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Oceanic realms in the central part of the Balkan Peninsula during the Mesozoic

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Abstract. During the Mesozoic three oceanic realms existed on the Balkan Peninsula: the **Vardar Ocean** (NW part of the main Tethys) forming the main basin with a marginal basin at the west. Since the Maastrichtian these basins with the intervening units build the Vardar Zone Composite Terrane, embracing the Main ophiolitic belt, the Kopaonik block-and-ridge unit, and the Western ophiolitic belt. The **Dinaridic oceanic tract** (Dinaridic Ophiolite Belt as its present scar) formed the second basin, with a continuation into Mirdita. The **Civcin-Severin oceanic realm** was the third basin at the east. These basins or realms were opened at different times, had different life spans, and were closed during different parts of the Mesozoic. These oceanic realms thus show different general characteristics and display different rock complexes connected with their closing.

Key words: oceanic realms, Mesozoic, Vardar Zone, Main Belt, Western Belt, Dinaridic Ophiolite Belt, Civcin-Severin Zone, Balkan Peninsula.

Introduction

At the beginning of this century two regional ophiolite zones were already noted in the central part of the Balkan Peninsula. These are very well expressed at the Geological map of SFRYu 1:500,000 (Geological Institute of SFRYu, Belgrade, 1970), and numerous papers (especially in the last thirty years) deal with parts or some members of these zones. Only those papers offering new exact data will be quoted here. The real importance of the zones has been understood only with the birth of recent ideas on the development of the Earth, when it became clear that ophiolite belts represent relics of ancient oceanic realms, as remains after the collision of adjacent continental plates and blocks.

The elaboration of the Basic Geological Map of SFRYu (references are given in Dimitrijević, 1997) made it possible to understand specific, essentially different characteristics of these ophiolitic belts. They resulted from differences in geotectonic framework, age of birth and duration of oceanic expanses from which these belts originated, being reflected in presence/absence of specific sedimentary and magmatic members, as well as in their sedimentologic, petrologic and geochemic characteristics. Correlative synthesis of the main belts, based on modern geological approaches was first given by Dimitrijević and Dimitrijević (1973). Recently comparative presentations were given by Karamata, Dimitrijević and Dimitrijević (1998a,b). The present paper is a further elaboration of thoughts developed in these papers.

This paper was submitted for the International geological conference „Carpathian Geology 2000“ (October 11th-14th, Smolenice), and a short version appeared as an abstract in the Conference publications.

General subdivision

Existing data point to the existence of three different ophiolitic belts (Fig.1), resulting from separate oceanic realms and basins, and differing clearly in the features summarized in the Table 1. These ophiolitic belts are:

1. The Vardar Zone, including relics of (at least) two oceanic areas:

   1a. Main ophiolite belt of the Vardar Zone (MVZ) as scar of the main Vardar oceanic realm - the Tethys, and

   1b. Western ophiolitic belt of the Vardar Zone (WB) as scar of the western marginal basin of the Vardar ocean.
2. The Dinaridic Ophiolite Belt (DOB), with the continuation to the south into the Mirdita Zone, as relic of the oceanic tract which ran through the Dinarides and was along its southeastern border locally connected with the Tethys.
3. The Civcin-Severin ophiolite belt at the east.

The Main Ophiolite Belt of the Vardar Ocean (MVZ)

This ophiolite belt represents the suture zone of a vast Mesozoic oceanic realm - the Vardar Ocean i.e. the NW part of the Tethys. This oceanic basin had a long existence as the successor of the Prototethys and later on as the continuation of the Lower Paleozoic (or older?) oceanic realm between the Gondwana and continental blocks at the north, from Permian building one unit the Eurasia (this oceanic realm was formerly named the „Zvornik ocean“ by Dimitrijević and Dimitrijević, 1973). This realm was highly complex, with island arcs (terranes docked to the Moesian massif, as well as the Veles series

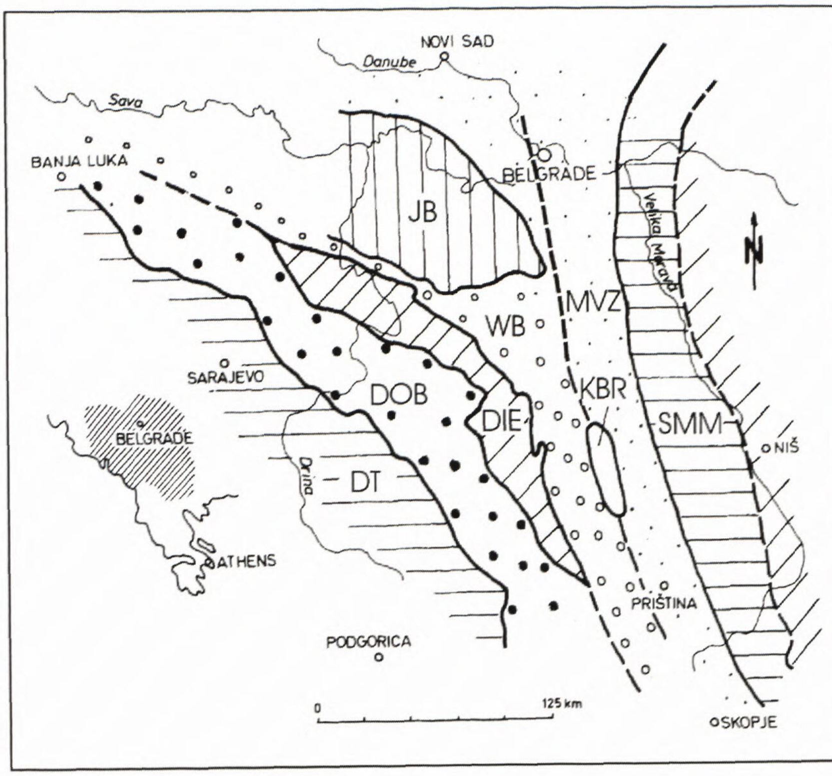

Fig.1. Present position of ophiolite belts, relics of Mesozoic oceanic realms in the central part of the Balkan peninsula. DT - main Dinaridic trunc; DOB - Dinaridic Ophiolite Belt; DIE - Drina-Ivanjica Element; WB - western belt of the Vardar Zone; KBR - Kopaonik block-and-ridge unit; MVZ - main Vardar Zone belt; JB - Jadar Block; SMM - Serbo-Macedonian massif.

units occurring in the belt) and oceanic basins between detached continental blocks. The original position of these terranes and units is presently still impossible to identify, but they were before and during Devonian separated units far at the south. It is assumed that after the Carboniferous the eastern border of this realm was represented by the Serbian-Macedonian Massif, which most probably already docked to the Carpathian units at the east. At the west of this realm lay the Apulian plate and the still unsolved set of Dinaridic microplates. During the Upper Triassic a slice separated from its eastern margin forming the Kopaonik block-and-ridge unit (KBR) and behind it a new back-arc basin or the Western oceanic basin, as the precursor of the Western ophiolite belt of the Vardar Zone (WB) originated.

The Main Vardar Ocean closed before the end of the Jurassic through eastward subduction beneath the Serbian-Macèdonian Massif and the pre-Permian Carpathian-Balkan collage.This closing ensued in a collisional metamorphism, visible less at the north where erosion was lower, and more visible to the south, where (in the area of Valandovo, Macedonia) lowP/highT metamorphic rocks and even S-granites are exposed. In the subduction trough in front of the Serbian-Macedonian massif an olistostrome of regional extension generated. This has blocks and lenses (some of kilometer length) of basalt, gabbro, ultramfites, turbiditic greywackes, dark limestone of unknown origin, red radiolarites, and sparse Tithonian limestone, embeded in a silicious siltstone matrix. This complex is mainly covered by an overstep-sequence of Tithonian reef limestone and Lower Cretaceous paraflysch.

A characteristic member of this ophiolithic complex are kilometer-long lenses of the „Veles series“ (Devonian?-Carboniferous, Grubić and Ercegovac, 1975; Stojanović 1997). They probable represent fragments of island arcs existing in the oceanic basin. Some high-grade metamorphic blocks and lenses in the southwestern part of the zone most probably represent late tectonic inclusions.

In the recently studied framework of the Vardar Zone products of this ocean encompass the main, eastern part of the Vardar Zone (the Central Vardar Subzone, Dimitrijević in Mahel, 1973), and are in the figure 1 to the east of the Kopaonik block-and-ridge unit.

Western Ophiolitic Belt of the Vardar Zone Composite Terrane

Along the western border of the Vardar Zone composite terrane, presently between the Kopaonik block-and-ridge unit (KBR) to the east and the Dinaridic Drina-Ivanjica Element (DIE) to the west, an olistostrome zone rich in large basaltic masses runs, representing the relic of a marginal basin of the Vardar (Tethys) ocean. It is mainly covered by Tertiary nappes of peridotite, expelled from the scar east of the Kopaonik block-and-ridge unit and pulled over it to the west as far as the boundary of the DIE (Ibar ultramafites - masses of Troglav, Stolovi, Ra{ka, southern Kopaonik etc.). This zone is thus visible in the deep erosional cuts only (the Studenica gorge, Maglić, Leposavić-Banjska area) only. Toward the north the zone extends over the Jelica Mt. along the Zvornik suture to Majevica and further on to northwestern Bosnia.

According to present data this zone represents the relic of a marginal basin of the Vardar (considered as the main Tethys) ocean, probably opened toward the border of the Dinarides during the Upper Triassic. The basin had to be very wide, with the formation of (probabaly immature) island arcs. The closing of the basin began probably during the early Upper Jurassic (age of the metamorphic rocks in the base of ultramafic slices; Balogh, oral communication), taking a very long time to the final closure. According to Milovanović et al. (1995) Barremian crossite schists at Fruska gora (northern part of the western marginal basin) are also related to this long lasting subduction.

In the subduction trench an olistostrome was formed. In this olistostrome rounded greywacke blocks are common together with pillow-lavas of MORB, as well as of

IAB affinity. Other constituents are blocks of gabbro, metamorphites of various grade from the ophiolite sole from basaltic as well as from sedimentary protoliths, rare ophiolite related albite-granites, abundant Karnian to Kimmeridgian cherts, Triassic limestone lenses and blocks, and fragments of Upper Cretaceous limestones. Fragments of rocks from the „Veles Series", as well as of locally Paleozoic granites, were not found. The matrix is silicious, argillaceous-silty.

The final phases of subduction and the closure of the basin are reflected by relations of the mélange and rudist limestones east of Ivanjica (Brković et al., 1977), by the occurrence of fragments of globotruncana limestones in the mélange of the Jelica Mt. (Vandjel and Marić. 1956; Brković et al., 1978), senonian fragments in the mélange of Sokolska Mt. (D.Ljubović-Obradović, 1985; A.Djuricković and V.Orsolić, 1988, written commun.), and fragments of the Upper Cretaceous limestone in the pillow-lavas near Krupanj (Filipović, oral commun.). To the subduction process can be related the Upper Cretaceous(-Paleogene) magmatic rocks in the Southern Carpathians and in East Serbia, as well as the Upper Senonian volcanics in the Belgrade area, in South Backa and Central Banat (Karamata et al., in print).

The basin existed to the upper Senonian, when it was closed, most probably by the early Maastrichtian. This is also indicated by the Campanian age of metabasalts and interlayered sandy limestones north of Kozara Mt. (Karamata S., Pushkarev Yu., Sladić M., 1999, written commun.).

The Dinaridic Ophiolite Belt

The Dinaridic Ophiolite Belt (DOB) extends between the Sarajevo Sigmoid with the Central Bosnian Schist Mountains Terrane and the East Bosnian-Durmitor block (as parts of the main Dinaridic trunc) to the west and southwest, and the Drina-Ivanjica element/terrane (DIT) to the northeast. Toward the northwest this belt comes into immediate contact with the Western belt of the Vardar Zone, the boundary between may be recognized only due to the differences in age of the mélange. Toward the south the belt continues into the Mirdita Zone, with a conspicuously different character; the relations of this region with the Vardar ocean are still uncertain.

Alongside the main exposed suture zone, olistostromes of a similar composition as those in the main zone occur inside the main Dinaridic trunc, e.g. beneath the front of the Durmitor nappe. They may represent either relics of some smaller oceanic tracts, or parts of the main olistostrome zone separated from it by the emplacement of the East Bosnian-Durmitor terrane. It is highly indicative that this olistostrome lies in a belt characterised by high thermal flux during the Upper Jurassic (*in situ* granitization in the Junik knot), and by extremely complicated relations of the adjoining nappes and terranes.

The Dinaridic Ophiolite Belt is the relic of an oceanic tract which opened during the Middle Triassic between the main Dinaridic trunc and the Drina-Ivanjica element (Dimitrijević and Dimitrijević, 1973). The extensional phase of this ocean probably lasted to the mid-Jurassic, the closure by subduction toward the present northeast lasted up to the end of Jurassic. The beginning of closure is indicated by the Middle Jurassic emplacement of ultramafics. The age of metamorphism of the ophiolie sole is arround 170 Ma for Vijaka and Bistrica (Lanphere et al., 1975), Brezovica (Karamata and Lovrić, 1978), and for Banija (Majer et al.,1979), mostly over already existing trench olistostromes. The final phases of influence of the already subducted oceanic crust or its ridge are shown by the Vallanginian metamorphism in the deeper horizons of the Drina-Ivanjica Paleozoic (Milovanović, 1984). The Pogari series, of Tithonian partly Lower Cretaceous age, was deposited over the ophiolitic mélange. This series contains as constituents in conglomerates and sandstones all members of the olistostrome.

The composition of the ophiolitic olistostrome, exposed in large outcrops, differs from the mélanges of the Vardar Zone. Prevailing are blocks and olistoliths of Triassic limestone and greywackes (turbiditic in places) or silicious siltstones, together with Ladinian (Obradović and Gorican, 1989), Carnian (Gostilje; Gorican 1998, oral commun.), Carnian-Norian (west of Sjenica, Gorican et al., 1999), and Callovian to early Kimmeridgian oceanic cherts (Bistrica; Obradović and Gorican, 1989). Abundant are basalts (pillow-lavas mostly) of MORB character north of the line Peć-Goles and of both MORB and IAB affinity south of this line, some gabbro, enigmatic lenticular bodies of Carboniferous granite of unknown provenience (Karamata et al., 1996), and fragments of metamorphites connected with the emplacement of hot peridotite slabs. An outstanding feature of the belt are kilometer-sized, composite, olistoplakae (large plate shaped olistoliths) of Triassic limestone, gravitationally transported from the Drina-Ivanjica element, less frequently of Triassic to Jurassic limestone from the slope of this element, large bodies of red oceanic silicious globigerina slate („Zlatar chert"), and huge masses of obducted ultramafites which metamorphosed the mélange along their base (Zlatibor, Konjuh, Brezovica etc) or diapirically intruded the oceanic crust (Ozren west of Sjenica; Popević, 1985) with metamorphism of surrounding rocks. Upper levels of these ophiolite complexes are only rarely preserved in continuity (e.g. Visegrad, Brezovica etc).

The Civcin-Severin Ophiolite Belt

This belt is situated at the east of the Serbian-Macedonian massif and the Ranovac-Vlasina-Osogovo terrane. In the lower Alpine nappe of the South Carpathians (Severin Nappe), in Eastern Serbia and southwestern Romania serpentinites are found at several places, together with small amount of other ophiolitic rocks and deep-sea sediments. These rocks, well studied in Romania, are connected with a Jurassic-Cretaceous oceanic basin, the position of which is still highly problematic. The features of this belt and of the former oceanic basin are insufficiently known due to the very complex cover of Alpine nappes, but existing data point to some similarities with the western belts.

Table 1. Correlation of oceanic realms in the present Balkan Peninsula
DBO=Dinaridic Ophiolitic Belt; WB=Western Ophiolitic Belt of the Vardar Zone; MVZ=Main Ophiolitic Belt of the Vardar Zone

A G E				TRENCH ASSEMBLAGES		
	DOB	WB	MVZ	DOB	WB	MVZ
TT		*Ca- alk* FLYSCH				
Mastr.		rudist				
Camp.		limestone → ←			GRAYWACKE RADIOLARITE T_2, T_3, J_3 basalts of MOR and IA type gabbro ultramafic (harzburgite) limestone (T_2-T_3, K_2) ophiolitic sole metamorphics ≈155 Ma MATRIX SILICIOUS AGILLACEOUS- SILTY	
K_2		Basalt *mag-* 80 Ma *ma-* (K/Ar) *tism*				
K_1		crossite schist	PARAFLYSCH REEF LIMESTONE → ←			
Tith.	POGARI SERIES → ←					
J_3	emplacement of ophiolite slabs ≈170 Ma (K/Ar)	emplacement of ophiolite slabs ≈155Ma (K/Ar)		GRAYWACKE RADIOLARITE T_2,T_3,J_3 basalt (MORB) gabbro Ab-granite ultramafic blocks limestone T,J CARBONIFER. GRANITE LIMESTONE T_1,T_2,T_3 OLISTO- PLAKAE ULTRAMAFIC SLABS (Lherzolite) OBDUCTED or INTRUDED with METAMOR- PHIC AUREOLE ≈170 Ma MATRIX ARGILLA- CEOUS -SILTY		BASALT gabbro ultramafics graywacke (turbiditic) radiolarite limestone (dark, white) METAMORFICS OF THE VELES SERIES of low to medium grade MATRIX SILICIOUS SILTSTONE
J_2						
J_1						
T_3		← V →				
T_2	← V →					
T_1						
P						
C			ISLAND ARC RELICS – THE „VELES SERIES“			
D						
O						
S			TRANSPORT OF TERRANES towards N-NE			

Time-scale is not linear ← V → opening; → ← closing
Capitals – characteristic members

Conclusions

In the central part of the Balkan Peninsula relics of several Mesozoic oceanic basins are exposed, they differ in their opening and closing times.The main oceanic realm was the Vardar ocean, which existed from the lower Paleozoic (or even earlier), while other basins represented marginal basins or oceanic crust generated between the dispersing continental fragments. The composition of olistostromes, deposited in subduction troughs of these oceanic basins, differs significantly, depending on the width of the oceanic area, geotectonic setting and lithology of the trough margins.

Main characteristics of these oceanic basins are as follows:

The main Vardar basin: long continuous existence as continuation of the lower Paleozoic (or older?) oceanic realm; closing at the end of the Jurassic; presence of Paleozoic island arc relics („Veles Series“); prevalence of material from the higher parts of the oceanic crust in the olistostrome; absence of limestone olistoplakae;

Western basin of the Vardar Zone: existence from the Late Triassaic to the latest Senonian; prevalence of greywackes and basalts of MORB and IAB affinity in the olistostrome; absence of Paleozoic metamorphites and granites; presence of Upper Cretaceous oceanic-crust basalts with limestone fragment of similar age;

Dinaridic Ophiolite Belt: existence from the Middle Triassic to the end of the Jurassic; prevalence of greywacke and Triassic limestone as olistoliths in the olistostrome; lenticular bodies of Carboniferous granite and kilometer-siyed lenses of deep-sea silicious rocks; olistoliths of Middle Triassic to Upper Jurassic chert; large olistoplakae of Triassic limestone; ultramafic slabs and diapirs with contact-metamorphism in the floor or sides respectively;

Civcin-Severin belt: insufficiently known, but limited existing data point to some similarities with western belt of the Vardar Zone.

These data show that ophiolitic belts of the central Balkan Peninsula represent relics of different Mesozoic basins with oceanic crust. They could not be regarded as products of the one sole oceanic realm, but evolved from a complex area which included more as one oceanic regions during the time of the closure of this large and complex oceanic area.

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Precipitation and chemical composition of iron ochres in the pyrite and stibnite deposits in the Malé Karpaty Mts.

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Abstract. Exploitation of pyrite ores and hydrothermal Sb deposits in the Malé Karpaty Mts. culminated at the turn of the 20[th] century. The host rocks of mineralizations are black shales in the actinolitic schists and amphibolites. Intensive weathering occurs mostly on the dumps and outcrops in the pyrite-bearing black shales of the Augustín deposit. Acidofilic lithotrophic bacteria was isolated from mine water with pH 2.5–3, which indicates their partition in weathering processes. As solid secondary phases, gypsum and jarosite occur in weathered black shales. The formation of young iron ochres prevail in drainages with pH 5.5–8. They are collected on the bottom of effluents and sometimes fill extentive marshes. According to chemical composition, several types were recognized: depending on the primary mineralization As-, Si-, Al-, and SO_4-rich ochres are formed. Sb, Zn, Ni, Pb, Cd, Ti, P, Ca, Na and P are accummulated as minor or trace elements. Our research refers to the presence of poorly crystalline goethite, schwertmannite, Fe arsenate-sulphate in precipitates, the occurence of ferrihydrite and Si-phases can not be excluded. Based on our and previous investigations, the ecological problem of the area actually may be toxic metals pollution (As, Sb, Al and SO_4) rather than acidification itself.

Key words: Malé Karpaty Mts., mine drainage, iron oxyhydroxides, sorbtion

Introduction

Economically the most important ore mineralizations in the Malé Karpaty Mts. were emplaced in the lower- to middle Devonian (Planderová & Pahr, 1983) volcano-sedimentary formation, later metamorphosed in the amphibolite facies. According to Chovan *et al.* (1992) two mineralization types occur:

I. Metamorphosed, primarily exhalation-sedimentary pyrite mineralization.
II. Hydrothermal mineralization, which is subdivided into three sub-types:
1 - molybdenum in granitoides;
2 - copper-base metal with silver: (a) Cu-Pb, Ag, (Ni); (b) Pb-Zn; (c) Pb-Ag;
3 - antimony-gold: (a) gold-sulphidic; (b) gold-quartz; (c) stibnite.

Two of these subtypes were subjected to mining. First, the pyrite-pyrrhotite mineralization, which is a product of exhalation and sedimentary processes of submarine volcanism. Pyrite exploitation dates to the late 18[th] century, flourishing between 1850 and 1896 (Cambel, 1959). Second, the hydrothermal Sb-As-Au mineralization cuts through the pyrite-pyrrhotite one. Dominant ore minerals are arsenopyrite, pyrite, stibnite and gudmundite. Intense carbonatization took place as well. Two Sb deposits bound to this mineralization were exploited: the Pernek deposit (1790 – 1922, Koděra ed., 1990) where several abandoned dumps remained, and the Kolársky Vrch deposit (1790 – early 1990's). The latter was equipped with flotation processing since 1906 (Cambel, 1959). Waste was deposited in three tailings impoundments. Both of these mineralizations are bound to the lenses of black shales in the actinolitic shists and amphibolites.

Weathering of open deposits, dumps and tailings impoundments is the cause of the following processes:

– pollution of water, soil, and alluvium in the surrounding area, main contaminants being As and Sb (Letko *et al.*, 1992; Veselský *et al.*, 1996);

– local acidification (Šucha *et al.*, 1996; Trtíková *et al.*, 1997);

– precipitation of supergene products – a wide variety of secondary minerals has been reported: allophane, azurite, cervantite, gypsum, halloysite, hyalite, jarosite, kaolinite, kermesite, malachite, Mn oxides and hydroxides, senarmontite, siderite, Sb ochre, schafarzikite, valentinite, limonite (Cambel, 1959; Andráš & Chovanec, 1985; Koděra ed., 1990), chapmanite (Polák, 1983), and scorodite (Uher, 1990);

– local degradation of phytocenoses, especially in close proximity of dumps with black shales (Banásová in Šucha *et al.*, 1996).

The most distinctively pronounced hypergenous process is the precipitation and deposition of iron ochre. Our research deals with these young hypergenous products in the pyrite-pyrhotite bearing ore district of Augustín and Michal and in the region of Sb - Au deposits of Kolársky Vrch and Pernek. The aim of our study is to identify iron oxyhydroxides - basic components of iron ochres (Bigham *et al.*, 1996a), to investigate their chemical composition and to specify the conditions of their formation.

Analytical Techniques

Places of acid mine effluents were recorded, as well as areas where iron ochres precipitate and settle. The pH of waters was measured and samples of water and solid secondary phases were taken for further analysing.
The sampling methods were chosen according to those described in Bigham *et al.* (1996b). Plastic bottles were filled with material after a wash in the local water. Organic matter (remains of plants) and inorganic detritus were removed by sieving and sedimentation. Solid samples were then air-dried and properly packed to prevent oxidation. The contents of dissolved elements in mine waters was measured by Atomic Absorption Spectrometry (AAS). In the X-ray powder diffraction (XRD), CoKα radiation with a Philips PW1710 goniometer was used and solid samples were scanned from 4 to 80° 2Θ with increment of 0.02° 2Θ and 0.5-0.8 s per step (Geological Institute of Slovak Academy of Sciences). The total Fe content (Fe_{tot}) in the precipitates was determined using dissolution in concentrated HCl. Oxalate soluble Fe (Fe_o) was determined after dissolution in ammonium oxalate (Schwertmann, 1964). Samples were analysed for Al (atomic emission spectroscopy with inductively coupled plasma), As, Sb (AAS with hydride generation), Fe (AAS with atomization in air-acetylene flame), SO_4 (gravimetry), Cd, Cu, Mn, Pb, Si, Zn, Na and K (AAS) (Laboratories of analytic methods of the Faculty of Natural Sciences, Comenius University, Bratislava and the Slovak Academy of Science, Banská Bystrica). Transmission electron micrographs (TEM) for studying morphology of precipitates (JEOL 2000, Faculty of Nature Sciences, Comenius University, Bratislava) were acquired. Infrared absorption spectra (IR) were obtained with a Nicolet Magna 750 Fourier transform infrared spectrometer equipped with a DTGS detector (Institute of Inorganic Chemistry, Slovak Academy of Sciences, Bratislava). For each sample 256 scans were recorded in the 4000 - 400 cm^{-1} spectral range with a resolution of 4 cm^{-1}. The KBr pressed disc technique (1 mg of sample and 200 mg of KBr) was used. The acidofilic chemolithotrophic bacteria from mine drainages were identified through isolation tests, where selective nutrient medium after Silverman, Lundgren and Vaksman (Silverman & Lundgren, 1959) was applied (Geotechnical Institute of the Slovak Academy of Sciences, Košice).

Results and discussion

Pyrite deposit Augustín

The pyrite-pyrrhotite deposit of Augustín is situated in the upper part of the Hrubá dolina valley (Fig. 1), 8.2 km NW from Pezinok. The most intensive oxidation, dissolution and acidification occur in the ore-bearing black shales of dumps and outcrops. The sites 1 and 2 were situated where the pH of stream and pit water is 2.5-3 (Tab. 1). Chemolithotrophic bacteria, *Thiobacillus ferrooxidans*, *Th. thiooxidans* and *Leptospirillum ferrooxidans* (Fig. 4) was isolated under these acid conditions. The occurrence of bacterial cells confirms that intensive biooxidation of the sulphides takes place. Stream water in the surroundings of this locality does not contain large amounts of dissolved contaminants (Tab. 1). Secondary phases precipitate from acidic solutions directly in the dump débris, predominantly gypsum and jarosite. Young iron oxyhydroxide precipitates form in very small amount on the bottom of brooks, usually close to the source, i.e. right below the dumps. The phases are poorly ordered, nevertheless, they exhibit features typical for schwertmannite - $Fe_8O_8(SO_4)(OH)_6$ (Fig. 2a, b). Although the conditions are favourable for its precipitation (Bigham *et al.*, 1994; Bigham *et al.*, 1996a; Bigham *et al.*, 1996b) and major part of the material is oxalate-soluble (Tab. 4) it does not contain appropriate amount of SO_3 (Tab. 2 and 4). According to Bigham *et al.* (1994), the Fe_{tot}/S_{tot} mole ratio for schwertmannite can vary from 5 to 8. Other elements and SiO_2 show low or trace contents, too, (Al, Ti, P, Zn, Cu, Pb etc.) in the samples (Tab. 3). Due to the common paucity of precipitates, it was not possible so far to obtain samples for further analyses.

