Mediterranean (Tyrrhenian and Aegean) basins, they classified the Pannonian Basin as an inter-arc, sialic basin. Basin formation and subsidence was controlled by the sub-crustal erosion of the underlying lithosphere which was caused by an active mantle diapir generated by the subduction of the European and Apulian (Adriatic) Plates beneath the Pannonian plate fragment (Fig. 6b). Moreover, the intense Neogene–Quaternary volcanic activity, the extremely high heat flow, the presence of an anomalous upper mantle and the greatly thinned crust in the Intra-Carpathian region are all closely related phenomena. An average subsidence of about 3 km was calculated for the Pannonian Basin, which was to a great extent the result of the combined effect of sub-crustal erosion and the loading effect of the Neogene basin fill.

These pioneering results served as an excellent starting point for further subsidence analysis. During the 80s this research led to the recognition of the depth-dependent, heterogeneous stretching of the lithosphere (Royden and Dövényi 1988) and the subdivision of the evolution of the Pannonian Basin into a synrift (Early to Middle Miocene) and postrift (Late Miocene), i.e. thermal phase (e.g. Royden et al. 1983; Horváth and Rumpler 1984). Further increase of available stratigraphic data made possible the refinement of this subdivision. According to Tari (1994), Horváth (1995) and Tari et al. (1999), a regional Middle Badenian unconformity indicates the termination of the synrift period, which is followed by a much quieter postrift phase. Moreover, as suggested by Horváth and Cloetingh (1996), the Pannonian Basin exhibits an anomalous subsidence pattern during Pliocene through Quaternary times that marks a new tectonic style. The onset of the synrift period is traditionally connected to the widespread occurrence of the lower rhyolite tuff horizon of early Ottnangian age (18.2 Ma – Árva-Sós et al. 1983), whereas the second, postrift phase sensu Tari (1994) started during the Middle Badenian (14.8 Ma). One may argue that from the viewpoint of kinematics it would be better to accept the traditional Sarmatian/Pannonian boundary for the end of the synrift phase (e.g. Royden et al. 1983). Other groups of dynamic models also suggest the key importance of subduction in the Pannonian region (e.g. Royden and Karner 1984; Csontos et al. 1992; Horváth 1993; Csontos 1995; Linzer 1996; Fodor et al. 1999). It is to be noted, however, that subduction below the Dinaric chain became gradually locked along the Dinaric chain and continental collision has been taking place there since Early Miocene (e.g. Dercourt et al. 1986; Wortel and Spakman, 2000). Thus, the Pannonian Basin had a back-arc location behind the Carpathian Arc (Horváth and Berckhemer 1982) rather then an inter-arc position (Stegena et al. 1975) (see Fig. 1). Also, extension and lithospheric stretching in these models are caused by the hinge retreat of the subducting European Plate along the Carpathians (Fig. 6c). Although no clear evidence has been found so far for the presence of oceanic lithosphere in the Carpathian embayment, this hinge retreat was most probably driven by the negative buoyancy of a subducting oceanic (?) plate (Royden and Karner 1984). An alternative explanation was given by Doglioni (1993) who pointed out that an eastward mantle flow may have been pushing the subducting slab and, hence, can be responsible for the same hinge retreat (Fig. 6d). Nevertheless, both models account for the passive rifting of the Pannonian Basin where tension is facilitated by trench suction forces exerted at the contact zone between the overriding and subducting plates.

Subduction was already active along the East Alpine–Carpathian front during Paleogene times. Thus, a fundamental problem must be solved: why and how did these trench suction forces overcome the collisional forces affecting the Pannonian Basin from the south? In other words: why did compression and intense convergence change to lithospheric extension and basin formation in the Intra-Carpathian area? Furthermore, due to the irregular shape of the European

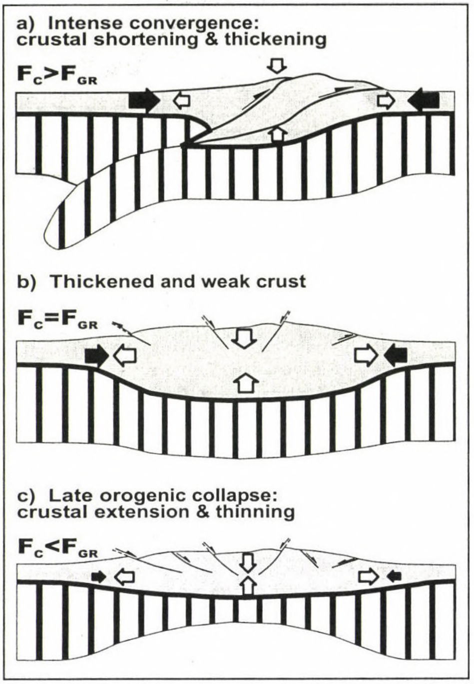

Fig. 7
(a) Intense convergence and shortening can result (b) in a thickened and rheologically weakened crust in zones of continental collision. (c) If, for some reason, gravitational forces (FGR) due to the increased potential energy of the thickened and elevated orogen exceed the magnitude of collisional forces (FC), late orogenic collapse of the mountain belt is initiated that can lead to crustal extension and thinning. White and black arrows indicate gravitational and tectonic (collisional) forces, respectively. Moho is depicted by heavy black line

plate margin, the termination of subduction shows a characteristic temporal pattern along the Carpathian Arc, becoming younger from the Eastern Alps toward the southeastern Carpathians. This is proved by the termination of major nappe emplacement in the outer Carpathian flysch belt (Jiříček 1979), and the Late Oligocene to Pliocene depocenter migration in the Carpathian foredeep (Meulenkamp et al. 1996).

More recently, Ratschbacher et al. (1991) offered a new explanation for the Miocene extension in the East Alpine–Pannonian region. It has been widely recognized that in regions of continental collision intense shortening and crustal thickening can eventually lead to the gravitational instability of the axial zone of an orogen (e.g. Tapponnier et al. 1986; England and Houseman 1988; Platt 1987; Molnar and Lyon-Caen 1988; Bird 1991) (Fig. 7). Moreover, the central belts of orogens are very often thermally weakened and, hence, are favorable areas of strain localization. If, under proper conditions, gravitational forces exceed compression exerted by the convergence between the colliding plates, the process of late orogenic collapse of the weakened crust can be initiated (Fig. 7c). In fact, orogen-parallel extension in the Eastern Alps has been very well documented (e.g. Selverstone 1988; Ratschbacher et al. 1989) and the idea of a weakened Alpine lithosphere was later confirmed (e.g. Cloetingh and Banda 1992; Genser et al. 1996; Okaya et al. 1996). This led to the formulation of the idea of Neogene lateral extrusion in the Eastern Alps toward an eastern, unconstrained margin, i.e. the Carpathian subduction belt. In a strict sense, lateral extrusion is a ductile flow of the lower crust confined between the brittle upper crust and mantle lithosphere (Ranalli 1995) that leads to the relaxation of the topography of both the surface and the Moho (Bird 1991). Ratschbacher et al. (1991) used this term in a wider sense and combined the processes of gravitational collapse and tectonic escape, assumed to be active simultaneously during the Early to Middle Miocene. In their model, a large amount of material is expelled eastward from between a rigid indenter (Southern Alps) and an even stiffer foreland buttress (Bohemian Massif). This process was coeval with a further shortening in the central zone of the Eastern Alps and the rapid uplift and exhumation of several Penninic units of lowermost structural position in the Alpine nappe pile. A similar reasoning led Tari (1994), Tari and Horváth (1995) and Tari et al. (1992, 1999) to the conclusion that the first (Karpatian) stage of deformation took place in the form of core complex extension (sensu Buck 1991). This resulted in the formation of several metamorphic core complexes in the Pannonian Basin during Early Miocene times. Moreover, Ratschbacher et al. (1991) also assumed that the material originating from the axial zone of Alpine convergence filled up the free space in the Carpathian embayment and, consequently, this process was eventually responsible for the formation of the Pannonian Basin. The model of lateral extrusion is especially popular among earth scientist dealing with the structural evolution of the Eastern Alps and adjacent areas (Linzer et al. 1997; Peresson and Decker 1997; Frisch et al. 1998). However, simple geometric considerations, i.e. the balance between shortening along the Carpathian orogen (Roure et al. 1993; Ellouz and Roca 1994) and extension inside the Pannonian Basin (Tari et al. 1995), suggest that the amount of the possible extrusion from the region of the Eastern Alps was in the order of only several tens of kilometers. Hence, it cannot explain large extension and crustal spreading in the Intra-Carpathian area. Furthermore, the surface morphology in the Pannonian region was relaxed within 1–2 Ma, still during the Early Miocene and, thus, gravitational collapse was not a viable mechanism for generating extension later in the Middle and Late Miocene. Therefore, it appears that lateral extrusion played a limited although important spatial and temporal role in the PANCARDI region.

Physical properties and rheology of the lithosphere in PANCARDI system

The topographic features of the Alpine–Carpathian–Pannonian–Dinaric system well reflect the main characteristics of the Late Cenozoic evolution of this northern segment of the Mediterranean puzzle. The flat and low-altitude Pannonian Basin, where major subsidence has been taking place, is surrounded by elevated fold-and-thrust belts where the general vergence of the nappe piles is pointing outward from the basin center. In the Dinarides, this corresponds to SW-directed thrusting, while all along the Carpathian Arc one can observe tectonic transportation being nearly perpendicular to the actual curvature of the mountain chain. This structural zonation is even traceable in the basement of the more internal areas; however, it is intensely overprinted by structural elements related to the formation and evolution of the Pannonian Basin during the last 18 million years (Horváth 1993; Csontos 1995; Fodor et al. 1999). There is no doubt that this period has definitely had the most profound influence on the present-day lithospheric architecture of the entire Intra-Carpathian area.

From a geomechanical point of view, topography also provides valuable clues for the physical properties of both the crust and the whole lithosphere. In general, there is a very good correlation between Moho depth (Fig. 8) and elevation in both the internal and external sectors of the PANCARDI region. Thin crust is an essential attribute of the subsided parts (Horváth 1993; Posgay et al. 1995). Here, thickness values in the range of 22.5–30 km indicate moderate to high amount of crustal stretching and thinning. Model calculations suggest that the ductile lower crust was attenuated rather uniformly, while the upper, brittle part of the crust is much more extended beneath the deepest sub-basins (Tari et al. 1999; Lenkey 1999). In the Transylvanian Basin a local minimum of 32.5–35 km indicates only slight crustal extension. On the other hand, although with strong lateral variation, areas in the Alps, Dinarides and Carpathians exhibit thick crust as a result of intense shortening during Alpine orogeny. The overall increase of Moho depth along strike of the Carpathian Arc from NW to SE suggests temporal differences in the main phases of deformation history (Tomek and Hall 1993). Yet the principal trends are similar; differences in the lithospheric thickness are much more dramatic than crustal thickness variation. Sixty km in the basin center vs. more than 100 km in the peripheral areas strongly supports the idea of non-uniform (i.e. depth-dependent) mode of lithospheric extension where the mantle lithosphere is significantly more attenuated than the crust (e.g. Royden et al. 1983; Lankreijer 1998; Lenkey 1999). Beyond spatial variation, it was also recognized (Royden and Dövényi 1988) that extension systematically increases toward the basin interiors. Below the mountain arc of the Alps a pronounced lithospheric root with more than 200 km depth is detectable as an expression of the relicts of the subducted European and Adriatic forelands (Laubscher 1990; Panza and Suhadolc 1990). In the southeastern bend zone of the Carpathians, similar high values of lithospheric thickness are inferred (Horváth 1993). However, this root is probably detached at the depth of about 50 km (Oncescu 1984; Spakman 1990).

The Late Cenozoic evolution of the area is reflected in the geothermal properties of the lithosphere. As a result of intense Miocene stretching, the Pannonian Basin exhibits anomalously high surface heat flow with values of about 90 mW/m^2 an average and 120–130 mW/m^2 as local maximum (Dövényi and Horváth 1988; Tari et al. 1999). Interestingly enough, the area of the Transylvanian Basin has a quite cool underlying lithosphere which, together with its close to normal lithospheric and crustal thickness, suggests a distinct structural evolution. The mountain belts around the Pannonian Basin show low surface heat flow with the exception of the Eastern Alps where the observed 80–100 mW/m^2 values are probably due to intense uplift and erosion.

All the mechanical parameters of the lithosphere in the PANCARDI region are notably reflected in its strength distribution. The casual relationship, however, is

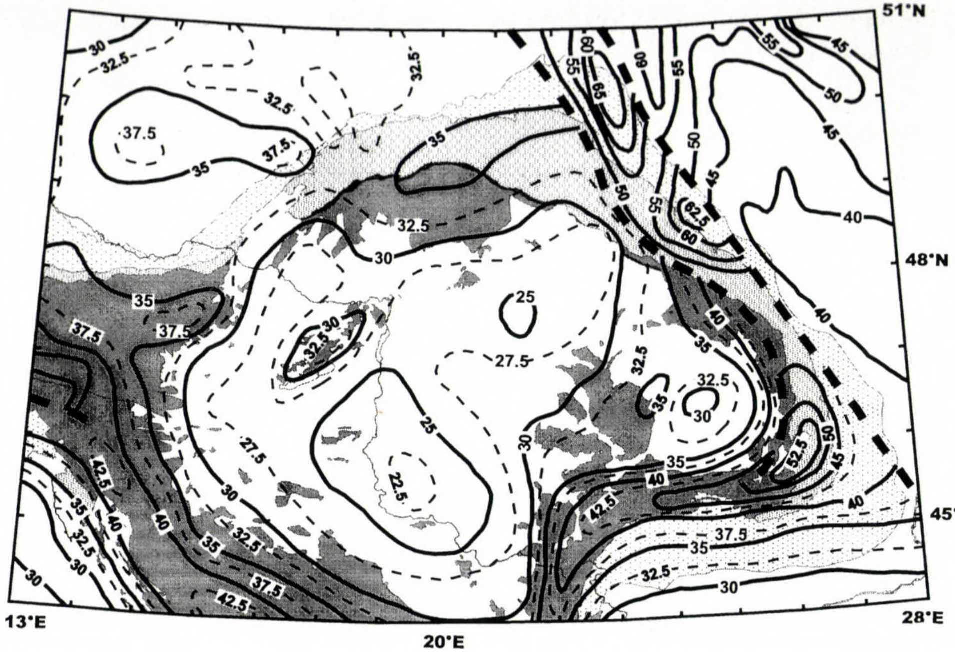

Fig. 8
Crustal thickness variation (values in km) in and around the Pannonian Basin. Both the crust and the lithosphere were stretched and thinned beneath the central sectors, while the values are considerably higher in the peripheral areas (after Horváth 1993). Dashed lines indicate sharp changes of Moho depth

not that obvious. As pointed out by Cloetingh et al. (1995) and Lankreijer (1998), the structural evolution of the Pannonian Basin system was very much influenced by its pre-rift rheology. The weakened lithosphere of the Alpine-Carpathian orogenic belt subjected to tensional stresses was an ideal medium to suffer large-scale deformation, i.e. the formation of the Pannonian Basin. On the other hand, this led to the further weakening of this thermally disturbed (i.e. heated) lithosphere and, hence, deformation was intensified in the already extended and warm regions of the Pannonian Basin. Within less than 20 Ma, this positive feedback mechanism resulted in remarkable differences between the Pannonian Basin and its surroundings in terms of the rheological strength of the lithosphere (Lankreijer et al. 1998). In the extremely weak Intra-Carpathian area the only strong part of the lithosphere is represented by the brittle upper crust, whereas in the Bohemian and Moesian foreland of the surrounding Carpathians the upper part of the mantle lithosphere also has a high level of strength. The presence of these two rigid buttresses had an important impact on the evolution of the study area throughout the entire Cenozoic. The coexistence of the weak and, thus, deformable structural domains in the Alpine-Carpathian orogenic belt and their relatively strong foreland on the southern edge of Eurasian plate has significantly influenced the style of deformation in the PANCARDI system.

Discussion and conclusions

Formation of the Pannonian basin: Interplay of gravitational and collisional forces

After a long period of convergence, crustal shortening and thickening in the PANCARDI region during the Cretaceous through Paleogene, the lithosphere beneath the Pannonian Basin was considerably extended and subsided within a time period of about 12 Ma, from the initiation of extension during Ottnangian times until the onset of general inversion of the Intra-Carpathian area at about the Miocene/Pliocene boundary. Results of numerical modeling (Bada 1999) suggest that the formation of the Pannonian Basin was initiated and to a great extent controlled, by orogen-parallel tension due to the presence of a thickened, elevated, thermally weakened and laterally unconstrained ALCAPA block. This finding raises a question: what sort of mechanism accounts for the commonly observed orogen-parallel tension in regions of active collision where the main trend of elevated topography is almost exclusively orthogonal to the direction of convergence? In other words, the fact that E–W tension (Fodor et al. 1999) is mostly perpendicular to the maximum topographic and Moho depth gradient requires some additional (i.e. non-gravitational) forces or boundary conditions. Furthermore, there is a theoretical necessity of an orogen-parallel gradient in paleoelevation (Frisch et al. 1998) and Moho depth (Ratschbacher et al. 1991). According to the quoted authors and employing simple physical considerations, one can indeed expect maximum topography and crustal thickness in the axial zone of Adria convergence, i.e. the central part of the Eastern Alps. Aside from this region toward the east, a thinner and less elevated crust can be assumed. However, it is to be noted that buoyancy forces alone are not capable of producing orogen-parallel tension, even in the case when this terrain was tilted to the east.

Therefore, as has been shown by modeling, orogen-parallel (i.e. E–W in paleo-co-ordinates) tension can only be explained upon the fulfillment of either of the following two conditions. First, the orogen-normal component of tension can be efficiently reduced by the collisional forces exerted by the pushing effect of the Adriatic Microplate as indenter. This might eventually lead to the formation of a strike-slip type of paleo-stress field where compression and tension are parallel and perpendicular to the general direction of convergence, respectively. Furthermore, this compression was less dominant at a greater distance from the axis of collision, which might explain the large extension in the Pannonian area (transtension) and further shortening in the Eastern Alps (transpression). The combined mechanism of buoyancy and collisional forces appears to be applicable for a period of about 1 Ma during the formation of the Pannonian Basin system

(Ottnangian). Then the entire Pannonian basin subsided below sea level very rapidly (i.e. by Karpatian times); thus most of the topographic relief was masked. Consequently, E–W tension had to be increased by the rollback process of the subducting Magura Ocean in the east. As a result, trench suction forces associated with the hinge retreat at the subduction zone became of dominant importance from Karpatian times.

Importance of subduction related processes along the Carpathian trench system

Numerical modeling of the paleostress and paleostrain field (Bada 1999) suggest the predominant role of trench suction forces along the Carpathian subduction system that induced tensional stresses in the Intra-Carpathian area from Karpatian times onward. Modeling results show that these trench suction forces are generally normal to the local curvature of the trench system. The physical explanation is that the overriding plate in a subduction zone tends to passively follow the retreating hinge of the descending lithosphere, which is marked by the strike of the trench zone. In other words, the upper plate is extruding in the direction of maximum potential energy difference, i.e. normal to the local trend of the trench axis. It is admitted, however, that the location and orientation of the trench zone along the Carpathian subduction belt can only be assessed by means of various kinematic indicators, the depocenter migration of the foreland basin system and other palinspastic considerations. Nevertheless, in the Pannonian region the retreat of the hinge zone of the descending plate created open space for the gravitational spreading of the overriding plate (e.g. Chemenda 1993). A set of these trench suction forces acting normal to the curvature of the Carpathian Arc, in combination with collisional forces exerted at the southern boundary of the area of interest, can indeed reproduce the observed paleostress pattern. Due to the finite strength of the lithosphere underlying the Pannonian Basin, tension was transmitted far behind the arc region and, as a consequence, nearly the entire Intra-Carpathian area was stretched to a large extent. Moreover, this scenario can also account for the gradual eastward translation of both the North and South Pannonian Units into the Carpathian embayment, delineated by the greatly irregular shape of the European foreland. The opposing rotation of these units was driven by the combined effect of the "pull-from-the-front" process along the Carpathian subduction belt and the "push-from-the-back" process along the Adria–Alps and Adria–Dinarides interface.

Similar to the key importance of subduction-related forces driving extension in the Pannonian Basin, the termination of subduction along the Alpine–Carpathian Belt had a great impact on the Miocene deformation history in the whole study area (e.g. Sperner 1996; Bada 1999; Fodor et al. 1999). As marked by a great variety of structural data and subsidence pattern in the Alpine foredeep, a slowdown of continental convergence occurred during the Middle Oligocene (25 Ma) in the

Central Alps and during the Early Miocene in the Eastern Alps (22–18 Ma). This temporal pattern is traceable further east along the Carpathian Arc: the final collision between the North Pannonian Unit and its foreland commenced at the Vienna basin during the Karpatian (17 Ma) and the NPU reached its present-day position at the end of the Middle Miocene (11.5 Ma). This process continued during the Late Miocene when the South Pannonian Unit became docked and fixed against the rigid, thick and buoyant European foreland. Paleostress data show that the direction of compression nearly exclusively pointed to the location of the docking segment of the Carpathian arc, although occasionally with a short delay (e.g. Tokaj-Zemplén Mts. – see Fodor et al. 1999). The combined effect of a northward Adria-push and a roughly perpendicular, eastward pull due to hinge retreat of the Carpathian arc can account for the observed clockwise rotation of the paleostress axes in either time or space (Bada 1999). On the other hand, along with the gradual vanishing of trench suction forces, the inactive arc segments became fixed and welded to their rigid foreland. Then compression became oriented normal to the structural trend of the arc due to the 'push from behind' effect of the continuously converging Adriatic indenter in the south-southwest.

On the possible importance of lateral migration of slab detachment along the Carpathian arc

Since subduction of the North Penninic and Magura Oceans along the Alpine and Carpathian arcs, respectively, was already active during Paleogene and Early Miocene times, one fundamental question should be answered concerning the origin of transition from an advancing to a retreating style of subduction that took place during the Early Miocene. According to Royden (1993), where the rate of overall plate convergence is greater than the rate of subduction (advancing subduction boundaries), the regional deformation of the overriding plate is manifested in horizontal shortening. This might be applicable for the Paleogene–Early Miocene evolution. When, for some reason, the rate of subduction exceeds the rate of convergence, the principal style of deformation becomes horizontal extension. This was a clear characteristic of the late Early through Late Miocene evolution of the Pannonian Basin. Sperner (1996) proposed a model why the style of subduction changed and the retreat of the Carpathian hinge zone accelerated. She applied the theoretical results of hydrodynamic modeling by Dvorkin et al. (1993) for the Alpine–Carpathian system and argued that rapid hinge retreat occurred when the along-strike length of the subducting slab was reduced to a critical value. In the case of a wide slab, the upward-directed hydrodynamic suction forces above the descending slab can balance the weight of the slab, i.e. the slab pull and, hence, trench suction forces are compensated. Upon the decrease of the along-strike width of the slab, a sideways asthenospheric flow into the slab/overriding corner can eliminate this suction effect. This can lead to the rapid sinking of the subducting slab and, consequently, the initiation and/or acceleration of hinge retreat. Consequently, the overriding (Pannonian) plate stretches and extends (rifting) and intense subsidence takes place.

It is also assumed (Sperner 1996) that the effective reduction of the slab width was facilitated by the process of slab detachment which is, at least according to Wortel and Spakman (2000), a typical attribute of the subduction systems in the Mediterranean region (see Fig. 1). When a non-subductable (i.e. continental) lithospheric fragment is entering into the subduction zone, hinge retreat terminates and the slab breaks down under its own weight. Due to the fact that convergence between the Adriatic indenter and the European Plate was not orthogonal and, furthermore, as a result of the highly irregular shape of the foreland, the slab break-off was not coeval along the whole arc. Instead, a lateral migration of slab detachment was assumed, which was initiated during the Paleogene in the Alps and then reached the southern spur of the Bohemian Massif by Ottnangian–Karpatian (18–17 Ma) times, coeval with the onset of extension in the Pannonian area (Fig. 9a). This zipper-like process appears to become accelerated when the hinge zone passed the Bohemian Massif that formed a gate toward the free space in the Carpathian embayment. By Middle Badenian times (14.8 Ma) the central part of the Western Carpathians became the area of possible slab break-off (Fig. 9b). The North Pannonian Unit reached its present-day position by the end of the Middle Miocene, which is also marked by the final thrusting in the northeastern flysch Carpathians at that time. The docking effect of NPU might partly explain the transient inversion in the northern and southern parts of the Pannonian basin at about the Sarmatian/Pannonian boundary. The lateral migration of slab detachment is then assumed to have reached the area in front of the Tisza–Dacides block (Fig. 9c). It was finally terminated below the Vrancea Zone as late as Pliocene–Quaternary times, as indicated by a characteristic gap between the shallow and deep level of seismicity in this region. The lateral variation of the rheological properties in the foreland area, however, strongly influenced the process of trench retreat that had a first-order effect on the structural evolution of the Carpathian fold-and-thrust belt. Moreover, it is also suggested that slab break-off was probably not a single and continuous process but, rather that the slab was probably further segmented into smaller columns along vertical fracture zones (Fig. 9c).

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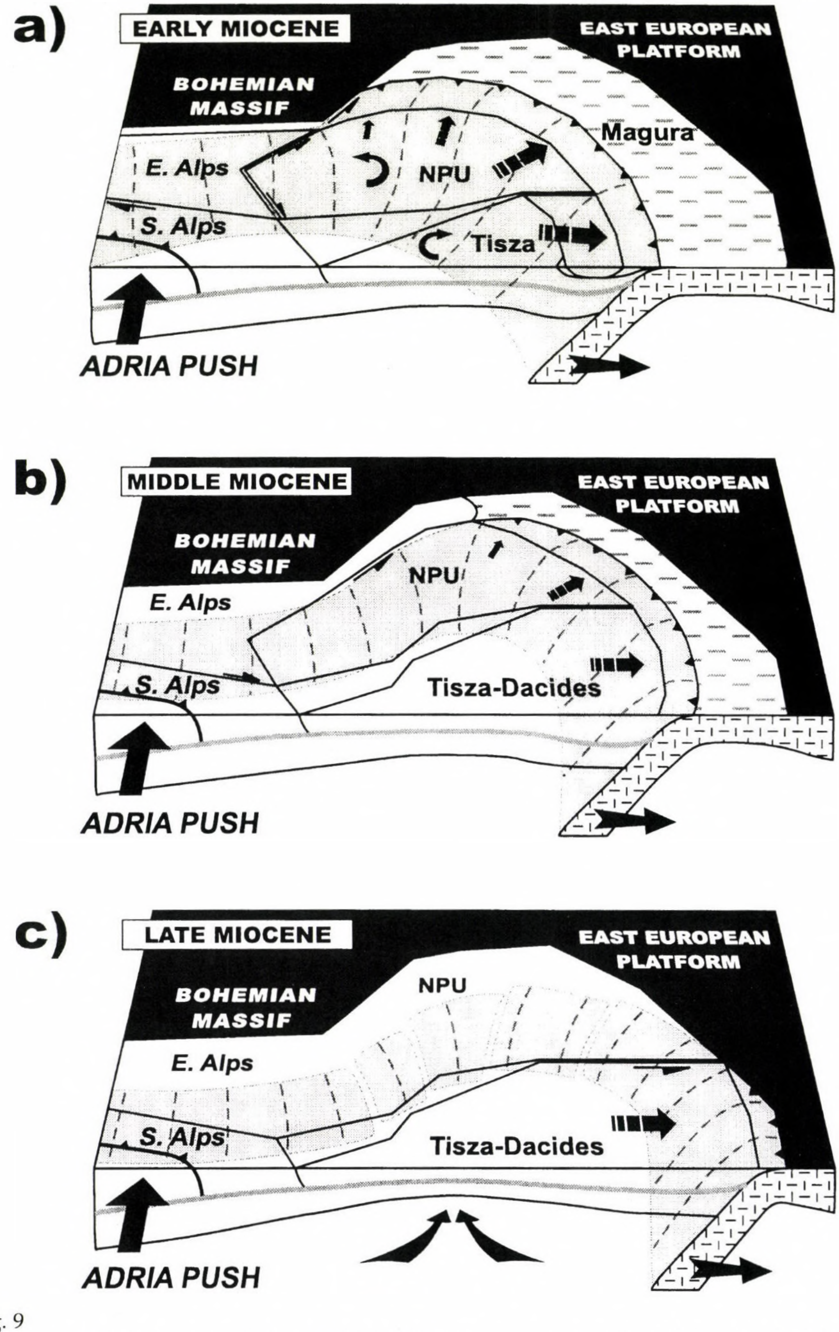

Fig. 9
Cartoon showing the main features of the Miocene tectonic evolution of the PANCARDI region in three steps. The figure is meant to depict the genetic relationship between gradual termination of subduction along the Carpathian Arc, eastward hinge retreat and associated translation of the Intra-Carpathian units. Basin formation and frontal accretion was coeval both in the North Pannonian Unit (NPU) and the Tisza–Dacia block. (a) Late Early Miocene (17–16 Ma); (b) Middle Miocene (15–11 Ma) and (c) Late Miocene (11–9 Ma) configuration. For a Mediterranean outlook, compare with Fig. 1

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Alpine regional metamorphism in the main tectonic units of Hungary: a review

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New data on the spatial and temporal distributions and physical (P/T) conditions of Alpine regional metamorphic events that were active in the main tectonic units of Hungary are summarized. These data were obtained from mineral paragenetic and thermobarometric investigations carried out on polymetamorphic formations and complex mineral facies, phyllosilicate crystallinity and coal rank studies of incipient metamorphic rocks, integrated with isotope geochronological data. Subducted parts of the Neotethyan oceanic realm, such as the Mesozoic of the Kőszeg–Rechnitz (Penninic) Window and the South Gemeric Meliata Unit, exhibit high-pressure blueschist facies metamorphism of various ages, overprinted locally by a younger medium-pressure event. In the Lower Austroalpine Unit (Sopron Mts) and in the Vepor Unit the eo-Alpine medium-high pressure metamorphism even reached the amphibolite facies. The composite Pelso Unit shows no or only slight signs of eo-Alpine metamorphism. In the Bükkium and in the Igal Zone low to medium-pressure type, anchizonal and epizonal regional recrystallization was proved. In the vicinity of the Alpine overthrusts the Mesozoic formations of the Tisza Unit were locally metamorphosed under anchizonal and epizonal conditions in medium to low-pressure systems. Preliminary results suggest that Cretaceous metamorphism might have caused amphibolite facies recrystallization in certain parts of the pre-Alpine polymetamorphic basement in the Tisza Unit.

Key words: Alpine metamorphism, low-grade metamorphism, polymetamorphism, Hungary, Carpathian Basin

Introduction

Striking geomorphological features including the sharp termination of the Alps in the East and the bifurcation of the orogenic belt (Carpathians and Dinarides) already led to the elaboration of the "median mass" or "Tisia" theory of the Carpathian Basin at the beginning of the 20th century. Naturally the knowledge of geologic formations, structural conditions, magmatic and tectono-metamorphic evolution of the basement has increased exponentially during the last decades. Thus, the early concepts on the "stable" or "refractory" nature of the basement during Alpine orogeny have radically changed (for recent reviews see Fülöp 1989; Haas 1996; Kovács et al. 1996–1997; Árkai 1999 and Kovács et al. 2000). Although the decisive role of Alpine tectonics has been acknowledged by most of the researchers, regional metamorphic processes connected to the Alpine tectonocycle were considered sporadic and of low intensity, even in the first metamorphic map of the Carpatho-Balkan-Dinaride area published under the

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auspices of the Carpatho-Balkan Geological Association 25 years ago (Szádeczky-Kardoss et al. 1976). Integrated metamorphic petrogenetic and geochronological researches carried out in this region since the first regional synthesis have basically changed our view on Alpine metamorphism. The present paper provides a brief review of the pressure–temperature–time (P–T–t) relationships of Alpine regional metamorphic events, which affected the geologic formations found in various tectonic units of Hungary, and occupying a central geographic position in the Carpathian Basin. Concerning the isotope geochronological results and models of late cooling (exhumation), the author refers to the review of Balogh and Pécskay (this volume).

Methods

Due to a relatively small uplift and moderate erosion, most of the Alpine metamorphic rocks in the Carpathian Basin belong to the incipient (= very low-grade = subgreenschist facies) realm of metamorphism. In this realm, the temperature of which ranges between c. 150 and 450 °C, and the pressure of which varies up to several (5–7) kbars, the P–T-indicative minerals and mineral assemblages are restricted to narrow ranges of bulk rock and fluid chemistries. Therefore, clay mineralogical (illite and chlorite crystallinity), coal petrographic (vitrinite reflectance) and other specific methods have been applied in addition to classic mineral paragenetic (mineral facies) studies, in order to obtain a reliable database on metamorphism. The methods generally applied are given in Árkai (1983, 1991a) and Árkai and Kovács (1986). The newest results of methodological developments are summarized by Árkai et al. 1995a, b; 1996; 1997; 2000b and Árkai and Sadek Ghabrial 1997).

Geotectonic framework

According to recent plate tectonic reconstructions the basement of the Hungarian part of the Carpathian Basin is built up by crustal blocks (also called micro-plates or terranes) that originated in various parts of the Neotethyan oceanic branch and its southern (African) and northern ("stable" European) border areas (Fig. 1). These blocks have come to occupy their present tectonic position by large-scale, mostly horizontal displacements that culminated in the Tertiary and ended c. in the Middle Miocene (Géczy 1973; Kovács 1982; Kázmér and Kovács 1985; Csontos et al. 1992; for a recent review see Kovács et al. 2000).

Figure 2 displays the schematic sketch of the main tectonic units of Hungary. Different authors use various names even today. For simplicity, the most common terms are applied here. The identification of the various units can also be easily made by their bounding tectonic lines, also given in Fig. 2. Both the northern (ALCAPA = Alpine–Carpathian–Pannonian) and the southern (Tisza) units are composite terranes consisting of various units with different metamorphic histories.

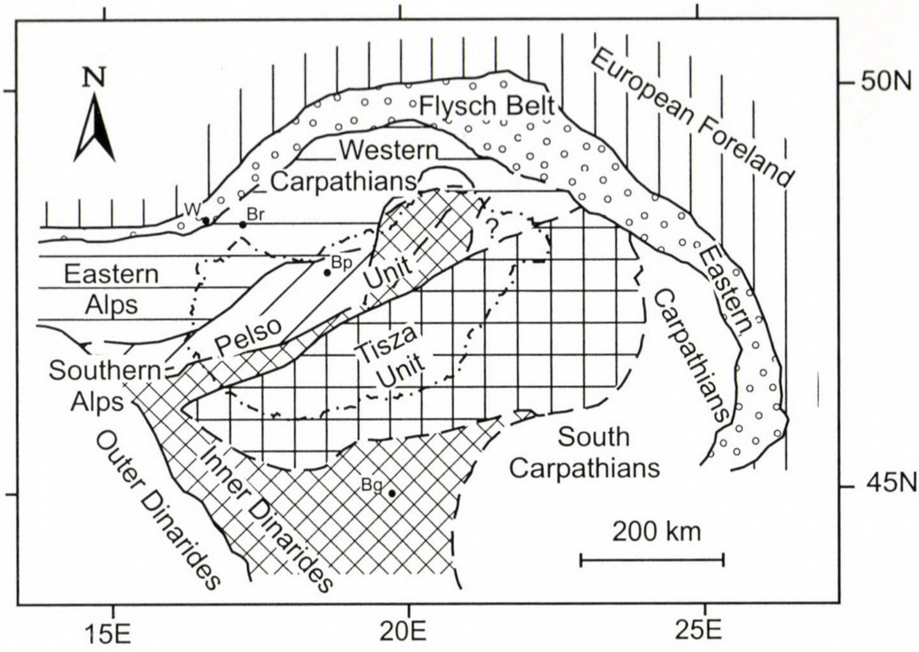

Fig. 1
The Alpine–Carpathian–Dinaric tectonic framework. Dashed–dotted line indicates the state boundary of Hungary. Abbreviations: Bg – Beograd; Bp – Budapest; Br – Bratislava; W – Vienna

Characterization of Alpine metamorphism in the main tectonic units of Hungary

The effects of Alpine regional metamorphism are manifold (Fig. 3). In part, the pre-Alpine (Variscan, presumably also pre-Variscan) formations suffered mostly retrograde polymetamorphic alterations, while some of the post-Variscan formations show clear evidence of prograde monocyclic (mono- or polyphase) metamorphism. Concerning the pre-Alpine metamorphic histories, the author refers to the summaries of Árkai (1991b), Lelkes-Felvári et al. (1996) and Kovács et al. (2000). The numbers in the following part correspond to those displayed in Fig. 2 indicating the localities described.

1. Rechnitz Window Group (Penninic Unit). The Rechnitz Window Group belongs to the Penninic Unit, forming its easternmost outcrop. According to Lelkes-Felvári (1982) a low-T/high-P blueschist facies Alpine event was overprinted by medium-pressure greenschist facies metamorphism in the Mesozoic magmatic and sedimentary formations of the Vashegy–Kőszeg Mountains. The exact age of the Alpine blueschist facies event is unknown. The K-Ar dates obtained on muscovite from phyllites by Balogh (1983, unpubl. internal report) scatter between 28 and 31 Ma, indicating the age of the meso-Alpine medium-pressure

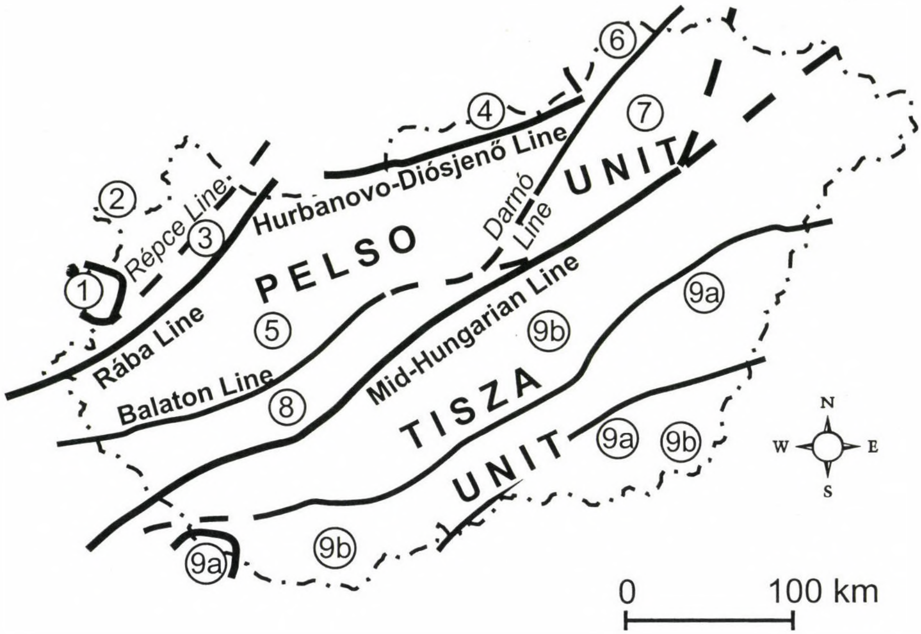

Fig. 2
The main tectonic units and their boundaries in Hungary after Fülöp (1989), strongly simplified. Numbers refer to those of the localities/units characterized in the text

greenschist facies event. Demény (1987–1988), using zircon morphology and tourmaline mineral chemistry, determined alkaline granitic, plagiogranitic, monzogranitic and granodioritic source rocks for the Ca-poor metapelites. Stable carbon isotope ratios of the graphite indicate mixing of a normal marine component with a detrital carbonaceous component and show regional trends within the Penninic Unit (Demény and Kreulen 1993). Although the exhumation mechanism and history of the Kőszeg–Rechnitz Windows are well constrained (Dunkl 1992; Dunkl and Demény 1997; Dunkl et al. 1998; Kuhleman et al. 2001), quantitative thermobarometry of the metamorphic events and age relations of the high-P (oldest) event are still lacking, at least in the Hungarian part of the formations in question.

Austroalpine (East-Alpine) units

2. *Sopron Mountains.* The polymetamorphic complex of the Sopron Mountains belongs to the Lower Austroalpine Nappe system. Its pre-Alpine metamorphic history is characterized by an older (pre-Variscan or early Variscan amphibolite facies event producing a kyanite + staurolite + sillimanite assemblage at <40 °Ckm^{-1} geothermal conditions. This event was overprinted by a Variscan,

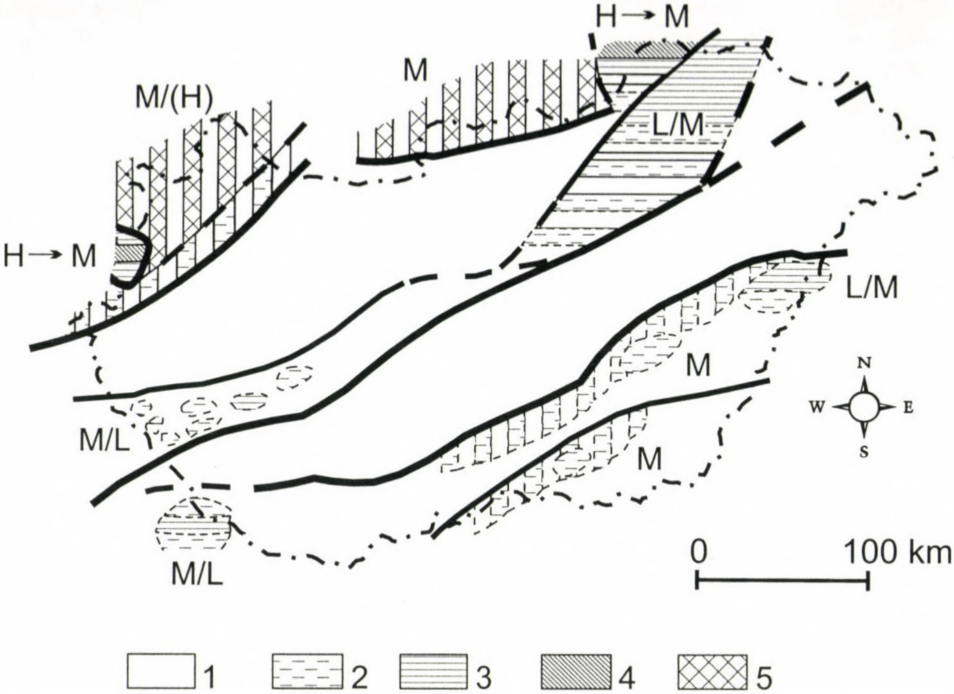

Fig. 3
Alpine regional metamorphism in the main tectonic units of Hungary after Árkai (1991b), strongly modified. Vertical strips indicate Alpine overprint on pre-Alpine metamorphic complexes. Legend: 1. not affected by Alpine regional metamorphism; 2. subgreenschist facies (~anchizone); 3. greenschist facies (~epizone); 4. blueschist facies; 5. amphibolite facies. L - low; M - medium and H - high pressure range

high (>40 °Ckm^{-1}) thermal gradient (andalusite + sillimanite) amphibolite facies event (Lelkes-Felvári et al. 1984). According to these authors, the Alpine overprint that effected the amphibolite facies Variscan complex was essentially of retrograde character, producing a new kyanite generation after andalusite, and displaying alterations of staurolite into chloritoid and white K-mica, plagioclase into clinozoisite, garnet and biotite into chlorite. Judging from the extensive rejuvenation of biotite and partially rejuvenated Variscan muscovite as compiled from the available Rb-Sr and K-Ar data, Lelkes-Felvári et al. (1996) concluded that the temperature of the Barrovian-type Alpine overprint might have been between c. 300 and 500 °C. (A similar conclusion was drawn on the Alpine overprint that affected the Fertőrákos Complex, located NE of the Sopron Mountains.)

Török (1996) reported Alpine high-pressure/low-temperature, blueschist facies metamorphism from a garnet-bearing orthogneiss occurrence of the Sopron

Mountains with peak conditions of c. 12 kbar at 450-500 °C. Involving various occurrences of the orthogneisses, Török (1998) characterized this eo-Alpine high-pressure metamorphism with a clockwise P-T path that culminated at 13–14 kbar and later, at 520–600 °C. Thus – according to Török (1998) – the gneisses from the Sopron Mountains were metamorphosed at higher pressures and temperatures than the other occurrences of the Grobgneiss series, and consequently might have been involved in the Cretaceous subduction. Demény et al. (1997) explained the formation of Alpine Mg-chlorite – muscovite–quartz phyllites (leucophyllites) found in shear zones by Mg-metasomatism at P/T conditions of c. 13 kbar/560 °C. On the basis of stable isotope data, dehydration of hydrothermally altered oceanic crust, i.e., the underlying Penninic Unit, seems to be a plausible candidate for the Mg-rich metasomatic fluids. Nagy and Árkai (1999) gave a detailed description of various generations of monazite found in various metamorphic formations of the Sopron Mountains. Part of these monazite generations might form during Alpine (Cretaceous?) metamorphic overprint. From borehole materials of an isolated area of the Sopron Mountains (Fertőrákos area) Frank et al. (1996) described Alpine overprint in the two-mica gneiss – amphibolite sequence characterized by fluid-driven static alteration of plagioclase, formation of fine-grained garnet and Ti loss from biotite. Biotites and partly also white micas that formed during the early Alpine overprint yielded $^{39}Ar/^{40}Ar$ ages mainly in the range of 170–220 Ma. These plateau-like patterns were interpreted by the authors as strong evidence for surprisingly homogenous ^{40}Ar overpressure during the mica formation. Only very few Cretaceous ages were obtained on biotite and muscovite. However, the 170–220 Ma range has been interpreted as a cooling interval after the Permo-Triassic crustal extension and related high-T/low-P metamorphic imprint documented recently in wide areas in the southern Austroalpine and South Alpine units (Schuster et al. 1999a, b).

3. *Upper Austroalpine Unit (basement of the Little Plain between the Rába and Répce Lines).* The basement rocks covered by a thick Neogene sedimentary pile were correlated with the Graz Paleozoic of the Eastern Alps by Balázs (1971). Metapelites, metasiltstone, carbonate schist, metasandstone and marbles intercalated with meta-volcanoclastics of basic to intermediate compositions made up the basement (Balázs 1971; Árkai et al. 1987). Using illite crystallinity, white K-mica geobarometric and coal rank methods, low-pressure epizonal (greenschist facies chlorite zone) regional metamorphism was proved (Árkai et al. 1987). K-Ar dates obtained on the white K-mica-rich, <2 μm fraction samples are typical mixed ages, scattering between 106 and 203 Ma. These data have been interpreted by Árkai and Balogh (1989) as Variscan ages partially rejuvenated by Alpine (Cretaceous?) metamorphic effects. Neither metamorphic petrographic nor isotope geochronological data indicating the eventual eastern, north-eastern continuation of the Penninic Window in the basement of the Little Plain supposed by Balla (1993) have been documented so far.

4.The southern continuation of the Vepor Unit (Inner Western Carpathians) Mica-schist, gneiss, amphibolite, phyllite and greenschist have been encountered in boreholes located north of the Hurbanovo-Diósjenő Line for a long time (see, e.g. Ravasz-Baranyai and Viczián 1976). They were interpreted as parts of the Veporic Unit (Szádeczky-Kardoss et al. 1976; Fülöp 1990; Bezák 1994). On the basis of detailed petrographic studies carried out by P. Kisházi and J. Ivancsics, Fülöp (1990) summarized the geologic and petrographic characteristics of these formations. Out of the metamorphic features described, only the locally occurring low-temperature (greenschist) facies retrogression was considered to be Alpine, supported by sporadic K-Ar ages of 96 to 116 Ma from biotites and muscovites of paragneiss and micaschist (Balogh 1984, in Lelkes-Felvári et al. 1996).

