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# *SLOVAK GEOLOGICAL MAGAZINE*

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# Pyrenees, Alps, Northern Carpathians, Greater Caucasus: Essay of comparison

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*Abstract.* A brief analysis of the Alpine evolution and modern structure of the outer chain of the Alpine-Himalayan belt in Europe shows the main features of evolution and structure of these edifices, along with some important differences. The Pyrenees and Greater Caucasus, on the one hand, have much in common, as do the Alps and Carpathians, on the other hand. Transcurrent movements along a fault separating stable Eurasia in the south from the microcontinents detached from Gondwana – Iberia, Apulia, Tiszia and Transcaucasia, played a substantial role in the evolution and formation of these edifices, thus, their origin is due not to a simple orthogonal divergence and convergence, but to transtension and transpression.

*Key words:* Pyrenees, Alps, Northern Carpathians, Greater Caucasus, geotectonic evolution

## Introduction

The Pyrenees, Alps, Northern Carpathians and Greater Caucasus (including the Crimea Mts. and Kopetdagh) form the outer row of young fold-nappe edificies of the Alpine-Himalayan belt in Europe and Western Asia (Fig. 1). The Pyrenees, Alps and Northern Carpathians lie in direct continuation of one another, and the Crimea, Greater Caucasus and Kopetdagh, also form a continuous chain that is displaced relative to the first three, to the south along the Teisseyre-Tornquist lineament, active during all the time of their Alpine evolution (Kopp 1996). Despite this, all four edifices considered in this paper have much in common, not only in their position, but also in their Mesozoic-Cenozoic evolution and in their modern structure. However, notable differences exist between them, especially between the Pyrenees and Greater Caucasus on the one hand, and Alps and Carpathians on the other. The purpose of this paper is just to stress the common features and to show the differences between these edifices and to try to explain some causes of the latter.

## Evolution (Table I)

### *Nature of Pre-Mesozoic basement*

All four edifices concerned possess a Pre-Mesozoic metamorphic basement. This basement consists of three main entities:

1) Upper Paleozoic molasse complex, comprising Carboniferous coal-bearing lower molasse and Permian red, coarse, clastic upper molasse, both with some volcanic material, slightly metamorphosed and deformed;

2) Lower Paleozoic carbonate-terrigenous suite, comprising Ordovician-Devonian ophiolites and island arc volcanics, all metamorphosed to the greenschist or amphibolite facies, and participating in a Hercynian nappe structure;

3) Cadomian (Baikalian, Pan-African) metamorphic assemblage, rifted during the Cambrian – Early Ordovician. At the beginning of Late Paleozoic the Alps were situated between the rising Central Europe Variscan orogen and the Gondwana passive margin. The same situation is seen to have occurred in the Greater Caucasus because a Cadomian age now is established for the Dzirula Massif of Georgia (Zakariadze et al. 1998) and could be suggested for the Northern Carpathians, forming the direct continuation of the Alps.

A special case is represented by the southern-slope zone of the Greater Caucasus. There a deep–marine, mainly terrigenous sequence of Devonian or Silurian, to Triassic sediments is developed. It probably represents sediments derived from the continental slope and rise of the epi-Cadomian Gondwana margin and was deformed and slightly metamorphosed at the Triassic-Jurassic border time, during the Early Cimmerian orogeny.

### *Early Mesozoic quasi – platform cover*

After the end of the Hercynian orogeny, the area of all four edifices became for some time a part of the epi-Variscan West European or Scythian platform or, more strictly speaking, of their southern margin. In the Alps, Northern Carpathians and the Greater Caucasus we see in the Triassic a transition from a purely epicontinental, mainly lagoonal and clastic sedimentation to the conditions of an outer carbonate shelf, characteristic of the Austroalpine and Inner Carpathian domain and of the northern slope of the Greater Caucasus.

### *Rifting stage*

By the beginning of the epoch this quiet platformal evolution was interrupted by extension and rifting. This occurred at different times in each of the edifices and had different consequences. In the Alps and Carpathians rifting

Table 1

| | PYRENES | ALPS | | N. CARPATHIANS | | GREATER CAUCASUS | |
|---|---|---|---|---|---|---|---|
| | | N | S | N | S | N | S |
| $N_1^3$ - Q | | | | | Volcanism | Foredeep formation, Volcanism, Granites | |
| $P_3$ - $N_1^{1-2}$ | Foredeep formation, Molasses | Foredeep formations, Molasses, Granites | | Foredeep formation, Molasses | Molasses | Foredeep formations Molasses | |
| $K_2$ - $f_2$ | Flysch | Flysch | HP/LP metamorphism<br>Flysch | Flysch | Shallow marine deposits | Epicontinental shallow marine deposits | |
| $J_3$ - $K_1$ | Black shales, Rifting | Rifting, Spreading<br>Epicontinental shallow marine deposits | HP/LP metamorphism<br>Schistes lustres | Spreading<br>Epicontinental shallow marine deposits | Molasses<br>Neritic and bathyal deposits | Epicontinental shallow marine deposits | Flysch |
| $J_{1-2}$ | | | Spreading<br>Neretic a. bathyal deposits | | | Paralic coal-bearing formation, Volcanics | Black shales, Bimodal volcanics, Rifting |
| T | Epicontinental lagoonal deposits | Epicontinental lagoonal deposits | Rifting<br>Shallow marine deposits | Epicontinental lagoonal deposits | Rifting, Spreading | Shallow marine deposits | Accumulation of continental slope a. rise, Terrigenous deposits |
| $PZ_2$ | Molasses, Volcanics, Granites | Molasses, Volcanics, Granites | Molasses + shallow marine deposits, Volcanics, Granites | Molasses, Volcanics, Granites | Molasses, Volcanics, Granites | Molasses, Volcanics, Granites | |
| $PZ_1$ | Metamorphic complex | Metamorphic complex, Ophiolites | Metamorphic complex | Metamorphic complex, Ophiolites | Metamorphic complex | Metamorphic complex, Ophiolites | |
| $PR_3$ | | Metamorphic complex | | | | Metamorphic complex | |

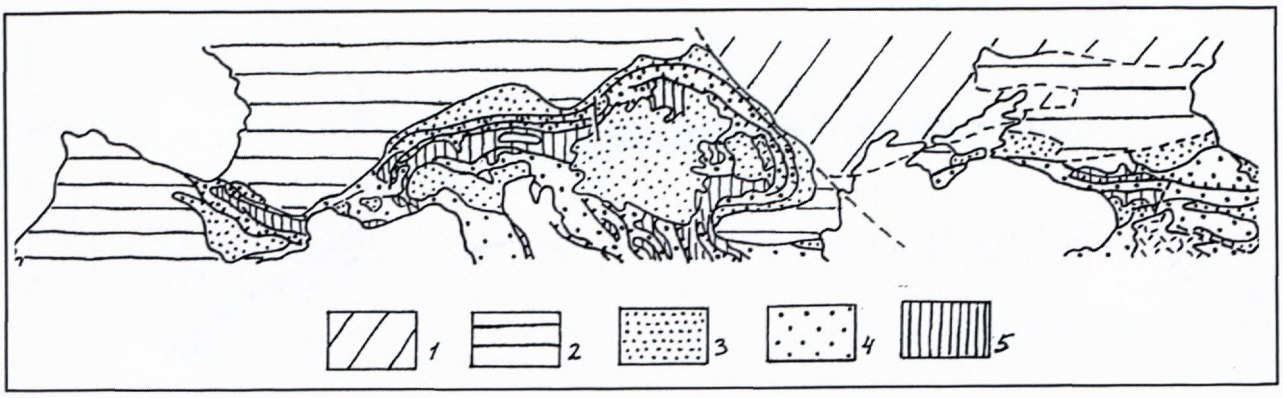


*Fig. 1 The outer row of the nappe-fold edifices of the Alpine-Himalyan belt of Europe and Western Asia*

already began during mid-Triassic time and led to the differentiation of their sedimentation area into relatively elevated and depressed blocks, with an accumulation in the latter of bathyal carbonate and siliceous (radiolarites) deposits. In the Pyrenees rifting took place during Aptian-Albian time, synchronously with the opening of the Bay of Biscay; this was accompanied by the deposition of black shales. In the Greater Caucasus rifting occurred at the very beginning of the Jurassic and ultimately created the deep marginal basin of the Neotethys, the main trunk of which passed through the central Lesser Caucasus.

Only in the Alps and Carpathians was continental rifting succeeded by spreading, documented by the presence of ophiolites. The rift-drift transition occured most early in easternmost Alps and Western Carpathians, manifested by the development of the Hallstatt-Meliata Ocean during the Middle Triassic. Then came the turn of the Liguro-Piemont Ocean, appearing in the Bathonian after two rifting phases, Hettangian-Sinemurian and Late Toarcian – Middle Jurassic (Froitzheim and Manatschal 1996). Spreading in this oceanic basin lasted until the end of Early Cretaceous. Meanwhile, during the Late Early Cretaceous, another narrower basin with oceanic crust, the Valais Basin appeared, presumably in the western prolongation of the Bay of Biscay and Pyrenees rift. During the Late Jurassic and Early Cretaceous, the Liguro-Piemont Ocean became the site of deposition of the „Schistes lustres" (Bundnerschiefer) formation.

In the Carpathians also, a second narrow oceanic basin (in addition to the Meliata one) was formed by the end of Jurassic. Its scarce remnants are preserved in the form of ophiolite detritus in the Pieniny Klippen belt (Mišík 1978). It disappeared later due to subduction of its crust under the Inner Carpathians.

In the Greater Caucasus rifting continued during the Early and Middle Jurassic but it is doubtful whether there was true spreading. The very thick black shale formation of that age contains volcanics of the spilite-keratophyre type and is similar to the Devonian of the Rhenohercynian zone of the European Variscides. Some basalts are even petrochemically akin to MORB. It seems most probable that extension of the pre-Alpine continental crust led only to its attenuation and transformation into crust of transitional type between continental and oceanic (suboceanic type).

Recently Lomize (1996) proposed a model of the development of the Greater Caucasus Jurassic basin (Fig. 2), based on the Wernicke pure-shear model, originally proposed for the Great Basin structure of the North American Cordillera. It is interesting that the same model was used by Froitzheim and Manatschal (1996) for the Central Alps, but for this domain they were able to show the transition from rifting to spreading.

*Beginning of convergence and basin closure*

Extension, creating the basins which later became the site of the fold-nappe edifices considered in this paper was followed by convergence and compression, leading, finally, to the closure of these basins and deformation of their infilling.

If we disregard the problematic manifestation of compression in the Greater Caucasus at the end–Middle Jurassic, which allegedly produced a ridge separating the flysch trough of the southern slope from an epicontinental basin of the northern slope, and envisage it rather as an effect of the prolonged rifting-upheaval of the rift shoulder – the earliest true compression began in the Eastern Alps and Inner Carpathians during the second half of the Early Cretaceous – the so called Austrian orogenic phase. This was also the beginning of the HP/LT metamorphism, provoked by the subduction of the Liguro-Piemont oceanic crust under the Apulian and Austroalpine margin (Marchant and Stämpfli 1997). According to these authors, the beginning of the closure of the Liguro-Piemont ocean was connected with the opening of the Valais ocean. This was also the time of the closure of the Meliata-Transylvanian Ocean on the southern border of the Austroalpine microcontinent and its Tiszia eastern prolongation. Subduction of the crust of this ocean already began during the Late Jurassic.

The next compression impulse was manifested in the Alps during the Senonian, and it attained its culmination in the Pieniny Klippen belt in the Carpathians at the Cretaceous/Paleogene border. But these Cretaceous compressive phases did not affect the Pyrenees and probably also the Greater Caucasus, where in the Early Cretaceous a new phase of rifting occured, as well as in the Crimea (Kopp and Khain 1996).

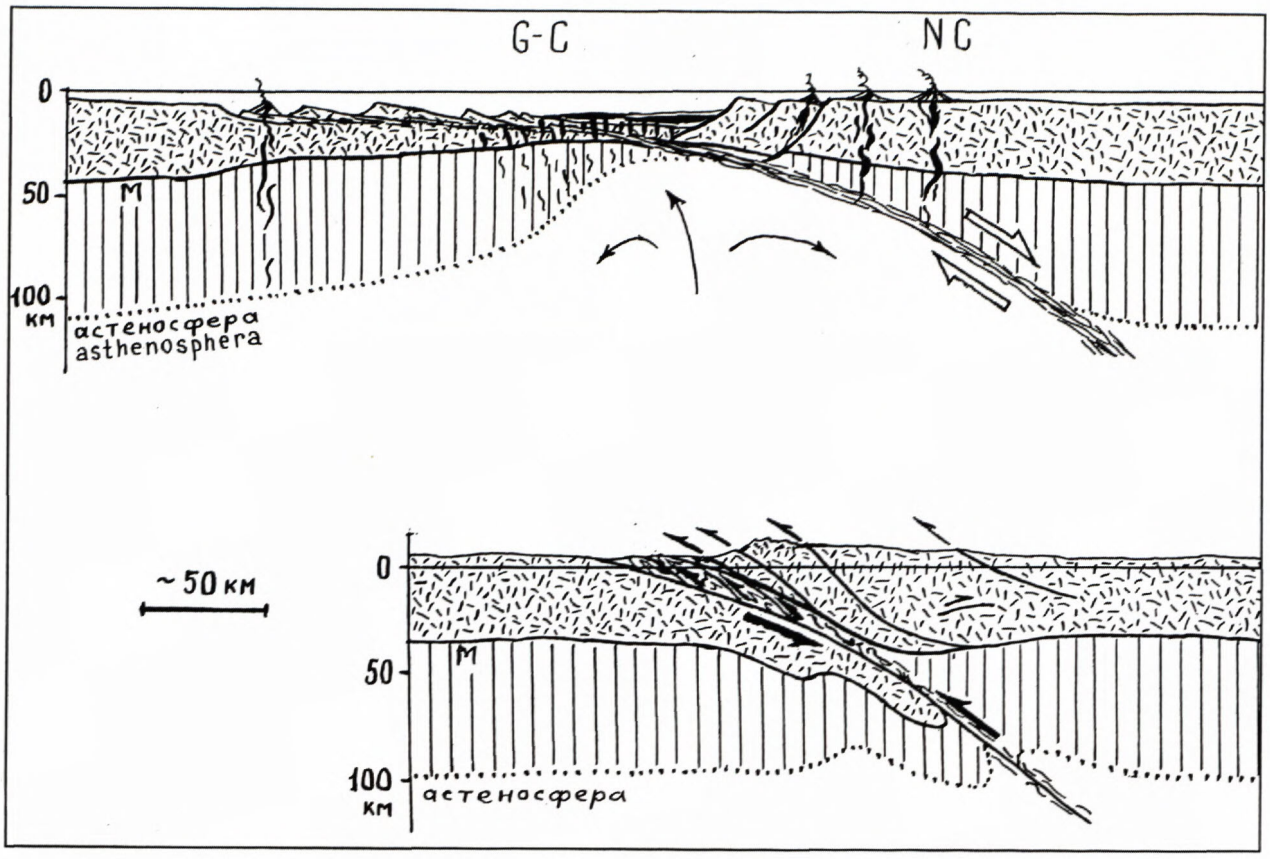


*Fig. 2 Evolution of the Greater Caucasus according to Wernicke model (after M. G. Lomize)*
*GC – greater Caucasus, NC – northern slope*

*Collision and orogeny*

The quiet evolution of the Pyrenees lasted until the Middle Eocene and was succeeded in the Late Eocene by an episode of strong compression as a result of the collision of the Iberian and Eurasian plates – the Pyrenean orogenic phase of Stille. The end-Eocene compression also marked the collision of Apulian and Eurasian plates and was most important in the Alpine tectonic history of the Alps, coinciding with the final creation of a thermal dome in their center that was closely followed by granite plutonism. However, in the Outer Carpathians the deformation only began during the Late Oligocene and attained its culmination during the Middle Miocene. The main phase of compressive deformation was manifested in the Greater Caucasus even later, by the end of Miocene. There is was produced by the northward shift of the Arabian plate after its separation from the African plate and opening of the Red Sea rift. But at the end of the Eocene the Greater Caucasus was already affected by a compressive impulse, leading to the beginning of its inversion, of the formation of molasse basins on its periphery, and of huge olistostromes descending from the flanks of the rising orogen toward these basins.

By the latest Miocene the formation of the fold-nappe mountain edifices of the Pyrenees, Alps and Carpathians, and of their molasse foredeeps was practically completed except for the burst of subaerial volcanism on the southern flank of the Carpathians. In the Greater Caucasus, especially its eastern part, the Pliocene-Quaternary evolution was quite different and very active. The formation of northern foredeeps and southern intermontane basins continued and at the end of the Pliocene and the beginning of the Quaternary strong fold–and–thrust deformation, and even of nappe development took place in the Terek and Kura basins on both flanks of the Greater Caucasus. Also volcanism occurred in its central part and granite intrusions were formed in places. The youngest of these igneous events dated at 2 Ma ago.

**Structure**

*Crustal structure*

Among the four edifices considered in this paper, the Pyrenees have the simplest and most symmetric structure. They form an axial ridge with exposures of Paleozoic basement rocks at their flanks composed of Mesozoic and Paleogene rocks that have folds, thrust faults and nappes, directed toward the Aquitaine and Ebro molasse basins to the south. But the magnitude of outward transport was notably larger on the southern than the northern flank.

The structure of the Greater Caucasus is more complicated than that of the Pyrenees. The dominant vergence is southward in the east, in the direction of Kura Basin.

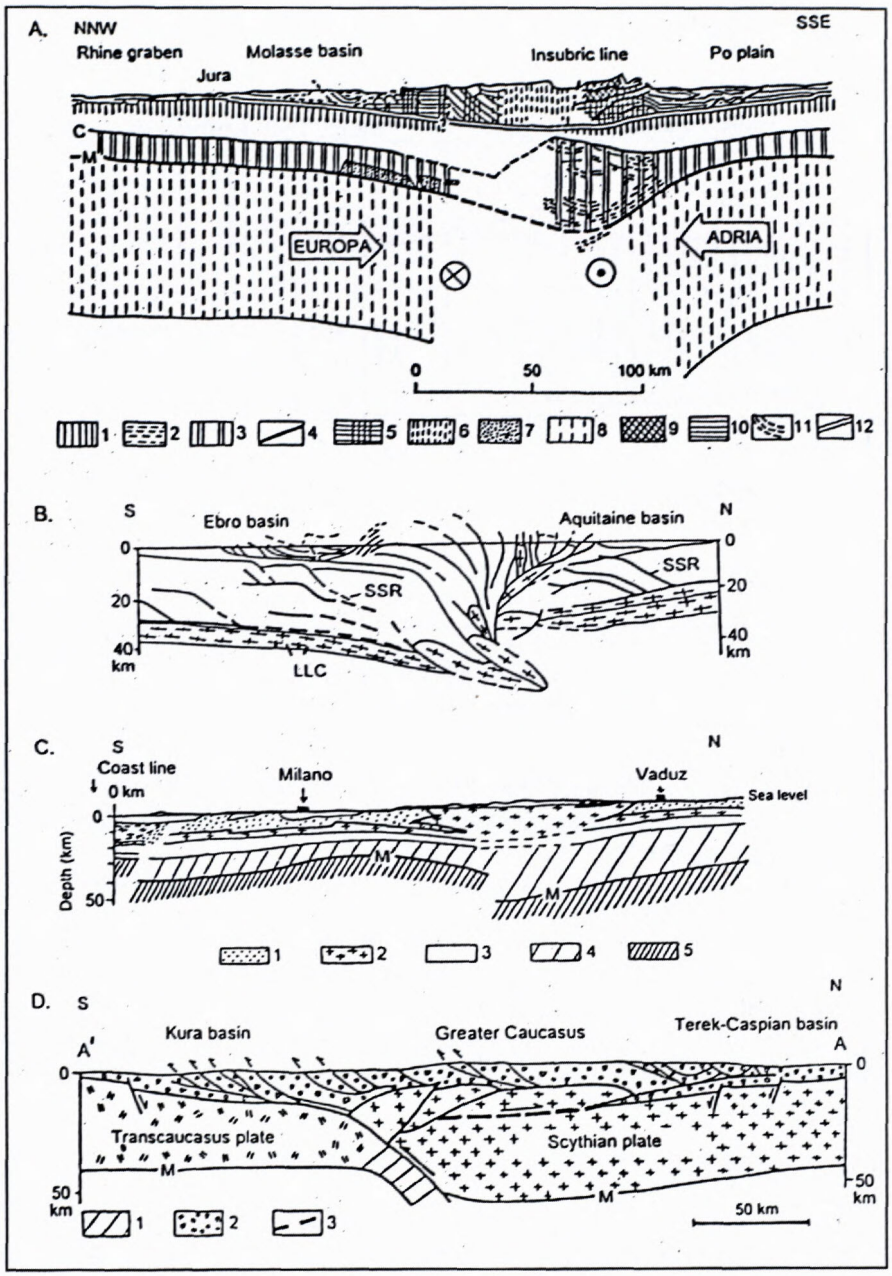


*Fig. 3 Deep seismic profiles through some orogenic edificed of the Alpine-Himalayan belt.*

***A.*** *Central Alps, according to the geological interpretation of Laubscher (1994).*
*Legend: 1 – crystalline basement of foreland, 2 – Helvetic nappes, 3 – lower crust from refraction, 4 – refraction Moho, 5 – external massifs of Jura phase, 6 – Penninic domain, 7 – European lower crust, 8 – mantle lithosphere, 9 – external massifs of earlier phases, 10 – post-Tortonian, 11 – lower crust imbrications stack, approximate, 12 – Jura phase basal front (brittle-ductile transition).*

***B.*** *Pyrenees, after Choukroune et al. (1990).*

***C.*** *Alps and Northern Apennines, after Cassonis et al. (1990), highly simplified.*
*Legend: 1-3 – upper crust ( 1 – sediments, 2 – granites), 4 – lower crust, 5 – mantle.*

***D.*** *Eastern Greater Caucasus, after Sobornov (1994).*
*Legend: 1 – oceanic crust, 2 – sediments, 3 – wave guide*

These was also some retro-thrust faulting at the western and eastern ends of the northern slope. The amount of tectonic transport of nappes on the southern slope attains 70 km. Nappes are also known in the Paleozoic basement; they are north-vergent as opposed to the southerly direction of the vergence of the alpine structure.