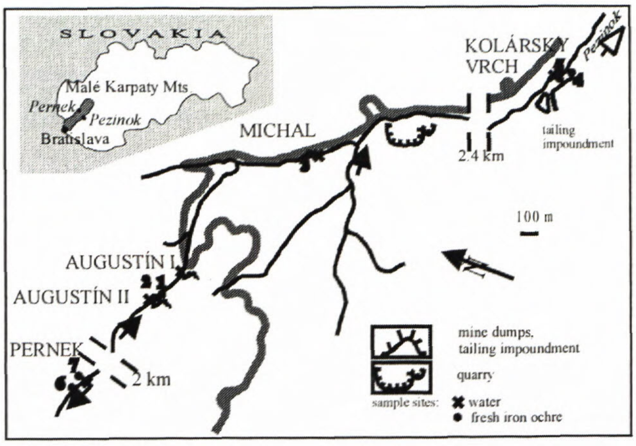


Fig.1 Schematic map of the mining area in the Malé Karpaty Mts. showing location of the main sampling sites

Pyrite deposit Michal

Pyrite deposit Michal (Fig. 1) is situated 1 km southeast of the Augustín locality. These two deposits are connected by old adits which are inundated by waters containing dissolved metals. The chemical composition of the water is similar, however, with pH of 6.3-6.6 (Tab. 4). The most massive sedimentation of iron ochres takes place in the Michal gallery effluent stream (site 3, Fig. 1), and forms a red-coloured wetland covering the area of about 400m^2. Ochres absorb considerable amount of Al_2O_3 and SiO_2 (12 - 13 wt%) but are poor in SO_3 (Fe_{tot}/S_{tot} = 156.67, Tab. 4). X-ray diffraction pattern of the sample consists of one broad line at 2.82 A (Fig. 2a) and indicates that the precipitates are very poorly crystallized. The infrared spectra shows the coincidents of

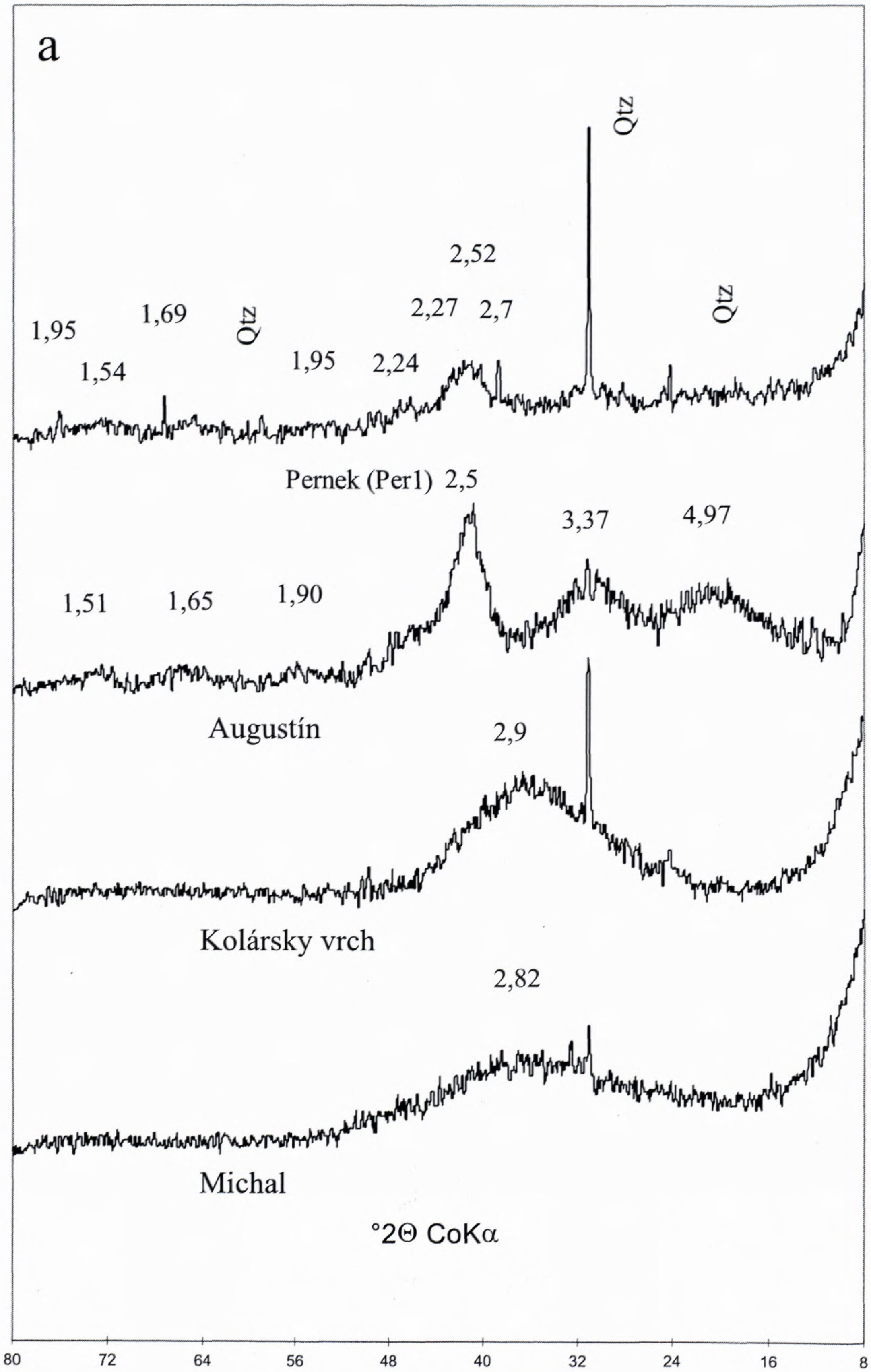


Fig. 2a - X-ray diffractograms of poorly crystalline natural iron ochres from Malé Karpaty Mts.; X-ray diffractograms of Fe oxyhydroxides from acid mine drainages, published data: b - schwertmannite (Sh), c - goethite (Gt) (Bigham et al., 1996b), d - ferrihydrite (6 and 2 lines ferrihydrite) (Schwertmann & Cornell, 1991).

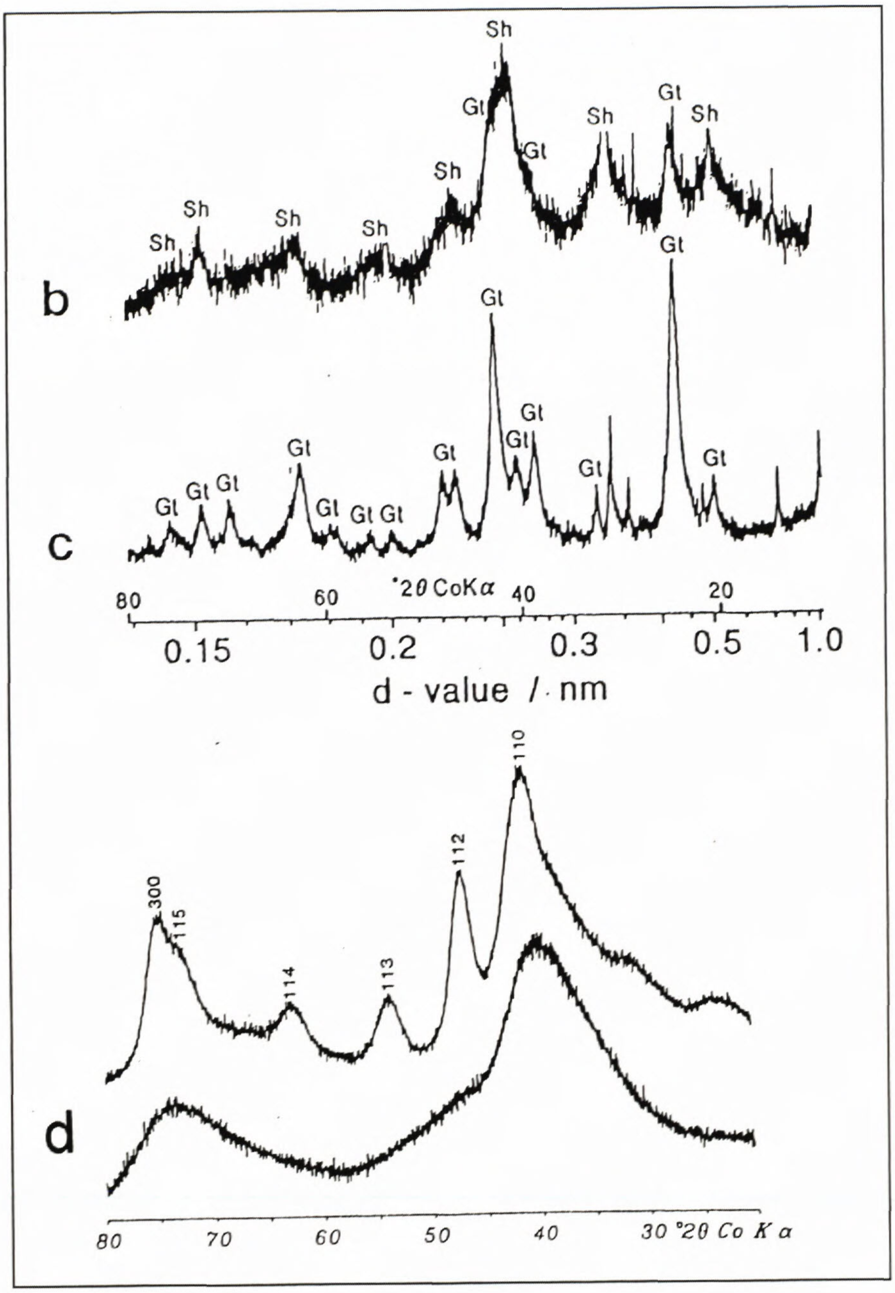


Fig. 2b-d

absorption bands of oxyhydroxides with more intensive bands of kaolinite (Fig. 3a). But the occurrence of Al and Si in the solution after selective dissolution (Tab. 2) is due to sorption of elements on oxyhydroxides. Ochres consist mainly of oxalate-soluble phase (Fe_o/Fe_{tot} = 0.95) and ferrihydrite - $Fe_5HO_8 \cdot 4H_2O$ then could be their main component.

Sb deposit Kolársky vrch

The Sb deposit with tailings impoundment is situated 3.5 km northwest of Pezinok (Fig. 1). Increased contents of metals in the mine waters of this area refer to intensive oxidation (Letko *et al.*, 1992). The pH of this water is 7.5-8 (Tab. 4) because the acidity here is buffered by the high content of carbonates in the tailings impoundment. The concentrations of SO_4, As, Sb, Mn, Na and K (Tab. 1) in the water of seepages (site 4) and tailing effluent (site 5) are significantly higher than in other deposits. The contents of As_2O_5 and SO_3 in iron ochres is very high (Tab. 2), sum of As_2O_5 and SO_3 is higher than Fe_2O_3 contents, Fe_{tot}/S_{tot} mole ratio is 2.53–2.86 (Tab. 4). The material which consists of colloid-size particles (Fig. 5) is poorly crystalline, an X-ray diffractogram shows a wide maximum at 2.9 A, in the region of Fe oxyhydroxides (Fig. 2a). This sample is highly soluble in ammonium oxalate (Tab. 4). The infrared spectra of a precipitates (Fig. 3a) show a sharp feature at 814 cm^{-1}, which is

Tab. 1 Water chemistry of the district (location of sample sites 1 - 7 are shown on Fig. 1). Former data (*) are from Šucha et al. (1996).

Sample site	Fe_{tot}	SO_4	Al	As	Sb	Cu	Zn	Mn	Ca	Na	K
	ppm										
Augustín II*	*3.38	*295.30	*2.11	*0.00	*0.00	*0.13	*0.14	*0.41	-	-	-
Augustín II	0.75	-	-	-	-	4.15	1.90	0.75	9.02	6.35	0.80
Augustín I	*28.61	*761.70	*31.03	*0.00	-	*0.77	*1.27	*0.79	-	-	-
Michal *	*10.44	*280.20	*2.84	*0.01	*0.00	*0.06	*0.37	*0.42	-	-	-
Michal	0.00	-	-	-	-	0.00	0.10	0.20	12.17	8.40	0.80
Kolársky vrch	0.08	2349.60	<0.10	3.35	2.51	0.10	0,10	3.15	24.19	23.20	22.40
Pernek	0.03	662,50	<0.10	0.00	0.31	0.08	0.00	0.08	107.50	4.93	0.90

Tab. 2 Chemical composition of iron ochres, main major components, 1 and 2 - Augustín deposit, 3 - Michal deposit, 4 and 5 - Kolársky vrch deposit, 6 and 7 - Pernek deposit, (location of sample sites 1 - 7 are shown on Fig. 2).

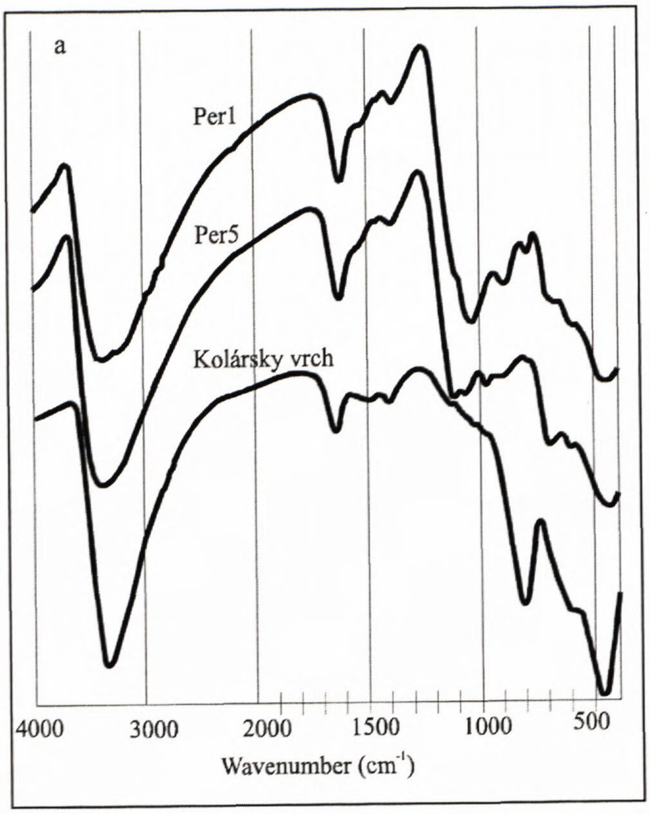
Sample site	Fe_2O_3	SiO_2	SO_3	Al_2O_3	As_2O_5
	wt %				
1.	50.34	1.84	1.12	0.68	0.05
2.	46.33	2.52	0.82	0.59	0.02
3.	37.18	12.88	0.23	12.47	0.02
4.	34.68	2.57	13.31	0.32	24.07
5.	36.56	6.91	12.51	1.08	21.40
6.	51.19	19.17	3.31	8.88	0.06
7.	55.63	10.10	2.32	3.74	0.06

Tab. 3 Chemical composition of iron ochreous precipitations, minor and trace elements,1 and 2 - Augustín deposit, 3 - Michal deposit, 4 and 5 - Kolársky vrch deposit, 6 and 7 - Pernek deposit, (location of sample sites 1 - 7 are shown on Fig. 2).

Sample site	Sb	Cu	Zn	Pb	Cd	P	Ti	Mn	Ca	Na	K
	ppm										
1.	239	80	400	25	13	3200	2800	0	396	600	40
2.	236	80	600	18	4	2700	2500	0	136	640	80
3.	191	720	840	24	12	2000	2400	80	4352	640	80
4.	527	0	606	19	5	2900	2300	1980	2832	727	323
5.	490	0	825	16	4	1600	1800	412	10837	701	124
6.	322	523	243	37	0	1200	1100	597	1800	467	131
7.	320	516	536	39	0	1000	600	1231	0	318	60

Tab. 4 Properties of natural Fe ochreous precipitations. (*) – mole ratio, Fe_o – Fe extractable in oxalate, Fe_{tot} - Fe extractable in HCl, 1 and 2 - Augustín deposit, 3 - Michal deposit, 4 and 5 - Kolársky vrch deposit, 6 and 7 - Pernek deposit, (location of sample sites 1 - 7 are shown on Fig. 2).

Sample site	pH	Fe_{tot}	Feo	SO_3	Fe_o/Fe_{tot}*	Fe_{tot}/S_{tot}*
	wt %					
1.	2.5-3	35.20	34.16	1.12	0.97	45,00
2.	2.5-3	32.40	29.28	0.82	0.90	58.00
3.	6.3-6.6	26.00	24.75	0.23	0.95	156.67
4.	7.8-7.9	24.24	24.36	13.31	1.00	2.53
5.	7.5	25.57	24.19	12.51	0.95	2.86
6.	6.3	35.80	32.83	3.31	0.92	16.00



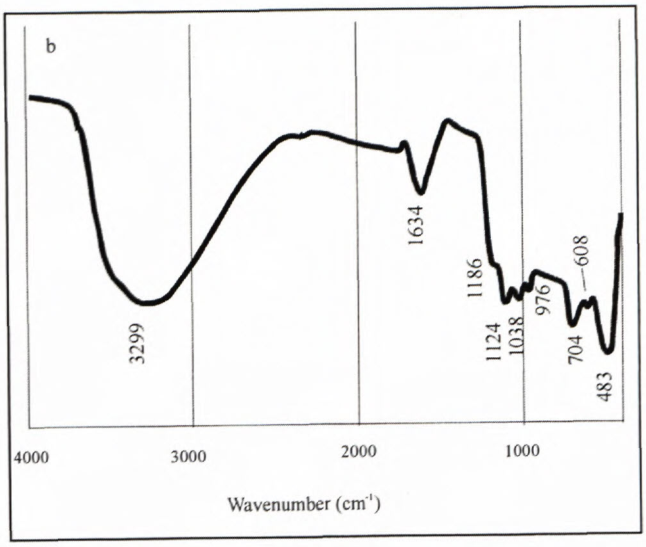


Fig. 3 Infrared spectra of the mine precipitates: a: IR spectra of natural schertmannite (Per1) and goethite (Per5) from Sb deposit Pernek and Fe arsenate-sulphate from Sb-Au deposit Kolársky vrch; b - IR pattern of schwertmannite (Bigham et al., 1994)

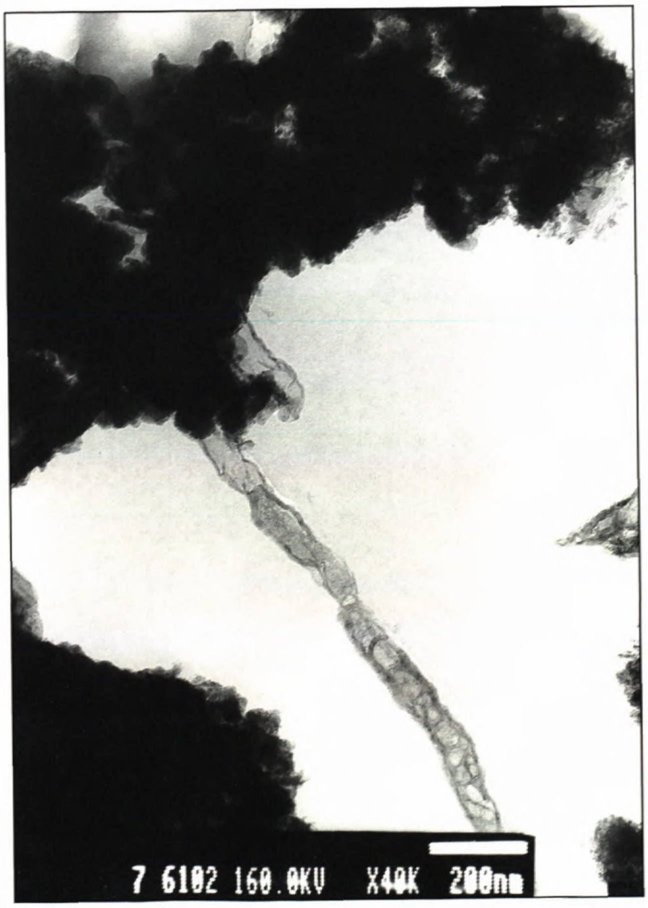


Fig. 4 Cells of acidofilic bacteria identified in acid mine water, Augustín deposit.

assigned to AsO^{3-} or AsO^{4-} compounds. The weak absorption bands at 606 and 982 cm^{-1} and in the 1060-1180 cm^{-1} range likely are attributed to SO_4. The spectrum is dominated by a very intensive feature at 486 cm^{-1}, which can be attributed to Fe-O stretch. It is likely that this material is composed of a mixture of Fe arsenate with Fe sulphate. The infrared spectrum of this sample is comparable to the infrared pattern of synthetic and natural precipitates described by Carlson & Bigham (1992). According to them, AsO_4^{2-} displaces SO_4^{2-} from the surface of oxyhydroxysulphates under acidic conditions.

Sb deposit Pernek

The Pernek-Krížnica locality (Fig. 1) is the oldest Sb deposit in the Malé Karpaty Mts. and it is situated about 3 km northeast of Pernek. No distinct acidification in the surface water was recorded and mine effluents did not exhibit consequential pollution (Tab. 1). Inundated groves where pH varies between 5.0-6.5 (Tab. 4) „produce„ a huge amount of iron ochres rich in SiO_2 and Al_2O_3 (Tab. 2). X-ray diffraction analyses consist of weak and broad lines and of distinct peaks due to detritical quartz (Fig. 2a). The infrared spectrum of sample PER 1 (site 6, Fig. 1) is likely due to the presence of schwertmannite and is comparable with one described by Bigham *et al.* (1994) (Fig. 3b). The spectrum is dominated by bands at 420-460 cm^{-1} (Fig. 3a) and together with broadened feature at 695 cm^{-1}, it can be attributed to Fe-O stretch. Absorption bands at 978 cm^{-1} are likely due to ν_1 (SO_4) and at 603 cm^{-1} are due to ν_4 (SO_4), which belong to structual SO_4. Features at 1118 and 1074 cm^{-1} can be assigned to the splitting of the ν_3 (SO_4), due to the formation of the bidendate bridging complex between SO_4 and Fe (Bigham *et al.*, 1994). A band at 1186 cm^{-1} is not present in spectrum of the sample and it is likely coincided. The promi-

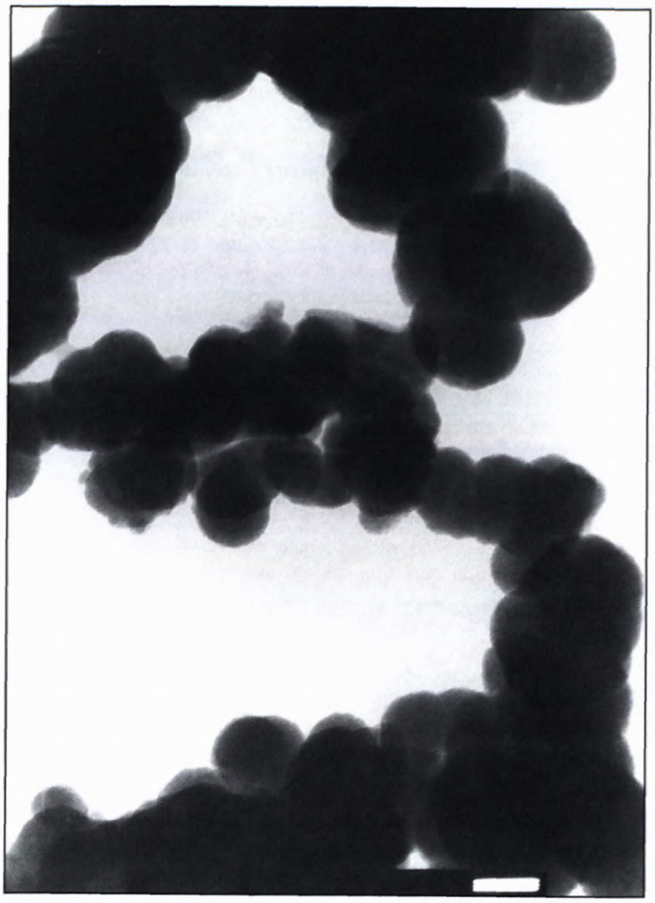


Fig. 5 Colloid particles of As bearing poorly crystalline precipitates, Sb deposit Kolársky vrch (sites 4 and 5, Fig. 1). Bar scale corresponds to 100µm.

Fig. 6 Aggregates of schwertmannite from ochreous precipitates, Sb deposit Pernek (sites 6, Fig. 1). Bar scale corresponds to 100 µm.

nent absorption band at 3382 cm^{-1} is attributed to the OH stretching band and the feature at 1625 cm^{-1} to the H_2O deformation. The Fe_{tot}/S_{tot} mole ratio of a mixture with probable schwertmannite is 16 (Tab. 4) and it occurs in more acidic mine effluents. This mineral consists of globular aggregates (Fig. 6), that are a typical feature of schwertmannite. The infrared spectum of sample PER5 (Fig. 3a) is due to the presence of goethite. The characteristic goethite bands with wavenumbers 899 and 800 cm^{-1} are assigned to the OH deformation (Farmer, 1974). The position of these features is shifted to the higher wavenumbers related to Al-Fe substitution (Schwertmann & Cornell, 1991). The intensive band centered at 3382 cm^{-1} may be assigned to the OH stretching band. The occurrence of ferrihydrite and Si-phases or SO_4 compound is not excluded from this sample.

Conclusions

Generally speaking, the environmental impact of the acidification is not very pronounced in the study area. Acidity there is buffered by carbonates of hydrothermal origin in the Sb deposits. Contamination of mine drainages in the surroundings of deposits is low except for the Sb deposit of the Kolársky vrch region, where high concentration of SO_4, As and Sb in mine water was determined.

The presence of chemolithotrophic bacteria (*Thiobacillus ferrooxidans, Th. thiooxidans* and *Leptospirillum ferrooxidans)* in the acid mine drainages with a pH of 2.5–3 indicates their contribution to the oxidation and precipitation processes. The formation and deposition of young iron ochres along the bottom of streams prevails in mine waters with a pH of 5.0–8 (Kolársky vrch, Michal and Pernek deposits).

The chemical and mineral composition of the precipitates varies widely as well, as a consequence of different formation conditions and the type of deposit. Considerable sorption of Al and Si in the ochres was registered in the old pyrite and Sb deposits. Ochres, precipitating at the tailing impoundment of the Sb deposit are highly enriched in As and SO_4. The Sb contents in the mine drainage precipitates never exceeded 530 ppm, even not in the stibnite deposits. Relatively homogeneous is the concentration of Zn and Pb – hundreds and tens ppm. All the samples exhibit a certain sorption of Ti and P (600-3200 ppm), in some samples Na, K, Mn, and Cu also were recorded.

This study refers to the occurrence of poorly crystalline goethite, schwertmannite, and Fe arsenate and Fe sulphate compounds in the mine ochreous precipitates. The formation of ferrihydrite and Si-phases can not be excluded. For more accurate identification of the mineral phases further analytical techniques will be used, such as thermogravimetric analysis and Mössbauer spectroscopy as described by Carlson & Schwertmann (1981); Carlson & Bigham (1992); Bigham *et al.* (1996a); Bigham *et al.* (1996b); Bigham *et al.* (1990); Bigham *et al.* (1994) etc.

The investigation also revealed that iron ochres might be a latent stock of dangerous pollutants, such as As, Al, Sb. This reservoir is growing and constantly supplied with these toxic metals.

Acknowledgments

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Geology and tectonics of the NE part of the Komjatice Depression

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Abstract. The Komjatice Depression is the northeasternmost branch of the Danube Basin. The geologic and tectonic evolution may be followed since the Middle Badenian. The sedimentary fill of the depression consists of the Neogene megacycle. Depositional environment passes from the marine to brakish, kaspibrackish, lacustrine and swamp during the Miocene. The overlying Pliocene cycle is characterized by lacustrine, deltaic and fluvial deposition.

The general NE-trend direction of the axial part of the depositional area was preserved during the entire evolution of the depression. The deposition was controlled by an NW-SE extension. The main controll for depocentres development were the NE-trending Mojmírovce and Šurany fault systems. Brittle faults are probably determined by extensional rejuvenization of the Veporicum thrust plane (Čertovica line).

Key words: Danube Basin, Komjatice Depression, Miocene and Pliocene sedimentation, Mojmírovce and Šurany faults, palaeostress orientation, Čertovica line

Introduction

According to the regional-geological division (Vass et al., 1988) the Komjatice Depression represents the northeasternmost part of the Danube Basin. It is confined by the Tríbeč Mts. in northwest and by the Central Slovakia Neovolcanics area in northeast. From the viewpoint of the regional-geomorphologic division (Mazúr & Lukniš, 1980) the investigated area is a part of the Podunajská pahorkatina Hills and it is restricted by geomorphologic units of the Tríbeč Mts., Pohronský Inovec Mts. and Štiavnické vrchy Mts., (Fig. 1).

As a basic material for evaluation of the geologic-tectonic evolution of the Komjatice Depression we used geological maps at scale 1:50 000 (Ivanička et al., 1998, Nagy et al., 1998, Harčár & Priechodská, 1988), published and archive materials (Ivanička et al., 1988b, Nagy et al., 1998b. Priechodská & Harčár, 1988, Zbořil et al., 1987, Tkáč et al., 1996, Kováč et al., 1994, Kováč et al., 1997) supplemented by our own research. The publication summarizes ideas on the tectono-sedimentary evolution of the area during the Neogene.

Methods and study objectives

The field work was based on maps at a scale of 1:50 000 by Harčár & Priechodská (1988), Nagy et al. (1998), Ivanička et al. (1998). It included a collation of basic sedimentologic, paleontologic and structural data.

The sedimentologic study was performed on suitable outcrops at all places where no data had been collected previously. Structural research was aimed on collecting and analysis of prevailingly brittle deformations. They performed as a basis for interpretation of tectonic regime during the individual periods of the sedimentary record. The methodology of Angelier & Gougel (1979) and Angelier (1979) was applied for the structural data processing.

Results and their Interpretation

Geological setting

The region is underlain by Paleozoic magmatic rocks, Late Paleozoic and Mesozoic sediments and Tertiary sediments and volcanics. The pre-Tertiary sediments are assigned to several tectonic units thrust and faulted in the Middle Cretaceous, during the Alpine orogeny. According to the tectonic division of the West Carpathians (Mišík et al. 1985) the area of the Tribeč Mts., represents the contact between the Tatricum and Veporicum tectonic units (c.f. Ivanička et al., 1998, Hók et al., 1998).

The uplift of the crystalline core of the Tríbeč Mts., which according to apatite fission track dating reached a level of about 4 km below the surface before 28±1 Ma (Kováč et al., 1994), is assigned to the oldest Neoalpine tectonic movements. The inferred thick covering Mesozoic units (the Krížna and the Choč nappes) and the Paleogene sediments together are 3800 - 4200m. The uplift resulted in gravity sliding of the Tatricum toward SE into the area of the present Komjatice Depression. The sliding occurred under brittle-ductile deformation conditions (Hók & Ivanička 1996).