Koroknai et al. (2001a) – proving also the traces of a pre-Alpine (eventual Variscan) amphibolite facies metamorphism of these rocks – provided a complex characterization of the Alpine tectonometamorphic history of the Hungarian part of the Veporic basement. K-Ar and Ar-Ar data on biotite, muscovite and amphibole scatter between 87 and 95 Ma. On the basis of mineral facies data and thermobarometric calculations, the peak conditions of this eo-Alpine event were 550±°C and 9±1 kbar. Zircon fission track ages indicate very rapid cooling (uplift) during the late Cretaceous, most probably due to the extensional unroofing of the Veporic core complex. This evolution path outlined by the authors is very similar to that described by Plasienka et al. (1997) and Janák et al. (2001a) for the main (Slovak) part of the Veporic Unit, and resembles many Austroalpine basement complexes in the Eastern Alps (Janák et al. 2001b).

The various parts of the Pelso Unit (composite terrane)

5. The Transdanubian Central Range. Both the Paleozoic and Mesozoic formations of this unit escaped Alpine regional metamorphism. No petrographic and geochronological data are available on Alpine metamorphism in this region so far (for clay mineralogy of the diagenetic alterations see Viczián 1975; Árkai 1991b; Árkai 2000).

6. The South-Gemeric (Aggtelek-Rudabánya) Unit. Between the Rožnava Line and the Tertiary Darnó Line the Late Permian–Triassic and Jurassic formations are stacked in mostly south-vergent nappes (Silica, Meliata and Torna Nappes), most of which also represent distinct lithofacial, paleogeographic and -tectonic environments. The Meliata Unit consists of ophiolitic magmatic rocks associated with deep-water sediments (marly, pelitic slate, metachert, etc.).

According to Árkai and Kovács (1986) the grade of regional alteration of these formations in the Hungarian part of the South-Gemeric Unit (i.e. in the Aggtelek–Rudabánya Mountains) varies from late or deep diagenesis through the anchizone up to the epizone (greenschist facies chlorite zone). In the upper Silica Nappe mostly diagenetic alterations can be observed. Higher temperature (anchizonal) conditions were detected only in certain tectonic zones. Low anchizonal conditions were recorded by Árkai and Kovács (1986) at the type locality of the Meliata Unit (village of Meliata) in Slovakia, while c. 400 °C/7 kbar conditions were estimated on the basis of amphibole chemistry and mineral assemblages for the glaucophane schist of the Rožnava Zone. In northern Hungary the lowermost unit, the Torna Nappe, was metamorphosed under anchizonal to epizonal temperature, and transitional medium to high-pressure conditions. The age of the metamorphism was put between the Late Jurassic "Eohellenic" and the Cretaceous Austrian phases.

As summarized by Faryad (1997) on the basis of his earlier work (1988, 1991, 1995a, b), the blueschist of the Meliata Unit originated from mostly MORB-type basalt, limestone, pelites, psammites and reworked amphibolite facies-metamorphic rocks. Thermobarometric calculations show metamorphic peak conditions of 350–460 °C and 10–12 kbar. Faryad and Henjes-Kunst (1997) dated this high-pressure metamorphic event by K-Ar and ^{39}Ar-^{40}Ar methods applied on white K-micas. They obtained Middle Jurassic ages: a somewhat older one for the continental wedge (172 Ma) and somewhat younger ones for the oceanic crust (152–155 Ma).

Dismembered and metamorphosed parts of the Meliata oceanic crust are also found in northern Hungary, in the Aggtelek–Rudabánya Mountains. The metamafic rocks of the Bódva Valley Ophiolite Complex, from which only Middle Triassic stratigraphic ages are known, do not form a single tectonic unit, as they are imbricated with the unmetamorphosed Permian–Mesozoic sedimentary formations of the Silica Nappe, displaying a melange-like appearance of the metamafic and meta-ultramafic blocks, and indicating that the regional metamorphisms of the ophiolitic rocks preceded the formation of melange. Using mineral paragenetic and microstructural observations, mineral chemical analyses and thermobarometric calculations, the P-T-relative time paths in the various blocks were reconstructed by Horváth (1997, 2000) and Horváth and Árkai (2001). The first recognizable blueschist facies event was documented in the metagabbros (350–500 °C/7–8 kbar). This event was followed by a quasi-isothermal decompression to 4–6 kbar under greenschist facies conditions. Increase in temperature up to c. 500–600 °C (albite-epidote amphibolite facies) under isobaric conditions is characteristic of the third event. In the metabasalt an early greenschist facies event was followed by a blueschist facies recrystallization. This path resembles the polyphase evolution of glaucophane-free phyllites and phengite quartzite (group 3) described by Faryad and Henjes-Kunst (1997). It seems probable that the various blocks of the Bódva Valley Ophiolite Complex originated from rather distinct, different parts of the subducted oceanic crust – continental wedge complex and preserved various stages (paths) from their polyphase metamorphic evolution.

7. The Bükk Unit. The Bükkium or Bükk Unit consists of three main units, namely the Bükk, Uppony and Szendrő Mountains. Its Paleozoic and Mesozoic formations show clear lithological, biostratigraphic and paleogeographical affinities to the northwestern part of the Inner Dinarides. For details of regional geology and tectonics the author refers to Kovács (1989), as summarized also by Fülöp (1994), and to Csontos (1988). Metamorphic petrological characterization and interpretation of the Bükkium were given by Árkai (1973, 1977, 1979, 1983), Árkai et al. (1981), while the chronology of metamorphism was outlined by Árkai et al. (1995a). Using indirect (tectonic, stratigraphic and metamorphic petrological) evidence only one (Cretaceous, pre-Senonian) metamorphic event could be proved. This affected both the Paleozoic and Mesozoic formations of the Bükkium (Árkai 1979, 1983). On the basis of the results of K-Ar dating on <2 μm grain-size fraction illite-muscovites this orogenic metamorphism culminated between the Eohellenic phase (c. 160–120 Ma), connected to the subduction in the Dinarides, and the Austrian phase (c. 100–95 Ma), characterized by compressional crustal thickening. Transitional low-intermediate pressure epizonal (mainly greenschist facies chlorite; locally, maximum biotite zone) conditions prevailed in the Lower Devonian to Middle Carboniferous Szendrő Paleozoic and low-pressure type transitional anchizonal-epizonal conditions in the Uppony Paleozoic (with protolith ages ranging from Upper Ordovician? to Middle Carboniferous), while the Middle Carboniferous to Jurassic formations of the Bükk Mountains suffered predominantly anchizonal regional metamorphism, characterized by slightly varying low to medium-pressure characters. The metamorphism in the so-called Fennsík Parautochthon in the eastern part of the Bükk Mountains reached the greenschist and pumpellyite-actinolite facies and resulted in mainly higher anchizonal metasedimentary rocks. In contrast, the Szarvaskő-Mónosbél Nappe of the southwestern Bükk Mountains suffered only late diagenetic and lower anchizonal orogenic alterations. In the Jurassic incomplete, dismembered ophiolite sequence a high thermal gradient ocean-floor hydrothermal metamorphism, exceeding the amphibolite facies (550–600 °C), preceded the orogenic very low-grade metamorphism (Sadek Ghabrial et al. 1996; Árkai and Sadek Ghabrial 1997), providing an example of Alpine plurifacial metamorphism. Recently, Koroknai et al. (2000, 2001b) described chloritoid schist from the Paleozoic formations of the Uppony and Szendrő Mountains. Confirming the earlier statement of Árkai et al. (1981) on the metamorphic origin of chloritoid in the Uppony Paleozoic, they provided detailed microstructural, mineral structural and chemical characterizations of the chloritoid, formed after the major Alpine (Cretaceous?) folding of the Lower Paleozoic sequences.

8. Mid-Transdanubian (Igal) Unit. This unit is a complicated tectonic zone that connects the Inner Dinarides with the Bükk Unit. The basement of the Central Hungarian Unit is built up by predominantly Mesozoic (mostly Triassic, partly Jurassic), subordinately Late Paleozoic formations. Recently, the geodynamic role of this unit was interpreted in a complex way by Haas et al. (2000). The only work on diagenesis and regional metamorphism of the basement was published by

Árkai et al. (1991). In accordance with the strongly disturbed tectonic conditions, the grade of regional alteration is rather variable. The Transdanubian Central Range and Dinaric-type Mesozoic formations suffered diagenetic, anchizonal and epizonal regional alterations in transitional low-medium pressure conditions. On the basis of K-Ar dating of newly formed illite-muscovite, the age of regional metamorphism is Cretaceous. Locally, polyphase retrograde metamorphic phenomena connected to cataclastic deformation, contact metamorphism and low-T hydrothermal alterations are the signs of younger (meso-Alpine) tectogenesis.

9. The Tisza Unit (Tisia Composite Terrane)

9a. Prograde Alpine metamorphism in the post-Variscan complexes of the pre-Neogene basement. In the Barcs-West area (Drava Basin, SW-Transdanubia) hydrocarbon exploration boreholes prove Inner Dinaridic-type Mesozoic formations in overthrusted tectonic position on the pre-Alpine polymetamorphic complex. The Mesozoic formations consist of metarhyolite and -tuffs, metasandstone, phyllite, marble and cipollino-like lithologies found in regular sequence in the different boreholes. Similar rock types are also found in the Croatian part of the Tisza Unit (in the Drava Basin) as well as in the Mid-Transdanubian (Igal) Unit that represents a tectonic link between the Inner Dinaridic Zone and its detached northwestern part, i.e. the Bükk and South Gemeric Units. The Mesozoic sequence in the Barcs-West area suffered mainly anchizonal, partly epizonal, mostly medium-pressure Cretaceous regional metamorphism, which (according to the K-Ar dates obtained on the <2 μm grain-size fraction illite-muscovite) was overprinted by a meso-Alpine (c. 30 Ma) thermal effect (Árkai 1990; Balogh et al. 1990).

In the pre-Neogene basement of the Great Plain (eastern Hungary) some pre-Alpine polymetamorphic complexes were thrusted over a low-temperature, prograde metamorphic sequence (borehole Sáránd-I) along the southern border of the so-called Mecsek–Northern Great Plain Zone. Here the Mesozoic "parautochthon" is built up by an upper (dolomitic – fine siliciclastic), a middle (calcareous – fine siliciclastic), and a lower (calcareous – basic to intermediate volcanoclastic) sub-unit. The upper part suffered anchizonal, the middle and lower parts epizonal Cretaceous regional metamorphism that proved to be older than the overthrusting (Árkai et al. 1998). New data gained from a systematic study show, prograde Cretaceous metamorphism was not an isolated phenomenon in the post-Variscan basement of the Great Plain (Árkai et al. 2000a). In contrast, it affected considerable parts of the Late Paleozoic and mainly Mesozoic formations located beneath the overthrusted parts of the poly-metamorphic Variscan – pre-Variscan (?) complexes or which are in imbricated, allochthonous tectonic positions along the main Alpine thrust zones. The climax of the orogenic prograde metamorphism reached the epizone (greenschist facies chlorite zone) in the easternmost part (Săránd area). In addition, anchizonal rocks were also described from various parts of the basement. The age of the c. 200–350 °C prograde, medium – transitional low-medium pressure-type metamorphism is Cretaceous.

9b. Alpine regional metamorphism in the mostly Variscan polymetamorphic basement complexes in the Tisza Unit. Before the metamorphic studies outlined in point 9a, the Tisza Unit had been traditionally considered a very stable, refractory crustal block or microplate that – disregarding locally occurring, mostly Cretaceous retrograde overprints in the pre-Alpine polymetamorphic basement – practically escaped Alpine prograde metamorphism (for a historical summary of this view that was rooted in the "median mass" concept see Árkai 1999). Although the presence of low-temperature Alpine (Cretaceous) retrogression has been generally accepted (see, e.g. Szádeczky-Kardoss et al. 1976; Lelkes-Felvári and Sassi 1981; Árkai 1991b; Lelkes-Felvári et al. 1996) no systematic research has been carried out so far in order to constrain the spatial, temporal and physical conditions of Alpine retrogression. In unpublished research reports prepared for the Hungarian Oil and Gas Plc. (MOL), Árkai (1995, 1996) stated that simultaneously with the Cretaceous prograde effect found in the Mesozoic, the related parts of the polymetamorphic complex experienced greenschist and subgreenschist facies overprints. This conclusion was based on the unpublished K-Ar data obtained on the fine fractions of retrogressed gneiss and micaschist by K. Balogh.

The available K-Ar and Ar-Ar data from the polymetamorphic sequences of the Tisza Unit were summarized by Balogh and Pécskay (this volume). It seems to be very plausible that certain areas (e.g. southern Transdanubia: Mecsek-Mórágy Mountains, eastern part of the Drava-Somogy Basin) escaped the Alpine metamorphic overprint. In other regions (see point 9a) Cretaceous metamorphism produced intense retrogression. However, further research is needed to differentiate between Alpine and pre-Alpine retrograde events at a regional scale. Sporadic Ar-Ar data, especially from the southern part of the basement (Ferencszállás, Újszentiván, Algyő, near the town of Szeged; see Balogh and Pécskay, this volume; Lelkes-Felvári et al., in preparation) together with the mineral paragenetic and thermobarometric calculations suggest that the grade of Cretaceous medium-pressure regional metamorphism might even here reach the amphibolite facies (Horváth and Árkai, in preparation). Thus, the Tisza Unit formed an integral part of the Alpine-Mediterranean orogenic belt, also from the point of view of regional metamorphism.

Conclusions

Summarizing the metamorphic petrological results achieved mostly during the last two decades, the following conclusions can be drawn. Alpine (commonly, eo-Alpine) regional metamorphism played important roles in the tectono-metamorphic evolution histories of the Mesozoic and older formations in most of the main tectonic units of Hungary.

Considering the Alpine units, the Penninic Kőszeg–Rechnitz Mesozoic suffered high-pressure blueschist facies metamorphism that was overprinted by a meso-Alpine medium-pressure greenschist facies event. In the Hungarian parts of the Lower Austroalpine Unit and the Inner Western Carpathian Vepor Unit the pre-Alpine medium-grade metamorphic complexes suffered Cretaceous medium- to high-pressure metamorphism that culminated in the amphibolite facies. The Upper Austroalpine Paleozoic formations (basement of the Little Plain, NW of the Rába Line) prove only slight traces of weak Alpine (Cretaceous) thermal overprint.

The signatures of Alpine metamorphism are strongly varying in the different parts of the composite Pelso Unit. High to medium-pressure (blueschist, greenschist and subgreenschist facies) Jurassic metamorphism is characteristic of the middle and lower parts of the South Gemeric Unit. The Bükkium, together with the Mid-Transdanubian (Igal) Zone, were subjected to low-medium-pressure, anchizonal to epizonal (subgreenschist to greenschist facies) regional metamorphism. In contrast, the Transdanubian Central Range escaped the Alpine regional metamorphism at the present erosional level.

Recent studies show that considerable areas of the Tisza Unit were also affected by Alpine (predominantly Cretaceous) metamorphism. The Mesozoic and Late Paleozoic formations show the signs of low to medium pressure-type anchizonal and epizonal metamorphism along the main Alpine thrust zones. Deciphering Alpine retrograde (and partly prograde) effects in the polymetamorphic complexes of the Great Plain is one of the greatest challenges Hungarian petrologists must face in the nearest future.

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The Dinaridic–Alpine connection – as seen from Hungary

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Comparative analyses led to the conclusion that prior to major displacements, certain elements of the basement mosaic of the Pannonian Basin belonged to the Dinaridic and the Alpine systems, or even to the junction zone between them. This review focuses on two key areas: the Alpine–Dinaridic–Pannonian triple junction in SW Hungary and NW Croatia, and the displaced Dinaridic terranes in NE Hungary. Between the Periadriatic–Balaton and Mid-Hungarian Lineaments, in the Sava Composite Unit, blocks of South Alpine and Internal Dinaridic origin are juxtaposed. The non-metamorphosed nappes of the South Alpine South Karawanken and Julian–Savinja Units are thrust over the Alpine metamorphic complex of the Medvednica and South Zala Units and the ophiolite melange of the Kalnik Unit, which was considered to be the prolongation of the Vardar Zone. The Bükk Composite Unit (Bükk Parautochthon, Szarvaskő and the Darnó Ophiolite Complexes), with its south-vergent structure and typical Dinaridic development, represents a displaced fragment of the Dinarides. These ophiolite complexes do not belong to the "Meliaticum" tectonostratigraphic unit, which is involved in the north-vergent Austroalpine nappe system. 400–500 km of dextral offset along the shear zone at the southern boundary of the Pelso system is well constrained. However, as a result of block movements, there is a sinistral facies offset of similar scale between the Pelso and the Tatro-Veporic system that should have preceded the Late Cretaceous nappe stacking.

Key words: tectono-stratigraphic units, structure evolution, Alps, Dinarides, Pannonian Basin

Introduction

One of the fundamental questions of the connection of the Dinaridic and Alpine systems is the northwestern termination the Dinaridic ophiolites and the nature of the contact between the Dinaridic and the Alpine units. For analysis of these questions and any other problems concerning the relationship of the Dinaridic and Alpine systems the complex evolutionary history of the Pannonian region must be taken into consideration. The present-day mosaic structure of the basement of the Pannonian Basin only came into existence by the Middle Miocene; before that, certain blocks (terranes) belonged to the Dinaridic, others to the Alpine systems and some of them were located just in the junction area. Due to collision of the large lithosphere plates and microplates, these terranes were extruded eastward and it cannot be excluded that the "missing links" were hidden at the bottom of the Pannonian Basin.

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Extensive exploration in the past half century revealed rock sequences of Dinaridic and Alpine affinity in the Pannonian region. In the last years much effort was made, through international co-operation, to correlate these sequences and reconstruct their paleo-position (see Haas et al. 2000; Pamić 2000; Protić et al. 2000; Dimitrijević et al. 1999 for latest reviews). According to the results of the comparative analyses, the transition/connection between the Dinarides and the Alps has been displaced about 400–500 km from the NW Dinarides to NE Hungary (Bükk and related units). The displacement took place along the Zagreb–Zemplin (or Mid-Hungarian) fault zone (Csontos and Nagymarosy 1998), forming the northwestern border of the Tisza Mega-Unit (Tisia Terrane).

In the present review we summarize the most important results of these efforts, focusing on two key areas: the Alpine–Dinaridic–Pannonian triple junction in SW Hungary and NW Croatia and the displaced Dinaridic terranes in NE Hungary.

The present contribution results from our cooperation with Dinaridic geologists (representatives of Croatia, Serbia and Slovenia; see acknowledgements and references cited) in the framework of the former IGCP Project No. 276 ("Paleozoic geodynamic domains in the Tethys and their Alpidic evolution"; final results published in Papanikolaou (Ed.) 1996/97) and its continuation, the ongoing Circum-Pannonian terrane map project.

Historical review

The necessity of a Late Paleozoic marine connection between the Dinarides and the Bükk was already recognized by Heritsch (1942, 1944) and Schréter (1936, 1943) referring to the striking similarities in their lithofacies and faunal content. Schréter (1948) already mentioned the facies relations between the Bükk and Jadar areas. Based on facies analysis and paleobiogeographical evaluation, Schréter (1959) and Balogh (1964) (updated and complemented by Pešić et al. 1988) proved a remarkable similarity between the Late Paleozoic development and fossil assemblages of the Jadar (W Serbia), Žažar (Slovenia) and Bükk (NE Hungary) areas.

From the early 1960s several boreholes encountered marine Permian formations and ophiolites in a narrow NE–SW trending zone between the NW Dinarides and the Bükk Mts. Based on these data, Wein (1969) introduced the term "Igal–Bükk Eugeosyncline" for an assumed seaway from the Dinarides to the Bükk. In a review on the ophiolitic rocks in Hungary Szepesházy (1977) arrived at a similar conclusion.

With the extension of the mobilistic plate tectonic concept it was not necessary to invoke a seaway for explanation of the close facies affinities. In his last work, Wein (1978) reinterpreted the notion of the "Igal–Bükk seaway", and following Laubscher's (1971) proposal, it was considered to be a major wrench-faulting zone.

Fülöp et al. (1987) defined the complicated, composite structural unit between the Balaton and the Mid-Hungarian Lineaments, suggesting the name Mid-Transdanubian Unit.

Comparative facies analyses carried out in the 1970s and 1980s led to conclusion that the Mid-Transdanubian units were originally located somewhere between the Southern Alps and the NW Dinarides from where they were extruded reaching their recent setting in the Tertiary (Kovács 1982; Balla 1988; Haas et al. 1995). According to Csontos et al. (1992) and Csontos and Nagymarosy (1998) a large-scale transpression zone was formed at the contact of the Alcapa and Tisia–Dacia terranes (which rotated in the opposite sense) during the Late Oligocene–Early Miocene.

The latest reviews (Haas et al. 2000; Pamić 1997, 2000; Kovács et al. 2000) assumed about 400–500 km dextral offset along this zone.

Units of Dinaridic and Alpine origin in the basement of the Pannonian Basin

The ENE–WSW trending Mid-Hungarian (Zagreb–Zemplin) Lineament subdivides the basement of the Pannonian Basin into two fundamentally different parts (Fig. 1). South of this lineament the Tisza Mega-Unit (Tisia Terrane) occurs. This is a large dismembered fragment of the Eurasian plate margin, which was a part of the Variscan Orogenic Belt (Buda 1996) and its evolutionary history was fundamentally different from that both of the South Alpine and the Dinaridic Units (Dercourt et al. 1993).

North of the Mid-Hungarian Lineament are located units of Alpine and Dinaridic relationships. These units belong to the Penninic, Austroalpine, Slovakocarpathian, and Pelso systems. They juxtaposed gradually and amalgamated into a large composite terrane (ALCAPA – Csontos et al. 1992) during the Early Tertiary.

The Penninic unit appears in the Rechnitz window in the territory of Hungary. In the basement of the Little Plain (Kisalföld), crystalline complexes of the Lower Austroalpine and equivalents of the Upper Austroalpine Graz Paleozoic series were encountered. The Slovakocarpathian system can be considered as the continuation of the Austroalpine system (Plašienka 1998). The upper nappe of its southernmost unit (Silica Nappe, in the sense of the cited author) extends over the territory of Hungary in the Aggtelek Mts. In contrast, Vozárova and Vozár (1992, 1996) consider only the Tatro-Veporic units (Tatro-Veporic Composite Terrane) as the direct continuation of the Austroalpine domain. According to the latter concept the Gemer-Bükk domain is mostly of Dinaridic affinity and belongs to the Inner West Carpathian–North Pannonian orogenic collage (which corresponds to the Pelsonia Composite Terrane, in sense of Kovács et al. 1997).

The Pelso system (Pelsonia Composite Terrane) is made up of units of South Alpine and Dinaridic facies affinity. They are the following: Transdanubian Range Unit, Sava Composite Unit, Bükk Unit and Aggtelek–Rudabánya Unit.

Fig. 1
Major structural units of the basement of the Pannonian Basin and the surrounding region

Problems of the latter are discussed in greater detail below. As to the Transdanubian Range Unit we note that due to its pre-Tertiary paleogeographic setting (see Majoros 1983; Császár and Haas 1984; Haas et al. 1995; Vörös and Galácz 1998) it shows mainly South Alpine facies relations, and according to paleomagnetic data it moved in coordination with the Southern Alps until the Late Cretaceous (Mauritsch and Márton 1995). However, as it supposedly rests above the Lower and Middle Austroalpine units (Tari 1996), its structural position appears to be similar to that of the Upper Austroalpine nappes (Haas et al. 2000).

The most recent characterization of the tectonostratigraphic terranes/units involved in the present contribution from the territory of Hungary are given in Kovács et al. (2000). In the present paper we only briefly summarize the characteristics of the units relevant to the Dinaridic–Alpine connection.

Alpine–Dinaridic–Pannonian triple junction in SW Hungary and NW Croatia

In a relatively narrow zone, between the Balaton and the Mid-Hungarian (Zagreb–Zemplin) Lineaments, displaced and sheared blocks were encountered beneath Neogene cover. Reflecting their significantly different origin, these blocks are made up of different rock successions. Based on evolutionary history and metamorphism they could be correlated with the structural units at the northwestern termination of the Dinarides in Croatia and parts of the South Alpine system in Slovenia, respectively. The Hungarian part of the composite unit between the Balaton and Mid-Hungarian lineament systems was named Central Transdanubian Unit (Fülöp et al. 1987) or Mid-Transdanubian Unit (Haas et al. 1988). Pamić and Tomljenovic (1998) suggested the name Zagorje–Mid-Transdanubian Zone for the entire composite unit (including the Croatian and Slovenian continuation of the Mid-Transdanubian Unit as originally defined). In a paper on the definition and correlation of the structural units and the bounding lineament systems Haas et al. (2000) suggested the simple term *Sava Composite Unit* to designate this heterogeneous zone.

Joint work of Slovenian, Croatian and Hungarian geologists revealed that large segments of the transition area of the Dinaridic and the South Alpine systems, including the Periadriatic Zone, were displaced during the Tertiary and they are presently located in the Sava Composite Unit (Haas et al. 2000). The northern boundary of the Sava Unit is the Balaton lineament system, which can be considered the direct continuation of the Periadriatic lineament system that separates the South Alpine and the Penninic–Austroalpine systems. The southern boundary of the Sava Composite Unit is the Mid-Hungarian lineament, which separates it from the Tisza Unit (Fig. 2).

In the northern part of the Sava Composite Unit the *South Karawanken* and the *Julian–Savinja Units* occur, consisting of non-metamorphosed, predominantly shallow marine Upper Paleozoic–Mesozoic sequences. They can be considered as the eastern continuation of the South Alpine system. Thin-skinned nappes of the Julian–Savinja Unit overthrust the Internal Dinaridic units (Fig. 3).

Alpine metamorphic complexes of the Medvednica and South Zala Units and ophiolite mélange of the Kalnik Unit belong to the Inner Dinaridic system. The Medvednica Unit is made up of low-grade metamorphic rocks. The Silurian to Triassic sedimentary and magmatic complex (Pamić and Tomljenovic 1998) was affected by Cretaceous metamorphism (110–120 Ma – Belak et al. 1995). It can be assumed that these metamorphic rocks were encountered in some wells in the Barcs area (Árkai 1990), in a nappe emplaced over the Tisza Mega-Unit (Haas et al. 2000).

The *South Zala Unit* is made up of Triassic deeper marine carbonates and Jurassic hemipelagic shale, and volcanoclastics affected by Cretaceous metamorphism (93–97 Ma – Árkai et al. 1991). Since character and age of metamorphism of the Medvednica and South Zala units are similar one cannot exclude that they are actually parts of a single structural unit (Haas et al. 2000).

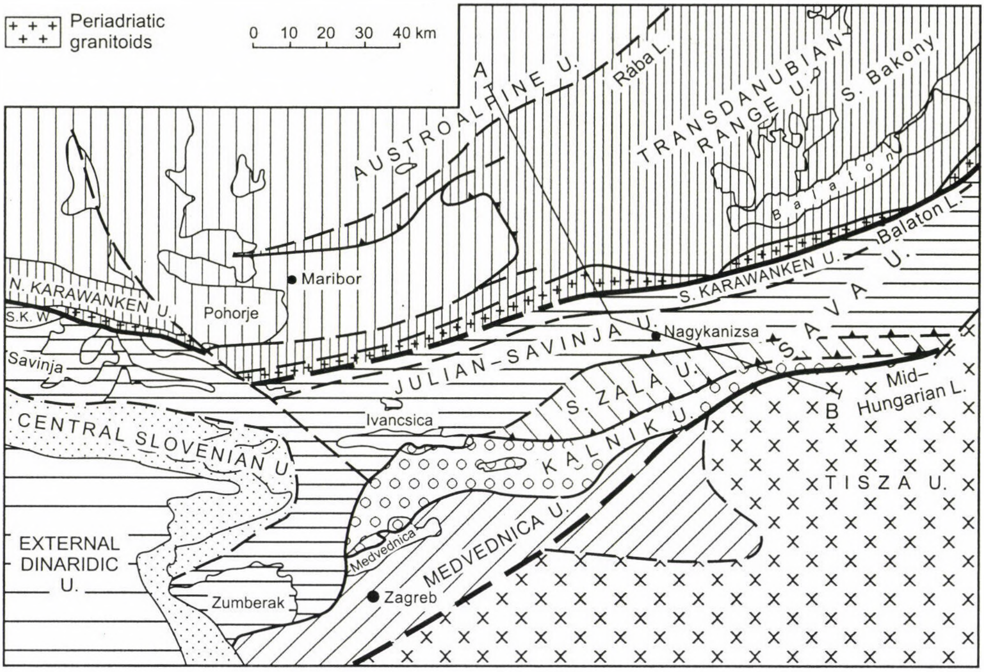

Fig. 2

Structural units in the basement of the Alpine–Dinaridic–Pannonian junction area (after Haas et al. 2000, modofied). A–B line of cross-section in Fig. 3

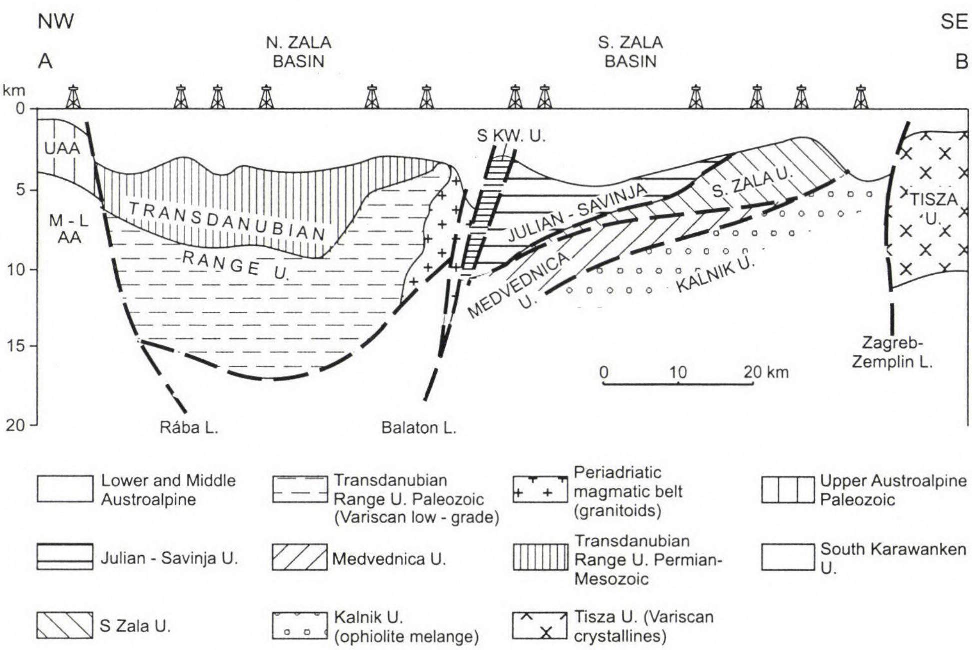

Fig. 3

Conceptual cross-section of the Zala Basin showing position of the structural units in the basement (after Haas et al. 2000). S.KW.U – South Karawanken Unit

The *Kalnik Unit* consists mainly of ophiolite mélange, which has been considered to be the prolongation of the Vardar Zone (Pamić 1997, 2000). In the outcrop areas (Medvednica, Ivanšcica, Kalnik Mts) the mélange complex is made up of smaller or larger blocks of basalt, gabbro, serpentinite, radiolarite, shale and limestone of varying age (the youngest ones are Late Cretaceous/Paleocene in age) occurring in a strongly tectonized shaly-silty matrix, probably of Cretaceous age (Pamić and Tomljenovic 1998). In the neighborhood of the Mid-Hungarian Lineament, in the surroundings of Inke, a thick complex of extremely varied lithology (acidic volcanoclastics, serpentinite, limestone, radiolarite and black shale) with an anchizonal metamorphic overprint was encountered (Árkai et al. 1991; Harangi et al. 1996), which may belong to the ophiolite mélange of the Kalnik Unit.

Terranes of Dinaridic and Alpine origin in NE Hungary

In NE Hungary two composite units (or composite terranes), wholly or partially of Dinaridic origin occur: the Bükk and the Aggtelek Composite Units (Fig. 4).

The *Bükk Composite Unit* (Bükkia Composite Terrane) is made up of the Bükk Parautochthon Unit, Szarvaskő and Darnó Ophiolite Complexes. Dinaridic as well as South Alpine (Carnic Alpine) characteristics of the Bükk Unit are as follows:

Bükk Parautochthon Unit: Variscan flysch in the Carboniferous, marine development of the Late Paleozoic (with a hiatus in the Lower Permian), marine passage form the Permian to the Lower Triassic, emersion and terrestrial conglomerate deposition in the Late Anisian, intense volcanism in the Ladinian and Carnian (first andesitic – variously interpreted as subduction or early rift-related – then WP-type basaltic; Harangi et al. 1996) and flysch-type Upper Jurassic representing the Eohellenic tectogenesis. The unit shows S to SE vergency, with emplacement of the Szarvaskő and Darnó Ophiolite Complexes from the NW over it (Geyssant and Lepevrier 1984; Csontos 1988, 1999) (Fig. 5). However, taking the large-scale Tertiary anticlockwise rotation of the ALCAPA unit into account (90° recorded in the Bükk area – Márton 1990; Márton and Fodor 1995; Csontos 1999, 2000), it should have been emplaced from the NE to SW, akin to the situation in the Dinarides. Most of the unit was affected by Late Jurassic (?) – Cretaceous anchi- to epizonal metamorphism, whereas some parts remained in the diagenetic zone (Árkai et al. 1995)

The *Szarvaskő Ophiolite Complex* records a Jurassic back-arc basin evolution (Balla et al. 1983). In the *Darnó Ophiolite Complex* both Triassic and Jurassic constituents occur in an accretionary complex. According to the recent interpretations, subsequent to the Triassic ocean opening an intraoceanic subduction zone came into being in the Jurassic and the spreading was renewed behind it (Józsa 1999; Dimitrijević et al. 1999; Migiros et al. in prep.). Middle Jurassic ooidic

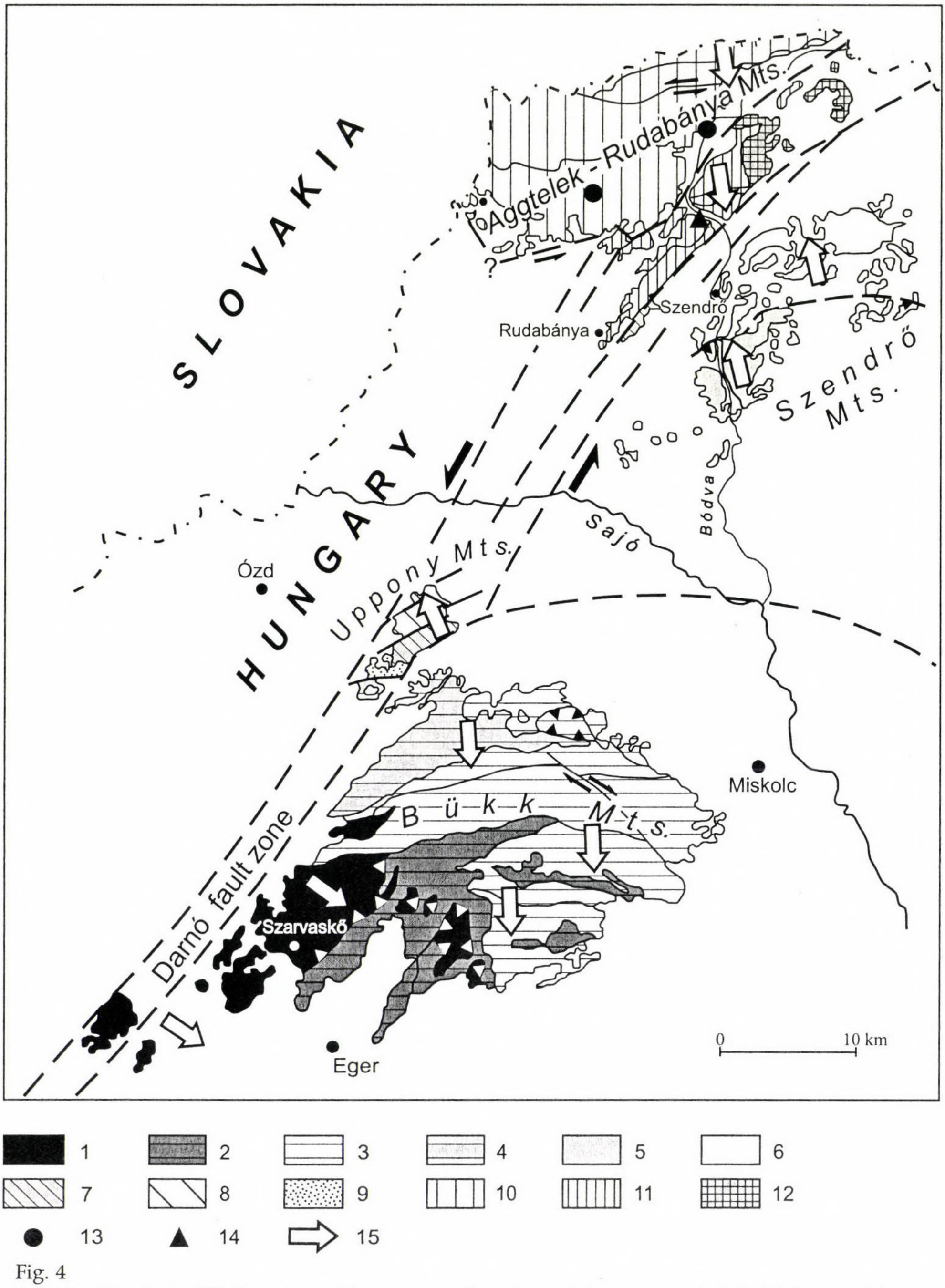

Fig. 4
Structural units in NE Hungary with vergence directions of the nappes. 1–4: Bükk Mts and Darnó Hill: 1. Szarvaskő and Darnó Opholite Complexes; 2. Jurassic formations of the Bükk Parautochthon Unit; 3. Triassic formations of the Bükk PA Unit; 4. Upper Paleozoic formations of the Bükk PA Unit; 5–6: Szendrő Unit: 5. Abod Subunit; 6. Rakaca Subunit; 7–9: Uppony Subunit; 7. Tapolcsány Subunit; 8. Lázbérc Subunit; 9. Upper Ctretaceous Gosau-type conglomerates; 10–12: Aggtelek–Rudabánya Mts: 10. Aggtelek Unit. s.l.; 11. Bódva (+Szőlősardó Unit); 12. Martonyi (or Torna s.s.) Unit; 13. Bódva Valley Ophiolite Complex (in boreholes); 14. Upper Jurassic(?) rhyolites; 15. structural vergencies

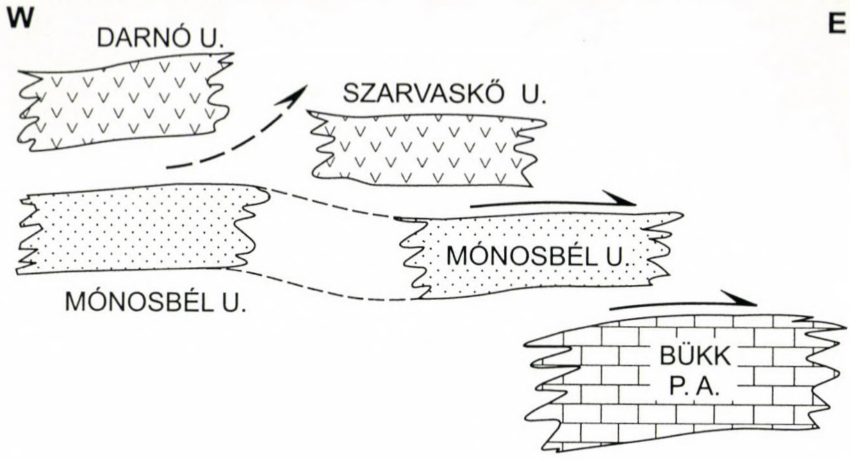

Fig. 5
Structural scheme of the Bükk Composite Unit

turbidites containing the foraminifer species Protopeneroplis striata (Pelikán and Dosztály 1999; Bérczi-Makk 1999) occur in both complexes. They are very peculiar formations, since no coeval carbonate platforms are known either in the ALCAPA or the Tisia units, which are at present in the neighborhood of the Bükkia Composite Terrane. Redeposited oolites could hardly be derived from an area outside of the Dinaridic Platform (including the Friuli Platform, which is the northwesternmost segment of the Dinaridic Platform) (Dimitrijević et al. 1999). Other specific constituents are slide blocks (olistothrymmata) of Ladinian–Carnian red, cherty limestone (resembling the Bódvalenke Limestone of the Bódva Unit of the Rudabánya Mts), radiolarite and basalt in Jurassic shale. Similar settings occur in the Dinarides (Dimitrijević et al. 1999) but are much more common in the Othrys and North Pindos ophiolite complexes of the Hellenides (Skourtsis-Coroneou et al. 1995; Migiros et al. in prep.). Although ultramafic rocks have not been preserved in the Darnó Ophiolite Complex, their former presence is indicated by serpentinite pebbles in the Miocene Darnó Conglomerate (Sztanó and Józsa 1996; Józsa 2000).

The Szendrő and Uppony Units are made up almost exclusively of Paleozoic rocks affected by Middle Cretaceous anchi- to epimetamorphism (Árkai et al. 1995). Based on their facial and supposed stratigraphic relationships, they were formerly assigned to the "Bükkium" (e.g. Balogh 1964; Kovács and Péró 1983). However, they show northern vergency (Austroalpine-type; cf. Koroknai and Frisch 1998), opposite to that of the Bükk nappe system, a fact not previously taken into consideration. Therefore this former assignment became ambiguous.

The *Szendrő Unit* represents a typical Austroalpine/Dinaridic–South Alpine transition in terms of its Variscan succession. Whereas its preflysch formations are correlatable with those of the Upper Austroalpine Graz Paleozoic (especially the

Middle Devonian, coral-bearing mixed carbonate-siliciclastic shelf sediments in both units), its flysch-type Carboniferous is a typical South Alpine–Dinaridic formation (cf. Ebner et al. 1991, 1998).

The Lázbérc Subunit of the *Uppony Unit* corresponds to the Graz Paleozoic, whereas the Tapolcsány Subunit (except for its volcanism) to the Bischofalm Facies of the Carnic Alps (Ebner et al. op. cit.).

The tectonostratigraphic units of the Aggtelek–Rudabánya Mts (Aggtelekia Composite Terrane) are made up by Upper Permian to Upper Triassic successions (Fig. 6). Differentiation of their sedimentary environments began in the course of the Middle–Late Anisian rifting. Jurassic rocks are known only in the Bódva Unit. As they have a certain importance concerning the paleogeographic setting of the highly debated uppermost Upper Austroalpine (Juvavic) units, they are discussed herein in more detail.

Located entirely within the Darnó Fault Zone, the Rudabánya Mts are made up of the non-metamorphosed *Bódva Unit* and the anchi- to epizonal metamorphosed *Martonyi* (or Torna s.s.) *Unit* (Less 2000). Although the latter should originally have been in a lower structural position, younger movements obliterated the original superposition of the units (Fodor and Koroknai 2000). The Bódva Unit of deep-water facies interfingers with the small Szőlősardó Unit of slope facies.

The Aggtelek Mts are built up by the non-metamorphosed Aggtelek Unit (consisting of the Aggtelek s.s., Alsóhegy and Derenk Subunits). Its sole thrust was formed on Upper Permian evaporites, containing slices of the Bódva Valley Ophiolite Complex (Less et al. 1998 and Less, 2000 for latest review).

The *Aggtelek* s.s. and *Alsóhegy Subunits* were in an outer shelf setting until the Late Carnian, with southward-facing Upper Anisian reefs at Aggtelek and Carnian ones at Alsóhegy (Fig. 6). The Wetterstein carbonate platform was drowned in the Late Carnian and the onset of pelagic Hallstatt Limestone indicates downfaulting of the shelf margin and back-stepping of the rimmed shelf to the north (Kovács 1984), as indicated by Dachstein reefs in the adjacent Slovakian territory (Mello et al. 1997; Less 2000). The *Derenk Subunit* was overthrust by the Alsóhegy Subunit from the north (cf. Less et al. 1998). It is characterized by Ladinian–Norian Hallstatt-type limestone with slope character until the Late Carnian. However, for evaluation of the facies polarity the large-scale Tertiary anticlockwise rotation (a 70–90° post-Triassic rotation was recorded in the Aggtelek and Bódva Units – Márton et al. 1988) must be taken into account.

Lower Jurassic shale and Middle Jurassic black, argillaceous-siliceous rocks make up the *Telekesvölgy Subunit* of the Bódva Unit (Grill 1988). Rhyolite subvolcanic bodies of the *Telekesoldal Subunit* are considered to represent remnants of a probably Late Jurassic magmatic arc (Kubovics et al. 1990; Harangi et al. 1996), whereas specific, limestone-rhyolite olistostromes (Kovács 1988) in the same subunit are viewed as a coeval trench complex.

Although the presence of Triassic Hallstatt-type limestone in the Aggtelek–Rudabánya units recalls the much debated Lower Juvavicum of the Northern Limestone Alps (Kovács 1980; for the controversial views about the latter see Schweigl and Neubauer 1997 versus Gawlick et al. 1998), it is important to emphasize that the southward facies polarity of the Alsóhegy reef toward the deep-water Bódva facies has been preserved in a predominantly south-vergent thrust-system (Less et al. 1998; Less 2000). Although no exact dating of the thrusting is available for the time being, it obviously took place after the

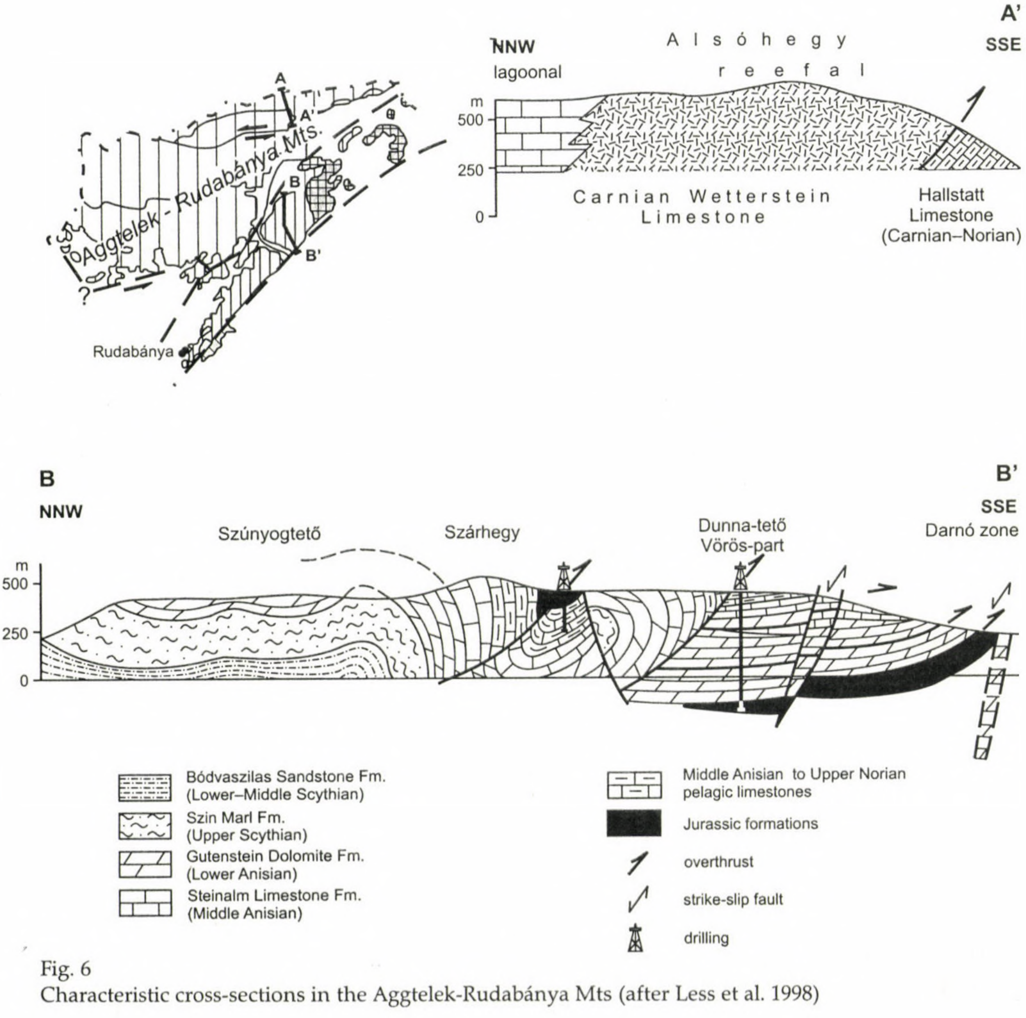

Fig. 6
Characteristic cross-sections in the Aggtelek-Rudabánya Mts (after Less et al. 1998)

obduction of the ophiolites (which is considered by most authors to be Late Jurassic) found in the sole thrust of the above-mentioned units. On the other hand, it is worth mentioning that Hallstatt-type limestone also occurs in the Dinarides and, further on, in the Hellenides (Skourtsis-Coroneou et al. 1995). The deep-water Bódva facies with red, cherty limestone (Bódvalenke Limestone Fm.) and related radiolarite, is not known in the Northern Limestone Alps, but similar deposits are very common in the Othrys and North Pindos ophiolite melanges of the Inner Hellenides.