The scale of nappe development in the Alps and Carpathians is known; its amplitude is much more than 100 km and the amount of shortening attains several hundred kilometers. These differences between Pyrenees and Greater Caucasus, on the one hand, and the Alps and Carpathians, on the other, are reflected in the shape of the respective orogens – linear in the case of the two former and arcuate, convex toward northwest, north and northeast, of the two latter. And, in their turn, these differences are due to different conditions of plate convergence. This convergence was orthogonal in the case of Pyrenees and Greater Caucasus, with crustal subduction of the Iberian plate under the Eurasian in the Pyrenees and of the Transcaucasian plate also under the Eurasian in the Greater Caucasus. In the case of the Alps the Apulian (Adriatic) plate played the role of indenter, and in the case of Carpathians, the Austroalpine – Tiszia plate, pushed to ENE by the Apulian plate.

*Deep structure*

From the reflection and refraction profiling of the last decades we now possess adequate representation of the deep lithospheric structure of the Pyrenees, Alps and Western Carpathians, for the Alps we also have a tomographic profile. For the Eastern Carpathians and Greater Caucasus however, only refraction DSS profiles are available. Despite this lack of data, we may infer some analogy with the deep structure of better explored edifices.

It seems that the main common feature of the construction of all edifices, especially the better studied ones, is the structural disharmony between the three principal layers of the lithosphere – the upper crust, the lower crust and the lithospheric mantle. While the lithospheric mantle and lower crust of the lower of the

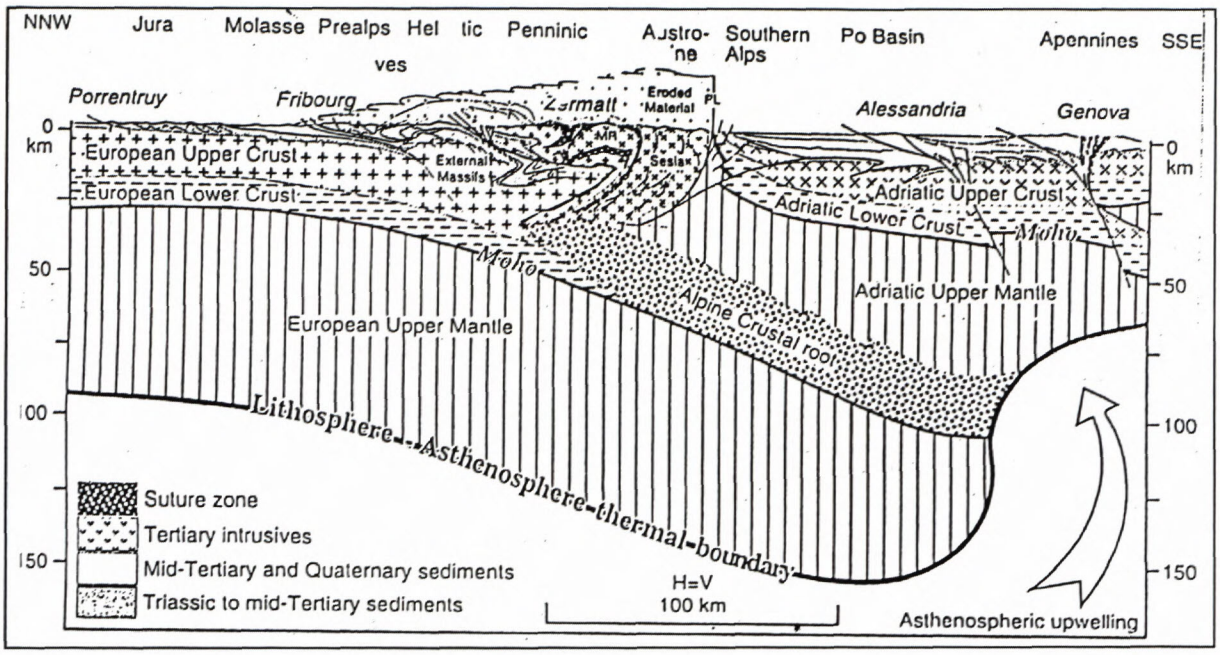


*Fig. 4 Lithosperic cross-section along the NFP-20 Western traverse (after Marchant, 1993).*
*MR – Monte Rosa nappe, PL – Periadriatic line*

colliding plates were subducted and partly absorbed in the astenosphere, the lower crust of the lower plate was partly squeezed and expulsed up even to the surface, and the upper crust of both plates was piled up, or stacked, thereby producing the antiformal structure of the mountainous edifice. This is most clearly expressed in the Pyrenees and Alps profiles (Figs. 3, 4). In detail, this picture is very complicated, with even some lateral indentation of layers. It must also be stressed that there has been a large amount of the subduction of the European continental crust under the Alps and Carpathians.

*Role of strike-slip faulting*

The existence of large longitudinal strike-slip faults is well recognized in the Pyrenees; the North Pyrenean fault and the Kopetdagh border fault on the other side of the Caspian Sea are similar structures. These faults are more conspicuous in the Alps and especially in the Carpathians, where they must be invoked to explain the formation of their oroclinal bend and are involved in the evolution of their earlier connection with the oceanic basins of the Alpine and Carpathian domains (Dal Piaz et al. 1995). The sense of movement along these faults changed; it was sinistral until the Mid-Cretaceous and dextral thereafter.

**Conclusion** (Table II)

All four edifices considered in this essay are situated along the southern margin of the Eurasian lithospheric plate, which was a passive margin during all of their Mesozoic-Cenozoic evolution. The southern border of these edifices at the time of the beginning of their Alpine evolution was represented by microplates – fragments of Gondwana (Africa) – Iberia, Apulia, Transcaucasia. The Austroalpine – Tiszia microplate primarly belonged to Eurasia, but later they played the role of hinterland, with respect to Eastern Alps and Carpathians.

Initially, all four edifices possessed a metamorphic Hercynian-Cadomian basement and, during part of the Mesozoic evolution, acted as a part of the epi-Hercynian West European or Scythian platform. One exception was the actual Greater Caucasus southern slope, which from Devonian or even earlier time showed conditions of continental slope and rise, probably of the Gondwana margin subsided.

This platformal regime was the shortest-lived, it lasted only to the Middle or Late Triassic in the Carpathians (Southern slope), to Early Jurassic in the Greater Caucasus, to Bathonian in the Western and Central Alps and to Aptian in the Pyrenees. Thereafter rifting and destruction of the epi-Hercynian continental crust began and thus led to its extension and to the formation of deep basins on oceanic (Alps, Carpathians) or of transitional – suboceanic (Greater Caucasus) or of attenuated continental (Pyrenees) crusts. A second episode of rifting–spreading took place during the Late Jurassic (Pieniny Klippen Belt) or mid-Early Cretaceous (Valais) in the Carpathians and the Alps, respectively.

The first manifestations of compression were felt in the Carpathians and possibly Greater Caucasus at the end of Middle Jurassic and the Late Jurassic, but an important deformation began in Eastern Alps and Inner Carpathians during mid-Early Cretaceous and was repeated in the Senonian. This compresion was the result of the subduction under the microplates the oceanic crust to the south. These deformations of were followed by the beginning of flysch accumulation, which attained its largest expansion during the Late Cretaceous and Early Paleogene.

Collision began during the Late Eocene in the Pyrenees and in the Alps, during the Late Oligocene–Early Miocene in the Carpathians and during the Miocene in the Greater Caucasus. In the eastern part of the Greater Caucasus, as well as in the Eastern Carpathians, the deformation was renewed at the Pliocene – Quaternary boundary.

The amount of extension, and the width, depth and character of the crust of the original marine basins had a fundamental implication on the future architecture of the orogenic edifices: these determined the degree of shortening and the complexity of this structure. The degree of shortening was approximately 100 km for the Pyrenees, 200-250 km for the Greater Caucasus and several hundreds of kms for the Alps and Carpathians. In this respect, the Alps and Carpathians could be called hypercollision orogens.

Table 2

| Edifices<br>Features | PYRENEES | ALPS | N. CARPATHIANS | GREATER CAUCASUS |
|---|---|---|---|---|
| Initial opening of the Alpine basin | Late Aptian | Bathonian (S)<br>Aptian (N) | Mid-Triassic (S)<br>End-Jurassic (N) | Sinemurian |
| Nature of basin crust | Attenuated continental | Oceanic | Oceanic | Suboceanic |
| Ophiolites | Absent | Present | Present | Absent |
| Time of main deformation | Late Eocene | Late Eocene | Mid-Eocene | Late Miocene |
| Regional metamorphism | Absent | Present<br>(inc. HP/LT) | Present<br>(inc. HP/LT) | Absent |
| Alpine granite plutonism | Absent | Present | Present | Present |
| Shape | Linear | Arcuate | Arcuate | Linear |
| Vergence | Mainly Southern | Northern | Northern | Southern |
| Nappe amplitude | n x 10 km | > 100 km | > 100 km | n x 10 km |
| Sense of plate convergence | EU<br>IB | AP<br>EU | TI<br>EU | EU<br>TR |
| Recent volcanism | Absent | Absent | Present | Present |
| Basement | Hercynian | Hercynian + Cadomian | Hercynian + Cadomian | Hercynian (N)<br>+ Cadomian (S) |

Finally, the important role played by the transcurrent faulting in the evolution of the Alpine edifices considered here must be stressed again.

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# Mafic dykes in Variscan tonalites of the Malá Fatra Mts. (Western Carpathians)

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*Abstract*: Dykes of mafic rocks occur in the Variscan tonalites in the Malá Fatra Mts. They contain numerous xenoliths, or desintegrated parts of the tonalite host rock. Porphyric phases in dykes are mainly clinopyroxenes, while amphiboles, and biotites are less common. On the basis of prevailingly porphyric clinopyroxenes and their chemical composition, we assign these rocks to the group of calc-alkaline lamprophyres. Taking their mineral composition as a criteria we term them as pyroxene spessartites. Preserved mylonites/blastomylonites of these rocks indicate their Late Variscan/Early Alpine age.

*Keywords*: calc-alkaline lamprophyres, clinopyroxenes, feldspars, petrology.

## Introduction

Dykes of mafic rocks located in the pre-Upper Carboniferous complexes of the Central Zone of the Western Carpathians belong to the rock types of scarce occurrence. They are known to occur in the Malá Fatra Mts., Nízke Tatry Mts., Suchý Mts., Považský Inovec Mts. and the Veporic units of the Slovenské rudohorie Ore Mountains, but only in the first two mountains their occurrences are more abundant, while in the others only scarce bodies were reported.

## Geological setting

Dykes of mafic rocks are known to occur exclusively in the Variscan tonalite environment. Their thickness is as much as a few meters. On the floor of a productive quarry Dubná Skala near Vrútky there crops out (September 1996) as a few meters thick dyke-like body. Very sporadically observed contacts between mafic veins and the surrounding tonalites are characterized by direct contact planes and as much as 2 cm thick zones of intense chloritization of the host rock. Within thicker veins, the transitions from porphyric and amygdaloidal types into the holocrystalline, evenly granular types in the central parts of veins, can be observed. The dips of dykes are variable, but steep to vertical ones predominate. Characteristic, numerous tonalite xenoliths are randomly distributed not only within individual dykes, but also among the veins in general. The tonalites are intersected by zones of (Alpine) mylonitization and blastomylonitization, in which also the mylonites and blastomylonites of discussed mafic rocks occur. They have a character of distinctly schistose, mostly by naked eyes observed aphanitic fabric.

Characteristic features including the geological setting of dykes and their general petrographic description are presented in a summary paper by Hovorka (1967). This author characterized the dyke rocks in tonalites of this mountain as 4 basic types as follows:

a) porphyric,
b) evenly granular,
c) amygdaloidal,
d) schistose types.

While Ivanov and Kamenický (1957) reported that the rocks under study are akin to cuzelites with some features in common with the odinites and kersantites, Hovorka (1967) described them as transitional rocks of the series porphyrite - lamprophyre (diorite porphyrite - kersantite - spessartite).

To date available general information about the mafic dyke rocks located in the Variscan tonalites of the Malá Fatra Mts. can be summarized as follows:

- discussed mafic rocks are known only from the magmatite/tonalite environment of Variscan age,
- while Ivanov and Kamenický (1957) reported the presence of mafic dykes in both parts i.e. in the group of Veľký Fatranský Kriváň and in the Minčol group, the authors of this paper record only a few occurrences in the Minčol group (Martinské hole ridge),
- described in the mafic dykes are the features of assimilation of surrounding tonalites (Hovorka, 1967) in a processes that at least to some degree modified the composition of the dyke melt,

5 analyses of petrogenic elements (Hovorka, 1967) were used to characterize the composition of mafic dykes. The content of $SiO_2$ in the analyzed rocks ranges from 44.5 to 50.5 wt. %.

The above summary of information indicates that exact data (REE, analyses of discriminatory minerals in the rocks etc.) which would allow for an assignment of the rocks under study to one of the general magmatic types, or to compare them with those known from the Mesozoic sequences of the Central Zone of the Western Carpathians, are not yet available.

This paper presents the results of study of clinopyroxene, a mineral of genetic importance and of distinctly magmatic origin. On the basis of clinopyroxene composition (zoning, twinning, two generations) we assign these rocks to a magma group and characterize the development of these rocks in time and space.

## Mineralogy and petrology

The dyke rocks are dark-green to grey-green, massive, predominantly aphanitic, with evenly granular, less porphyric textures. Most phenocrysts (as much as 5 mm across) are clinopyroxenes, while biotites, amphiboles and plagioclases are of scarce occurrence. However, the latter phases form phenocrysts of microscopic dimensions (as much as 0.5 mm across). Most tonalite xenoliths are partly altered and measure as much as a few cm in diameter. They can best be observed on the cutting, or fracture surfaces. The xenoliths are: a) compact, b) desintegrated (so, that on the flat surfaces, the desintegration planes of xenoliths are evidently visible) and c) epidotized. The latter process affected mainly the original plagioclases in tonalite. Most samples have amygdaloidal structure (amygdales measure as much as 7 mm, but most have 2-5 mm in diameter). In blastomylonites of the mafic rocks, the amygdales have flat, distinctly elongated shape. They are made up of dark-green to almost black, fine-scaly chlorite aggregate, but in thin sections, a carbonate mineral was also identified in the amygdale filling.

Our additional microscopic study (of some 25 thin sections) confirmed that the assignment to the basic rock types as stated at the beginning of this paper, was justified.

Clinopyroxene stands for the principal phyric phase of these mafic rocks. Its distribution in the rocks is random. The crystals are euhedral, as much as a few milimeters across, locally magmatically corroded and of short-prismatic habit. Glomeroblasts of several clinopyroxene crystals were scarcely observed. The cpx crystals have a fresh outlook, but signs of carbonatization may also be observed along the cleavage (110) planes. Optical and microprobe studies confirmed the presence of "hourglass" texture/twinnings of phyric Cpx.

Apart from the distinctly magmatic, phyric Cpx I, the plagioclase, quartz and biotite, which also attain a size of phyric phases (as much as 1, scarcely 2 mm), also occur in the rocks.

Characteristic features of biotite flakes are their freshness and intense brown pleochroism. According to its relative abundance, this mineral belongs to the subordinate category, but in the majority of cases it falls within the category of accessory silicates. Most porphyric plagioclases have a diameter of 2-4 mm, but some may have as much as a few cm (!) across. Their crystal morphology is variable, either euhedral, or irregular. Genetically, they belong to: a) xenocrysts representing desintegrated tonalite or other plutonic rocks xenoliths and to b) magmatic neomorphs. The xenocrysts form the cores of compositionally complicated crystals that were totally replaced by a mixture of fine-grained minerals (biotite, carbonate, quartz, epidote-zoisite group of minerals, chlorite). These irregularly shaped "cores" are overgrown by water transparent plagioclase ($An_{30}$) characterized by a repeated albitic twinning. It is analogous to the plagioclase crystals that occur as much smaller (2-4 mm) phenocrysts. Together with the distinctly laths-like, and mostly intensively "filled" (desintegrated) plagioclasses of the rock matrix, they are a predominant magmatic phase of the rock. The plagioclase laths in the matrix are either simple, or albite twinned. In most thin sections, the transitional types of plagioclases of magmatic provenance, i.e. fine, slender laths and platelets of plagioclases can be observed, whose size (0.X - 1 mm) indicates a transition to the magmatic, porphyric plagioclases.

Quantitatively, quartz belongs to the accessory mineral phases. Genetically, (it only forms xenocrysts of the porphyric development) it corresponds to xenoliths of magmatic rocks (of tonalitic series!). It is to various degrees corroded, intensively pigmented and narrow reaction rims develop along the quartz crystal margins. Scarce poikiloblasts of apatite and zircon are present in quartz.

The matrix of these mafic rocks is predominated by plagioclase laths and by fine-grained chlorite aggregate, metallic ore pigment and grey, low birefringence mass (recrystallized volcanic glass). Plagioclases are intensively altered into a submicroscopic aggregate of white mica and clay minerals. Sphene, hercynite (?), ilmenite and other metallic ore phases are present in accessory amounts (besides of other above mentioned phases).

Another type is represented by the evenly granular varieties that occur only as the dyke facies of the above type. Besides of predominant clinopyroxenes and platy plagioclases, the rocks also contain monoclinic amphibole with a distinct, green pleochroism that is scarcely segmented to form splinter-like features. These holocrystalline types also contain accessory amounts of biotite, quartz, apatite, sphene, ilmenite and secondary minerals (calcite, chlorite, sericite, minerals of clinozoisite-epidote group).

Amygdaloidal types are the most common variety of the mafic rocks that occur in the Malá Fatra Mts. tonalite. The amygdales measure 5, but sporadically as much as 8 mm across. They are either isometrical, or of a distinctly elongated (biscuit) shape. Their colouration is deep-green to green-black. The amygdale fillings are submicroscopic and granular. They are homogeneous, composed of fine lathy aggregate of intensively pleochroic Fe-chlorite. In other types, compositional zoning of amygdales can be observed; in such case, the amygdale rims are filled up by chlorite with local indications of radial arrangement. Towards the centre of the amygdale a calcite zone occurs, which is scarcely replaced by a quart

Tab. 1 Selected analyses of clinopyroxene

| Sample | DS 19/1 | | DS 29/1 | | | | | | | DS 29/3 | | | DS 18/1 | | DS 18 | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Number | 1N | 1N | 2pr | 3py | 2pr | 3py | 4m | 4m | 2pr | 3py | 2pr | 3py | 1N | 1N | 4m | 4m |
| $SiO_2$ | 48,98 | 49.56 | 48,68 | 52,54 | 48,58 | 52,24 | 52,21 | 51,12 | 49,32 | 51,75 | 48,98 | 52,15 | 49,42 | 50,97 | 50,37 | 51,28 |
| $TiO_2$ | 2,51 | 2,26 | 2,20 | 1,32 | 2,16 | 1,22 | 1,16 | 1,71 | 1,95 | 1,14 | 2,29 | 1,27 | 1,66 | 1,29 | 1,78 | 1,36 |
| $Al_2O_3$ | 5,22 | 5,50 | 6,22 | 2,70 | 6,14 | 3,09 | 2,79 | 3,56 | 5,54 | 3,01 | 6,22 | 2,86 | 4,54 | 3,61 | 4,62 | 4,33 |
| $Cr_2O_3$ | 0,00 | 0,10 | 0,31 | 0,00 | 0,00 | 0,06 | 0,09 | 0,00 | 0,44 | 0,12 | 0,32 | 0,00 | 0,17 | 0,19 | 0,29 | 0,46 |
| $FeO^+$ | 8,30 | 7,51 | 7,57 | 8,34 | 8,59 | 7,39 | 7,60 | 8,64 | 7,29 | 7,50 | 7,21 | 8,72 | 7,55 | 7,27 | 7,54 | 6,87 |
| MnO | 0,30 | 0,26 | 0,19 | 0,26 | 0,26 | 0,23 | 0,16 | 0,23 | 0,23 | 0,20 | 0,22 | 0,25 | 0,17 | 0,17 | 0,08 | 0,12 |
| MgO | 12,57 | 13,24 | 13,08 | 14,72 | 12,99 | 15,01 | 14,96 | 13,63 | 14,01 | 15,33 | 13,16 | 14,20 | 14,31 | 15,08 | 13,35 | 14,04 |
| CaO | 20,35 | 20,12 | 20,39 | 20,24 | 20,48 | 19,00 | 20,20 | 19,85 | 19,73 | 19,42 | 20,69 | 19,69 | 20,18 | 20,04 | 21,08 | 21,35 |
| $Na_2O$ | 0,43 | 0,38 | 0,68 | 0,50 | 0,72 | 0,47 | 0,59 | 0,54 | 0,62 | 0,43 | 0,40 | 0,31 | 0,58 | 0,47 | 0,44 | 0,43 |
| $K_2O$ | 0,00 | 0,00 | 0,00 | 0,01 | 0,02 | 0,01 | 0,00 | 0,00 | 0,00 | 0,00 | 0,02 | 0,00 | 0,01 | 0,01 | 0,02 | 0,00 |
| TOTAL | 98,66 | 98,93 | 99,32 | 100,63 | 99,94 | 98,72 | 99,76 | 99,28 | 99,13 | 98,90 | 99,51 | 99,45 | 98,59 | 99,10 | 99,57 | 100,24 |
| Formula based on 6 O | | | | | | | | | | | | | | | | |
| $Si^{IV}$ | 1,85 | 1,85 | 1,82 | 1,93 | 1,81 | 1,94 | 1,93 | 1,91 | 1,84 | 1,93 | 1,82 | 1,94 | 1,86 | 1,90 | 1,87 | 1,89 |
| $Al^{IV}$ | 0,15 | 0,15 | 1,18 | 0,07 | 0,19 | 0,06 | 0,07 | 0,09 | 0,16 | 0,07 | 0,18 | 0,06 | 0,14 | 0,10 | 0,13 | 0,11 |
| $Al^{VI}$ | 0,08 | 0,09 | 0,09 | 0,05 | 0,08 | 0,08 | 0,05 | 0,07 | 0,08 | 0,06 | 0,10 | 0,06 | 0,06 | 0,06 | 0,08 | 0,08 |
| Ti | 0,07 | 0,06 | 0,06 | 0,04 | 0,06 | 0,03 | 0,03 | 0,05 | 0,05 | 0,03 | 0,06 | 0,04 | 0,05 | 0,04 | 0,05 | 0,04 |
| Cr | 0,00 | 0,00 | 0,01 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,01 | 0,00 | 0,01 | 0,00 | 0,01 | 0,01 | 0,01 | 0,01 |
| $Fe^{3+}$ | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 | 0,00 |
| $Fe^{2+}$ | 0,26 | 0,23 | 0,24 | 0,26 | 0,27 | 0,23 | 0,24 | 0,27 | 0,23 | 0,23 | 0,22 | 0,27 | 0,24 | 0,23 | 0,23 | 0,21 |
| Mn | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,01 | 0,00 | 0,00 |
| Mg | 0,71 | 0,74 | 0,73 | 0,81 | 0,72 | 0,83 | 0,82 | 0,76 | 0,78 | 0,85 | 0,73 | 0,79 | 0,80 | 0,84 | 0,74 | 0,77 |
| Ca | 0,82 | 0,81 | 0,82 | 0,80 | 0,82 | 0,76 | 0,80 | 0,79 | 0,79 | 0,77 | 0,83 | 0,78 | 0,81 | 0,80 | 0,84 | 0,84 |
| Na | 0,03 | 0,03 | 0,05 | 0,04 | 0,05 | 0,03 | 0,04 | 0,04 | 0,04 | 0,03 | 0,03 | 0,02 | 0,04 | 0,03 | 0,03 | 0,03 |