The Middle Badenian clastic sediments filled the Komjatice Depression after erosion of the Paleogene

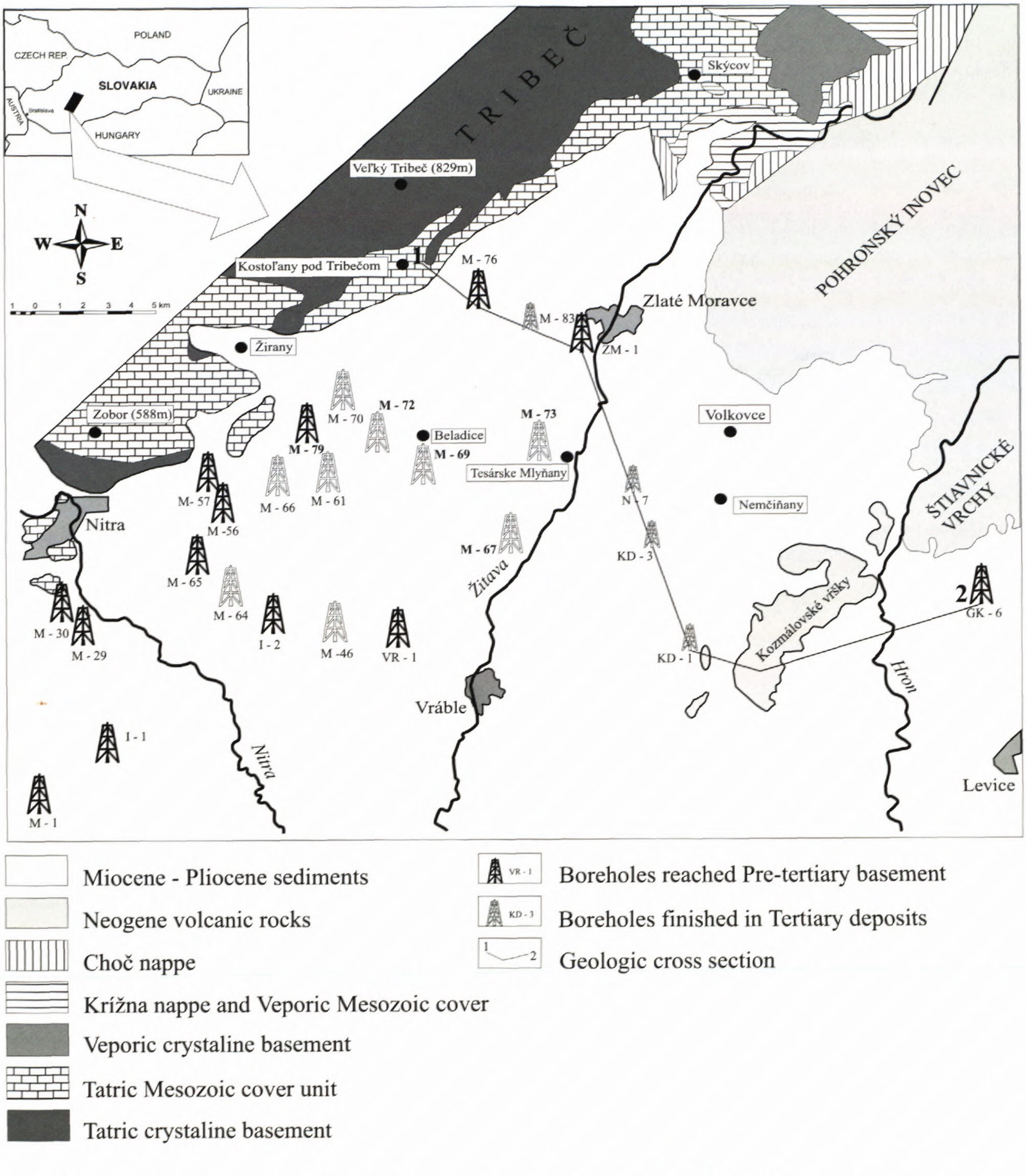

Fig. 1 Generalized geologic map of the Komjatice Depression with locations of boreholes and a geological cross section 1-2 (Fig. 9)

sediments. This implies that the mean exhumation and erosion rate was about 0.1mm/year. The transgressive sediments of the lower part of the Pozba Formation, described from the boreholes VR-1, ZM-1 and GK-6 (Gaža, 1968; 1970; Karolus et al., 1971, in Biela, 1978), exclusively overlying rocks of the Choč Nappe. The lowermost part of the stratigraphic succession consists of coarse-grained deposits overlain by a unit of volcano-sedimentary and andesitic volcanic rocks resembling the lower volcanic units of the Štiavnica stratovolcano. The Middle Badenian sedimentary area is bounded by northeast-trending faults and has the shape of a narrow marine bay whose axis parallels the bounding faults (Fig. 3). Except the occurrence in the borehole GK-6, the volcanics occur in a graben along the axial part of the Komjatice Depression which is about a 10 km wide. The narrow graben filled by the Middle Badenian deposits spread south- or southwestward. This is well documented by borehole and geophysical data (Biela, 1978, Zbořil et al., 1987, Gaža & Beinhaureová, 1976).

The marine sedimentation of the Pozba Formation continued during the Late Badenian, when gradual

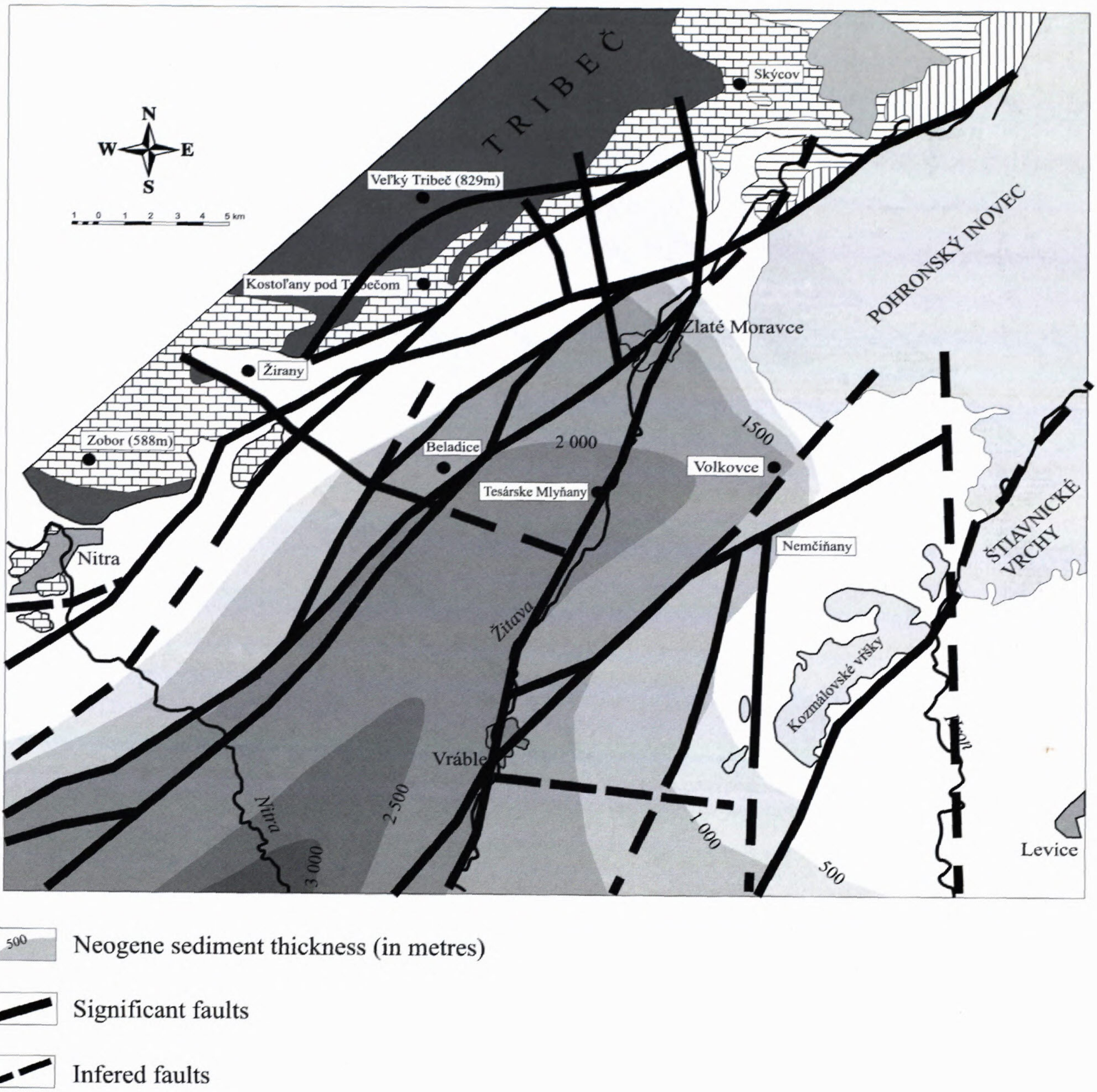

Fig. 2 Thickness of Tertiary deposits of the Komjatice depression and location of faults (after Adam & Dlabač, 1969; Gaža & Beinhauerová, 1976; Zbořil et al., 1987; compiled by Hók, 1999). Pre-Tertiary rocks explanations see (Fig. 1).

upward shallowing of the basin and a lowering of salinity during the final part of deposition took place (Kováč et al., in press), as recorded in the boreholes ZM-1, VR-1 and M-1 (Priechodská & Harčár, 1988).

Stratigraphically younger Miocene deposits gradually transgressed onto various tectonic units of the pre-Tertiary basement. Those tectonic units were denuded and segmented by faults during the Miocene.

The Sarmatian sedimentation area, whose axis trends northeast consistent with the Badenian depocenter axis (Fig. 4). The Vráble Formation sediments transgressively overlie the southeast slopes of the Tríbeč Mts. and volcanics of the Kozmálovské vŕšky Hills. Sarmatian sediments are most probably of the Late Sarmatian age, because they overlie Early Sarmatian volcanic and volcanoclastic rocks. The age of the volcanics is paleontologically determined from the relicts of thermally reworked sediments incorporated to the pyroclastic products of volcanism (Fordinál in Kováč et al. 1997).

The Pannonian deposits (Ivánka Formation) transgressively overlie an angular unconformity above the Sarmatian deposits (Fig. 5). The Ivánka Formation sediments mostly consist of clays, silts and fine-grained sands. During the Late Pannonian the kaspibrackish basin environment gradually degraded and changed to a lacustrine one. In the Pontian the freshwater Beladice Formation was deposited in the environment of a marginal shallow-water bay. The Pontian deposits overly

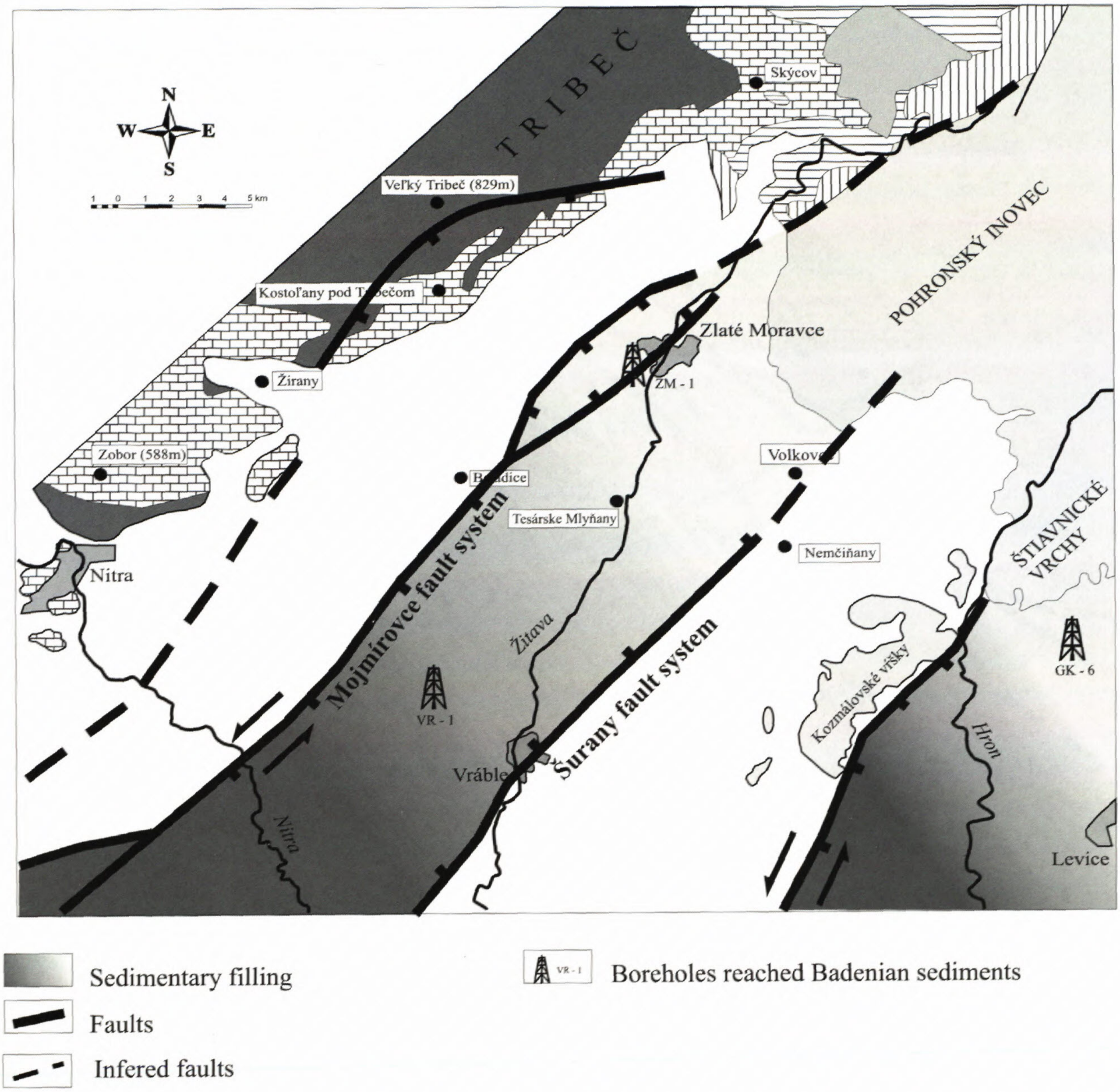

Fig. 3 Middle and Late Badenian - supposed situation

partly transgressively volcanic rocks of the Kozmálovce Hills. However, the sedimentary area narrowed during the Pontian and it did not reach the area of the Pannonian deposits expansion (Fig. 6).

The Pliocene-Dacian Volkovce Formation (Priechodská & Harčár 1968) was formed in a fluvial-limnic environment and especially by development of a fluvial delta (palaeo-Hron river) flowing southwest. This fan delta with a Gilbert delta front was built into the Komjatice Depression (Baráth & Kováč, 1995). Mostly deltaic sediments occur in the area of Mochovce (Fig. 7) while further southwest lacustrine deposition gradually prevailed. The Dacian sediments suggest an enlargement of the depositional area of the Komjatice Depression toward its marginal parts. They transgressively overlie both the southeast part of the Tríbeč Mts. and northwest part of the Kozmálovské vŕšky Hills. The past boundary of the Komjatice Depression was almost identical to the present one.

The Panonian and Pliocene sediments are absent in the area located southeast of the Kozmálovské vŕšky Hills. This suggests a different post-Sarmatian history of northeast flank of the Komjatice Depression comparing to its central part. The Pannonian deposits thick 930 m in the borehole I-2, 915 m in the borehole VR-1, 770 m in the borehole I-1 and 693 m in the borehole ZLM-1, all in the central part of Komjatice Depression (Biela, 1976; Priechodská & Harčár, 1988). The area southeast of Kozmálovské vŕšky Hills was most probably dry land during the Pannonian and Pontian. It again diving during the Quaternary as documented by 40 m thick Quaternary deposits (Tkáč et al., 1996). Based on analysis of the fluvial sediment thickness the calculated average

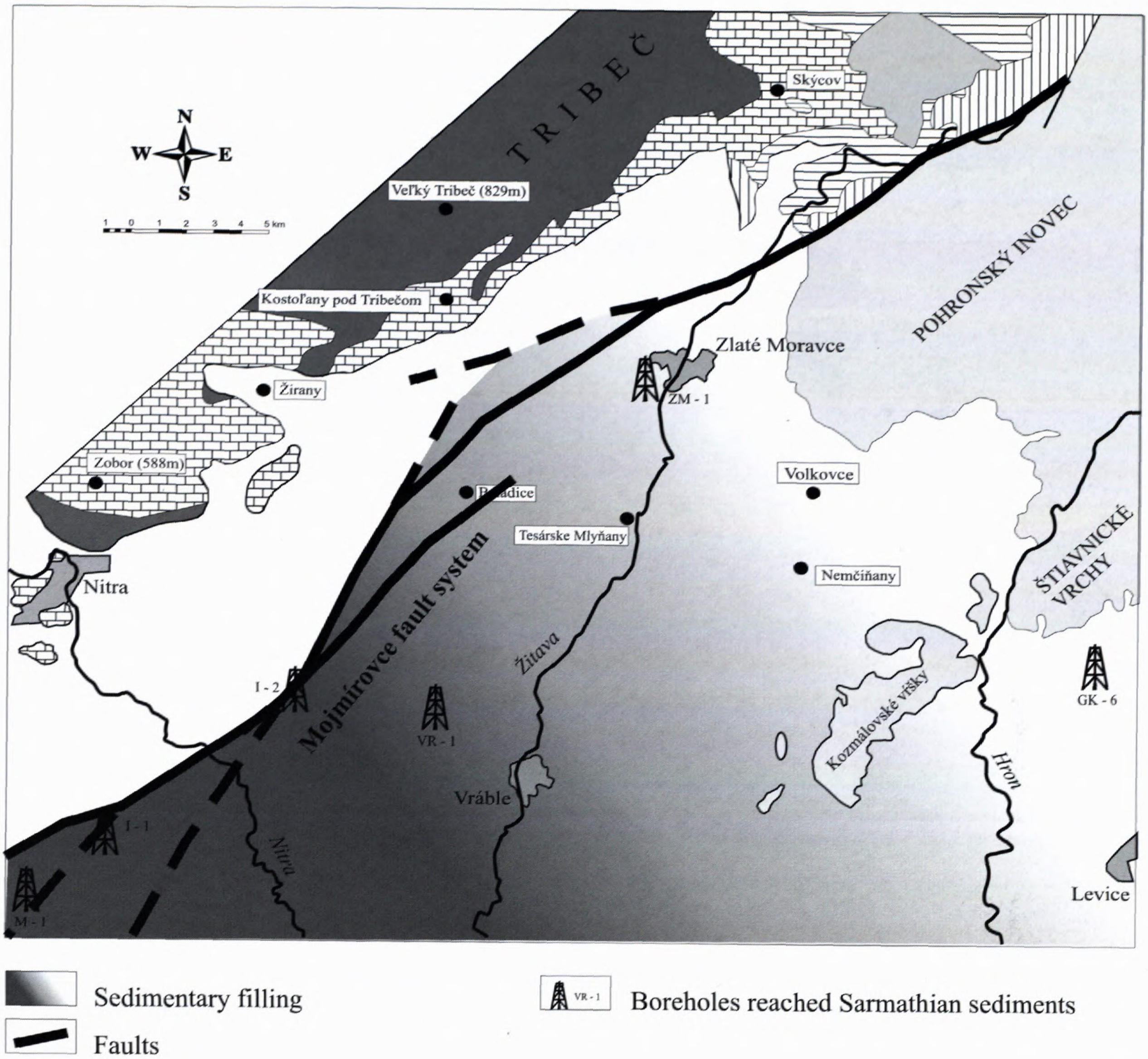

Fig. 4 Sarmatian - supposed situation

subsidence rate is 0.34 mm/year during the Latest Pleistocene (Würm) and Holocene (Durmeková,1998).

Sedimentation

The sedimentary fill of the Komjatice Depression during the Neogene megacycle was characterized by a gradual decrease of salinity of the depositional environment upward. There are several lower-order shallowing upward cycles in the sedimentary record. The depositional environment changed from a marine to brackish, kaspibrackish and lacustrine-swamp with coal deposition during the Miocene. This succession is overlain by a Pliocene cycle, composed of lacustrine, deltaic and fluvial deposits.

The first, marine deposition cycle, is represented by Pozba Formation, with clastic conglomerate and sandstone facies at the base. A facies of gray calcareous clays and silts, locally well bedded, prevails in the overlying Middle Badenian sedimentary succession. Coaly plant remnants are common. This unit is overlain by a clay and silt facies of Late Badenian age in which a sandy admixture increased upward. A vulcanic component appears. A gradual shallowing and salinity lowering is characteristic for the Late Badenian depositional environment. A maximum thickness of the Badenian sediments is 600 m in the investigated part of the Komjatice Depression.

Because the described deposits comprise one sedimentary cycle, we do not consider its original division to Špačince and Pozba Formations (Priechodská & Harčár, 1988, Nagy et al., 1998b) to be convinient. The Špačince Beds (Jiříček, 1985) and Madunice Beds (Jiříček, 1978) were named and defined for the area of the Blatné Depression (the Trnava branch of the Danube Basin), and they represent the Middle-Late Badenian

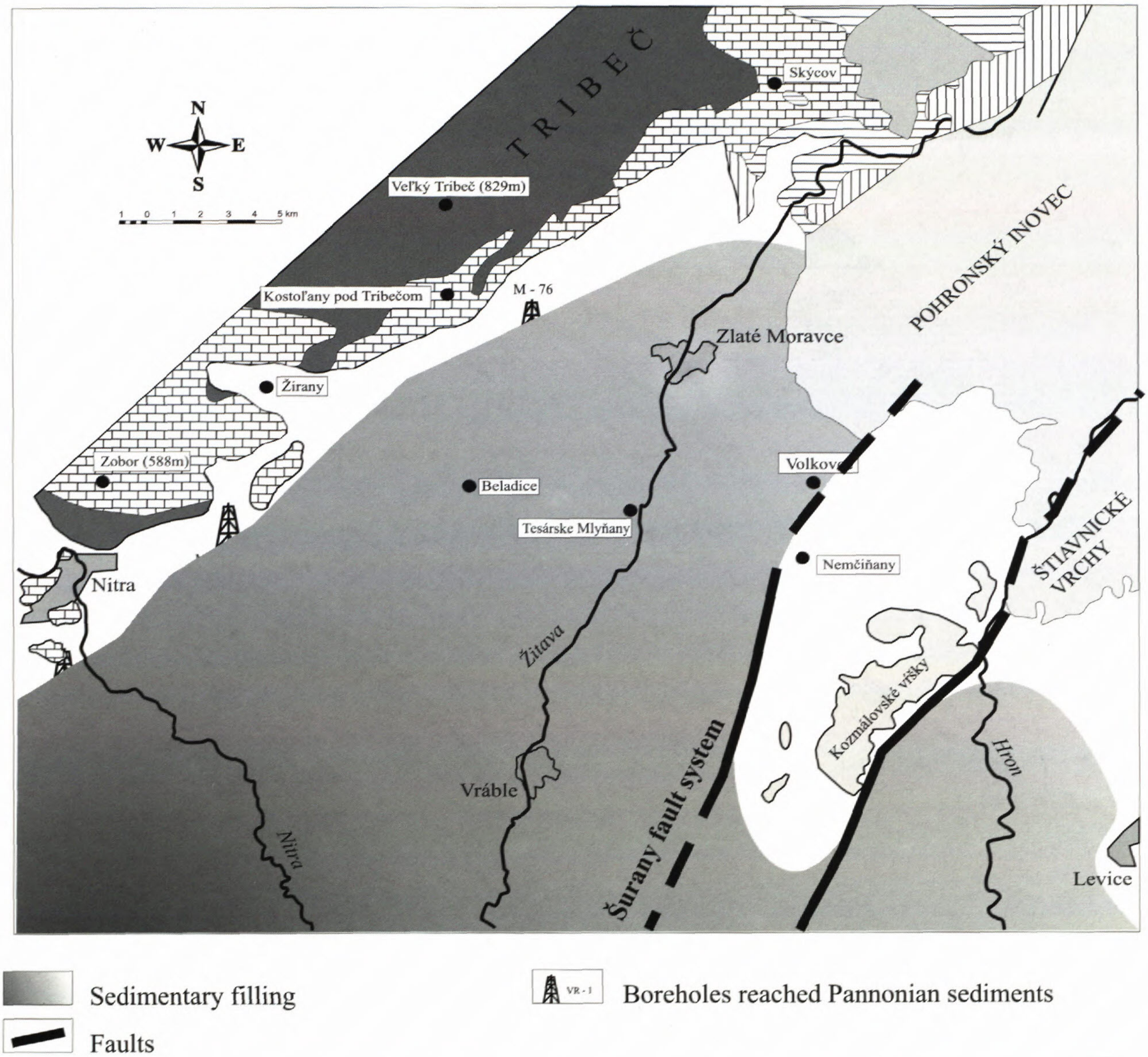

Fig. 5 Pannonian - supposed situation

succession of pelites and sands reflecting depositional conditions like those which occurred in the Komjatice Depression (Zlaté Moravce branch of the Danube Basin) at that time. The basic difference between the above described formations is the occurrence of volcaniclastics in the eastern part of the Danube Basin. For this reason, we suggest a redefinition of the Pozba Formation which would be a time equivalent of the Špačince Formation and Madunice Sands in the western part of the Danube Basin.

Next sedimentary cycle, deposited in a brackish environment, is mostly represented by shallow water deposits of the Vráble Formation (Priechodská & Harčár, 1988). The lower and upper boundaries of the formation are commonly sharp, the uppermost part of the Sarmatian deposits was often eroded before deposition of the Pannonian deposits. The lithology of the formation is very variable in the Danube Basin and it is characterized by the occurrence of volcanoclastics in the Komjatice Depression. It is composed of marginal gravels and sands, organogenic sandy limestones, and basinal sandy and calcareous clays. The maximum 600 m thickness of the formation in the Komjatice Depression was found in borehole VR-1 (Priechodská & Harčár, 1988).

The third, upward-shallowing sedimentary cycle was deposited in the kaspibrackish and fresh-water environments. It is represented by the Pannonian and Pontian sediments of the Ivánka and Beladice Formations (Priechodská & Harčár, 1988). The Pannonian sediments (zones A-E) have a maximum 1 200 m thickness revealed in the borehole Ivánka-1 (Biela, 1978) located in the Komjatice Depression. They are composed of facies of gray pelites commonly containing as much as 80 m thick sand and gravel bodies at the base. The Pontian sediments (zone E-F) were the product of the fresh-water environment and they are also characterized by coal seams. They mostly consist of gray and greenish-gray

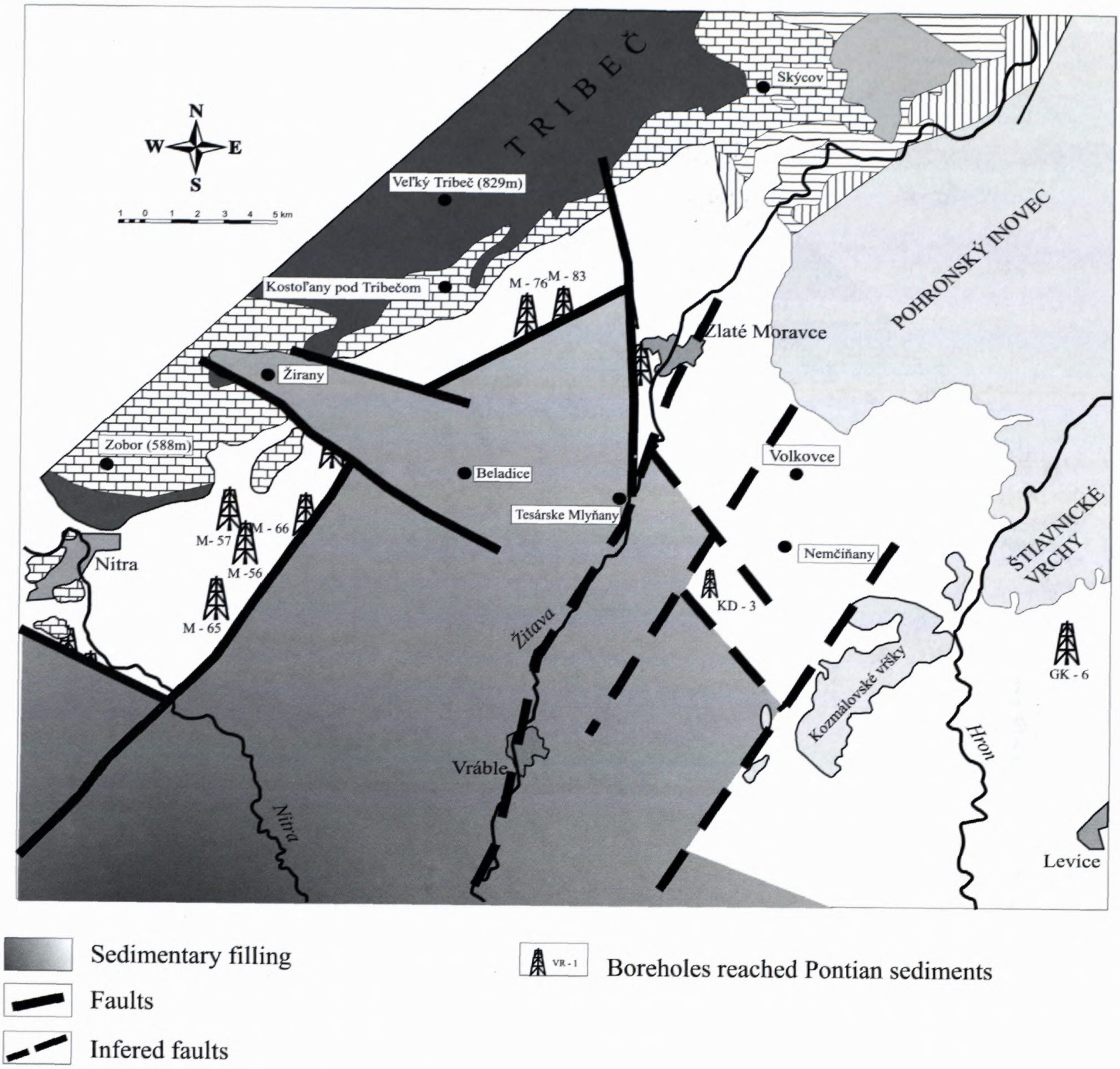

Fig. 6 Pontian - supposed situation

sandy clays and clays. In the boreholes Mojmírovce-1 and Ivánka-2 they are as much as 400 m thick (Biela, 1978). However, based on the recent biostratigraphic data from the Vienna and Danube Basins (Rögl et al., 1993, Fordinál & Nagy, 1997, Kováč et al., 1998), both formations are of Pannonian age and they have a maximum 1 600 m thickness in the Komjatice Depression (Adam & Dlabač, 1969). Considering the lithologically identical upper part of the Beladice Formation and overlying Volkovce Formation consisting of variegated, yellowish-brownish mottled claystones with spare organic remnants and a lack of micro- and macrofauna in both formations, the boundary between the Pannonian (Pontian) and Pliocene is very vague.