Conclusions

Subsurface mapping of the critical southwestern Pannonian region revealed that abrupt northwestern termination of the Internal Dinaridic units is apparent. Sizable tectonic movements led to displacement of the transitional paleogeographic zones between the Dinaridic and the Alpine systems, i.e. the area surrounding the western termination of the one-time Vardar (Neotethys) branch. Sheared remnants of the ophiolite melange can be found along the major transpression zone, in the interior of the Pannonian Basin.

Between the Periadriatic–Balaton and Mid-Hungarian Lineaments, in the Sava Composite Unit, blocks of South Alpine and Internal Dinaridic origin are juxtaposed. The non-metamorphosed nappes of the South Alpine South Karawanken and Julian–Savinja Units are thrust over the Alpine metamorphic complex of the Medvednica and South Zala Units and the ophiolite melange of the Kalnik Unit, which was considered to be the prolongation of the Vardar Zone (Pamić 1997). The south-vergent nappe movements took place in the Oligocene–Miocene interval. This was followed by intense strike-slip movements in the younger Miocene, leading to lateral splitting, shearing and displacement of the nappes.

There is a consensus on a 400–500 km dextral displacement in the shear zone between the Periadriatic–Balaton and Mid-Hungarian Lineaments at the southern margin of the Pelso system (Pelsonia Composite Terrane). In contrast, there are different interpretations of the sinistral offset at the northern margin of Pelsonia, which was proposed by Kázmér and Kovács (1985). However, facies relationships of both Paleozoic (Ebner et al. 1998) and Hallstatt-type Triassic sequences (Haas et al. 1995) clearly indicate this offset. According to the latest structural investigations the displacement should have preceded the Late Cretaceous nappe stacking (Neubauer et al 1996; Plašienka 1997). As to the Eastern Alps, a model for the Late Jurassic–Early Cretaceous closure of the "Vardar–Meliata" ocean branch was already proposed by Frank (1987) and Tollmann (1987).

The Bükk-Szarvaskő–Darnó Block in NE Hungary is a displaced fragment of the Dinarides. The Bükk Parautochthon Unit was broken from the Dinaridic platform margin (Protić et al. 2000; Filipović et al. in press). The Szarvaskő and

Darnó ophiolite complexes rest upon it; it is part of a southerly vergent thrust system (cf. Geyssant and Lepevrier 1984; Csontos 1999; Dimitrijević et al. 1999 and in press).

The Aggtelek–Bódva couplet is either a fragment of the southern, marginal part of the North Tethyan margin representing the shelf margin, slope and pelagic basin formed on an attenuated and downfaulted continental margin (cf. Less 2000, for latest review), or the southern margin and slope (according to present coordinates) of a microcontinent/intraoceanic carbonate platform, facing a southerly-lying oceanic domain. In the latter case the Drina–Ivanjica Block of the Dinarides, situated within two ophiolite zones (Dimitrijević and Dimitrijević 1991; Karamata et al. 1999) provides a good analogue.

The Szarvaskő and Darnó Ophiolite Complexes cannot be fitted into any ongoing, Austroalpine-type Meliata-model (cf. Neubauer et al. 1996; Plašienka et al. 1998). That means that the northern, "Gemeric" part of the "Gemer–Bükk Composite Terrane" in the sense of Kovács et al. (1996/97), due to its Austroalpine-type structure could not be involved in the same paleotectonic scenario as the southern, Bükkian part, characterized by Dinaridic-type structure (cf. Geyssant and Lepevrier 1984; Csontos 1988, 1999). Eventually, Kozur and Mock (1995, 1997) raised the possibility of two, distinct zones with oceanic floor, although this version is under discussion (cf. Plašienka et al. op. cit.). However, the northern part is outside of the scope of the present review.

In contrast to the highly debated Lower Juvavicum of the Northern Limestone Alps, in NE Hungary the Hallstatt limestone facies is located south of the southward-rimmed carbonate platforms (Aggtelek, Alsóhegy), in a predominantly south-vergent thrust system, thus preserving the original facies polarity, even if the Tertiary anticlockwise rotation of the ALCAPA Composite Unit (Márton et al. 1988) is taken into account. This fact raises questions about the paleogeographic interpretation of the Hallstatt facies in the NW Tethyan realm (cf. Gawlick et al. 1998).

The interpretation of the paleotectonic position and structural evolution of the Szendrő, Uppony, Martonyi (or Torna s.s.) units and the Bódva Valley Ophiolite Complex (Réti 1984) requires further clarification. However, it is clear that the previously published "orthogonal" paleotectonic models for the Gemer–Bükk units (see, for example, in Kovács 1984 and Grill et al. 1984) are oversimplified and cannot provide a suitable explanation for the complicated setting. These blocks were only juxtaposed (at least in the territory of Hungary) in the Late Oligocene-Early Miocene, as constrained by their overstep sequences (cf. Szentpétery 1998; Less 2000).

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The SI (System International) should be used for all units of measurements.

References

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

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New possibilities for the evaluation of uncertainties in safety assessment of radioactive waste disposal

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In the first part of paper the authors discuss the general problems of uncertainties in safety assessments. The main types of potential errors are reviewed. Limitations of the traditional methods of evaluation – deterministic and probabilistic – are discussed. The authors review the main "uncertainty oriented" mathematical methods that are more suitable for safety assessments than the traditional ones. Possibilities of their application to safety assessments are discussed, with special attention to spatial and temporal uncertainties and risk analysis. As a conclusion, new steps and contents are suggested for safety reports, taking into account the uncertainty of the input data.

Key words: safety assessments, uncertainty analysis, risk analysis

Introduction

It is well known that safety assessment reports are crucial for the establishment of radioactive waste repositories (NEA Reports 1991, 1992). The Performance Assessment Advisory Group (PAAG) of the OECD Nuclear Energy Agency (NEA) executed a detailed comparison of ten selected safety assessments from seven countries (1997). A further report was published in 2000 by an International Review Team of the NEA for a deep repository of spent fuel in Sweden. These two reports presented excellent overviews of the methods applied in these safety assessments and they discussed existing problems as well. Based on these statements and on other published documents, such as SANDIA Reports 2364 (1994), 2240 (2000), and the papers of the journal "Risk Analysis" No. 5, 1999, we have applied new methodological approaches for safety assessments. These new approaches are outlined and discussed in this paper.

Significance and types of uncertainties in safety assessments

There is a general agreement among experts that the basic problem of safety assessments is their *reliability*, in other words the *uncertainties* related to their

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statements[1]. Much attention was paid to the identification and determination of relevant features, events and processes, called FEPs. This procedure helped to diminish uncertainties, but in our opinion, in order to achieve significant results, the reasons and the types of uncertainties – related to safety assessments – must first be clarified and determined. In this respect two main types of uncertainties can be distinguished:

1. *The natural variability* of the relevant geologic features, events and processes (FEPs), existing independently of our investigations, which is part of natural reality. According to geologic experience the degree of natural variability may range from very low to high levels in the different geologic environments. However, no completely homogenous (zero variability) or completely heterogeneous geologic objects have ever been observed. Another important aspect of variability is its *degree of regularity*. Experience shows that every geologic environment contains structured (ordered) and unstructured (disordered) elements. The structured ones can be mathematically described; thus, their uncertainty can be significantly reduced, e.g. regular transitions in space and time, or cyclic features and events. On the other hand, unstructured elements may appear unexpectedly in space or time, and their position and magnitude cannot be predicted. There are no perfectly ordered or disordered FEPs in geology, rather a mixture of the two types. Obviously, the higher the degree of variability and the larger the proportion of disordered FEPs, the larger is the uncertainty related to the given geologic object. Consequently, natural variability is site specific.

2. *Uncertainties of the investigations* related to the safety assessments. They are called *incertitude* in some publications. According to the generally accepted *multibarrier concept* engineered and geologic barriers and the surface environment (biosphere) are distinguished. In this paper only the uncertainties of the geologic barrier are discussed. Engineering problems, such as waste form, waste packages, packing materials, underground openings and their support, backfills, treatment of the EDZ, are beyond our competence.

The 1997 NEA Report stressed that the process of site characterization and that of safety assessment cannot be discussed separately. Thus the related uncertainties will be discussed in this paper from the beginning of the site characterization process. In this context even the composition and the total activity of the stored radioactive waste are uncertain, as representative sampling is difficult and limited. Furthermore, the analytical error can be high for certain radioisotopes.

2a) *Observation errors.* Most FEPs are first observed in the field. Due to unfavorable circumstances, such as unfavorable climate, vegetation cover and

[1] Uncertainty, in the broadest sense, is the recognition that the results of our investigations may deviate more or less from natural reality. The term error expresses quantified uncertainty, that is "the quantified difference between a true value and an estimate of that value" (Merriam-Webster Collegiate Dictionary, 10th edition 1997). In this context accuracy expresses the closeness of the estimate to the true value. Finally, bias is the consistent under- or overestimation of the true value.

sometimes haste, several observations may be incomplete, superficial or even partly or completely erroneous. Even the most thorough observational activities are often influenced by special personal fields of interest or curiosity, and as a result are biased.

2b) *Sampling errors*. There is very limited possibility to carry out adequate sampling of the geologic object in space and time. Only boreholes, pits, shafts and galleries allow underground sampling. In our opinion it is almost impossible to achieve perfect representative sampling for the selected host rock during the site characterization process. This inevitably leads to a certain amount of uncertainty.

2c) *Measurement errors*. During site characterization a large number of measurements are carried out at the site or in laboratories. Measurement errors occur due to the imperfection of the measurements. Random and systematic errors are distinguished. The main sources of measurement errors are the imperfection of the instrument and of measurement techniques, imperfection of the method applied (including calibration and sample preparation) and the incomplete skill and attention of the measuring person. These uncertainties are discussed in detail in analytical chemistry (Day and Underwood 1991), in geochemistry and in environmental science (Ferson et al. 1999)

In the NEA reports mentioned above these three error types are discussed together under the name "parameter uncertainty" or "parameter value uncertainty". We prefer to treat them separately in order to achieve better transparency.

2d) *Conceptual model uncertainty*. In safety assessments models should represent the various aspects of the barriers and the main processes. Unfortunately, according to the 1997 NEA Report the term conceptual model was found to have different meanings in different safety assessments. When constructing the models pre-existing conceptual geologic ideas are applied of necessity. Are they always adequate to the given geologic object of our site selection? Are the model boundaries right? Moreover, the incomplete or faulty knowledge of geologic phenomena and processes may lead to essential errors in geologic modeling. Even complete misidentifications may occur. *Natural analogs* broadly applied to site characterization are generally imperfect, as they cannot take into account undetected local, site-specific features. Particularly large hydraulic flow model uncertainties are encountered in fractured crystalline rocks, e.g. granite. Simplifications and generalizations made in conceptual modeling result in additional uncertainty. The selection of the appropriate scale of spatial and temporal representation is a further source of model uncertainty.

2e) *Scenario uncertainties*. According to the 1997 NEA Report a scenario represents a combination of FEPs that may affect the safety of the repository. Scenario uncertainties result from the incomplete knowledge of future events. They comprise the uncertainty of the determination of FEPs for each future scenario and their interactions, the identification of all possible scenarios at the selected site (a justification of the choice is often lacking) and the possible changes of the scenarios as time passes. Limitations of the scope of the given scenario influences the results and leads to further uncertainties. From the mathematical point of view these are all *predictions* becoming more and more uncertain with increasing time from the present situation. In accordance with the 2000 NEA Report we consider it to be a good idea to start with a "base case" and then develop the further, alternative scenarios.

2f) *Errors in the mathematical evaluation*. From the point of view of mathematical evaluation, *statistical* and *non-statistical* uncertainties can be distinguished (Ferson et al. 1999). Non-statistical uncertainties are not based on measurements but express experts' opinion. Thus, no probability distributions and statistical parameters can be calculated upon them. Mainly model uncertainties belong to this group, such as forms of rock types and deposition, forms and location of tectonic disturbances, climate types, etc. All other types of uncertainties listed above are of statistical character.

The most common source of error is the insufficient number of samples collected. According to leading statisticians, the minimum number of a reliable statistical evaluation is about 30 samples in cases not too different from the standard conditions (Tukey 1977). The solutions to this problem will be discussed later among the new mathematical methods.

Another source of error is the neglect of mathematical rules, e.g. that several statistical calculations require normal distribution of the variable. Nevertheless, often variables with skewed distribution are evaluated by these methods. Obviously the results will be biased.

The *propagation of errors* is also an important error component, as the errors of the individual measurements propagate differently, depending on the interdependencies of the variables. Moreover, the propagation of errors through mathematical calculations is also different. When applying the new, uncertainty-oriented methods to be discussed later, error propagation becomes a highly important subject, significantly influencing the results, as it is absolutely necessary to avoid repetition of the parameters in the calculations. For example, equation $a(b+c)=ab+ac$ does not hold for fuzzy numbers. If "a" contains some error, then the left side of the equation should be used instead of the right side where "a" appears twice. *Back-calculation* is another point where errors can be generated (Ferson and Kuhn 1992). Neglect of error propagation may significantly increase the overall uncertainty of the safety assessment. Even the programs ("codes") written for the different models and scenarios may contain erroneous assumptions, or programming errors.

The choice of the mathematical procedures should correspond to the given conditions of the geologic barrier. This choice may be difficult and may contain further uncertainty.

Finally, uncertainties are present in the calculation of *radiological risk* to hypothetical individuals at the site in the future. In this context the identification of the "critical group" is a further source of uncertainty (2000 NEA Report).

The first type of uncertainty, variability, is a property of Nature, being completely independent of us, whereas all other types of uncertainty are consequences of incomplete knowledge and imperfect human activity. Conceptual model uncertainty seems to be the major source of errors when predicting the future behavior of radioactive waste repositories. In order to distinguish the uncertainties listed above, safety assessments must assure complete *traceability* and *transparency*, as emphasized by the 1997 NEA Report.

The effectiveness of uncertainty analysis is highly dependent on the mathematical methods applied, for which reason a critical discussion of these methods is presented. First the current, traditional methods are discussed.

Traditional (current) approaches and methods of handling uncertainty

A variety of geologic investigations for site selection and site characterization are furnishing a growing number of observations and measurement results. Their proper evaluation is impossible without the application of appropriate mathematical methods. When applying mathematical methods to geologic problems, two types of approaches can be followed. The first is the *best guess* or *best estimate* approach. In this case point estimates are made, indicating what the author considers as the best result. We suspect that about 99% of all geologic investigations currently operate using the best guess approach. Variability is expressed by some statistical parameters, such as variance or standard deviation, but no special attention is paid to the quantification of uncertainty. Qualitative discussion of the uncertainties is obviously not a satisfactory solution.

The second is the *uncertainty-oriented* approach. In this case, from the beginning of the investigation, at each step, the uncertainty is a matter of special attention, and if possible appropriate methods are applied to quantify the uncertainty. Even the initial *input* data should reflect their uncertainty! In this way error propagation is also taken into account.

In case of purely scientific geologic investigations the best guess approach and the neglect of uncertainties had no serious consequences. One could expect that, with the overall development of geology, the effects of this neglect would be eliminated. On the other hand, in radioactive waste disposal severe short-term effects may occur if the uncertainties are not properly treated.

So far, the best guess approach has been applied in the calculations for radioactive waste disposal and particularly for safety assessments. The importance of uncertainty has been recognized, but only deterministic and probabilistic (stochastic) methods were used in all cases (Storck 1993; PAGIS 1988). However, a number of new mathematical methods were elaborated by theoretical mathematicians over the last decades. Some of them were found to be particularly suitable for solving the problems of uncertainty and risk in health, biology, ecology and environmental protection (Ferson et al. 1999). The main difference between the above-mentioned traditional and the new methods is in the type of their *input data*. The input data of the traditional methods are real numbers, whereas the new, uncertainty-oriented methods apply special input characteristics (uncertainty intervals, fuzzy numbers, probability bounds, etc.) to express their type and degree of uncertainty. These new methods have been broadly discussed by Ferson et al. (1999) for medical, biological, land use, and some environmental applications, but not for geologic problems. In the following we discuss the application of these methods to site characterization and safety assessments, showing their benefits as well as their limitations.

There are three main circumstances for the application of mathematical methods in safety assessments:

a) The evaluation does not include the spatial and temporal position of the input data, e.g. the determination of the chemical and mineralogical composition of rock samples
b) Evaluation of *spatially determined* input data, e.g. fractures, tectonic structures, ranges of influence of some FEPs, etc.
c) Evaluation of *spatially and temporally determined* input data, e.g. hydrogeologic flow data, predicted future climatic conditions, etc.

Different mathematical methods must be applied and combined for the evaluation of data under these three circumstances. So far this need has been neglected in several cases.

Limitations of the deterministic and probabilistic methods

The *deterministic methods* apply fixed (single) parameter values determined by the best guess approach, generally one of the measures of central tendency (Rock 1988). Based on these parameter values the geologic object or process is described by differential equations. These methods are straightforward but do not take into account the possible errors and error propagation. However, in the absence of perfect knowledge of the geologic object we cannot be sure about the exact values of the input numbers. All our calculations, although seemingly precise, harbor some degree of uncertainty. Consequently, deterministic methods can furnish unbiased results only if all the variables influencing the end results are known; furthermore, their proportion must also be well established and the relationships (positive and negative dependencies) among the variables be known perfectly. In geology, where perfect representative sampling is almost impossible and the influence of the different variables is only approximately detected, it is almost impossible to fulfill these requirements. We therefore consider the deterministic methods to be least suitable for safety assessments.

Computer-based repetitions of deterministic calculations with modified parameters and a choice of the "best estimate" may decrease the uncertainty of these calculations, but it is by no means a satisfactory solution, due to its subjective judgement and because it does not eliminate the basic shortcomings outlined above.

The *worst case analysis* (Morgan and Henrion 1990) is an approach that acknowledges the presence of uncertainty without modeling it explicitly. It works with the upper (or lower) bounds of distribution, attempting to insure that no larger (or smaller) value of the parameter may occur in the given system. This method has been applied for the safety assessment of several potential nuclear waste repositories by designating "realistic" and "pessimistic" values (2000 NEA Report). Obviously, the former expression corresponds to a best guess, and the latter one to a worst case value. Their calculation method is generally not explained. The application of a multiplication factor of 2, 5, 7 or 10 to arrive at the pessimistic value – as mentioned in the NEA Report (2000) – lacks any geologic justification and should be avoided. Experiences in other fields of research showed that many worst case analyses produce biased, hyper-conservative results (Ferson et al. 1999). Furthermore, the level of "conservatism" is not consistent from analysis to analysis, e.g. 5% and 95%, or 2% and 98% percentiles to be applied.

The well-known *probabilistic* (stochastic) methods have found broad application in safety assessments and in many cases furnished acceptable results (1997 NEA Report). However, these methods also operate with scalars, that is with integers and real numbers as input data (in uncertainty analysis scalars are often called *crisp numbers*). Degrees of uncertainty can be expressed to a certain extent by types of data, such as dichotomous, nominal, ordinal, interval and ratio-type data (Rock, 1988). *Bayesian* (*conditional*) *probabilities* represent an additional tool to detect and to describe uncertainties by applying information from prior investigations (Freeze et al. 1989; Stone 1990). In all such cases probability distributions must be defined. However, often the number of the input data is not sufficient to carry out these calculations (NEA Reports 1997, 2000). *Sensitivity analyses* help quantify how much an uncertain parameter may influence the annual dose rate and the radiological risk. In this way FEPs can be ranked according to their weight in the production of end-results of the safety assessment.

A further possibility for using probability distributions for uncertainty analysis is supported by *Monte Carlo analysis*, broadly applied to performance assessments in the last decade, e.g. the RIP software of Golder Associates. This method consists of computer-based, repeated, random sampling from each probability distribution of the FEPs, taken into account for safety analysis. Up to 10,000 realizations can be easily produced by appropriate computer programs. The method requires a lot of information, such as knowledge of the probability distribution of each variable, their mean and variance and even their interdependencies (correlations and cross correlations). Risk analysts usually assume the variables to be independent. However, according to our experience the variables of the geologic barrier are practically always interdependent; moreover correlation is often of non-linear character. Ferson (1994) stated that "an analysis that assumes independence among the variables could potentially under-estimate the probability of high exposures and any high-consequence effects of such exposures". He therefore suggested applying *dispersive Monte Carlo sampling* under the assumption of linear correlation among the variables, and *dependency bounds analysis* when no assumptions were made about the type of correlation, with the aim of yielding conservative results. We estimate that for geologic barriers the latter method is more suitable. Finally, the Monte Carlo method is inappropriate for semi-quantitative and qualitative data and for non-statistical uncertainty (Ferson 1996).

Instead of Monte Carlo analysis *Latin hypercube sampling* was applied by Iman and Shortencorier (1984) with the aim of insuring more complete exploration of the sample space. Recently, it has been applied in the 2000 SANDIA Report as well. A detailed comparison of the benefits and shortcomings of the two methods would be highly useful. Recently the theory of *copulas* (Schweizer and Sklar 1983) was applied by Haas (1999) for correlated random variables.

It should be stressed that joint probability distributions can also be calculated by *convolution methods*. They furnish more exact results but are cumbersome to perform.

The application of probabilistic methods has some theoretical limitations as well:

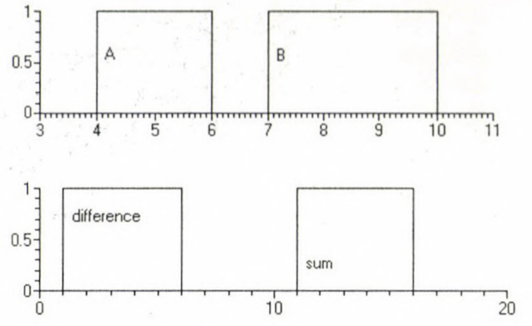
A. The additivity axiom – a fundamental law of probability theory – recognizes only mutually exclusive populations. As a consequence the method works with well-defined, sharp boundaries between the populations and no transitions are admitted. However, in geologic objects, including the host rocks of repositories, sharp boundaries are rare and gradual transitions with mixed features are more frequent. This circumstance should be taken into account in safety assessments, particularly in modeling of the scenarios.

B. Several statistical procedures require data from *repeated sampling*. In the investigations for site selection or site characterization this is often unrealizable, e.g. to repeat drilling of boreholes for transmissivity measurements. In such cases confidence intervals cannot be calculated.

C. *Closed systems*, that is systems characterized by fixed sums, such as chemical analyses of rocks expressed as percentages, furnish biased results when calculating correlation among their variables. There are several closed systems among the FEPs of safety assessments.

D. Several FEPs are not exactly defined. Their characteristics and boundaries can be indicated only in a semi-quantitative or even qualitative way. Traditional probabilistic methods are not suitable for the mathematical evaluation of such

Fig. 1
Two intervals and their sum (A+B) and difference (B–A)



FEPs. However, several important FEPs, significantly influencing the results of the safety assessment, belong to this group.

For all of the reasons listed above we consider that the deterministic and probabilistic methods are mathematically correct, but not optimal for the goals of safety assessments.

Review of the main uncertainty-oriented methods

A common feature of the uncertainty-oriented new methods is that they dissolve the above-listed limitations of the traditional methods. The most important point, however, is that they are able to mathematically describe the uncertainty of the input data by different types of "uncertain numbers". Finally, they insure the correct propagation of the errors throughout all the calculations of safety assessment.

Interval analysis (Moore 1979) replaces scalars by uncertainty intervals (Fig. 1). It is assumed that the true value is somewhere within the interval. Interval analysis lacks gradations and is the simplest method to express and calculate propagating uncertainty through arithmetic calculations. The method guarantees that the true value will always remain within the interval, but this goal is achieved at the expense of precision. Thus, the main shortcoming of interval analysis is that through the calculations the intervals become wider and wider and the final results become too conservative. According to our knowledge interval analysis has not yet been applied to safety assessment of radioactive waste disposal. The method seems to be most applicable in screening assessments.

Possibility theory, a generalization of interval analysis, provides a suitable model for the quantification of uncertainty by means of the degree of possibility of an event (Zadeh 1978). The theory acknowledges that not all types of uncertainty can be handled by probability distributions. Instead it uses intervals and fuzzy numbers to represent non-quantified uncertainty. The theory has been applied successfully in biology, health and medicine (Ferson and Ginsburg 1996; Ferson et al. 1999) and in different branches of industry and economy (A. Bárdossy and Duckstein 1995; Fodor and Roubens 1994).

The related *fuzzy* set and *possibility theory* expresses uncertainty by the use of *fuzzy numbers*. They represent estimates of uncertainty at different levels of possibility. Fuzzy numbers are by definition unimodal and they must reach, at least in one point, possibility level one, that is the full possibility. In geology mainly triangular and trapezoidal fuzzy numbers are applied (Fig. 2) and this is

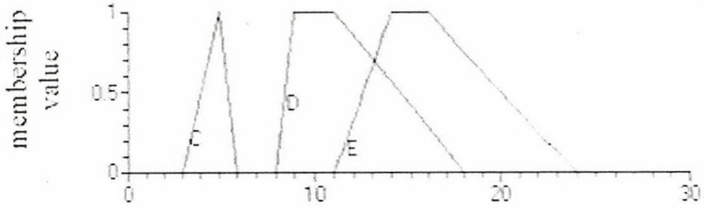


Fig. 2
Triangular (C) and trapezoidal (D) fuzzy numbers and their sum (E)

also the case in radioactive waste disposal. The smallest and the largest possible values of the given variable represent the lower and the upper bounds of the fuzzy number. All values of the variable must be within these boundaries. The values reaching possibility level one are considered as the most optimistic estimates of the given variable. The fuzzy numbers are generalizations of the scalars and the intervals. They are robust descriptions of the uncertainty of the given data.

All arithmetic calculations applied to safety assessments can be carried out with fuzzy numbers. One of their great advantages is that they do not require the knowledge of the correlations among the variables and the type of their probability distribution. For the sake of numerical comparisons and ranking, fuzzy numbers can be reconverted into crisp numbers. This calculation is called *defuzzification*. But the main advantage of the fuzzy method is that prior geologic experience can be incorporated into the construction of fuzzy numbers. This goal can be achieved by a joint constructing of the fuzzy numbers by the geologists or hydrogeologists involved and by the mathematician. In our opinion this is a significant advantage of this method over the traditional ones, particularly for safety assessments, as the definition of the fuzzy set incorporates the semantic content of the chosen set. The method allows the appropriate evaluation of semi-quantitative and qualitative input data as well. It is easy to communicate the results of the calculations with fuzzy numbers and uncertainty propagation is also transparent. The frequent transitions of the geologic populations, as mentioned before, can also be represented by fuzzy numbers (Fig. 3). The method has not yet been applied to safety assessments. We consider it to be the most suitable and efficient method for the problems of site selection, site characterization and safety assessment of radioactive waste disposal.

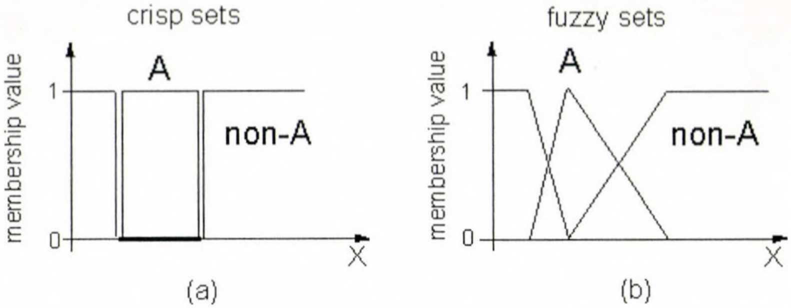


Fig. 3
a) Crisp set A and its complement non-A. Their intersection is empty, and their union is the set of all elements of the universe. b) Fuzzy set A and its complement non-A. They overlap

The manner of constructing fuzzy numbers raises the problem of their robustness. Imagine that several well-trained and experienced experts are asked to construct fuzzy numbers based on the same crisp data. It is certain that the resulting fuzzy numbers will not be exactly identical. However, the differences are expected to be rather small. Fortunately all of the mathematical operations to be carried out with these numbers are stable, that is, small changes in the input data yield only small changes in the output. As a consequence the final results are not sensitive to small differences in the initial fuzzy numbers.

In the last decade joint probabilistic and possibilistic methods have been developed as well:

· Probability bounds analysis (Smith 1996; Tessem 1992; Ferson et al. 1999)

· Neural networks (Kosko 1992; Singer and Kouda 1999; Fullér 2000)

· Hybrid arithmetic (Cooper et al. 1996; Ferson and Ginzburg 1996)

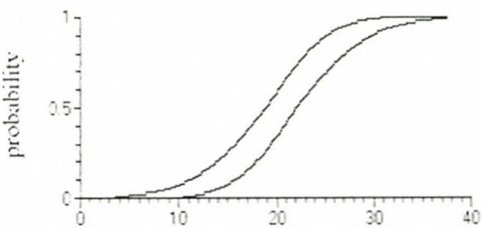


Fig. 4
Probability bounds

The *probability bounds method* is a combination of probability theory with interval analysis. It expresses uncertainty by two cumulative probability distributions. The area between the two curves represents the extent of uncertainty of the given variable (Fig. 4).

Probability bounds are considered to be a generalization of scalars, intervals and classical probability distributions. The great advantage of this method is that it can apply different probability distributions (e.g. normal, lognormal, Poisson, exponential, etc.), different parameter values and even correlations for the characterization of the given uncertainty, obtained from existing prior information. However, the method also works without making any assumption about the type of distribution, etc. The probability bounds become narrower with more empirical prior information on the given FEPs of the geologic barrier. Its disadvantage is that it is more complicated than the preceding methods. Nevertheless it seems to us to be a highly efficient method for safety assessments, when prior information is abundant.

In modeling of complex systems and processes, it is usually assumed that an analytical system model can be defined. However, many processes are so complicated that a global model cannot be built. *Neural networks* build models directly based on measurements. They consist of adaptable nodes, which through a process of learning from task examples, store experimental knowledge and make it available for use (Alexander and Morton 1990). Neural networks, with their remarkable ability to derive meaning from complicated and imprecise data, can be used to extract patterns and to detect trends that are too complex to be understood by humans or other computer techniques. A trained neural network can be thought of as an "expert" and used to provide predictions for future scenarios in a safety assessment. It can also distinguish areas within the site of the study that satisfy regulatory safety standards, and other areas that do not and should be excluded from further considerations. The use of neural networks, however, requires that a large amount of good-quality data describing the process are available. Recently neuro-fuzzy systems were developed (see Fullér 2000), thus enlarging the method with complementary aspects of uncertainty.

The method of *hybrid arithmetic* combines probability distributions with intervals, fuzzy numbers and probability bounds. It allows the use of all kinds of input numbers, including scalars, which is its greatest advantage. Arithmetic calculations can be carried out by convolving the elements according to fuzzy arithmetic and probability theory. We do not have direct experience in the suitability of this method for safety assessments, but theoretically it should work well. We foresee test calculations in the near future.

All of the methods listed above require for their calculations at least some 30 cases as a sample size. Below this number the results become more and more uncertain (Tukey 1977). This circumstance presented difficulties in several safety assessments (NEA Reports 1997, 2000). The recently developed *bootstrap method* (Efron and Tibshirani 1993; Davison and Hinkley 1997) allows diminishing the sample size to about 10 cases by performing computerized random replicate sampling and by calculating the statistics of each replicate. Up to 1000 replicates can be quickly obtained by adequate computer programs. The method has been tested recently in Hungary on the potential host rock of HLW, the "Boda Claystone Formation", situated in southern Hungary. The results proved the mathematical correctness and the efficiency of the method. Another method to calculate with small sample size is the *jackknife method* (Rock 1988). It is similar to the bootstrap method with the difference that it produces replicates by omitting one data in turn and then averaging the statistics of the trimmed replicates.

Calculations of spatial uncertainty

The calculations using the new uncertainty-oriented methods outlined above did not extend to spatial problems, as they were applied mainly to medicine, health, biology, social sciences and chemistry (Ferson et al. 1999). Many FEPs of the safety assessments require the evaluation of their spatial position as well. Consequently the uncertainties related to them must also be evaluated within a spatial context. As is well known, for any spatial evaluation spatial coordinates (X, Y, Z) must be added to each input data involved and the spatial position must be part of all further calculations.

The theory of *regionalized spatial variables*, also called "geostatistics" (a term leading to many misunderstandings) and developed by Matheron (1971), solved the problem of spatial evaluations within the framework of traditional probability theory. The method of calculating *variograms* allowed quantifying spatial variability and the determination of the "range of influence" in space for the given variable, that is the range of spatial autocorrelation. In our opinion, variograms are highly important tools for understanding the spatial variability of any variable and thus for diminishing their spatial uncertainty. Spatial predictions required at safety assessments can be performed by *point-* and *block-kriging* and even the standard error of the prediction can be calculated. The results can be represented on isoline maps. Matheron's theory was a real breakthrough for spatial calculations in geology but also has some limitations, such as the requirement of first and second-order *stationarity* in the study area. A further limitation is that semi-quantitative and qualitative input data cannot be evaluated by this method. Geostatistics have been broadly applied during the last decades in petroleum and mineral exploration and mining. Unfortunately, they have not yet found application in radioactive waste disposal and in safety assessments.

Matheron's theory is the best solution for spatial averaging and its variance, but does not take the uncertainty of the input data into consideration. The development of *fuzzy geostatistics* (A. Bárdossy et al. 1990) and particularly of fuzzy variograms and fuzzy-kriging was an essential step for the handling of spatial uncertainty in geology. It should be applied to safety assessments as well.

In our opinion, spatial equivalents can be developed for all non-spatial, uncertainty-oriented methods discussed in the preceding chapter. As many FEPs are spatially determined it would be very important to also carry out test calculations for safety assessments by these methods.

Calculations of spatial and temporal uncertainty

The study of temporal processes requires the introduction of another dimension, that of time. Methods of time-trend analysis are well known and applied in geologic investigations (Davis 1986), but their uncertainties have so far been treated only by the traditional deterministic and probabilistic methods.

Experiences have shown that several temporal processes, e.g. in hydrogeology, are not of linear type but nonstationary and only quasi-periodic. Their mathematical treatment is therefore different from simple spatial evaluations. Uncertainties of joint spatial-temporal processes have so far attracted very limited attention.

Temporal evaluations are of great importance for safety assessments when calculating the effects of different future scenarios. Particular attention has been paid to groundwater-flow models, transport and sorption of radionuclides in fractured crystalline rocks (1997 NEA Report). So far only deterministic and probabilistic methods have been applied. Further theoretical studies are needed, in our opinion, to elaborate the uncertainty analysis of spatial-temporal events and processes in the disposal of radioactive waste by taking the new uncertainty-oriented methods into account.

Uncertainty of risk analysis in safety assessments

Risk analysis is one of the final steps of safety assessments to meet the specific requirements of the regulatory criteria. These calculations are carried out at present by traditional deterministic and probabilistic methods and were discussed in several NEA Reports (1997, 2000). The basic requirement for the calculations is to exclude the possibility of under-estimation of risk at the given repository conditions. There are still open questions as well, e.g. a more exact definition of risk in safety assessments, as emphasized by the 2000 NEA Report and the special issue of "Risk Analysis" (1999, No. 5). As Ferson (1994) pointed out, it is generally not the given measure of central tendency that is of paramount importance in risk analysis, but the tails of the distribution, which represent risks of low probability but of severe consequences. Dependency bounds analysis seems to guarantee sufficiently conservative estimates of tail probabilities. Furthermore, the adverse consequences of risk should be calculated separately.

In risk analysis arithmetic procedures are necessary among two or several variables, represented by their probability distributions. To perform these calculations *convolutions* must be applied (Williamson and Downs 1990; Ferson et al. 1999). The necessity of this type of calculations was ignored on many occasions so far. The 1997 NEA Report stated that "there are differences in how probability estimates were combined into a joint risk estimate".

We are convinced that by application of the outlined uncertainty oriented methods the uncertainties related to risk calculations could be significantly diminished and the final joint risk and its adverse consequences can be presented in the form of fuzzy numbers or probability bounds. Corresponding test calculations will be presented by us in the near future.

Conclusions

A. Site characterizations and safety assessments for future radioactive waste repositories still contain large amounts of unquantified uncertainties.

B. The traditional deterministic and probabilistic methods have theoretical limitations that do not permit evaluating all the input data (semi-quantitative and qualitative ones) and all of the situations characteristic of the geologic barrier (e.g. transitions). Expert judgement is certainly important but cannot replace the uncertainty-oriented mathematical methods outlined above because no expert has the right to present his opinion as a fact.

C. By applying the new, uncertainty oriented mathematical methods a large proportion of the uncertainties can be quantified and diminished and all available semi-quantitative and qualitative information can be included in the calculations. We admit that even by applying the methods outlined above some model uncertainties may remain non-quantifiable.

D. Traceability and transparency must be indispensable requirements of the safety assessments and reports.

E. In a preliminary manner the following *uncertainty-oriented steps* can be suggested for future site characterizations and safety assessments (we suggest applying the contents of the *safety reports* according to the recommendations of the 1997 NEA Report):

1. *Selection* of site-specific FEPs for the site characterization and safety assessment
2. Collection of the *crisp input data*, including the semi-quantitative and qualitative ones
3. Construction of *uncertainty-oriented input data*, such as uncertainty intervals, fuzzy numbers, probability bounds and hybrid numbers, and calculation of the main statistics
4. Calculation of *ranges of influence* (variograms) for the main FEPs
5. Calculation of conceptual and site-specific *models* for the different barriers, main processes and mathematical expression of the models
6. Calculation of *model uncertainties* taking error propagation into account
7. Identification of the possible future *scenarios*, their effects and calculation of their uncertainties
8. *Mathematical simulations* to estimate the conformity of the models and scenarios.
9. *Sensitivity analyses* to examine effects of changing certain parameters of the models
10. Overall verification of the models and scenarios by *back-calculations*
11. Uncertainty-oriented *risk calculations* (adverse consequences expressed by fuzzy numbers or probability bounds)
12. Compliance with *regulatory criteria*, for annual radiation dose limits and radiological risk limits

13. Suggested *decisions* for regulatory authorities (approval, further investigations or rejection of the site)

F. Further theoretical mathematical research is needed to meet all the calculation requirements of the above-listed steps, particularly for spatial and spatial-temporal calculations of uncertainty. Its application should be tested both on conceptual safety assessments, like SR97 in Sweden, and on site-specific assessments

G. We agree with the 2000 NEA Report that repeated, iterative safety assessments should be executed in order to increase the staff and expert experience and the skills to execute the investigations and mathematical calculations.

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Petrochemical database of the Cenozoic volcanites in Hungary: structure and statistics

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Since the beginning of geologic research in the region thousands of major element chemical analyses have been performed on Cenozoic volcanites found in Hungary. These data are scattered in numerous published studies, technical reports and partial databases. In the present project, most accessible (more than 3000) analyses have been collected and organized into a computerized database. In the database are registered a complete reference of all data sources, the most possible exact location of each sample, ages and lithostratigraphic units of the rocks analyzed, original major element and related chemical data, component values recalculated on a common basis, and some relevant background information. In this paper a comprehensive description of this database is given, based partly on a many-sided statistical characterization of the figuring geologic formations. Some of these statistics suggest that major element data may have preserved effects of genetic elements common for all these formations. Some methodological problems related to data homogenization and the effect of closure are also discussed. A few suggestions for further applications of the database are provided at the end.

Key words: database, volcanite, major oxides, statistics, Cenozoic, Hungary, calc-alkaline rocks, alkali basalts, Carpatho-Pannonian Region

Introduction

Areal distribution of the Cenozoic volcanites, as well as their stratigraphy, petrography, petrology, age, and geodynamic aspects have been discussed in a great number of studies since the beginning of geologic exploration in the Carpatho-Pannonian Region, which includes the territory of Hungary. Meanwhile, hundreds of major element, and, later, trace element and isotope analyses have been made of bulk rock and constituent minerals. Trace element and isotope data serve as the most up-to-date geochemical tool for solving genetic questions like that of magma origin, age, evolution, and relationship with tectonic processes. Major element analyses are traditionally used for calculation of normative composition and various petrochemical indices, as well as for making bivariate and triangular diagrams which, in turn, are useful in comparing and classifying particular samples, volcanic bodies or regions.

Statistical elaboration of compositional data from volcanites in Hungary began with pioneering works of Dienes (1971), Vogl (1979, 1980), Embey-Isztin (1980, 1981a, b), and Embey-Isztin et al. (1985), dealing mainly with basic rock types. A

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large dataset embracing most of the Neogene and Quaternary formations was compiled and statistically processed by Póka (1985, 1988). Production of statistically significant, high quality analytical data of Cenozoic basalts was reported by Embey-Isztin et al. (1993). More sophisticated multivariate methods were applied by Kovács and Ó.Kovács (1990) and Ó.Kovács and Kovács (1994) for a statistical evaluation of major element data from the young alkali basalts of Hungary. From the point of view of multivariate methodology helpful in chemical classification of volcanites, studies of Harangi (e.g. 1990), Szabó et al. (1992) and M. Tóth et al. (1998) are also relevant.

The present paper aims at introducing a recently compiled digital database attempting to incorporate all major element data from these geologic formations accessible in the literature.

Geologic background

The immediate history of the rock formations represented in the database began with the generating of the magmatic material from which they were formed. Generation of magma, although there are a lot of hypothetical points in the relevant theories, is usually related to the tectonic evolution of the region concerned. The Cenozoic tectonic and/or magmatic history of the Carpatho-Pannonian Region is discussed and modeled in numerous papers (Balla 1980, 1981, 1987; Csontos 1995; Doglioni 1993; Downes et al. 1995; Embey-Isztin and Dobosi 1997; Harangi et al. 1995b; Horváth 1993; Póka 1988; Royden 1988; Szabó et al. 1992; Tari et al. 1993, etc.). From the point of view of initiating magma generation, three major motifs figure in these interpretations: a Late Cretaceous-Eocene-Miocene lithospheric subduction, a partly accompanying or subsequent extensional tectonic regime, and an effective mantle plume in the Middle Miocene to Pleistocene. Simplifying the suggested genetic ideas, subduction of the European-Outer Carpathian lithosphere beneath the Carpathians roughly from N to S is responsible for an Eocene to Miocene intermediate to acid volcanism, while the extension coupled with mantle upwelling in the Pannonian Basin for a younger and subordinate in volume basaltic volcanism. The actual genetic models are, of course, more sophisticated and accurately elaborated; besides, the birth of a number of rock varieties (e.g. the Bár and Balatonmária potassic rocks) has been given more intricate explanations (Harangi et al. 1995a).

Events of volcanic activities resulting in the formations covered by the database are traditionally combined to make the next three major stages. (1) Eocene andesites associate intermediate predominantly pyroclastic and lava products, and subordinately subvolcanic and intrusive, often metasomatized ones of a multi-cycle volcano-magmatic activity, of calc-alkaline chemical character and larger masses concentrated in three areas revealing stratovolcanic structures. (2) In the Miocene calc-alkaline volcanism became active in multiple cycles affecting a number of regions at different times (Fig. 1 and Table 1), and

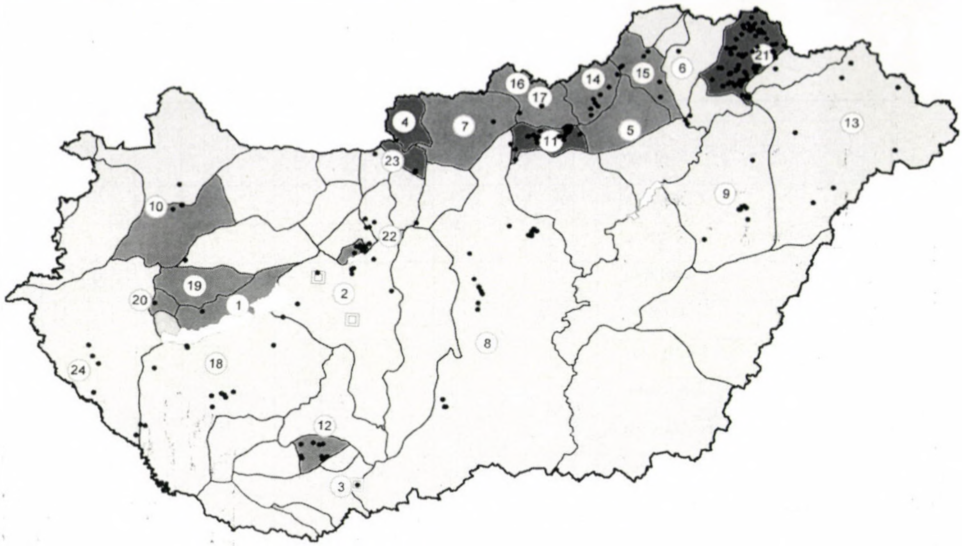

Fig. 1

Location of the samples stored in the database. Legend: dark grey – mountain range built up by Cenozoic volcanites, grey – geographic-geologic unit with abundant outcrops of Cenozoic volcanites, light grey – geographic-geologic unit with no or minor outcrops of Cenozoic volcanites; square – sampled minor outcrop of Cenozoic volcanites, dot – borehole with analyzed sample(s); numbers indicate the concerned geographic-geologic units as follows: 1. Balaton Highland, 2. Balatonfő–Mezőföld area, 3. Baranya Hills, 4. Börzsöny Mts, 5. Bükk Foreland, 6. Cserehát Basin, 7. Cserhát Hills, 8. Danube-Tisza Interfluve, 9. Hajdúság area, 10. Little Hungarian Plain–Kemeneshát Hills, 11. Mátra Mts, 12. Mecsek Mts (East), 13. Nyírség–Szatmár Basin, 14. Ózd Basin, 15. Sajóvölgy Basin, 16. Salgótarján Basin (Karancs area), 17. Salgótarján Basin (Medves area), 18. Somogy area, 19. South Bakony Mts, 20. Tátika Hills, 21. Tokaj Mts, 22. Velence Mts and surroundings, 23. Visegrád Mts, 24. Zala Basin.

resulting in large, thick (up to hundreds, in places to thousands of meters) patches of andesitic through dacitic to rhyolitic, sometimes trachytic tuffs, agglomerates, and, (locally) lavas or ignimbrites. They were deposited in a terrestrial or aquatic environment, and are often stratified and/or interbedded with clastic sediments, and in places form large stratovolcanos. In certain occurrences this stage is only represented by subvolcanic bodies and dykes. (3) Products of the Middle Miocene to Pleistocene basic volcanism are represented by Na-alkaline and, in limited occurrences, K-alkaline basalts and their pyroclastics, and form hill-sized lava domes, tuff rings, volcanic remnants, or smaller lava, lava-breccia and tuff body layers. The division of the given volcanites into these three associations is convenient but somewhat arbitrary as it masks important differences within the associations and possible relations between them.