Explanation: 1/N = number in figure symbols/type of pyroxene, N = new formed rims, pr = prismatic sector, py = pyramidal sector, m = microlite

Tab. 2 Selected analyses of feldspars

| N. anal. | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 67,318 | 68,322 | 67,578 | 66,943 | 66,607 | 62,83 | 63,59 | 62,33 | 66,93 | 65,208 | 66,07 |
| $Al_2O_3$ | 18,161 | 18,363 | 17,745 | 17,748 | 18,128 | 22,88 | 22,00 | 23,10 | 17,94 | 17,447 | 17,264 |
| $FeO^+$ | 0 | 0 | 0 | 0 | 0 | 0,00 | 0,03 | 0,00 | 0,00 | 0 | 0 |
| CaO | 0,224 | 0,382 | 0,055 | 0,198 | 0,443 | 5,63 | 5,23 | 5,49 | 0,23 | 0,003 | 0 |
| $Na_2O$ | 5,342 | 5,58 | 5,55 | 5,473 | 5,52 | 7,57 | 7,91 | 7,56 | 0,66 | 0,532 | 0,312 |
| $K_2O$ | 8,884 | 7,993 | 9,152 | 8,781 | 8,096 | 1,48 | 1,74 | 1,64 | 14,06 | 15,083 | 15,732 |
| TOTAL | 99,929 | 100,64 | 100,08 | 99,143 | 98,794 | 100,38 | 100,50 | 100,12 | 99,84 | 98,273 | 99,378 |
| Formula based on 8 oxygens | | | | | | | | | | | |
| Si | 3,029 | 3,036 | 3,04 | 3,036 | 3,023 | 2,786 | 2,819 | 2,774 | 3,052 | 3,044 | 3,055 |
| Al | 0,963 | 0,962 | 0,941 | 0,949 | 0,97 | 1,196 | 1,15 | 1,212 | 0,962 | 0,96 | 0,941 |
| Na | 0,466 | 0,481 | 0,484 | 0,481 | 0,486 | 0,651 | 0,68 | 0,652 | 0,058 | 0,048 | 0,028 |
| K | 0,51 | 0,453 | 0,525 | 0,508 | 0,469 | 0,084 | 0,098 | 0,093 | 0,818 | 0,898 | 0,928 |
| Ca | 0,011 | 0,018 | 0,003 | 0,01 | 0,021 | 0,267 | 0,248 | 0,262 | 0,011 | 0 | 0 |
| or | 51,69 | 47,59 | 51,89 | 50,86 | 48,05 | 8,36 | 9,58 | 9,24 | 92,16 | 94,93 | 97,09 |
| alb | 47,24 | 50,51 | 47,83 | 48,16 | 49,76 | 64,96 | 66,23 | 64,77 | 6,58 | 5,07 | 2,91 |
| an | 1,07 | 1,9 | 0,29 | 0,97 | 2,19 | 26,69 | 24,19 | 25,99 | 1,27 | 0 | 0 |

zose (chalcedony) aggregate in the central parts of the amygdales. Locally, the chalcedony has an indistinctly radial orientation.

Of little interest for solving the genetic problems are the Alpine recrystallized types (of mylonitic and blastomylonitic character), such as those exposed in a forestry track 2 km SSE of the Kunerád castle, on a slope of the triangulation point 815.1 m, north of Sučany and elsewhere. Their general characteristics are described in the paper by Hovorka (1967).

A characteristic feature of the mafic dykes located in tonalites of the Malá Fatra Mts. is the presence of numerous xenoliths of magmatic rocks of tonalite series. They measure as much as several centimeters across, and many are mechanically desintegrated due to their incorporation into the mafic melt. The desintegration into individual crystals can best be observed on the cutting faces of the rock. As seen in thin sections, this phenomenon is due to fluidal arrangement of lath-like plagioclases in the rock matrix. The desintegration of incorporated xenoliths (of tonalites) took place when they became overheated and the mafic magmatic melt moved within the tonalitic body. Most tonalite xenoliths are to various degrees hydrothermally altered - mainly epidotized (plagioclases and biotite in tonalite). Intense epidotization affects preferably the endocontact zones in individual xenoliths. This is subsequently made up of fine-grained aggregate of epidote minerals. In the plagioclases of tonalite xenoliths, the carbonatization also took place. Of interest is that most xenoliths have in the marginal parts the fan-shaped arrangement of alkaline feldspar needles (Table 2). The needles of alkaline (natrium-potassium) feldspars intimately overgrow with the potassium feldspars. These minerals probably formed due to thermic-metasomatic processes during the interaction between xenoliths and magma. The occurrence of such types of feldspars were not yet reported from the similar rock types.

The above review indicates that most mafic rocks have porphyric textures, ophitic matrix and originally intersertal texture. Local arborescent arrangement of plagioclase laths can be observed. Typical is also the fluidal arrangement of plagioclases in the matrix, as to attain a trachytic features (texture) where the porphyric phases are missing.

**Composition of minerals**

Genetically, the best evidence for the origin of these rocks can give the clinopyroxenes which occur as a distinctly porphyric phase, as well as a part of the fine-grained matrix. Besides of clinopyroxenes we have also studied the composition of plagioclases.

We used the JEOL Superprobe 733 installed at the Geplogical Survey of the Slovak Republic, Bratislava, operated by Dr. P. Konečný to analyze the above minerals at standard conditions(Tables 1 and 2).

The study of clinopyroxene composition is of importance when the genetic type of magma is to be determined and when the evolution of magma is to be deciphered.

On the basis of geological setting, composition and origin, three types of clinopyroxenes can be distinguished:

- clinopyroxene phenocrysts,
- newly formed rims around older pyroxenes,
- microlites in the matrix.

The pyroxene phenocrysts in the rocks under study are almost always zonal, but the most common zoning is that of the "hourglass" type, while the oscilatory zoning is scarce and relatively indistinct. The sector-like zonal structure is characterized by two distinctly differing sectors, a pyramidal one and a prismatic one. These sectors differ not only optically, but mainly by their chemical composition. Relative to prismatic sector, the pyramidal one is commonly enriched by Si and Mg and depleted by Al, Ti and Fe. This feature is usually explained as the concurrent vector growth of all parts of a crystal, when each plane has its own growth rate and creates the partial equilibria between the crystal surface and the melt (Hollister & Gancarz 1971, Leung 1974). Since the attachment of diopside type units runs relatively easier way, the pyramidal sectors grow faster and the melt at the contact with them grows relatively depleted by Mg a Si. The slower growing prismatic sector is being enriched by other elements, such as Ti, Al, Na and K (locally also Fe). As regards the stability of the crystal structure, the sector zonality represents a disequilibrium state, thus, it is subject to obliteration during later diffusional processes. Its preservation depends on the rate of crystallization and diffusion in the crystal. Hence, the sector-like zonal structure is usually preserved in the fast solidifying rocks, which is the case of mafic dykes located in the Malá Fatra Mts. tonalites.

Of another type are the newly formed rims, or water-transparent margins of large pyroxene phenocrysts (symbol 1). These probably developed immediately prior to the ascension of magma to the surface, under changing pT conditions in a magmatic chamber. Their composition is similar to that of the prismatic sectors.

Yet another type of pyroxenes, typical for the fast-cooling rocks, are the microliths in the matrix. Their shapes are irregular and composition is similar to that of the pyramidal sector. They mark the latest crystallization stage during the ascension of magma towards the surface, i.e., they indicate a reduction of pressure and a relatively fast melting.

The study of pyroxene composition is of great importance when the origin of the protolith is to be determined. In the classification diagram (Morimoto et al. 1988, Fig. 1), some figurative points for selected clinopyroxenes fall within the field of augite, while the others fall within the field of diopside (especially microliths, newly-formed margins and prismatic sectors). Generally, the distribution of figurative points of clinopyroxenes from the samples studied differs from that of the clinopyroxenes that occur in Mesozoic alkaline basalts of the Malá Fatra Mts. (Hovorka & Spišiak 1988, hatched field). In the Ti:Al diagram (Fig. 2), the figurative points of clinopyroxene analyses are widely scattered indicating relatively good correlation be-

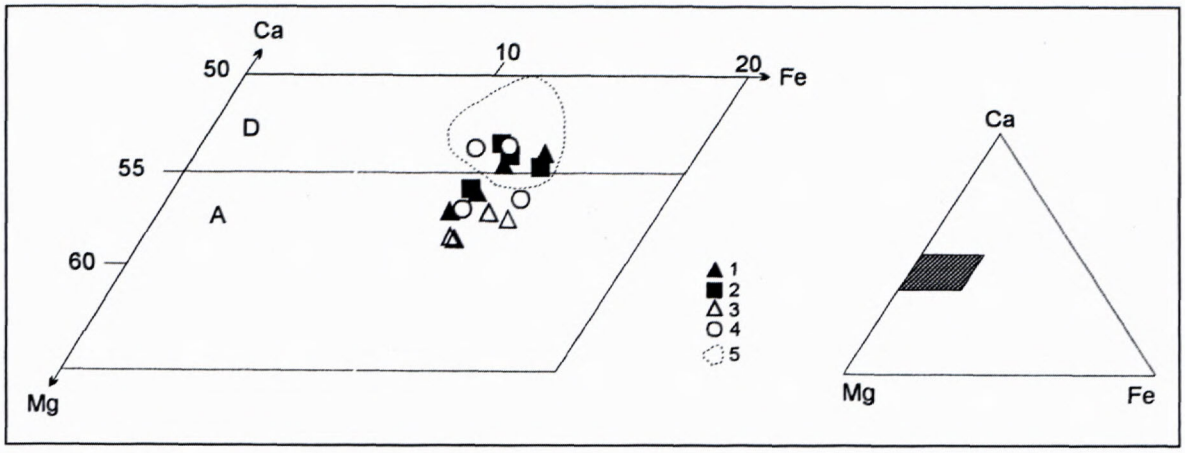



*Fig. 1. Classification diagrams of clinopyroxenes (Morimoto et al. 1988). 1 - 4 = analyses of clinopyroxenes from Table 1; 5 - field of clinopyroxenes from Mesozoic basanites (Hovorka - Spišiak 1988; A= augite, D = diopside.*

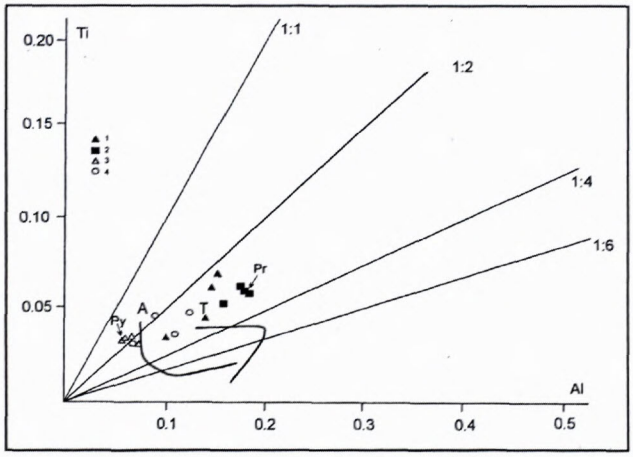



*Fig. 2. Ti vs. Al in pyroxenes; plotted field show the maximum content of Ti in tholeiitic basalts (T) and the minimum content in alkaline basalts (A) (fields after Maruyama 1976 in Takeda 1984), 1- 4 analyses of clinopyroxenes from Table 1, m Py = pyramidal sector, Pr = prismatic sector*

tween Ti and Al. Projections of mean analytical points approach the Ti:Al ratio of 1:3, whereas the individual types, or parts of clinopyroxenes have different projections. Relatively lowest contents of Ti and Al have the pyramidal sectors and microliths, while the prismatic sectors and newly-formed Cpx have relatively highest contents. To classify the clinopyroxenes more precisely, we used the diagram showing the dependence $SiO_2$ vs. $Al_2O_3$ (Fig. 3).

In this diagram, the projection points of analysed samples fall within the fields S and A, i.e., in the field of low, or normal alkalinity rocks. Almost all analyses fall within the field coincident with the field of clinopyroxenes of the oceanic floor basalts.

Therefore, a mechanical application of any of the discrimination diagrams without taking account of the given geological position, would probably be a failure. The of the Veľká Fatra Mts. (Hovorka & Spišiak l.c.) are projections of the clinopyroxenes from the Mesozoic basalts - distinctly different and located exclusively within the field of increased alkalinity rocks (field P). Recently, Letterier et al. (1982) attempted to statistically classify the Cpx from basalts of various provenances. In the classification diagram (Fig. 4), the projection points of clinopyroxene analyses appear in the field of calc-alkaline basalts Cpx. Such position would comply with the geochemical classification of basalts from which the pyroxenes in question were derived.

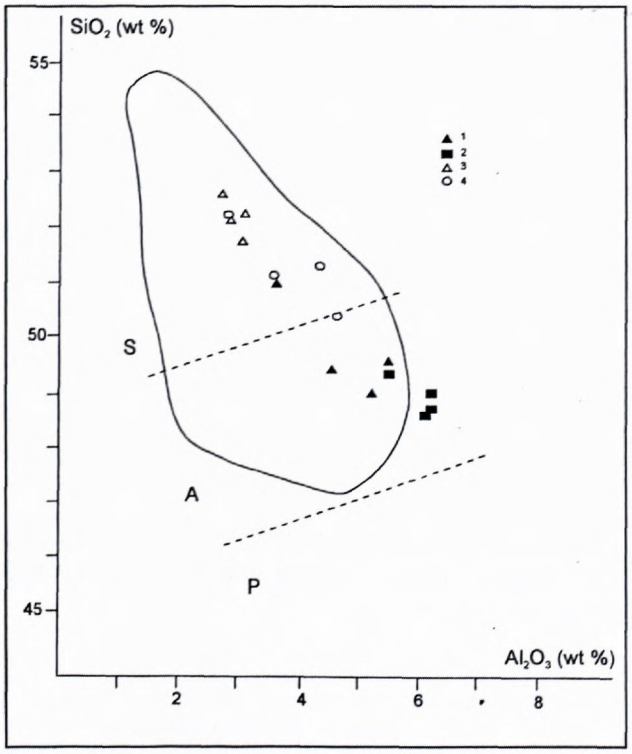



*Fig. 3. Diagram $SiO_2$ vs $Al_2O_3$ in pyroxenes; boundaries according to Le Bas (1962): S - clinopyroxenes from low alkalinity rocks, A - clinopyroxenes from normal alkalinity rocks, P - clinopyroxenes from increased alkalinity rocks, plotted field - clinopyroxenes in oceanic floor basalts. 1- 4 analyses of clinopyroxenes from Table 1.*

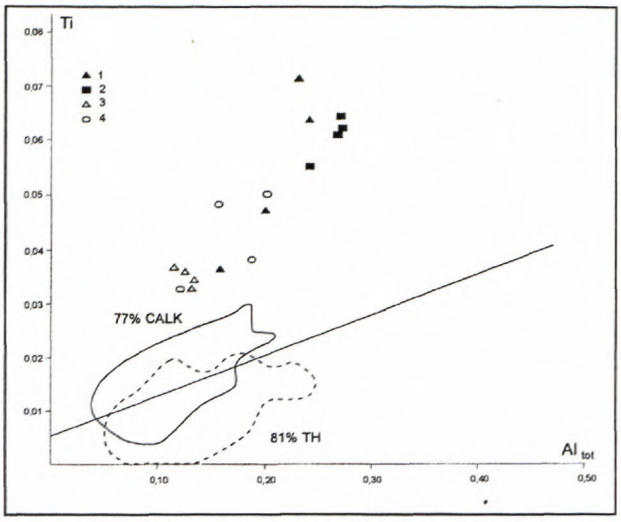



*Fig. 4 Discrimination diagram of clinopyroxenes after (Letterier et al. 1982). TH -field of tholeiitic basalt clinopyroxenes, CALK - field of calc-alkaline basalt clinopyroxenes. 1- 4 analyses of clinopyroxenes from Table 1*

Table 3 Chemical analyses of rocks

| | Malá Fatra | | | | |
|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 |
| $SiO_2$ | 44,46 | 45,22 | 45,91 | 48,46 | 50,47 |
| $TiO_2$ | 1,60 | 2,00 | 2,60 | 2,50 | 1,60 |
| $Al_2O_3$ | 15,30 | 16,13 | 13,62 | 15,71 | 14,92 |
| $Fe_2O_3$ | 4,29 | 3,28 | 2,71 | 2,31 | 1,92 |
| FeO | 7,68 | 7,39 | 6,20 | 5,73 | 6,72 |
| MnO | 0,16 | 0,20 | 0,15 | 0,15 | 0,14 |
| MgO | 5,86 | 4,07 | 5,70 | 5,37 | 5,25 |
| CaO | 9,15 | 8,01 | 8,89 | 8,04 | 7,86 |
| $Na_2O_3$ | 2,00 | 4,15 | 3,46 | 3,59 | 3,93 |
| $K_2O$ | 0,80 | 2,05 | 1,70 | 1,85 | 1,66 |
| $P_2O_5$ | 0,43 | 0,36 | 0,47 | 0,38 | 0,28 |
| $H_2O^+$ | 3,56 | 6,36 | 8,37 | 5,31 | 4,98 |
| $H_2O^-$ | 0,12 | 0,10 | 0,12 | 0,13 | 0,13 |
| $CO_2$ | 4,72 | 0,34 | 0,32 | 0,27 | 0,28 |
| spolu | 100,13 | 99,66 | 100,22 | 99,8 | 100,14 |

1 – Ivanov and Kamenický, 1957
2 – ridge Kalužná, Hovorka, 1967
3 – N of Sučany, Hovorka, 1967
4 – Dubná Skala near Vrútky, Hovorka, 1967
5 – Dubná Skala near Vrútky, Hovorka, 1967

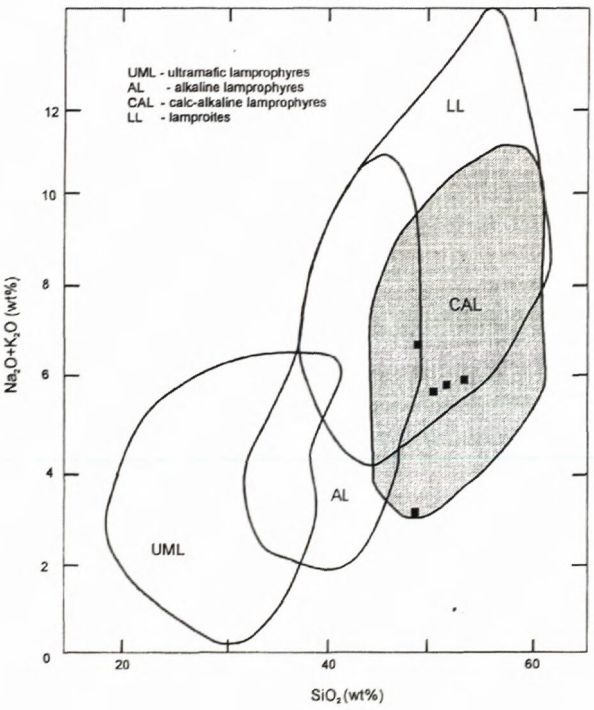



*Fig. 5. ($K_2O+_{NaO}$) vs. $SiO_2$ plot of lampropyres. Field of lamprophyres according to Rock (1987). Analyses of dykes from Table 3.*

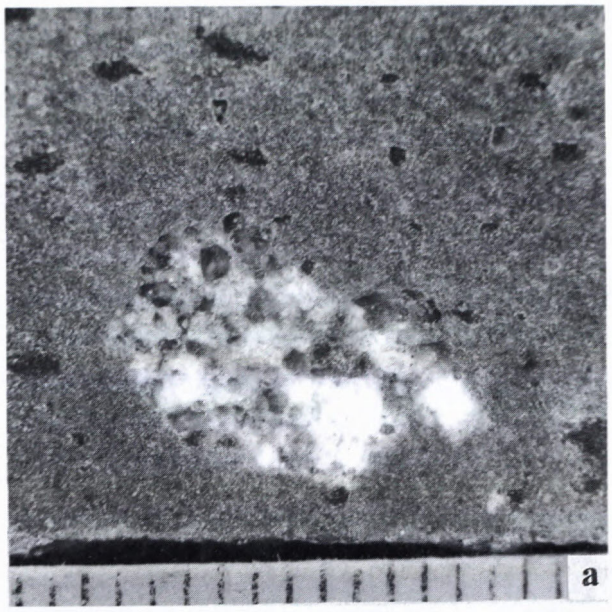
## Geochemical features

In spite of the fact that the basaltic dykes use to be strongly altered, we attempted to determine the character of basaltic magma from the analyses of freshest samples. We used the published analyses (Table 3), recalculated them to fit the waterless form and plotted them in the classification diagram for dyke rocks (Fig. 5, Rock 1987). The projection points for analyses of dyke rocks of the Malá Fatra Mts. lie within the field of calc-alkaline lamprophyres. This classification complies with the character of clinopyroxenes, thus, we can assume that they are calc-alkaline dyke (lamprophyric) rocks.