The Pliocene sedimentary cycle, deposited in a prograding delta and lacustrine environments (Baráth & Kováč, 1995), is represented by the Volkovce Formation (Priechodská & Harčár, 1988). The lower lithologic boundary in the Komjatice Depression is given above the last lignite or siderite occurrence. Due to a close resemblance to the underlying variegated beds, the variegated beds of the Volkovce Formation are differentiated by heavy mineral content (change of the source area). Sands and gravelly sands are assigned to the proximal deltaic environment, distal deltaic facies are mostly represented by sandy clays, silts and clays.

The cyclicity of sedimentation in the Komjatice Depression is also well substantiated by curves of basin subsidence history. The curves show gradually fading out trend since the Late Badenian with a small rejuvenization of a tectonically controlled subsidence at the beginning of the Pliocene (Lankreijer et al., 1995).

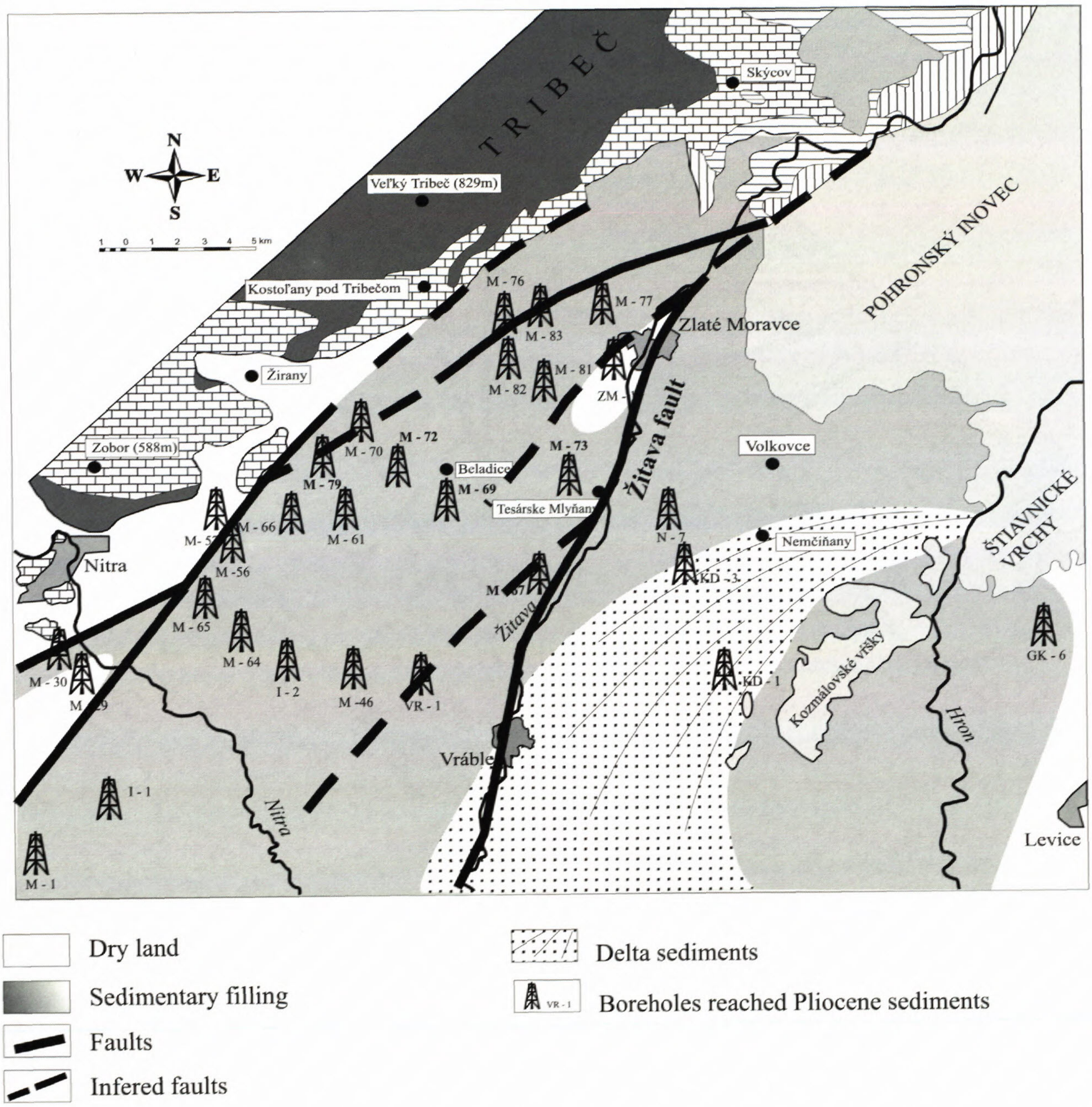

Fig. 7 Pliocene - supposed situation

Tectonics

The northeast trending faults are most notable in the tectonic structure of the Komjatice Depression. These faults played the most decisive role in development of the partial depocenters (branches / bays) on the northern margin of the Danube Basin in the Miocene.

The Čertovica line occurring in the Tríbeč Mts. (at the northeast end of the Komjatice Depression), plays an important role in a tectonic setting of the Komjatice Depression. The Čertovica line is a superficial scar of the Veporicum thrust plane which overlies the Tatricum. Recent structural and geological analysis (Hók et al. 1997), focal earthquake mechanism studies (Pospíšil et al., 1992) and geophysical indications (Bezák et al., 1993, Bezák et al., 1995, Šefara et al., 1998) we interpret the Čertovica line to be extensionally activated overthrust plane of the Veporicum (Fig. 8). The distribution of recent seismic activity, indicated support of this interpretation is only proved in the area of the Middle Slovakia (Šefara et al., 1998). The continuation of this thrust plane from the Tríbeč Mts. to the south/southwest is most likely located beneath the Tertiary sediments of the Komjatice Depression. The location of brittle deformations above the proposed course of the extensionally activated Čertovica line in the Komjatice Depression (Hrušecký et al., 1996, Šefara et al., 1998) plays an important role for the origin of the Komjatice

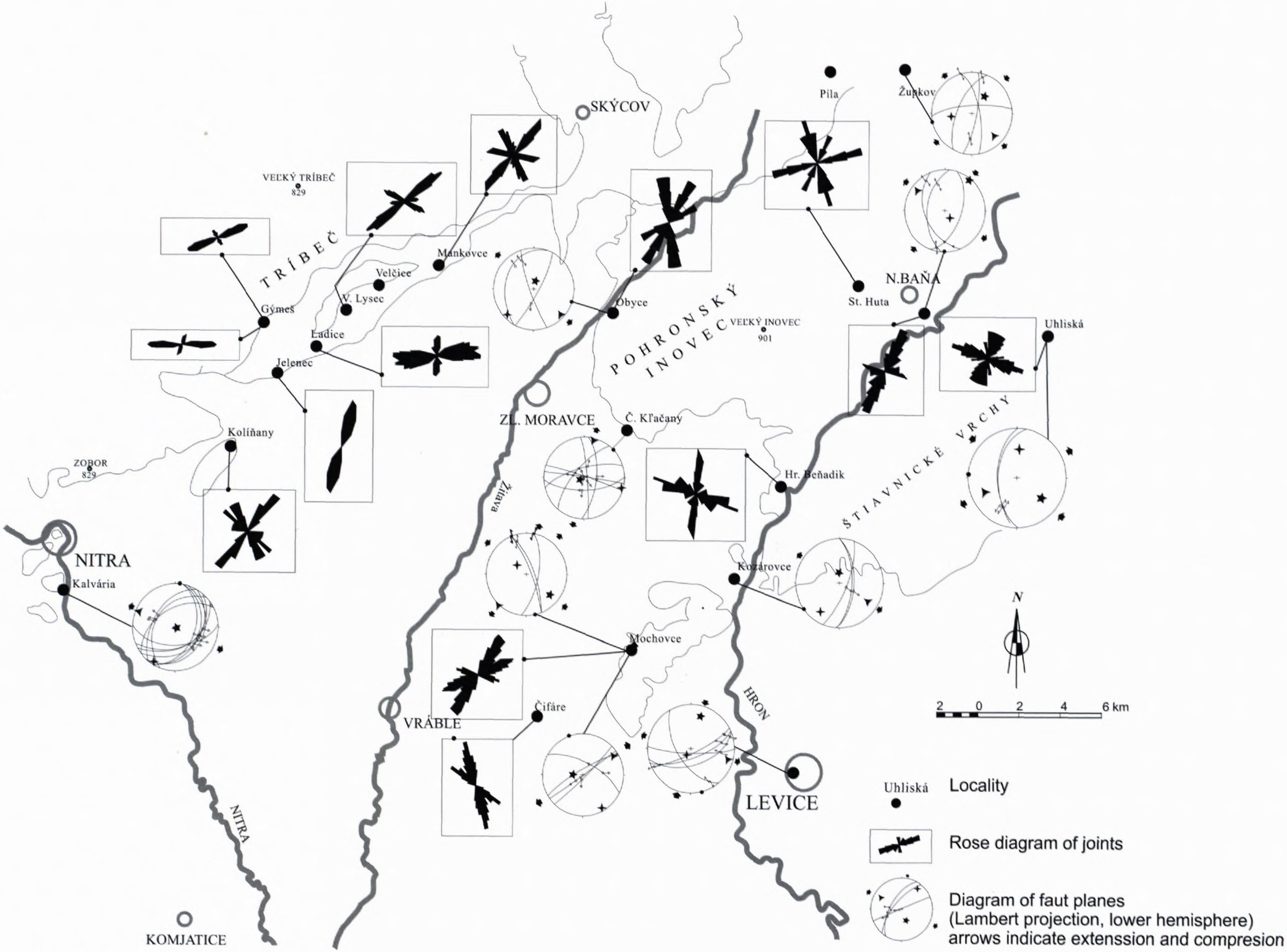

Fig. 8 Map of structural measurement (Kováč, P. & Hók, J., 1999)

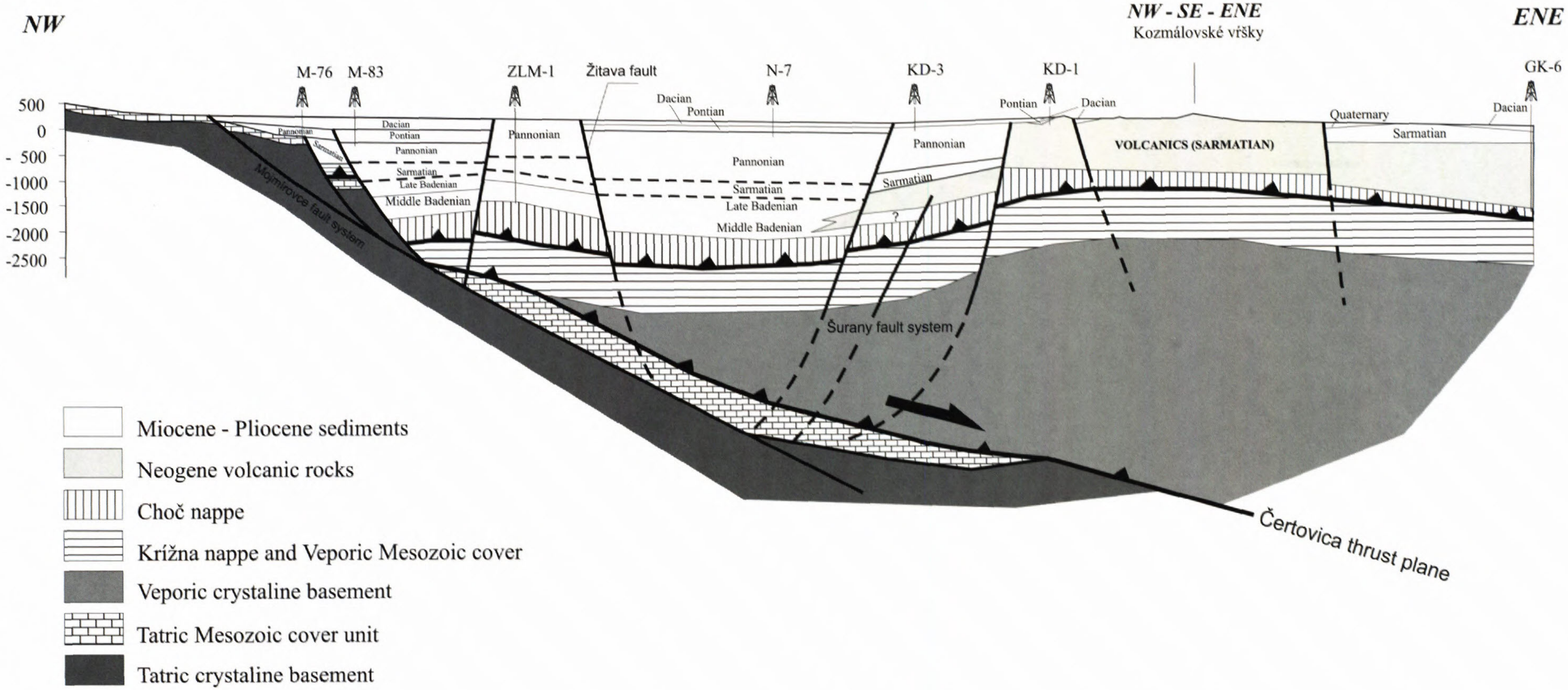

Fig. 9 Geologic cross section 1-2 see (Fig. 1), (Hók, J. & Kováč. P., 1999)

AGE					Lithostratigraphic Unit	Lithology
3,0 Ma	NEOGENE	PLIOCENE	Romanian		Kolárovo Formation	Gravels and sands with sandy clays Gravels and sands with variegated sandy clays
5,4 Ma	NEOGENE	PLIOCENE	Dacian		Volkovce Formation	Light grey and light green clays, brownish sandy clays, sands and gravels
7,1 Ma	NEOGENE	MIOCENE	Pontian		Beladice Formation	Green, bluegreen clays, sandy clays and sands, coal clays and lignite
11,5 Ma	NEOGENE	MIOCENE	Pannonian		Ivánka Formation	Grey calcareous clays, grey sands dark grey claystones greygreen calcareous clays
13,0 Ma	NEOGENE	MIOCENE	Sarmatian		Vráble Formation	Grey calcareous clays, calcareous sands and gravels andezites lava flovs and volcanoclastic rocks
14,0 Ma	NEOGENE	MIOCENE	Badenian	Late	Pozba Formation	Grey calcareous sandstones, marls, sands and sandstones with tuffitic material reefs, Rhyolitic tuffs and andezite lava flows Gravels and sands
15,0 Ma	NEOGENE	MIOCENE	Badenian	Middle	Pozba Formation	
16,4 Ma	NEOGENE	MIOCENE	Badenian	Early		
17,5 Ma	NEOGENE	MIOCENE	Karpathian			
18,8 Ma	NEOGENE	MIOCENE	Ottnangian			
20,5 Ma	NEOGENE	MIOCENE	Eggenburgian			
PRE-TERTIARY BASEMENT						limestones, dolomites, quarzites granitoides

Tab. 1 Lithostratigrapghic column of the tertiary fill of the Komjatice depression (compiled by Kováč, M. & Hók, J., 1999 - according Priechodská & Harčár, 1988)

Depression. This extensionally reactivated thrust plane could generate normal brittle faults in the overlying rock sequences during the Miocene similar to the generation of normal faults above original overthrusts in the central part of the Pannonian Basin (Tari et al., 1992, Horváth, 1993).

The brittle structures of the Mojmírovce and Šurany fault systems were the most important of the faults activated during the development of the Komjatice Depression. The Mojmírovce fault system bounds the Tertiary deposits of the Komjatice Depression with the pre-Tertiary rocks of the Tríbeč Mts. From the Middle Miocene to the Pliocene it is possible to date the age of the fault system, determine its movement direction and the activity of its individual segments. The Šurany fault system is a comparative, partly antithetic fault system relative to the Mojmírovce fault system. It restrains the southeast margin of the Komjatice Depression depositional area and also the horst of the neovolcanic rocks comprising the Kozmálovské vŕšky Hills.

The Žitava fault can be followed from the NE part of the Tríbeč Mts. in the north as far as Vráble in the south. According to the asymmetric development of the Žitava river fluvial terraces (Priechodská & Harčár, 1988) the Žitava fault was active during the Quaternary.

The Mojmírovce fault system was the most significant of the tectonic structures of the area. We can follow its main activity since the Middle Badenian when it controlled the extent of the Pozba Formation deposits in relation to the paleofoothill of the Tríbeč Mts. Considering the approximately NE - SW orientation of palaoestress compression and the perpendicularly oriented extension in the Middle Badenian (Nemčok et al., 1998),we assume that the Mojmírovce fault was active as a sinistral normal fault during the initial rifting of the Komjatice Depression.

A palaeostress field with uniaxial horizontal compressionalstress axis oriented northeast-southwest accompanying subsidence of the depression synrift stage along normal northeast trending faults was documented in the Komjatice Depression during the Sarmatian. The transgressive characteristics of deposits and modelling suggest a moderate subsidence during this period (Lankraijer et al., 1995).

The angular unconformity between the Sarmatian and Pannonian deposits, occurring in marginal parts of the Komjatice Depression, may be regarded as a manifestation of the next phase of a wide rifting markedly applied only in the central and southern part of the Danube Basin (Lankreijer, 1998) at the beginnning of the Pannonian. The Late Miocene palaeostress field, with a NE-SW compression and perpendicularly oriented extension resulted in following subsidence of the depression. Deposition during the Pannonian and Pontian suggests a rapid transition of the synrift stage evolution to final thermal subsidence (postrift evolutionary stadium) of the Danube Basin.

The Pliocene Volkovce Formation deposits, reaching to 1 000 m in thickness toward the Danube Basin centre (Gaža, 1984, Baráth & Kováč, 1995) points to a rejuvenization of the tectonic activity as the Late Miocene paleostress field (late rifting phase development).

However, the end of the Pliocene and the beginning of the Quaternary represent structural reworking of the Komjatice Depression during the period of tectonic inversion. The central part of the depression between the Tríbeč Mts. and Kozmálovské vŕšky Hills is characterized by uplift resulting in erosion and denudation of older units and the formation of typical relief forms, particularly the Late Pliocene peneplenization surface (Priechodská & Harčár, 1988).

The area southeast of the Kozmálovské vŕšky Hills has been subsiding since the Mindelian. This is shown by the age of the oldest preserved terrace of the Hron River considered by Halouzka (1968) as the Mindelian, as well as by the 40 m thick acummulations of Quaternary sediments (Tkáč et al., 1996). During the Quaternary we infer the presence of NW - SE oriented extension. This extension is documented by synsedimentary faults on the southeast side of the Kozmálovské vŕšky Hills (Tkáč et al., 1996; Durmeková, 1998).

Conclusion

The Komjatice Depression is a northeast branch of the Danube Basin. It is possible to follow its tectono-sedimentary evolution since the Middle Badenian clastic deposition, which trangsressively overlies rocks of the Choč Nappe. Sedimentary fill of the depression comprises a Neogene megacycle, consisting of gradually shallowing-upward sedimentary sequences. During the Miocene the depositional environment changed from marine to brackish, kaspibrackish and lacustrine-swamp with coal deposition. The Pliocene cycle, consisting of lacustrine, deltaic and fluvial deposits, overlies the described succession.

The general direction of the depression – the northeast direction of the axial part of the depositional area, was retained throughout the evolution of the depression. The deposition was controlled by NW-SE extension from the Middle Badenian to the Pliocene / Quaternary. The Mojmírovce and Šurany fault systems played the main role during the depocentres generation which was related to the initial rifting phases and synrift and postrift stages. The fault systems acted as normal faults with the exception of the Early – Middle Badenian period when we infer that they had a component of horizontal movement. The brittle deformations are inferred to be partly determined by the extensional rejuvenization of the thrust plane of the Veporicum (Čertovica line).

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Storm-dominated mixed siliciclastic-carbonate „Szin„ ramp (Gľac Unit of the Silicicum Superunit, Inner Western Carpathians): implication for Lower Triassic eustacy

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Abstract. The upper part of the Lower Triassic Szin Fm. in the Gľac Unit (Silicicum) comprises atypical upwards shallowing ramp-slope succession consisting of thin bedded horizontally – to wavy – laminated deposits overlain by storm-dominated, hummocky, cross stratified deposits and capped by channelised intervals of ?intratidal deposits. In the upper part of the succession, small scale compensation cycles were identified.

Bioclastic shoals growing upward in response to inherited sea-floor topography are supposed to have taken part in the carbonate factory action both by means of bioproduction and of modification of the ramp-top relief. The carbonate material was further redeposited due to wave, and especially to storm wave activity.

The overlying Szinpetri Fm. was interpreted to represent transgressive deposits at the base of a Middle Triassic carbonate sequence. Based on sequence stratigraphic analysis, the succession was correlated with the southern Silicicum and the Italian Dolomites.

Key words: ramp, tempestite, compensation cycles, sequence stratigraphy, Lower Triassic, Western Carpathians

Introduction

Origin of a ramp as a depositional environment in a passive margin setting is conditioned by a relatively low ratio of inshore to offshore production and by storm and/or wave induced offshore resedimentation (Aurel et al., 1995). Other factors, such as inherited topography of the basal surface, sediment supply, sediment type and fabric, rate and character of bioproduction, and marine circulation also affect the ramp geometry.

A prograding ramp creates a readily diagnostic facies association. The slope is that part of a ramp with the highest accumulation rate, building up a characteristic upwards shallowing sequence.

According to Burchette and Wright (1992) the „slope crest„ („ramp shoulder„, break-of-slopw, offlap- or ramp break) could correspond either to the shallowest water sediment, as in most rimmed shelves, or to the fair weather wave base transition from inner to outer shelf. The shallow flats behind the slope crest can easily be affected by storm-driven processes. Basinward streams induced by violent storm events are responsible for basinward transport in deeper neritic zone. Landward transport dominates during the vanishing stage of the storm (Rudowski, 1986, Michalík, 1997).

The resemblance to the Bouma sequence is not fortuitous in the storm induced lags, since both storm deposits and turbidites are deposited from suspension by waning currents (Nelson, 1982). While in turbulence, the transported mass behaves as a turbidity current and can create micro- and macrostructures equivalent to those of the deep-water turbidites, even in a relatively shallow water setting (Mutti & Sonnino, 1981). This suggestion remains the matter of an unjustified permanent controversy for all that. Logically, the primary sedimentary structures are responses to depositional agents rather than depositional environments, and the behaviour, not genesis, of the depositing medium is the critical aspect (Swift et l., 1987).

Compared with the ramp slope, sedimentation rates are much lower at the top flats. Generally, they can be related to the subsidence rates of the passive margin. The extremely low-angled extensive shallow flats are exposed to and might be affected by any, even small relative sea-level changes. Basin- or landward shift of facies belts would cause resedimentation by erosion or by accommodation. During periods of minor sea-level falls and lowstands, when ramps experience a loss of accommodation, sediments may be remobilized and rapid pulses of progradation and beveling of the inshore area may occur (Aurell et al., 1995). Infilling of minor highstands accommodation space is read as a parasequence of lower order in the sedimentary record.

In the case of major sea-level changes, the top ramp flats can serve as ideal basal surfaces for the development of superjacent sedimentary bodies.

General setting

The Silicicum Megaunit of the Inner Western Carpathians comprises a complex of N-NW-vergent superficial nappes. At present it is preserved in several isolated parts over an area of about 200 km by 80 km in S-SE Slovakia and NE Hungary.

During Lower Triassic time, a large, homoclinal, low-angled to distally steepened ramp evolved on the passive margin of the Palaeo-Europian shelf (Roniewicz, 1966; Brandner, 1984; Mihalík, 1993 a, b;, Hips, 1996 a, b, 1998 a, b and references therein; Kovácz and Hips, 1998). Extensive marginal seas rimmed the Mediterranean shelf in a belt several hundred km wide. The Silicicum represents the most basinward part of this shelf domain, with a connection to the open marine environment to the south.

The lack of reef-builders in the Lower Triassic time, due to the Permian-Triassic boundary extinction event, was the crucial factor for a prolonged maintenance of ramp conditions (Hips, 1996 a, 1998 a).

Generally, the studied Lower Triassic succession has a fairly uniform lithology over the entire Silicicum. This lithology upwards changing ramp system composed of initially siliciclastic through mixed siliciclastic-carbonate an ultimalely to undiluted carbonate (marlstone) deposits served as a basement for the evolution of an extensive Middle Triassic shallow-water , carbonate, ramp association.

Lithostratigraphy

The Lower Triassic succession of the Gľac Unit in the Stratená Mts. (Inner Western Carpathians, part of the Silicicum Megaunit (Havrila, 1995), (Fig. 1) is represented by Bódvaszilas Fm. (sandy-shaley formation by Maheľ (1957), traditionally known as „Campil Beds„) (Richthoffen, 1859; Bystrický in Andrusov & Samuel, eds. 1983; Kovács et al., 1989; Hips, 1996 a, b, 1998 a, b; Mello et al., 1997). The Szinpetri Fm. comprises the uppermost part of the Lower Triassic succession. Kovács et al. (1989 cf. Hips (1996 a, 1998 a) introduced this lithostratigraphic division for the Lower Triassic sediments (overlying the Permian Perkupa Evaporite of the Silicicum in the Aggtelek – Rudabánya Mts.

Kovács et al. (1989; Hips, 1996 a, 1998 a) interpreted the Bódvaszilas Sdst. (Middle to Upper Griesbachian) as a shallow subtidal to intratidal, partly restricted, flat, coastal sediment. The Szin Marl (Nammalian – Spathian) has been interpreted as open subtidal (shallower and deeper) sediment with varying terrigenous influx.

The Lower Triassic sandy-marly sequence is overlain by Middle – Upper Triassic carbonate sequence, beginning with the Gutenstein Lmst. (Bystrický, 1982; Kovács et al., 1989). Kovács et al. (1989) divided the Szinpetri Lmst. (uppermost Spathian – lowermost Anisian) into the underlying Szin Marl and the overlying Gutenstein Lmst. They interpreted it as a subtidal – intertidal restricted lagoon facies with a decreasing terrigenous influx and with a rich benthos of low diversity, consisting only of worms. Hips (1998 a) reinterpreted the Szinpetri Lmst. in which three main facies were recognized: a storm-sheet mid-ramp facie, a dysaerobic outer ramp facies (Szinpetri Lmst.) and an anaerobic outer ramp facies (Jósvafő Mbr.).

According to Hips (1998), three succeeding depositional systems can be distinguished in the Lower Triassic succession: (1) wave- and storm-dominated siliciclastic shallow-shelf (Bódvaszilas Sdst.; (2) a mixed siliciclastic-

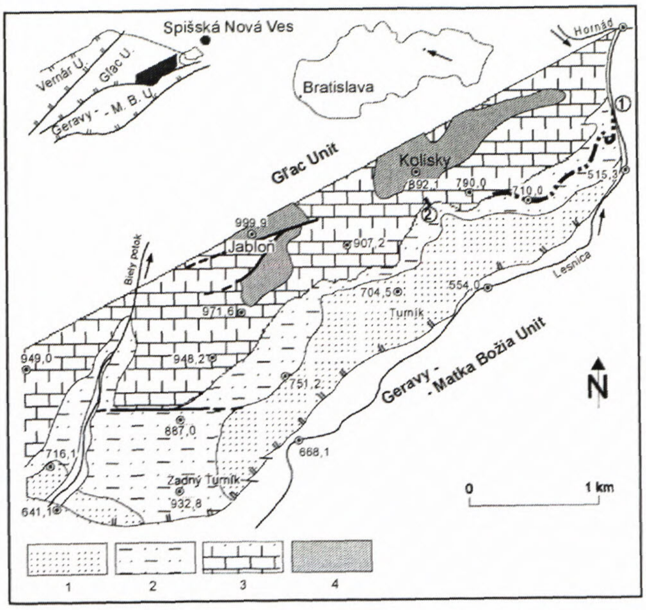

Fig. 1: Geologic map of the study area. 1. Bódvaszilas Sandstone Fm. (?U. Griesbachian-Nammalian) 2. Szin Marl Fm. (Spathian) 3. Gutenstein Fm. (L. Anisian) 4. Steinalm Limestone Fm. (U. Anisian). 1. Pf. Kolísky 1. 2. Pf. Kolísky 2. (traces with dotted line) (Because the Szinpetri Limestone Fm. is relatively thin in the study area it was mapped together with the Gutenstein Fm.).

carbonate, high energy, storm-dominated, ramp system (Szin Marl); (3) and an outer, low-energy zone of a storm-related, carbonate, ramp system (Szinpetri Lmst. ?and Gutenstein Lmst.).

The paleoenvironmental interpretation of the Szinpetri Lmst. will be subject to further discussion in this paper.

As seen from the regional correlation (Fig. 6), this paper deals with the upper part of the Szin Marl and the Szinpetri Lmst. to the base of the overlying Middle-Upper Triassic carbonate sequence.

In the study area (Fig. 1), several outcrops were logged in a detail. Their sedimentologic and palaeoenvironmental interpretation is the aim of the present paper. Well-documented logs and interpretations, such Maheľ & Vozár (1972), Vozárová (1977), Fejdiová & Salaj (1994) , Kovács et al. (1989), Hips (1996 a, 1998 a) from all over the Silicicum's Lower Triassic have been taken under consideration.