Table 1
Number of samples from each formation and area represented in the database

FORMATION	AREA																								TOTAL
	Balaton Highland	Balatonfő-Mezőföld area	Baranya Hills	Börzsöny Mts.	Bükk Foreland	Cserehát Basin	Cserhát Hills	Danube-Tisza Interfluve	Hajdúság area	Little Hungarian Plain - Kemeneshát	Mátra Mts.	Mecsek Mts. (East)	Nyírség-Szatmár Basin	Ózd Basin	Sajóvölgy Basin	Salgótarján Basin (Karancs area)	Salgótarján Basin (Medves area)	Somogy area	South Bakony Mts.	Tátika Hills	Tokaj Mts.	Velence Mts. and surroundings	Visegrád Mts.	Zala Basin	
Balatonmária basalt																		7							7
Bár Basalt F.			11																						11
Börzsöny Volcanite (F.)				124																			52		176
Cserehát F.						5									3										8
Felsőnyárád F.															3										3
Galgavölgy Rhyolite Tuff F.		2			5	4								4	6										21
Gyulakeszi Rhyolite Tuff F.					19		4				20	8		10	1										62
Hasznos Andesite F.							6				27														33
Kecel Basalt F.								6																	6
Mecsek Andesite F.												19						1							20
Mátra Andesite F.							57				346					20								3	426
Nadap Andesite F.		5																15				208			228
Nyírség Volcanite (F.)									10				26												36
Pásztori Trachyte F.										9															9
Recsk Andesite F.											381														381
Sajóhídvég Trachyte (F.)						3																			3
Sajóvölgy F.															7										7
Salgóvár Basalt F.																	70								70
Szentmihály Andesite F.																		1						4	5
Sárospatak basalt																					9				9
Tapolca Basalt F.	125									94									84	77					380
Tar Dacite Tuff F.					27		2	5			26	22	1	6	7			7			92				195
Tokaj Volcanite F.																					825				825
unassigned Cenozoic volcanite					2			3			2				1									1	9
unassigned Miocene volcanite								22																	22
TOTAL	125	7	11	124	53	12	69	36	10	103	802	49	27	20	28	20	70	31	84	77	926	208	52	8	2952

For explanation see sections "Names of geologic formations and frequencies by area vs. formation".

Collection of data

Collection of data was simultaneously launched from several starting points. Comprehensive studies on the three major volcanic areas of Hungary, namely the Mátra Mts (Varga et al. 1975), Tokaj Mts (Gyarmati 1977), and Börzsöny Mts (Pantó 1970), with a great number of samples in each, were processed first. Several existing specialized databases were also adapted or acquired and, after revision and modification, incorporated. These include the database of young alkali basalts (Ó.Kovács and Kovács 1994, largely based on Jugovics 1976), the Cenozoic volcanites from the database of Transdanubian igneous rocks compiled by Darida-Tichy (1991), analyses from the Eocene andesites and related rocks of the Recsk area (Gasztonyi and Holló 1991), and a database of unpublished data from the Tokaj Mts (created by Szanyi (1998), using Gyarmati's (1978) manuscript register). About 650 out of the 22,000 samples in the international igneous rock database IGBA (Brandle and Nagy 1995) originate from Hungary, around 300 of them from Cenozoic formations. However, the structure of that database is so different from the one described below that it was easier to re-collect the data stored there, making use of IGBA's reference list, than to adapt them.

Furthermore, Hungarian and international bibliographic databases were queried seeking for related studies. The promising references in each reviewed publication on the hit lists were also checked for major element data. Unpublished technical reports were also inspected if they were explicitly referred to as a data source (the number of available manuscript reports dealing with Cenozoic volcanites is close to one thousand, their comprehensive review was beyond the scope of this project). Thus, several hundreds of publications and dozens of reports were looked over for analytical data, about two hundred of them successfully (only some of these were mentioned on the Reference section of this paper).

As a result, data of well over 3000 samples of Cenozoic volcanites with major element chemistry were collected. Only individual samples were selected, i.e. average compositions were not considered. At the final stage of data search, more and more often samples already stored in the database were encountered in newly reviewed publications. This led to an estimation that not more than 500 further samples could remain hidden in rarely cited works, mainly in unpublished reports.

Quality of data

Reliability of numerical data is always important in a computerized environment. The collected data are rather heterogeneous regarding both the place and time of chemical analyses performed; i.e. different laboratories, analysts, and partly, methods have been involved. The dates of analyses roughly span the twentieth century. Luckily, all these analytical data have a common feature, namely, the expected sum of major oxides expressed in weight % should be equal to 100. This fact, causing severe problems in the mathematical analysis of certain properties of this type of data (see section Correlations and the problem of closure), is beneficial from the point of view of data quality, as it has served for a self-control of analysts at all times. An eventual damage of this inherence has urged chemists to rethink their work (Csajághy 1959). Hence, the overwhelming majority of the encountered data could be entered into the collection. Nevertheless, some erroneous data may be hidden in the database. For instance Harangi (1999) noted that his new analytical data on the rather homogenous dacites of the Csódi Hill showed a remarkable difference in the SiO_2 and Al_2O_3 contents compared to those published 65 years before. The revealing of such cases and an appropriate treatment thereof should be part of particular applications relying on the given database, although no statistically significant error-related anomalies have been detected.

The quality of data entry has also been verified by comparing the calculated sums of components (corrected for S, Cl and F, if present) with the published ones. Even if there was no divergence, totals less than 99% and more than 101% have all been checked. This has assured an essentially error-free data input.

Names of geologic formations

As Fig. 1 shows, in most of the areal units concerned, rocks other than volcanites are also present. On the other hand, these areal units usually host more than one volcanic formation. As stratigraphers often prefer presenting interpretations in terms of formally-defined lithostratigraphic units (formations), all efforts have been made to assign each sample to an "official" formation (Császár 1997). The latter gives a good summary of the system of currently accepted formations with a short and informative description of each.

A few Cenozoic volcanite occurrences represented in the given database, however, are not covered by these lithostratigraphic units (e.g. Balatonmária basalt), or, despite their significance, they are incorporated into other formations (e.g. Nyírség into Tokaj). They are those without the letter "F." (Formation), and with the sign "(F.)" in Table 1, respectively. The silica-saturated, potassic Balatonmária basalt (Middle Miocene) found in a borehole was described by Harangi et al. (1995a), revealing its subduction-related petrochemical character. The Upper Miocene Sárospatak basalt (borehole Sárospatak-10.) is considered a member of the Miocene calc-alkaline suite, wide-spread in the Pannonian Basin (Gyarmati 1977; Downes et al. 1995). The Börzsöny Volcanite (F.) represents all volcanic rocks in the Börzsöny and Visegrád Mts., as defined in Korpás (1998). Rocks of Sajóhídvég Trachyte (F.) are known from several boreholes (Mauritz and Tolnay 1953). Juhász (1971) reports of, among others, predominantly Middle Miocene intermediate to acidic pyroclastics and lavas of uncertain stratigraphy traversed by boreholes in the Danube-Tisza Interfluve; here they are referred to as "unassigned Miocene volcanite". "Unassigned Cenozoic volcanite" denotes those samples that could not be unambiguously attached to any formation. The Felsőnyárád Formation is a special case because it is given in Császár (1997) but with no volcanites in the description there, despite the observations of Radócz (1975), whose interpretation is respected in the database.

Description of the database

For simplicity, the data are organized into one table (created in "dbf"-format). Each row (record) of the table represents an analyzed rock sample; columns (fields) correspond to attributes of samples. The attributes form three groups: information on the data source(s), main qualitative properties of the rock, and compositional data.

Information on the data sources consists of bibliographical references of a maximum of 5 items (books, reports, databases, etc.) where the chemical data of a sample were published. Why more than one source? Because a lot of samples were presented in several places, and sometimes a later paper gives more chemical components, or corrected values, or more background information for the same sample. In such cases, the better reference is taken as source even if it is later, but there are also samples the attributes of which were collected from several sources.

For a short qualitative characterization of the sampled rocks, the following pieces of information are registered: the area (a more or less standard name of the geographic-geologic unit) where the sampling site is located, "exact" location of the sample (including geodetic co-ordinates for boreholes), sampled level within the rock body if provided in the source, name of the rock type given by the source, a more standardized rock name, type of alteration if any, age of the rock as it stands in the source, a revised age according to recent stratigraphic interpretations, and the currently used name of the lithostratigraphic unit (geologic formation).

The list of major components analyzed and used in different publications is not absolutely standard which is especially true for the earlier sources. For example, components like SO_3, Ba, Sr, S, Cl, etc. are rarely present. Although they are all included in the database, the basic set of components comprises the most widely used major oxides: SiO_2, TiO_2, Al_2O_3, Fe_2O_3, FeO, MnO, MgO, CaO, Na_2O, K_2O, P_2O_5, H_2O^+, H_2O^-, and CO_2. All other, non-volatile components are combined under the name "others", both in the database and throughout this paper. Even some of these components are not given in all sources, and in a number of samples they may have "values below detection limit", or be given simply as 0.00% which is neither possible (in principle) nor favorable for statistics. All this information is maintained in the database. Mainly in earlier analyses, for Fe total-iron is given (also in not a standard form, i.e. either as Fe_2O_3 or as FeO); therefore (a uniform) total-iron has been calculated for all of the samples and is considered as a basic component when Fe^{2+} and Fe^{3+} are not determined.

In the modern database management programs the structure, indexation, field names, etc. of a database can easily be changed. The current structure of the given database is represented by a list of fields and their main attributes (Table 2). Hopefully, most of the names are self-explanatory and give an idea about the organization of information stored. Any usual database-handling function (search according to criteria, selecting records/fields and creating subsets, combining fields, sorting, etc.) may obviously be performed. Due to its simple structure and universal format the database can directly be accepted by most statistical, spreadsheet, etc. programs.

Frequencies by area vs. formation

The function of the database is to maintain data on the Cenozoic volcanites located in Hungary. A number of processed sources, however, contain data from those parts of the volcanic complexes that are actually outside the state limits. These analyses were also entered into the database, yielding a total of more than 3000 of which 2952 originate in Hungary.

In order to characterize the geographical and lithostratigraphic distribution of the samples in the database, these 2952 samples are broken down by area vs. formation and their counts are cross-tabulated in Table 1. Several samples could not be classified with certainty into the formations (grouped in "unassigned..." sets in Table 1). On the other hand, some of the formations (e.g. Kecel, Felső-nyárád, Sajóhídvég) are represented by a small number of analyses. Any statistics of these formations can only be treated as rough estimations. In the framework of the current project new samples have been collected and sent for analysis from several of these obviously under-represented formations. As soon as the analyses are ready they will be published and incorporated into the database.

Univariate statistics

In this section some basic descriptive statistics are given in order to characterize the representativeness of the database, and also to provide petrochemical reference statistics for further studies. As the database covers all sets of the known Cenozoic volcanites, at this stage it is attempted to treat the data together as much as possible. In this regard a major concern is how to assure a consistent data assemblage, which is not obvious because the quality and structure of analyses are slightly (sometimes notably) different. In most of the cases, the only way of checking the quality is to calculate the sum of component values. This cannot help when one or two major components are omitted in the source which is, luckily, not a frequent case. In this paper such samples are not taken into consideration. As to the criterion based on the sum, in the petrological literature the interval 99–101% is frequently used, but is still subjective. For the descriptive purposes here a less severe (and also subjective) condition, namely the range

Table 2
Fields and their main attributes in the database

Field Name	Type	Width	Dec	Field Name	Type	Width	Dec
CODE	Character	8		CO2_ORIG	Numeric	6	2
LABEL	Character	9		SO3_ORIG	Numeric	6	2
REF1	Character	30		LOI_ORIG	Numeric	6	2
PAGE1	Numeric	5		OTHER_ORIG	Numeric	6	2
TABLE1	Character	4		WHAT_OTHER	Character	48	
NUMBER1	Numeric	4		S_ORIG	Numeric	6	2
REF2	Character	30		CL_ORIG	Numeric	6	2
PP_TBL_NO2	Character	10		F_ORIG	Numeric	6	2
REF3	Character	30		O_MINUS_OR	Numeric	6	2
PP_TBL_NO3	Character	10		TOTAL_ORIG	Numeric	7	2
REF4	Character	30		TOTAL14_OX	Numeric	7	2
PP_TBL_NO4	Character	10		TOTALVOLAT	Numeric	6	2
REF5	Character	30		YR_ANALYS	Character	7	
PP_TBL_NO5	Character	10		SUMCALC	Numeric	7	2
SOURCE_REF	Character	30		SUMCTRL	Character	3	
SAMPLE_NO	Character	15		REMARKS	Character	75	
AREA	Character	23		SIO2	Numeric	6	2
ROCK_ORIG	Character	60		TIO2	Numeric	6	2
ROCKTYPE	Character	25		AL2O3	Numeric	6	2
ALTERATION	Character	15		FE2O3	Numeric	6	2
LEVEL	Character	3		FEO	Numeric	6	2
SAMPLESITE	Character	70		FE2O3_TOT	Numeric	6	2
FORMATION	Character	30		MNO	Numeric	6	2
AGE_ORIG	Character	30		MGO	Numeric	6	2
AGE	Character	20		CAO	Numeric	6	2
SIO2_ORIG	Numeric	6	2	NA2O	Numeric	6	2
TIO2_ORIG	Numeric	6	2	K2O	Numeric	6	2
AL2O3_ORIG	Numeric	6	2	P2O5	Numeric	6	2
FE2O3_ORIG	Numeric	6	2	OTHER	Numeric	6	2
FEO_ORIG	Numeric	6	2	TOTAL_100	Numeric	7	2
FE2O3_T_OR	Numeric	6	2	H2O_P	Numeric	6	2
MNO_ORIG	Numeric	6	2	H2O_M	Numeric	6	2
MGO_ORIG	Numeric	6	2	CO2	Numeric	6	2
CAO_ORIG	Numeric	6	2	LOI	Numeric	6	2
NA2O_ORIG	Numeric	6	2	FEO_TOT	Numeric	6	2
K2O_ORIG	Numeric	6	2	BOREHOLE	Character	1	
P2O5_ORIG	Numeric	6	2	X_KOO	Numeric	10	2
H2O_P_ORIG	Numeric	6	2	Y_KOO	Numeric	10	2
H2O_M_ORIG	Numeric	6	2	Z_KOO	Numeric	6	2
H2O_T_ORIG	Numeric	6	2				

98–102%, is used. As a result, 2795 analyses are labeled "good quality" and marked out for further analysis.

Recalculation of component values

As the set of major components determined has not been absolutely standard, even the "good quality" samples are not directly comparable. The problem of iron can be circumvented, to a certain extent, by computing total iron expressed in a unique form (here as $Fe_2O_3^{total}$) for all samples, but this also means losing part of information, namely the relative abundance of Fe^{2+} and Fe^{3+}. Obviously, when investigating processes that should be reflected in the ratio of Fe^{2+} and Fe^{3+} this is not acceptable. Here, descriptive statistics for all three forms of iron are calculated.

Similar, but a little more complex, is the problem of volatile components. The final common denominator might be loss-on-ignition (LOI), although the ignition procedure is not standardized among laboratories. But even LOI is not unambiguous, as sometimes H_2O^- – water released on heating up to 105 (or 110) °C – is included (although usually not), but sometimes this remains unclear to the reader. Furthermore, in the literature, unfortunately, H_2O^- is often used as a regular component in spite of the fact that it is nothing else but the dampness of the sample. The only appropriate way of using H_2O^- would have been to recalculate all analyses to 100% on a H_2O^--free basis, before interpretation and publication.

There are cases when volatiles are just not given and not mentioned. If, in addition, the sum of other components is not far from 100%, one cannot be sure if they have been recalculated in some way, or whether the sample was essentially volatiles-free. In other cases, beside LOI, one of the volatiles is also given. In such a situation one can only guess what the other volatiles could be, not to mention such embarrassing cases when the only specified volatile has a higher (!) value than total LOI. In the database, for each sample it was attempted to calculate as many volatiles as possible, from the information given in the source data. However, in order to gain statistics for as many samples as possible, all analyses were recalculated also on a LOI-free (i.e. volatiles-free) basis, thus assuring comparability of the other major components.

Separating atypical samples

It is clear that for different applications different atypicality conditions might be defined. Although there are a number of techniques suggested and traditions followed in this respect, they can never be absolutely objective. Nevertheless, it is widely accepted to judge the freshness of the sample by checking if H_2O^+ is less than 2% and/or CO_2 is less than 0.5% (e.g. Le Maitre 1984), which is important when dealing with the primary petrology of the rocks. Here, representative statistics of the volcanic complexes are searched for, hence altered (but not totally) variants of the rocks should also be considered. For this reason, petrological assessment is coupled with statistical reasoning, and deliberately tolerant thresholds of atypicality are defined. They are based on a joint analysis of the percentiles (Table 3) and histograms (Fig. 2) of the components, and it was attempted to set them close to the 99th percentiles often used by statisticians, but they are also adjusted to be petrochemically reasonable. The accepted atypicality thresholds are still subjective but, hopefully, not meaningless.

Of course, one could also suggest lower thresholds for atypicality, based straight on data from Table 3. This practice is not common in the literature; however, it can be defended by pointing out that extremely low values of one component are usually accompanied by elevated values of other components. As this is not absolute, using lower thresholds of atypicality should be encouraged in many applications. Here, the most significant lower threshold could be set up for SiO_2, but in fact, it (namely 40%) has already been considered in the process of data collection. Therefore, no (other) lower thresholds are defined for the statistical descriptions below. Also, for the traditionally main types of rocks (acid, intermediate, basic) one could set up different thresholds in case of some components, but it has been found that the above-defined general criteria used simultaneously perform well enough. Applying the upper thresholds marked in Table 3, all samples that have a higher value in any of the components have been excluded from further analysis, which has resulted in 2560 "typical" (and "good quality") samples. At the same time, the filtered-out samples have been found to be better investigated separately (in fact, they are too much altered).

Descriptive statistics

From the database, univariate statistics can easily be computed for any subsets of the samples (e.g. samples clustered by age, or area, or the SiO_2 content, etc.). Here, statistics are given for the geologic formations concerned, as they are more and more often referred to in comparative studies. From among the great number of descriptive statistics offered by statisticians, petrochemists most frequently use those characterizing the central tendency and the spread of component values. Given the largely varying number of samples in the formations (see Table 1) and the frequent bimodal (see below) or asymmetrical empirical distributions of component values, only robust estimations are worth considering.

First, the minima, maxima and medians are tabulated (Table 4). These values are thought to be characteristic statistics of the geologic formations concerned, and it should be emphasized that they are based on a number of analyses assumed to be close to the total number of analyses ever made. On the other hand, it can be seen that the component "others" behaves in a rather erratic manner: it is present in only a small number of formations, and it has large values

Table 3
Percentiles of components in the set of "good quality" samples, and the accepted atypicality thresholds

		SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	Fe_2O_3tot	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	others	H_2O^+	CO_2	LOI
		wt%															
	Minimum	41.40	0.01	0.87	0.01	0.01	0.07	0.01	0.01	0.03	0.01	0.04	0.01	0.01	0.03	0.01	0.04
Percentiles	0.10	42.06	0.01	1.39	0.01	0.02	0.11	0.01	0.01	0.07	0.03	0.04	0.01	0.01	0.03	0.01	0.08
	0.20	43.37	0.01	2.84	0.01	0.03	0.17	0.01	0.01	0.09	0.05	0.06	0.01	0.01	0.05	0.01	0.16
	0.30	44.30	0.01	3.74	0.01	0.03	0.22	0.01	0.01	0.10	0.06	0.08	0.01	0.01	0.07	0.01	0.24
	0.40	44.67	0.01	4.16	0.02	0.03	0.26	0.01	0.01	0.11	0.07	0.09	0.01	0.01	0.10	0.01	0.30
	0.50	44.76	0.01	4.58	0.03	0.04	0.28	0.01	0.01	0.13	0.08	0.12	0.01	0.01	0.14	0.01	0.33
	1.00	45.60	0.03	5.51	0.09	0.07	0.35	0.01	0.03	0.19	0.10	0.20	0.01	0.01	0.23	0.02	0.48
	2.00	46.07	0.04	7.24	0.15	0.10	0.58	0.01	0.06	0.27	0.15	0.32	0.01	0.02	0.35	0.02	0.64
	3.00	46.36	0.05	8.32	0.19	0.13	0.73	0.02	0.07	0.35	0.19	0.44	0.01	0.02	0.43	0.02	0.76
	4.00	46.54	0.07	9.42	0.22	0.16	0.93	0.02	0.10	0.44	0.24	0.53	0.02	0.03	0.51	0.03	0.86
	5.00	46.78	0.08	10.77	0.25	0.20	1.03	0.02	0.13	0.57	0.28	0.63	0.02	0.03	0.58	0.03	0.94
	10.00	48.39	0.16	12.91	0.46	0.34	1.39	0.03	0.28	0.98	0.70	1.04	0.03	0.05	0.82	0.05	1.39
	20.00	53.42	0.31	14.35	0.89	0.64	2.30	0.05	0.63	1.96	1.68	1.55	0.07	0.11	1.15	0.10	1.96
	30.00	56.79	0.45	15.51	1.36	1.10	3.80	0.08	1.18	3.32	2.15	1.85	0.10	0.23	1.45	0.16	2.47
	40.00	58.95	0.57	16.17	1.87	1.73	5.17	0.10	1.94	4.82	2.42	2.05	0.13	0.83	1.81	0.27	2.98
	50.00	61.15	0.67	16.83	2.35	2.47	6.06	0.12	2.54	6.03	2.66	2.28	0.16	2.60	2.16	0.45	3.63
	60.00	63.28	0.80	17.50	2.89	3.19	6.91	0.14	3.08	6.94	2.91	2.54	0.19	3.73	2.63	0.76	4.46
	70.00	67.14	0.97	18.19	3.49	4.03	7.94	0.16	3.72	7.80	3.15	2.95	0.23	6.89	3.29	1.30	5.77
	80.00	71.99	1.31	18.94	4.20	4.93	9.24	0.18	5.01	8.68	3.42	3.83	0.36	9.53	4.23	2.33	7.79
	90.00	75.78	2.08	20.16	5.62	6.19	10.46	0.23	7.72	9.56	3.91	4.71	0.69	13.25	5.97	4.23	11.07
	95.00	76.93	2.29	21.69	6.73	6.92	10.90	0.31	8.43	10.16	4.38	5.75	0.80	16.72	7.56	6.78	14.04
	96.00	77.48	2.32	22.14	7.05	7.09	11.09	0.33	8.66	10.39	4.51	6.53	0.83	17.29	7.96	7.39	14.99
	97.00	78.16	2.40	22.64	7.49	7.26	11.33	0.39	8.97	10.74	4.71	7.84	0.87	18.32	8.45	8.31	16.59
	98.00	79.54	2.47	23.86	8.18	7.45	11.70	0.47	9.49	11.85	4.96	9.13	0.92	21.76	**9.53**	**9.45**	**19.02**
	99.00	**81.44**	**2.62**	**28.51**	**9.51**	7.84	13.17	0.59	10.32	13.47	5.52	10.28	1.12	**23.50**	13.37	12.72	22.29
	99.50	83.52	3.06	39.00	10.67	8.06	14.48	0.98	11.98	14.85	6.09	11.23	1.28	38.75	15.86	14.63	24.17
	99.60	84.13	3.16	40.11	11.11	**8.11**	**15.14**	1.10	12.85	16.55	6.15	11.60	1.36	41.35	16.09	15.32	24.66
	99.70	85.98	3.39	40.40	11.38	8.62	15.98	**1.35**	13.52	**18.72**	6.32	11.96	1.58	43.78	17.13	15.88	25.62
	99.80	86.52	3.65	40.96	16.21	9.20	18.52	2.35	13.84	21.46	**6.62**	**12.17**	1.75	43.78	18.41	17.88	28.23
	99.90	87.63	3.77	42.21	23.44	10.37	23.24	2.71	**14.78**	23.24	7.42	13.18	**1.95**	43.78	23.32	19.13	34.98
	100.00	96.74	4.52	45.83	27.92	11.08	30.92	7.10	40.69	29.34	8.39	13.53	7.15	43.78	24.00	19.83	48.65
N	Valid	2795	2664	2795	2631	2611	2794	2534	2730	2791	2782	2793	2520	325	2412	1515	2722
	Missing	0	131	0	164	184	1	261	65	4	13	2	275	2470	383	1280	73

Explanation see in section "Separating atypical samples".

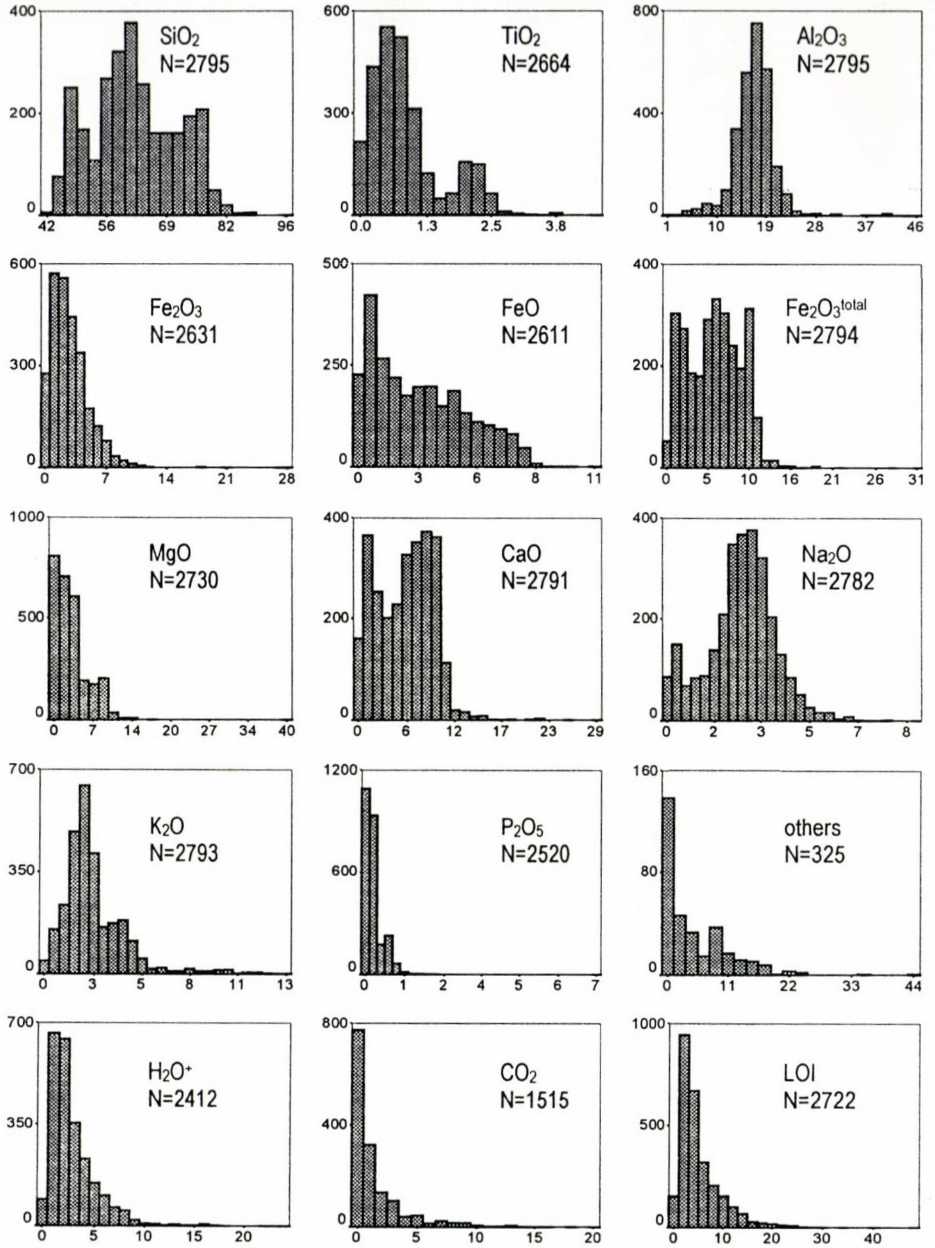

Fig. 2
Histograms of components in the set of "good quality" samples. N – number of samples for which a component was given or could be calculated. For more explanation see section "Univariate statistics".

Table 4
Medians and ranges of components in the studied formations

FORMATION	SiO_2				TiO_2				Al_2O_3				Fe_2O_3			
	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N
Balatonmária basalt	52.07	53.71	56.97	7	0.98	1.03	1.22	7	13.80	15.23	15.55	7	.	.	.	0
Bár Basalt F.	48.38	48.92	52.81	9	1.18	2.08	2.41	9	14.31	15.05	17.13	9	1.30	3.72	6.77	9
Börzsöny Volcanite (F.)	45.69	58.54	71.32	166	0.04	0.78	1.26	154	13.56	18.71	22.88	166	0.29	3.68	8.08	146
Cserehát F.	70.08	72.72	73.93	8	0.14	0.46	0.53	8	13.01	14.50	15.44	8	0.19	0.57	4.26	8
Felsőnyárád F.	58.79	60.72	62.65	2	0.42	0.47	0.52	2	22.02	24.96	27.89	2	3.27	4.66	6.05	2
Galgavölgy Rhyolite Tuff F.	67.88	74.73	76.78	21	0.07	0.24	0.60	20	12.85	14.01	21.68	21	0.41	1.36	5.31	21
Gyulakeszi Rhyolite Tuff F.	62.30	73.55	77.92	56	0.02	0.23	0.70	56	11.71	14.88	19.22	56	0.10	1.19	4.61	56
Hasznos Andesite F.	50.33	58.35	65.89	25	0.53	0.94	1.44	25	16.21	19.20	23.30	25	0.13	3.69	6.79	24
Kecel Basalt F.	44.76	50.83	60.34	4	0.68	2.22	2.51	4	15.19	16.55	18.03	4	2.01	5.68	7.53	4
Mátra Andesite F.	49.39	57.72	76.68	399	0.09	0.84	1.98	396	8.57	18.46	28.01	399	0.27	2.85	9.03	390
Mecsek Andesite F.	59.47	62.38	67.40	20	0.10	0.67	1.80	20	15.94	18.20	20.34	20	1.72	3.31	6.49	20
Nadap Andesite F.	41.85	60.17	78.35	184	0.12	0.73	2.22	182	10.24	18.12	25.28	184	0.20	3.29	9.04	177
Nyírség Volcanite (F.)	61.66	72.17	80.12	32	0.08	0.28	1.17	27	12.47	15.14	17.06	32	0.20	1.27	4.06	32
Pásztori Trachyte F.	50.83	61.46	65.89	6	0.70	1.00	2.43	6	17.03	18.37	20.51	6	0.34	1.89	5.68	6
Recsk Andesite F.	46.88	64.52	81.17	243	0.04	0.44	2.15	243	1.47	14.07	25.06	243	0.01	0.53	8.05	240
Sajóhidvég Trachyte (F.)	67.44	69.15	70.85	2	0.45	0.47	0.48	2	14.37	15.42	16.46	2	0.75	1.75	2.74	2
Sajóvölgy F.	57.86	60.80	65.59	6	0.54	0.66	0.80	5	17.35	18.74	20.41	6	3.48	4.25	5.30	6
Salgóvár Basalt F.	44.19	47.22	52.23	66	0.69	1.96	2.45	66	13.71	17.65	20.79	66	1.66	3.83	8.18	55
Sárospatak basalt	48.40	49.66	51.26	9	0.63	1.25	1.48	9	16.52	17.55	19.08	9	2.22	2.83	5.29	8
Szentmihály Andesite F.	53.72	61.95	70.05	5	0.42	0.56	0.85	5	16.30	17.94	19.28	5	0.79	2.25	4.00	5
Tapolca Basalt F.	42.11	47.91	55.93	335	0.58	2.14	2.61	334	12.12	15.92	21.57	335	0.19	3.59	9.43	270
Tar Dacite Tuff F.	58.27	69.96	79.38	179	0.02	0.30	1.79	172	11.38	16.04	21.60	179	0.05	1.54	8.82	178
Tokaj Volcanite F.	42.29	67.03	81.44	748	0.01	0.52	1.73	661	8.60	16.17	28.29	748	0.02	1.66	8.97	734

FORMATION	FeO				$Fe_2O_3^{total}$				MnO				MgO			
	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N
Balatonmária basalt	.	.	.	0	7.35	8.22	9.50	7	0.11	0.15	0.16	7	3.62	5.47	6.30	7
Bár Basalt F.	2.89	4.18	6.76	9	7.21	8.79	10.42	9	0.08	0.13	0.16	9	6.83	7.83	8.56	9
Börzsöny Volcanite (F.)	0.04	3.17	7.43	146	2.62	7.37	12.95	166	0.01	0.14	0.99	153	0.08	2.06	6.34	166
Cserehát F.	0.25	1.64	5.33	8	1.82	3.15	7.14	8	0.08	0.13	0.20	8	0.21	0.29	0.94	8
Felsőnyárád F.	0.84	0.86	0.88	2	4.25	5.62	6.98	2	0.02	0.03	0.03	2	0.94	1.27	1.60	2
Galgavölgy Rhyolite Tuff F.	0.14	0.61	2.49	21	0.79	1.76	5.98	21	0.02	0.05	0.72	19	0.06	0.55	2.93	21
Gyulakeszi Rhyolite Tuff F.	0.19	0.83	2.05	56	0.61	2.33	6.44	56	0.01	0.05	0.46	50	0.10	0.46	3.00	55
Hasznos Andesite F.	0.99	2.99	6.54	24	4.45	7.40	9.31	25	0.03	0.13	0.36	23	0.54	2.14	3.50	25
Kecel Basalt F.	2.60	3.48	3.67	4	6.08	9.55	10.42	4	0.13	0.18	0.27	4	1.78	5.60	9.07	4
Mátra Andesite F.	0.03	4.17	7.81	390	0.88	7.62	12.12	399	0.01	0.14	1.05	381	0.06	2.42	12.29	397
Mecsek Andesite F.	0.10	1.74	3.57	20	2.83	5.59	6.62	20	0.03	0.10	0.23	17	1.10	3.03	5.02	20
Nadap Andesite F.	0.09	2.80	7.17	177	1.15	6.73	14.47	184	0.01	0.15	1.15	180	0.04	3.14	8.84	183
Nyírség Volcanite (F.)	0.06	1.18	4.70	32	0.30	2.70	8.92	32	0.01	0.06	0.23	24	0.07	0.64	2.48	31
Pásztori Trachyte F.	2.31	3.57	4.69	6	4.66	5.17	10.33	6	0.14	0.17	0.19	6	0.23	0.45	4.03	6
Recsk Andesite F.	0.01	1.75	7.84	241	0.07	3.35	10.61	243	0.01	0.06	0.51	235	0.26	3.01	10.17	243
Sajóhidvég Trachyte (F.)	0.63	1.70	2.77	2	3.44	3.63	3.82	2	0.01	0.04	0.06	2	0.19	0.29	0.38	2
Sajóvölgy F.	0.37	2.41	3.26	6	3.89	7.50	8.04	6	0.07	0.11	0.27	4	0.77	2.01	2.60	6
Salgóvár Basalt F.	1.34	5.93	7.49	55	8.15	9.91	13.18	66	0.06	0.18	0.42	66	2.79	6.34	10.11	66
Sárospatak basalt	2.21	4.53	5.60	8	7.63	8.07	8.87	9	0.12	0.18	0.23	9	6.51	9.81	10.58	9
Szentmihály Andesite F.	2.37	4.34	4.86	5	4.88	6.18	8.86	5	0.05	0.16	0.21	5	1.83	2.49	8.05	5
Tapolca Basalt F.	1.39	6.44	8.08	270	7.25	10.49	14.03	335	0.03	0.17	0.59	326	3.77	8.10	13.96	335
Tar Dacite Tuff F.	0.03	1.20	5.28	178	0.27	2.86	10.84	179	0.01	0.07	0.29	163	0.01	1.17	4.33	174
Tokaj Volcanite F.	0.01	1.47	7.51	720	0.19	4.48	13.08	747	0.01	0.10	1.31	616	0.01	1.18	6.97	711

Explanation see in section "Descriptive statistics".

Table 4
(cont.)

FORMATION	CaO				Na_2O				K_2O				P_2O_5				others				H_2O^+				CO_2				LOI			
	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N	Minimum	Median	Maximum	Valid N
Balatonmária basalt	7.60	8.19	9.08	7	1.75	2.01	2.50	7	4.58	4.97	5.47	7	0.75	0.88	1.30	7	.	.	.	0	.	.	.	0	.	.	.	0	0.67	0.99	2.29	7
Bár Basalt F.	5.61	7.28	8.95	9	2.21	3.30	4.78	9	2.08	5.46	6.71	9	0.76	0.98	1.51	9	.	.	.	0	0.65	1.29	3.71	8	0.03	0.23	0.81	6	0.83	1.58	4.99	9
Börzsöny Volcanite (F.)	0.94	6.71	14.72	166	1.29	3.08	6.45	166	0.72	2.11	6.54	166	0.02	0.18	0.74	131	0.12	0.22	1.36	7	0.21	1.55	5.31	142	0.02	0.41	8.53	91	0.23	3.00	13.32	164
Cserehát F.	1.57	1.97	2.49	8	1.68	2.75	3.19	8	2.58	3.94	5.41	8	0.02	0.05	0.06	8	.	.	.	0	3.21	4.74	7.35	8	0.22	0.55	3.42	8	4.80	7.21	10.66	8
Felsőnyárád F.	2.96	3.41	3.85	2	2.84	2.89	2.94	2	0.58	0.73	0.87	2	0.01	0.02	0.03	2	.	.	.	0	5.20	5.65	6.09	2	0.33	0.37	0.41	2	10.43	11.09	11.75	2
Galgavölgy Rhyolite Tuff F.	0.36	1.95	7.54	21	0.67	1.93	3.54	21	1.31	3.72	7.48	21	0.02	0.06	0.23	19	0.10	0.10	0.10	1	1.10	5.51	8.19	16	0.07	0.18	2.73	11	0.91	6.44	15.11	21
Gyulakeszi Rhyolite Tuff F.	0.87	2.44	6.56	56	0.24	2.17	3.66	56	1.15	3.33	7.50	56	0.01	0.05	0.20	54	0.03	0.05	1.21	11	1.38	3.98	9.20	44	0.02	0.17	7.13	23	1.43	6.00	14.04	56
Hasznos Andesite F.	2.32	7.55	14.80	25	1.35	2.54	3.31	25	0.85	1.80	3.89	25	0.06	0.20	0.72	25	.	.	.	0	1.15	2.76	7.39	23	0.04	0.39	6.45	19	1.80	5.09	17.87	24
Kecel Basalt F.	5.12	8.32	16.56	4	0.25	3.14	4.88	4	0.95	1.35	7.84	4	0.13	0.28	0.47	4	.	.	.	0	3.79	5.97	9.32	4	0.08	3.55	7.28	4	10.96	14.14	17.64	4
Mátra Andesite F.	0.18	7.40	13.39	398	0.09	2.44	6.10	399	0.05	2.09	12.10	399	0.01	0.15	1.75	383	0.01	0.01	0.01	1	0.03	1.70	8.88	390	0.01	0.21	8.87	216	0.34	2.75	17.84	399
Mecsek Andesite F.	1.24	5.01	6.27	20	1.95	3.49	4.90	20	0.85	2.33	3.18	20	0.01	0.19	0.34	17	.	.	.	0	0.86	1.43	9.03	19	0.08	0.91	3.76	14	1.73	4.58	12.24	20
Nadap Andesite F.	0.03	6.31	12.41	184	0.08	2.76	6.09	180	0.09	1.86	5.64	184	0.01	0.18	0.84	177	0.03	0.41	12.35	31	0.03	3.00	8.88	141	0.03	0.92	6.93	148	0.60	5.91	16.90	181
Nyírség Volcanite (F.)	0.14	2.26	5.96	32	0.21	3.46	5.04	32	0.66	3.12	5.48	32	0.02	0.07	0.26	27	0.06	0.10	0.38	3	0.53	2.44	6.00	32	0.03	0.30	4.80	28	0.57	4.44	9.86	32
Pásztori Trachyte F.	0.77	1.64	8.56	6	4.59	6.08	6.27	6	1.82	4.27	4.91	6	0.08	0.18	0.52	6	.	.	.	0	1.05	1.24	1.29	3	2.61	3.07	3.86	3	3.71	4.31	5.94	6
Recsk Andesite F.	0.23	5.46	16.53	243	0.05	2.31	6.35	243	0.09	1.20	11.23	243	0.01	0.17	0.60	224	0.01	5.04	22.46	172	0.05	2.02	7.96	226	0.05	2.00	9.43	232	0.38	6.23	16.10	243
Sajóhídvég Trachyte (F.)	0.13	0.24	0.34	2	3.93	4.56	5.18	2	5.40	6.37	7.33	2	0.03	0.07	0.10	2	.	.	.	0	0.77	1.12	1.47	2	0.11	0.23	0.35	2	1.46	1.96	2.46	2
Sajóvölgy F.	5.46	6.27	8.79	6	2.05	2.30	2.99	6	1.54	1.70	2.57	6	0.06	0.13	0.23	5	.	.	.	0	0.25	2.79	5.69	6	0.07	0.26	1.87	5	1.80	5.37	10.62	6
Salgóvár Basalt F.	7.20	9.59	14.60	66	2.90	4.57	5.87	66	1.30	2.30	3.16	66	0.01	0.58	0.97	65	0.04	0.04	0.04	1	0.41	1.05	3.79	40	0.02	0.29	3.55	32	0.07	1.90	6.58	62
Sárospatak basalt	8.48	10.04	10.78	9	2.24	2.60	3.16	9	1.08	1.48	1.75	9	0.19	0.28	0.38	9	.	.	.	0	1.71	2.41	3.46	8	0.07	0.40	1.14	7	2.83	3.89	6.64	9
Szentmihály Andesite F.	1.26	4.56	9.74	5	0.89	1.64	2.61	5	1.58	2.39	3.48	5	0.10	0.12	0.14	5	0.03	0.03	0.03	1	1.02	2.14	6.07	5	0.15	1.53	2.39	4	1.82	5.09	8.12	5
Tapolca Basalt F.	7.73	9.17	18.68	335	1.41	3.50	6.09	335	0.55	2.14	3.24	335	0.01	0.71	1.75	329	0.12	0.14	0.23	5	0.32	1.89	8.35	168	0.01	0.23	8.14	120	0.52	2.34	14.81	274
Tar Dacite Tuff F.	0.13	2.92	9.04	179	0.17	2.10	4.71	179	0.85	3.64	12.17	179	0.01	0.06	0.78	164	0.02	0.03	0.09	4	0.29	2.83	7.81	156	0.02	0.78	6.74	82	0.50	5.65	12.99	179
Tokaj Volcanite F.	0.07	3.59	11.44	746	0.03	2.54	6.08	742	0.04	3.22	12.17	746	0.01	0.10	1.18	617	0.01	0.09	9.04	56	0.05	2.16	9.53	724	0.01	0.16	5.14	314	0.04	3.13	16.58	747

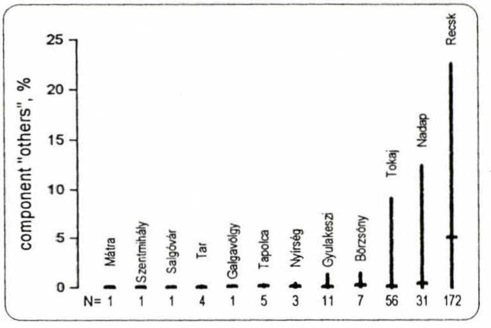

Fig. 3
Ranges and medians of component "others" in the studied formations. For explanation see section "Descriptive statistics".

only in three formations (Recsk, Nadap, and Tokaj). "Others" are mainly represented by FeS_2 (pyrite), which is a well-known postvolcanic phase in these formations. As this component is included in the LOI-free set, its high values may disturb comparison and interpretation of the other non-volatile components. Figure 3 suggests that with a little refining of the sample selection the (probably) problematic samples may be avoided, and, at the same time, the non-volatile components of the concerned formations still preserve clues of secondary alterations which might be desirable at the level of description. 5% is chosen for selection limit (it is undoubtedly subjective), resulting in the omission of a further 98 samples. Obviously and inevitably, in this respect, requirements of representativeness and comparability work against each other.

Even these simple and robust statistics should be referred to with caution, because the observed frequency distributions are often bimodal. The most spectacular cases are given in Fig. 4. Representation of such distributions with just one central tendency value is only justifiable if neglecting the heterogeneity does not confuse the given application. A special case is the Tokaj Volcanite Formation. Practically each of its components is characterized by two populations of values. This bimodality has long been known in the literature (e.g. Gyarmati 1961, 1977); nevertheless, dividing this formation into two (or more) petrochemical types should be a more common practice in related studies.

Comparison of the univariate statistics across formations can be done in many ways. In order to visualize the main similarities and differences, box plots of most components are displayed with formation medians in ascending order (Figs 5 and 6). It can be observed how poorly the formations separate in one dimension. Below, it will be shown that the multivariate space (expectedly) performs much better.