## Discussion and conclusions

Although, the laboratory investigations were made (Hovorka, 1967) and the results summarized in this paper are available, an essential problem remains how to classify and call the mafic rocks in question. Considering the analyses made after the removal of visible tonalite xeno-liths or their desintegrated parts, undoubtedly, these rocks belong to the mafic group (Table 2). Another point is that in accord with the content of $K_2O$ and $Na_2O/K_2O$ ratio, the analyzed mafics should be ranked with the natrium series. Using the recently valid discrimination criteria (TAS diagram, Le Maitre et al., 1989) they are subalkaline rocks of basaltic and/or trachybasaltic type.

a) Partly desintegrated tonalite xenolith in lamprophyre. Scale = 1 mm. Sample: DS-13/1.
b) Size of irregular chloritic and calcite-chlorite amygdales grows towards the dyke margin (in Fig. upwards). Scale = 1 mm. Sample: DS-23.
c) Incipient melting of tonalite xenolith in lamprophyre. Sample: DS-26/4; magn. 45x, X pol.
d) A plane of recrystallized amygdale with isometric metallic ore minerals; ilmenite platelets predominate in the matrix of lamprophyre. Sample: DS-26/4; magn. 45x, X pol.
e) Subophitic structure of lamprophyre. Sample: DS-18/2; magn. 45x, X pol.
f) Fluidal arrangement of plagioclase laths in lamprophyre. Sample: DS-13/1; magn. 45x, X pol.
g) Development of needle-like alkaline and potassium feldspars at the margins of a plagioclase in the xenolith. Sample DS-22 magn. 45x, X pol.
h) Development of needle-like alkaline and potassium feldspars (radially arranged) at the margins of a plagioclase in the xenolith. Sample DS-22 magn. 45x, X pol All samples come from the Dubná Skala quarry near Vrútky.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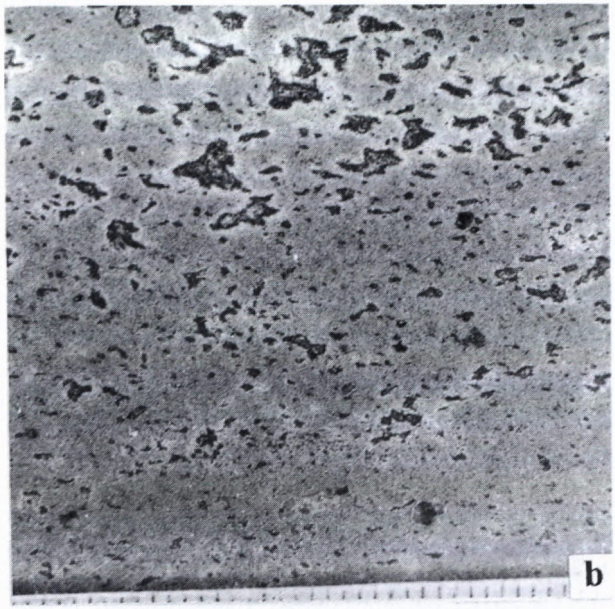
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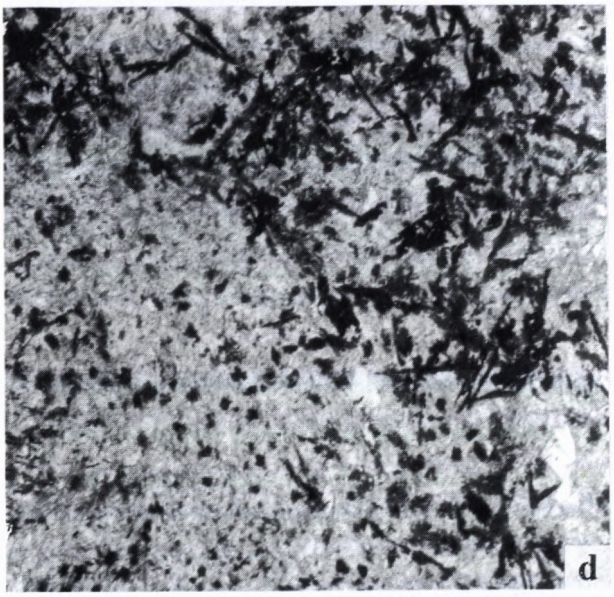
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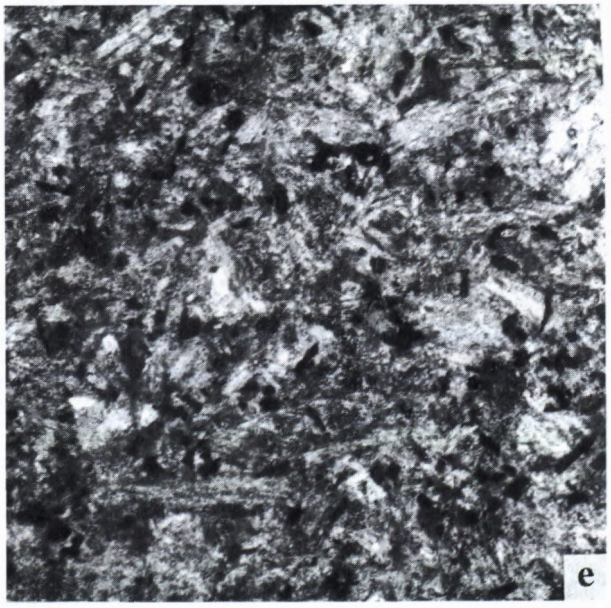
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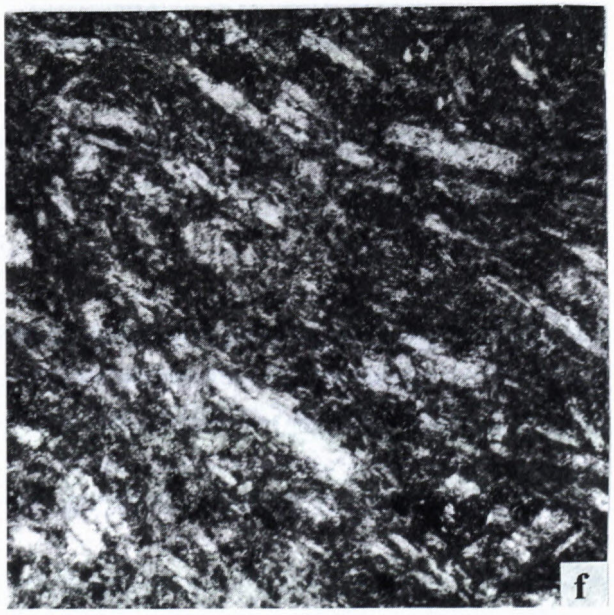
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*According to currently effective classifications (Le Maitre et al. 1989, Rock 1991, Woolley et al. 1996), the lamprophyres of the calc-alkaline group are characterized by porphyric black micas and by amphiboles, or pyroxenes that also occur in the rock matrix. The calc-alkaline porphyries are characterized by the presence of calc-alkaline feldspars (l.c.).*

We assume that broad platelets of Plg I of distinctly magmatic origin are characteristic for only a relatively small part of thin sections of mafic rocks. This is also a characteristic feature of the porphyry, or porphyrite rocks (gabbro porphyrites). However, this finding applies to only a small part of rocks, while the others do not contain porphyric plagioclases, but the porphyric clinopyroxenes instead. Such types (with fine pseudomorphs after olivine) have the composition similar to odinite. However, the international classification recommends not to use this term (Le Maitre 1989). Lack of good exposure of individual dykes of mafic rocks in the mountains in question does not allow to define the spatial relationships between the mafic varieties containing scarce, fine, porphyric plagioclases or porphyric clinopyroxenes, respectively, but it does allow to take the mineral composition of most mafic dykes that occur in the Variscan tonalites of the Malá Fatra Mts. as a criteria to assign them to the group of calc-alkaline lamprophyres. Since the clinopyroxene is its predominant porphyric phase we coin the term pyroxenic spessartite.

The amygdaloidal development of most mafics, with the amygdales measuring 5, but sporadically as much as 8 mm across, indicates an emplacement of the magmatic melt during its consolidation shallow beneath the surface. At the same time, the character of xenoliths indicates that these were incorporated in the mafic melt within the tonalite body, hence, the xenoliths of the lower horizons of the continental type of crust are missing.

Subalkaline character of the mafic melt proper, evidenced not only by its chemical composition, but also by the composition of magmatic clinopyroxenes (Table 1), rules out the existence of any genetic relationship between the mafic dykes and the effusives/extrusives that occur in some Mesozoic units of the Central zone of Western Carpathians, because the latter have a distinctly alkaline character (Hovorka and Spišiak 1988).

Calc-alkaline character of the magma of mafic dykes eliminates their genetic linking with the products of alkaline volcanism in the Mesozoic units of the Western Carpathians (Hovorka-Spišiak 1988, Spišiak - Hovorka 1997). Absence of others but tonalitic types of xenoliths indicates that the magmatic reservoir of mafic rocks was adjacent to magmatic reservoir of tonalites, or to its basal part, respectively. Thus, we can visualize the enclosing tonalites and mafic dykes in them as reverberations of magmatic activity (generation of tonalites) that took place during the Variscan period. In tis process, the mafic dykes filled the contraction joints in the cooling tonalitic body. Therefore, we can consider the tonalites and the mafics in these mountains as consanguinic rocks. To verify these findings, an exact geochronological dating of both, dykes and tonalites, should be made.

*Acknowledgements*

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# Geologic and tectonic evolution of the Turiec depression in the Neogene

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*Abstract*: The Neogene fill of the Turiec Depression represents a halfgraben filled above all in the Middle and Late Miocene. The material transport was from W toward E, the basin axis was situated in the NE-SW direction. The sediment distribution was determined by altered palaeostress field. During the Early Miocene a compressional component was oriented in the NW-SE direction and gradually it rotated into NE-SW direction in the Late Miocene. During the Pliocene a compressional component was again oriented in NW-SE direction. The paper contains lithostratigraphic division of Neogene deposits and we define a new lithostratigraphic unit - Turiec Formation.

*Key words:* Neogene tectonic, geology, sedimentology, Neogene basin, Western Carpathians.

## Introduction

The Turiec Depression is bounded to the west to Kriváň part and to the north to Lúčany part of the Malá Fatra Mts. according to the regional-geologic division (Vass et al. 1988). From the eastern part the boundary separates the depression from the Veľká Fatra Mts. From the south the Turiec Depression is bounded to the Žiar Mts. and volcanics of the Kremnice hills (Fig. 1).

The basic data dealing with Tertiary structure and fill of the Turiec Depression were published in the works of Buday (1957, 1962) who also suggested the first stratigraphic division of the Neogene deposits (Tab. 1). The work of Andrusov (1954) mainly concerned the more detail stratigraphic and tectonic assignment of the Neogene deposits. The tectonics and spatial distribution of the main rock types were mainly analysed in the works of Gašparik (1989, 1995), Gašparik & Halouzka (1989) and Gašparik et al. (1991). The basic works on paleontologic research and stratigraphic assignment of the Neogene deposits was submitted by Němejc (1957, 1957a), Rakús (1958), Pokorný (1960), Sitár (1969, 1976, 1982), Ondrejíčková (1974), Brestenská & Planderová (1979) and Gašparíková (1987).

## Methods and objectives of the work

The field research was based on the published map at the scale 1:50 000 Gašparik & Halouzka (1989). It consisted of the gathering of the basic sedimentological, paleontological and structural data. If the characteristics and the sufficient exposure of rocks allowed, the outcrops were evaluated complexly applying all the methods. The results and data obtained by field research were subsequently processed and synthetized in laboratories.

The sedimentological study was done on the suitable outcrops at all locations where data on deposition character, direction and mode of clastic deposition and depositional environment were missing. At the same time we have taken samples for macropaleontological and micropaleontological study. The structural research was aimed at the gathering and analysis of mainly brittle deformations from which a tectonic regime for individual periods of the depositional record was intepreted. The methodology of Angelier & Gougel (1979) and Angelier (1979) respectively, were applied for the structural data processing.

## The obtained results and their interpretation

The Turiec Depression represents an intermountain depression filled by the Paleogene, Neogene and Quaternary deposits. In general, the deposits dip toward west (Fig. 1). The uniform dip of the sedimentary sequence point to the long-term activity of faults near the western margin of the depression.

The Turiec Depression, similarly to other Neogene basins in the Western Carpathians, contains older Tertiary rocks in its fill, which represent preserved relics of the original depositional areas. Into this category we assign Paleogene and Early Miocene deposits cropping out on the eastern and southeastern margin of the depression as well as in the basement of the Miocene basin fill (Gašparik & Halouzka 1993, Gašparik et al 1995).

The Paleogene deposits are exclusively exposed on the eastern and northeastern margin of the basin and they are assigned to the Subtatric Group (Gross et al.

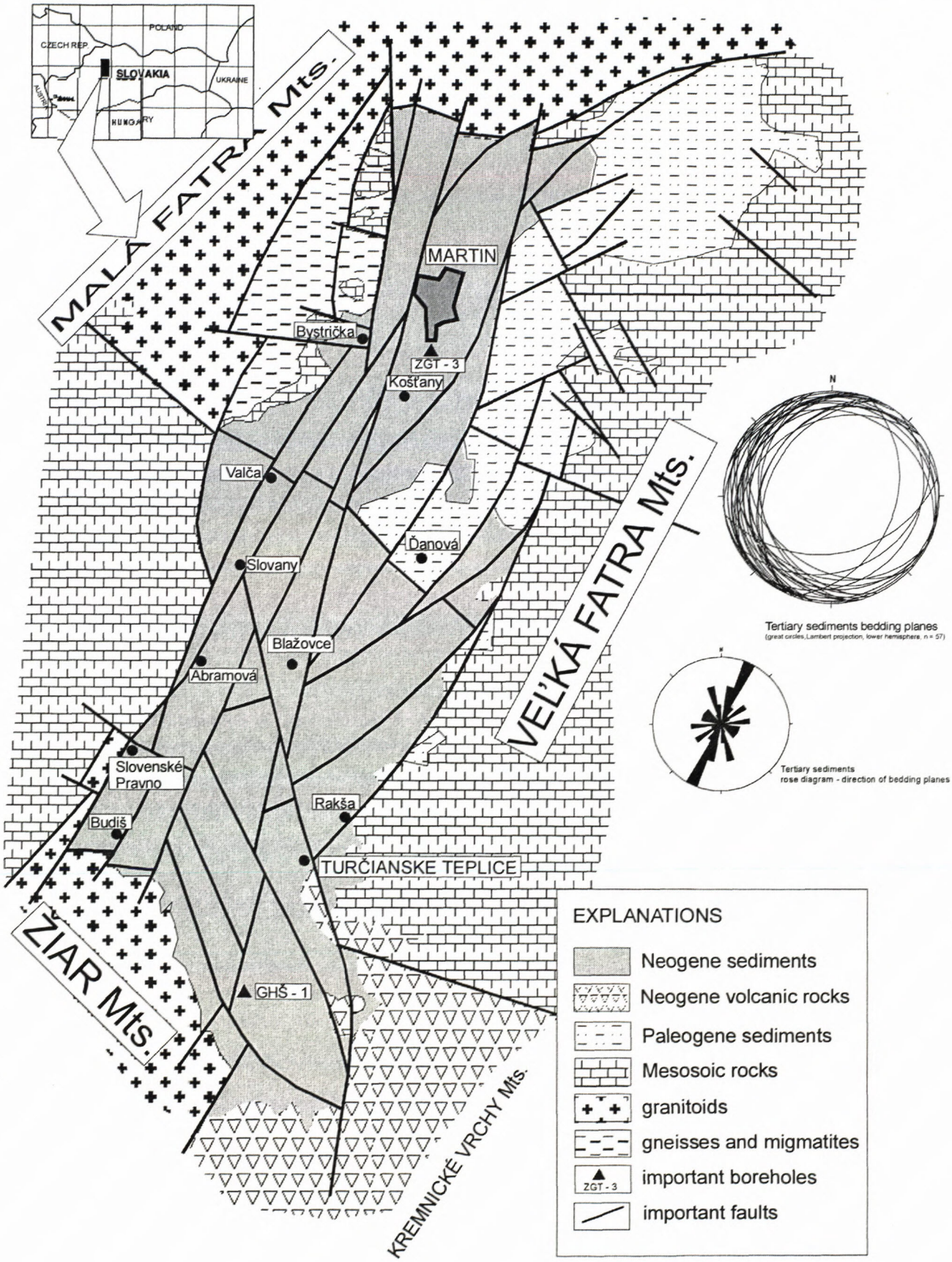



*Fig. 1 Schematic geological map of the Turčianska kotlina depression (compiled by Hók, 1997, after Maheľ et al., 1964 and Gašparik and Halouzka, 1989)*

| | | | | | | |
|---|---|---|---|---|---|---|
| NEOGENE | PLIOCENE | Dacian | Diviaky member | clays, andesites gravels (50 - 60m) | freshwater facies | last movements along west situated marginal fault<br>3rd transgression |
| | | Pontian | | missing | | |
| | MIOCENE | Pannonian | | missing | | |
| | | Sarmatian | Martin member | gravels, clays (50 - 300m) marls,coaly clays | freshwater facies | strong subsidence, shalow water, movements along NW oriented faults |
| | | | | marls, limestones and rare coaly clays (200 - 500m) | brackish facies | strong subsidence, andesite volcanics, rhyolite volcanics |
| | | | | gravels, limestones, volcanic tufits | | 2nd transgression |
| | | Badenian ? | | marls, sands, sandstones | marine facies | 1st transgression |

*Tab. 1 Litthostratigraphic column of the Neogene fill of the Turčianska kotlina depression (compiled by Hók, 1997 – according to Buday, 1962*

1984). According their lithologic content and stratigraphic level they may be divided into the following formations:

- Borové Formation (Lutetian - Early Eocene)

It is a basal formation and it only occurs in the north-eastern part of the Turiec Depression. It contains coarse-grained deposits - breccias, conglomerates, locally also limestones and sandstones. The composition of breccias and conglomerates usually reflects lithologic character of the immediate basement.

- Huty Formation (Priabonian - Late Eocene)

In the greater part of the area it substitutes the basal facies and it lies directly on the Mesozoic basement. The thickness of the formation reaches up to 1 000 m (the borehole ZGT-3, Fendek et al. 1990). The Huty Formation is the main lithofacial member of the Subtatric Group in the Turiec Depression. Lithologically it consists of clays and claystones predominating sandstones.

- Zuberec Formation (Early to Late Oligocene)

In the Turiec Depression it is only developed in the denudation relics in the surroundings of Daňová. Lithologically it represents deposits of turbidity flows - alternation of claystones, sandstones and sandy shales.

The tectonic deformation of the Paleogene bed succession (locality Daňová) by folding with the fold axis ENE - WSW and vergency to SE is consistent with the deformations observed in the area of Malé and Brezovské Karpaty Mts. or in the area of Zázrivá (Kováč et al. 1989, 1993, Marko et al. 1995) and they are comparable with deformations observed in the Rajec Depression (locality Kľače). The structures of the folded Late Eocene and Early Oligocene deposits in the depression show general palaecompression direction NW-SE. The deposits of the Krížna nappe are deformed by a similar compression in the Turiec Depression (Dolné Jaseno). It is necessary to note that thrusts indicate displacement toward NW. However, it does not refuse the NW-SE orientation of the palaestress compressional component. At the same time the compression is younger and it has different characteristics as the compression responding the displacement of the Krížna nappe. The deformation of the Paleogene rocks probably originated during the desintegration of the Paleogene depositional area after the Oligocene and before the deposition of the Early Miocene clastics similarly to the area of the NW margin of the Malé Karpaty Mts. (Kováč et al. 1989, Marko et al. 1995).

The compressional tectonic regime of the active Central West Carpathian front having palaeostress field characterized by compression of NW-SE direction controlled origin and palaeogeographic distribution of sedimentary areas during the Early Miocene. The denudation relic of their fill is Rakša Formation of Eggenburgian age in the Turiec Depression. It consists of basal carbonate lithofacies containing marine macro- and microfauna (Gašparik et al. 1995).

The uplift of the Žiar Mts. is documented from 46 ± 5 to 52 ± 7 Ma by FT ages of apatites (Kováč et al. 1994). The uplift of the Žiar Mts. occurred on NW-SE oriented faults as it is well documented by distribution of the sedimentary sequence in the Horná Nitra depression (Hók et al. 1995). We assume sedimentation in the southernmost part of the depression (Fig. 2). The occurrence of the Early Miocene deposits in this part of the depression is undirectly indicated by redeposited microfossils of the Early Miocene in the borehole GHŠ-1 (Gašparik et al. 1995). The orientation of the Early Miocene sedimentary environment axis probably was in NW-SE direction as it was in the Horná Nitra depression (Hók et al. 1995).

The activity of the NW-SE oriented faults was also connected with the uplift of crystalline complex of the Žiar Mts. Mostly coarse-grained, micaceous, quartzite sandstones (arkoses) and fine-grained conglomerates of Early? to Middle Badenian Budiš Formation (Gašparik et al. 1995) or Budiš Member was deposited on the foothill of this mountain. The deposits represent gravity sediments on the mountain slope or alluvial fans respectively, which is well documented by the outcrop in the area of Rudno composed of sheet flow deposits.