Lithology

This contribution explores further the interpretation of the studied succession as a low-angled southerly open ramp system. We term the Szin ramp in accordance with the name of the lithostratigraphic unit into which most of the studied succession belongs.

Because of the general paucity of fossil record, due to a slow repopulation of the shallow seas after the Late Permian world-wide extinction (Hips 1996 a, 1998 a) and thus the lack of an exact biostratigraphy, as well as the still uncertain palaeogeographic and (pre) tectonic position of individual tectonic wrecks of the Silicicum, we are not able to state the spatial relationships of the individual

Fig. 2: Profile Kolísky 1. (At the locality, marly limestone intercalations were post-depositionally changed into diagenetic quartzarenites).

sections with certainty. The profiles at Kolísky 1, 2 (Figs. 2, 3) are shown to illustrate the general facies succession of the uppermost part of the Szin Marl and its gradual transition into the overlying Gutenstein Fm. Based on the lithologic criteria, we also infer the relationship between Szin Marl and the underlying Bódvaszilas Sdst.

The Sin Marl comprises of greenish-gray, purplish-red and dark-gray silty shale, marly shale and schistose clay-marl, which are interbedded with beige, light- to dark-brown, also pink and grey thin bedded (2-5 cm) to platy (5-10 cm) marlstone layers that have marly drapes on the bedding planes. The marly interbeds show evidence for higher energy level, including positive gradation, horizontal, flat wavy to cross ripple lamination, undulating upper surfaces and coarse bioclastic coquina lags on the bedding planes.

At a larger scale (several tens of meters), intervals of predominantly siliciclastic (shaley) deposits with ubiquitous but subordinate marly layers and of almost purely carbonate (marly) deposits occur rhythmically (Figs. 2 and 4). Coquina lags (0.5 – 2 cm) are common within the prevailingly marly intervals.

Generally, the studied succession shows clear evidence for a relative increase of energy upwards. The succession can be divided into three parts on the basis of the energy indicating sedimentary features. The marly intervals single out the parts of different energy levels.

In the lowermost part, predominantly horizontally laminated silty shales with marly intercalations are present (Plate 1, a, b). Above it, an interval of strom-dominated deposits is found (Plate 1 c, d). Farther up, the positive hummocky stratified structures are less frequent and scouring features are common. In the uppermost part of the succession there is a series of major erosion surfaces weith a mud-and-marl multiple-filled channel (3 m deep and 15 m wide) being the most prominent (Plate 1, e). The channel fill represents the topmost part of the last interval of predominantly siliciclastic deposits. It is overlain by ca 30 m complex of platy, horizontally laminated marlstone with finer drapes. Towards the top, considerably thicker (up to 20 cm) coquina intercalations occur. These possibly represent wash-over sheets derived from adjacent bioclastic shoals.

The last marlstone succesion is completely overlain by micritic - biointrapelmicritic limestone with no terrigenous influx. Vermicularly textured limestone layers are interbedded with 2-8 cm thick biointrapelmicritic redeposits and with dark-gray, thick-bedded to massive micrtites of Gutenstein facies (Bystrický, 1982; Kovács, 1989; Hips, 1996 a, 1998 a; Fig. 3).

The vermicular structure (nodular habit or burrow-mottled appearence (Hips, 1996 a 1998 a), „pseudo-nodular„ structure (Fejdiová and Salaj, 1994) may result from several genetically different processes: disturbances due to reverse density gradient (Kasinski et al., 1978), sliding (Kotanski, 1995), bioturbation or subsequent erosion (Michalík, 1997). In this case, the vermicular appearence of certain layers is obviously due to bioturbation by benthic infauna (worms) (Plate 2, b).

Fig. 3: Profile Kolísky 2.

The Szinpetri Limestone Fm. represents a completely different microfacies association from that of the underlying beds. It shows a close lithofacies proximity to the Gutenstein Fm. Its genetic interpretation also indicates that it is related to the overlying Gutenstein Fm. rather than to the Szin Marl Fm.

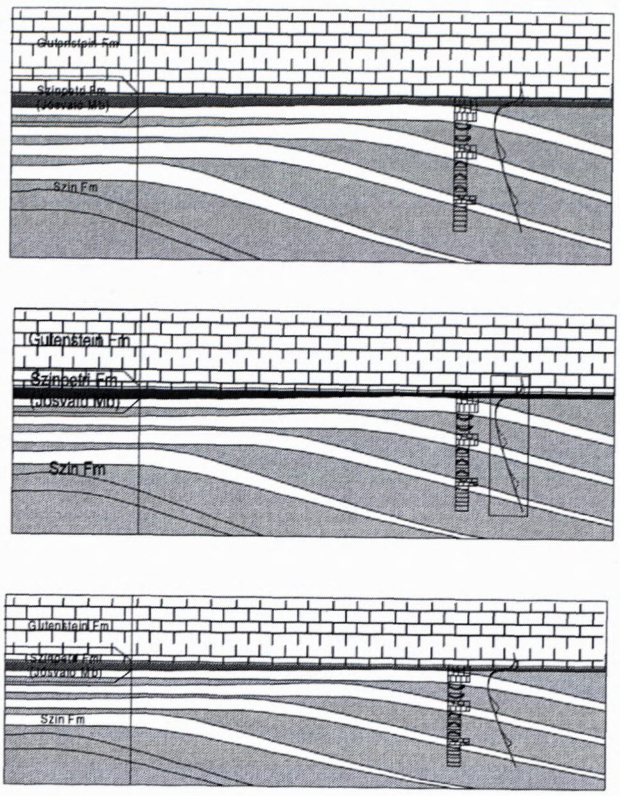

Fig. 4: Schematic cross section through the slightly inclined "Szin ramp" showing position of the profiles Kolísky 1 & 2 within its architectural framework. Marly limestone accumulations (white) are suggested to point out small-scale highstands. The Szinpetri Limestone Fm. represents main transgression.

Evidence for storm induced sedimentary processes

Remobilization of the bottom sediment and its redeposition controlled by storms is criticial in maintaining the ramp profile through time (Aurell et al., 1995; Michalík, 1992, a1997; Lavoi, 1997). Both basin- and landward sediment transport may occur during a storm event. Sea-bed topography depth, distance and sediment type and fabric control the final depositional architecture of storm deposits.

The shallow flats behind the slope crest are the place of major remobilization and erosion while slope (middle ramp) is the place of major accumulation of storm deposits. In the studied section, the marly interbeds show evidence of a rapid, single deposition. In the basal part of the succession, the abrupt change in lithology and grain size, sorting, positive gradation, horizontal, flat wavy and also cross ripple lamination are the basis for interpreting the marly interlayers as distal storm deposits.

Farther upwards, positive domical structures (hummocks) are found. These are very complex in their internal architecture and grain size (Plate 1, c, d).

Above, sedimentary structures are found whose internal architecture shows a close genetic proximity to compensation cycles, which we know from deep-water turbiditic systems (Mutti and Sonino, 1981; the pinch-and-swell structure of Swift et al., 1987).

The remobilized material is transported during storms and post-storm surges and is redeposited inshore or in a deeper region of the ramp in response to seabed topography. While reflecting the topography, the redeposited sediments build up a new small synoptic relief. Thus, an autocyclic process may develop in which each storm-induced sediment sheet is deposited in response to the previous one and predicts the depositonal architecture of the next one. This features indicate the depositon in compensation cycles as illustrated in Plate 2, a.

A sequence of sediments several meters thick that were deposited in compensation cycles is found in the upper part of the present section. Note that this horizon is situated between the hummocky-cross stratified and the scouring/redeposition dominated parts of the section. The channelized uppermost part of the succession represents the environment with the highest energy level.

Note that while the mixed but prevailingly siliciclastic deposits show obvious increases of energy level upwards, the platy marlstones remain structurally the same within the marly intervals throughout the entire succession. Although poorly outcropping in the present profile Kolísky 1, the predominantly marly intervals are well documented in many other sections. Structurally, the planar to flat wavy lamination, positive gradation, pertinent bed thickness, finer muddy marl drapes and especially the common bivalve and gastropods shells coquina lags on the tips of the beds are the evidence for which these can be precisely described as redeposits, most probably related to strom events.

It is inferred that the fine grain size of the silty shales and especially the lithological heterogeneity controlled the origin of well developed, architectonically variable storm redeposits within the prevailingly siliciclastic intervals (Plate 1). The lithologically homogenous, relatively high viscose marl did not allow such morphological variability of final accumulations and thus the well-bedded marl sheets were deposited.

Sediment source

The location of the direct sources of both siliciclastic and carbonate material is ambiguous. An extensive southerly thicking wedge of siliciclastics is thought to have rimmed the passive margin of the Palaeo-European shelf during the Upper Permian - Lower Triassic times (Roniewicz, 1966; Brandner, 1984; Michalík, 1978, 1993 a, b, 1994; Kázmer and Kovács, 1985; Hips, 1996 a, 1998 a and references therein). This idea is constrained by evidence of facies grading of fluvial and coastal to shallow and deep water open marine sediments within the Permo-Triassic successions of the superimposed nappe systems of Western Carpathians. The rapidly eroded „Vindelic Land„ is believed to have been the primary source area. (According to petrographic analysis by Vozárová (1977) and Fejdiová and Salaj (1994) the provenience falls between the recycled orogeny / magmatic fields when plotted in the triangular diagrams (c.f. Dickinson and Suczek, 1979).

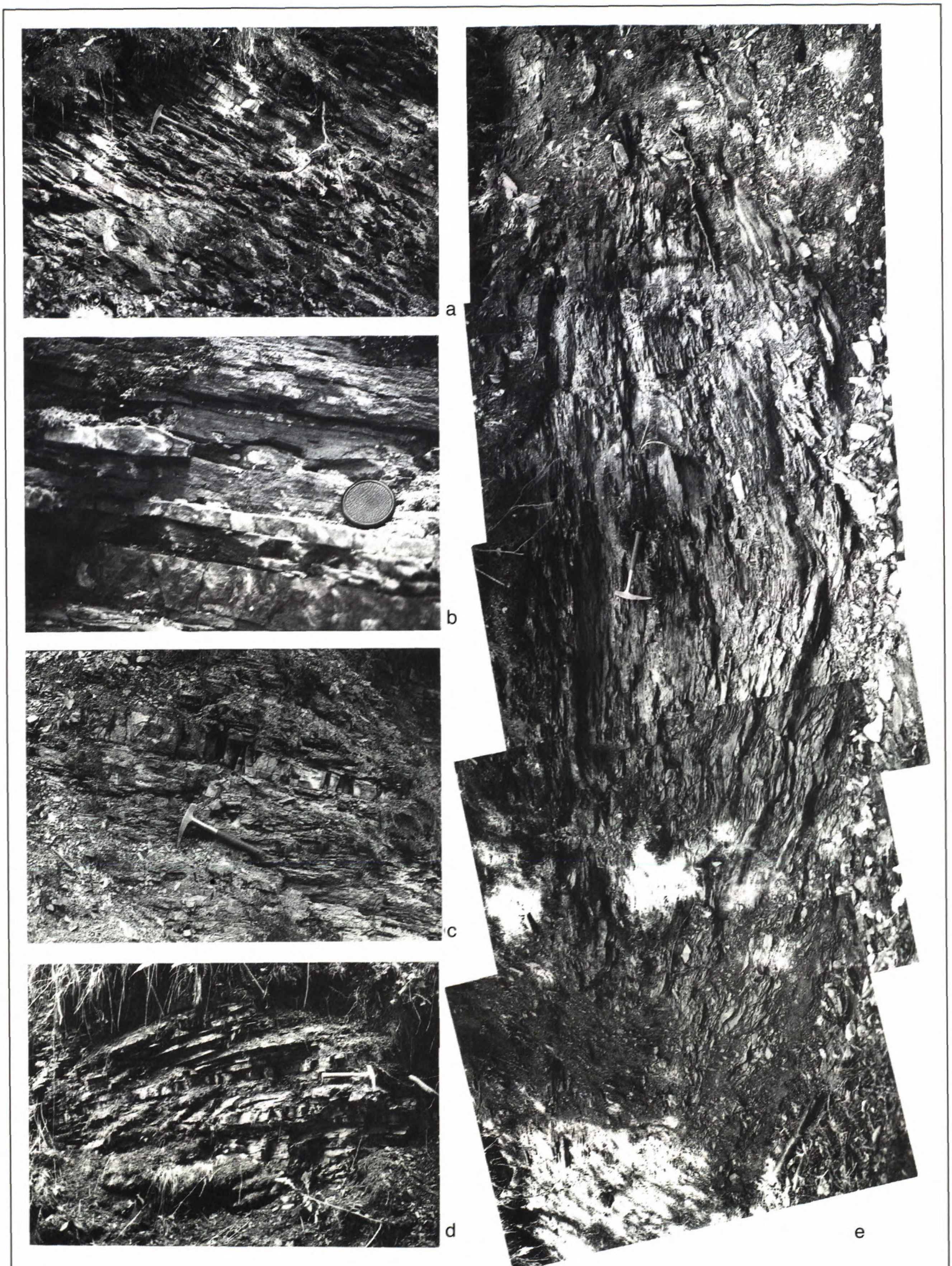

Plate I

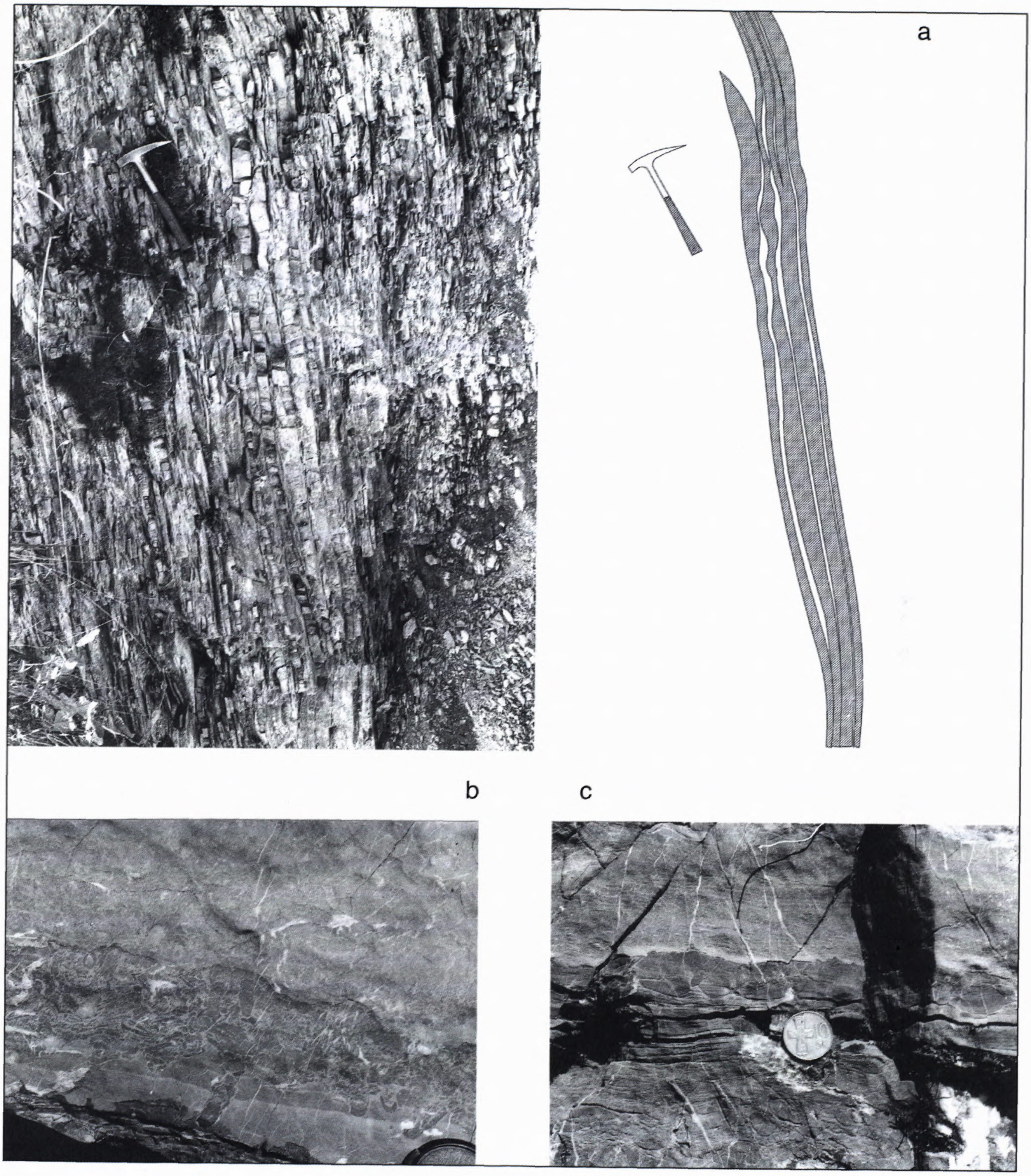

Plate II a. Detail view on a small scale compensation cycle in the upper part of profile Kolísky 1 (see Fig. 2 for exact localization). Sediments carried by marl-laden debris flows and/or turbidity currents induced by storms from adjacent lime mounds were deposited in dependence on autocyclically conditioned changes in sea bed topography. Note the lateral shift of apical, thickest parts of successive undulating layers (?turbidites). Deep scour of unknown origin on the right side of most such domatic structures is chracteristic in this part of the outcrop. **b, c:** Szinpetri Limestone Fm. (Jósvafő Limestone Mb) in the overlier of the Szin Marl Fm. **b.** Vermicular appearance is due to intense bioturbation by benthic infauna. **c.** Thoroughly bioturbated layer is scoured by biointrapelmicritic redeposite.

←

Plate I Sedimentary structures and depositional geometries of deposits preserved in profile Kolísky 1. See Fig. 2 for exact locallization at profile. **a.** Thin-bedded horizontally laminated shales with marly (diagenetic quartzarenite) intercalations. **b.** Marly (diagenetic quartzarenite) intercalations with undulating upper surface. **c, d.** Hummocky cross-stratified deposits in the middle part of the profile. Some hummocks are as large as 5 m wide and 1 m high (hammer for scale is 33 cm long) showing very complex internal structure. **e** About 15 m wide and 3 m deep channel in the uppermost part of the profile is filled mainly with shales, with marlstone (diagenetic quartzarenites) and biodetritic marly limestone intercallations becoming more common and thicker upwards.

In the Permian Perkupa Fm. and Lower-Triassic Bódvaszilas Sdst., which underly the studied Szin Marl, there is a clear evidence for terrestrial sedimentation. Most of the evaporitic Perkupa Fm. was deposited in a sabkha and playa type setting (Vozárová, 1979, 1996; Kovács et al., 1989; Hips, 1996 a, 1998 a). The sediments were frequently redeposited, possibly due to seasonally flooding streams. In the above, the Bódvaszilas Sdst. shows features typical for shallow subtidal to intertidal sediments. Among them, rain drops imprints on the bedding plains, dessicaton cracks, heringbone cross stratification and conglomerate-filled channels are common (Vozárová, 1997; Kovács et al., 1994; Hips, 1996 a, 1998 a).

Thus, local sources in negligibly elevated parts of the shelf domain can be taken under consideration. These local elevations could have served as additional sources of redepoisted siliciclastic material or as the places suited for later shallow-water carbonate-producing shoals evolution.

Origin of the cyclic marlstone-shale alternation

The effect of a periodic alternation of prevailingly siliciclastic (shaley) and prevailingly carbonate (marly) intervals (Fig. 2) can be linked to various controls. Progressive aridization succeeded the Lower Triassic humid event. The periodicity may thus reflect the oscillatory change of climatic conditions. The relative increase in fine-grained siliciclastic influx may have been related to brief partial shutdown of the carbonate factory during the relatively humid periods. Carbonate production may have recovered on the ramp top in response to subsequent humidity decrease.

As noted above, the shallow flat inner ramp areas are well exposed and would be affected by minor sea-level changes. Land- or basinward shift of facies belts related to erosion/accommodation effects must be considered while discussing the variable relationships among possible sediment sources, sea-level fluctuation and seabed tography. The role of relative sea-level fluctuation is constrained by two lines of evidence: a) the abrupt but long-lasting change in lithology of b) obviously redeposited sediments. The considerably different lithologies of interbedded redeposits appear to be related to the source, rather than to climatic changes at all scales on a small, as well as on a large scale. In the present section each marly interval represents a significant change in energy level. This relationship is not so obviously seen in deeper facies sections.

We do not know much about the complex geometry and the sequence-scale depositional patterns of the supposingly extremely low angled ramp system studied. This is mainly due to its considerable areal extent: the pericontinental „clastic wedge„ is estimated to have been 750 - 800 km long and 250 - 300 km wide (Michalík, 1993 a, b). The depositional area of the studied formations comprise approximately 1/3 of its extent. Significant tectonic displacements and relatively poor exposures are also problems.

Hips (1996 a, 1998 a) in her comprehensive studies of analogous formations in the Aggtelek-Rudabánya Mts. in NE Hungary suggested a complex model of the „Aggtelek„ ramp depositional system. However, the presumption that the geomorphologically determined study area would offer a complete lateral section throughout the ramp body from foreshore down to outer ramp and basin would probably be false. The vertical genetic subdivision of the Aggtelek ramp into depositional units is correlatable with relative successions from more northward (15 and 50 km) situated parts of the Silicicum megaunit (Fig. 6). Nevertheless, the position of the particular ramp (or ramps?) mentioned in the depositional system of the Lower Triassic Alpine-Carpathian shelf domain remains uncertain.

At present we do not have enough data on the ramp - slope - basin successions to be able to set up a spatial genetic subdivision of the ramp complex into depositional units. Thus, we are not even able to correlate certain sediments (shales/marls) in the studied section with specific stand of sea-level with any degree of certainty.

Two diametrically different paradigms may be chosen in order to solve this problem. In one opinion, the slope experiences a considerable increase of redeposition of eroded and remobilized inner shelf lime muds and calciclastics during the relative sea-level fall and lowstand (Hunt and Tucker, 1992, 1993; Lavoie, 1995), whereas sea transgression is the time when siliciclastic input may prevail on the drowned shelf. This concept is analogous to those used in siliciclastic systems and its application to carbonate systems may be false (Hunt and Tucker, 1992, 1993). From another point of view,Crevello and Schalger, 1980; Boardman et al., 1986; Hunt and Tucker, 1992, 1993), slope and basinal sedimentation may decrease during the lowstand inasmuch as the emerged shelf produces no or little carbonate sediment and may be affected by terrigenous siliciclastic input. Highstand is the time when the maximum growth potential of the platform (ramp) is realized and the shelf margins typically prograde basinwards (Hunt and Tucker, 1992, 1993).

As discussed later, we consider the shale/marlstone (?marlstone/shale) cycles in the lower part of the studied succession to represent parasequences (Vail et al., 1977). The problem mentioned above is where to put the parasequence boundaries.

A scheme marking the positions of the studied succession within the Szin ramp system is shown in figure 4.

Further speculation on possible sediment sources

Hips (1998 a) considered the origin of carbonate mud supplied to the ramp to remain enigmatic: it could presumably be winnowed and transported from the shallower ramp by storm-generated currents, or produced by the outer ramp (?).

Four possible relationships among contemporarily acting sources of fine siliciclastics and marls are illustrated in figure 5. The idea of carbonate-dominated shallow (e.g. warm-water, higher saline etc.) nearshore flats and shale-dominated offshore areas is hindered by about 50% of non-carbonate terrigenous shales (unmixed with the inner ramp marls) building up the ramp (a; see Fig. 2).

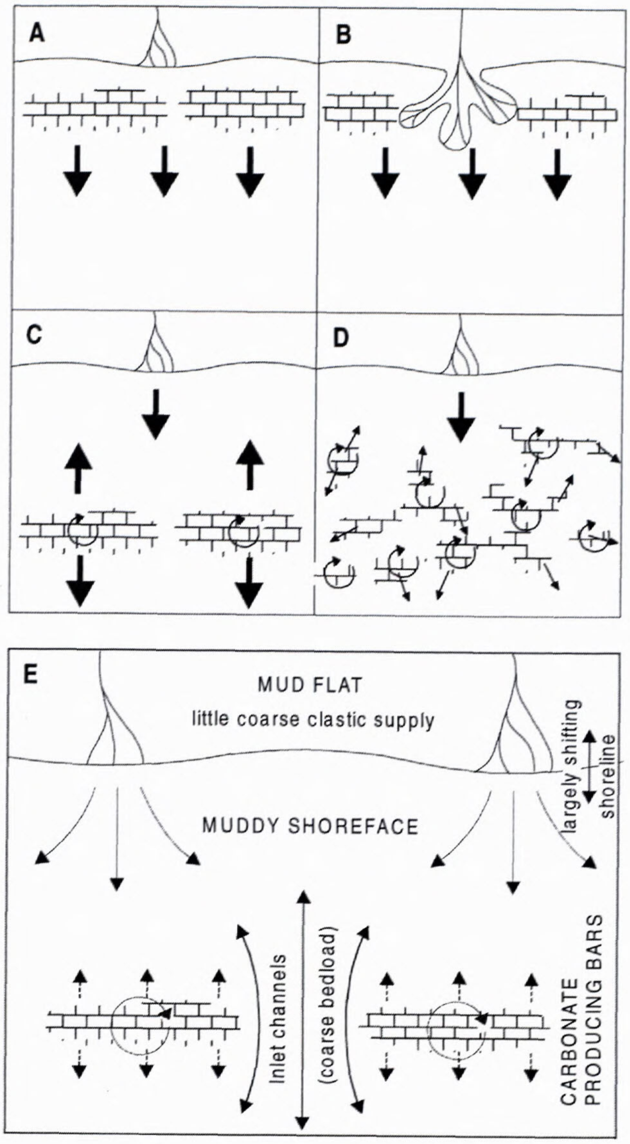

Fig. 5: Schematic view on four (A, B, C, D) possible relations between the sources of terrigenous clastics and carbonates. See E for explanation of symbols and text for discussion.

In the case (b), parallel, independent near-shore sources of both siliciclastics and carbonates, being further mixed in the more distal parts of the shelf are inferred. It would be hard to simply relate to sea-level changes the siliciclastics/carbonate ratio in a sediment supplied from such a system. Other controlling factors such as directions and rates of siliciclastic/carbonate progradation, climate, ecology, bioproduction and marine circulation should be considered to explain the described periodicity features. Aurell et al. (1995) described a similar case from the Upper Jurassic Deschambault carbonate ramp in Spain. Sand and microgravel-fed delta prograded over middle ramp but remained separate and did not induce the origin of a largely dispersed mixed carbonate-siliciclastic sequence. In figure 5 (c), the source of carbonate is located at a more distal pat of the inner ramp area.

Barrier bars may have originated in the high agitated fair weather-wave (FWW) break zone of the flat-topped ramp. The relatively high water circulation and moderated depth of the bars might have supported their subsequent colonization by carbonate-producing organisms. Thus, bioclastic shoals may have evolved in these zones. The carbonate mud and biodetritus could easily have been transported and redeposited by FWW and especially by storm weather-wave (SWW) activity, nearshore as well as in the offshore ramp areas.

An identifiable rimmed margin is lacking in the studied section and a non-rimmed ramp depositional environment is suggested (Hips, 1996 a, 1998 a). However, Fejdiová and Salaj (1994) described a thick-bedded cross laminated gastropod shells-rich oolitic grainstone accumulation („gastropodenoolite„, Mišík, 1966, 1977; Bystrický in Andrusov and Samuel et al., 1983) within an analogous section through Szin Marl Fm. to Gutenstein Fm. (Fig. 6). Accompanied hardgrounds suggest periods of arrested accumulation, probably because of an increase in water energy (Jones and Desrochers, 1992). Similar characteristics are typical of carbonate shoals accumulating above the FWWB in an agitated shallow-water setting (Wilson, 1986; Jones and Desrochers, 1992). Hips (1996 a, 1998 a) considered the ooids to have been formed in the subtidal, highly agitated surge zone on the inner ramp and piled up in fringing shoals, together with bivalve and gastropode shells. However, there were no in-situ shoals found throughout the Szin Marl Fm. The later author Hips further presumed that oolitic sand was washed out and redeposited by storms in the intershoal, shallow subtidal areas. The blanket appearence of the redeposited , amalgamated lobes of shoals reflect the rapid migration of the storm affected shoals.

An idea of independent bioclastic shoals developing in response to sea-floor topography is illustrated in figure 5 (d). Any negligible sea-floor elevation shallow enough to serve as a standing crop for carbonate producing organism could have evolved into a bioclastic shoal.

As in the case (c), breaking of FWW and thus a higher water circulation in certain zones related to sea-floor topography could have supported bioproduction on elevated domains. It also could have affected the further distribution of the produced carbonate. Naturally, storm-driven resedimentation would have played a significant role in such system.

In the „ramp-ooid-barrier complex„ facies model of Hips (1993 a, 1998 a; Read, 1985), moving ooid sand shoals were built up on the edge of the inner ramp, in the breaker-surf zone, as a result of strong, continuous wave agitation. Inner ramp and restriced lagoon were separated by shoals, from which washover fans spread out into inner ramp during major stroms. On the mid-ramp crinoidic, proximal storm sheets or hummocks bordered the outer flanks of the ooidic shoals. More distally, siliciclastic and lime sands formed flat storm sheets, with the grainsize and thickness decreasing towards the outer ramp. (The model is correlatable with the case (c) of the discussion above).