Bivariate statistics

TAS-diagram

In petrological applications bivariate data analysis is usually performed for descriptive/diagnostic purposes (see e.g. reviews of Rollinson 1993, or Rickwood 1989), or in order to reveal correlations of components. Sometimes these objectives mix, due to the fact that the ultimate cause of relations of both the

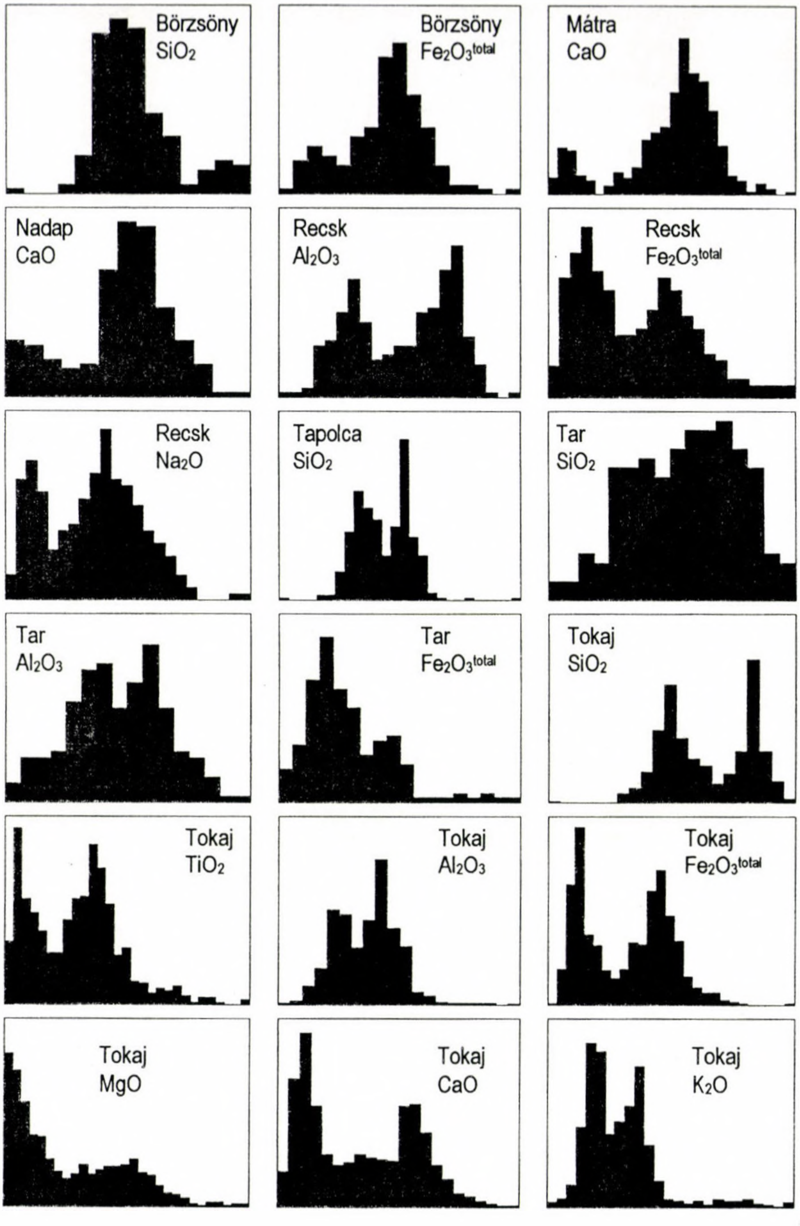

Fig. 4
Histograms revealing bimodality of components in the studied formations

samples and components is the same: the set of rock-forming processes. In the current study, which aims at presenting the given database, emphasis is laid on description. As there are about one and a half dozen components in the system, a great number of bivariate (sub)spaces exist. Probably the most often investigated is the TAS-plane, which was also suggested for diagnostics (Le Maitre 1984, adjusted by Le Bas et al. 1986). Every sample that could be assumed

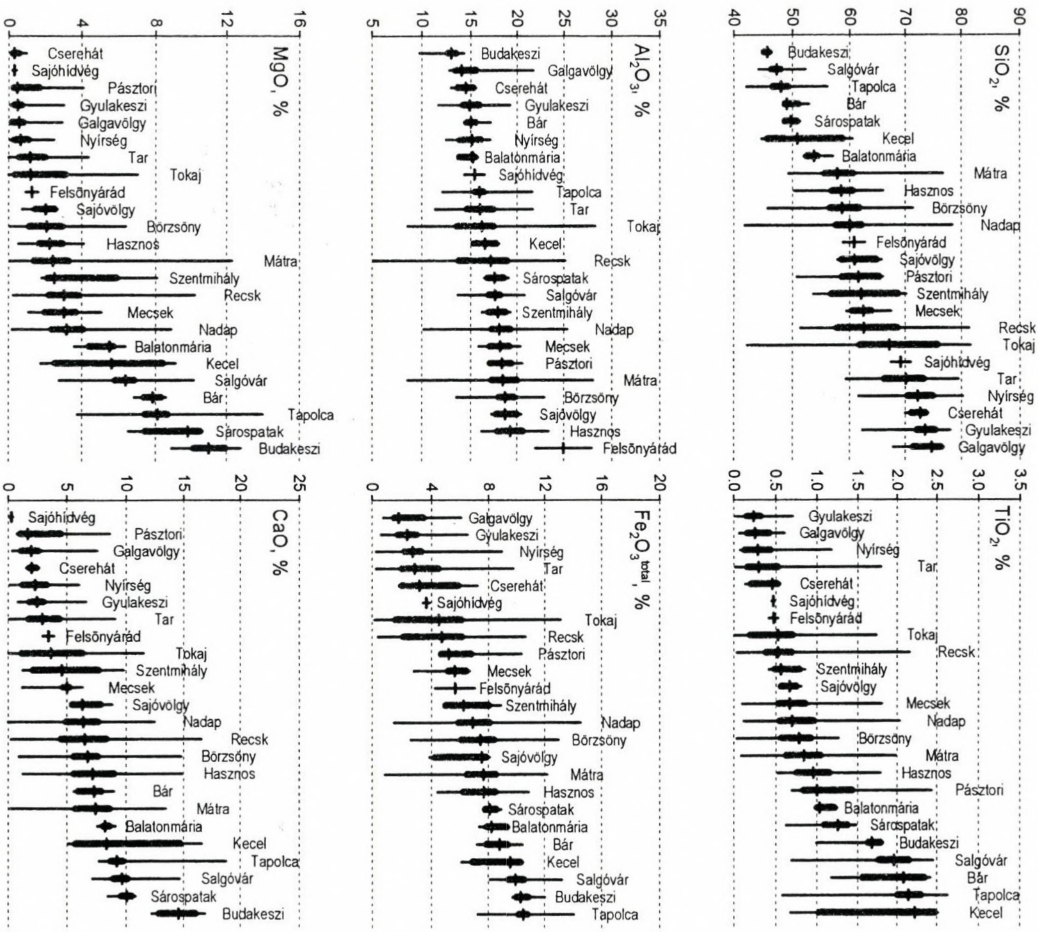

Fig. 5
Box plots of component values in the studied formations. Thin bars represent empirical ranges, thick bars spread from the 25th to the 75th percentile, the crossing strokes denote the medians.

to come from a fresh rock (964 pieces) is displayed on this plot (Fig. 7). It is remarkable that the given analyses spread to most of the fields of the diagram.

As SiO_2 and alkalis are good discriminators they are also suitable for visualizing an important part of the variability of samples. One would certainly be interested in the behavior of formations on that plot. In Fig. 8 the most numerous formations are presented using all their samples that are not atypical (but not necessarily fresh). In accordance with what is seen in Fig. 4 some of the

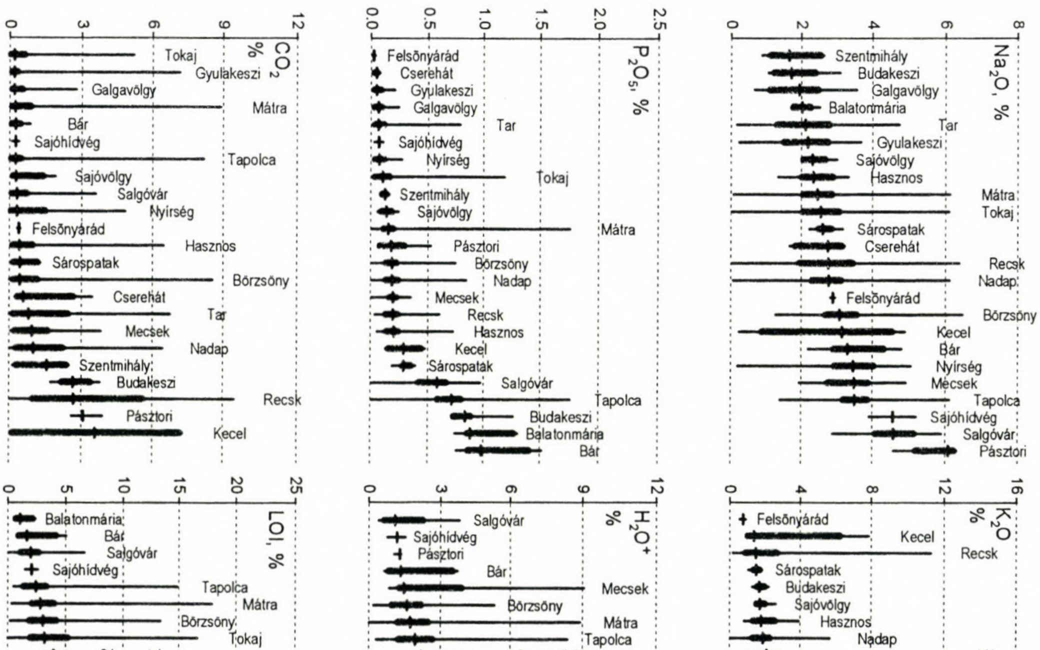

Fig. 6
Box plots of component values in the studied formations. Thin bars represent empirical ranges, thick bars spread from the 25th to the 75th percentile, the crossing strokes denote the medians.

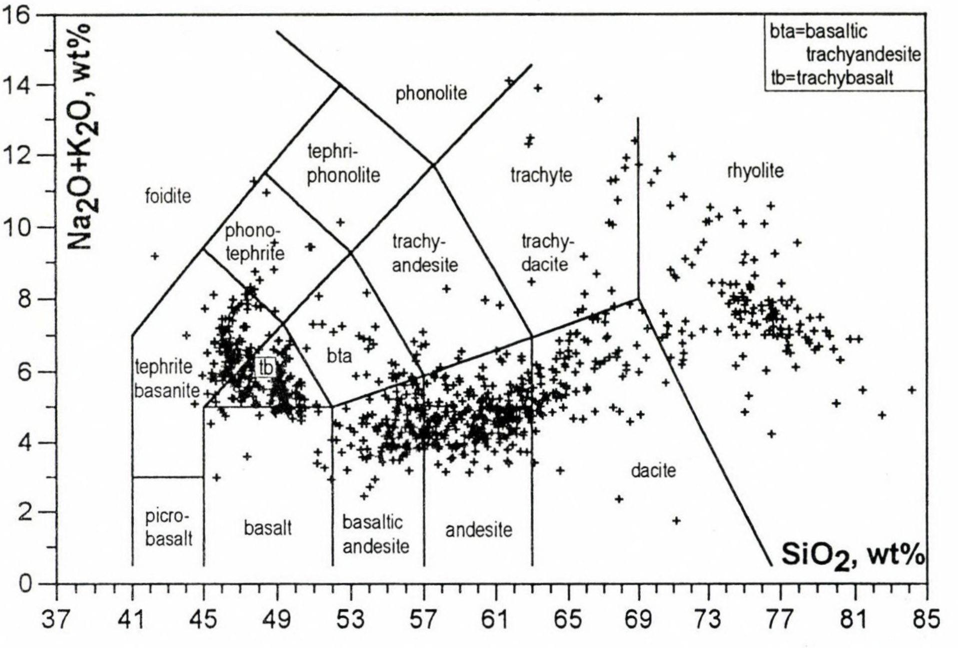

Fig. 7
TAS-diagram, as suggested by Le Bas et al. (1986), of the 964 fresh rock samples. The samples were selected according to criteria: $H_2O<2\%$, $CO_2<0.5\%$, and LOI<3%.

formations reveal bimodal distributions. This is expressed in this way in case of the Tokaj Volcanite and, somewhat surprisingly, the Tapolca Basalt Formations, so that it seems reasonable to divide them each into two sub-populations. For simplicity, they are cut apart at $SiO_2=70\%$ and 48.5%, respectively (values expressed on a volatiles-free basis), although in more sophisticated analyses a multivariate separation would be more elegant. After dividing, every component of Tokaj 1 and Tokaj 2 becomes unimodal (Fig. 9). The other, less remarkably bimodal (or multimodal) formations are also left undivided here for simplicity.

In Fig. 10 the medians of each formation (after the previous dividing) are plotted with all samples in the background. Although they cannot equally incorporate all parts of the point cloud, they are able to represent the majority of samples. Besides, as formations are complexly justified (or at least defined) rock assemblages, one may want to see the behavior of formation medians on other bivariate plots as well. A few of such plots are given in Figs 11 and 12 (purpose: diagnostics), and in Fig. 13 (description).

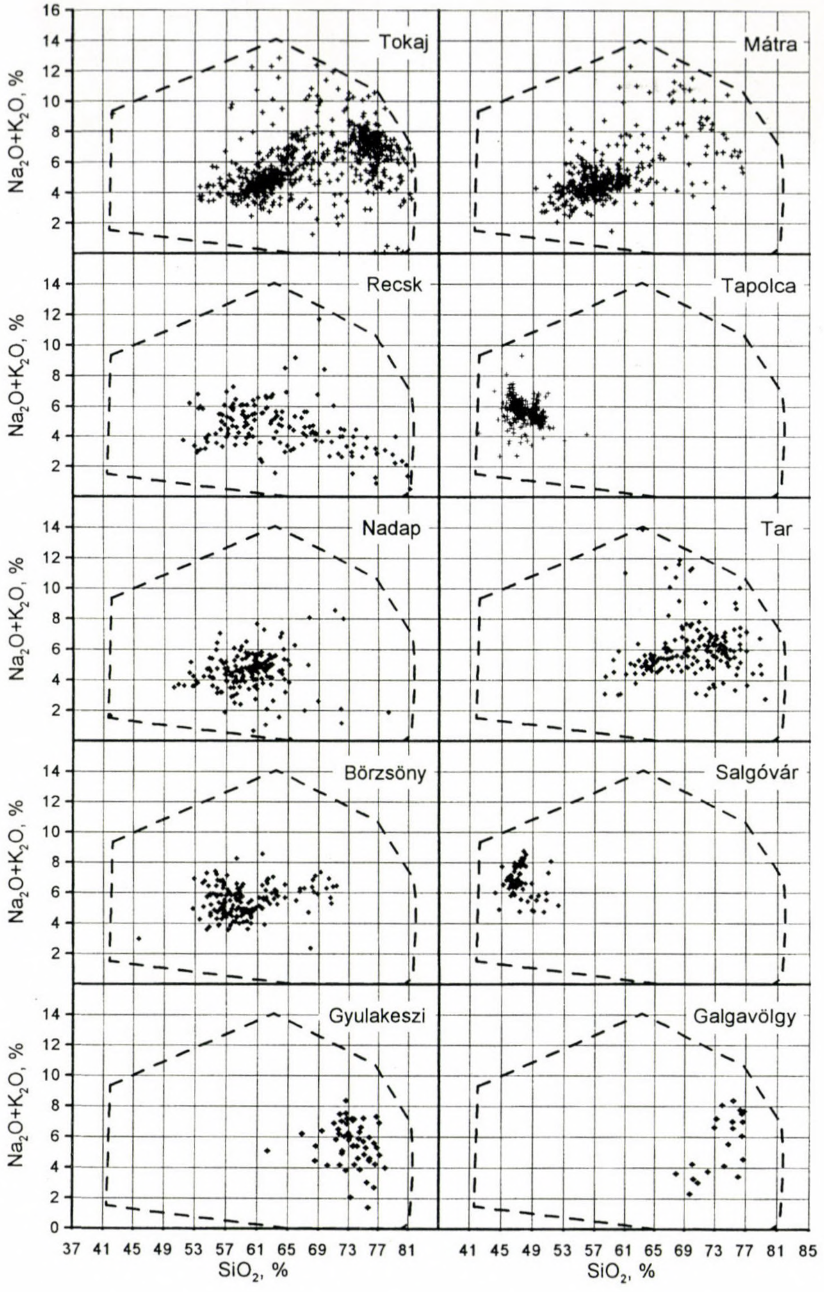

Fig. 8
Scatters of selected formations with a polygon enveloping all sample points on the TAS-plane. All "good quality" (not only fresh) samples are displayed.

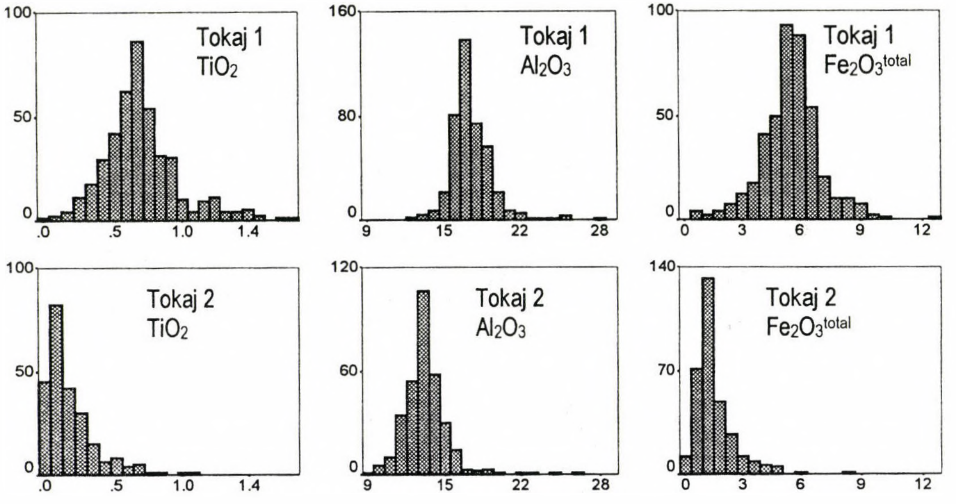

Fig. 9
Histograms of several components in the Tokaj Volcanite Formation divided at SiO_2=70% into Tokaj 1 and Tokaj 2. Bimodality of each component is related to the same subsets of samples. (Compare with Figs 4 and 8.)

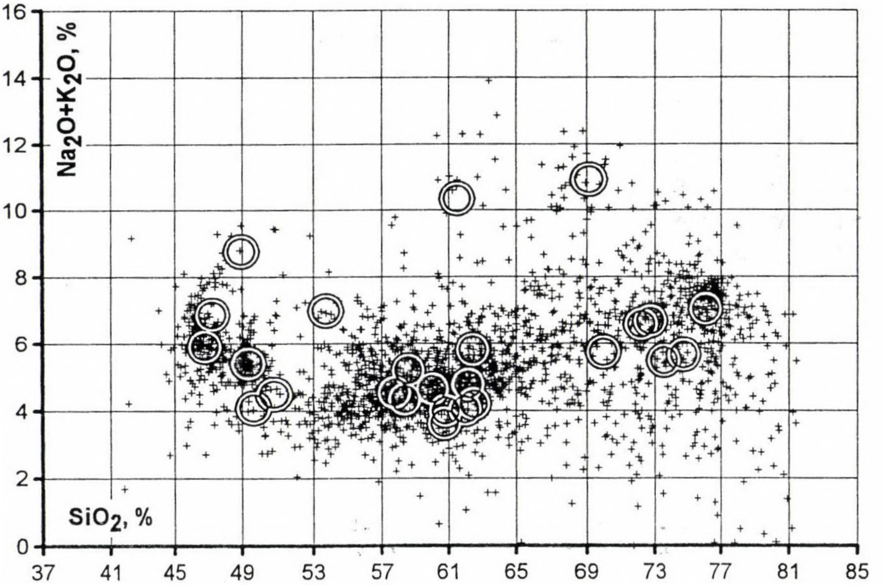

Fig. 10
TAS-diagram of all (2462) comparable samples and medians (circles) of the figuring formations. Tokaj Volcanite F. and Tapolca Basalt F. have been subdivided at SiO_2=70% and 48.5%, respectively. For more explanation see section "TAS-diagram".

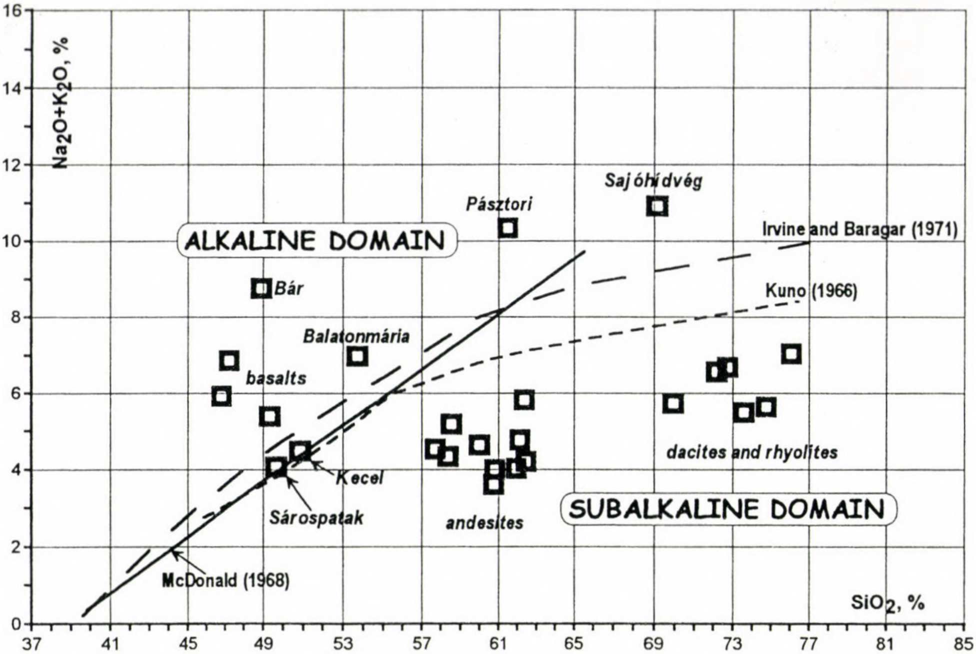

Fig. 11
Scatter of formation medians on a TAS-plot with the boundary lines, as suggested by different authors, between the alkaline and subalkaline series. Boundary lines' co-ordinates are taken from Rickwood (1989).

Correlations and the problem of closure

Some of the plots (f, g, l) in Fig. 13 show an unordered spread of formation centers (medians), others (b, h, k) allow seeing a link among the medians (and hence a correlation of components) in a part of formations, but several (a, c, d, e, i, j) display a linear or slightly curved pattern of points. What is the meaning of these regularities (and irregularities)? Since the works of Sarmanov and Vistelius (1959) and Chayes (1960) more and more attention has been paid to the problem of correlations in compositional (closed) data and the related problem of ratio correlation (Pearson 1896–1897; Chayes 1949), and a number of mathematical techniques have been suggested or tried out (e.g. Vistelius and Sarmanov 1961; Chayes and Kruskal 1966; Koch and Link 1971, pp. 153–183; Skala 1979; Drooger 1982; Aitchison 1986, etc.) to overcome the related evaluational uncertainties. There have been warnings that closure may distort even univariate distributions (Miesch 1969), and trends seen on ternary diagrams (Butler 1979). Of course, multivariate analysis may also suffer from the closure effect. The International Association for Mathematical Geology devoted an Annual Meeting to the problem (Pawlowsky-Glahn 1997). In short, the nature of the problem is very well known. The present paper does not aim at discussing any theoretical aspects of related phenomena, nor does it attempt to search for interpretational tools; this will be done elsewhere. However, a few remarks are made below in order to indicate the potential of the database in investigating bivariate and multivariate correlations.

In general, a frequently mentioned side-effect of the closure problem is that the relative values of components (e.g. expressed in percentages) that are known, do not provide enough information about the processes changing the absolute

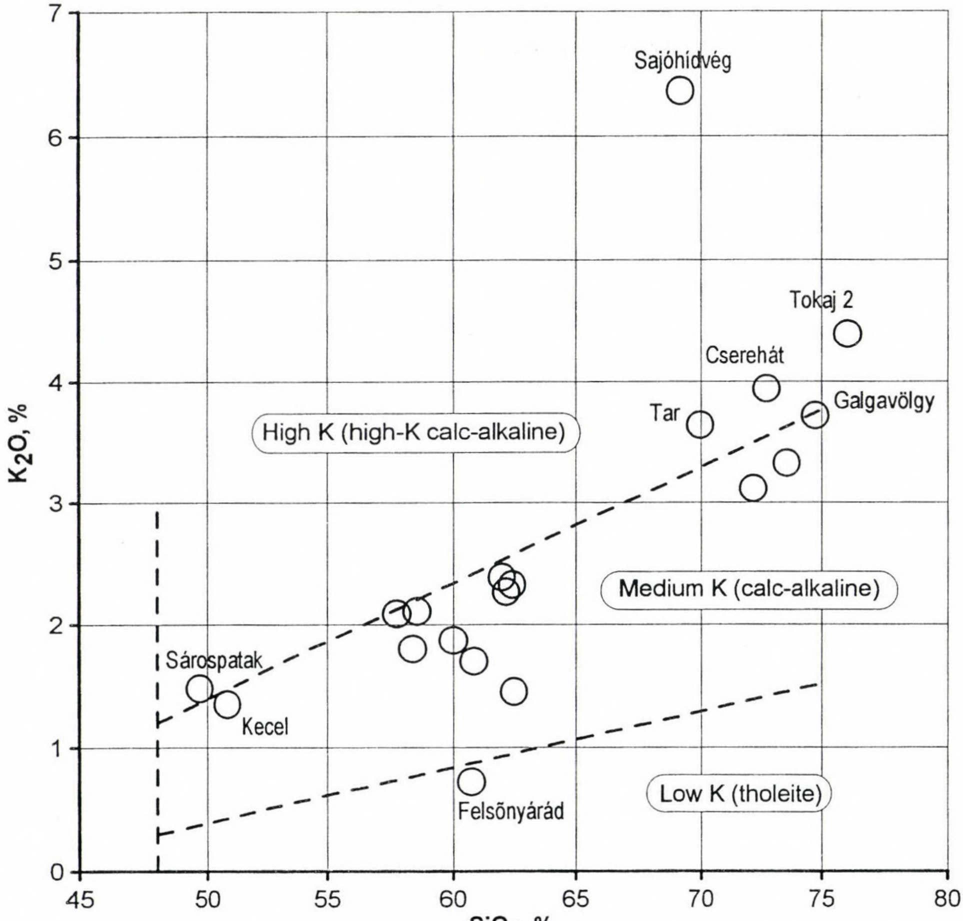

Fig. 12
Subdivision of the subalkaline formations (see Fig. 11) on a K_2O vs. SiO_2 plot. Boundary lines' coordinates are adapted from those given by Rollinson (1993).

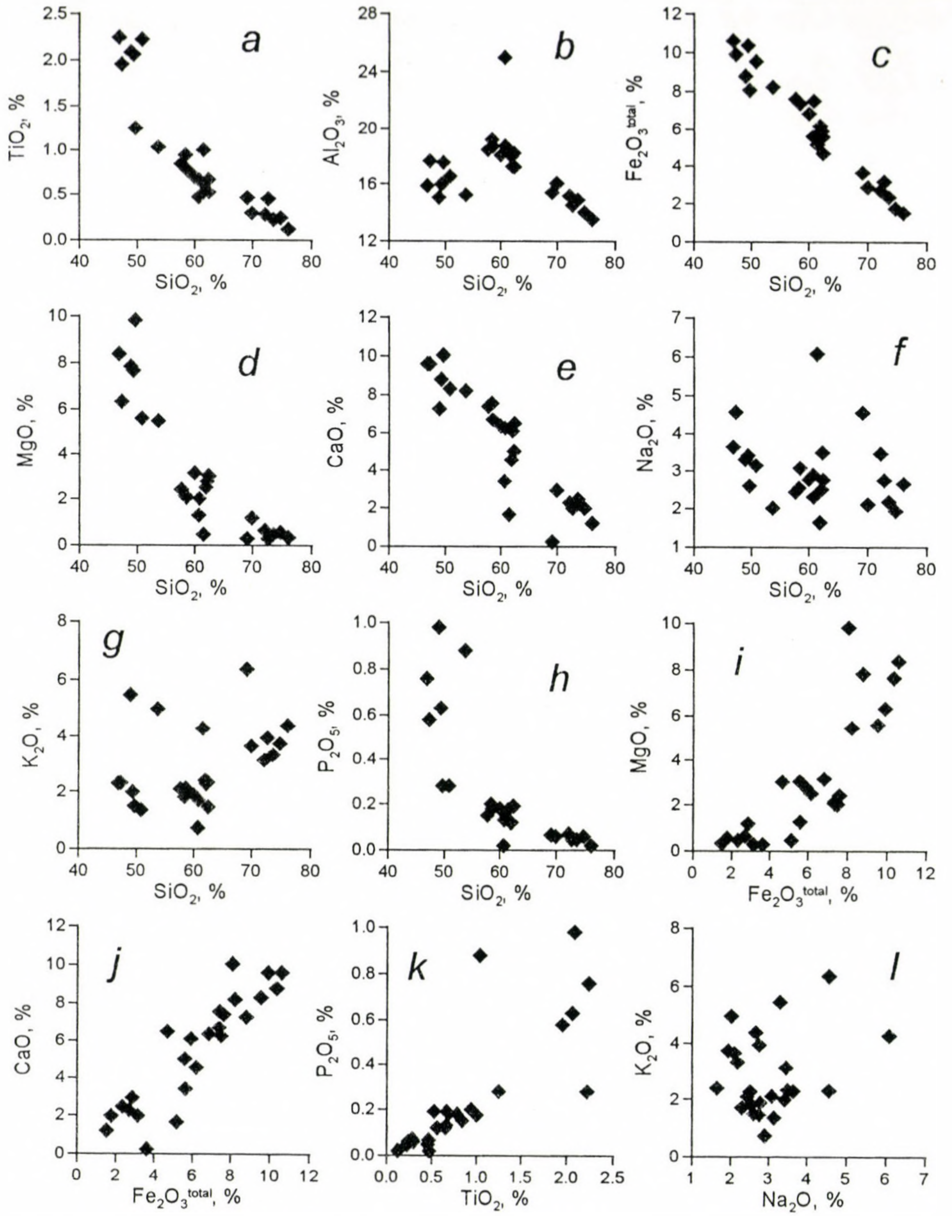

Fig. 13
Selected bivariate plots of formation medians. For remarks see section "Correlations and the problem of closure".

values of components, which always remain unknown. In petrology, the problem is even more complex, as there are at least three types of possible processes with different relations to the compositional data: those understood in terms of relative values (e.g. mineral crystallization from magma related to saturation), those increasing (e.g. entering the magma due to melting) or decreasing (e.g. gravitational crystal fractionation) the absolute value of one or more components, and those adding or subtracting material in post-magmatic stages (e.g. hydrothermal processes). All these processes, possibly mixed and/or superimposed, may change the relative values that are measured in the sampled rock. As changing of one component causes changing of all other components, in terms of relative values every observed process becomes multivariate, no matter how many components participate in the real process.

Therefore, the empirical bivariate (and multivariate) interrelations of components may only be regarded as reflections of possible *natural* interrelations. In other words, the observed correlations should only be accepted if they are consistent with other (petrographical, mineralogical, etc.) data. Hence, the plots in Fig. 13 generate a number of exciting questions: Are the patterns seen on these plots caused by a real petrological factor? Or, are they just a consequence of closure? If they are a numerical consequence, why are not they similar for all the components plotted against silica, for example? If they reflect natural relations (e.g. in the SiO_2–$Fe_2O_3^{total}$ plot) what is the explanation of fitting the same trends by so different (in character and age) formations? How would the picture change if using other central tendency estimates or entire populations instead of medians (that are, in fact, very concise representations of formations)? If a particular pattern observed is a combination of numerical and natural processes, what is the actual contribution of each? And so on.

As there are a lot of excellent studies on these rocks in the literature, almost any observed numerical relation may be checked for reality, or at least for probability, by confronting it with pieces of other, maybe less indirect petrological information. Classically, interpretations have been based on a limited number (usually dozens, and just rarely more than a hundred) of samples, which may cause overlooking general (supra-formational) factors. The relatively large number of analyses in the database allows investigations of both suitably selected subsets and the entire set together. Experimenting with different sample sets might be useful as different factors may have different scopes. Once a numerical pattern is there, it is expected to be evaluated in some way or other. And of course, fully petrological hypotheses and conclusions should not contradict the numerical observations, either.

Relations with time

Although the recently proposed tectonic models for the Pannonian Basin (Csontos 1995; Horváth 1993, etc.) and the petrologic interpretations of the given formations (Dobosi et al. 1995; Harangi et al. 1995a; Szabó et al. 1992; Pécskay et al. 1995, etc.) do not allow a simple, time-progressive genetic model for the whole assemblage of these volcanites, one might want to see the change of their compositional range through time. This means that, to an essentially multivariate system, time is also added. Bivariate plots of components against time, however, may give insight into this change. Plots in Fig. 14 portray the distribution of formation medians in geologic time. The points denoting the medians are enveloped by hatched polygons in order to help to catch possible trends. Although the fitting of polygons to the points is a bit subjective, three or four different patterns can probably be distinguished: those of Si-LOI, Ti-Fe-Mg-P-Na-K(-Ca), and Al.

The relative shape of these polygon types might be understood recalling the bivariate correlations of components (Fig. 13); their observed time-related stretch, however, is not obvious. At the level of description, there is no problem with treating all formations together, but the outlined patterns again raise the question: are they petrologically meaningful "trends", i.e. is there any common genetic factor that would lead to that kind of trend? Of course, there are remarkable differences between the formations, and probably all of them have undergone unique genetic events. This, however, does not necessarily eliminate the effect of common genetic elements.

Multivariate statistics

For understanding many, if not all, petrological processes, a simultaneous consideration of several/all variables is necessary. However, as a corresponding numerical evaluation requires more sophisticated mathematical procedures and always has results less easily defendable, most of the "classical" classification and diagnostics schemes rely on two-dimensional representations (TAS, K_2O-SiO_2, FeO^{total}/MgO-SiO_2, AFM, etc. diagrams), even those using trace elements and/or isotopes (Ba/Nb–La/Nb, $^{143}Nd/^{144}Nd$–$^{87}Sr/^{86}Sr$, and other diagrams sometimes found to be very powerful), while fully worked-out multivariate tools are rare (e.g. Pearce 1976). The potential of multidimensional techniques in revealing compositional similarities and/or dissimilarities of volcanites in Hungary and identifying the responsible petrological processes, however, has been shown (e.g. Ó.Kovács and Kovács 1994). The analyses in the given database may be grouped based on different principles, but the applied system of lithostratigraphic formations are so basic that one would certainly like to see their relationships in multivariate space.

Multivariate discriminant analysis

From among the formations it is not difficult to select a few that completely separate even in two dimensions (as an example see Fig. 15). Those with less exotic or less confined compositions, however, are lumped together and it is not

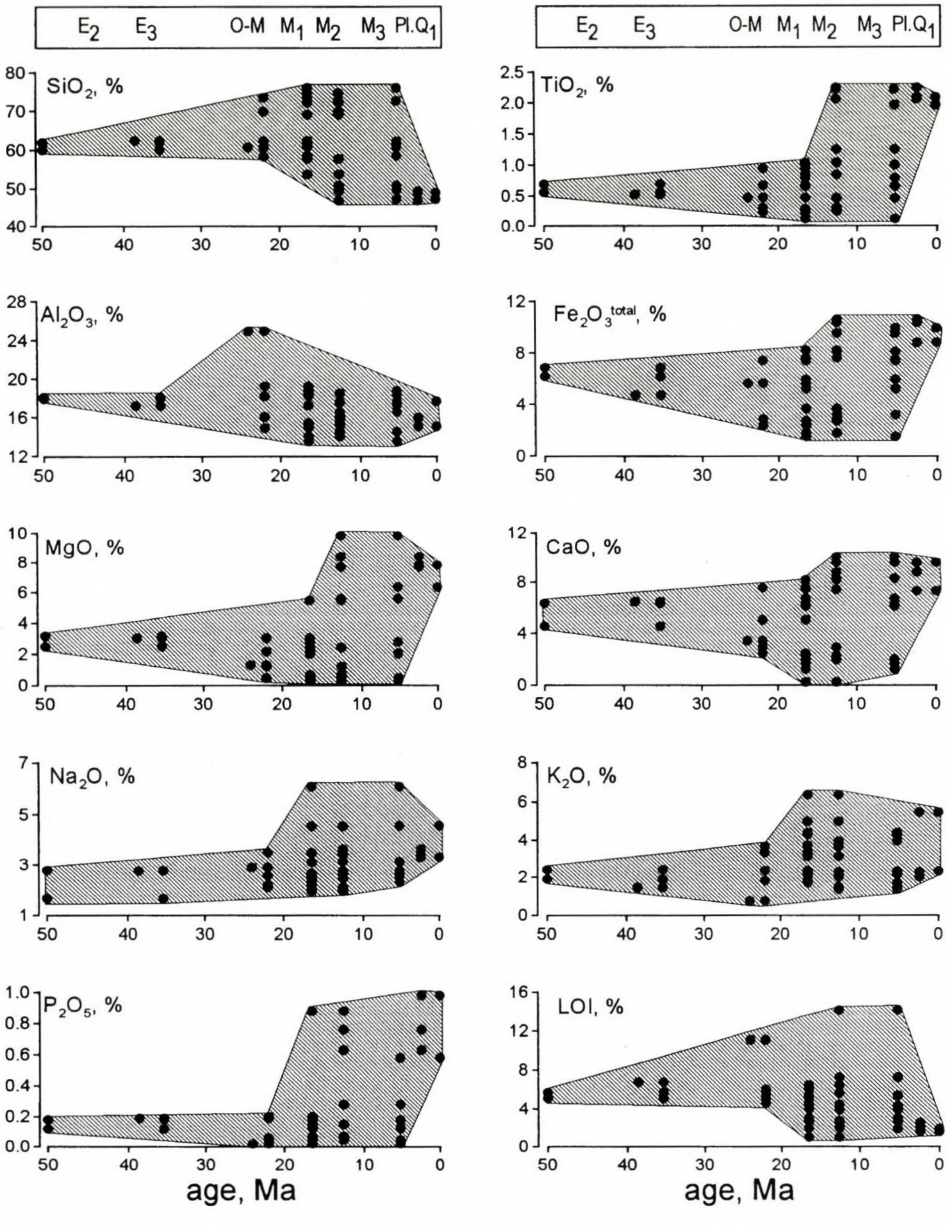

Fig. 14
Scatter of formation medians in time. Age data are adapted from Császár (1997). Each formation is represented by one point (in case it is bounded within one chronostratigraphic section), or by two points (in case it spans over more than one chronostratigraphic section). For details see section "Relations with time".

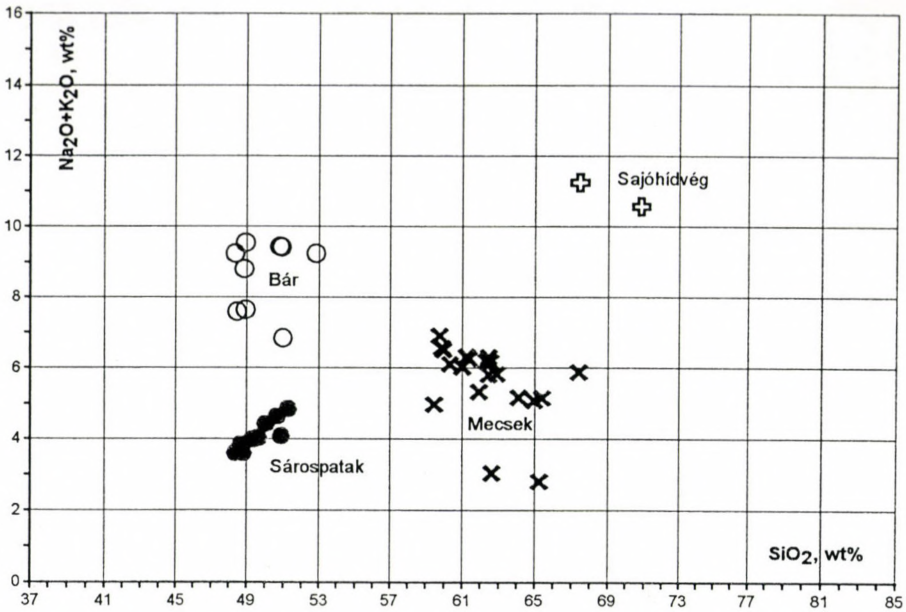

Fig. 15
Sample points of selected formations on the TAS-plane. The formations perfectly separate in two dimensions. For more comments see section "Multivariate discriminant analysis".

easy to see any difference between them (all the andesitic formations, for example). A lot of mathematical methods have been suggested to investigate groups of multidimensional objects (e.g. Davis 1986, pp. 468–620; Le Maitre 1982, pp. 58–177), but their success depends mainly on intra-group and inter-group similarities of the objects, i.e. of the rock samples here. Below, plots of samples based on multivariate discriminant analysis, a relatively simple method of evaluating multidimensional similarities, are given for selected pairs of formations.

Anyone dealing with a particular lithostratigraphic unit for some time would certainly have the feeling that the given formation is different from any other formation. And obviously, *any* group of rocks has some specialty. Nevertheless, it is not obvious with how much certainty it is possible to separate this specialty from a similar unit, and how reliably an unknown sample can be classified, especially in terms of petrochemical data. Having compared the overlapping (in one and two dimensions) formations it was found that the chances in the multivariate space are rather good. For the pairs Tapolca-Salgóvár, Recsk-Mátra, and Nyírség-Cserehát (Fig. 16) efficient discriminatory functions can be presented, as they separate with a success of well over 90%. In case of Tokaj 1–Mátra, Tokaj 2–Gyulakeszi, and Recsk–Nadap (Figs 16 and 17) an automatic classification based on the calculated discriminant functions was more than 80%

correct. The multidimensional separation of even the most intermingling pairs (Mátra–Börzsöny, and Gyulakeszi-Galgavölgy, Fig. 17) was 73% and 67%, respectively. All this means that each studied formation has a quantifiable petrochemical specialty; it does not prove, however, that each formation has been defined the most adequate way in a geologic sense.

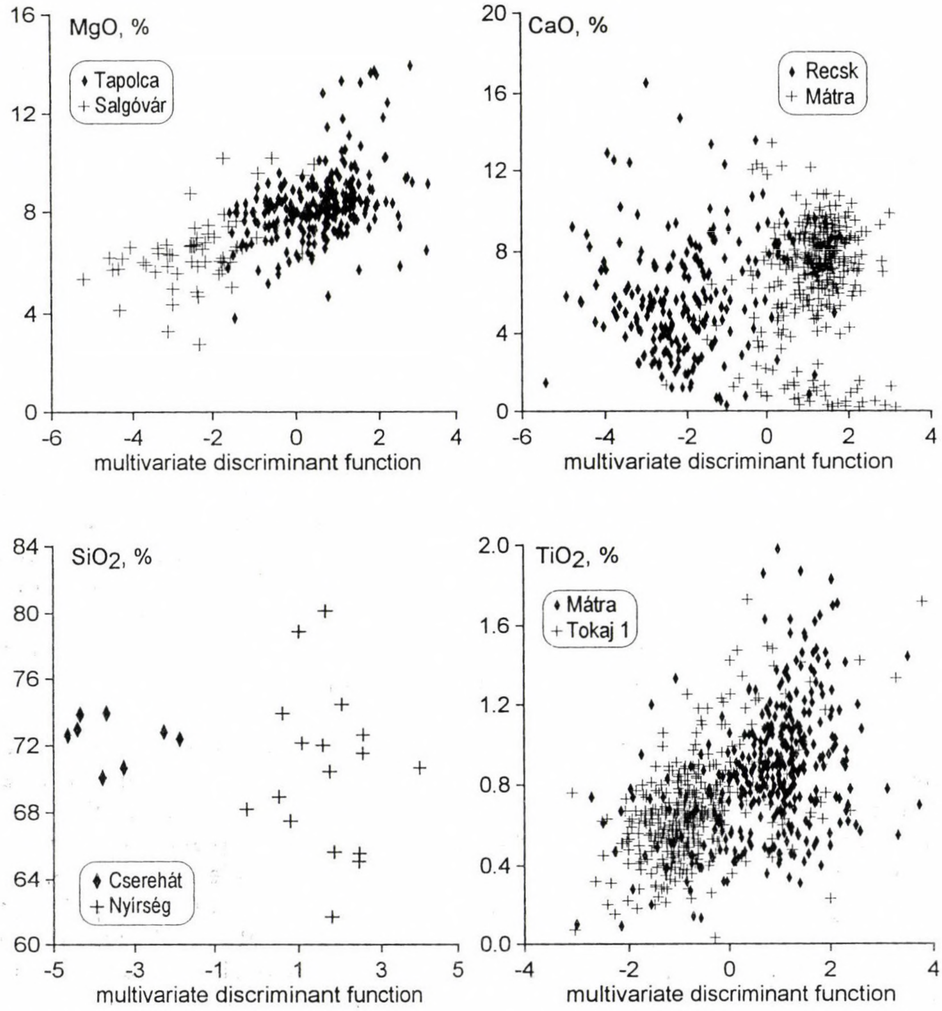

Fig. 16
Bivariate plots of selected overlapping (in 2D) formations on the plane of a multivariate discriminant function and a major component. For comments see section "Multivariate discriminant analysis".

The problem of subsets (do formations represent natural groups?)

In the database each chemical analysis is just a record. After recalculations (on a volatiles-free basis, with total iron and total loss on ignition) they have become more or less comparable. Selection of data sets for numerical or other processing can be done in many ways (e.g. based on age, rock type, the values of a particular

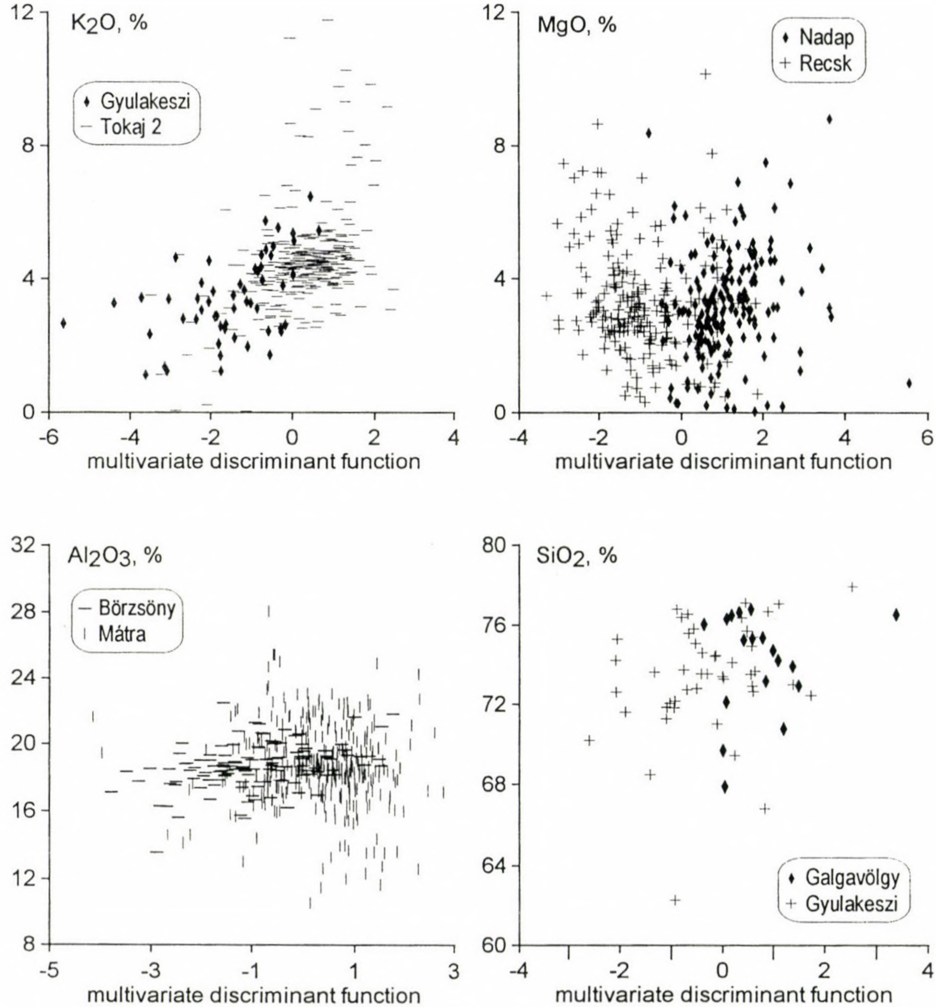

Fig. 17
Bivariate plots of selected overlapping (in 2D) formations on the plane of a multivariate discriminant function and a major component. For comments see section "Multivariate discriminant analysis".

component) and is technically easy. The question is: which is the most appropriate selection for a particular statistical (or other) analysis?