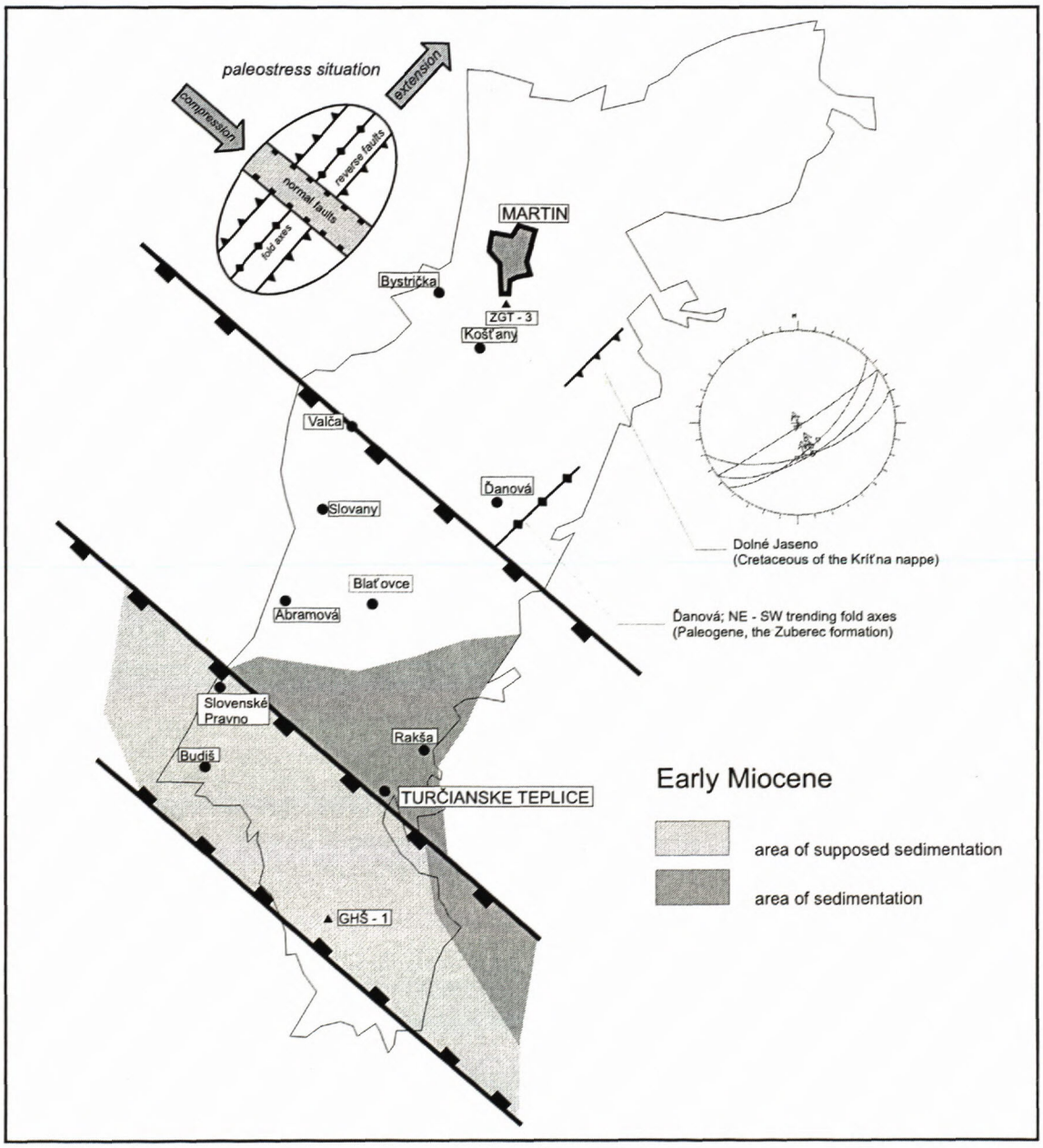



*Fig. 2 Supposed paleogeographic and paleotectonic situation in Early Miocene*

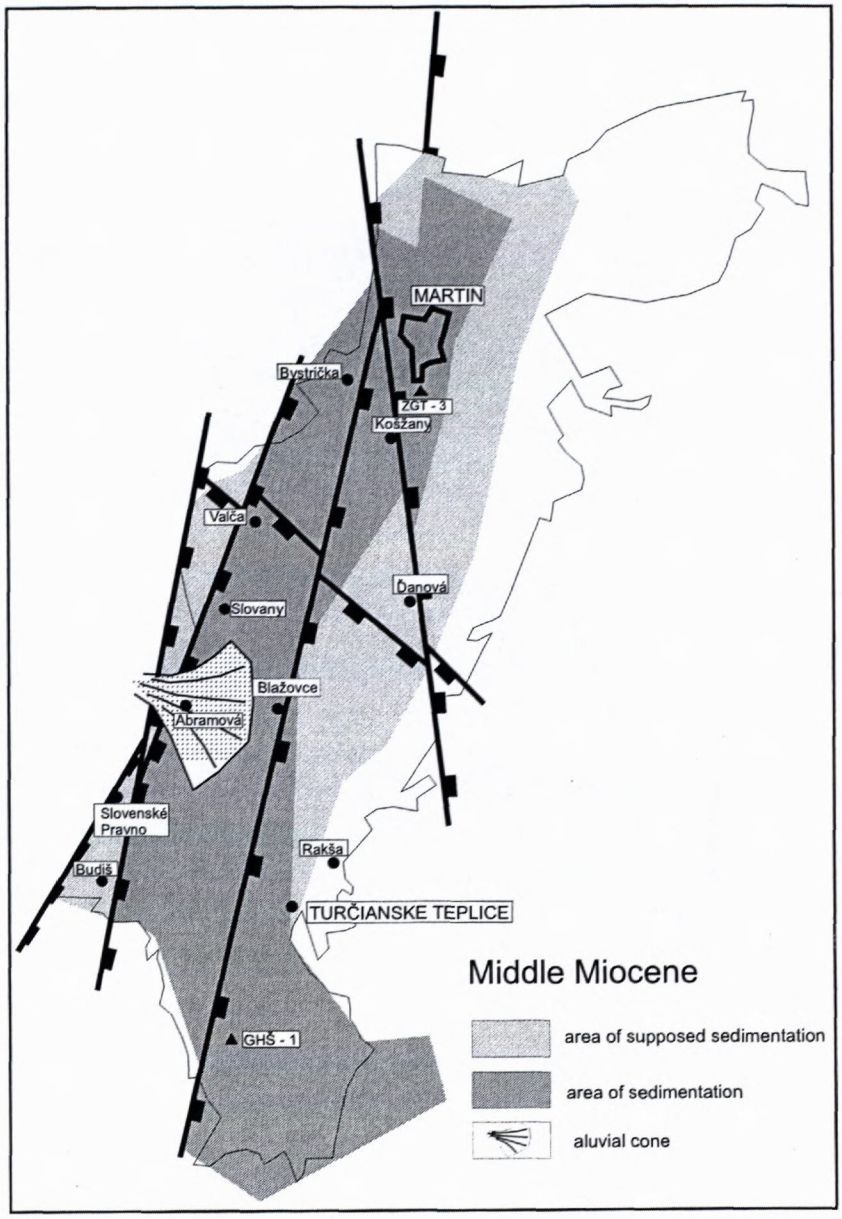



*Fig. 3 Supposed paleogeographic situation in Middle Miocene*

The orientation of the palaeostress was principally changed in the Middle Miocene. It determined the configuration of the sedimentary area from the NW-SE to N-S and NE-SW direction. The abundant findings of the endemic fauna (Rakús 1958, Pokorný 1960) point to the separation of the Turiec Depression from other intermountain depressions of the Western Carpathians in that time.

The continuation of the Žiar Mts. uplift as well as rejuvenation of subsiding NE-SW faults on the western margin of the Turiec Depression was determined by mantle diapirism accompanied by volcanic activity in the middle Slovakian area (Nemčok & Lexa 1990).

Volcano-sedimentary formations (Turčok, Kremnický Štít and Rematy) reflect the volcanic activity in the southernmost part of the depression in the Late Badenian to the Early Sarmatian age. We assign coarse-grained marginal development of the Abramov Member (Gašparik et al. 1995) and correlable fine-grained basin facies studied at Kolísky, Abramová, Trebostovo, Ležiachov, Valča, Moškovce outcrops into this lowermost etage.

Abramov Member consists of coarse grained deposits of debris flows deposited as alluvial fans on the margin of uplifted mountain (Fig. 3). The marginal, coarse-grained development is toward the basin interfingered by fine-grained lacustrine deposits containing fossil remnants of plants. The plant association is represented by numerous termophile elements of species *Lauraceae – Daphnogene polymorpha* and *Ficus cf. Lanceolata, Sequoia langsdorfii, Celastrus cassinefolius.* Species *Equisetum parlatorii, glyptostrobus europaeus, Taxodium dubium, Salix tenera, Quercus pseudocastanea, Castane atavia, Acer tricuspidatum, A. pseudomonspessulanum, A. integerrimum, liquidambar europaea, Betula prisca, Alnus ducalis, species of genus Carpinus, Carya denticulata, Ulmus longifolia, Zelkova zelkovaefolia* are common forms in the Slovakian Late Badenian and Sarmatian (Sitár 1982). Palaeotrope elements, especially *Echnatisporites miocaenicus, Leiotriletes pseudomaximus, Lygodiosporites sp., Platycaryapolllenites miocaenicus, Symplocoipollenites sp., Engelhardtia type, Sapotaceoidaepollenites sapotoides, Quercoidites henrici, Q. microhenrici, Myricipites bituitus* dominated in the palynospectrum. Occasionally representants of the arctotertiary geoflora, especially pines of *Pinus* genus and genus of the species *Taxodiaceae*, occurred. The pollen analysis of deposits indicated extremely warm, humid subtropic climate. Fauna is represented by a group of ostracoda suggesting fresh water environment of swamps and small lakes in the depression area.

The coarse-grained development is expressed by high-density gravity flow mechanism in subaerial and subaquatic environment. The fine-grained lacustrine deposits does not show traction current features in the basin (for example cross bedding), on the contrary some indices suggest deposition from turbulent suspensions (pebbly mudstone, grading). They consists of tuffs, siltstones and fine-grained sandstones with tuff admixture (Kolísky, Abramová, Trebostovo and Ležiachov), passing upward into siltstones and fine-grained dolomitic sandstones (Valča and Moškovec).

The alluvial fan bodies consist of fine-, middle- and coarse-grained conglomerates (locally of block size) transported on a short distance which are poorly sorted

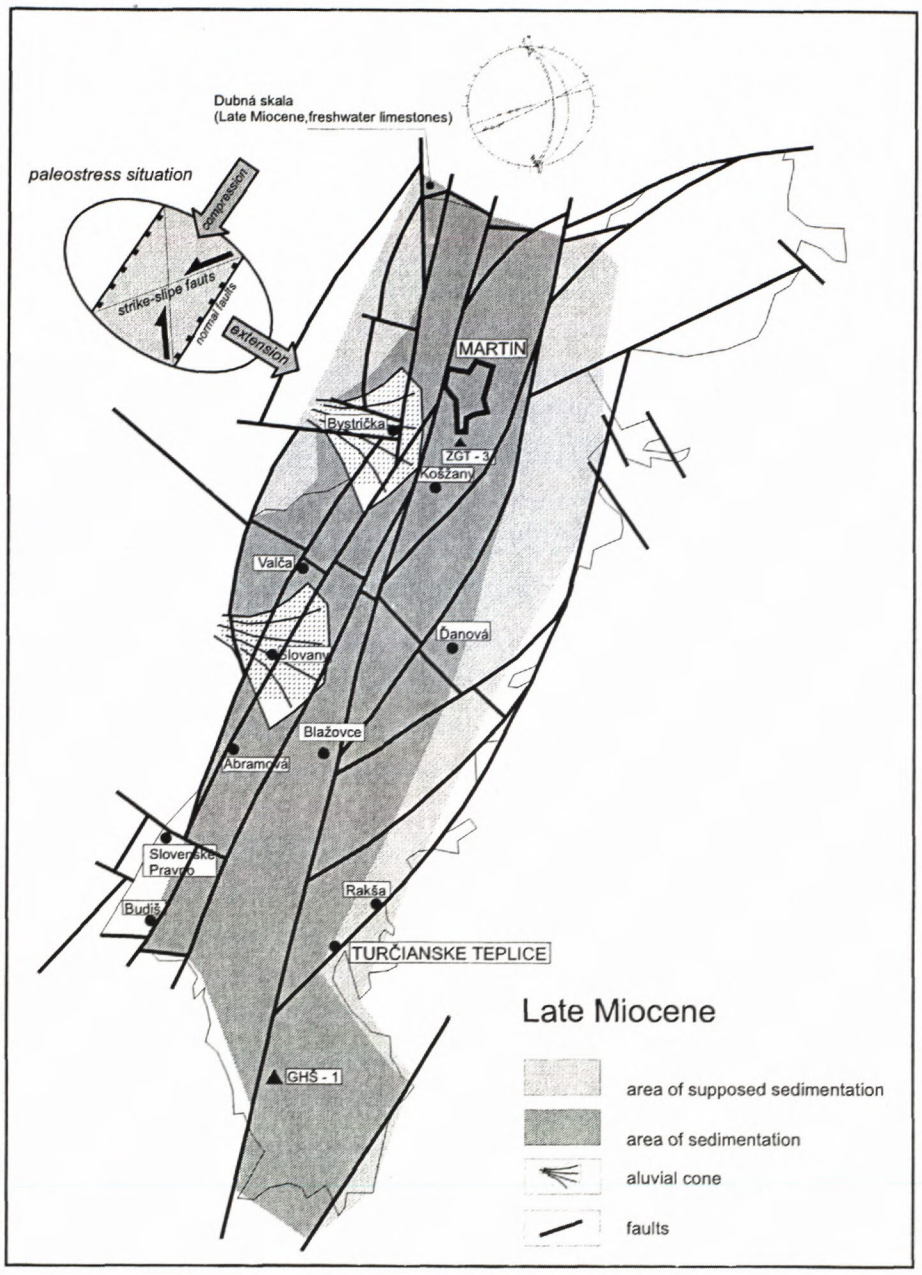



*Fig. 4 Supposed paleogeographic and paleotectonic situation in Late Miocene*

and weakly rounded. The occurrence of subangular clasts, mainly dolomites, is very frequent. The conglomerate clasts are mainly composed of Triassic sequences of Choč nappe (limestones and dolomites). The matrix is carbonaceous and consists of sand or clayey sand and clay. The coarse-grained development varies in a wide range from matrix-supported to clast-supported conglomerates. The bedding is only emphasized on boundaries with great grain-size difference (siltstones/conglomerates), otherwise it is faint and in most cases the beds are amalgamated. The conglomerate structure is mostly chaotic with local faint inverse grading, deposits of pebble ab toward west, normal grading. The large-scale trough cross-bedding was also observed (Valča and Moškovec).

The clast grain size decreases from the west toward the east, e.g. from the basin margin toward its axial part.

The general strata dip to the west as well as the increasing of the marginal clastics thickness in time suggests synsedimentary as well as postsedimentary activity of N-S to NNE-SSW normal slip faults (fig. 3). Based on analogues from other Neogene basins, particularly of neighbouring Horná Nitra Depression (Hók et al. 1995) we assume the orientation of the compressional palaeostress axis in N-S to NE-SW direction.

Abramov Member is partly overlain by deposits of so called Slovany Member, which Eggenburgian age (Gašparik et al. 1995) seems to be not substantiated on the basis of the field observation. The Slovany Member overlies Abramov Member at localities Slovany and Moškovec. The Eggenburgian deposits are paleontologically only proved from the surroundings of Rakša. The mudstone type, occasional imbrication with plane dip deposits of the Rakša Formation lie on the Middle Triassic Dolomites or on the basal Paleogene (Gašparik 1989).

The Slovany Member consists of carbonate gravelly conglomerates, coarse-grained gravels with interbeds of breccias conglomerates and sandstones. Clays and claystones comprise 20 – 30 cm thick beds in the conglomerates. The matrix consisting of dolomites, illite with kaolinite, chlorite, quartz ± calcite admixture is sandy-clayey, rusty brown. A layer of brownish-red shales of bauxit composition is described from the borehole GT-12 (Gašparik 1989) showing terrestrial environment. The immediate basement of a part of Slovany Member consists of crystalline complex of the Lúčany part of the Malá Fatra Mts. The fact that the deposits lie directly on the crystalline complex and the different lithologic composition of the Slovany Member evokes an substantiated assumption that they have originated after the Malá Fatra uplift and were transported on a short distance in subaerial environment deposited directly on the exposed crystalline basement. The Lúčany Malá Fatra Mts. uplift is according to FT apatite ages 28 ± 15 Ma (Kováč et al. 1994). Slovany Member as well as underlying Abramov Member (Valča and Moškovec) represent subaerial and subaquatic deposits of an alluvial fan.

The body of marginal clastics of the Slovany Member consists of sandstones and fine-, medium- and coarse-grained conglomerates (locally even blocky conglomerates) transported for a short distance by a dense, water-saturated gravity flow mechanism. The conglomerates are poorly sorted, the clasts, consisting of dolomites and Triassic limestone, are often subangular. The sandstones are coarse- and medium-grained, the sandstone beds with floating clasts in sandy matrix occur (Slovany). The carbonate material of clasts is mainly derived from th Triassic sequences of the Choč nappe. The matrix is sandy or clayey-sandy and clayey, locally it is rusty to red coloured (Trebostovo). The coarse-grained development varies in a wide range from matrix- to clast-supported conglomerates. The beding is emphasized on boundaries between different grain size (sandstone/conglomerates). The conglomerates are mostly chaotic, locally faint imbrication with a general ab plain dip toward west occurs. The faint trough-cross bedding was observed. The general dip of beds is again toward the west.

Slovany Member, represented by a coarse-grained marginal facies passes into finer-grained sandy to silty-clayey and clayey development of the upper part of the Late Sarmatian – Pannonian Martin Member (Trebostovo to Hrádok nad Bystričkou, outcrops along tank drome). The marginal coarse-clastic body of Bystrička Member was deposited above the mentioned deposits (Gašparik et al. 1995).

Bystrička Member consists of coarse-grained deposits of subaerial and subaquatic gravity flows (debris flows) deposited as alluvial fans. Marginal, coarse-grained development is in the today's erosive cut occuring above the basinal development of Martin Member. We can not exclude its basinward interfingering with fine-grained lacustrine deposits containing fossil remnants of Pannonian plants and molusks. We assume its Late Sarmatian to Pannonian age based on the underlying sandy-clayey beds of the rusty-brown colour containing prints of arctotertiary leaf mainly *Carpinus grandis, Betula prisca*, more species of genus *Quercus, Fagus, Castanea, Parrotia fagifolia, Zelkova zelkovaefolia, Acer tricuspidatum, Platanus platanifolia.* Even if the mentioned assemblage contains Badenian-Sarmatian elements (*Parrotia, Fagus, Quercus, Acer*), The Late Sarmatian and Pannonian forms (*Carpinus, Betula, Ulmus*) prevail. Probably it is an equivalent of the lower horizon of Martin brick yard (Sitár 1969).

The alluvial fan body consists of fine-, medium- and coarse-grained conglomerates (locally blocky conglomerates) and poorly sorted and weakly rounded sandstones transported on a short distance. The clast grain size decreases from the west toward the east e.g. from the basin margin to its axial part. The studied fanglomerates often contain subangular clasts of Krížna nappe rocks (mainly radiolarites and fleckenmergels). The conglomerate composition is mainly derived from carbonates (dolomites, limestones, fleckenmergel) and radiolarites of Krížna nappe. Conglomerates and sandstones probably of the Paleogene age occur relatively frequently.

The matrix of conglomerate is carbonaceous, clayey-sandy and clayey. The coarse-grained development varies in a wide range from matrix- to clast-supported conglomerates. The conglomerates are mainly chaotic, nearby the elevation point Hrádok nad Bystričkou pebbly mudstones occur. The bedding is unconspicuous, mostly amalgamation of more beds occurs.

Comparison of the clast composition of up till now described alluvial fan bodies located on the western margin of the Turiec Depression suggest a conclusion that the body of Bystrička Member was deposited from rivers which erosive base reached the deeper level of the thrust structure of the Central Western Carpathians and the area of the preserved relics of the Paleogene deposits.

The basinal development of the Turiec deposits consists of grey calcareous clays alternating with siltstones, fine- to coarse-grained sandstone and fine-grained conglomerate beds. The beds containing increased amount of the coalificated plant detritus, coal beds and fresh water limestones occur. The beds are rich on endemic fauna of molluscs and ostracods. The occurrence of swamp plant species (*glyptostrobus europaeus, Potamogeton martinianus, Nelumbium protospeciosum*), fossil roots in grew position as well as fresh water molluscs point to relatively shallow, fertile lacustrine depositional environment.

The grey calcareous clays with tuff beds are considered as the oldest deposits of the bed succession of Martin Member. They may be correlated with the marginal coarse-grained development of the Abramov Member of the Late Badenian – Early Sarmatian age containing rich microflora with dominant palaeotrope component represented by *Trilites multivallatus, Castaneoideaepollis sp.* and representants of swamp vegetation of swamp *Myricaceae, Nysaceae* and *Taxodiaceae* when the representants of the family *Taxodiaceae* grew directly in swamps. A warm subtropic climate dominated during the deposition of the lower etage deposits from the bed succession of the Martin brick yard.

The overlying grey calcareous clays alternating with siltsone, sandstone and fine-grained carbonate conglomerates, coal beds and endemic fauna of molluscs (Martin brick yard) represent a bed succession of the Late Sarmatian to Middle Pannonian. They are equivalent of the upper part of the coarse-grained Abramov and Slovany Members. The relative rise of lake level in that time is documented by a transgressive character of Martin Formation clays inspite of a fast subsidence of individual basin depocenters. A gentle increase of arctotertiary elements (*Ulmipollenites undulosus, Untratriporopollenites instructus, Tsugaepollenites sp., Cedripites sp., Betulaepollenites betuloides, Alnipollenites verus, Carpinipites carpinoides, Chenopodipollis sp., Graminidites sp.* in the palynospectrum of the central part of the bed succession suggests gentle cooling, besides the climate seems to be gently more arid – (*Chenopodiaceae, Gramineae*). The climatic change was also reflected in the vegetation character which composition has more afinity to the mountain type of vegetation (*Tsuga, Cedrus, Sequoia*).