Discussion

The studied section is interpreted as an upward-shallowing succession of the uppermost part of a prograding,

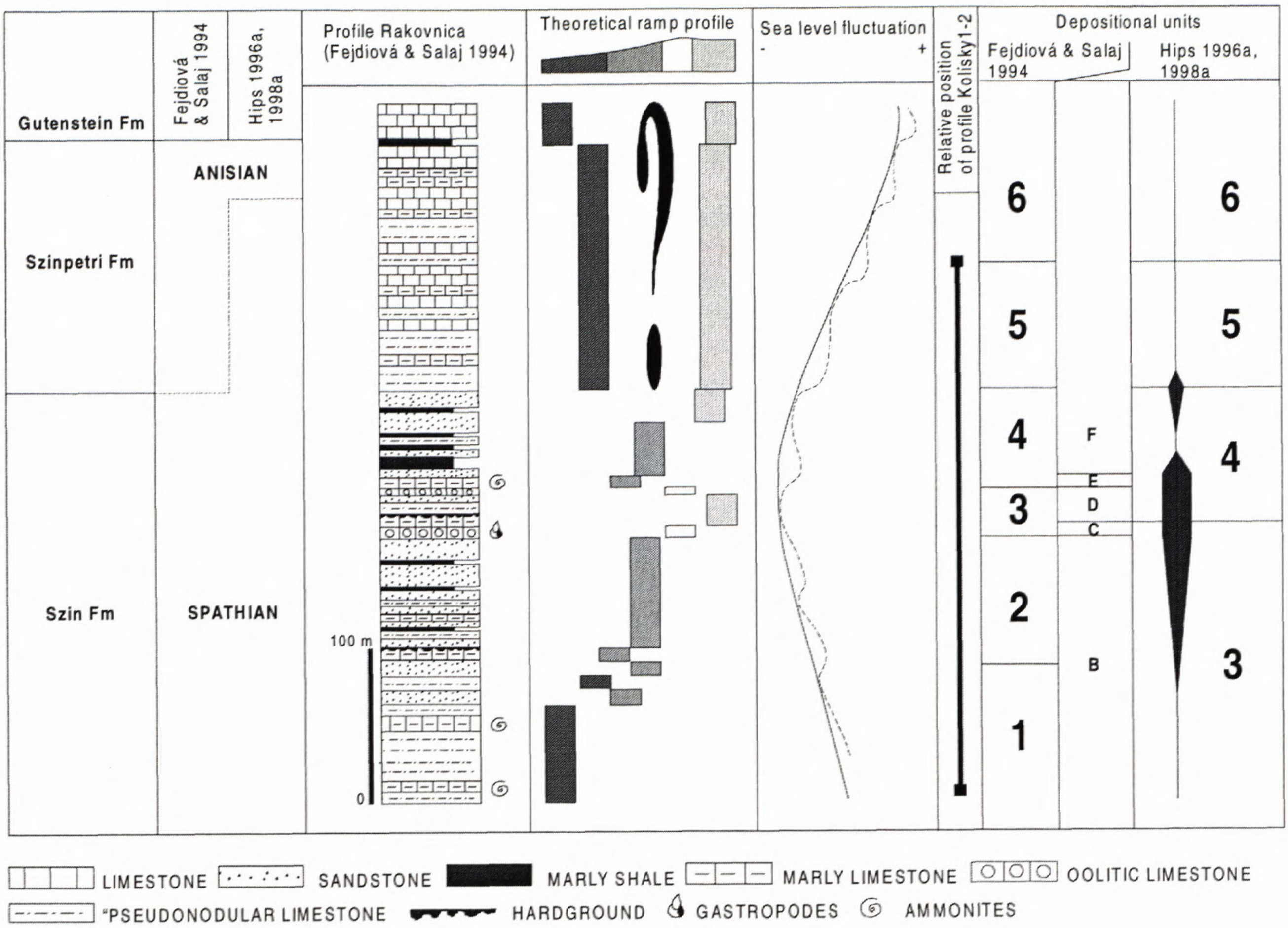

Fig. 6: Lithostratigraphic, paleoenvironmental and sequence stratigraphic interpretation of profile Rakovnica (Fejdiová & Salaj 1994) and its correlation with profiles Kolísky 1& 2 and with the synthesis on the Slicicum's Lower Triassic by Hips (1996a, b, 1998a, b). Note that the Szinpetri Limestone Fm. is much thinner (10–15 m) in the study area.

storm-dominated, gently-sloping mixed carbonate-siliciclastic ramp. The thin bedded, horizontally laminated shales/marlstones in the lower part of the present section represent a lower-middle slope facies. The storm-dominated middle part is linked to middle-upper slope facies and the scoured, erosion-resedimentation dominated, channelized, upper part of the succession is interpreted as a slope-crest-related and ramp-plain facies.

The last described part of the section is overlain by an about 30 m thick complex of well-bedded marlstone intercalated with coquina lags as much as 20 cm thick. This marlstone succession is conformably overlain by thin, well-bedded, partly bioturbated micritic limestone that is intercalated with a positively graded, horizontally laminated, biointrapelmicritic, limestone layers. These layeres commonly have often an erosive base and are clearly redeposits (Plate 2, a, b).

The well-preserved lamination and bioturbation features indicate a restricted, storm wave surges-protected, depositional environment. The redeposits represent the influence of an adjacent relatively high bioproductive area. Both the micrite and the biodetritus and other allochems have a completely different character from that of the subjacent rocks and show a close proximity to the overlying Gutenstein Fm.

The Szinpetri Limestone (Josvafő Limestone Mb.) is thus interpreted to have originated as an inner-ramp, restricted lagoon, transgressive sediment. Occurrence of foraminifera *Meandrospira deformata* (Salaj), indicates a hypersaline lagoonal environment (Salaj and Polák, 1994). Fejdiová and Salaj (1994) described a 4 m thick layer of bituminous shale in the uppermost part of the Szinpetri Limestone Fm. We interpret this as a maximum flooding surface that developed when the shelf experienced maximal drowning and the shelf lagoon became overdeepened (Hunt and Tucker, 1992, 199).

Considering the width of the Lower Triassic shelf, the biointrapelmicritic redeposits are not inferred to have been derived from either the sea- or landward shelf margin but the reverse; the existence of a local source seems probable. Topography-related bioclastic shoals are inferred ho have participated in supplying carbonate into the Szin ramp system. Locally, these could have been able to keep up with the rate of the sea-level rise and serve as a local sources from which the storm-induced redeposits were derived. Finally, these also could have served as the centers from which the superjacent Gutenstein carbonate ramp began to grow under the conditions of early highstand. The present secton Kolísky 2 may record the sedimentary fill of a locally restricted depression between the adjacent aggrading initial ramps. The inferred depression could occasionally have been influenced by storm induced sediments derived from the ramp fronts.

In contradiction to the previous interpretations of the Szinpetri Limestone as a sediment of a progressively restricted lagoon (Kovács et al., 1989; Fejdiová and Salaj, 1994; Hips, 1996 a, 1998 a), Hips (1998 a) reinterpreted it as dysaerobic to anaerobic outer ramp facies reflecting restriction due to rapid flooding of the distally steepened ramp. In this interpretation, the overying dark-grey to black, platy to massive micrites of the Anisian Gutenstein Fm. are supposed to represent the subsequent gradual filling of a restricted basin. However, the nature of the basin suggested in front of the Aggtelek ramp and the cause of its restriction remains ambiguous, due to the uncerain position of the studied ramp within the Lower Triassic Alpine-Carpathian shelf depositional system.

In the Gl'ac Unit the Szinpetri Limestone Fm. clearly overlies inner ramp deposits at the top of the upward shallowing ramp-slope succession. In both settings they are interpreted as transgressive deposits reflecting rapid flooding of the antecedent ramp. The overlying Gutenstein Fm. thus represents deposits of the subsequent highstand.

A turn in the ramp geometry and development of aggrading/?retrograding margin in response to the relative sea level rise was inferred in order to explain the contemporaneous restriction and limey mud production and its supply into the overdeepened lagoon. Although no such aggraded margin was found among the studied successions, it should be kept in mind that the ramp system constributed to a much more extensive shelf domain (Roniewicz, 1966; Brandner, 1984; Michalík, 1993a, b; Hips, 1996 a, b, 1998 a, b and references therein; Kovács and Hips, 1998).

Although relatively thin (10 – 15 m) and represented exclusively by the Jósvafő Limestone Fm. in the study area, the Szinpetri Limestone Fm. thickens southwards to about 100-150 m in the Slovak Karst and the Aggtelek-Rudabánya Mts. (Fejdiová and Salaj, 1994; Kovács et al., 1989; Hips, 1996 a, 1998 a). This may reflect the inner-shelf paleoposition of the Gl'ac Unit relative to more southward located successions, or it may support the spatial facies distribution suggested by Hips (1998 a). The latter case inferrs the deposition of dysaerobic to anaerobic Szinpetri Limestone Fm. in both drowned inner ramp lagoon and outer ramp restricted basin.

In general, we infer that relatively deeper water, upwards-shallowing Szin Marl Fm. is capped by the transgressional basal sequence of the Middle Triassic carbonate ramp complex (Szinpetri Limestone Fm. underlying the Gutenstein Fm.).

From all the evidence described we proposed a scheme of major relative sea-level movements (Fig. 6). As the Upper Permian-Lower Triassic is widely believed to have been a tectonically calm periods at the passive margin to Western Tethys, we inferr the curve to reflect eustacy to a high extent. The studied succession from the Gl'ac Unit is a possible analog of the units HST/3 to 6 of Hips (1996 a, 1998 a) and HST/Sc.5 to TST/An.1 of De Zanche et al. (1993) (Fig. 6).

The overall lack of fossils does not allow for exact dating either of the lithostratigraphic units of the sea-level fluctuation related events. Earlier it was accepted as a compromise that the base of the Gutenstein Fm. coincided with the Scythian/Anisian boundary. But it seems to be more realistic that the lithostratigraphic boundaries are not coeval and are generally time transgressive southwards (Bystrický 1985, Hips 1989). We involve the Hips´ (1996 a, 1998 a) litho- and biostratigraphic division of the Lower Triassic of Silicicum and use the sequence stratigraphic analysis for approximate inter-regional correlation.

Conclusions

The Lower Triassic (Spathian) Szin Marl Fm. of the Silicicum megaunit of Western Carpathians is interpreted as a low-angled, storm-dominated, mixed siliciclastic–carbonate ramp. Its upwards shallowing slope succession is subdivided on the basis of progressively increased energy of the depositional environment. The intervals of different energy levels are separated by structurally monotonous marly horizons. Several meters-thick layers were found in a certain level of the upper ramp slope succession–in between hummocky cross stratified deposits and the scouring/redeposition dominated deposits in the upper part of the present succession–in which the sediments were deposited in compensation cycles.

It is inferred that the fine grain size of the silty shales and especially the lithological heterogeneity controlled the origin of well developed, architecturally variable storm redeposits within the mixed shaley/marly intervals.

Bioclastic shoals growing up in response to an inherited sea-floor topography are believed to have taken part in the carbonate-factory action, by means of both bioproduction and modification of the ramp-top relief. The carbonate material was further redeposited due to wave and especially the storm wave activity.

From the sedimentologic and lithofacies evidence we develop a scheme of major relative sea-level movements. One major transgressive event was assumed, separating the Szin Marl Fm. from the overlying Middle Triassic carbonate complex. The Szinpetri Limestone Fm. was interpreted to represent transgressive deposits on the base of the Middle Triassic carbonate complex.

The studied succession of the Gl'ac Unit is possibly analogous to the units HST/3 to 6 of Hips (1996 a, 1998 a) and HST/Sc.5 to TST/An.1 of De Zanche et al. (1993).

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Fresh-water limestones of the Hlavina Bed in the Rišňov furrow and Bánovce Depression

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Abstract: Carbonate deposits (lacustrine carbonate, fresh-water limestone, travertine) crop out in the Rišňov Furrow and Bánovce Depression. They were studied at localities Malé Kršteňany, Klížské Hradište and Sádok. Based on the study of gastropods *Aegopinella orbicularis* (Klein), *Fortuna clairi* (Schlickum-Strauch), *Klikia* cf. *goniostoma* (Sandberger) and *Tropidomphalus* cf. *doderleini* (Brusina) showing their Late Panonian age they were assigned to the Hlavina Bed. The content of trace elements (Cu, Zn, Sr, Mn, K, Na, Ti , V a B) showed their fresh-water origin. Based on the carbon isotope analyses it is possible to state that a part of the fresh-water limestones is enriched in light ^{12}C isotope suggesting their organic origin. This is supported by the occurrence of fresh-water algae of *Rivularia* genus. The limestones, containing heavy carbon isotope, originated by precipitation of hydrothermal solutions enriched in $CaCO_3$. This is shown by increased Fe content in the rocks. Oxygen isotope content suggested the limestone origin in an environment 5 – 10° C warm. Occurrence of numerous onkoids in fresh-water limestone suggests a shallow-water, dynamic depositional environment.

Key words: West Carpathians, Danube Basin, Panonian, fresh-water limestone, stratigraphy, lithogeochemistry, genesis, C isotope, O isotope

Introduction

Fresh-water deposits of the Hlavina Bed, stratigraphically assigned to the Late Panonian (zone H), prevailingly consist of carbonates (lacustrine limestone, fresh-water limestone, travertine, Fordinál & Nagy 1997). They occur nearby marginal faults in the Tribeč and Považský Inovec Mts. According to the regional-geologic division (Vass et al. 1988) they occur in the Rišňov Furrow and Bánovce Depression.

In the mentioned area fresh-water limestones were studied at localities Malé Kršteňany (Bánovce Depression), Klížské Hradište and Sádok (Rišňov Furrow, Fig. 1).

We studied structure of fresh-water limestones, oxygen and carbon isotope content, trace element contents (Cu, Zn, Sr, Mn, K, Na, Ti, V, B) and we performed ma-

Fig. 1 Situation of studied localites

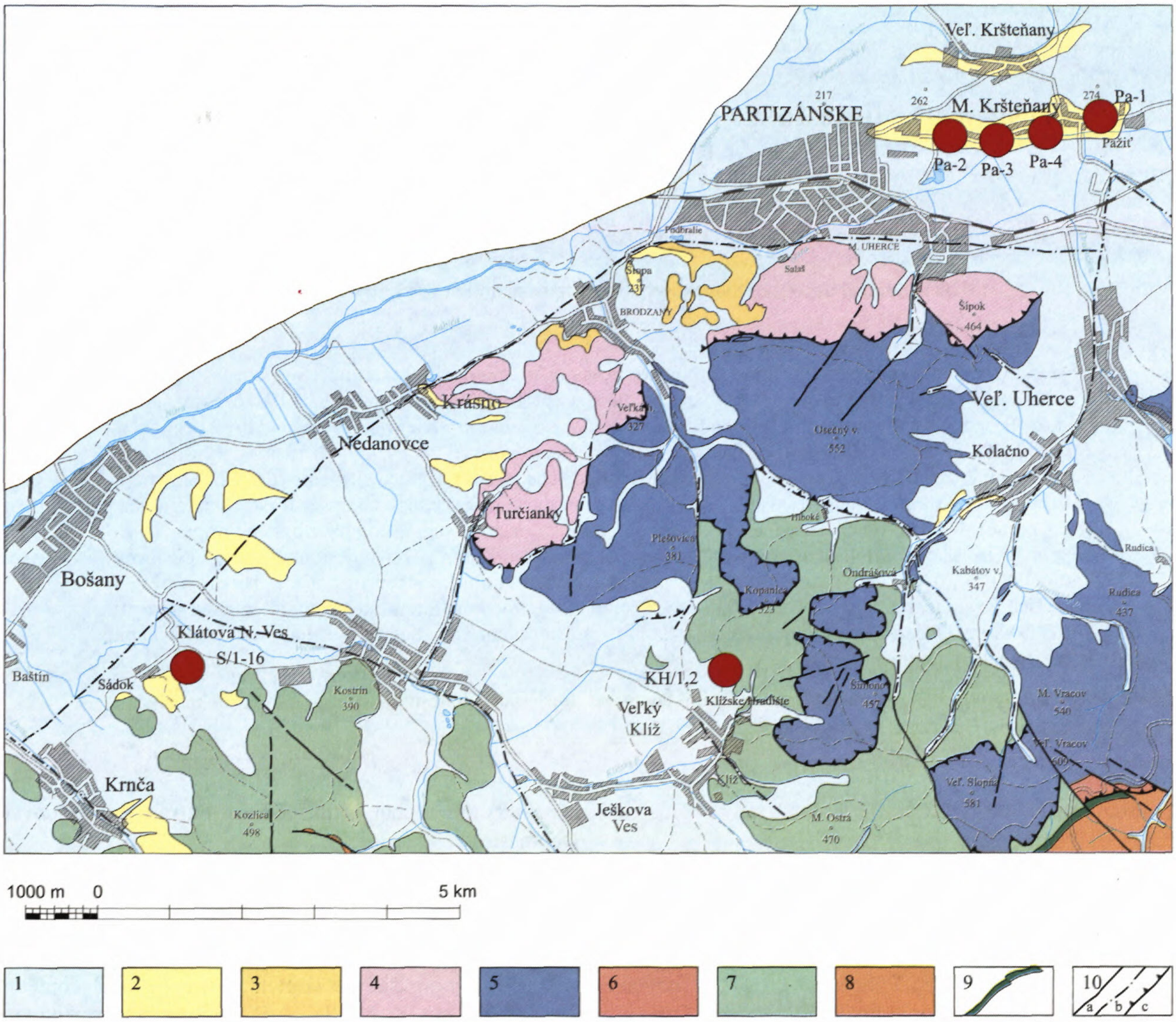


Fig. 2:Schematic of geologic map and location of studied localities
1 – Quatenary, 2 – Tertiary, 3 – Paleogene, 4 – Hronicum, 5 – Veporicum, 6 - Crystalline rocks, 7 – Tatricum (envelop unit), 8 – Crystalline rocks of Tatricum, 9 – Metamorfic rocks, 10 - Faults: a) faults assumed, c) trust lines.
The map modified based on the geological map of Tríbeč 1 : 50 000, Ivanička et al., 1998

nometric analysis. The oxygen and carbon isotope content of limestones was done in the isotope laboratory at the Geologic Survey of Slovak Republic in Mlynská dolina in Bratislava by RND. Ivan Repčok. The results were interpreted by I. Töröková (Törökova 1998).

Gastropode fauna from limestones occurring at locality Malé Kršteňany (Tab. 1) was studied by K. Fordinál.

Overview of previous research

The oldest information on fresh-water limestone occurrence in the Bánovce Depression and Rišňov Furrow originate from the second half of the 19th century.

Stache (1865) refers an occurrence of Neogene deposits consisting of clay and fresh-water limestone from the Bánovce Depression. The limestone contains gastropods of genus *Helix, Bulimus, Planorbis* and *Lymnaeus.* He found it at localities Partizánske – Šimovany (Ssimnowany), Malé Bielice (Male Bilice) and in the area between villages Malé Kršteňany (Male Kŕstenany) and Veľké Kršteňany (Welke Kŕstenany). At the end of 70-ties the region of Bánovce Depression was geologicaly completely processed (Brestenská et al. 1980). The fresh-water limestone described from surroundings of Pravotice, Bielice, Malé and Veľké Kršteňany were assigned to the Dakian.

Occurrence of fresh-water limestone in the Rišňov Furrow was known from the surroundings of villages Kovarce (Kowarz), Sádok, Bošany (Bossany), Brodzany (Brogyán), Krásno (Szeplak) and Nedanovce (Nedanocz) (Winkler 1865). At the break of century Schafarzik (1900) refers an occurrence of the deposits from Sádok (Szádok) and from the area between the villages Kovarce (Kovarcz) and Čeľadince (Családka). He assigns them in the Pliocene.

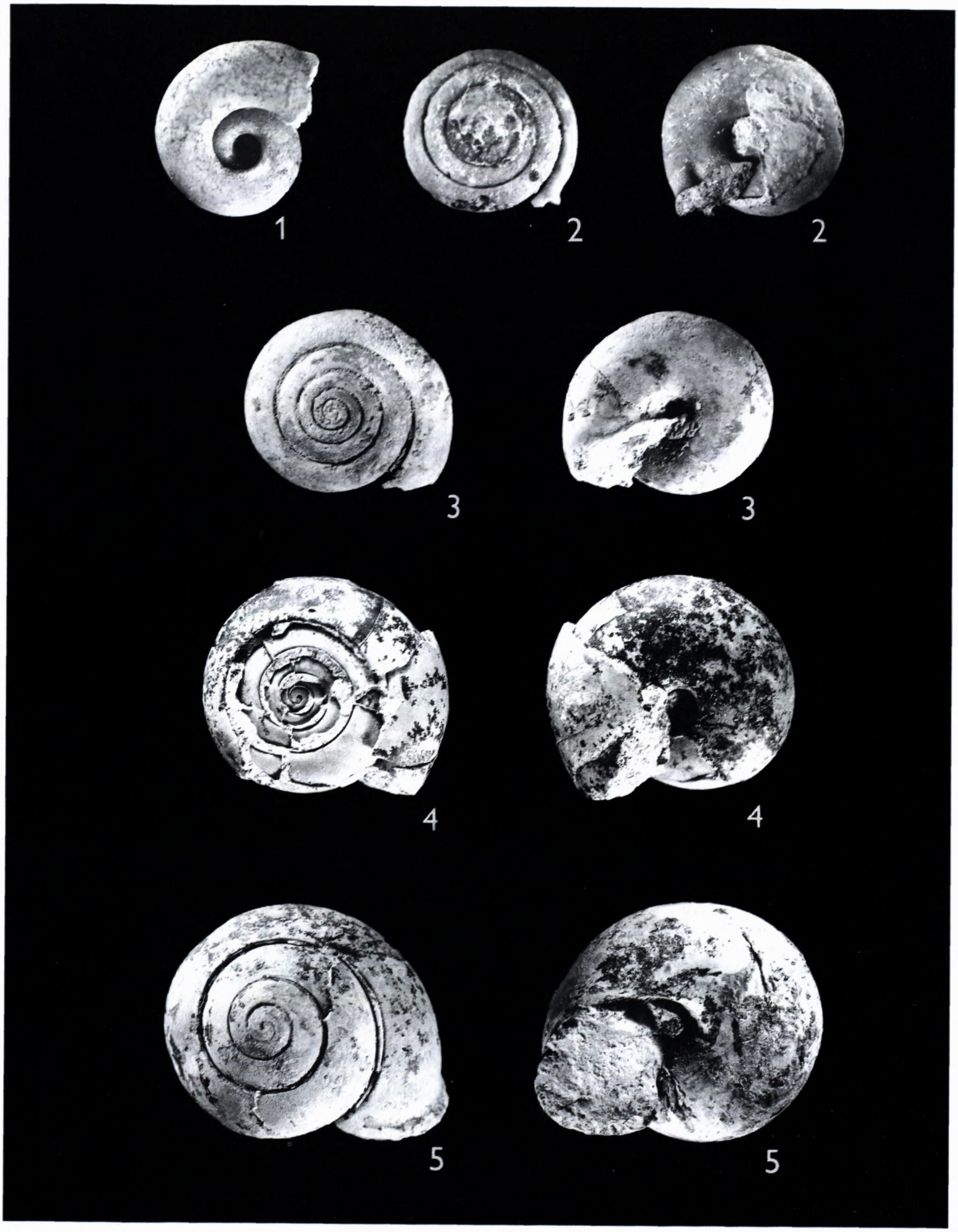


Table 1

1. *Planorbis* sp., Malé Kršteňany, Pa-3, enlarged 3,8x; 2. *Isognostoma* sp., Malé Kršteňany, Pa-2, enlarged 1,7x; 3. *Leucochroopsis kleini* (KLEIN) Malé Kršteňany, Pa-1, zv. 3x; 4. *Klikia* cf. *goniostoma* (SANDBERGER), Malé Kršteňany, Pa-2; enlarged 2,8x; 5. *Cepaea* cf. *etelkae* (HALAVÁTS), Malé Kršteňany, Pa-2, enlarged 2,7 x
Photo C. Michalíková

The western margin of the furrow (nearby Považský Inovec) was mapped by Brestenská (1962). She found remnants of fresh-water and terrestrial molluscs in the fresh-water limestones and pelitic sediments in the area. She assigned the sediments in the Pontian (Brestenská l.c.). The recent knowledge on the fresh-water limestone occurrence and molusk fauna from the area were mainly obtained by field research in the last years (Fordinál 1994, Fordinál in Maglay et al. 1997).

The eastern margin of the Rišňov Furrow (neighbouring the Tribeč Mts.) was mapped by Brestenská and Priechodská (1969). They assigned the studied deposits in the Pontian. The latest knowledge from the area are summarized in the Explanation to geologic map of Tribeč 1:50 000 (Ivanička et al. 1998).

Characteristic of localities

Malé Kršteňany

In the Malé Kršteňany village and in the quarry localized NE of the village Hlavina Bed deposits consisting of fresh-water limestone crop out. A layer containing blocks of fresh-water limestones, floating in the unconsolidated calcareous matrix of lacustrine limestone type, occurs at the base of outcrops in the village (from Pa-2 to Pa-4, Fig. 2).

The limestones have micritic and biomicritic structure. Micrite consists of fine-grained calcite and it comprises the main part of limestone matrix. The sparite cement, consisting of coarse-grained crystalline calcite, fills numerous pores, gaps and veins. Numerous sections of gastropod fragments filled by crystalline calcite occur in the limestone structure (Figs. 3 and 4). Clastic grains are represented by quartz and occasional plagioclas. Dark spots of clay admixture occur in the micrite. Fe coatings occur in pores and fissures of limestones. They also occur in the form of sphere bodies.

Malé Kršteňany – quarry
(Pa – 1, Fig. 2)

Solid fresh-water limestone occur in the upper right section of the quary. The limestone locally pass to loose rocks resembling lacustrine limestones. Cores of terrestrial and fresh-water gastropods were found at the section. The following species were identified:

Terrestrial gastropods: *Leucochroopsis kleini* (Klein), *Tropidomphalus (Mesodontopsis)* cf. *doderleini* (Brusina)
Fresh-water gastropods: *Aplexa* cf. *subhypnorum* Gottsch, *Anisus sp., Viviparus sp.*

Malé Kršteňany – village

Fresh-water limestones crop out on slope in the village at three places (Pa-2 to Pa-4). The limestone contains gastropods fauna in the core form.

Pa-2

At this locality (Fig. 2) cores of terrestrial gastropods occur in the fresh-water limestones. The following gastropods were identified: *Aegopinela orbicularis* (Klein), *Fortuna clairi* Schlickum-Strauch, *Klikia* cf. *goniostoma* (Sandberger), *Cepaea* cf. *etelkae* (Halaváts), *Cepaea sp.*, *„Helix" richarzi* Schlosser, *Isognostoma sp., Claussilidae indet.* Monospecific assemblage consisting of tests of *Abida* species was found in a fragment (Figs. 5 and 6).

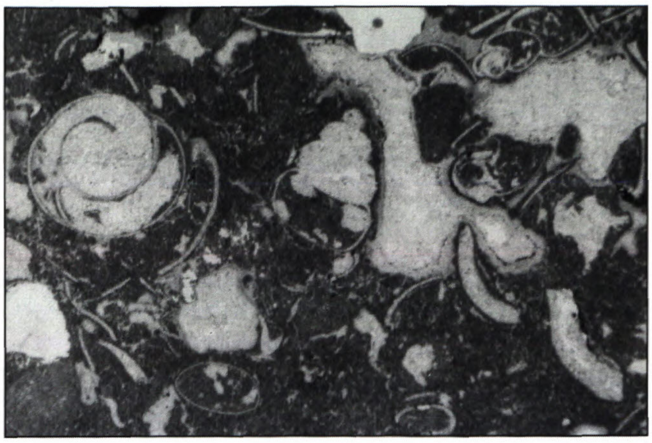


Fig. 3: Cross sections of gastropode and lamellibranchiata shells (locality Malé Kršteňany)

Fig. 4: Cross section of gastropode shell filled by crystalline calcite (locality Malé Kršteňany)

Pa – 3
(Figs. 7, Tab. 2)

Similarly to the previous locality cores of terrestrial and fresh-water limestones were found at the locality (Fig. 2). The following species were recognized:
terrestrial gastropods: *Aegopinella orbicularis* (Klein), *Leucochroopsis kleini* (Klein), *„Helix" richarzi* Schlosser, *Succinea* sp., *Cepaea* sp.

fresh-water gastropods: *Planorbis sp.*

Pa – 4
(Fig. 8, Tab. 3)

In the fresh-water limestones occurring at the locality (Fig. 2) cores of terrestrial gastropods of the *Cepea, Helicigona* and *Klikia* genus were found.

Fig. 5: Fresh-water limestone containing rich fauna of gastropodes (Abida sp.), locality Malé kršteňany Pa2, natural size, photo C. Michalíková

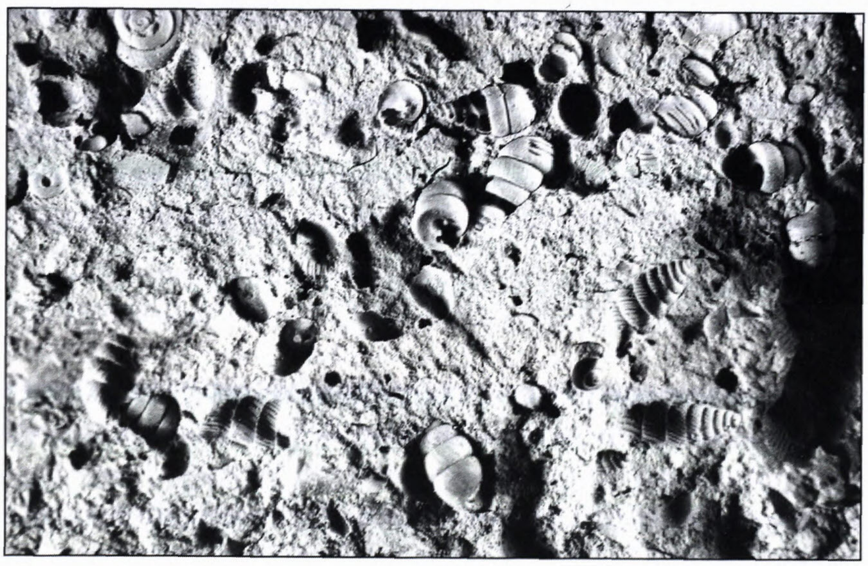


Fig. 6: Close-up of the Fig. 3, enlarged. 2x; photo C. Michalíková

Kližské Hradište

The studied locality represented by a quary wall occurs some 2 km NW of village Klížské Hradište (Fig. 9). In the lower part of the outcrop light-brown, baige, solid and compact fresh-water limestone with occasional porous layers occur (Fig. 10). Toward the overlying, about 20 cm thick bed, porous layers increase and the upper part is composed of weathered travertines containing red-coloured karst loams. The loams fill a few karst holes (Ivanička et al. 1998).