In this paper the basic division used is the system of lithostratigraphic formations expected to be complex and informative units (with individual appearance, genetics, age, etc.). However, the definition of the category of formation is tolerant enough to allow linking together very different, in some sense or other, rocks, just for the sake of having good (local) mapping units. Geologic maps and formation columns are so powerful that once a rock sample has been assigned to a formation it tends to loose any special feature different from the generalized image of the given formation. If it is different one is inclined to treat it as an extreme and incidental case. That means one is tempted to overlook certain natural variations.

The solution is easy when a set of samples comes from the same natural rock structure (a volcanic edifice, a lava layer, a volcanic body, etc.) and no formal geologic correlation with other structures is necessary. There are splendid studies in the literature with that kind of scope (a great part of the References of this paper). Clearly, complications begin with correlation, and no general solution exists. One should decide which is more important in a given analysis or synthesis: representation of the total observed variation, or emphasis of that part of variation found to be significant from a certain point of view. Again, experimenting with different sample selection schemes should help to capture interlacing subdivisions of different nature.

Uncertainties about the delineation of a formation are also documented by the relatively frequent changes in the system of formations: formations are cancelled or redefined, and new ones are introduced from time to time, and not exclusively due to the development of regional geology. Even within one book (Császár 1997), aiming to define the current lithostratigraphic units, formations may "partly correspond" to others (e.g. Salgóvár to Tapolca). Therefore, when investigating questions that concern several formations one should not expect more (or less) similarities or differences, or any relations than are incorporated with the (actual) definitions.

Suggestions for use

Due to its simple structure, the database can easily be used by everyone (at present it is maintained at the Hungarian Geological Survey by the authors). Numerical data evaluation and the establishment of links to other research projects have already begun. Development of the database by extending it to trace element and isotope data is being considered. For future applications a few suggestions have already been made above; below a few more are added.

Analytical data of a rock sample can only be given a (geologic, genetic, etc.) meaning in comparison with those of other samples. The simultaneously regarded data should obviously be expressed on the same basis. When the rocks are fresh, raw data might be directly comparable, but, as seen above, a large part of the sampled rocks is altered to some degree. It might be unjustified to use them in certain classification; they belong, however, to real rock formations. In order to compare recalculation to an adequate basis might be attempted. An adequate basis might be different in various applications. For example, if secondary alterations affected only certain components (say, CO_2 and CaO) one may try to look at data recalculated on a basis free of those. Or, when secondary sulfides are contained in a few percent, a recalculation on a sulfide-free basis will certainly better approximate the igneous composition even if the other components have also suffered from the hydrothermal process. On the other hand, the ore grade of porphyry copper, for example, is expressed relative to the whole-rock weight; that is, the primary basis is the adequate one in this case. For general descriptions like these, a recalculation on a volatiles-free basis and expressing the volatiles relative to the same basis is a recommendable procedure, which makes the non-volatile components more suitable for traditional petrological comparisons and correlations (but certainly not for normative calculations), and, at the same time, the values of volatiles are also correct in a relative sense. For that matter, elevated contents of volatiles may disturb the very basic classification, the one based on the silica concentration, of a sample as well. Approximating too highly altered rocks to the primary magmatic composition is, of course, not that simple, if possible at all; improving comparability, however, might help to preserve for a joint analysis as many samples as possible.

In multidimensional space it is usually difficult to reveal bimodality (or multimodality). Besides, geologic processes often produce overlapping distributions, or distributions whose differences may not be described in simple statistical terms. Therefore, any detected statistical bimodality (multimodality) should be appreciated and, possibly, reflected in the suggested divisions. Especially respectable should be univariate bimodality (e.g. those in Fig. 4), as they can easily be visualized. They are, of course, expected to be geologically validated to the extent the actual level of investigation requires. For example, a desirable subdivision of the Tokaj Volcanite Formation has already been mentioned and related to earlier observations (Gyarmati 1977). Similarly, bimodality of the rocks in the Tapolca Basalt Formation might be attempted to be correlated with described petrographic types, e.g. with the two major types ("basanitoid" and "feldspar basalt") of Vitális (1911), or with the two main multivariate petrochemical types revealed by Kovács and Ó.Kovács (1990). And so on.

Given compositional (closed) data, revealing or justifying "petrological" correlations among variables might be problematic. Geometrically, correlations are a kind of scatter of data points in the variables' space. Although numerically true, any kind of scatter may bear the influence of the fact of closure. This means any pattern, trend, etc. observed on scattergrams (bivariate, triangular or other) should not be *ab ovo* taken as a petrological one, but be always geologically validated, explained or, at least, understood.

A set of so different rocks like those in the given database are seldom treated together (for a nice exception see Grunsky et al. 1992). Usually, the problem of smaller and more homogeneous subsets is complex enough. What is the basis for still keeping them in one? Here, it is the geographical proximity of the sampled occurrences and the relatively long-lasting characteristic tectonic regime, as suggested in the above referred papers. Moreover, trace element and isotopic compositions suggest that all these rocks, from the Eocene to the Pleistocene, are genetically related to some degree: both subduction-related calc-alkaline and extension-related alkaline rocks have a common source component, the asthenospheric mantle, modified to a variable extent by mixing with different contaminating components (probably the upper crust and lithospheric mantle, respectively), as suggested for instance by Salters et al. (1988) and Downes et al. (1995). An ultimate documentation of the common genetic elements might require considering all relevant data. At the same time, it would be unwise to announce once and for all that major elements could not preserve any corresponding information.

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Contribution of volcanic material? A new aspect of the genesis of the black shale-hosted Jurassic Mn-carbonate ore formation, Úrkút Basin, Hungary

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This study is a complex interpretation of the results, achieved in the past 80 years, connected with the Toarcian black shale-hosted manganese mineralization of the Úrkút Basin from a genetic point of view. As up to now no final explanations for the formation of the deposit have been proposed, the aim of this study is to collect and summarize those arguments which support the idea of the contribution of fine-grained vitreous volcanic matter (glass, tuff) during the formation of the deposit (Polgári and Kecskés 1999).

This assumption is supported by the following peculiarities:

– the joint occurrence of smectite, zeolite (clinoptilolite, heulandite, gismondite?...), and potassium feldspar in the black shale and manganese ore samples;

– the presence of strongly weathered basic plagioclases described in several horizons of the formation, as well as the basic igneous and metamorphic heavy mineral assemblages in the thin marl underlying the manganese ore formation (Szabó-Drubina 1957, 1959, 1961);

– the Al-depletion of the authigenic clay minerals in the black shale;

– the relatively high Ti-content compared to poor sedimentary systems without any terrigenous contribution;

– silica access in the section;

– the occurrence of idiomorphic zeolite (clinoptilolite) crystals in rounded, Mn-bearing calcite fragments from the black shale horizon;

– the peculiar, delayed reduction of a portion of the iron content occurring only during diagenesis, where the increasing iron content of clay minerals caused the second separation step of Mn and Fe (possible Fe-bearing silicate mineral phase).

Key words: black shale, Tethys, Toarcian, genesis, vitreous tuff, clay, geochemistry, manganese carbonate, Hungary

Introduction and overview of the previous models

In the past 80 years several attempts have been made to explain the complex formation of the black shale-hosted manganese carbonate ore of Úrkút, but none of them could reliably resolve the contradictions (Cseh Németh 1965, 1966, 1967; Vámos 1968; Konda 1970; Szabó 1977; Varentsov et al. 1988; Polgári et al. 1991;

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Polgári 1993). It was Szabó-Drubina (1957) who first published the contribution of volcanic material on the basis of micromineralogical studies.

The investigations carried out so far have shown the relationship of the ore deposit to the Toarcian oceanic anoxic event, the role of marine currents (upwelling), the remarkable organic matter content accumulated due to algal boom on the shelf areas and the concentration of manganese in the basin. During diagenesis reduction of Mn^{4+} resulted in the decomposition of organic matter. The subsequent calcification of radiolaria tests indicate intense bacterial activity. All these processes have changed remarkably the original features of the formation; thus the reconstruction of the original state of the sediment is rather problematic.

Recently, the formation conditions of the manganese carbonate ore (Polgári et al. 1991; Polgári 1993) were explained by derivation of manganese from distant hydrothermal centers, the transport by oxygen-poor water in the course of the oceanic anoxic event as well as by precipitation-accumulation at the redox interfaces. The model has been accepted on a global scale (Dickens and Owen 1993). Previously, the derivation of the clay material proved to be problematic. The Liassic clays of Úrkút are partly so-called green clays (celadonite) and nontronite of authigenic origin (Kaeding et al. 1983; Grasselly et al. 1985, 1990). Varentsov et al. (1988) completed the interpretation of the clayey rocks of the Úrkút Formation taking also into account the geochemical characteristics. They explained the formation of celadonite by diagenetic transformation of Fe-smectite (nontronite) into Fe-mica (celadonite) following deposition. The same could be observed in the Galapagos Rift zone (Varentsov et al. 1983) where the hydrothermal character was considerably modified by hydrogenetic and diagenetic effects.

It is well known that in young sediments the traces of contemporaneous volcanism can be usually well recognized. These can be either directly observed or can be indirectly followed. With increasing geologic age the indicators may become overshadowed due to the diagenetic processes.

This study presents an overview on literature and some new aspects of results at the beginning of further research, and summarizes the indirect evidence of smectite formation process during transformation of the volcanic vitreous material. A further aim of studies can be to identify the possible volcanic explosion center.

Geologic setting

Jurassic formations of the Bakony Mountains

Úrkút is located in the central part of the Bakony Mountains belonging to the North Pannonian unit of the Alps–Carpathians–Pannonian region (Csontos et al. 1992).

The Jurassic of the Bakony Mountains, similar to the sequence of the Southern Alps (Bernoulli and Jenkyns 1974; Winterer and Bossellini 1981) was deposited on the basement of differentiated substrate of the passive southern continental margin of the Jurassic Tethyan Ocean (Galácz et al. 1985). The paleotectonic evolution of the strata sequence of the area is highly similar to that of the Trento Platform of the Southern Alps (Kázmér 1987; Kázmér and Kovács 1985) and to the Jurassic of Western Sicily (Jenkyns and Torrens 1971) and of the Northern Calcareous Alps.

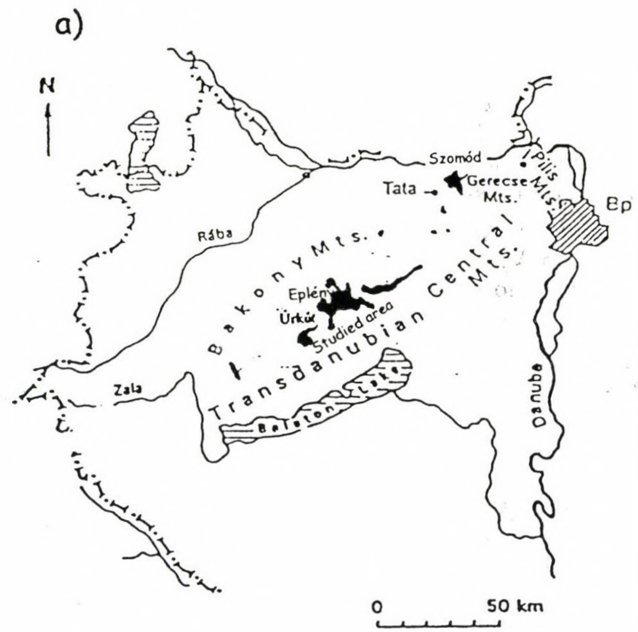
According to the distribution of ammonite and brachiopod fauna the area was separated by widening marine trenches, both from the European and from the African shelf (Géczy 1984; Vörös 1987, 1988).

The Upper Triassic Dachstein Limestone constituting the main mass of the mountains is overlain by a varied Jurassic sedimentary sequence (Fig. 1a, b), the lithological, sedimentological and paleontological characteristics of which were caused by the differentiated subsidence of the dissected uniform carbonate platform (Galácz and Vörös 1972).

Generally the Jurassic formations are characterized by condensed sedimentation, by reduced thickness and rapid horizontal change of the lithological units. According to the studies of Galácz and Vörös (1972), Galácz (1988) and Vörös (1991), the dissected morphology developed during the Liassic (ridges and inter-ridge basins) which determined the characters of sedimentation: in the deeper areas the sedimentation was interrupted by short-term unconformities, in the uplifted areas the sequences are dissected by considerable unconformities, in the transitional zones on the steep slopes synsedimentary breccias and Hierlatz Limestone were formed. Evidence of block tectonics related to the rift formation of the passive margin are the fissure fillings of Neptunian dykes related to the broken slopes of ridges (Fülöp 1975; in the South Alpine formations: Winterer et al. 1991), the synsedimentary breccias accumulated at the foot of ridges as well as the intercalation of redeposited material in the pelagic sequences of basins (Galácz and Vörös 1972; Galácz et al. 1985).

The Úrkút Manganese Deposit

Manganese mineralization occurs in the Úrkút basin, which was formed by north–south trending block faulting that characterized the Late Triassic and Jurassic of this region (Cseh Németh et al. 1980). The Úrkút deposit of Late Liassic age is found within marine sedimentary rocks composed mainly of bioclastic limestone, radiolarian clayey marlstone and dark-grey to black shale (Grasselly and Cseh Németh 1961; Polgári 1993). Accordingly the development of the Úrkút basin is placed in the period when accelerated crustal extension propagated the rifting from the Tethys toward the Atlantic Ocean that led to the initiation of



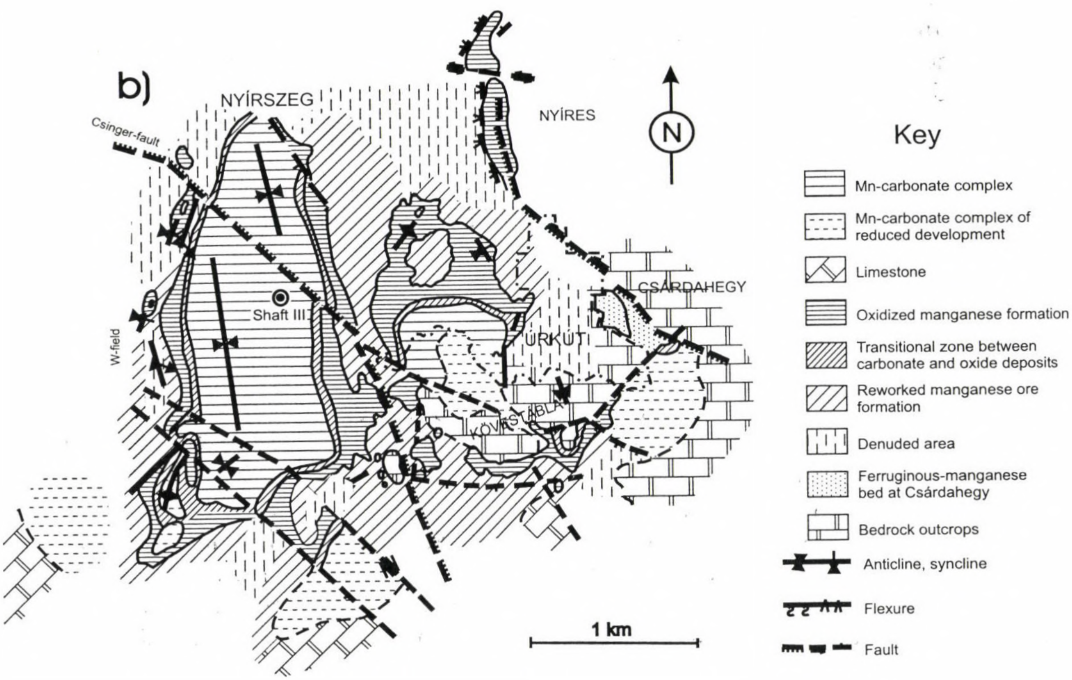


Fig. 1
a) Location map (after Dosztály, 1998) and b) geologic sketch map of the Úrkút area, Hungary (after Cseh Németh et al. 1980). Sample location: Shaft III. Note: the main ore bed and ore bed No. 2 are the thickest and laterally most extensive where indicated as fully developed; both beds are present, but thinner where indicated as less developed. The denuded areas are predominantly, but not completely, eroded to basement

seafloor spreading in a narrow east–west trending trough up to the end of the Middle Jurassic (Ziegler 1988).

Mn-mineralization occurs in a radiolarian clayey marlstone (black shale) of Toarcian age. The marlstone rests conformably on Middle Liassic carbonate rocks formed at the center of the depositional basin. The marlstone is about 40 m thick at the center of deposition and thins out toward the basin margins (Szabó et al. 1981). The manganese deposits form a NE–SW trending unit of approximately 12 km length and 4–6 km width. According to a recent survey the economically important ore deposit covers an area of 8 km^2 (Szabó et al. 1981).

Manganese mineralization is restricted to two intervals within the marlstone. The lower main bed is about 8–12 m thick and is underlain by 0.5 to 1 m-thick radiolarian clayey-marlstone; the upper mineralized zone, Bed 2, is 2 to 4 m in thickness and is separated from the main ore bed by 10 to 25 m-thick radiolarian clayey-marlstone. Figure 2 shows two Mn-carbonate ore profiles from different localities of shaft No. III from the Úrkút Mine. Section 1 is from the Mn-carbonate gallery of the northern part (+256 mBf) of the western mine area, and Section 2 is from the section between 245 and 320 m of the deep level exposure (+175 mBf). At the base the ore sections begin with a thin, greenish, organic-rich, pyritiferous radiolarian clayey-marlstone containing enrichments of trace elements, namely Co and Ni (Polgári 1993). Concretions and thin layers of phosphate and chert are common at the boundary of the marlstone and the underlying limestone. The rhodochrosite ore section is composed of alternating grey, green, brown and black sections of finely laminated, very fine-grained clay mineral–carbonate mixtures (Cseh Németh and Grasselly 1966). Fine-grained (1–2 μm) rhodochrosite rock lacks coarse detrital clastics (Szabó-Drubina 1957). The formation is microlaminated lenticular, microlenticular or microstratified detrital. Microlaminae are 15 mm thick.

From a sedimentological point of view the Mn-carbonate ore and black shale of Úrkút form a finely laminated clayey-marl/calcareous-marl sequence of primary depositional stratification that was emphasized by the separation of more carbonatic and more clayey parts caused by diagenesis (Mindszenty in Grasselly et al. 1990). Laminites may reflect different origins. It is only the proportion of main mineral components changing in them that is macroscopically represented by the different colors. The grain-size is uniformly a few mm, sometimes locally somewhat coarser. Reasons of this stratification can be seasonal changes, the change of the rainy/dry periods, productivity cycles, the periodicity of Mn-inflow and precipitation, respectively.

Mineralized sections show a lack of fossils or traces of benthic fauna, and contain only rarely fish scales and bone fragments as well as coalified plant fragments. According to palynological investigations the presence of smaller island(s) in the surroundings cannot be excluded (Kedves 1990).

Radiolarian chert contains abundant clay minerals and occurs stratigraphically within or immediately above the Mn-beds.

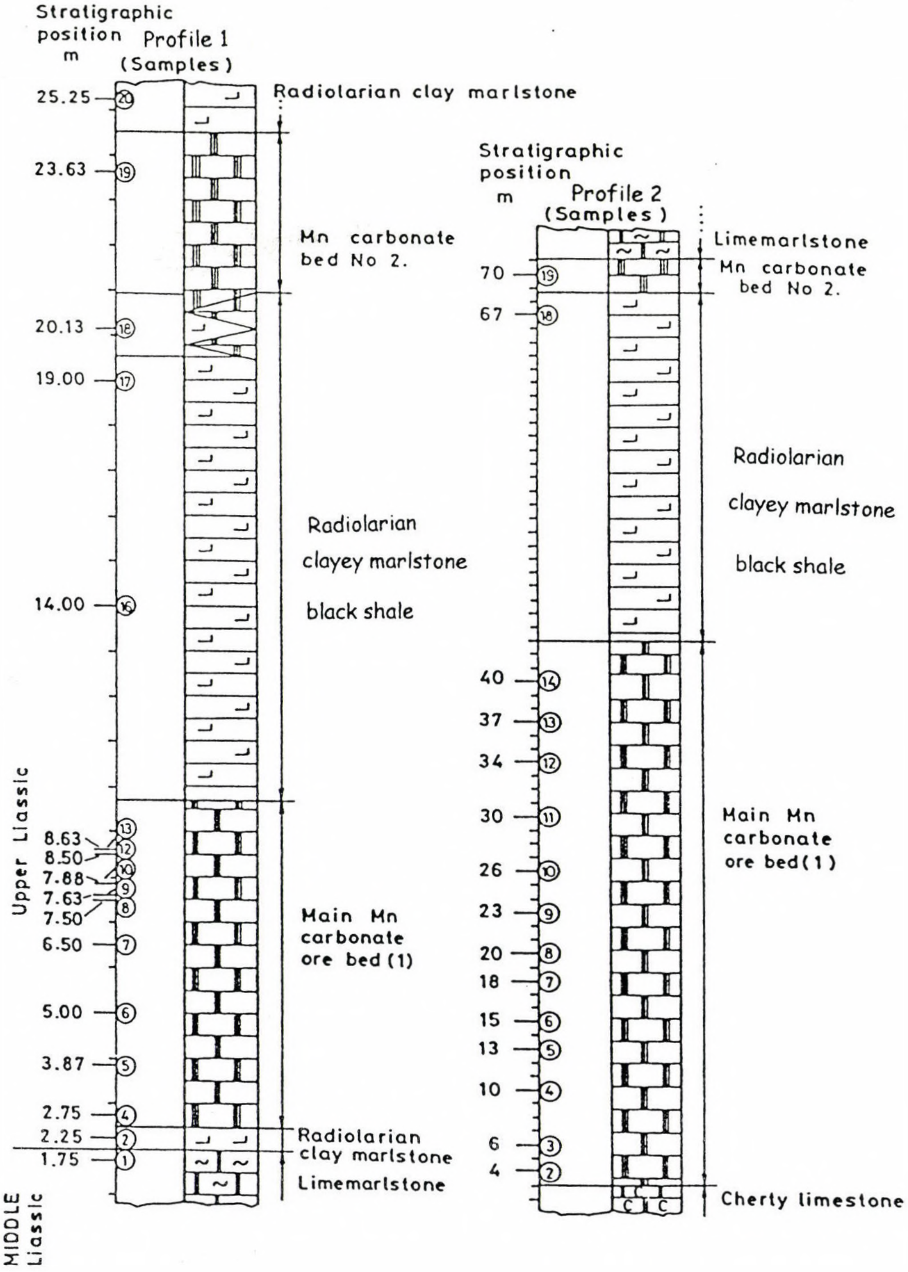


Fig. 2
Section of two typical unaltered Mn-carbonate sequences from two localities from shaft III of the northern part of the western mine field, Profile 1 (+256 mBf) and Profile 2 (+175 mBf) (after Szabó et al. 1981) (for sample locations see Fig. 1b)

Basic data on the geology and mineral, chemical and trace element compositions of the manganese ores are presented mainly in the works of Hungarian geologists (Grasselly and Cseh Németh 1961; Cseh Németh and Grasselly 1966; Cseh Németh et al. 1971, 1980; Szabó and Grasselly 1980; Szabó et al. 1981; Polgári et al. 1991; Polgári 1993).

The age of the formation falls into the falciferum ammonite zone (Toarcian; Géczy 1972). The accumulation rate of the deposit was very low, about 50 m/My as estimated by Vető et al. (1995). The part of the Toarcian Tethys where the manganese ore formation was deposited was probably characterized by high biological productivity. During deposition of the formation the productivity slowly increased (Vető et al. 1997).

Mineralogical and geochemical characteristics in the view of genetic models

The interpretation of the black shale and manganese ore mineralization of Úrkút requires a complex mineralogical and geochemical approach.

Mineralogical aspects

General composition

The mineralogical composition of the formation studied is the following (Pápai and Tóth in Grasselly et al. 1985):

1) carbonate ore, main ore bed and bed No. 2: rhodochrosite, 10Å-phyllosilicate (celadonite), smectite (nontronite), quartz, pyrite, chlorite (tr), zeolite, feldspar (tr),

2) black-shale: quartz, calcite, pyrite, smectite, 10Å-phyllosilicate (illite, celadonite), chlorite, zeolite (tr), rutile.

1) According to Viczián (1995) clays in Lower Toarcian manganese ores can be characterized by the local enrichment of smectites within a limited time interval. In other areas, however, in the same stratigraphic horizon and in other Liassic formations the normal detrital illite-dominated clay mineral assemblage is found.

Origin of REE containing minerals

Considering the genesis of the mineralogical composition, first and foremost it should be emphasized that neither is the source of the main elements, of the Mn and of the clay in the Toarcian black shale and manganese carbonate ore, by all means the same, nor are the processes forming them.

Previously the two formations were treated together from the aspect of source material. Authors, on the basis of the REE distribution pattern measured in the clay, believed them to be of terrestrial origin (Grasselly and Pantó 1988). The particular chemical and trace element analyses proved that rare earths accumulated originally in thin phosphorite lenses together with Th and U at the base of the deposit, though it is true that these lenses are imbedded in the clay.

REE were bound to the clay as a result of subsequent oxidation (element mobilization) and finally precipitated in carbonate form (bastnäsite). The geochemical reactions took place during the diagenetic processes; pyrite decomposition and Mn-carbonate oxidation was displaced toward the low pH range that modified the original distribution of REE (Wright et al. 1989).

These phosphorite bands of high REE (2000 ppm) and high Th and U (up to 4/Th and 181/U ppm) contents are characteristic of the upwelling systems of shelf areas (Hein et al. 1999) where the nutrient-rich water masses of the upwelling marine currents promote the production of algae. The Eplény and Úrkút-Csárdahegy occurrences in the Bakony Mountains (Fig. 1) can be considered to represent these upwelling formations, these differing to some extent from the mineralization of carbonate character at the main Úrkút deposit. Features characteristic of upwelling can also be observed in the lower region of the Úrkút deposit, so this could also affect the development of that deposit.

Mineralogical composition, similarity to volcanic rocks

Disregarding the Mn-ore and organic contents of the formation, the remaining mineral assemblage (smectite, zeolite, potash feldspar) agrees fairly well with the typical case of decomposition of volcanic glass. In case of Úrkút it is obvious that no 100% tuff composition, but only a smaller portion, that could have been deposited in form of fine-grained and highly reactive vitreous form.

Authigenic clay minerals

When reviewing the chemical compositions, the Al-content of the Úrkút authigenic clay is low as compared to the terrestrial clays. The average value for Section 1 is 4.73 and for Section 2 3.03 wt.% Al content.

The samples of the deposit are Fe-rich, which is proved by the ratio Fe/(Fe+Mg+Al) = 0.70 (Varentsov et al. 1988). In the profiles different grades of the transformation of Fe-rich smectite into Fe-mica can be detected. The mobilization of a part of Fe proceeded in the late phase of diagenesis.

Based on the investigations the ferric iron survives in certain form its stay in the reductive zone where, in harmony with thermodynamic equilibrium, it ought to enter into reaction, and is actively reduced only later. The mechanism that preserves the metastable ferric iron is not yet clear but the following possibilities must be taken into account:

a) microbiological relations (e.g. at room temperature the sulfate ion is stable in presence of organic carbon without bacteria),

b) the mobility of ferric iron and organic carbon is problematic; thus their reaction requires certain preconditions, i.e. the soluble organic carbon has to be transported to the iron and the ferric-containing mineral has to be weathered (this difficulty does not exist in case of Mn-oxyhydroxide since it is a gel with high water content).

The ferrous iron formed in the delayed Fe reduction zone often enters the structure of clay minerals.

In the sediments the inhomogeneous distribution of organic carbon is important; this produces strong local diagenetic activity and diffusion processes. The concurrence of pore waters of different peculiarities or the episodic events may cause considerable changes in the system.

Thus, Fe-rich character (ferrous and ferric iron together) of the clay minerals in the Úrkút Basin can be explained by a peculiar Fe-bond. The presence of nontronite in some horizons of the ore as second step of the Mn-Fe fractionation as well as its joint occurrence with pyrite are related to the fact that a part of the ferric content survived the sulfate reduction occurring during early diagenesis and was reduced only later. The ferrous iron thus produced entered the clay mineral structure (Coleman 1985). Another evidence of the ferric iron survival may be the local occurrence of hematite and goethite. As to the K/Ar age determinations of Grasselly et al. (1994) celadonite formed subsequently to the Toarcian sediment accumulation, in a later phase of the temporally lasted diagenesis, i.e. 151 ± 6 Ma ago.

The source of Fe needs to be investigated. Iron deriving from the volcanic hydrothermal belts of submarine spreading centers is unable to get as far as Mn does. Because the inflow of Fe is hardly possible together with the volcanic detritus, more probably it is released during weathering. Finally it enters into the clay mineral structure.

The transportation of vitreous volcanic material into the Úrkút Basin could have occurred in different ways, e.g. by atmospheric means or also by marine currents.

Authigenic smectite is one of the important factors of geochemical equilibrium of oceans. According to Hein et al. (1979) four main types of authigenic smectite formation in marine environment can be distinguished:

1) direct precipitation from hydrothermal solution,

2) precipitation from solution at low temperature in vesicles of basalt,

3) alteration of volcanic rock detritus and volcanic glass in marine environment (transformed clay)

4) formation of "biogenic clay" (neoformed clay) by low temperature combination of Fe-oxyhydroxides and silica.

Among the listed processes the alteration of volcanic material is most frequent and most common.

Other clay minerals occurring in marine sediments (illite, chlorite, kaolinite) are terrigenous in origin, older and entered the sedimentary basin most probably through eolian or fluviatile transport.

In case of Úrkút processes 1) and 2) can be excluded, since there are no traces of hydrothermal centers and the terrigenous origin of celadonite and nontronite cannot be proved either. Liassic sediments were formed at ambient marine temperature; no traces of thermal effects have been identified so far.

The formation of "biogenic clay" (neoformed clay) cannot be excluded since the biological activity was remarkable. The dissolved tests of siliceous algae could provide the silica material, and iron could have been available in the system from the submarine hydrothermal centers. The origin of Al in biogenic clay is more complex and according to Hein et al. (1979) and Heath and Dymond (1977) its origin can be summarized as follows: 1) Al can be bound to the biogenic silica or can be adsorbed on it and released during dissolution; 2) being transported by rivers into the ocean it can be deposited on the sea bottom forming Al-hydroxide, or can be adsorbed by other particles; 3) it can be released during weathering of volcanic dust particles. Smectite formed in this manner is probably of high Fe, of low Al and probably of low Mg content in case of precipitation close to a Fe-oxyhydroxide source or, of moderately to high Fe, of moderate Al and low-to-moderate Mg contents in that of formation far from the Fe-source.

In fact, the internal part of some microfossils in the clayey formations of the Úrkút deposit is occasionally filled with a green clayey phase. Koritár et al. (1998) investigated the Lower Jurassic glauconitic clay mineral in the samples from the Tűzkőhegy of Szomód (Gerecse Mountains, Fig. 1a) where the green clay also filled the shells of fossils (though only partly of microfossils): ammonites, pelecypods, and ostracods. Based on X-ray diffractometric, chemical and Mössbauer spectroscopic analyses they qualified the material as high-Fe dioctahedral hydromica with about 20–40% intercalated smectite, i.e. as high-Mg and high-Al interstratified glauconite: i.e. a neoformed clay mineral. Nevertheless, the geochemical and chemical character of the celadonite in the black shale-hosted manganese carbonate ore of Úrkút is different from this glauconitic mineral. However, proving the possible volcanic origin of the neoformed part of the clay is a very difficult task.

Gottardi and Galli (1985) stated that there are two modes of origin of the widespread glassy detritus in the ocean bottom: 1) the acid explosive activity disperses the great amounts of acid glassy particles into the ocean through the atmosphere and 2) the rapid chilling of the hot basaltic lava material of the non-ridge-related submarine central volcanoes generates the granulation or pulverization of the hot lava in the seawater itself, and the currents spread the basic glassy particles. Basic-to-alkali explosive activity also exists but its contribution can be neglected as compared to the effect of the two mentioned factors.

Basic plagioclase

Regarding the weathered basic plagioclase occurrence in the underlying marl layers of the manganese ore bed (Szabó-Drubina 1957, 1959, 1961), the glass may have been of basic composition.

Iron-manganese nodules

In the Bakony Mountains, in several stages of the Jurassic, Fe-Mn nodules (pisoliths) were formed in oxidative environment of the ammonitico rosso facies. The mineral-chemical features and formation of these concretions was studied by Grasselly and Szentandrássy-Polgári (1985), Mindszenty et al. (1986), and Cronan et al. (1991). Reviewing the data again, in some of the nodules a joint smectite-zeolite occurrence is characteristic, associated with shells and cemented skeletons of sessile benthic carbonate-skeletal microorganisms. Between the clayey and carbonate phase Co-Ni enrichment occurs. In these cases the presence of volcanic glass debris is possible as a nucleus around which the benthic organisms lived and the Fe-Mn concentration proceeded. Similar Fe-pisoliths of Jurassic age are common in Western Sicily, some of them showing altered volcanic debris (calcitized trachyte) as a core (Jenkyns 1970a, b).

Zeolites (clinoptilolite)

In two unaltered black shale and manganese carbonate ore profiles the occurrence of minerals characteristic of the decomposition of volcanic glass is as follows (Table 1a, b).

In case of whole rock X-ray diffraction analysis the zeolites and potash feldspar were close to the detection limit but could be unambiguously defined; this means a proportion of several percent. Idiomorphic zeolite (clinoptilolite) crystals are shown in Mn-bearing, rounded calcite fragments from the black shale horizon (Plate I). The entire formation is a very fine-grained mixture; thus the separation and enrichment of individual mineral phases is difficult.

In the formation the occurrence of zeolite (clinoptilolite) can be interpreted as follows:

According to Hay (1978) zeolites are extremely widespread in volcanoclastic sediments, especially in vitreous tuffs, and may constitute even 80% of the weathered volcanic deposits.

Based on the comprehensive work of Gottardi and Galli (1985) the two most frequent zeolites occurring in marine sediments are phillipsite and clinoptilolite. The occurrence of these two zeolites in marine sediments is peculiar. Phillipsite is common in recent sediments, since the transformation of basic glass into phillipsite is a rapid process; thus the older rocks do not contain basic glass debris. The transformation of acid glasses into clinoptilolite, however, requires more time. It is well known that phillipsite never occurs in sedimentary rocks older than Jurassic, and generally not older than Miocene. In greater depths phillipsite is not stable; it is transformed into the more stable clinoptilolite. Nevertheless, clinoptilolite may form from basic glass in the presence of high silica activity. At the same time the crystallization of phillipsite in solution alongside the precipitation of Ca-carbonate is impossible for chemical reasons. Beside phillipsite, frequent accessory minerals are smectite, Fe-Mn-oxides and/or

Table 1/a
Minerals of possible volcanogenic origin in the northern part of the western mining field Úrkút, shaft No. III, Profile 1 (+256 mBf) on the basis of X-ray diffraction analysis

Sample No.	Rock type	Stratigraphy	Zeolites (clinoptilolite, heulandite)	Potassium feldspar	Smectite
1	red-brown, green marl st. (Middle Liassic)	Lime marlstone		*	
2	dark-grey, black radiolarian clayey marlst.	Rad. clayey marlstone (bs)	*	*	*
3	grey-green carbonatic Mn-ore	Main Mn-carbonate ore bed (1)	*	*	*
4	green-grey carbonatic Mn-ore		*	*	
5	brown-black carbonatic Mn-ore		*	*	
6	green, thick-banked carbonatic Mn-ore				*
7	brown rock part				*
8	grey rock part				
9	green, coarse-banded carbonatic Mn-ore				
10	grey, carbonatic Mn-ore				*
11	green rock part		*		
12	grey, carbonatic Mn-ore				
13	green, coarse-banded carbonatic Mn-ore				
14	grey, carbonatic Mn-ore		*		*
15	radiolarian clay-marlst, direct overlying part of the main bed	Rad. clayey marlstone (black shale)	*	*	*
16	radiolarian clay-marlst. (thick-banked)		*	*	*
17	radiolarian clay-marlstone		*	*	*
18	radiolarian clay-marlst. with $MnCO_3$-bands	Mn-carbonate ore bed No. 2	*	*	*
19	Mn-carbonatic bed No. 2.				
20	radiolarian clay-marlstone	Rad. clayey marlstone (bs)			

Analyzed and interpreted by M. Tóth, Lab. for Geochemical Research, Hung. Acad. Sci., 1983.

-hydroxides, in addition to the basaltic glass fragments. The formation of phillipsite follows after the hydratation of glass, i.e. after the transformation of glass into palagonite. The final phase of palagonitization is when palagonite is replaced by intragranular authigenic minerals (zeolite, smectite, Fe-Mn-oxide aggregates).

The occurrence of clinoptilolite in deep-sea conditions is frequent. It is much more frequent in carbonate-containing sediments under high sedimentation rate. It can be accompanied by opal and/or quartz.

There is general agreement that most deep-sea clinoptilolites crystallize from acid (mainly rhyolitic) volcanic glass as an alteration product thereof when the glass debris is deposited into the ocean from the atmosphere in relation with the explosive volcanic activity at the continental margins but there is no obvious relationship between the presence of glass and clinoptilolite.

Zeolites may also form through the reaction of poorly crystallized or X-ray amorphous clay. In fact, authigenic clinoptilolite occurs in many non-tuffaceous, brown, detrital pelagic clays. The common association of radiolarians or authigenic cristobalite with clinoptilolite suggests the contribution of biogenic silica to the formation of clinoptilolite (von Rad and Rösch 1972). Moreover, clinoptilolite casts of radiolarians and forams have been widely noted in non-tuffaceous sediments.

Table 1b

Minerals of possible volcanogenic origin in the deep-horizon profile in the western mining field (Úrkút, shaft No. III, Profile 2 +175 mBf) on the basis of X-ray diffraction analysis

Sample No.	Rock type	Stratigraphy	Zeolites (clinoptilolite, heulandite)	Smectite
1	limestone with chert	cherty limestone		*
2	grey, green, fine-banded carbonatic Mn-ore	Main Mn-carbonate ore bed (1)		*
3	brown-grey, fine-banded carbonatic Mn-ore			*
4	fine-banded brown, grey carbonatic Mn-ore			*
5	grey carbonatic Mn-ore			*
6	grey-green, coarse-banded carbonatic Mn-ore			*
7	green, grey thin-laminated carbonatic Mn-ore			*
8	brown, black carbonatic ore			*
9	black, fine-banded carbonatic ore			*
10	black, fine-banded carbonatic ore			*
11	brown-grey, coarse-banded fine-laminated Mn-ore			*
12	brown-green, coarse-banded carbonatic ore			*
13	green, coarse-banded carbonatic ore			*
14	grey, coarse-banded carbonatic ore			*
15	rad. clay-marlstone	Rad. clayey marlstone (black shale)	*	*
16	rad. clay-marlstone		*	*
17	rad. clay-marlstone		*	*
18	rad. clay-marlstone			*
19	grey, green, coarse-banded carbonatic ore, bed No. 2.	Mn-carbonate ore bed No. 2		*
20	lime-marl	Lime marlstone		*
21	ferrous chert	chert		

Analyzed and interpreted by Mária Tóth, Lab. for Geochemical Research, Hung. Acad. Sci., 1983.

X-ray amorphous materials forming from volcanic rocks, mainly from vitreous tuffs by weathering, are highly reactive in the marine environment and may contribute to the formation of phillipsite, nontronite, montmorillonite-illite and sepiolite.

In summary, the study of deep-sea sediments shows that phillipsite, clinoptilolite and other zeolites form at low temperature through the reaction of the volcanic glass and interaction between silica-enriched (biogenic silica) sea-water and poorly crystallized alumosilicate material.

Geochemical aspects

The geochemical analysis of manganese carbonate ores and clays of the Úrkút deposit is based mainly on studies of the international DSDP.

Based on (Fe+Mn)/Ti and Ti/Al ratios Varentsov et al. (1988) assumed transportation of fine-grained terrestrial debris. As to the heavy metal relationship of (Co+Ni+Cu+Pb) - (Fe+Mn)/Ti, the higher heavy metal content cannot be interpreted by hydrothermal processes, but hydrogenic effects can explain the accumulation. The Ni-Co relationship is unusual due to the accumulation of Co exceeding that of Ni; this was explained by the reductive character of bottom water.

Regarding the major element and trace element composition characteristics Varentsov et al. (1988) found that the deposit displays a hydrothermal effect accompanied by limited terrestrial hydrogenic and strong diagenetic impact.

Taking into account all these contradictions and comparing them to the features of the deposit, namely that the occurring minerals are almost solely syngenetic and epigenetic, the assumption of contribution of volcanic material resolves the problem, and may explain the higher Ti and lower Al contents (Al content of terrestrial clays is higher). The Si/Al ratio in some layers is higher than 3, which means an extra Si supply to the system (Arthur and Premoli Silva 1982). To explain the slightly enriched character of heavy metals the formation of the deposit in the peculiar black shale environment cannot be neglected, since the considerable organic matter content and its decomposition was favorable to the formation of organo-metallic complexes. Microprobe analyses showed the presence of dispersed Co-Ni-S and Ag-S phases in the black shale (Polgári 1993). According to the investigations of Vető (1998) in some samples the detected increased Mg content suggests that the flux of Mn-oxyhydroxides was accompanied by fluxes of Fe and Mg minerals (goethite and Mg-Fe silicates ?), which can also support the idea of volcanic (tuff) material contribution to the sediments.

Jurassic volcanism in the Tethyan Basin

It follows from the above that, although both the authigenic smectite and the zeolite formation can be deduced to have originated in greatest amounts from volcanic products, the joint occurrence of these minerals does not mean a single process of formation, especially in the presence of microbiologically-derived organic matter. This is why the comparison of features of the close-lying similar formations as well as the overview of main geotectonic processes of the given period, are of great importance.

In the Gerecse Mountains, in the manganiferous clay similar to the black shale-hosted manganese carbonate formation of Úrkút (but of higher organic matter content and only of 50 cm thickness), large quantities of brown, brownish-green, predominantly euhedral, pseudohexagonal, slightly altered biotite, dispersed amphibole and augite grains as well as part of the ilmenite and magnetite prevailing in the heavy mineral fraction, indicate airborne pyroclastics derived from volcanic activity. Based on the geographic arrangement of the volcanogenic fragments and taking into account the recent geographic directions, the Liassic – Early Dogger (?) volcanic activity was probably located northeast of the Gerecse Mountains. This is supported by increased ilmenite+magnetite frequency in this direction as against frequency increase of continental minerals to the west (Árgyelán and Császár 1998). On the other hand, the clay mineral composition does not show the characteristics of volcanic origin (Viczián 1995). Investigation results from the Gerecse Mountains may contribute to the understanding of the formation conditions of Jurassic units of the Bakony Mountains. Especially the accumulation of airborne crystal fragments, probably derived from volcanic activity, is worthy of mention; they were detected in the Toarcian strata of the Tölgyhát quarry, support the presence of the presumed volcanic island arc in Árgyelán and Császár (1998) and seem to strengthen the assumption discussed previously in relation with the Úrkút area, namely that contribution of vitreous volcanic material must be presumed. At the same time pollen analyses from the Úrkút area and the presence of siliceous wood trunks also seem to support the island arc theory.

Based on our most recent knowledge, in the Alps-Carpathians-Dinarides region the earliest occurrence of dispersed Cr-rich detrital spinels is in Late Hettangian to Early Sinemurian sedimentary formations, with increasing frequency from the Toarcian (Árgyelán and Császár 1998).

Conclusions

A number of conclusions on genesis were deduced from the analysis of the Úrkút manganese ores and on their previously-described occurrence. The structure of the ore section, the textures and mineral composition, the geochemistry of the major components, heavy metals and REE, as well as the paleofacies characteristics of the Liassic basin in Central Europe, were taken into account.

Consequently, on the basis of the evaluation of the mineral composition, geochemical and other analyses of the Upper Liassic black shale-hosted Mn-carbonate ore formation in the Úrkút Basin, it cannot be excluded that during the formation, fine-grained vitreous volcanic materials were added. Transformation products of this vitreous tuff material, i.e. zeolites (clinoptilolite, etc.), smectite, potash feldspar as well as its trace element content affected the general features of the deposit. This recognition allows an evaluation with fewer contradictions, explaining at the same time previously hardly interpretable features.

The features below support this proposition, or are not in contradiction with it:

– the laminar distribution of the green clay components in the black shale-hosted carbonatic Mn-ore formation;

– joint occurrence of Fe-rich clay minerals (nontronite, celadonite), zeolite (clinoptilolite, heulandite, gismondite?...) and potash feldspar in the black shale and Mn-ore samples;

– the presence of strongly altered basic plagioclase minerals as well as the basic igneous and metamorphic heavy mineral assemblage in the marl underlying the Mn-ore formation (Szabó-Drubina 1957, 1959, 1961);

– Al-depletion of the authigenic clay minerals of the black shale;

– a relatively high Ti-content compared to poor sedimentary systems without any terrigenous contribution;

– the peculiar, delayed reduction of Fe occurring only during diagenesis, where increasing the Fe-content of clay minerals causes the second separation step of Mn and Fe;

– the access of SiO_2;

– occurrence of idiomorphic zeolite (clinoptilolite) crystals in rounded, Mn-bearing calcite fragments from the black shale horizon.

The formation process of authigenic smectite and zeolite presumed in submarine sediments cannot be neglected, in the course of which biogenic silica and Fe-oxyhydroxide, as well as poorly crystalline or X-ray amorphous alumo-silicates of volcanic origin and biogenic silica react with each other.

The alteration process of vitreous volcanic material as well as the separation of the products deriving from volcanic and biogenic processes, with special regard to the micromineralogical analysis of the Úrkút formations, require further investigation.

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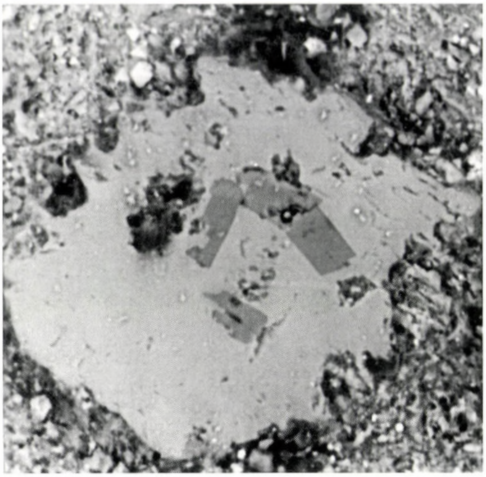
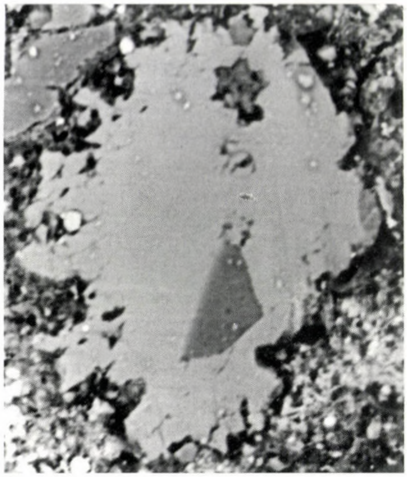
Plate I

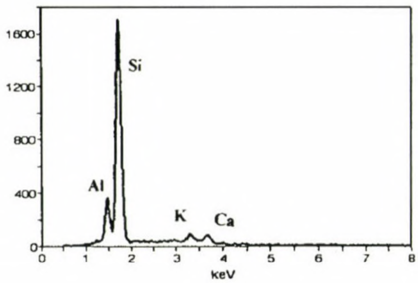
Photos of electron-microprobe study on radiolarian clayey-marlstone black shale. a), b), c), d) are backscattered electron images of samples from the black shale from Profile 2. In the granular groundmass of Si, Al, K, Fe the light rounded grains fragments are of Ca(Mn) content in which dark-grey idiomorphic minerals of Si, Al, Na, K, Ca composition occur zeolite, clinoptilolite. The photos and the EDS spectrum were made by Dr. Géza Nagy, Lab. for Geochem. Res., H.A.S., 1999)
The EDS spectrum was made from the mineral signed by arrow.