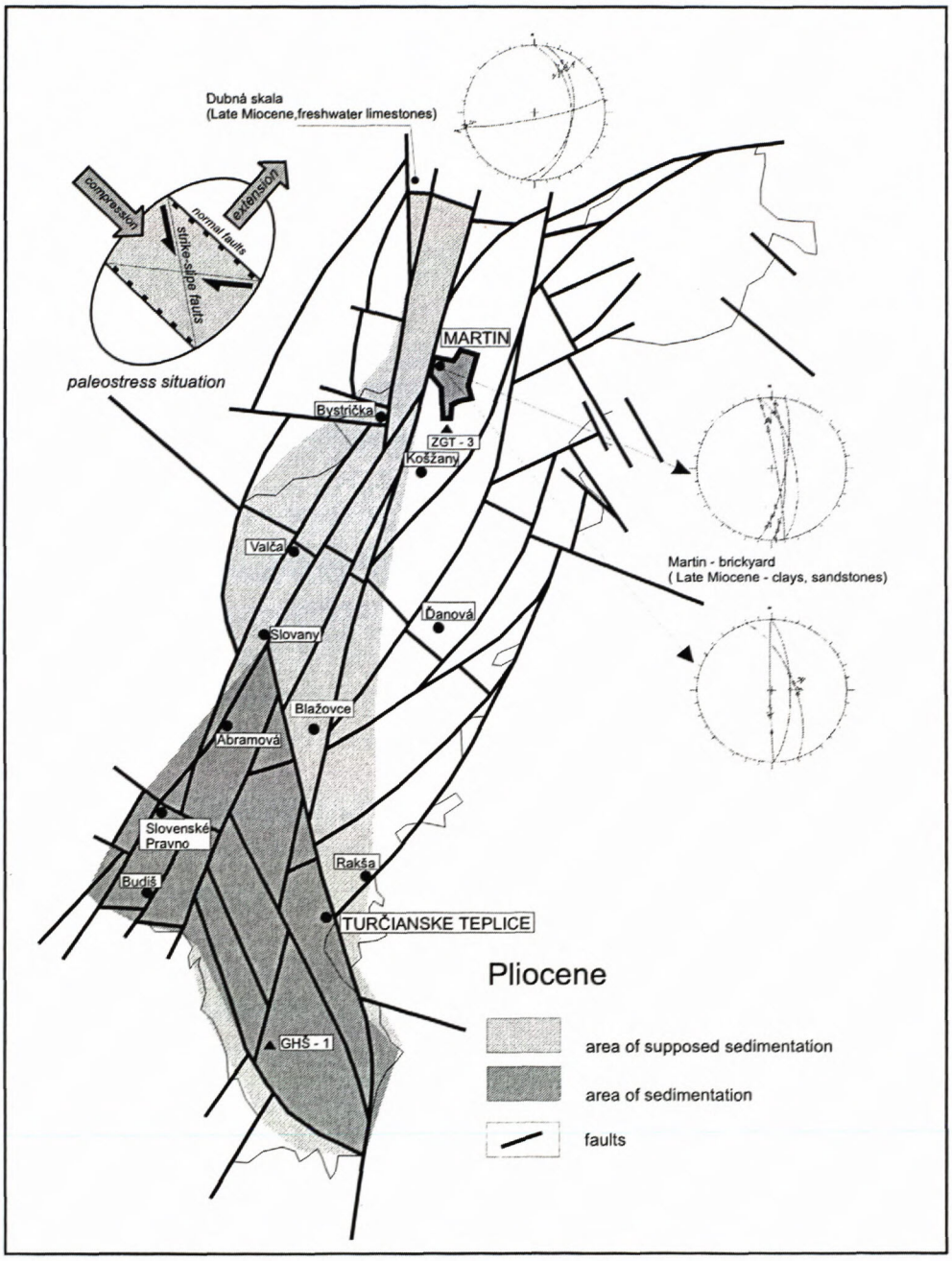



*Fig. 5 Supposed paleogeographic and paleotectonic situation in Pliocene*

The grey, calcareous clays with siltstone beds and fine-grained sandstones containing rich fauna of molluscs characterized by an occurrence of small congeria form are considered as the youngest, Late Pannonian deposits. The Pravno Member (Andrusov 1954) crop out in the southern part of the Turiec Depression (Slovenské Pravno) underlying fresh-water limestones with fresh-water and subaerial fauna of gastropods and lamellibranchata (Lúcky mlyn).

The body of fresh-water limestones, occurring in the area of Dubná skala in the northern part of the today's depression, is composed of thick beds of limestones and sandy limestones containing fauna and flora. Fossil remnants of organism suggest lacustrine depositional environment in the moderate zone of greenwood at the end of the Miocene. The carbonate conglomerates in the upper part of the bed succession document local source of clastics (clasts composed of Permian rocks). The clastics were not transported on the larger distance. Analogous to the Vienna and Danube Basins we assign the deposition of the greater fresh-water limestone bodies of Pravno Member to Pontian and Pannonian H, respectively.

From the Sarmatian to Pontian the stress field was stabilized and compressional component was generally oriented in the NE-SW direction enabling maximum extension in the NW-SE direction and continual subsidence of depositional areas (Fig. 4). The maximum thickness of the Neogene deposits occur in the immediate surroundings of Martin as it is well documented by borehole ZGT-3 and

geophysical data (Fendek et al. 1990, Panáček et al. 1991).

The peneplain development indicates a period of tectonic quiescence at the end of the Pannonian and in the Pontian. The peneplain relics are preserved in the southern part of the Žiar Mts. and in the western part of Kremnica Mts. (Mazúr 1964, Činčura 1969).

The fluvial deposits located in the central part of the Turiec Depression (Socovce – Stráža, Blážovce) are assigned to the Blážov Formation of the Pliocene age (Gašparik et al. 1995). The body of fluvial gravels and sands overlies by a sharp, probably erosive contact the underlying clays of Martin Formation (Socovce – Stráža).

The coarse-, medium- and fine-grained conglomerates mainly consists of carbonate clasts with various degree of roundness. They are relatively poorly sorted. The individual beds are 0,5 – 5 m thick, aften amalgamated . The bedding is emphasized by various grain-size or alternation of conglomerate, sandstone and siltstone beds. The sandstones and siltstones comprises 0,5 – 2,0 m thick beds, locally horizontal lamination occurs. The deposition by water current is documented by sporadic cross bedding and trough bedding, grading and clast imbrications. Based on the channel orientation and ab plain dip direction we assume deposition in an environment of smaller braided river flowing from the south toward north and northeast along the axial part of the basin (palaeo Turiec?).

After the Pannonian and more probably after the Pontian the orientation of the palaeostress changed in a substantial way when the compressional axis changed orientation from the NE-SW to the NW-SE direction. The changed palaeostress regime influenced clastic deposition mainly in the southern part of the Turiec Depression (Fig. 5). The depositional area axis of the Plio-Pleistocene deposits was oriented in the NW-SE direction. The Plio-Pleistocene deposits (clays overlying andesite gravels and sandstones) reach a thickness of 50 – 60 m and they are assigned to the Diviaky Formation (Buday 1962).

The Quaternary deposits were observed at the locality Ležiachov, Hrádok nad Bystričkou – creek bluff and Martin – part Močiar).

The locality Ležiachov represents Plio-Pleistocene deposits of alluvial fans deposited from smaller rivers emerging from valleys at the margin of the Turiec Depression from SW to NE. The carbonate conglomerates and sandstones containing beds of silt with clayey admixture pass upward into palaoesoil horizons. Their position in relation to the flood plain of Turiec river suggests valley incision in the Pleistocene. The locality Hrádok nad Bystričkou – creek bluff documents similar Plio-Pleistocene deposits of an alluvial fan deposited by a smaller river emerging from a valley at the margin of the Turiec Depression toward its axial part. It is possible to observe alternation of micaceous coarse- and medium-grained sandstones with coarse-, medium- and fine-grained conglomerates at the outcrops. The clasts are well rounded and sorted, grading and imbrication of ab clast plain occur. Besides the prevailing carbonates

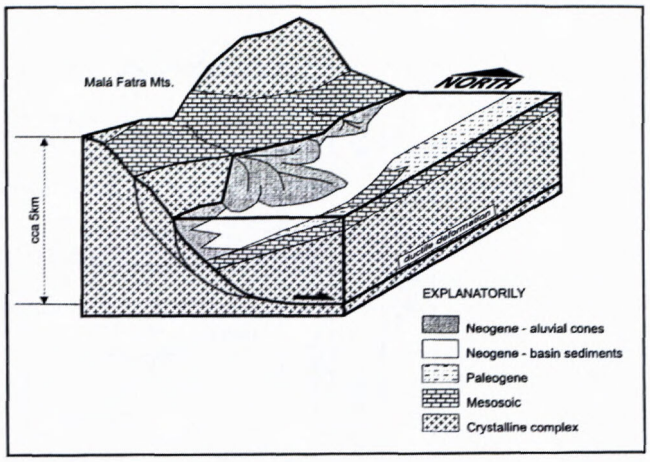



*Fig. 6 The model of tectonic evolution – Late Miocene*

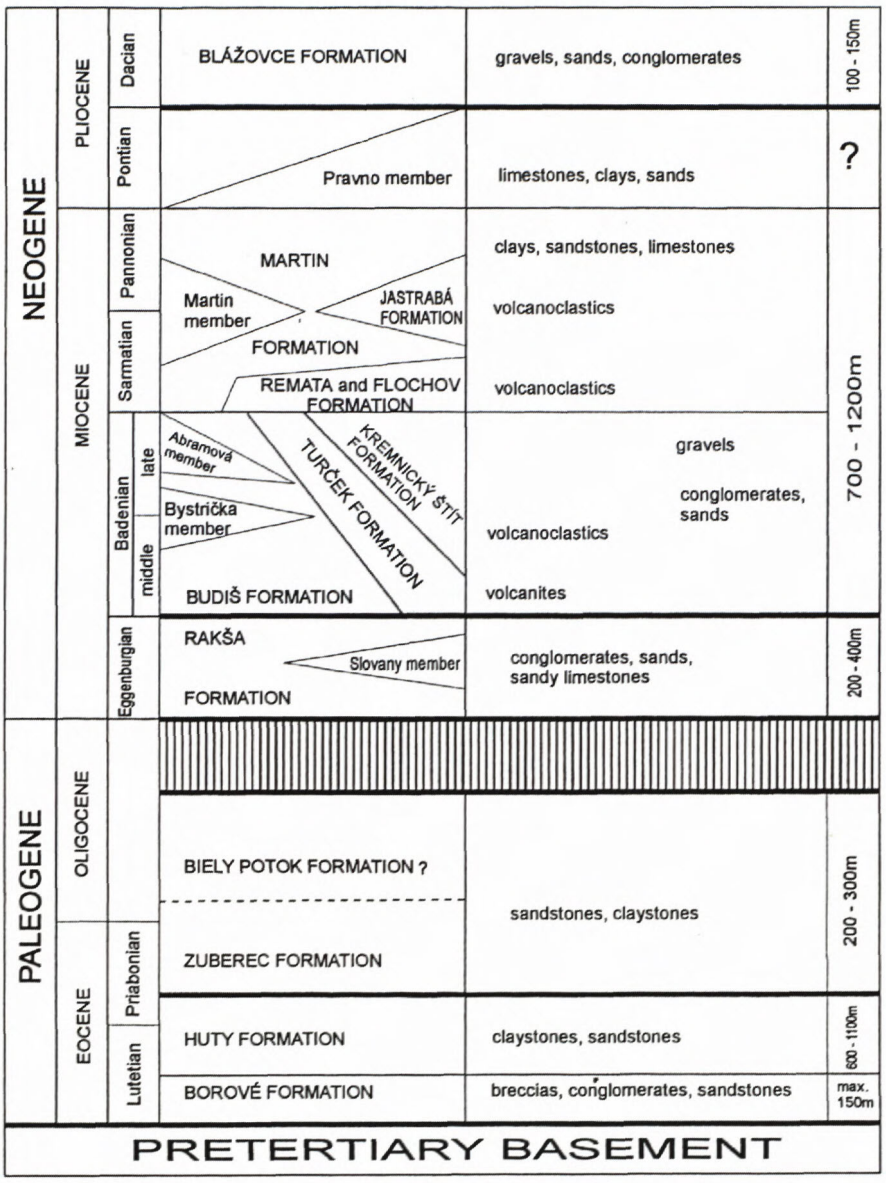
also crystalline clasts (granites 5% to 14%) occur. The locality documents Pliocene uplift of the Turiec Depression margins and erosive incision.

The locality Martin – Močiar represents the youngest – Holocene etape of the depression development. The blocky, coarse- to medium-grained conglomerates with sandy matrix predominantly consists of well rounded clasts of granites. The deposits of scree to alluvial fan are exposed at the root area in the fault zone of NW-SE direction. Similarly to the previous localities it documents a tectonic activity related to the uplift of the mountain margin in the palaeostress regime with compression in the NW-SE direction.

## Discussion and conclusion

On the basis of the results obtained we can state that after the deposition of the Paleogene rocks the Eggenburgian deposits and later probably also Early Badenian clastics were deposited. They were deposited in the depositional environment restricted by NW-SE trending faults. In contrast to Gašparik et al. (1995) we do not consider Slovany Member (Tab. II) as a part of Rakša Formation (Tab. III).

We assume that from the Badenian the Turiec Depression was an isolated sedimentary basin in the framework of intermountain Neogene depressions of the Western Carpathians. This assumption is confirmed by endemic fauna found at more localities (Rakús 1958). The main reason of its isolation was the uplift of Žiar Mts. in the south, dated on the basis of FT apatite ages to 46 ± 5 to 52 ± 7 Ma (Kováč et al. 1994), and an intense neovolcanic activity which products are preserved mainly in the southern part of the depression (borehole GHŠ-1). The main phase of subsidence is represented by Turiec Formation deposits which were deposited in a basin with the axis in NNE-SSW direction. The sediment input into the basin occurred from the west toward the east and deposits comprise Abramov, Slovany and Bystrička Members.




*Tab. II Lithostratigraphic column of the Tertiary fill of the Turčianska kotlina depression*

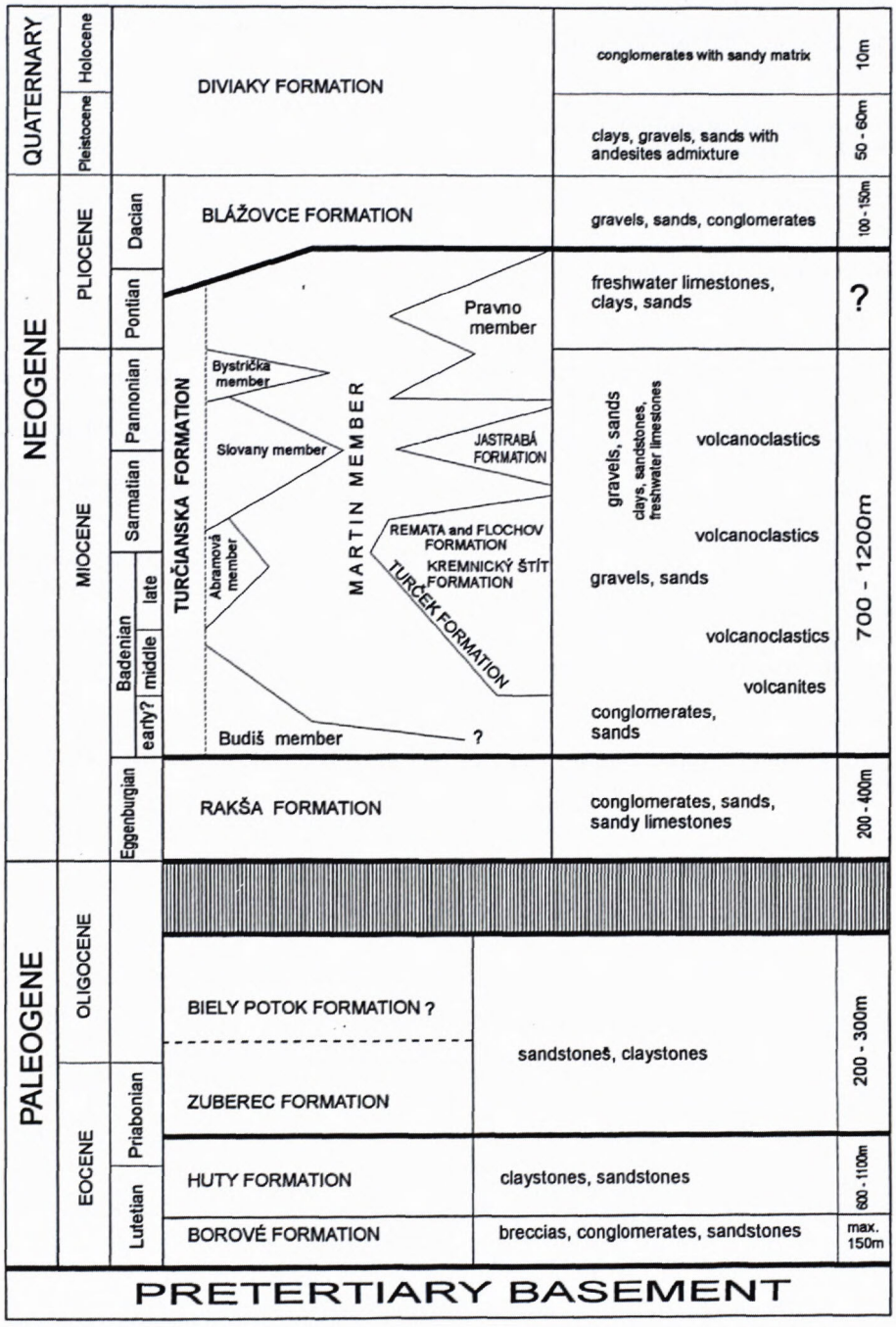
The verified thickness of the Neogene deposits in the basin centre is up to 1027 m (borehole ZGT-3). The compressional axis of the palaeostress field was during the Turiec Formation deposition (Early? Middle Badenian – Late Pannonian/Pontian) oriented in the NNE-SSW to NE-SW direction. The most conspicuous phenomenon of the Turiec Depression is tilting of its fill generally toward the west. The main role during the depression history had marginal listric faults nearby the west margin which were active during the whole Neogene development of the depression. The faults originated in an extensional regime and younger faults were activated gradually in the west direction (Fig. 6). The generally E-W oriented extension also determined an origin of the antithetic faults nearby the eastern margin of the depression. The marginal fault activities resulted in origing of huge alluvial fans of the Slovany, Abramov and Bystrička Members as well as in position of the Slovany Member, which material exclusively consists of carbonate rocks, over the crystalline basement. Such a position may be explained by strike faults along the north-south oriented fault systems. The result of these movements have been bridge structures. The Paleogene deposits hugging the whole eastern margin of the depression do not occur under the Neogene at the western margin of the depression.

During the Middle and Late Pannonian the connection between the southern sedimentary area of the Turiec Depression and sedimentary areas of Pannonian clastics in the other parts of the Western Carpathians occur. The connection of the sedimentary area was proved by the occurrence of congeria fauna (*Congeria exgr. Ornitopsis Brusina*) described by Andrusov (1954) and documented by our field investigation.

During the Pliocene the sedimentary area of the Turiec Depression is isolated which is probably related to the change of the palaoestress field. The compressional

component rotates from the NE-SW to the NW-SE direction. The maximum deposition in that time was concentrated in the southern part of the depression. In this part the clastic deposition continued to Pleistocene.

We assume new lithostratigraphic division of Tertiary deposits in the Turiec Depression where we introduced a new lithostratigraphic unit – Turiec Formation (Tab. III). We used a term Martin Member for the basinal pelitic facies. The lithostratigraphic division of mainly Neogene deposits sources from the previous works of Buday (1962, Tab. I), Gašparik et al. (1995, Tab. II) and, above all, from the conclusions of our research.




*Tab. III Lithostratigraphic column of the Tertiary fill of the Turčianska kotlina depression (compiled by Kováč, Hók, Rakús and Nagy, 1997*

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# Tectonic position of Veporicum and Hronicum Tribeč Mts.

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*Abstract.* In the northeastern – Rázdiel part of the Tribeč Mts. a position of the Tatricum, Veporicum and Hronicum tectonic units was determined more precisely. The Tatricum is an allochthonous unit consisting of crystalline fundament and Late Paleozoic and Mesozoic rocks. The Veporicum is composed of a newly defined sequence of Veľké Pole represented by a crystalline rock complex and subautochthonous rock sequence of the Late Paleozoic and Mesozoic and Krížna nappe. The topographic projection of the Veporicum fault sole is an equivalent of the Čertovica lineament which represents a tectonic boundary of the first order in the West Carpathians. The Hronicum comprises tectonically scaled sequences of the Late Paleozoic and subordinary represented Mesozoic rocks. The displacement of Veporicum on the Tatricum from SE and E toward NW and W was defined by the kinematic indicator analyses. The overthrust of Hronicum occurred from SW toward NE after the overthrust of Veporicum. The tectonic ramp originated by piling of the Tatricum and Veporicum rock complexes resulted in tectonic scaling of the Hronicum.

*Key words:* West Carpathians, Tribeč Mts., tectonics, Tatricum, Veporicum, Čertovica lineament

## Introduction

The Tribeč Mountain is the westernmost extremity of the inner zone of the core mountains exposed from the Tertiary deposits of the Danube Basin. Morphologically it comprises a NE-SW direction horst with the dip of the axial part toward SW.

From the point of view of regional-geologic division (Vass et al. 1988) the Tribeč Mts. is divided into Zobor and Rázdiel part (Fig. 1). The southern Zobor part consists of granitoids, imbricated Mesozoic cover sequence and Krížna nappe. The Rázdiel part is composed of the pre-Permian metamorphic rocks, granitoids and stratigraphically reduced sedimentary cover sequence with conspicuously represented Permian basal formation as well as Veporicum, Krížna nappe and Hronicum.

The basic conception of the geologic map of the Tribeč Mts. is expressed on geologic map at scale 1:50 000 (Biely 1974). The geological structure of the Tribeč Mts. was mainly studied by Biely (1962), Krist (1959, 1971), Rekošová (1987), Kopál (1989), Ivanička & Hók (1992), Hók et al. (1994), Hók (1997), Ivanička et al. (1992, 1994, 1995, 1996, 1998) and others.

## Working methods and results obtained

The submitted article sources from the results of the basic geological mapping at the scale 1:25 000 and 1:10 000. Taking into account the geologic structure of the Tribeč Mts., the problematics of the tectonic relations between the Tatricum, Veporicum and Hronicum was solved in the Rázdiel part. More types of granitoid rocks and crystalline schists were divided and more precious distribution of the Permian deposits was done. We defined the imbricated geological structure of the crystalline complex and cover sequence (Kopál 1989, Hók et al. 1994, Ivanička et al. 1996, 1998).

## Tatricum

More types of granitoid rocks were divided within the crystlline complex of the Rázdiel part (Ivanička et al. 1998). The tectonically lowermost structural unit is composed of leucocratic granitoids overlain by their normal sedimentary cover in the stratigraphic range of the Late Paleozoic and Middle Triassic. The Permian rocks only occur in the Rázdiel part and they are represented by Skycov and Slopňa Formations (Vozárová & Vozár 1988, Vozárová & Vozár in Ivanička et al. 1998). The Mesozoic complexes of the cover series mainly crop out in the area of Zobor massif. They are represented by formations of the basal Early Triassic clastics occurring up to the Early Cretaceous (Tab. 1). One of its characteristic feature is a shallow-marine development of the Jurassic lithostratigraphical units. The whole cover sequence of the Tribeč Mts. is typical by metamorphosed reworking of its members.