The limestone from the locality prevailingly has micrite structure which locally passes in to sparite structure. Numerous pores, holes and veins are filled by crystalline calcite. Onkoids, spheres of irregular form having concentric rims and central part filled by calcit, occur in the structure of the limestones (Fig. 11). Partly irregular form of micritic rim of onkoids suggests increased wave activity. According to classification of Logan et al. (1964) they may be assigned to the structural type SS – C, concentricaly growing spheroids. The clastic grains of quartz also have accretionary microcrystalline calcite consisting of thin layers. They have form of pizoids (Fig. 12).

Only fresh-water algae *Rivularia* cf. *haermatites* Shaffer & Stapf and traces after their activity represent fossil remnants. We found them in the form of clumps consisting of thin tubes filled by crystalline calcite (Fig. 13). They also form pillow forms (Fig. 14).

The mentioned algae are assigned to series *Cyanophytae* (blue-green alga) and family *Rivulariaceae*. The family is known from the pre-Cambrium.

The representants of the genus *Rivularia* prefer shallow lacustrine and fluvial environment with fresh-water. They also tolerate brackish water.

Sádok

SE of the village Sádok travertine pile occurs (Fig. 15, Tab. 4). The lower part of the pile consists of yellowish-brown coloured solid and compact layers of travertines having occasional up to 0.6 m thick interlayers of porous travertines. This passes into porous travertine with observable accretionary layers. The upper part is composed of variegated clay having 5 – 10 cm thick bed of loose sharp-edged quartz clasts without matrix at the base. Other clasts are composed of crystalline rocks. They are 1 – 2 mm in diameter. They probably represent deposits of a rapid wash of already sorted sediment from coast. The clay contains ostracods *Candona (Typhlocypris) roaixensis* Carbonell and *Candona sp.* (Fordinál in Ivanička et al. 1998).

The carbonate structure consists of micrite passing into sparite. Numerous pores are filled by crystalline calcite. In the unfilled pores calcite forms druse crystals. Occasionally pores are rimed by limonite pigment. Some pores are rimed by coarse-grained calcite. They form geopetal structures in which the lower part is composed of microsparite having a gradual transition to sparite.

Similarly to Klížske Hradište also onkoids, algae *Rivularia* cf. *haematites* Schaffer & Stapf and fragments of juvenile gastropode tests occur. Also quartz and plagioklas grains occur. A part of grains is of authigenic origin.

Fig. 7: Outcrop Malé Kršteňany Pa3

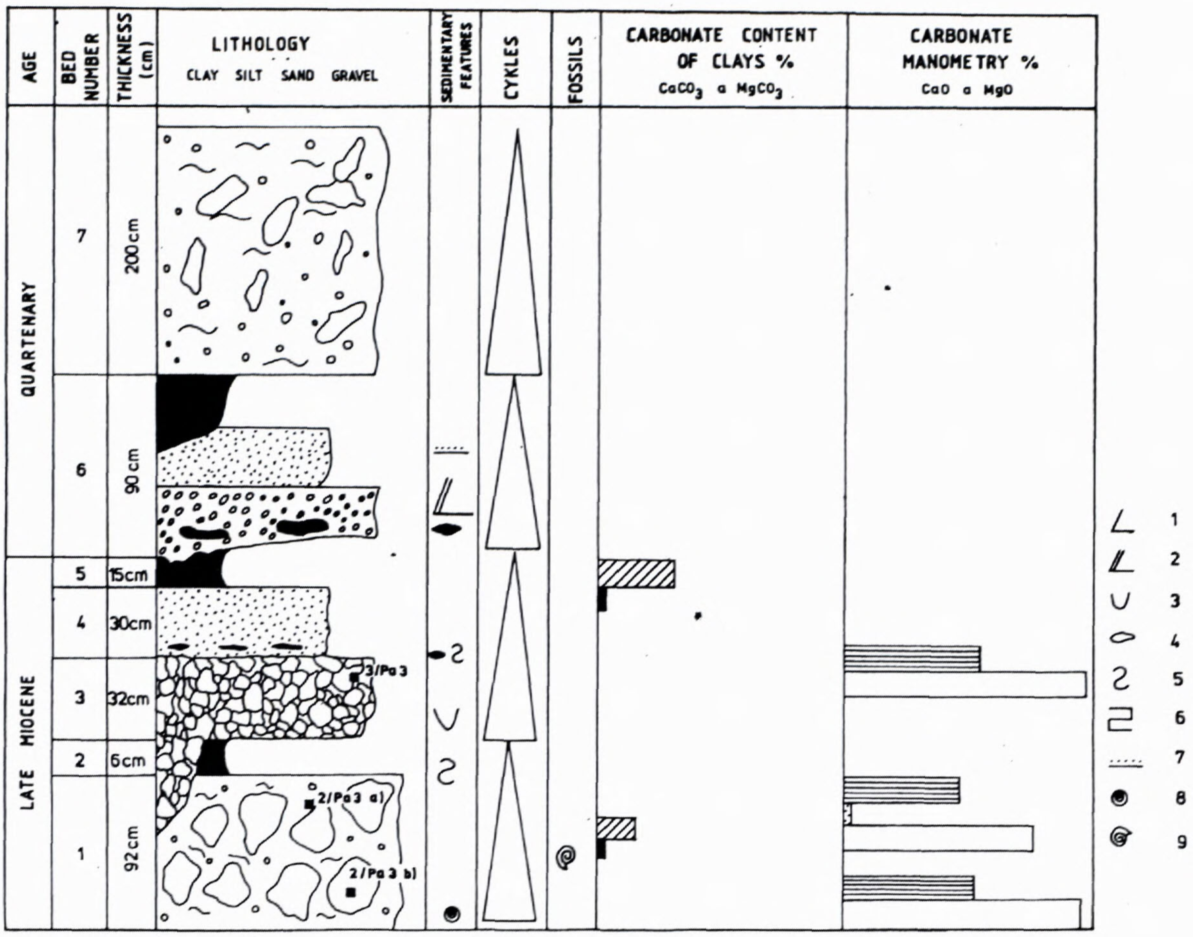
Fig. 8: Outcrop Malé Kršteňany Pa4

Biostratigraphic assignement

Only at locality Malé Kršteňany (Pa – 1 to Pa – 3) remnants of fresh-water and terrestrial gastropods were found which could by identified.

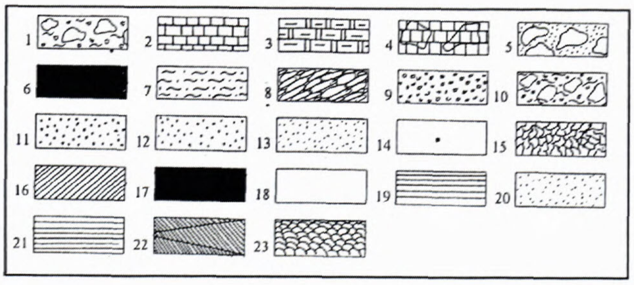
Gastropods with wider and narrower extent occurred in gastropod assemblages from the locality. For the biostratigraphic assignement of fresh-water limestone the occurrence of gastropods *Aegopinella orbicularis* (Klein), *Leucochroopsis kleini* (Klein), *Fortuna clairi* (Schlickum-Strauch), *Klikia* cf. *goniostoma* (Sandberger) and *Tropidomphalus (Mesodontopsis)* cf. *doderleini* (Brusina) are important.

Species *Aegopinella orbicularis* (Klein), *Leucochroopsis kleini* (Klein) generally suggest the Panonian age (sensu Rögl et al. 1993). The age is constrained by the occurrence of the species *Fortuna clairi* (Schlickum & Strauch), which stratigraphic range is the Late Panonian (zones G-H) and Pliocene (Lauger 1981). Based on the stratigraphic range of the species it is possible to state



Tab. 2: Profile of Pa3 outcrop

Sedimentary features and Fossils to the Tab. 2, 3,4: 1 – Cross stratification – normal, 2 – Cross stratification – large scale, 3 – Erosive channel, 4 – Intraclasts, rip-up clasts, 5 – Erosive surface, scor, 6 – Horizontal bedding, 7 – Gradded bedding - normal, 8 – Oolit, pisoid, 9 – Gastropode



Legend to the Tabs. 2, 3 and 4

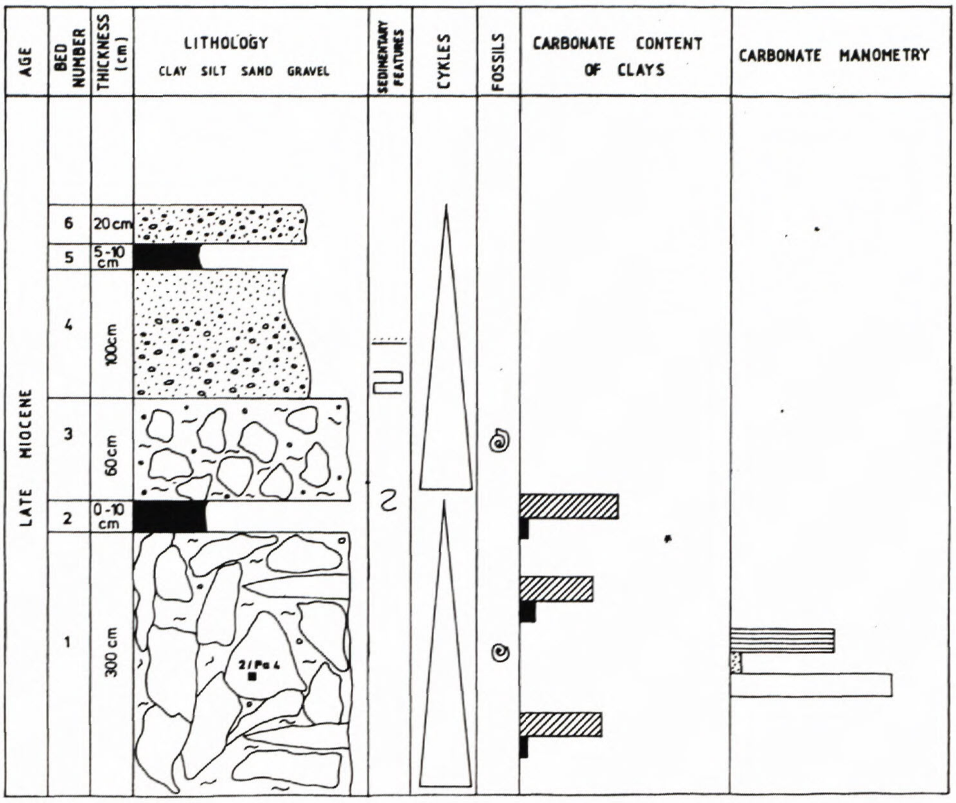
1 – oolitic limestone, 2 – massive carbonate beds, 3 – porous carbonate, travertine, 4, 5 – fine-grained carbonate with calcareous matrix, 6 – grayish clay, 7 – lake marl, 8 – conglomerate, 9 – fine-grained conglomerate, 10 – gravelly loam containing occasional pebbles, 11 – coarse-grained sand, 12 – medium-grained sand, 13 – fine-grained sand, 14 – massive sand without lamination, 15 – calcareous crust, duricrust, 16 – $CaCO_3$ content, 17 – $MgCO_3$ content, 18 – calcite content, 19 – CaO content, 20 – MgO, 21 – horizontal bedding, 22 – cross bedding, 23 – ripple-cross lamination

that the fresh-water limestones are of the Late Panonian age and they represent Hlavina Bed stratigraphically assigned to zone H (Fordinál & Nagy 1997). The stratigraphic assignement is also confirmed by the occurrence of gastropods *Klikia* cf. *goniostoma* (Sandberger) *and Tropidomphalus (Mesodonntopsis)* cf. *doderleini* (Brusina) which were up to now only found in the mentioned zone in the Danube Basin (Fordinál 1996).

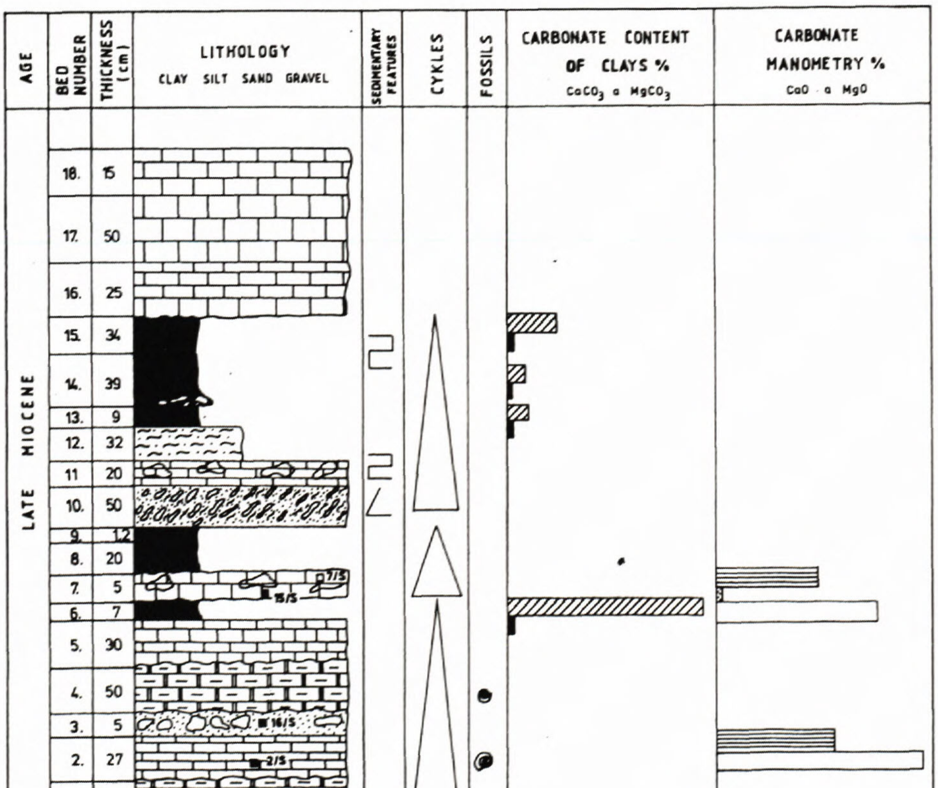
Lithogeochemistry of carbonates

Manometric analysis (Tab. 5) of studied limestones from the above mentioned localities showed that they contain 70 –98% of pure calcite. The part of them besides calcite also contains Fe dolomite (21% - 23%). The insoluble rest, represented by clay minerals, clastic quartz and feldspat grains, limonite and authigenic quartz, varies in volume from 0.30% to 6.25%. The non-carbonate part contains an essential amount of insoluble rest which does not release carbonates. Its content in limestone varies from 1.69 to 7.29%.

The $CaCO_3$ value varies from 45.59 to 54.06% while the content of molar calcium is relatively high (81.4% to 98%). The MgO content is low (1.11 to 1.20%). The molar amount of magnesium is also very low (2.7 – 3%). The FeO value is relatively high considering fresh-water



Tab. 3: Profile of Pa4 outcrop



Tab. 4: Profile of the Sádok outcrop

Fig. 9: Quarry Klížske Hradište

Fig. 10: A layer of porous limestone, locality Klížske Hradište

Table 5: Manometric analysis of carbonates

Sample No.	Locality	Calcite %	Fe - dolomite%	Nz %	Np %	CaO %	MgO %	FeO %	Co_2 %	Sr/Ca %	Mg/Ca %
1/KH	Klížske Hradište	70, 03	23, 53	5, 18	6, 43	45, 59	1, 20	6, 00	40, 77		0, 036
2/KH	Klížske Hradište	75, 26	22, 35	3, 33	2, 38	48, 20	1, 14	5, 70	42, 57	5, 345	0, 032
2/Pa3 a	M. Kršteňany	92, 71	0	3, 43	7, 29	51, 95	0	0	40, 77	60, 53	/
2/Pa3 b	M. Kršteňany	76, 56	21, 75	2, 15	1, 69	48, 77	1, 11	5, 54	42, 89	/	0, 031
3/Pa3	M. Kršteňany	97, 94	0	1, 60	2, 06	54, 88	0	0	43, 06	35, 81	/
2/Pa4	M. Kršteňany	71, 34	21, 74	6, 25	6, 91	45, 84	1, 11	5, 54	40, 59	66, 09	0, 032
2/S	Sádok	96, 48	0	1, 80	3, 52	54, 06	0	0	42, 42	45, 72	/
7/S	Sádok	74, 96	21, 74	0, 30	3, 29	47, 87	1, 11	5, 54	42, 18	30, 47	0, 031

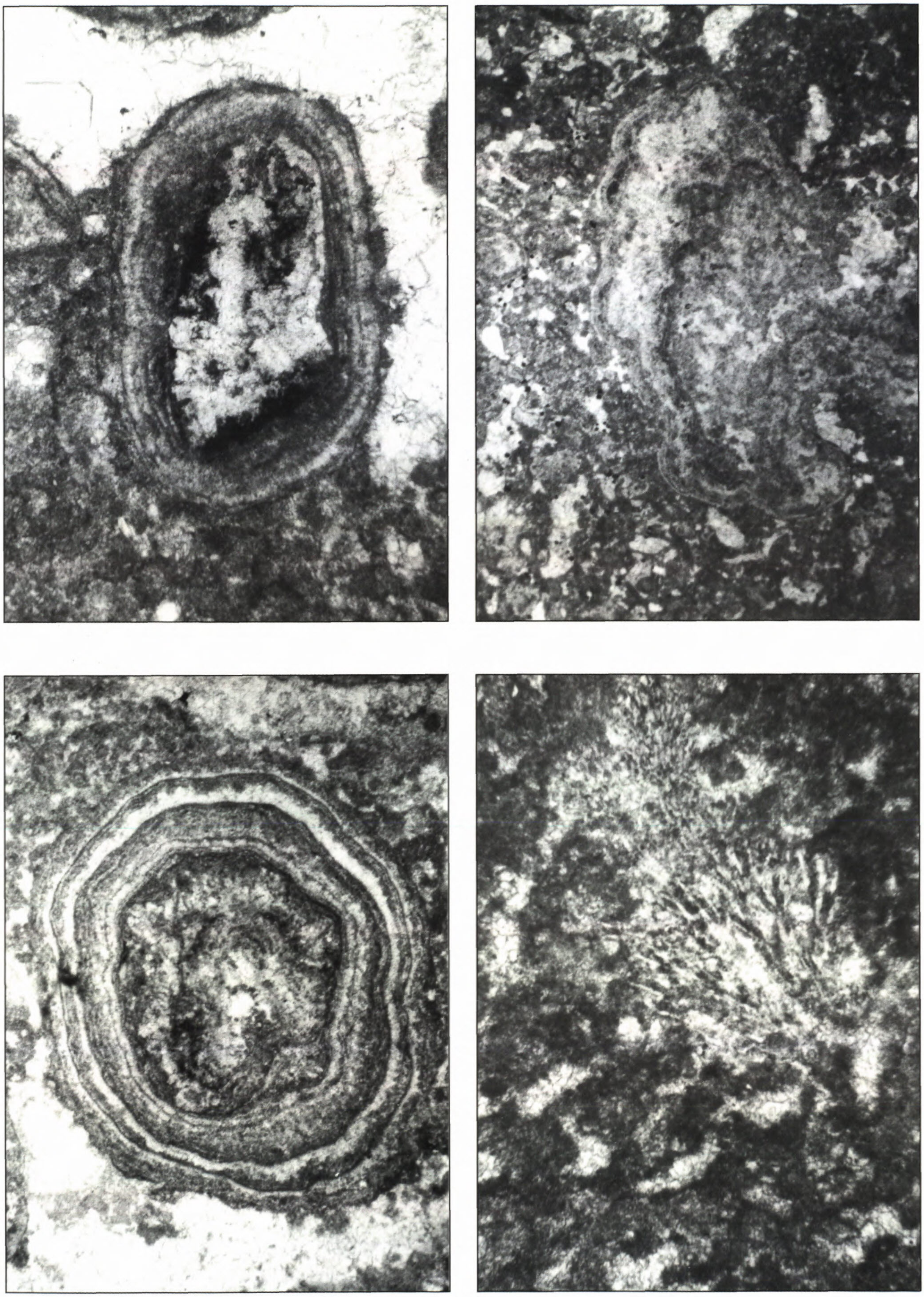
limestones (5.54% - 6.0%). FeO was probably brought by warm springs circulating on the bottom of sedimentary basin. The carbon dioxide content varies from 40.59 to 43.06%.

The mutual rate Sr/Ca shows high values up to 66.09. The reason may be high content of carboniferous part because Sr has tendency to bind to high-carboniferous components. The lowest value was observed in limestones from Klížske Hradište (5.34).

The rate Mg/Ca varies from 0.001 to 0.036 and the contents of individual samples does not show big differences. Mean value for marine carbonates is considerably higher (0.80 - 0.95) suggesting fresh-water origin of carbonates.

A layer of clayey sediment having white colour and high carbonate value with $CaCO_3$ up to 92.34% and $MgCO_3$ 0.92%. The sediment is identified as lacustrine limestone.

Slovak Geol. Mag., 5, 3(1999), 213–226



Also content of trace elements Cu, Zn, Sr, Mn, K, Na, Ti, V, B was analysed in limestones from studied localities (Tab. 6).

Cu – copper is biophillous element and it binds to rock-building minerals. It is a part of biologic processes. The mean values of copper in limestones are around 4 ppm (Rosler & Lange 1972). The copper content in the studied limestones varies from 1 to 6 ppm.

Zn – zincum is mainly bound with sheet silicates, oxids and Fe hydroxides. The mean value in fresh-water limestones varies from 16 to 19 ppm, the value for marine limestones varies from 18.6 to 31 pm (Wedepohl 1969). In the studied limestones the zincum value varies from 7 to 17 ppm suggesting their fresh-water origin.

Sr – stroncium is an element with afinity to high-carbonate components and it is diadochic with Ca and K elements. In the studied limestones the stroncium contents vary considerably. The lowest value occurred in the limestone from the Klížske Hradište (46 ppm). The highest value was obtained from the limestones from Malé Kršteňany (563 pm). The values in the rest of limestones varies from 259 to 542 ppm. Rosler and Lange (1972) refer the mean value for fresh-water limestone around 610 ppm. Wedepohl (1969) refers stroncium values for marine limestones form 452 to 765 ppm. The recorded values of Sr confirm fresh-water environment during the limestone origin and higher amount of Sr occurring in the samples from Malé Kršteňany may be resulted by high Ca content to which Sr has ability to bind.

Mn – manganium mainly concentrates in marine deposits what is suggested by high Mn contents more than 385 ppm in various areas (Rankama & Sahama 1952). In the studied limestones the Mn content is very low (5–76 ppm) suggesting their fresh-water origin.

Fig. 15: The outcrop Sádok

Tab. 6: Trace elements and their contents in fresh-water carbonates

No.	Sample label	Locality	(ppm) Cu	(ppm) Zn	(ppm) Sr	(ppm) Mn	ppm) K	ppm) Na	(ppm) Ti	(ppm) V	(ppm) B	(%) H_20
1.	2/Pa3	M. Kršteňany	6	12	563	7	80	20	16	<5	5	0, 24
2.	3/Pa3	M. Kršteňany	3	8	351	9	30	10	11	<5	3	0, 13
3.	2/Pa4	M. Kršteňany	5	17	542	8	230	20	34	6	12	0, 37
4.	1/S	Sádok	2	7	487	7	30	20	6	<5	<3	0, 24
5.	2/S	Sádok	1	8	439	5	30	20	4	<5	<3	0, 23
6.	7/S	Sádok	2	11	259	10	40	30	7	<5	<3	0, 27
7.	2/KH	Klížske Hradište	6	7	46	76	120	10	27	28	9	0, 32

12 14

11 13

Fig. 11: Onkoid (locality Klížske Hradište)

Fig. 12: Clastic quartz grain with concentric layers of calcite – pisoid (locality Klížske Hradište)

Fig. 13: Patches of fresh-water algae of genus Rivularia cf. haematites (locality Klížske Hradište and Sádok)

Fig. 14: Traces after activities of fresh-water algae (Klížske Hradište)

K, Na, Ti – content of the elements is generally very low. The values for kalium are 30–230 ppm (marine limestones 2700 ppm), natrium 10–30 ppm (marine limestones 370 ppm) and titanium 4–34 ppm. Concentration of the elements in marine sediments is essentially higher and the low values confirm a fresh-water origin of the studied limestones.

V – vanadium is diadochic with Fe^{3+} and it binds to organic matter and phyllosilicates. The vanadium values in the studied fresh-water limestones varies from 5 – 28 ppm. Rosler & Lange (1972) refer values obtained from Carboniferous and Mesozoic carbonates 2.5 to 10 ppm. The higher value of vanadium (28 ppm) was obtained from a sample from Klížske Hradište. It may be a result of admixture of organic matter in limestones to which vanadium is bound.

B – borum is an important element for the palaeoenvironment reconstruction and for the palaeosalinity indication. In the fresh-water environment the contents varies from 15 to 45 ppm and in marine environments it varies from 20–55 ppm (Rosler & Lange 1972). The borum value in the analyzed limestones is low (3–12 ppm) suggesting their fresh-water environment.

Content of trace elements in the studied limestones are considerably lower than referred by some authors for marine limestones. The low concentrations of individual elements in the limestones suggest their fresh-water origin.

Oxygen and carbon isotopes in limestones

Isotopic composition of oxygen (^{18}O) and carbon (^{13}C) was analyzed from four carbonate samples. Two samples were from the locality Sádok and two samples were from the locality Klížske Hradište. In the studied are limestones were also analyzed at localities Veľké Kostoľany, Bojnice, Sádok, Krásno, Záhrada nearby Veľké Tesáre, Veľký Kríž and Podhorany (Töröková 1988, Tab. 7).

Tab. 7: Values of oxygene and carbon isotopes in carbonates

Locality	Age	$^{18}O_{PDB}$	$^{13}C_{PDB}$
V. Kostolany	Quaternary	- 1, 387	2, 012
Bojnice	Riss - Würm	0, 747	- 0, 437
Sádok	Pannonian	2, 004	- 3, 285
Krasno	Pannonian	2, 838	- 3, 375
Záhrada pri V. Tesároch	Pontian - Quaternary	2, 866	- 8, 476
V. Klíž	Pannonian	3, 085	- 9, 126
Podhorany	Pannonian	3, 321	- 10, 478
Podhorany	Pannonian	3, 181	- 10, 251
Sádok 1 5	Pannonian	2, 542	- 7, 891
Sádok 1 6	Pannonian	2, 012	1, 117
Kližské Hradište	Pannonian	3, 092	- 9, 061
Kližské Hradište	Pannonian	3, 252	- 9, 165

Isotopic content of oxygen ^{18}O (Tab. 7, Fig. 16) from limestones sampled in Klížske Hradište has values δ^{18} O_{PDB} from 3.092 to +3.52 per mile. It suggest that the limestones are enriched in heavy oxygene isotope (^{18}O) and during their formation only small temperaturer fluctuation not exceeding 5°occurred. The limestones probably originated in cooler water having temperature 5–10°C.

Isotope content of ^{13}C carbon in the studied limestone, which values is δ^{13} C_{PDB} from – 9,061 to –9.165 per mile, suggests enrichment in light carbon isotope ^{12}C which is of organic origin. It showes more intensive input of organic carbon also confirmed by macrofauna occurrence and abundant occurrence of algae identified by microscopic study of limestones.

The limestones from the locality Veľký Klíž, Záhrada, Podhorany have very homogenous oxygen and carbon isotope composition (Tab. 4) and they probably originated in very similar environment like limestones from Klížske Hradište. It is possible to consider them as coeval (Pannonian) and they probably had similar origin. (Fig. 16)

Oxygen isotope composition at the locality Sádok (δ^{18} O_{PDB} from 2.012 to +2.542 per mile) shows that the limestones are enriched in heavy oxygen isotope (^{18}O) and they also originated in cooler waters having temperature 5 to 10°similarly to limestones in Klížske Hradište.

Carbon ^{13}C and its isotope composition in the limestones at the studied locality is different (δ^{13} C_{PDB} form – 7.891 to + 1.117 per mile). The values show big differences indicating possible change of the carbon source. The value (δ^{13} C_{PDB} + 1.117 per mile obtained from sample Sádok 16 shows that the limestone is enriched in heavy carbon isotope ^{13}C thus originally it is of organic origin. It is probable that it originates from warm springs. It is consisting with increased content of FeO in the limestones. The negative carbon value (δ^{13} C_{PDB} – 7.891 per mile obtained from sample S/15 (Sádok) shows the enrichment of the limestone in light carbon isotope ^{12} having organic origin. The organic carbon originates from found algae and fragments of gastropod tests.