Plate I

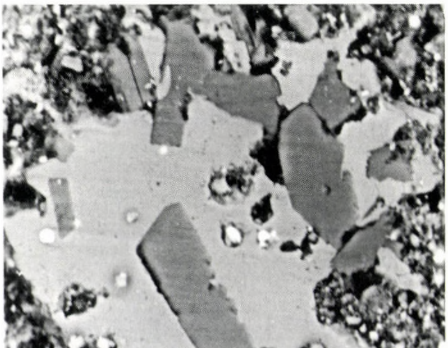


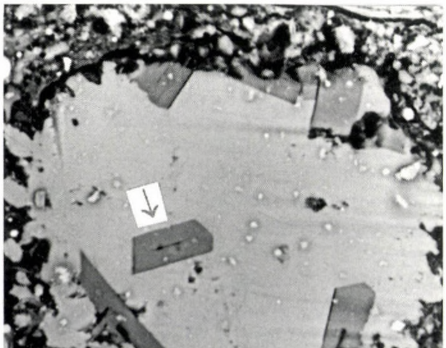


a) **b)**



c) ——— **= 100 µm** **d)**





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Preliminary results on Jurassic and Lower Cretaceous formations in the Karavanke Mountains and Lienz Dolomites, Austria

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The paper summarizes the preliminary results of field studies carried out in the Northern Karavanke Mountains and the Lienz Dolomites during the last few years. Two sections have been measured in the Lienz Dolomites (Figs 3, 4) and eight in the Northern Karavanke Mts (Figs 5, 6, 8, 10–14). While briefly describing the sections, two formation names are introduced: the Toarcian Kisgerecse Marl Formation, well known in the Transdanubian Range and the Toarcian (or Lower Dogger) Wildenstein Breccia Formation, named after its type locality.

Two submarine high(s) and an intervening basin are outlined and their evolution is evidenced from the late Early Jurassic till the Late Jurassic.

As preliminary results, a few megafossil taxa from the Lower Jurassic (bivalve, gastropod, brachiopods) are listed in the paper, also indicating their ecological and stratigraphic importance. Some newly collected Upper Jurassic (Tithonian) ammonites are also listed and figured.

Keywords: Karavanke Mts, Lienz Dolomites, Jurassic–Lower Cretaceous, Submarine high, Tithonian ammonoids

Introduction

For Hungarian geologists it was always very important to know as much about the geology of the Eastern Alps as possible, for several reasons: a) from the very beginning, geologic research in Hungary had been largely built on the geologic knowledge obtained in the Alpine area, b) the Alps are an open book to read compared to the flat and partly covered areas of the Pannonian region, c) it was suggested that the Pelso unit was a direct continuation of the Drau Range and that it originated from between the Northern Calcareous Alps and the Southern Alps (Kázmér and Kovács 1985).

Whereas the paleogeographic reconstruction of the Paleozoic and Triassic formations has been based on widely accepted, well-distinguishable facies zones, the distribution of the Jurassic formations within the Tethyan realm shows more uniform character, or at least a less regular pattern. The main facies differences within the Tethyan shelves depend basically on the local morphology of the

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bottom: in the basinal part more or less complete but thin, reduced successions, on the highs lacunose, even more condensed sequences have been deposited. As a consequence more detailed information is needed for finding the original facies connections within the Jurassic than the Triassic of the Tethys. In principle the distribution of the Lower Cretaceous formations offers a better chance for recognition of the regularities in the facies distribution, but owing to the subsequent erosion at the beginning of the Late Cretaceous only a very small proportion of the successions has been preserved.

During the last two decades, Austrian geologists paid limited attention to the Jurassic and Cretaceous formations of the Drau Range area. In addition to geologic maps and explanatory notes (Bauer 1981; Bauer et al. 1983) there are only a few papers dedicated to the Karavanke Mts (Bauer 1970; Holzer and Polting 1980; Schröder 1988; Kuhlemann et al. 1993; Császár and Dosztály 1994; Krystyn et al. 1994; Schlaf 1995; Császár et al. 1998) and the Lienz Dolomites (Faupl 1977; Blau and Schmidt 1988; Blau and Meister 1991; Blau and Grün 1995). The geology of the Karavanke Mts and the Julian Alps attracted the attention of the Hungarian geologists due to their supposed direct paleogeographic links with the Transdanubian Range (Kázmér and Kovács 1985; Schmidt et al. 1991). This is the reason why the present study, promoted by the Geological Survey of Austria in the framework of joint research, is dedicated to this narrow tectonic zone along the Periadriatic Lineament.

Geologic setting

The Drau Range (Fig. 1) is composed of the following east-west oriented tectonic units: Northern Karavanke Mts, Gailtal Alps, Lienz Dolomites, Winnebach Kalkzug and Aschbach Horst in the Hochpustertal (Blau and Grün 1995). These units consist mainly of thrust sheets and nappe structures of northerly vergence called Basal Nappe ("Sockeldecke") and Obir Nappe by Stini (1938) and Basal Zone and Central Zone by Schröder (1988) in the Northern Karavanke Mts (Fig. 2). According to Tollmann (1977) the maximum overthrust (4 km) occurred during the Early Tertiary. Jurassic and Cretaceous formations are preserved in the Northern Karavanke Mts and the Lienz Dolomites only from among the units listed above. Owing to the Late Miocene culmination of the multiphase deformations (Bauer 1970; Siewert 1984) the nappes and imbrications composed of Middle and Upper Triassic formations piled up very high above the Jurassic and Cretaceous formations in the northern part of the zone. In addition to the nappes, a blanket of Triassic dolomite and limestone fragments derived from the Obir Nappe may also cover a considerable part of the narrow Jurassic and Cretaceous stripe. The dip of the younger Mesozoic formations is variable.

The Northern Karavanke Mts have similar stratigraphic and tectonic character in its Slovenian continuation (Ramovš and Rebek 1970; Mioč and Šribar 1975).

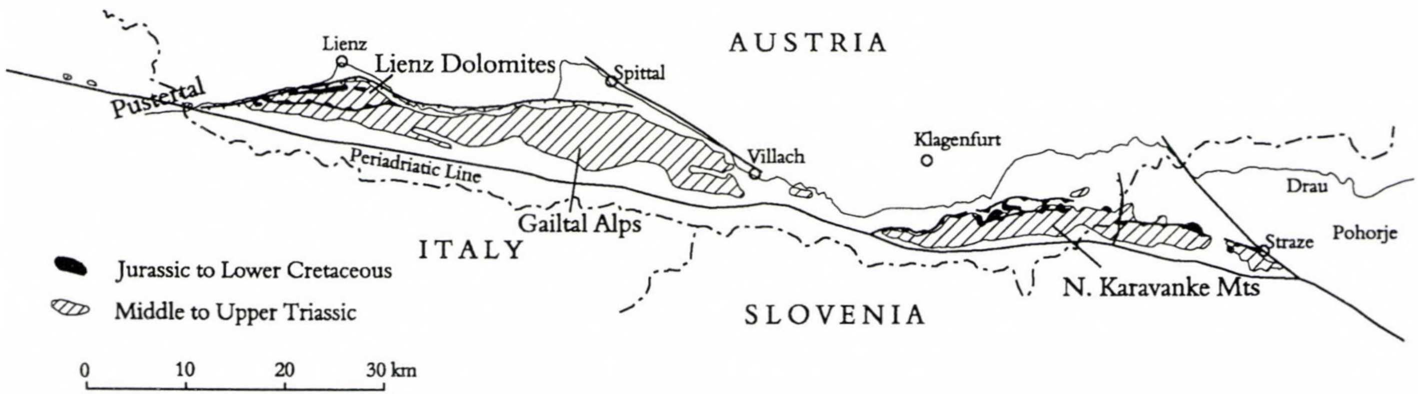

Fig. 1
Tectonic units and geologic sketch map of the Drau Range showing the extent of the Mesozoic formations (after Beck-Managetta and Braumüller 1964, simplified)

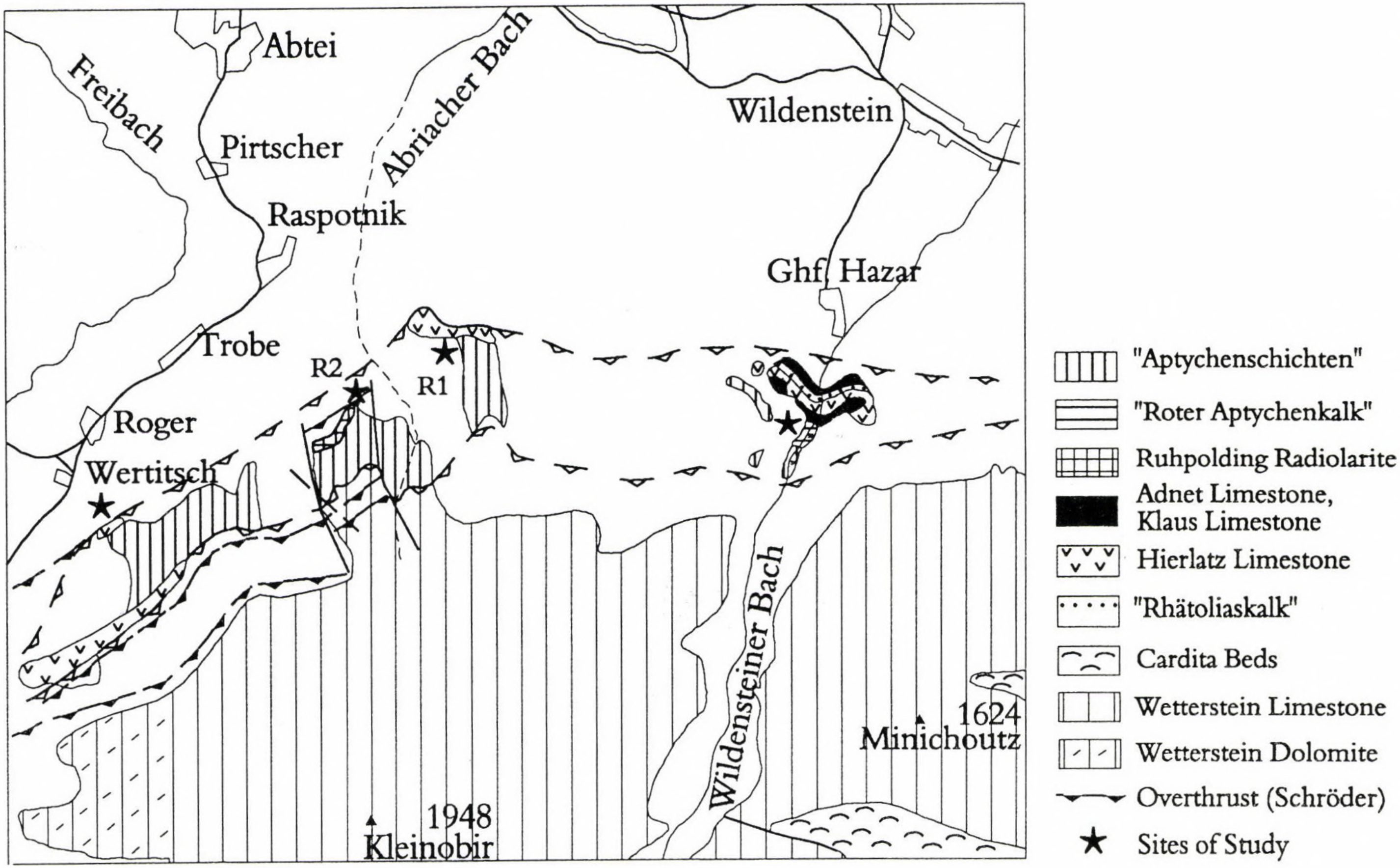

Fig. 2
Detail from the geologic map of the Northern Karavanke Mountains. Scale: 1:25 000 (after Bauer 1981, with the overthrust lines of Schröder 1988). R1 – site of Raspotnik section no 1; R2 – site of Raspotnik section no 2

Lithostratigraphy – description of the sections and localities under study

On the geologic map of the eastern part of the Northern Karavanke Mountains (Fig. 2; Bauer 1981) the following Jurassic and Cretaceous formations were distinguished: "Rhätoliaskalk" (Rhaetian to Liassic), "Hierlatzkalk" (Hierlatz Limestone – Liassic), "Adneter Kalk" (Adnet Limestone), "Klauskalk" (Lower to Middle Jurassic), "Ruhpoldinger Schichten" (Ruhpolding Radiolarite – lower Upper Jurassic), "Roter Aptychenkalk" (Red Aptychi Limestone – uppermost Upper Jurassic to Neocomian), "Aptychenschichten" (Aptychi beds – Neocomian), "Kalkbrekzie" (Albian). The Red Aptychi Limestone is subdivided into a "Saccocomakalk" and "Calpionellenkalk" by Bauer et al. (1983). They are called red Aptychi limestone s. str. and Biancone or Calpionella limestone respectively by Schröder (1988). He also named the Aptychi limestone Schrambachschichten (Schrambach beds). Császár and Dosztály (1994) noted Toarcian red marls (equivalent to the Kisgerecse Marl Fm. in the Transdanubian Range) with a rich ammonite assemblage at Raspotnik.

The Jurassic and Cretaceous formations in the Lienz Dolomites were described recently by Blau and Grün (1995), Blau and Meister (1991), Blau and Schmidt (1988). Blau and Grün (1995) distinguished a slope and basin and a ridge successions. In the former the following formations were recognized: "Lavanter Breccie" (? Hettangian–Sinemurian), "Bunte Kalke" (? Hettangian–Upper Sinemurian), "Allgäuschichten" (? Hettangian–Sinemurian), "Rotkalk" ("red limestone"–Pliensbachian to Upper Jurassic), "Biancone" (Upper Tithonian–Valanginian), "Kreidefleckenmergel" (Valanginian, Hauterivian–Aptian), Amlacher Wiesen Schichten (Aptian/Albian). According to Blau and Grün (1995) there are hardground(s) within the "red limestone" in the condensed ridge succession where no Middle Jurassic is evidenced.

The most significant difference between the Jurassic sequences in the Northern Karavanke Mts. and the Lienz Dolomites is that the Karavanke sequence is more calcareous and pelagic, whereas intercalation of the Allgäu Beds in the Lienz Dolomites indicates a stronger terrigenous influx.

Several Jurassic and Lower Cretaceous formations in the Drau Range area were either not known thus far or simply not named and described according to the rules of the international stratigraphic guide (Hedberg 1976; Salvador 1994). Since the sequences of the Transdanubian Range in Hungary are similar to those in the Drau Range, especially in the Northern Karavanke Mts, in the present paper we often use formal Hungarian lithostratigraphic names in addition to the informal local names, and also names defined in the Northern Calcareous Alps (NCA), if they are relevant (e.g. "Roter Aptychenkalk" = Pálihálás Limestone Fm or "Tegernseer Kalk", or "Agathakalk"). If a formation was thus far unknown in the Drau Range but was known in Hungary, the Hungarian formal name is used in addition to a name from the NCA (e.g. Kisgerecse Marl Fm. – "Saubachschichten"). If the newly recorded formation was previously not known in the NCA, or in the Transdanubian Range, a new local name is introduced as a formal lithostratigraphic one (e.g. Wildenstein Breccia Fm). It should be noted that the informal Jurassic names in Austria often traditionally carry a chronostratigraphic meaning that may cause trouble in their application. There are various names for formations of more or less the same lithology, just because their ammonite content and age are (in part) different. A good example is the Pálihálás Limestone Formation in Hungary, a "dark or light red limestone, «Ammonitico Rosso» type argillaceous, nodular limestone with Saccocoma and with variable bedding, maybe cherty" (Knauer 1997b). It is equivalent to the "Rotensteinkalk", the "Agathakalk", and the "Tegernseer Kalk", formations which differ only in their age: the first one is Oxfordian, the second one is Kimmeridgian and the third one is Kimmeridgian to earliest Early Tithonian.

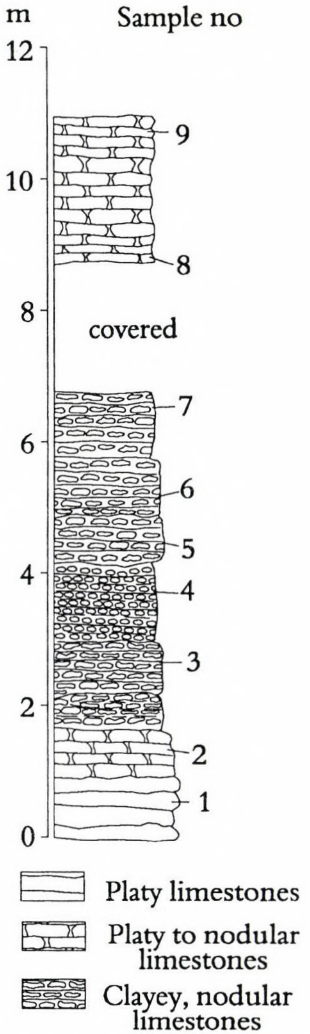

Fig. 3
Schematic columnar section (1) of the Lower Jurassic, Galitzen Klamm, Lienz Dolomites

Galitzen Klamm Section 1 and 2, Lienz Dolomites

There are only two short Lower Jurassic sections in the Lienz Dolomites (Fig. 4), which were rapidly surveyed in this project. According to Blau and Schmidt (1988) and Blau and Grün (1995) both of them can be placed within the basinal facies of the "Rotkalk" formation. Section 1 (Fig. 3) was measured at a rock fall, south of Galitzen Klamm, section 2 (Fig. 4) about 300 m away from the western end of the forest road going to Obertilliach. The basal and the topmost beds of section 1 are thick-bedded platy limestone of greyish and lilac brown color with fine-grained crystalline or micritic fabrics. The middle part of the section is made up of red or brownish-red, clayey, nodular limestone that turns both upsection and downsection into platy nodular limestone. No megafossils have been encountered in the section.

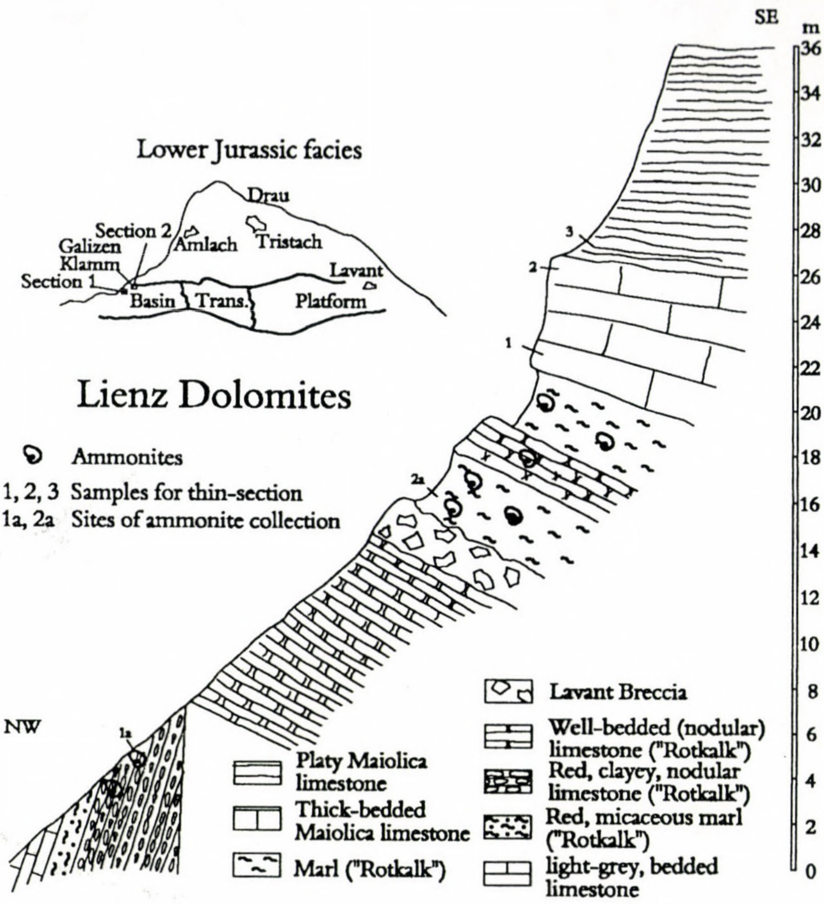

Fig. 4
Location map (after Blau and Grün 1995) and schematic lithological column (2) of the Jurassic, Galitzen Klamm, Lienz Dolomites

Section 2 is tectonically complicated. The steeply dipping lower part begins with limestone beds of grey color, followed by red, slightly micaceous marl and red, clayey, ammonite-bearing nodular limestone. Its heavy mineral content in % is as follows: epidote group: 7, garnet: 8, staurolite: 3, rhombic pyroxene: 6, clinopyroxene: 6, amphibole: 2, kyanite: 1, zircon: 1, rutile: 1, turmaline: 1, biotite: 5, chlorite: 4, hematite: 24, chromite: 1, magnetite: 26, pyrite: 2 (det. by B. Árgyelán). The association indicates mixed magmatic and metamorphic source rocks. Above the fault there is a well-bedded, grey or lilac limestone with red clay intercalations of a few cm thickness. The limestone is replaced by a limestone breccia bed in red, calcareous, clayey matrix (Lavant Breccia – Blau and Grün 1995). The overlying red marls and nodular limestone are rich in ammonites. The marl is overlain by thick-bedded white limestone and then by white, platy limestone. The last two limestone units may belong to the Maiolica, which is equivalent to the Mogyorósdomb Limestone Formation in the Bakony Mts.

Wertitsch, Northern Karavanke Mts.

The studied section is situated at the foot of a rather steep slope to the south of Wertitsch (Fig. 2). According to Bauer et al. (1981) the "Hierlatz Limestone" is in contact with the "Aptychi beds". The columnar section (Fig. 5 – Császár et al. 1998) clearly indicates a thick Lower Jurassic succession and a poorly exposed Middle and Upper Jurassic one.

The Lower Jurassic "Hierlatz Limestone" (Bauer et al. 1983) that can be compared to the Pisznice Limestone in the Gerecse Mts, can be subdivided into two lithological units. The lower 15-m interval (Wertitsch II in Fig. 5) is made up of pink or red, thick-bedded, crinoidal limestone of variable grain size with scattered manganese-coated intraclasts. The nodular character is subordinate. The upper half of the formation (12 m) consists of pink limestone of wackestone, occasionally mudstone, texture with scattered, fine-grained crinoid ossicles. The uppermost half meter is a clayey limestone with abundant manganese-coated intraclasts. The peculiarity of the formation is the lack of other macrofossils. Based on this fact – even if the redefinition of the Hierlatz Formation has not yet been completed (Vörös 1991; Böhm et al. 1998) – the authors suggest using another lithostratigraphic name for this succession because its composition does not correspond to the definitive pattern of the Hierlatz Formation (Trauth 1950). Lithostratigraphically it could be better correlated with the Pisznice Limestone Formation, described in the Gerecse Mts, Hungary (Császár et al. 1998).

The above-described beds are followed by a formation not indicated in the latest geologic map by Bauer (1981). It is a brownish-red, clayey, nodular limestone that can be correlated with the Adnet Limestone, described from the Karavanke Mts by Schröder (1988).

The upper 3 m of the Wertitsch section is mostly covered by soil and vegetation; therefore the contacts of the formations are uncertain. Its lower part is formed by a 1 m-thick breccia unit. The rock fragments of this breccia reach up to 30 cm in size and derive from the yellowish-grey, oncoidal "Rhätoliaskalk" (similar to the Kardosrét Limestone Fm. in the Bakony Mts, Hungary). The matrix is a red, mudstone-type limestone with a few Bositra shells. This type of breccia is not unique in the Karavanke Mountains; it was first reported from the Wildenstein Valley (Császár et al. 1998). Therefore we propose calling it Wildenstein Breccia Formation. Although the precise age of the formation is not known, the Bositra shells suggest that it cannot be older than Toarcian. So far no other breccias are known from the Northern Karavanke Mts. Early Jurassic (Hettangian to Sinemurian) breccias were reported from the Lienz Dolomites (Blau and Schmidt 1988; Blau and Grün 1995). This kind of breccia also occurs in the

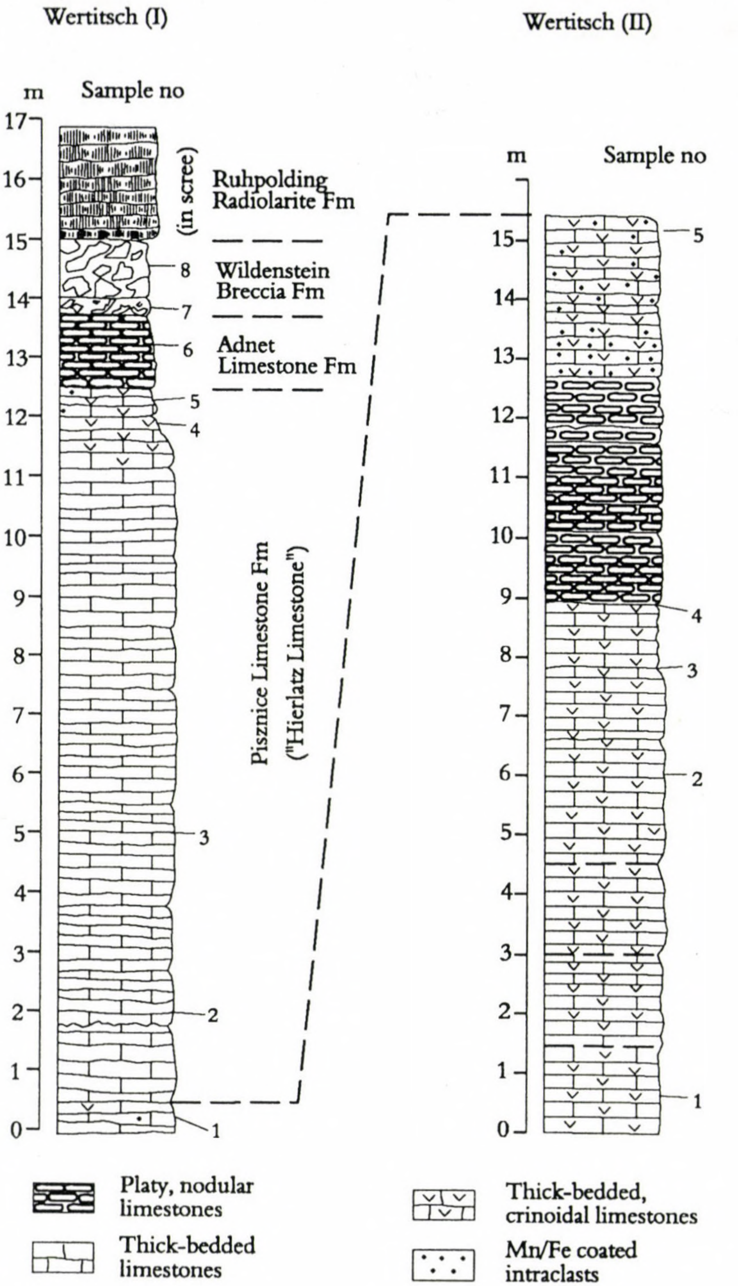

Fig. 5
Composite lithological column of the Jurassic sequence at Wertitsch, Northern Karavanke Mountains

Pliensbachian and Bajocian of the Transdanubian Range (Galácz and Vörös 1972; Vörös 1986; Vörös and Galácz 1998).

The section is capped by chert fragments of Ruhpolding Radiolarite (Fig. 5) found as scree on the surface. As chert fragments can move over considerable distances down the slope the supposed thickness of the Wildenstein Breccia may be underestimated in the columnar section.

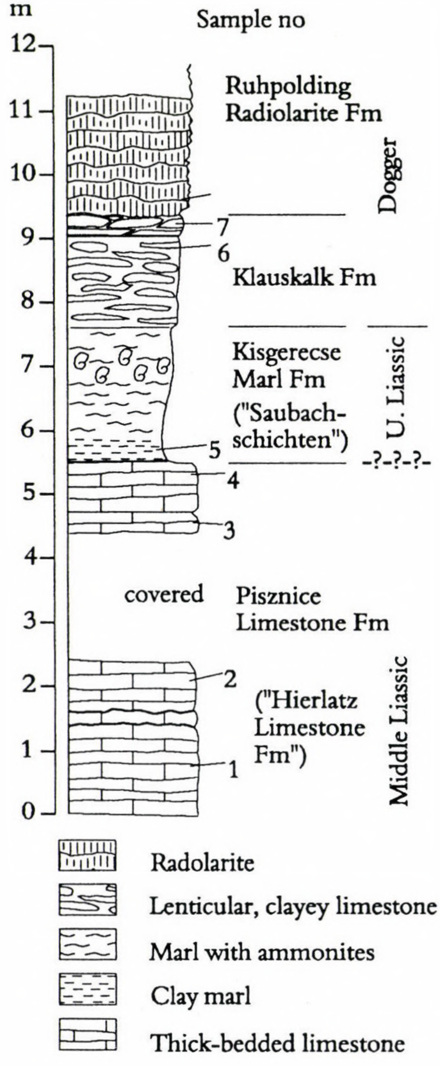

Fig. 6
Columnar section of the Lower to Middle Jurassic sequence in a road-cut, Raspotnik, Northern Karavanke Mountains (R 1 in Fig. 2)

Raspotnik section no. 1

Bauer (1981) indicates a direct contact between the "Hierlatz Limestone" and the Lower Cretaceous Aptychi beds on the detailed geologic map (Fig. 2). In the section (Fig. 6) the upper part of the "Hierlatz Limestone" crops out. Although it resembles the lower unit of the Wertitsch section its topmost beds, of slightly nodular character, contain a few ammonites and Fe/Mn-coated intraclasts in the micritic matrix; therefore it is better to call it either Pisznice Limestone or Adnet Limestone.

In an artificial trench, Császár and Dosztály (1994) discovered new formations above a hardground surface of the Pisznice or Adnet Limestone. Its lower 2-meter part consists of calcareous clay and marl and contains a rich ammonite assemblage belonging to the Hildoceras bifrons Zone (Géczy in Császár and Dosztály 1994; Császár et al. 1998). As in the Karavanke Mts this significant formation (indicating a rapid subsidence, and/or an overall sea level rise) was not specified or named, Császár et al. (1998) suggested using the name Kisgerecse Marl Formation, defined in the Transdanubian Range (Fig. 7) within the same tectonic zone. "The Kisgerecse Marl is a bright red, Ammonitico Rosso-type marl, and marl with limestone nodules" (Knauer

1997a). The "Saubachschichten", described from the NCA by Plöchinger (1982) and also by Böhm (1992), may be an equivalent formation.

The mineralogical composition of this formation at Raspotnik in wt.% is as follows: calcite: 61, (67), montmorillonite: 4 (3), illite/montmorillonite: (3), illite: 11 (8), kaolinite: 4 (4), quartz: - (9), K-feldspar: - (1), hematite: - (2), goethite 1 (trace), gypsum - (1). amorphous - (2) (the first numbers show the results of thermal analyses made by M. Földvári, numbers in brackets are results of X-ray analyses made by P. Kovács-Pálffy, both in the laboratory of the Geological Institute of Hungary).

The Kisgerecse Marl Formation is followed by a 2 meter-thick red marl and clayey limestone of lenticular bedding without macrofossils. Because of the gradual transition, it might be part of the Kisgerecse Marl (or "Saubachschichten") but more probably it is a clayey member of the "Klauskalk". These marly units are directly overlain by the Ruhpolding Radiolarite.

Raspotnik section no. 2

The section is situated south of the settlement along a road-cut, higher up the slope (Fig. 2). The section (Fig. 8) offers a good opportunity to study the transition between the following formations: Ruhpolding Radiolarite, Saccocoma limestone (equivalent to "Tegernseer Kalk" in the NCA and Pálihálás Limestone in the Transdanubian Range) and the Maiolica

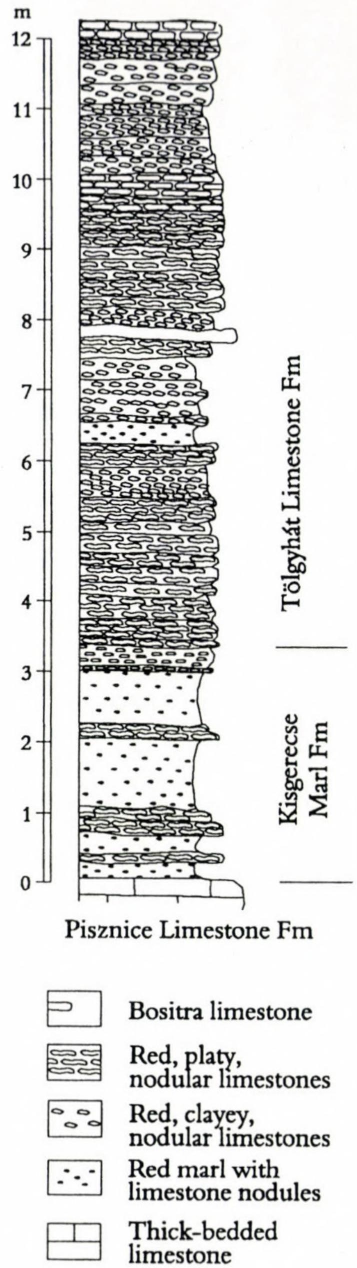

Fig. 7
Type-section of the Toarcian Kisgerecse Marl Formation, Mount Kis-Gerecse, Gerecse Mts, Hungary

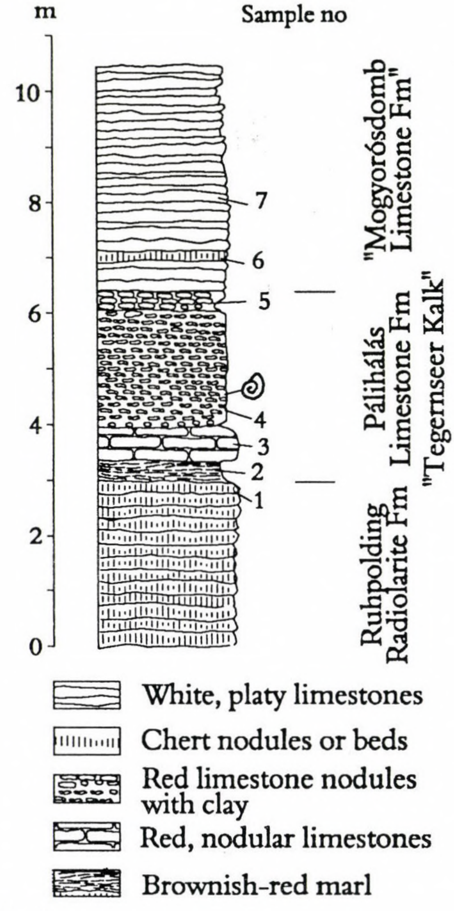

Fig. 8
Upper Jurassic at a forest road-cut, Raspotnik, Northern Karavanke Mountains (R2 in Fig. 2)

(Mogyorósdomb Limestone in the Bakony Mountains).

The upper 3 meter-thick part of the Ruhpolding Radiolarite is visible in the section. It consists of 10 to 15 cm-thick brownish-red cherty beds with thin clayey interlayers. The Radiolaria assemblage of the formation gave an Oxfordian age, according to Dosztály (in Császár and Dosztály 1994). The age of the radiolarite at Unterort (a few kilometers east of Wildenstein) is Kimmeridgian, providing evidence for heterochronous onset and termination of radiolarite sedimentation in the Karavanke Mts, just as in the Transdanubian Range.

The Pálihálás Limestone begins with a 40 cm-thick brownish-red marl layer followed by a 60 cm brownish-red, grey-spotted, nodular limestone interval. The remaining 2.5 m of the formation is lilac-red, clayey, nodular limestone with an ammonite-rich horizon. This level yielded a few very poorly preserved phylloceratids, numerous fragments of lytoceratids (most likely belonging to *Protetragonites quadrisulcatus* (D'Orbigny 1840), and some densely ribbed, strongly dissolved perisphinctids, insufficient for precise determination. Two specimens were determined as *Hybonoticeras hybonotum* (Oppel 1863). This species is the zonal index of the widely-used Hybonotum Zone of the lowermost Tithonian. The specimens were relatively large fragments of body chambers. The strong, distant, simple or bifurcating ribs, the ventrolateral tubercles and the bicarinate, concave venter are visible.

In the scree, another important ammonite, *Pseudowaagenia acanthomphala* (Zittel 1870) was found. The specimen – a moderately well-preserved body chamber – shows the typical internal row of tubercles; the flanks are flat and the venter is rounded with a flat band in the middle. According to Checa (1985) the species is characteristic for the middle-upper part of the Kimmeridgian, with a maximum abundance in the uppermost Kimmeridgian Beckeri Zone.

The upper half meter of the formation shows characteristics transitional to the Maiolica. This bed is reddish pale-grey, sometimes greenish in color and less nodular but lenticular, and resembles the "Anzenbach Schichten" in the NCA.

The Maiolica (equivalent to the Mogyorósdomb Limestone in Southern Bakony, Hungary and "Schrambach-schichten" – Schröder 1988) is here a greyish-white platy, micritic limestone with dark-grey chert nodules or thin beds cropping out only in 4 m of thickness.

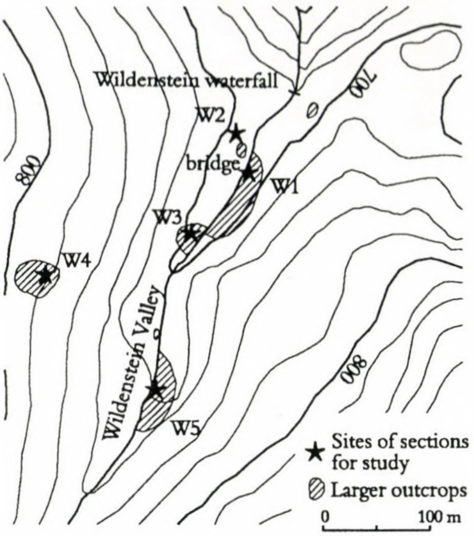

Fig. 9
Wildenstein Valley with outcrops of Jurassic and Lower Cretaceous formations and with the indication of the sites of sections studied, Northern Karavanke Mountains

Wildenstein Valley, section W 1

According to Bauer (1981), in the Wildenstein Valley and its closer surroundings (indicated as Wildersteiner Bach in Fig. 2) the Adnet Limestone is directly overlain by the "red Aptychi beds". In the creek level, close to the wooden bridge, a breccia unit was discovered in red, muddy and calcareous matrix (Császár and Dosztály 1994; Császár et al. 1998). The maximum size of the breccia components exceeds 1 m. The prevailing rock type of this Wildenstein Breccia is the "Rhätoliaskalk" but a few Lower Jurassic crinoidal limestone fragments of smaller size were also found. In an artificial trench at the base of the small cliff on the eastern side of the creek (Figs 9, 10), in addition to the Ruhpolding Radiolarite, the Wildenstein Breccia was also revealed below the Saccocoma limestone. Here the breccia contains smaller rock fragments (2–25 cm). The very poor heavy mineral content of the clayey, marly matrix of the Wildenstein Breccia in % is as follows: epidote group: 0.7, garnet: 7.1, staurolite. 2.9, rhombic pyroxene: 2.9, amphibole: 0.7, zircon: 0.7, magnetite: 85 (det. by B. Árgyelán).

The thickness of the brownish-red, platy radiolarite with thin clay intercalations is less than 1 m. The Ruhpolding Radiolarite is separated by a few cm-thick red clay of lenticular bedding from the Saccocoma limestone (Pálihálás Limestone). The red or brownish-red limestone is micritic, often fine-grained biodetrital, occasionally with red chert nodules or just siliceous impregnation. The lower part of this limestone yielded the following megafossils: *Ptychophylloceras ptychoicum* (Quenstedt 1847), *Phylloceras* sp., *Protetragonites* *quadrisulcatus* (D'Orbigny 1840), *Haploceras carachtheis* (Zeuschner 1846), *Virgatosimoceras micrum* Olóriz 1978, *Lamellaptychus* sp., *Punctaptychus* sp., *Triangope* sp. Approximately 3 m above the radiolarite in addition to phylloceratids and lytoceratids (*Protetragonites* sp.), *Haploceras* sp. and *Lemencia* sp. were found (the fossiliferous layer is some 0.5 m above the top of the column in Fig. 10 and indicated in the column of Fig. 6 in Császár et al. 1998). All fossils are poorly preserved, fragmentary and strongly dissolved. Ammonites are internal moulds, only the brachiopod shells and aptychi are preserved.

The lower assemblage (including the relatively well-preserved *Virgatosimoceras*, an adult, nearly complete body chamber) is characteristic for the lower part of the Lower Tithonian, probably for the Darwini Zone of Enay and Geyssant (1975). In certain horizons of this level, small aptychi are abundant.

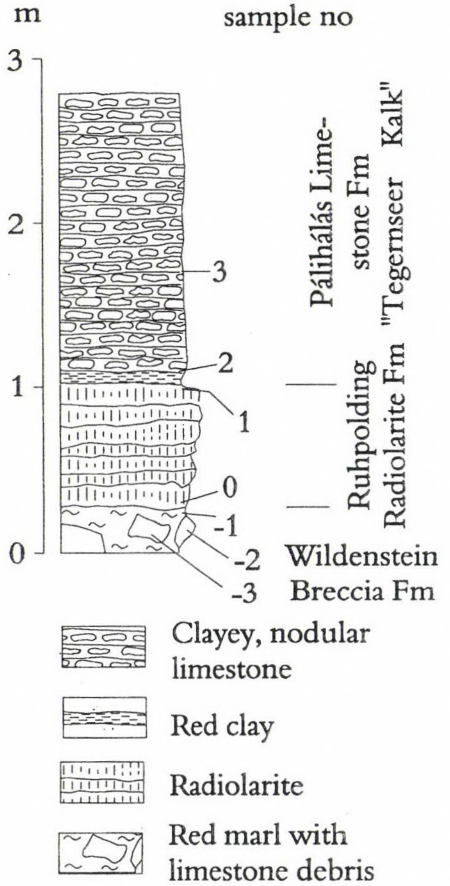

Fig. 10
Columnar section with Jurassic formations at the bridge, Wildenstein Valley, Northern Karavanke Mountains (W1)

Lemencia sp., collected higher in the section, represents the upper part of the Lower Tithonian – probably Ponti Zone (Olóriz 1978).

The section is clear evidence for the lack of "Klauskalk" in the valley. On the other hand both the Wildenstein Breccia and the Ruhpolding Radiolarite occur.

Wildenstein Valley W 2

The section is located on the western side of the valley, close to the base level (Fig. 9). It exposes the topmost beds of the Pálihálás Limestone "Tegernseer Kalk" and the lower beds of the Szentivánhegy Limestone. The upper beds of the succession can be considered as a transitional facies between the "Haselbergkalk" and the "Mühlbergkalk" (Fig. 11). The former is a fine-grained, reddish or pale-grey, nodular limestone with a few red chert nodules, while the latter is a micritic and stylolitic, lilac, reddish or white limestone with scattered crinoidal ossicles. The significance of this section is that the Upper Jurassic and lowermost Cretaceous part of the Calpionella limestone is biodetrital only here (akin to the Szentivánhegy Limestone in the Transdanubian Range) replacing the widespread Maiolica-type Mogyorósdomb Limestone that is barren of biodetritus.

Wildenstein Valley W 3

This section was measured on the western side of the valley, approx. 170 m to the south of the waterfall (Fig. 9). The 21 m-thick section exposes the basal part of the Neocomian Maiolica (equivalent to the Mogyorósdomb Limestone Formation) consisting of well-bedded, mainly platy, occasionally laminated micritic limestone (Fig. 12). Its lower 2.1 m thick part is pale lilac or brownish-lilac in color resembling the "Anzenbach Schichten" in the NCA, and provided a few poorly-preserved ammonites (*Berriasella* sp.). Above this level the color turns into light grey or greenish grey. Dark grey clay laminae or seldom cm-thick layers intercalate up to 4 m. Black or dark grey chert nodules are restricted to the 2.1 and 6.2 m interval. The thicker upper part of the section (approx. above 6 m) is called "Schrambachschichten" by Schröder (1988). Due to its considerable thickness the formation is well represented and recognized in the Northern Karavanke Mts. (Suette 1978; Holzer and Suette, in Bauer et al. 1983; Császár and Dosztály 1994) and also in the Lienz Dolomites (Blau and Grün 1995).

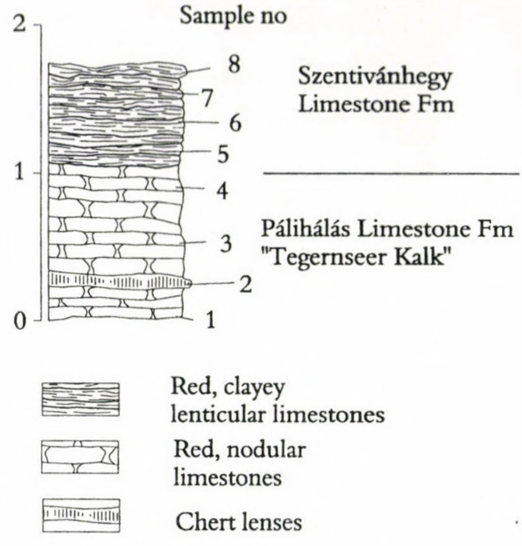

Fig. 11
Transitional beds of the Saccocoma limestone (Pálihálás Limestone Fm.) and the Maiolica beds (Szentivánhegy Limestone) at the bridge, Wildenstein Valley (W 2)

Wildenstein Valley W 4

The section is situated high on the western slope of the valley (Fig. 9) and it forms an upsection continuation of the Mogyorósdomb Limestone introduced in the previous section (W 3). Lithologically the sequence of this section (Fig. 13) is similar to that one in section W 3 but the stratification is more varied: there are several thicker beds and the laminated intercalations are also rather frequent. Black chert nodules are restricted to the lower 6 m and bioturbation, that made the limestone spotty, can be recognized throughout the sequence.