## Veporicum

The lowermost member of the higher positioned structural unit is represented by mylonitized granitoids

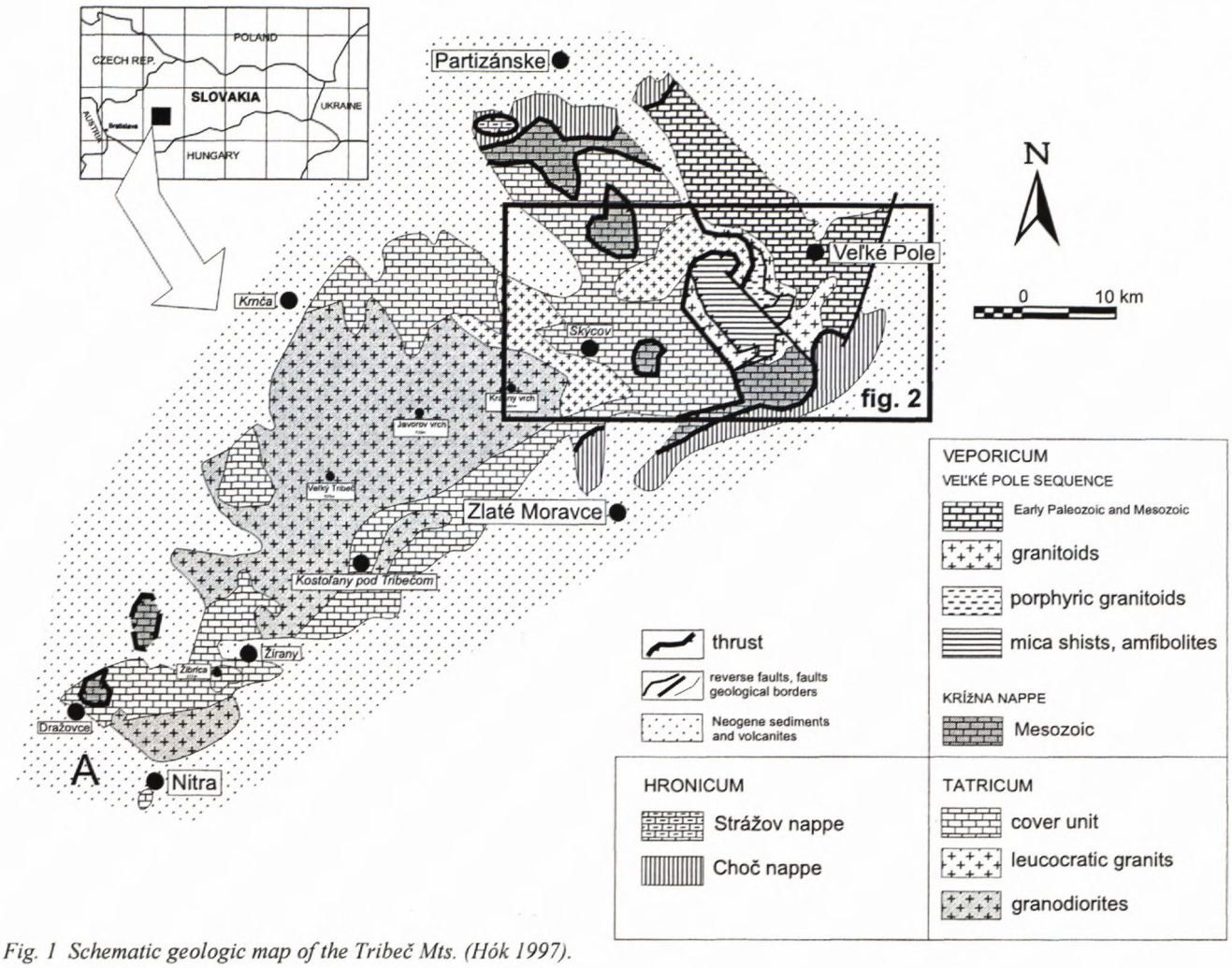



*Fig. 1 Schematic geologic map of the Tribeč Mts. (Hók 1997).*

(blastomylonites) and Permian deposits overlain by shist phyllites in tectonic position which are in turn overlain by deformed granitoids. The whole sequence is capped by deformed granitoids (blastomylonites) with subautochthonous sedimentary sequence in the stratigraphical range from the Permian up to the Middle Cretaceous. According to ambiguous tectonic position above the Tatric rock members, we assign this structural unit to the Veporicum.

The results of a new investigations, especially the cartographic division of the crystalline complex, give an answer to the tectonic position of Veporicum and Krížna nappe in Rázdiel part of the Tribeč Mts. Solving the problem of the tectonic position of the Krížna nappe Biely (1972) states: „ The mentioned facts imply different position of the Zliechov Serie in the different parts of the mountain. In the most part of the area it occurs in a „cover„ position, in the smaller part in a displaced position.„ The ambiguous interpretation was hindered by missing knowledge about the occurrence, tectonic position and relation of the crystalline fundament to the metamorphosed Late Paleozoic and Mesozoic as well as structural position of the Mesozoic, which is individualized (often stratigraphically reduced) and lies on the cover – tatrid unit, which apparently consists of a normal bed succession. The connection of the sedimentary sequences to the fundament was interpreted unambiguously in the past. Biely (1961a, 1962a) assumed that the Krížna nappe is autochthonous here similarily to the „Štiavnica island„ and „Staré Hory island„. Later Biely (1974), convinced about the occurrence of the cover replacement phenomenon (remplacement de couverture) revised his opinion and he located the fault sole between the Early Triassic formation and Middle Triassic carbonates. The conception is also expressed in the geologic map 1:50 000 (Biely 1974). The model solved above all the situation from the viewpoint of the Krížna nappe outliers occurring in the NW Rázdiel part and S of Skýcov (Fig. 2), which lie on the undoubtably Tatric basement and are not related to their fundament. However, the bed succession commences by either Early Triassic or even by Late Triassic (elevation point Kruh, 580 m) and only in one case it commences by the Middle Triassic which should represent basal formation above the decollement plane or fault sole, respectivelly. From the viewpoint of the knowledge level concerning the crystalline structure,

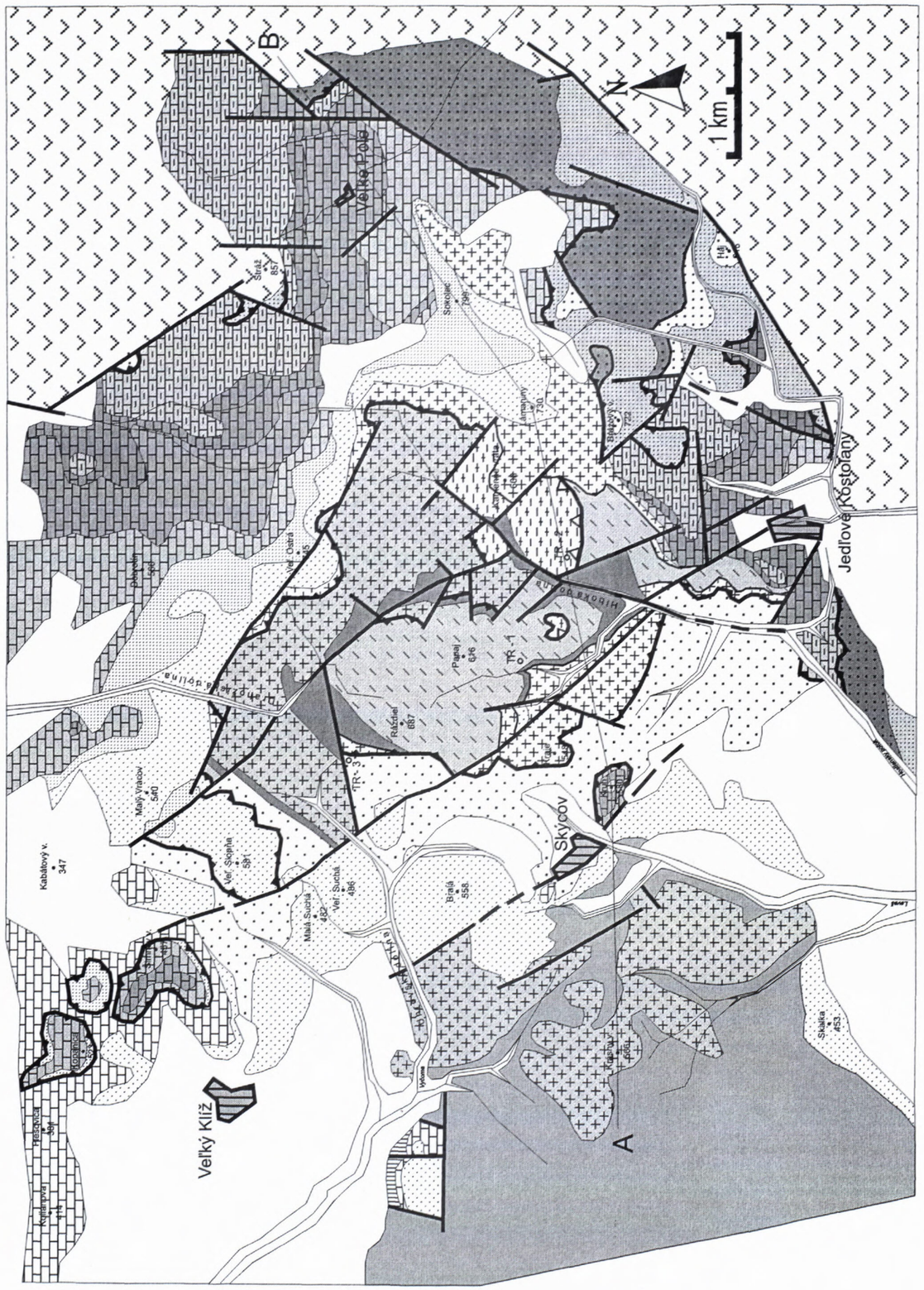



*Fig. 2 Schematic geologic map of the Rázdiel part of the Tribeč Mts. (constructed by Hók 1997 based on Ivanička, Polák & Vozár)*

QUATERNARY

- deluvial and fluvial sediments

TERTIARY

- volcanic and volcanosediments (Middle Badenian - Middle Sarmatian)

HRONICUM

- Benkovský potok Formation (Early Triassic)
- arcoses, greywacs and volcanits, volcanosediments - Permian
- dark grey shales, sandstones Carboniferous

VEPORICUM
VEĽKÉ POLE AND KRÍŽNA NAPPE

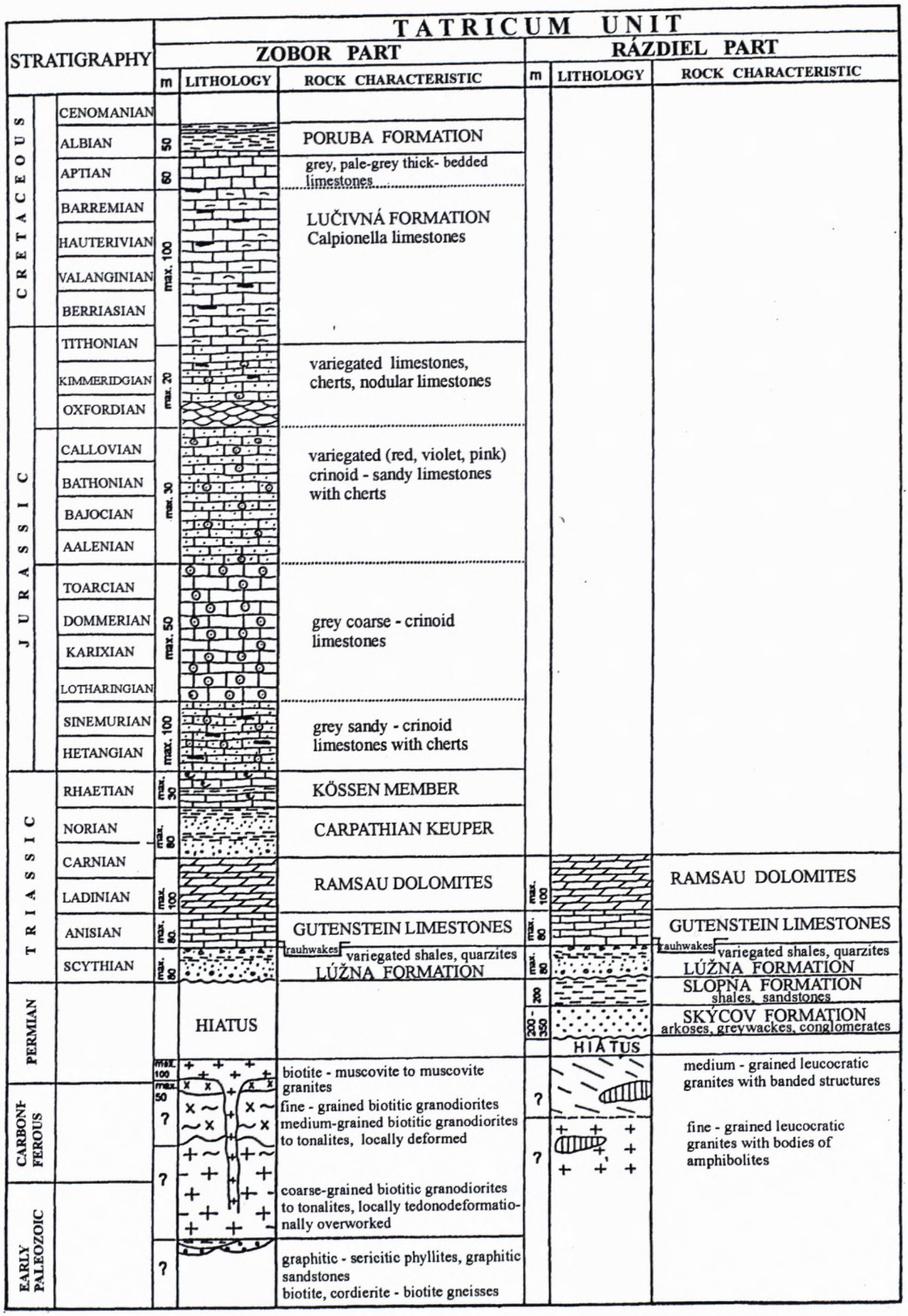
- marly shales, marls, sandstones and limestones (Tithonian - Albian)
- radiolarites,nodular limestones, shales,fleckenmergel (Hetangian - Kimeridgian)
- sandstones, shales, dolomites, dark limestones (Carnian - Rhaetian)
- light limestones and dolomites (Anisian - Ladinian)
- Lúžna Formation (Early Triassic)
- variegated arcoses, greywacks,shales (Permian)
- porphyric granitoides
- amphibolites
- micaschists
- granodiorites (partly mylonitized)

TATRICUM

- dark limestones with cherts (Lias)
- grey limestones and dolomites (Anisian - Ladinian)
- Lúžna Formation (Early Triassic)
- arcoses, greywacs and variegated shales (Permian)
- leucocratic granitoides, aplites
- granodiorites

general explanations

- a) geological boundaries
  b) thrust linie
- a) Hercynian thrust
  b) Alpine thrust
- a) faults observed
  b) faults supposed
- TR - 3 — boreholes




*Tab. I: Lithostratigraphic table of Tatricum (Ivanička et al., 1998)*

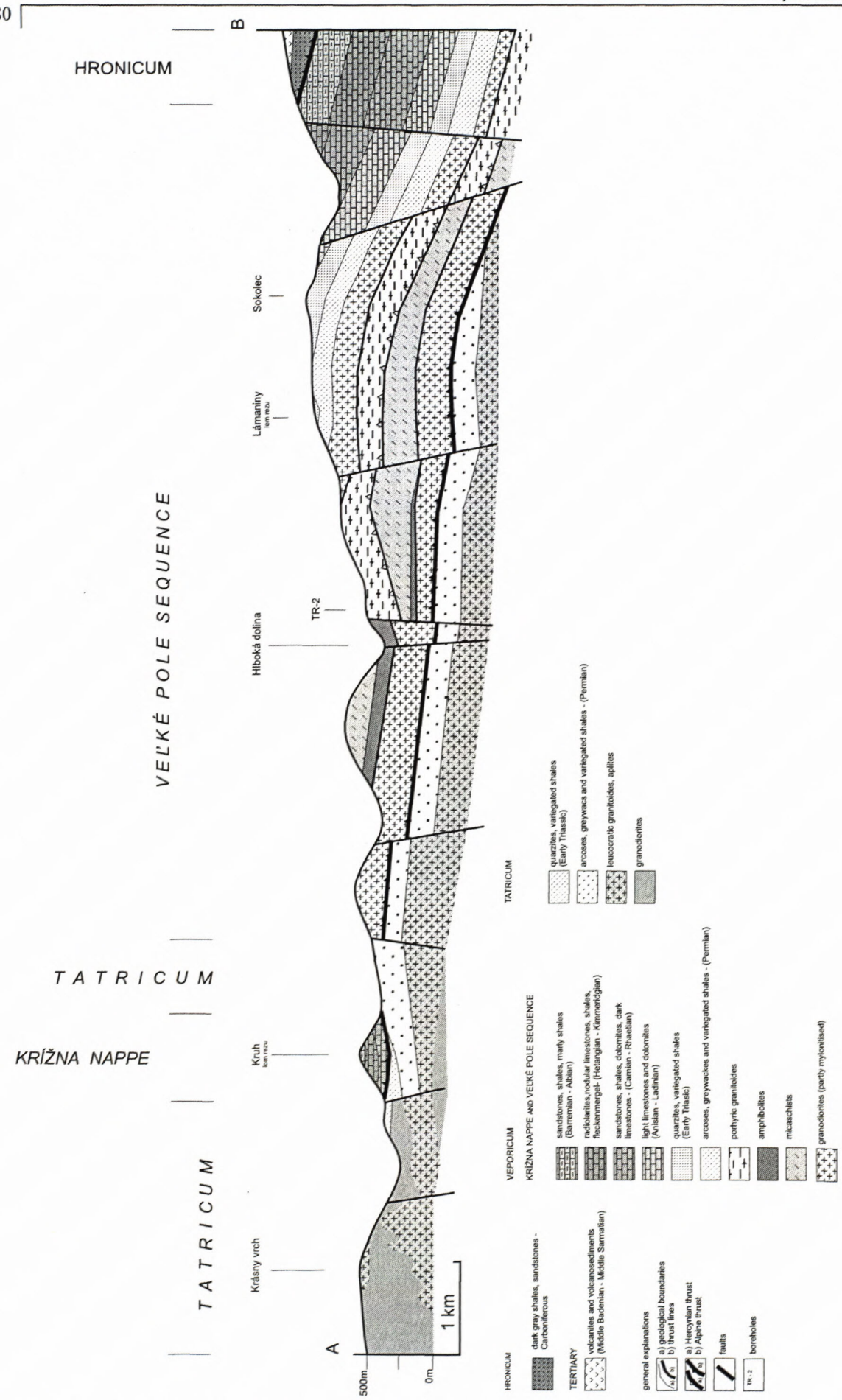
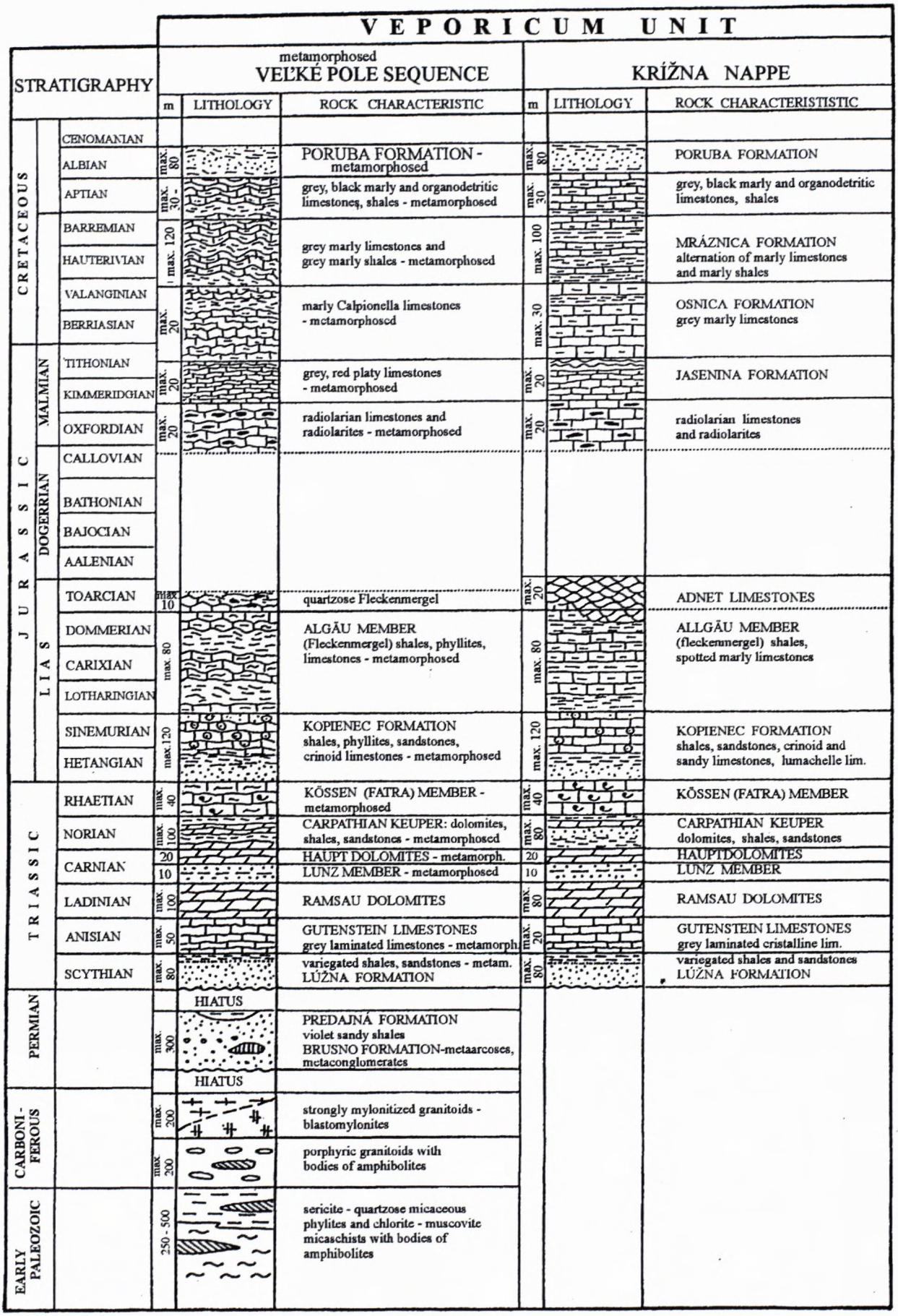



*Fig. 3: Simplified geologic profile of the Rázdiel part of the Tribeč Mts. (Hók, Ivanička & Polák 1998).*




*Tab. II: Lithostratigraphic table of Veporicum (Ivanička et al. 1998)*

the conception of structuralization and tectonic position of the Krížna nappe in the given area was fully acceptable.