The special distribution of oxygen and carbon isotopes in the carbonates have the samples of the Quaternary age from Veľké Kostoľany and Bojnice. The measured values of oxygen varies from (δ^{18} O_{PDB} form 0. 777 to –1.387 per mile showing their origin in warmer waters with temperature 15 to 20° C becuase they are enriched in heavy carbon isotope ^{13}C having anorganic origin. This type of carbon isotope prevailingly accumulates in warm waters (Hudson 1977). The values of carbon isotopes ^{13}C varies in the limestones (δ^{13} C_{PDB} from – 0.437 to +2.012 per mile suggesting their enrichment in heavy carbon isotope ^{13}C which is of anorganic origin. It may originate from warm hydrotherms similarly to the locality Sádok (sample no. S/16) which have content of heavy carbon isotope.

Kantor and Mišík (1992) studied travertines at locality Dubná Skala. They assigned them to the Pliocene. The recent research showed that they have originated since the Late Miocene (Hók et al. 1998). Based on carbon isotope composition they found (Kantor & Mišík 1992) that the origin of the travertines was influenced by input of anorganic carbon from warm springs and that the Quaternary

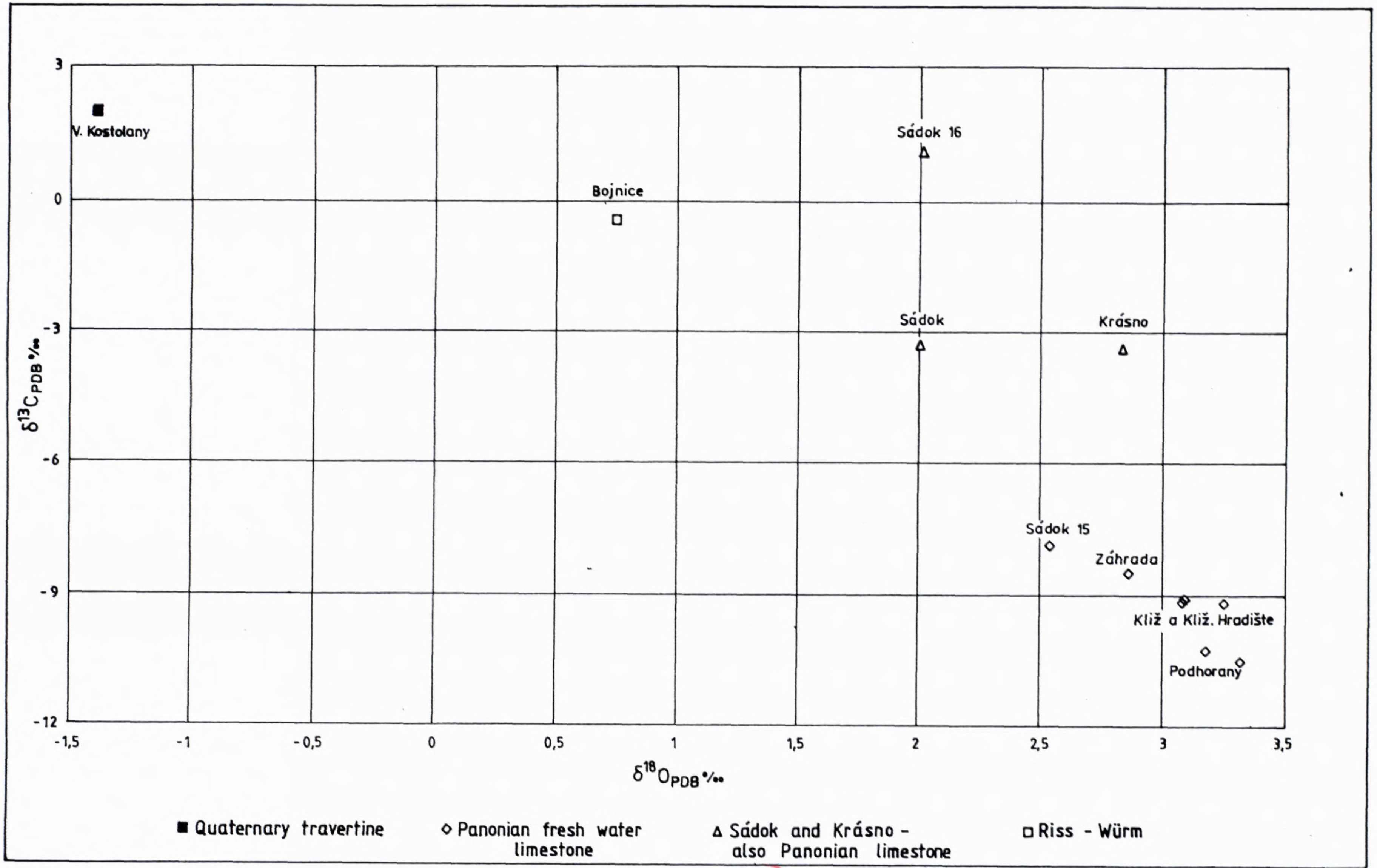


Fig. 16: Isotopic composition of oxygene and carbon in fresh-water limestones

Slovak Geol. Mag., 5, 3(1999), 213–226

travertines from localities Veľké Kostoľany and Bojnice and a part of the Late Pliocene travertines from locality Sádok where higher enrichment in heavy carbon isotope ^{13}C originated by similar way.

Conclusion

Studied fresh-water limestones and travertines from localities Malé Kršteňany (Pa-3, Pa-4), Sádok and Klížske Hradište originated by different ways.

Thick basal layers of boulder carbonates from Malé Kršteňany probably originated by their breaking and sliding. The occurrence of gastropode fauna in both limestone clasts and matrix proves synsedimentary origin of the boulder carbonates. Broken, chaotically arranged carbonate boulders may suggest a change of sedimentary conditions. The change might cause movement of carbonate beds and their subsequent breaking.

Beds of fresh-water limestones from localities Sádok an Klížske Hradište are prevailingly of organodetritic origin. However, a part of limestones has anorganic origin.

The organic origin of limestones is confirmed by occurrence of fresh-water algae of *Rivularia* genus and debris of gastropode shells. Layers of travertine limestone (Sádok and Klížske Hradište), having anorganic origin, occur between organogenic carbonates. They originated by precipitation of mineral springs and by hydrotherm effects at the bottom of sedimentary basin. Some percentage of Fe and limonite aggregates suggest their enrichment in Fe. Results of oxygene and carbon isotope analyses showed that except the limestones from Veľké Kostoľany and Bojnice, all the limestones probably originated in waters 5°C - 10°C warm. The first ones originated in 15°C - 20°C warm water. The water temperature could be influenced by warm mineral springs.

The depositional environment of limestones was influenced by increased wave activity. It is suggested by numerous bodies – onkoids with concentric structure. The cooler waters and climate is also evidenced by clay minerals in clays alternating with limestone layers. From clay minerals smectite prevails above illite and kaolinite. The content of trace elements in limestones proves their fresh-water origin.

Acknowledgement

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Slovak Geol. Mag., 5, 3(1999), 227–232

Numeric age of the Sarmatian boundaries (Seuss 1866)

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Abstract: The revision of the numeric age of the chronostratigraphic Sarmatian boundaries (Seuss 1866) according to the proposal of Kokay et al. (1991), Oszczypko (1997, 1998) and to the implications from the work of Gaždzicka (1994) is premature. The magnetostratigraphic interpretation of the Sarmatian/Pannonian boundary, which served to Kokay et al. (1991) as a base for the lowering of the boundary to 12.3 Ma, is not unambiguous. The chosen borehole in the part, where the contact between the Sarmatian and Pannonian is recorded, is not applicable for that purpose. There are not unambiguous reasons to get younger the numeric age of the Sarmatian/Badenian boundary from 13.6 Ma to 13.0 Ma or to 11.8 Ma. The biostratigraphic data yielded from the Polish Carpathian Foredeep are not convincing. The younger age is not consistent with numerous radiometric ages of biostratigraphicaly well dated volcanic rocks and it also is not supported by magnetostratigraphy.

Key words: Parathetys, Sarmatian, radiometric age, magnetostratigraphy, chronostratigraphy, biostratigraphy

Introduction

The Sarmatian is chronostratigraphic stage of the Paratethys Miocene. Its time range is different in the Central and Eastern Paratethys. It is because it was defined by curious way and by two authors – Suess in the Cetral Paratethys (in the Viena Basin, the stratotype profile was in the Viena city part Hernals, the lectostratotype is in Nexing NE of Viena) and Barbot de Marny in the eastern Paratethys. Both definitions were published by Suess (1866). Today it is widely known that the Sarmatian according to Barbot de Marny has larger time range and it also includes the sediments originated later than the youngest Sarmatian sediments sensu Seuss. Biofacial evolution of the Sarmatian in the central and eastern Paratethys provided its more detail division, and, what is more worthy, it also provided mutual correlation. Generally the biostratigraphic correlation is accepted according to which the Sarmatian sensu Suess correlates with the Volhynian and Early Bessarabian i.e. with the older part of the Sarmatian sensu Barbot de Marny (Steininger et al., 1985). This stratigraphic correlation was also confirmed by radiometric ages of volcanic rocks interfingering the Sarmatian deposits in both parts of the Paratethys (Vass et al. 1987, Chumakov et al. 1992 and somewhere else).

Based on the radiometric dating of neovolcanics, the Sarmatian sensu Suess (1866) was numerically calibrated as follows: the base 13.6+-0.2 and the top 11.5 +-0.5 Ma (Vass et al. 1987). According to the calibration the Sarmatian stage lasted around 2 Ma. In the last years several correlation schemes of the Paratethys Miocene were published where the authors presented the Sarmatian as a stage with essentially shorter duration. Some of them shift its top downward and the others rise its base. It results in an absurd situation when Kokay et al. (1991) suggested numeric age 12.3 Ma for the top et Sarmatian and Oszczypko (1998) calibrates the Sarmatian base to some 11.8 Ma (Fig. 1).

Critical analysis of the reasons for the change of numeric calibration of the Pannonian/Sarmatian boundary

The tendency of decreasing the numeric age of the Sarmatian or the Pannonian/Sarmatian boundary commences with the cited work of Kokay et al. (1991). The result was, besides something else, redefinition of the numeric age of the Pannonian/Sarmatian boundary and its shifting from 11.5+-0.5 Ma to 12.3 Ma. The redefinition found a response in other Hungarian authors (i.e. Hámor 1995 in Czászár ed. 1997).

Kokay et al. (1991) analysed lithostratigraphy, biostratigraphy and magnetostratigraphy of sediments from the borehole Berhida 3 (Bh-3). The borehole was drilled in the southern part of the Várpalota brown coal Basin on the northern margin of the Bakony Mts., south of the town Várpalota. In the borehole the Pannonian/Sarmatian boundary was determined immediately beneath the 5 cm thick layer of dacite redeposited tuff having the radiometric age 12.6+-0.5 Ma, but the authors suggested slightly younger numeric age of the boundary. This only one radiometric age became a reference datum for numeric dating of the Pannonian/Sarmatian boundary for several Hungarian authors. In fact, the numeric date is also supported by local bio- and lithostratigraphy as well as by correlation of magnetic measurements of the borehole core with magnetostratigraphic scale (Berggren et al. 1985).

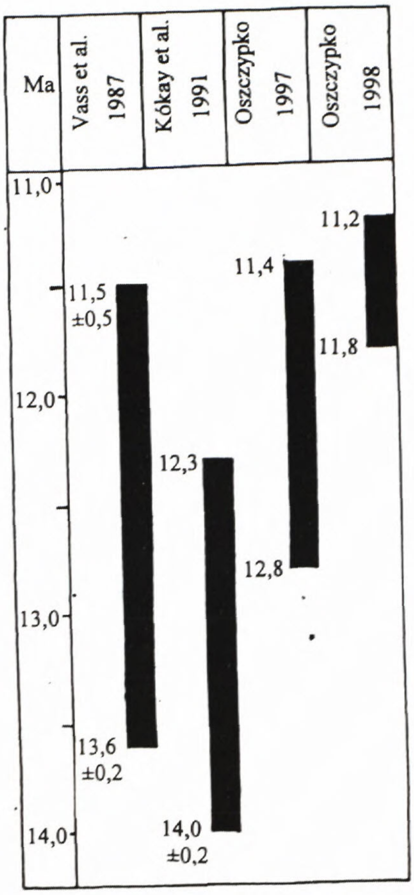

Fig. 1: Numeric age of Sarmatian stage according to Vass et al. (1987), Kokay et al. (1991) and Oszczypko (1997, 1998). The tendency to shortening of the Sarmatian time interval is evident.

It is possible to rise several substantial objections against the redefinition of the numeric age of the Pannonian/Sarmatian boundary based on the basis of the data obtained from the borehole Bh-3.

1) The thickness of the Sarmatian and Pannonian deposits in the borehole Bh-3 is very small. The Sarmatian, represented by three formations (Gyulafirátot, Kazárd and Tinnye), is thick only 138.4 m and it only represents a fragment of total thickness of the Sarmatian Stretava and Ptrukša Formations in the East-Slovakian Basin (3240 m; Král et al. 1990, Rudinec 1978, Vass et al. in lit.). In the Viena Básin, where the Sarmatian was described by Suess, the maximum thickness of its deposits is 1 000 m (Jiříček & Seifert 1990). Also in the Danube Basin is the thickness of the Sarmatian Vráble Formation two- to threefolds greater than in the borehole Bh-3. For example, in the borehole Ivánka-1 is the Sarmatian thick about 420 m and some authors assume that the borehole only penetrated the Early Sarmatian (Čermák 1972, fide Biela 1978).

Similarly the Pannonian, representing in the Hungarian chronostratigraphic terminology the Early Pannonian or Peremárton Group, is only thick some 70 m (69.4 m), while the Pannonian in other parts of the Hungary attains the thickness several thousands of meters (e.g. in the Zala Basin about 1680 m, W of Tisza river 2850 m, in the Makó Basin 2940 m; Nagymarosi 1981).

From the above mentioned it is possible to imply that the Sarmatian and Pannonian deposits in the borehole Bh-3 are very condensed and most likely they represent uncontinuous profiles shortened by disconformities or by faults.

2) Particularly the lithologic profile of the formation from Ös, comprising the lower part of the Pannonian in the borehole Bh-3, suggests an interpretation of the formation as chronologicaly uncontinuous sedimentary succession, interrupted by periods without sedimentation. The formation from Ös originated in a shallow lagune. The interruption of the deposition resulting from repeating dessication of the lagune is proved by alternation of grayish-green and yellowish-red pelites (oxidation Fe^{2+} to Fe^{3+}), dessication cracks and anhydrite crystals (Kókay et al. 1991).

3) Radiometricaly dated dacite tuff was also by Kokay et al. (l.c.) assigned to the Pannonian because they assume it to be equivalent of the dacite tuff lying in other profiles of the SW foothill of the Bakony Mts. on the Pannonian deposits as stated by Jámbor (1988, fide Kokay et al.). The correlation may be, but does not have to be, right. There is many manifestations of acid explosive volcanism in the form of rhyolite, rhyodacite or dacite tuffs and intercalations of glassy volcanic ash in the Sarmatian of the central and also of the eastern Paratethys. The redeposited tuff from the borehole Bh-3, which radiometric age is 12.0+-0.5 Ma may be equivalent of some Sarmatian tuff.

4) The mollusc assemblage occurring in the immediately underlying sediments of the redeposited tuff, which is interpreted belonging to the Sarmatian – Pannonian transition, consists of small, closely ribed, poorly preserved Cardium, small forms of *Modiolus incrassatus*; foraminifera assemblage is composed of numerous specimen of the species *Rotalia beccarii*. The assemblage does not need to indicate the Sarmatian – Pannonian transition as it is assumed by Kokay et al. (l.c.). It may represent an assemblage of shoreface, lower salinity, shallow-water facies of the Sarmatian.

5) Correlation of magnetic measurements with magnetostratigraphic scale may have more variants. One of them is correlation of two normal events measured in the borehole Bh-3 in the depth around 180 – 190 m with the upper part of chron C5r (events C5r 1n and C5r 2n; Berggren et al. 1995). The chron C5An, to which the events of normal polarization were correlated by Lantos in Kokay et al. (1991), is probably absent (Fig. 2).

6) For the proposal of modification or change of radiometric time scale it is unconditionaly necessary to substantiate the correction by more data, the best yielded by more laboratories. It is impervious to base this corrections only on one dating which questions the time scale based on numerous matching data.

Critical evaluation of new opinion on numerical age of the Sarmatian/Badenian boundary

As it already was indicated, several authors get younger the numerical age of the Sarmatian/Badenian boundary. I hold the following critical attitude to their opinions:

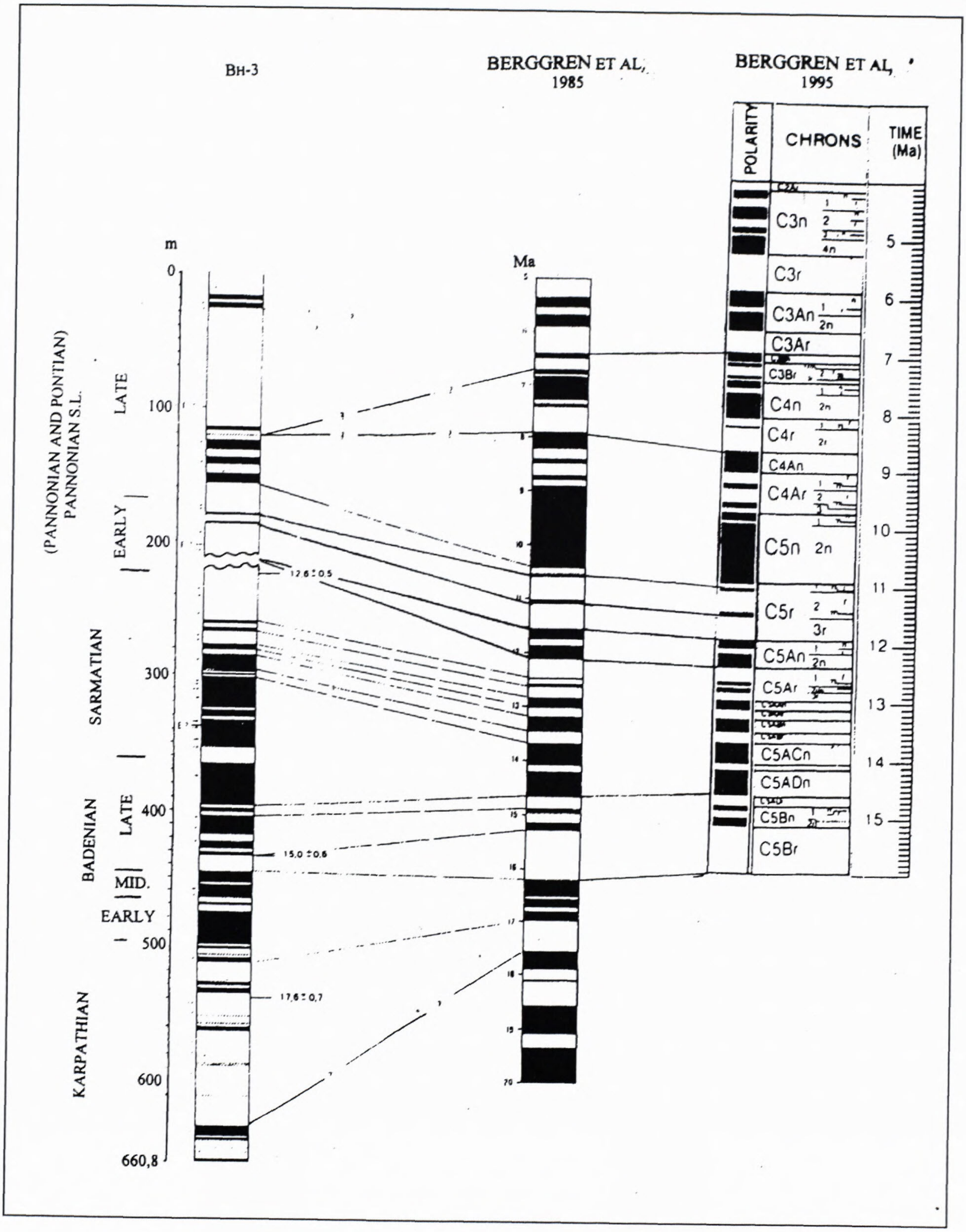

Fig. 2: Scheme showing the tentative correlation of the author of this paper. Two normal events found in the borehole Berhida – 3 (Bh-3) in the depth 180-190 m are corelated with two normal events in the upper part of the chron C5r (events C5r1n and C5r2n). Chron C5An with two normal events in the borehole Bh-3 is missing because of the hiatus.

1) Gaždzicka (1994) found in the Machov Formation – the Polish Foredeep- assemblage of calcareous nanoflora with *Discoaster calcaris* in sediments overlying evaporites (equivalent of the Wieliczka Formation). The first occurrence of this assemblage is in the zone NN 8. *Discoaster kugleri*, the leading form of the zone NN 7, does not occur in the assemblage. Based on this she deduced that the deposits above the evaporites are younger

as zone NN 7 and they may be assigned to the zone NN 8 and partly also to the zone NN 9. According to Gaždzicka the NN8/NN7 zone boundary occurs either above the evaporites (as she depicted in the Fig. 4) or within it, or beneath the evaporites (in the Figs. 2 and 3). The numeric age of the NN 8/NN 7 is represented on the time scale of Berggren et al. (1995) in two variants: 11.5 Ma or 10.8 Ma. The Sarmatian base in the Polish Foredeep, which occurs within the Machov Formation between brackish Krakowiec Bed (the Sarmatian) and marine Pecten Bed (the Late Badenian) should be according to Graždzicka younger than 11.5 Ma or 10.8 Ma. It seems, that Oszczypko (1997) considered the younger age of the Sarmatian base as not possible and therefore he also assigned evaporitic formation (Wieliczka) to the Sarmatian and he defined the Sarmatian base on the level around 12.8 Ma. A drawback of this correlation is fact, that the deposits of the halite associated with the evaporitic formation, should originate in saline lagunes of low-salinity sea providing they are of the Sarmatian age. The low-salinity environment is indicated by fauna occurring in the Sarmatian deposits everywhere in the Paratethys, suggesting brackish or even lower-salinity environment. The origin of halite deposits in brackish sea is unprobable and maybe from that reason Oszczypko (1998) later modified his stratigraphic correlation. He correlated the evaporitic formation from Wieliczka and Pecten marine beds with the Late Badenian and numeric age of the Sarmatian/Badenian boundary shifted higher on the level around 11.8 Ma. He defined the Pannonian/Sarmatian boundary on the level 11.2 Ma. Other authors in the last works i.e. Rogl (1996, 1998) Rogl, Krhovský & Hamršmíd (1997) define the Sarmatian base on the level 13 Ma.

2) Numeric age of the Sarmatian 11.2 – 11.8 Ma i.e. 0.6 Ma as suggested by Oszczypko (1998) is not realistic considering the thickness of the Sarmatian deposits, for example in the East-Slovakian Basin. The entire thickness of the Sarmatian is 3240 m and the thickness of Early Sarmatian deposits (zone of large Elphidia or zone with *Elphidium reginum*) is 2410 + 1/3 after decompaction. Totally it comprises 3 213 m. If the Early Sarmatian represents the half of the time of the entire Sarmatian sensu Oszczypko, i.e. 0.3 Ma, then the sedimentation rate in the East-Slovakian Basin at that time should be 10 710 m Ma^{-1}. Even the higher sedimentation rate should be in the East-Slovakian Basin if we accepted conclusions of Gaždzicka (1994), suggesting that the the Early Sarmatian is only a part of the zone NN 8. The maximum mean sedimentation rate in the epicontinental seas is only 500 m Ma^{-1} and in big ocean bays 383 m Ma^{-1} (Gulf of Texas). The sedimentation rate only increases to 70 000 m Ma^{-1} in large rivers (Volga) but these rates are measured on modern deposits and in deltas with big catchment area. The sedimentation rate calculated from the sediment column originated in longer period is essentially lower (all data on sedimentation rate are gathered by Kukal, 1964).

3) Against the tendencies suggesting shifting the lower boundary of the Sarmatian upward are up to now unquestioned radiometric ages of the Sarmatian volcanic rocks. The numeric calibration of the Sarmatian (sensu Suess 1866) in the last radiometric scale of the Paratethys Neogene (Vass et al. 1987) is supported by 47 ages of volcanic rocks from the central, but also from the eastern Paratethys. Their Sarmatian age is reliably proved by biostratigraphic data. From these analyses, 26 ages document radiometric age of the Early Sarmatian in range 12.2 – 14.2 Ma. It is also possible to include among these ages the three later datings of volcanic glass from the base and one dating from the top of the Volhynian from two localities on the river Dnester and another two datings from the Kerch Pennninsula (Chumakov et al. 1992). The radiometric ages were performed by K-Ar and F.T. methods in several laboratories (Pisa, Italy; Campinas, Brasil; Jerevan, Armenia; Vladivostok, Russia; Bratislava, Slovakia; Debrecen, Hungary). Based on these ages the Sarmatian/Badenian boundary was assigned the numeric age 13.6 ± 0.2 Ma (Vass et al. 1987) or 13.5–14 Ma (Chumakov et al. 1992).

4) The substantiation of the calibration is supported by magnetostratigraphy from the borehole Berhida-3. Above we questioned correctness of the paleomagnetic measurement interpretations of the Pannonian/Sarmatian boundary from the borehole. However, we did not find a reason to object the measurement correctness on the Sarmatian/Badenian boundary. The magnetostratigraphic measurements imply (Fig. 2) the Sarmatian/Badenian boundary in time interval of chrons C5ACn and C5ADn i.e. between 13.65 and 14.6 Ma (compare Kokay et al. 1991 and Berggren et al. 1995).

Discussion to numeric calibration of the Sarmatian/Badenian boundary

Biostratigraphic data of Gaždzicka (1994) concerning the Late Badenian and Sarmatian of the Foredeep in the Poland are based on the correlation of the identified calcareous nanoplankton assemblages with the standard zones of Martini (1971). The correlation with the zone NN 8 is not done on the base of index form *Catinaster coalitus* which does not occur in the assemblages identified by Gaždzicka. It was done on the base of the species *Discoaster calcaris*, which datum of the first occurrence (FAD) is in the zone NN 8. Considering the uncertainty around the numeric calibration of the zone NN 8 (Berggren et al. 1995), very short duration of the zone (Fig. 3) and basic contradictions with actual chronostratigraphic and numeric scales, one can suspect if the FAD of *Discoaster calcaris* should be not shifted down in the time scale or if just the Carpathian Foredeep is not one of the places where this species occurred much earlier than in tropic or subtropic zones of the Pacific or India ocean. The absence of the index species of the zone NN 7 *Discoaster kugleri*, except for one specimen as referred by Gaždzicka, may reflect unfavourable conditions for flourishing of marine calcareous nanoflora (brackish Sarmatian sea) resulting in the absence of the species *D. kugleri*.

On the other hand it is not possible to omit the discrepancy in radiometric data. The radiometric ages of

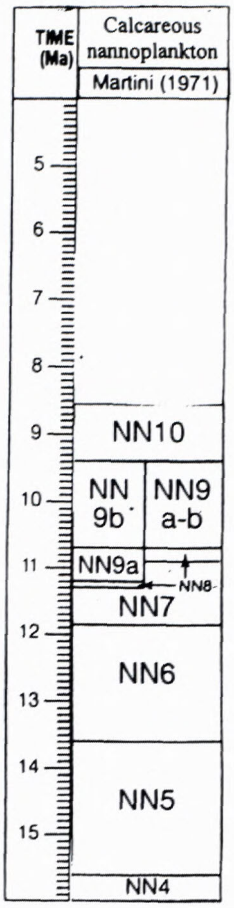

Fig. 3: Zones of the calcareous nannoplankton of the Middle and the early Late Miocene according to Martini (1971), numerically calibrated by Berggren et al. (1995). See the extremly short time interval of the zone NN8 as well as uncerteinly of its stratigraphic position.

redeposited tuffs in the Polish Foredeep, particularly the redeposited tuff from the uppermost part of the Skawina Formation (underlying evaporites from Wieliczka) 12.5 Ma (Banas & Bukowski 1997 fide Oszczypko 1997) and redeposited tuff from Chodenice Bed (overlying evaporites and underlying Krakow Bed, i.e. Sarmatian) around 12 Ma (Van Couvering 1981) are much younger than they should be according to the stratigraphic position and radiometric scale of the Neogene. However, it is necessary to add that redeposited tuff is the most unsuitable material for radiometric dating.

Conclusion

Revision of numeric age of the Sarmatian chronostratigraphic stage (Suess 1866) as suggested by Kokay et al. (1991), Oszczypko (1997, 1998) and as it is implied from the work of Gaždzicka (1994) is premature. Magnetostratigraphic interpretation of the Pannonian/ Sarmatian boundary, served as a base for Kokay et al. (1991) for proposal of the lowering of boundary age to 12.3 Ma, is not unambiguous. The chosen borehole in the part where the contact between the Sarmatian and Panonian is recorded, is not suitable for this purpose.

Biostratigraphic data from the Polish side of the Carpathian Foredeep are generally poor and unsuficient to introduce literally a revolutionary change of chronostratigraphy and numeric scale of the Neogene.

Radiometric ages, served as a base for the construction of the Neogene numeric time scale, are partly compromised by radiometric data from the Polish Carpathian Foredeep and they would totally compromised by biostratigraphy of the foredeep after cross correlation with the time scale of the Middle and Late Miocene constructed by Berggren et al. (1995). Radiometric dating of redeposited tuffs from the Carpathian Foredeep is not reliable according to the quality of the rocks dated. Also it is possible to doubt about the biostratigraphic correlation of the Middle Miocene in this area.

Because the statement against statement without new convincing evidence is unfruitful discussion, it will be very desirable to repeat or to do new radiometric dating of the Sarmatian volcanics by the technique of the 3rd millenium and also to continue in biostratigraphic and ecostratigraphic studies reaching by this way a new or more detail numeric calibration of the Sarmatian chronostratigraphic stage boundaries. Magnetostratigraphy applied to continuous profiles and profiles without hiatuses in the Sarmatian deposits would also be very helpful.

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