Wildenstein Valley W 5

The section (Fig. 14) represents the uppermost part of the Lower Cretaceous sequence of the Northern Karavanke Mts (Fig. 9). Between the grey or dark grey limestone beds dark grey clay and marl intercalations become more and more frequent upsection. The entire section was also assigned to the "Schrambach-schichten" by Schröder (1988). According to Császár and Dosztály (1994) these

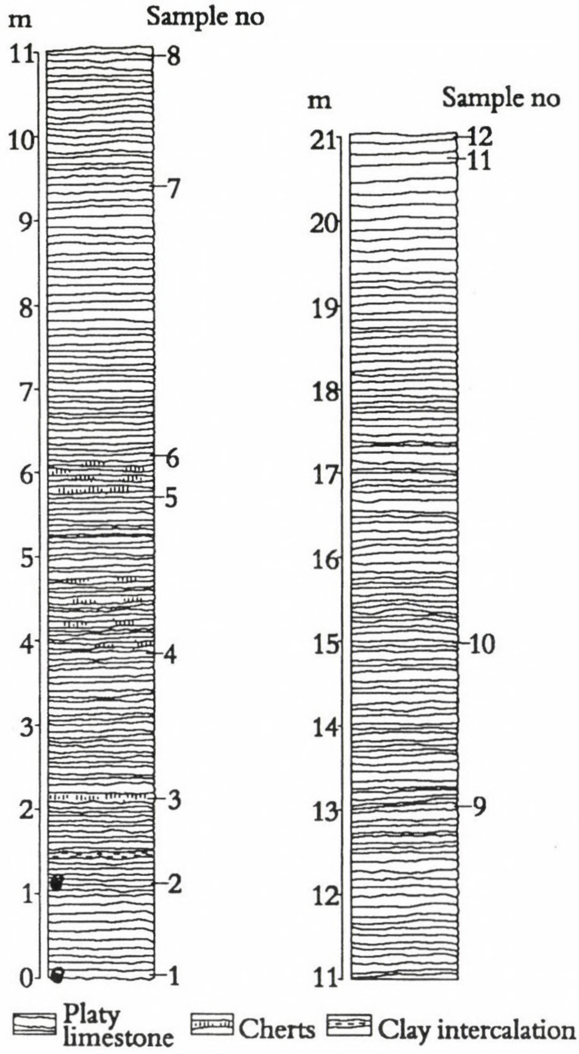

Fig. 12
Columnar section of the (Maiolica) Mogyorósdomb Limestone Fm. in the Wildenstein Valley (W 3)

beds correspond to the Sümeg Marl Formation in the Bakony Mts. By definition the Sümeg Marl is a grey, occasionally greenish, or brownish silty marl, siltstone and calcareous marl of shallow bathyal basin facies with sandy intercalations in the upper part in some cases with a significant amount of ammonites, radiolarians and planktonic foraminifers. The bottom part is calcareous marl, whereas the middle part is predominantly calcareous siltstone or silty marl [Császár and Haas in Császár (ed.) 1997].

Wildenstein Valley W 6

This is not a measured section but a locality, where particular and distinctive rock types, otherwise unknown in the area, occur. On the top of a nearby hill east of the Wildenstein Valley, a few large boulders have been found. One of them, a red, micritic, finely crinoidal, massive limestone block provided the following fossils: plenty of tiny gastropods (*Ataphrus*? sp., det. by J. Szabó); a dozen small brachiopods (*Koninckodonta* sp., *Phymatothyris cerasulum* (Zittel 1869), det. by A. Vörös) and one bivalve (*Praechlamys* sp., det. by I. Szente). This fauna points to a Pliensbachian age. The other block of pink, micritic limestone contained cm-thick, black Fe-Mn-oxide crusts and a large specimen of Lytoceras sp., corroded on one side and encrusted by manganese oxide. Though this ammonoid taxon has a very long stratigraphic range, this limestone can probably be assigned to the Toarcian or Bajocian, by analogy with similar rock types frequently found in the condensed sections of the Bakony Mts.

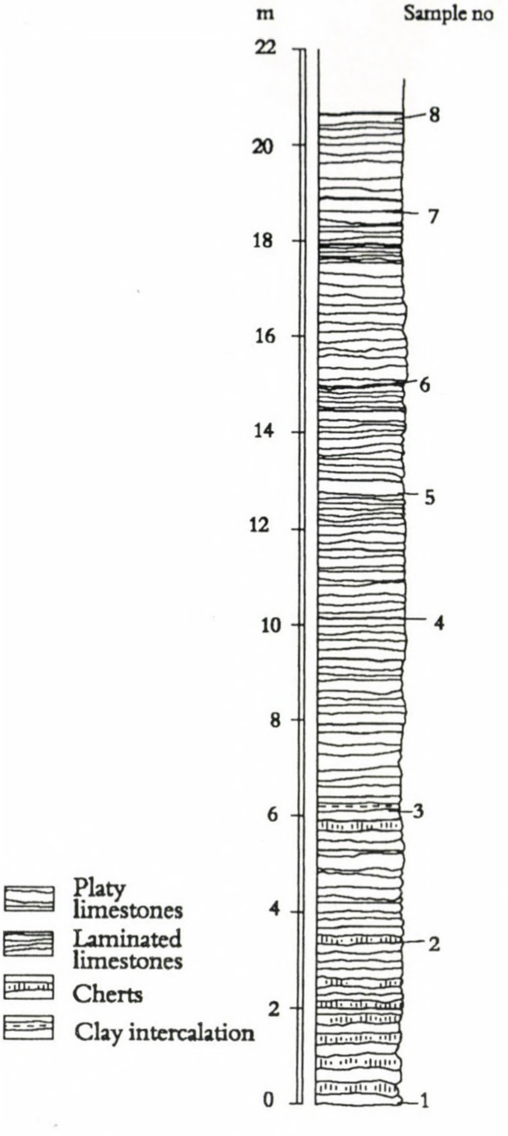

Fig. 13
Columnar section of the (Maiolica) Mogyorósdomb Limestone in the Wildenstein Valley (W 4)

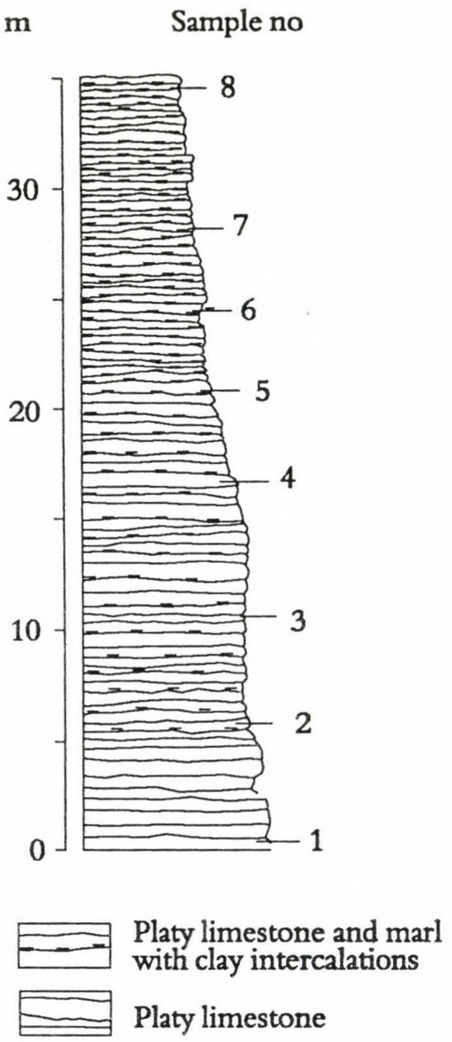

Fig. 14
Uppermost beds of the Lower Cretaceous sequence in the Wildenstein Valley. Transitional beds to the Sümeg Marl Fm. (W 5)

Discussion

The lowermost formation of the Jurassic, the Pisznice Limestone ("Hierlatz Limestone" of earlier authors) is widespread all over the region and shows relatively little thickness variation. The frequent crinoid ossicles of the limestone indicate a relatively shallow submarine source area. The topmost part of this limestone sequence is usually more clayey, reddish and nodular and may turn into the Adnet Limestone (see Wertitsch column I – Fig. 5), which may suggest a gradual deepening of the sedimentary basin. The predominance of less clayey layers within the Wertitsch section may indicate a short-term and less significant sea-level rise during the Early Jurassic epoch.

The silty and micaceous Lower Jurassic limestone units in their westernmost occurrence in the Drau Range (Lienz Dolomites) is a clear indication of a relative proximity of the land that is supposed to have been located to the west or northwest.

A sudden lithological change can be recorded in the Raspotnik R 1 column (Fig. 6) at the beginning of the Toarcian where the clayey, Kisgerecse Marl ("Saubachschichten") appears. It can be substituted by the coeval Wildenstein Breccia. Rhätoliaskalk extraclasts of the breccia occur in red, marly matrix in the Wildenstein Valley (Fig. 10) and in red calcareous marl and limestone matrix at Wertitsch (Fig. 5).

So far two breccia horizons have been recorded in the Jurassic sequences of the Drau Range area. The older one, the Lavant Breccia (Hettangian–Sinemurian, sometimes early Pliensbachian – Blau and Grün 1995), is restricted to the Lienz Dolomites; the younger one, the Toarcian (or perhaps Bajocian?) Wildenstein Breccia, occurs in the Northern Karavanke Mountains. The breccias in both cases are clear evidence for the tectonic differentiation of the basement within the sedimentary basin. The peculiarity of the phenomenon is the different age of the formation of breccia in two neighboring tectonic units within the Drau Range. In both tectonic units, the appearance of the breccia is in connection with changes in lithology. In the sequence of the Lienz Dolomites "the Bunte Kalke" and Allgäu beds are replaced by the red, clayey, nodular "Rotkalk", whereas in the Northern Karavanke Mts the "Hierlatz Limestone" or Adnet Limestone is followed by the "Saubachschichten" (Kisgerecse Marl). The lithological change indicates a deepening of the sedimentary basin that can be attributed primarily to tectonic subsidence. The overall and crucial lithological change at the beginning of the Toarcian might express a combined effect of a tectonic subsidence and a eustatic sea-level rise.

The Wildenstein Breccia indicates the margin of a submarine high at the Wildenstein Valley and at Wertitsch as well, from where the rock fragments of Rhätoliaskalk were transported gravitationally to the foot of the submarine high. There is evidence for an edge of a submarine high on the eastern side of the Wildenstein Valley. The rock types and fossils in the boulders found on the top of a nearby hill represent the facies characteristic to the top of submarine highs. The brachiopods (especially the koninckodonts) are typical for the submarine highs in the Transdanubian Range (Vörös 1986). According to Szabó (pers. comm.) the gastropods were herbivorous and lived in the photic zone. The Toarcian (or Bajocian) limestone, with corroded and Mn-oxide-encrusted ammonites, is also characteristic to the condensed facies of the submarine highs. This proves a submarine high environment on the eastern side of the Wildenstein valley from the Pliensbachian to the Toarcian (or perhaps to the Bajocian).

The lack of "Klauskalk" and the unusually thin radiolarite in the Wildenstein Valley (Fig. 10) is in accordance with the idea of a neighboring submarine high during the Middle Jurassic.

The lateral facies change in the Late Jurassic (crinoidal Szentivánhegy Limestone in the Wildenstein Valley sections instead of the eupelagic Maiolica facies in the more western areas) is another indication of a submarine high (as source of the coarser biodetrital material) to the east of the Wildenstein Valley.

Because the Jurassic beds are exposed only in a narrow strip in the Northern Karavanke Mts, the above data are insufficient for a detailed, local paleogeographic reconstruction. The submarine highs to the east of Wildenstein Valley and at Wertitsch could be two independent highs or parts of a larger, dissected one. In any case it can be stated that, between the two sites, at Raspotnik, a basinal succession was deposited throughout the entire Jurassic period.

Except for the Wildenstein Breccia, which is a new phenomenon at the top of the Lower Jurassic, both complete and incomplete Jurassic and also Lower Cretaceous successions of the Northern Karavanke Mountains highly resemble the successions of the South Bakony Mountains (Fig. 15).

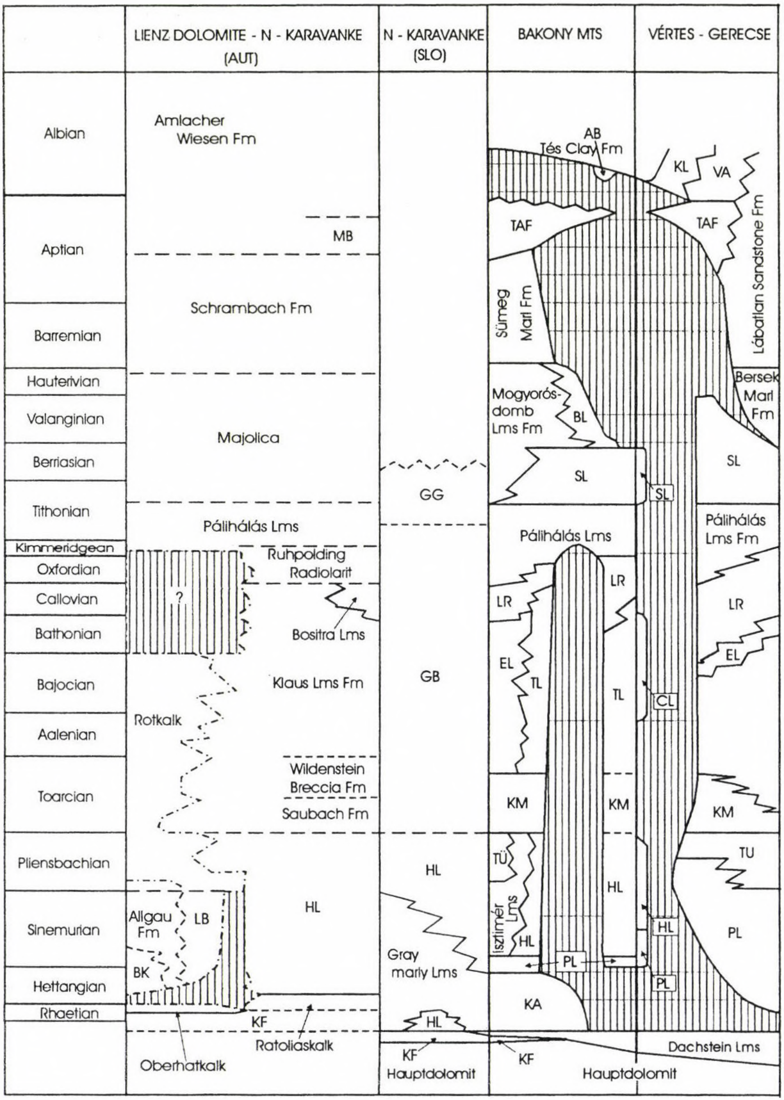

Fig. 15

Correlation chart of the Jurassic to Lower Cretaceous sequences between the Drau and Transdanubian Ranges. AB – Alsópere Bauxite Fmé BK – "Bunte Kalke"; BL – Borzavár Limestone Fm; CL – Csókakő Limestone Fm; EL – Eplény Limestone Fm; GB – Garyish-green and reddish brown platy, marly limestone; GG – Grrenish-gray and reddish platy limestone; HL – Hierlatz Limestone; KA – Kardosrét Limestone Fm; KF – Kössen Fm; KL – Környe Limestone Fm; KM – Kisgerecse Marl Fm; LB – Lavant Breccia; LS – Lábatlan Sandstone Fm; LR – Lókút Radiolarite Fm; MB – "Microbreccia" (N. Karavanke); PL – Pisznice Limestone Fm; SL – Szentivánhegy Limestone Fm; TAF – Tölgyhát Limestone Fm; TL – Tölgyhát Limestone Fm; TŰ – Tűzkövesárok Limestone Fm; VA – Vértessomlói Aleurolit

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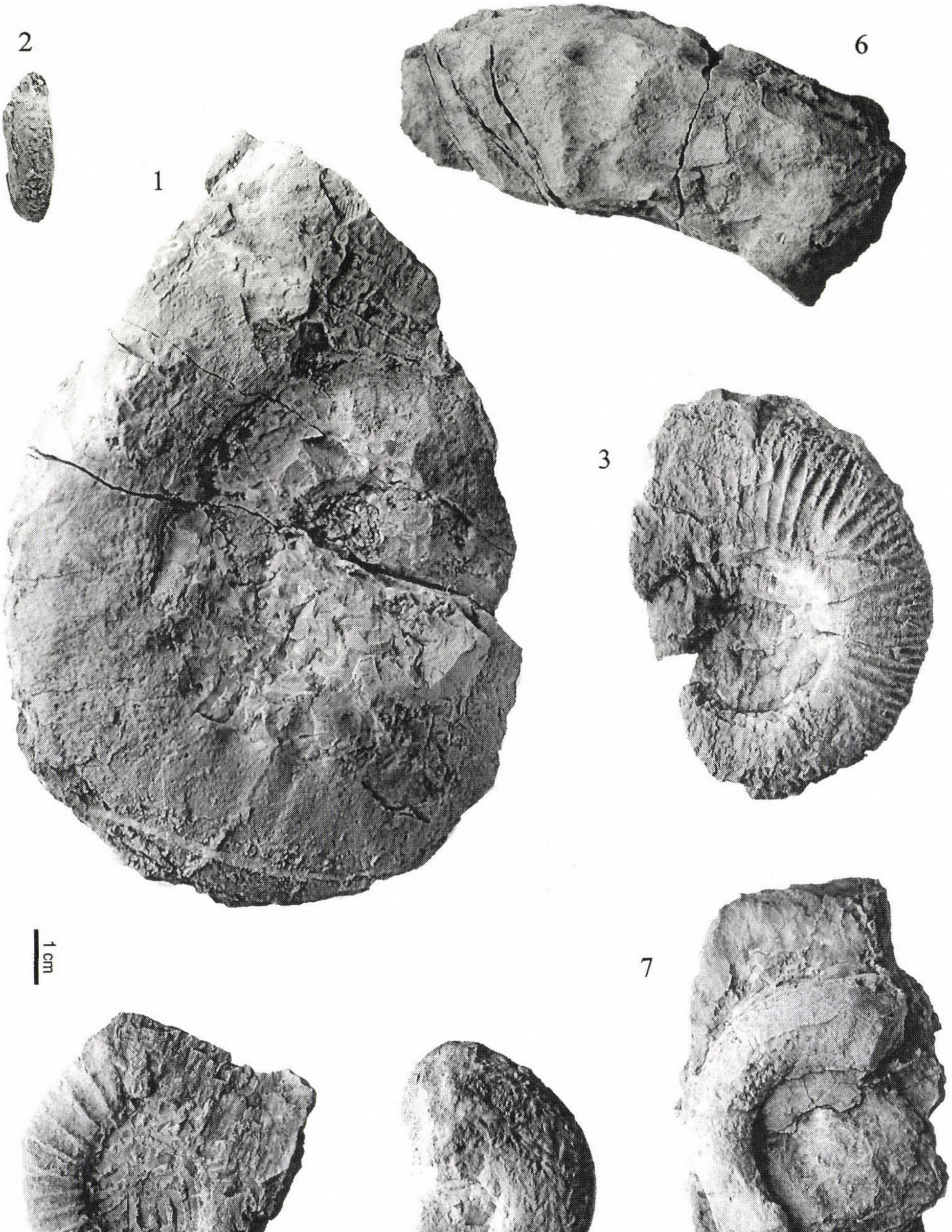
Plate I

1. *Pseudowaagenia acanthomphala* (Zittel 1870), Raspotnik 2, collected from debris, Kimmeridgian
2. *Haploceras carachtheis* (Zeuschner 1846), Wildenstein 1., Lower Tithonian
3. *Lemencia* sp. Wildenstein 1., upper part of the Lower Tithonian (? Ponti Zone)
4. *Ptychophylloceras ptychoicum* (Quenstedt 1847), Wildenstein 1., Lower Tithonian
5. *Virgatosimoceras micrum* Olóriz 1978, Wildenstein 1., Lower Tithonian (probably Darwini Zone)
6. *Hybonoticeras hybonotum* (Oppel 1863), Raspotnik 2, Lower Tithonian (Hybonotum Zone)
7. *Protetragonites quadrisulcatus* (D'Orbigny 1840), Raspotnik 2., Lower Tithonian

The specimens are housed in the Department of Geology and Paleontology of the Hungarian Natural History Museum.

Plate I



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Native gold in the epithermal HS mineralization of the Recsk Ore Complex

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The epithermal high sulfidization mineralization in Lahóca Hill is a part of the Recsk ore complex, which formed in the Upper Eocene andesitic sequence of NE Hungary. The Recsk ore complex consists of skarn-, porphyry Cu, high sulfidization and low sulfidization epithermal mineralizations. The Lahóca epithermal Cu-Au deposit was mined for more than 130 years, mainly for the copper ore, as gold was only a by-product. In this epithermal portion of the complex – as opposed to the former descriptions that mention gold to be present in pyrite – the native gold forms 1–30 μm large grains or irregular lamellae of high fineness. It is associated mainly with the second-generation pyrite but also occurs in the silicified matrix of andesites, pyroclasts and hydrothermal breccias. The presence of rutile is characteristic in the samples with higher gold grades. The Cu and Au mineralizations form two, distinct HS generations.

Key words: Recsk, epithermal, high sulfidization, native gold, silicification, second-generation pyrite

Introduction

Genetic interpretation of ore minerals is usually based on the examination of their morphological and mineralogical characteristics and the associated ore paragenesis. However, in case of native gold in HS epithermal gold deposits it is problematic, as the size of gold grains is generally only a few microns. They can be studied only by electron microscope, and because of this, earlier their presence was not realized in the rocks. Examinations included in this study were focused on Lahóca Hill. Earlier, during the Lahóca exploration project in the late nineties, examination of gold was limited to grade of gold content, not emphasizing the mineralogical characteristics of the gold. The mineralogical study of gold contributes to the better understanding of the genetic framework of Lahóca, and can also be an important factor in mineral processing.

Genetic framework of the Recsk ore complex

The larger part of the Mátra Mountains consists of Neogene (Middle–Upper Miocene) volcanic rocks. On its northeastern part the Recsk volcanic complex is represented by Paleogene volcanism. The unit was part of the internal island arc related to the subduction zone that was located between the Northern and

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Southern Alps during the Laramian–Pyrenean orogeny (Balázs et al. 1980). Due to a large-scale northeastward movement along the Darnó strike-slip fault, the Recsk Paleogene complex was emplaced in the recent position during the Lower Miocene (Csontos et al. 1992).

The pre-volcanic basement of the Recsk area consists of Triassic limestone, radiolarite and shale. Jurassic and Cretaceous rocks are absent. The Paleogene series begins with Upper Eocene, shallow marine limestone and marl, underlying, intermingling and overlying the dominant submarine intermediate volcanic rocks. The Paleogene volcanic cycle comprises four stages: (1) submarine lava flows, agglomerates, peperites (the rocks of this stage do not outcrop on surface); (2) stratovolcanic sequence of dacitic character, shifting gradually to subaerial environments; (3) stratovolcanic sequence of biotite-hornblende andesites, pyroclasts and reworked andesitic volcanic sediments, with the emplacement of diorite porphyry and quartz diorite intrusions hosting the porphyry copper mineralization; (4) development of the central explosive caldera in the area of the third-stage volcanism, formation of dyke bodies and laccoliths within and around the caldera. Oligocene sandstone, clay and marl cover the area except for the central andesitic horst (Baksa 1988; Gatter et al. 1999).

The Recsk mineralized complex consists of ore formations of different types linked to each other. In the deep intrusive body typical porphyry Cu-Mo mineralization was developed. Along the exocontact and endocontact with the Triassic limestone, skarn Cu-Zn-Fe mineralization was formed. In the Triassic limestone metasomatic and vein-type Zn-Pb ores occur. During the alteration of the deep intrusive body, pervasive silicification developed in the central apical part. Outward it is followed by a phyllic zone containing a quartz-sericite-anhydrite assemblage. The surrounding propylitic zone (with albite, chlorite, epidote, anhydrite and calcite) is not continuous and overlaps with the endoskarn, containing diopside, amphibole and phlogopite. The exoskarn is fringed by a metasomatic zone in which the limestone recrystallized to marble (Csillag 1975).

The porphyry Cu mineralization shows chalcopyrite-pyrite disseminations and stockworks. In the peripheral parts molybdenite occurs in siliceous-anhydrite veinlets. In the skarn mineralization the basic Cu-bearing mineral is chalcopyrite accompanied by pyrite, pyrrhotite, magnetite and hematite in the deeper horizons. In the skarn polymetallic deposit sphalerite is essential, associated with pyrite, chalcopyrite, galena and magnetite. In the zones of the hydrothermal-metasomatic alterations the polymetallic base metal ore deposits contain predominantly sphalerite, pyrite, galena and chalcopyrite (Csongrádi 1975).

Gold enrichment was recorded in several mineralization types. High associated gold content is present in the eastern flank of the porphyry copper ores. Elevated gold content was recorded in certain copper skarn, especially with magnetite-chalcopyrite-pyrite types. Gold is known in sediment-hosted form in

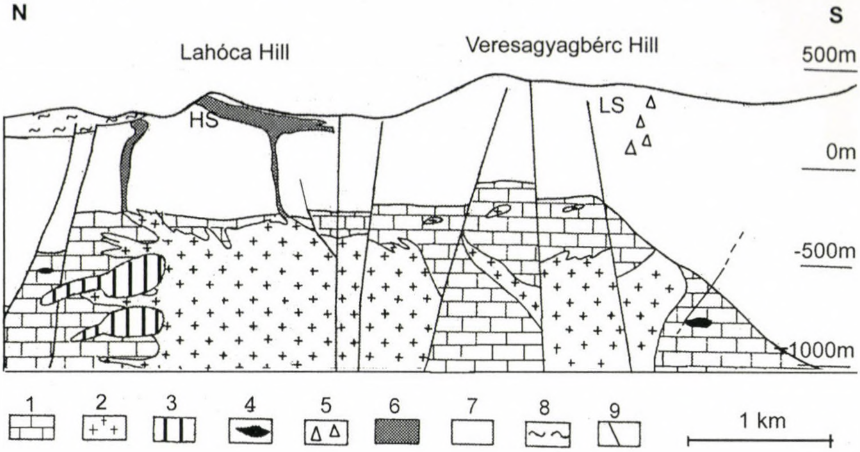

Fig. 1
Schematic cross-section of the Recsk mineralized complex. 1. Mesozoic basement; 2. mineralized copper porphyry; 3. skarn mineralization; 4. metasomatic Pb-Zn bodies and veins; 5. LS epithermal mineralization; 6. HS epithermal mineralization; 7. Upper Eocene stratovolcanic andesite; 8. Oligocene clastics; 9. tectonic zone

the upper Triassic shale horizon, close to and far from the intrusive centers. The relationship of these precious mineralizations to the epithermal near-surface enrichment of gold has not yet been investigated.

The deep-seated porphyry Cu deposit is in close spatial association with the epithermal Cu-Au deposits (Fig. 1). The epithermal mineralization appears at surface near Parádfürdő and on Lahóca Hill. In the Parádfürdő area the mineralization occurs in second-stage dacites and third-stage andesites and pyroclasts of the stratovolcanic complex. The altered bodies are strongly silicified, flat lenses and vein-like forms, surrounded by argillic alteration zones of kaolinite-alunite-(pyrophyllite)-smectite.

Lahóca Hill represents the typical high-sulfidization portion of the Recsk ore complex. There are several Au-ore bodies in the Lahóca system. The horizontal extent of the largest one (containing 95% of the ore reserves) is 500 × 900 m. This flat, brecciated body dips slightly southward and its thickness varies between 30–100 meters. The gold content is higher near the top and decreases with depth. Beside the central breccia zone, the Lejtakna "breccia pipe" and the northern breccia zone also show significant gold anomalies. The highest Au grade in the examined samples is 10.9 mg/kg (the formerly mentioned 180–300 mg/kg Au ore bodies were exploited in the twenties). The reserves of the Lahóca and Lejtakna areas are 35.8 mt ore with 1.47 mg/kg average gold content, at 0.5 mg/kg cut-off. The epithermal mineralization of Lahóca has a dominant pyrite–enargite–luzonite assemblage. Pyrite occurs in several forms and varieties. Gold is linked both to copper-enriched zones and virtually copper-free zones of the ore body. In certain smaller parts of the mineralized zone tetrahedrite enrichment is present, with high associated silver content (Földessy 1997).

Near Lahóca, south of Parádfürdő, polymict breccia bodies show K-metasomatic alteration with development of adularia and sericite. The ore minerals are galena, sphalerite, less pyrite, Pb-Se- and Ag-Sb sulfosalts, fahlores and rare Au-Ag-Bi-Te-Sb minerals. Gold appears both as tellurides and in native form (Nagy 1983). Gold enrichment is related to the adularia-bearing zones. According to Molnár and Gatter (1997) the Parádfürdő area can be considered as an early high-sulfidization deposit overprinted by low sulfidization mineralization.

The basic aim of this study was the examination of the Lahóca HS mineralization, revealing the composition, habitat and morphological characteristics of gold, and the genetic interpretation of the examination results. In order to clarify genetic questions, petrographic characteristics, rock alteration and the associated ore minerals were also examined. The samples represent all the rock types in Lahóca that are related to the gold mineralization.

Analytical methods

Thirty-four samples were collected from the following boreholes: R 368, R 370, R 371, R 372, R 377, R 378, R 390, R 396, R 404, R 408, R 416, R 417 and R 421.

The microscopic characteristics were studied both in transmitted and reflected light by with AXIOLAB A-type polarization microscope with MC 80 DX camera.

The scanning electron microscopic and EDX analysis was carried out with an AMRAY 1830 I scanning electron microscope with EDAX EDS detecting unit pv9700/36, (20 kV, SiLi detector, W cathode) at the Institute of Material Science, University of Miskolc by the help of Á. Kovács. The analyses were performed on the same thin sections that were subjected to the microscopic examinations.

XRD and thermoanalytical studies were carried out at the Hungarian Geological Institute, by P. Kovács-Pálffy and M. Földvári. The XRD instrument is a fully automated Philips PW 1730 diffractometer (Cu anticathode, 40 kV, 30 mA, graphite monochromator, goniometer-velocity 2º/min). The thermoanalytical instrument is a computer-controlled and evaluated Derivatograph PC with simultaneous TG, DTG, and DTA appliances, ceramic and corundum pot, 10 Cº/min heating rate and Al_2O_3 inert material.

Full chemical analyses were made at the Hungarian Geological Institute by A. Bartha. The AAS gold assay results were produced by ANALABS, Australia, Perth.

Samples, rock types and alterations related to HS gold mineralization

According to the examination results, five rock types were distinguished in the Lahóca area, related to the gold mineralization. These types in the order of genesis are as follows:

1) stratovolcanic andesite and pyroclasts (andesites: R372-76.5 m, R378-70 m, R417-68 m; pyroclasts: R372-85 m, R372-99 m, R372-111 m, R390-58 m, R396-14 m)

2) late andesite and pyroclasts (R370-101 m, R372-5,7 m, R408-14 m)

3) "blueschist" (R372-6.5 m, R372-8 m, R417-34.1 m, R421-86 m)

4) subvolcanic andesite (R368-103 m, R371-71 m, R372-121 m, R372-131 m, R372-134 m, R416-110 m)

5) hydrothermal breccia (R372-13.5 m, R372-16.5 m, R372-36 m, R372-41 m, R372-46 m, R372-52.5 m, R372-58 m, R372-62 m, R377-104 m, R416-262.5 m)

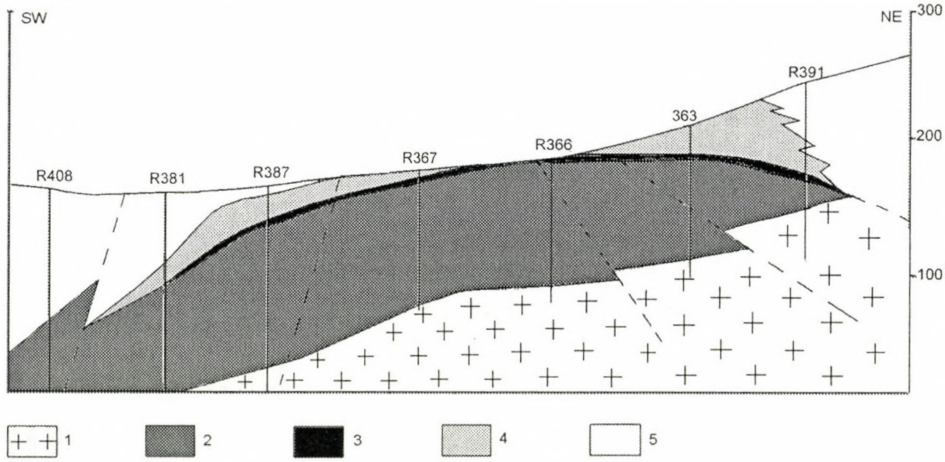

Fig. 2
Section across Lahóca Hill showing the lithological units related to gold mineralization. 1. subvolcanic andesite; 2. hydrothermal breccia and stratovolcanic andesite and pyroclasts; 3. "blueschist"; 4. late andesitic pyroclasts; 5. late andesite extrusives. Horizontal scale = vertical scale

The genetic framework of these rock types is as follows (Fig. 2):

The stratovolcanic andesite and pyroclasts formed the oldest unit in the area. As a subsequent event the late andesite extruded through and overlaid the former stratovolcanic sequence. The formation of this unit began with an initial eruption of pyroclasts. In the early phase the pyroclasts were deposited in a subaqueous environment, and a special variant of the pyroclastic rocks, locally called "blueschist" (a dark bluish-gray, clayey rock) was formed by the halmyrolytic decomposition of the volcanic fragments (Baksa 1975). This rock occurs along the boundary with the stratovolcanic series. In a later stage a subvolcanic andesitic body intruded the stratovolcanic sequence, and intrusive breccias formed in the apical part of the subvolcanic intrusion. Following this event, hydrothermal explosions produced hydrothermal breccia both in the stratovolcanic sequence and in the subvolcanic andesite. This hydrothermal breccia is the main host rocks of the gold mineralization. The "blueschist" formed an impermeable layer above the stratovolcanic and subvolcanic complex. Above this horizon the hydrothermal rock alterations are not significant and only a few veinlets containing Au enrichment were able to be formed.

Silicification and leaching is most intensive in the central part of the main ore body. The hydrothermal breccia frequently shows multiple brecciation and multiple silicification. The first silicification (matrix silica) displays colloidal or fine-grained quartz with slurred contours. The second and the following generations of quartz, which rim earlier fragments or appear in the form of younger fragments, have a mosaic-patterned texture, with definite grain contours. Pyrite is more abundant in the matrix silica than in the following quartz generations.

Outward from the central part argillic alteration becomes predominant. It is important primarily in the pyroclastic rocks. The strongest argillitization was observed in some parts of the blueschist, with dickite and montmorillonite. The argillitization in and around the central silicified breccia body displays zonality: closer to center dickite and kaolinite are abundant, outward they are followed by illite/montmorillonit-bearing zones leading to a predominance of montmorillonite and chlorite in the outermost zones (Table 1). It can be generally stated that gold (and pyrite) content decreases with increasing argillitization.

Table 1
Quantitative XRD phase composition of some of the samples studied

Rock type	Sample	montm	ill/mm	illite	Kaolinite	chlorite	quartz	K-feldspar	plagio-clase	pyroxene	pyrite	gypsum	calcite	amor-phous
Late andesite	R408-14	-	16	-	-	6	27	9	9	7	20	1	2	3
	R372-6.5	11	-	-	71*	-	-	-	-	-	15	-	-	3
Blueschist	R417-34.1	-	18	3	23	-	27	-	10	-	16	1	-	2
Stratovolc.	R372-99	-	11	-	9	-	56	4	-	-	17	-	-	3
and+pyrocl.	R390-58	-	18	-	3	3	25	4	25	-	6	2	11	3
subvolcanic	R368-103	-	-	2	23*	-	54	-	-	-	21	-	-	-
andesite	R372-121	-	15	-	1	-	60	5	-	-	15	1	-	3
Hydrotherm. breccia	R372-16.5	-	19	-	9	-	48	8	-	-	11	-	-	4

Amounts of minerals are given in %. Values <5 % are only of indicative role. Analyst: P. Kovács-Pálffy, Hungarian Geological Institute

Calcite was found only a in a few samples from the peripheral part of subvolcanic andesite and stratovolcanic tuffs, as replacement of porphyric minerals and fissures fillings. It represents the latest, post-ore phase of alteration as it cuts enargite and is free of pyrite. It indicates the late neutralization of hydrothermal fluids.

Ore mineralization

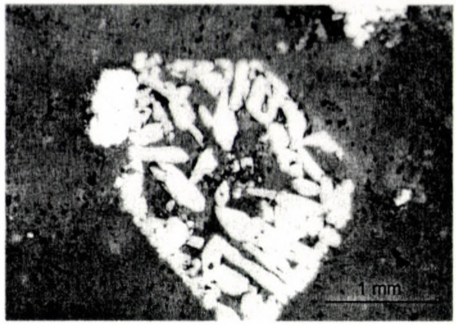

Fig. 3
Prismatic and elongated rutile crystals after ilmenite (light gray), subhedral pyrite (white) in silica matrix (dark gray). Au: 10.9 mg/kg. R377-104 m, hydrothermal breccia. Reflected light, 1N

The main opaque mineral is pyrite (about 95% of ore minerals). The rest is mostly enargite and luzonite, subordinately galena, sphalerite, tetrahedrite and complex sulfosalts. Three generations of pyrite can be distinguished: (1) 100–300 μm large, euhedral, corroded crystals; (2) fine dissemination in the silica matrix (2–20 μm); (3) collomorph, spherical or banded structures. Marcasite is even a later phase, following colloidal pyrite. Gold precipitation is related mostly to the second pyrite generation. Enargite and luzonite form 500–500 μm large subhedral or anhedral crystals. They occur as vein fillings, replacements of the porphyric minerals in the silicified andesites, or included in breccia fragments.

It was not recognized earlier that rutile is a relatively frequent mineral. It possibly formed by the decomposition of Ti-rich accessory minerals (ilmenite) and amphiboles, and reflects the rate of alteration. The unaltered late andesite is free of rutile, but contains ilmenite crystals. On the other hand, the altered rocks display extremely fine, dispersed rutile in the silica matrix and concentrated rutile crystals after ilmenite and amphibole. In the latter case euhedral pyrite crystals often rim the rutile-cluster. These pyrites probably utilized the Fe produced by the decomposition of the original crystals. During the hydrothermal processes, the multiple leaching and mineralization could have contributed to the Ti-enrichment. The highest Au-grade samples contain the largest, well-developed rutile crystals (Fig. 3), and are usually more abundant in Ti, although the correlation between Au and Ti is not always definite (Fig. 4).

Mineralogical characteristics of gold

The gold is concentrated in the highly silicified zones. Gold enrichment is the highest along the contact with the blueschist. In the argillic alteration zones the gold content is 0.1–0.2 mg/kg.

In the examined Lahóca samples gold is present in all rock types. Even in the samples with low assayed Au grade it could be examined by electron microscope. It occurs predominantly in native form, in high fineness (Au:Ag ratio). In a few cases the presence of Au-tellurides is assumed from the amount of tellurium which was detected in gold-enriched particles. Tellurium was also found in form of melonite ($NiTe_2$) in the gold-free luzonite (Gellért et al. 1999). The native gold appears both related to pyrite and in the silicified matrix.

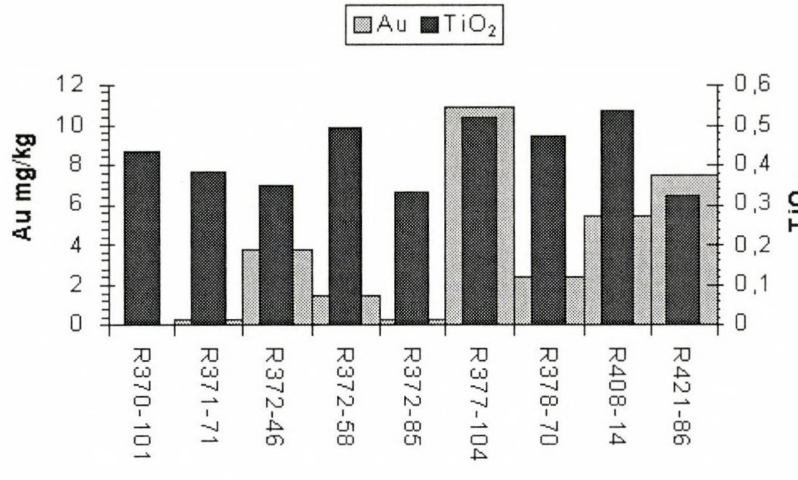

Fig. 4
Relation of Au and TiO_2 indicating that the higher-Au-grade samples are more abundant in rutile. An exception is sample R371-101 m (unaltered, late andesite) in which TiO_2 belongs to ilmenite. R377-104 m hydrothermal breccia contains the largest, well-developed rutile crystals

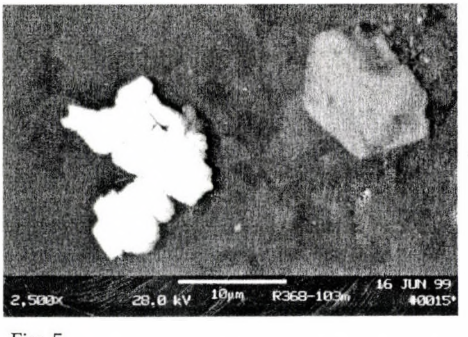

Fig. 5
Native gold with aggregate-like structure (white) and euhedral, second-generation pyrite (gray) in silicified matrix. Au: 0.237 mg/kg. R368-103 m, intrusive andesite. BSE image

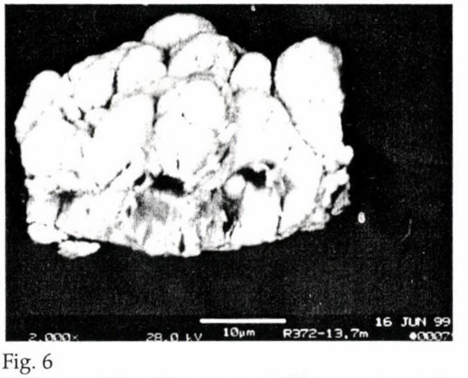

Fig. 6
Native gold with aggregate-like structure in silicifed matrix. Au: 2.33 mg/kg. R372-13.5 m, hydrothermal breccia. BSE image

The "free" gold in the silicified matrix occurs in form of 10–30 micron-large aggregates (Figs 5, 6) or cloud-like structures (Fig. 7). It always appears in a silicified environment, either in the glassy matrix of the volcanic rocks or in the breccia matrix.

The pyrite-related gold is more frequent, but these gold grains are smaller than those in the silicified matrix. Their average size is a few μm. It is the second-generation pyrite that is mostly associated with gold (Figs 8, 9). These pyrites are euhedral-subhedral crystals, a few μm to 20 μm across, and are disseminated in

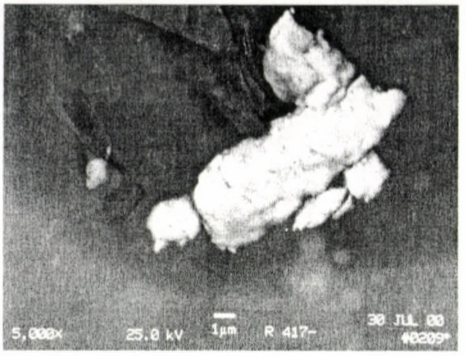

Fig. 7
Native gold with cloud-like structure in silicified matrix. Au: 4.14 mg/kg. R417-68 m, stratovolcanic andesite. BSE image

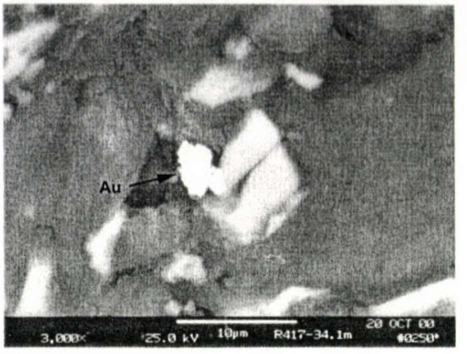

Fig. 8
Irregular native gold grown to the edge of a second-generation, subhedral pyrite crystal in the silicified matrix. Au: 4.79 mg/kg. R417-34.1 m, blueschist. BSE image

the breccia matrix or in the silicified rocks. The larger, corroded, first-generation pyrite that also occurs in breccia fragments is usually free of gold. Similarly, the third-generation collomorph pyrite and the marcasite are not associated with gold. In the investigated samples gold was not found in or related to enargite, luzonite or any other ore minerals.

Contrary to former assumptions, gold does not occur as solid solution in pyrite. It inevitably shows affinity to pyrite, but always grows on the crystals, either in small cavities or attached to any part of the pyrites, indicating a separate mineralization phase (Figs 10, 11).

Genetic interpretation of the examination results

The Lahóca deposit, formerly considered only as a copper ore mineralization, proved to be a typical high-sulfidization gold deposit as well. The formation of the two ore-types took place in different stages (Fig. 11).

Fig. 9
Irregular gold grains on pyrite. Au: 4.79 mg/kg. R417-34.1 m, blueschist. BSE image

According to the textural characteristics of the examined rocks and minerals, considering the paragenetic sequence and the development of hydrothermal fluids the following summary can be made:

The Lahóca HS mineralization has a definite spatial relationship with the porphyry Cu mineralization in the depth. In the early phase of the HS

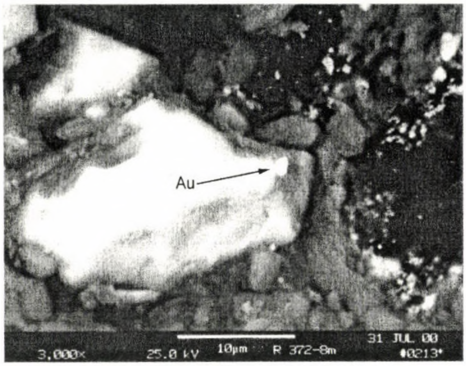

Fig. 10
Native gold on pyrite, silicified matrix. Au: 1.57 mg/kg. R372-8 m, blueschist. BSE image

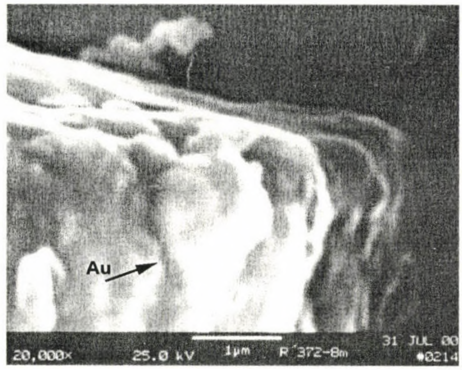

Fig. 11
SEM morphological image of the native gold grain visible in Fig. 10 indicating that native gold grows on the surface of pyrite. Au: 1.57 mg/kg. R372-8 m, blueschist.

mineralization the fluids had a strongly acid character that is represented by the formation of enargite and luzonite. Some luzonites contain Ni-tellurides as inclusions (Gellért et al. 1999). A slight increase in the pH of fluids resulted in the formation of first-generation pyrite and the free gold in the matrix. These pyrites are partially dissolved (corroded) by the later mineralization phases. Explosive brecciation followed the formation of enargite, luzonite and first-generation pyrite. The second-generation pyrite occurs mainly in the breccia matrix. Shortly after the second-generation pyrite, the main gold mineralization took place. The correlation between pyrite and gold content is definite. These pyrites promoted the formation of gold by providing a surface for precipitation. By the formation of the third-generation pyrite and the late marcasite the fluids had lost their gold content. Thin Au-bearing veinlets in the late andesite were produced by the late mobilization of the gold content of the main ore body.

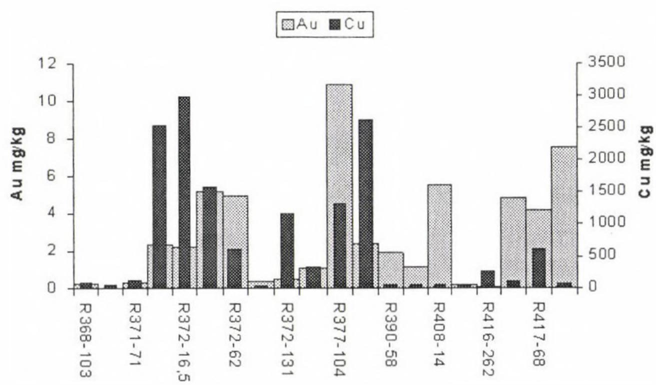

Fig. 12
Relation of the Cu and Au content in the examined samples showing poor correlation between the two elements

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