A new geologic mapping (Kopál 1989, Ivanička et al. 1992, 1996, 1998, Hók et al. 1994, Hók 1997) showed that the crystalline rocks underlying sedimentary sequences consist of more petrographic types and it has nappe/overlap characteristics from the tectonic point of view. The metamorphosed sedimentary beds of Krížna nappe (sensu Biely 1974), ranging from the Permian to Albian, are related to their crystalline fundament and together with it they lie in the nappe position on the crystalline rocks and Mesozoic which Tatric tectonic assignement is undoubtable. These sedimentary beds may be from the viewpoint of their tectonic position, lithology and metamorphosis correlated with the Veľký Bok sequence. From this reason we assigned to the Veporicum also the structure of the metamorphosed Late Paleozoic – Mesozoic bed succession together with the crystalline beds (shist phyllites, porphyric granitoids and blastomylonites) which during the displacement remain attached to its crystalline basement. We termed the whole sequence as a Veľké Pole Sequence (Polák in Ivanička et al. 1998). Unmetamorphosed Mesozoic bed succession of Zliechov type which covers the Tatricum basement without its crystalline fundament was assigned to the Krížna nappe (Tab. II).

The structural analysis of the Krížna nappe, Veľké Pole sequence and rocks of Tatricum suggests a system of relatively flat overfaults with repeating stratigraphic or tectonic sequence which does not show reverse bed successions (Fig. 3). Such an arrangement of individual structures is typical for imbricated and duplex structure, respectivelly. Thus the displacement of Veporicum and Krížna nappes occurs on the system of flat displacement/nappe faults following suitable rock boundaries.

Kinematic indicators, analysed in the framework of Krížna nappe and Veľké Pole sequence show the orientation of overlap from SE, ESE toward NW, which is confirmed by results of the previous works (Rekošová 1987, Kopál 1989, Hók et al. 1994).

## Hronicum

Similarly to the Veporicum, also rock sequences of Hronicum only occur in Rázdiel part of the Tribeč Mts. The tectonic unit of Hronicum includes Choč and Strážov nappe which are understood as tectonic analogue of the originally unified nappe body sensu Havrila & Buček (1992) and Havrila (1993).

Hronicum is mainly represented by the Late Carboniferous Nižná Boca Formation and Permian Malužiná Formation (Vozár & Vozárová 1988) occurring predominantly on the SE margin of the Rázdiel part. Mesozoic prevailingly consists of the Triassic members commmencing the Early Triassic Benková Formation and terminating by Haupt Dolomites of the Norian age (Tab. III). Hronicum comprises a duplex structure consisting of two scales (duplexes) which structure is predominantly formed by the Early Paleozoic members. Their position and occurrence of individual lithostratigraphic members implies an existence of a structure elevation or tectonic ramp before the displacement of Hronicum in the studied area. The elevation was most probably formed by piles of Veporicum structural units originated during the tectonic individualization of Krížna nappe. The ramp resulted in scaling of the basal

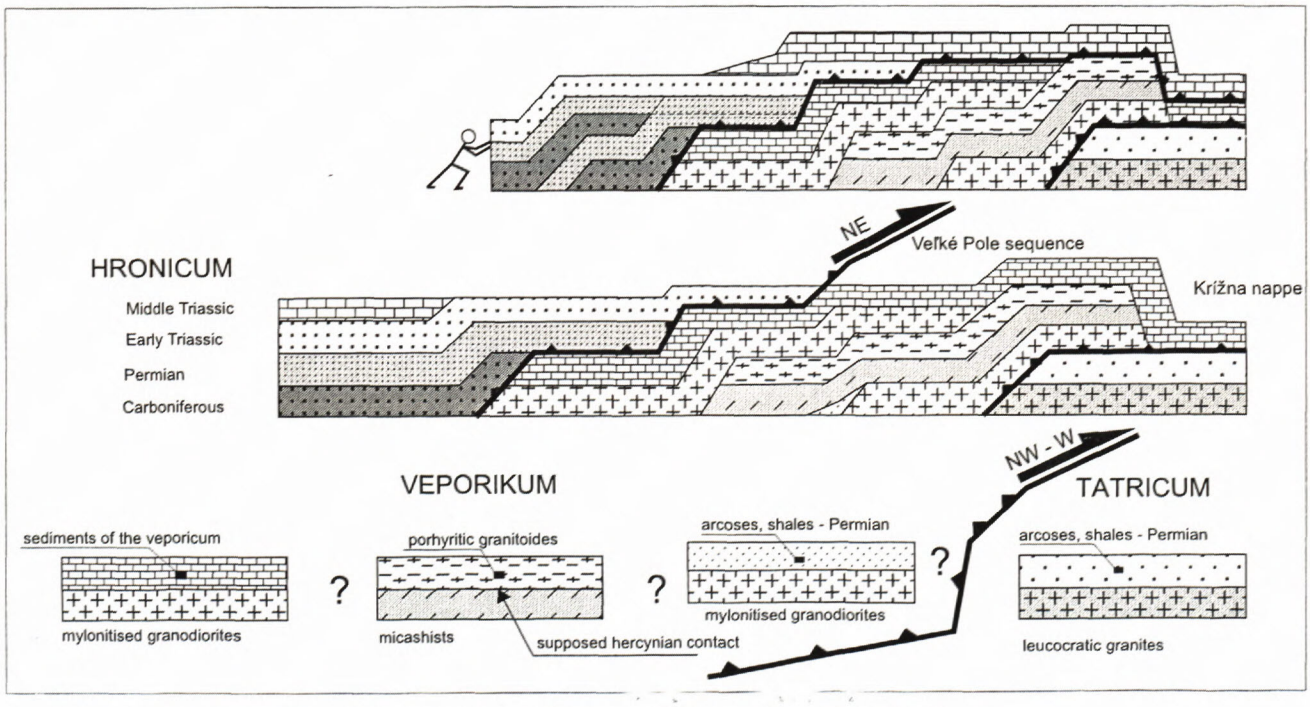
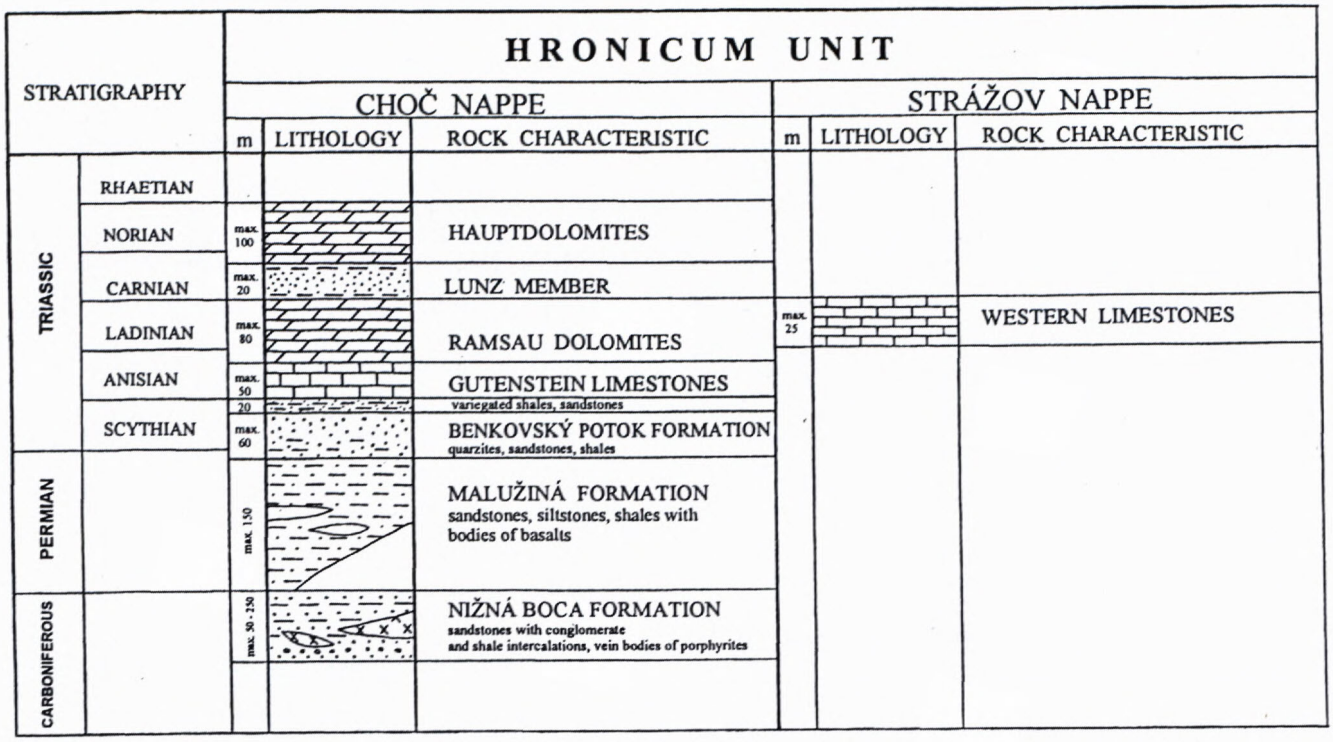



*Fig. 4 Model of overlapping of tectonic units in the Rázdiel part of the Tribeč MTs. (Hók, 1997*




*Tab. III: Lithostratigraphic table of Hronicum (Ivanička et al. 1998)*

Hronicum lithostratigraphic members. The upper members were detached and further displaced on their foreland. The stratigraphically uppermost rock complexes of the Late Triassic only occur in the NW part of Rázdiel area. It is also possible to untangle the assumed overlap mechanism from the geologic map (Fig. 2), c.f. also Biely (1974), Ivanička et al. (1998). Both scales of Choč nappe are present in the southern part of Rázdiel area. They are above all represented by Late Paleozoic members, folded Early Triassic formation and occasionally preserved rauhwacked , probably Middle Triassic carbonates. Toward the north and northwest only the upper scale is developed where the carbonates on the base of the nappe are substituted by the deposits of the overlying Permian (c.f. Vozárová & Vozár 1988). The obvious impliciation from the map is that the main overlap sole was originating gradually on the base of the upper stratigraphic levels of the Early and finally of the Middle Triassic.

## Discussion

On the basis of the facts obtained, we tried to reconstruct tectonic history of the Veporicum and Hronicum structures in the Rázdiel part of the Tribeč Mts. (Fig. 4). The original distribution of the individual structural elements assumes an autochthonous unit (Tatricum) generaly located north of the future allochthonous units of Veporicum and Hronicum. It is difficult to estimate the width of the space shortening between individual segments. We assume a shortening in an order of tens kilometers. In the subsequent etape the Krížna nappe starts to individualize as an allochthon. The Hronicum was overlapped on such an structural elevation and during its overpassing an inner structuralization underwent resulting in origin of partial scale structures. The lower structure was probably metamorphosed in higher conditions assuming mainly from the bigger deformation of the Early Triassic deposits and preserved relic of the Middle Triassic carbonates changed into rauhwacs. The overlying structure of the Choč nappe individualized gradually in the higher stratigraphic members during its advance toward the foreland.

## Conclusion

In the Rázdiel part of the Tribeč Mts. a position of the Tatricum, Veporicum and Hronicum units was made more precise by a detail geologic mapping. From the tectonic point of view Tatricum comprises the lowermost autochthonous unit consisting of granitoids, Late Paleozoic cover deposits and Mesozoic. Veporicum overlies Tatricum in a nappe position. It consists of tectonic structures of the Krížna nappe and a newly divided Veľké Pole sequence. The Veľké Pole sequence is a tectonic equivalent of the Veľký Bok sequence. It is composed of crystalline fundament and rock complexes of the Late Paleozoic and Mesozoic, which are tectonically bounded to it. Based on the kinematic analysis in the Veporicum rocks it is possible to determine its overlap from SE and E toward NW and W. The topographic projection of the displacement/nappe sole of the Veporicum on Tatricum represents an equivalent of the Čertovica lineament (sensu

Biely & Bezák 1997) in the Rázdiel part of the Tribeč Mts.

The Hronicum is mainly composed of the Late Paleozoic deposits and volcanics, Mesozoic deposits are less frequent. Similarly to Veporicum, its inner structure consists of two tectonic scales predominantly composed of Late Paleozoic rocks. The overlap of Hronicum occurred from SW toward NE.

Based on the distribution of the rock complexes of individual tectonic units and their tectonic deformation we assume that the overlap of Hronicum followed the overlap of the Veporicum. During the process of tectonic individualization of Veporicum a structural and probably also a morphological elevation (tectonic ramp) formed which made a barrier for overlapping Hronicum.The stratigraphic lower members of Hronicum stuck on the foreland of the tectonic ramp and displacement/nappe line gradually generated on higher stratigraphic levels responsible for origin of further tectonic structures/ scales.

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# Sequence stratigraphy approach to the Central Carpathian Paleogene (Eastern Slovakia): eustasy and tectonics as controls of deep-sea fan deposition

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*Abstract:* The Central Carpathian Paleogene shows a sequence stratigraphy pattern developed as follows: - alluvial-fan deposition; - transgressive onlap by shore zone and carbonate platform deposits, mainly Nummulitic banks (Upper Lutetian - Bartonian); - drowning of carbonate platform and highstand aggradation marked by high productivity of Globigerina Marls (Priabonian); - interference of lowstand and rapid tectonic subsidence in fault-controlled accumulation of marginal slope fans (Priabonian - Early Oligocene); - highstand deposition of mud-rich fans associated with condensation (manganese beds) and Menilite episodes (Lower Oligocene); - lowstand, progradational wedging out of sandy-rich fans (Late Oligocene - Early Miocene). Correspondence of sequence-stratigraphy events (e.g. Globigerina-rich, Menilite as well as Krosno) indicates connection of the basinal systems of the Central and Outer Carpathians accommodating the destructive plate margin, trench zone and accretionary terranes.

*Key words:* Central Carpathian Paleogene, deep-sea fans, sequence stratigraphy, eustasy, tectonics.

## Introduction

In the deep-sea fan envinonments, the application of sequence stratigraphy is still questionable. In the passive margin settings, the deep-sea fan deposition is clearly influenced by the eustatic sea-level changes (Vail et al. 1984, 1987 etc.). Hovewer, in active margins where the frequency of tectonic events is greater than that of eustatic changes, the sequence development responds to the integrated effects of both these parameters (Posamentier et al. 1991). The deep-sea fan system of the Central Carpathian Paleogene shows an organization responsible for the geodynamic setting of active margin-fans (cf. Shanmugam & Moiola 1988) having an elongate shape, development of attached lobes as well as suprafan lobes (cf. Marschalko 1981, 1987, Soták et al. 1996, Janočko et al. 1998). Therefore, the depositional stacking of the Central Carpathian Paleogene is constrained to be a result of eustasy and/or tectonics.

## Sequence stratigraphy approach[+]

The Central Carpathian Paleogene is commonly divided into lithostratigraphic formations of Subtatric Group (Gross et al. 1984) and Podhale Flysch (Golab 1959). From the base, the sedimentary sequence is developed as follows (Fig.1A): ***Borové Formation*** (Eocen Tatranski) - basal transgressive facies consisting of breccias, conglomerates, polymict sandstones to siltstones (Tomášovce Member - Filo et al. 1996), marlstones, organodetrital and organogenic limestones; ***Huty Formation*** (Zakopane Fm.) - claystone/siltstone lithofacies less frequently with interbeds of fine- to medium-grained sandstones and "Menilite"-type shales; ***Zuberec Formation*** (Chocholów Fm.) - a sandier, medium- rhythmical flysch sediments; Biely Potok Formation (Ostrysz Fm.) - massive sandstone banks. Each of the Central Carpathian Paleogene formations contain a coarse clastic fans named Pucov Member. The thickness of these formations is highly variable depending on bottom configuration and differential subsidence. As for the age, the sediments of the Central Carpathian Paleogene belong to the Bartonian - Lower Oligocene (e.g. Samuel & Fusán 1992, Gross et al. 1993), but their nannoplankton zoning extends up to the Latest Oligocene (Nagymarosi, Harmšmíd & Švábenická in Soták et al. 1996, Olszewska & Wieczorek, 1998).

In fact, the lithostratigraphic units of the Central Carpathian Paleogene appear to be depositional sequences, that developed from the continental alluvial-fan deposits, overlapped by shoreface sands and carbonate platform deposits, through synrift accumulation of shaly flysch deposits and scarp breccias (Šambron Beds), claystone subflysch deposits of mud-rich fans, progradational stacking of deep-sea fans with complex facial zones (slope and channel deposits, lobe-levee deposits, fan-fring lobes, basin-plaine deposits, etc.) to sandy-rich deposits of "suprafan" (Fig. 1B). Considering that, the basin-fill formations of the Central Carpathian Paleogene basin show a sequence stratigraphy pattern of systems tracks (Fig. 2).

The alluvial-fan sediments of the Central Carpathian Paleogene (Marschalko 1970) consist of subaeric and

[+] Communication of ESSE Conference, Bratislava 1997

subaquatic cycles of conglomerates and boulder breccias, which were deposited from stream flows, fluidal surge flows, debris flows, traction currents and high-density turbidite currents (Baráth & Kováč 1995). Continental footplain sediments of alluvial fans were flooded to subaquatic zone and than overlapped by shoreface sands (Tomášovce Beds) and carbonate platform deposits. Upper Lutetian transgression (Andrusov & Köhler 1963) led to shallow-marine deposition of Nummulitic banks developed in two 3rd order cycles of deepening-upward

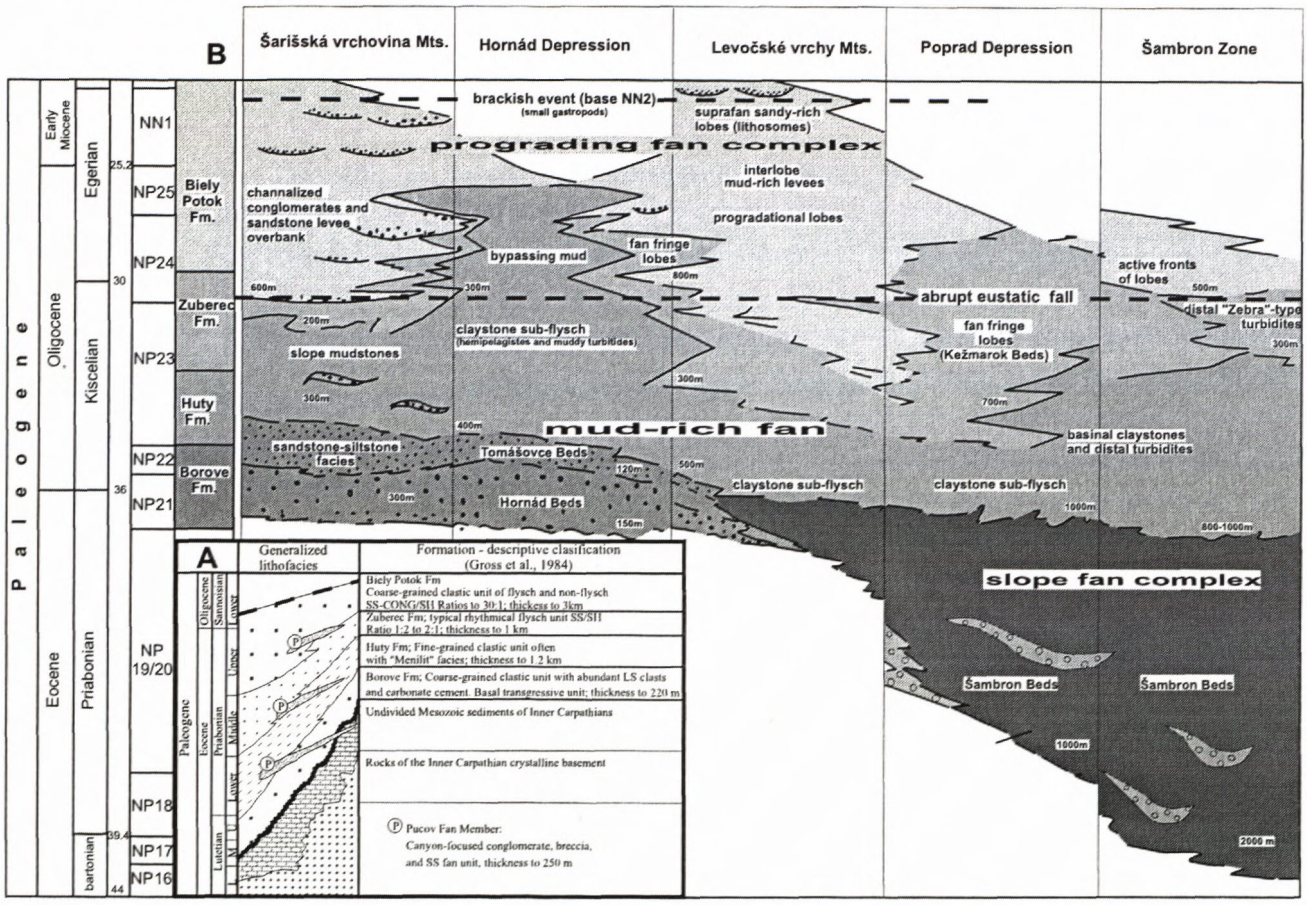



*Fig. 1 Schematic lithostratigraphy of the Central Carpathian Paleogene after Gross et al. 1984 (A) and depositional stacking of the Levoča Basin interpreted as a facies tracks of the deep- sea fans (B) Tomášovce Beds - sensu Filo and Siráňová, 1996; Hornád Beds - sensu Filo and Siráňová, 1998; Kežmarok Beds / sensu Gross, in press.*

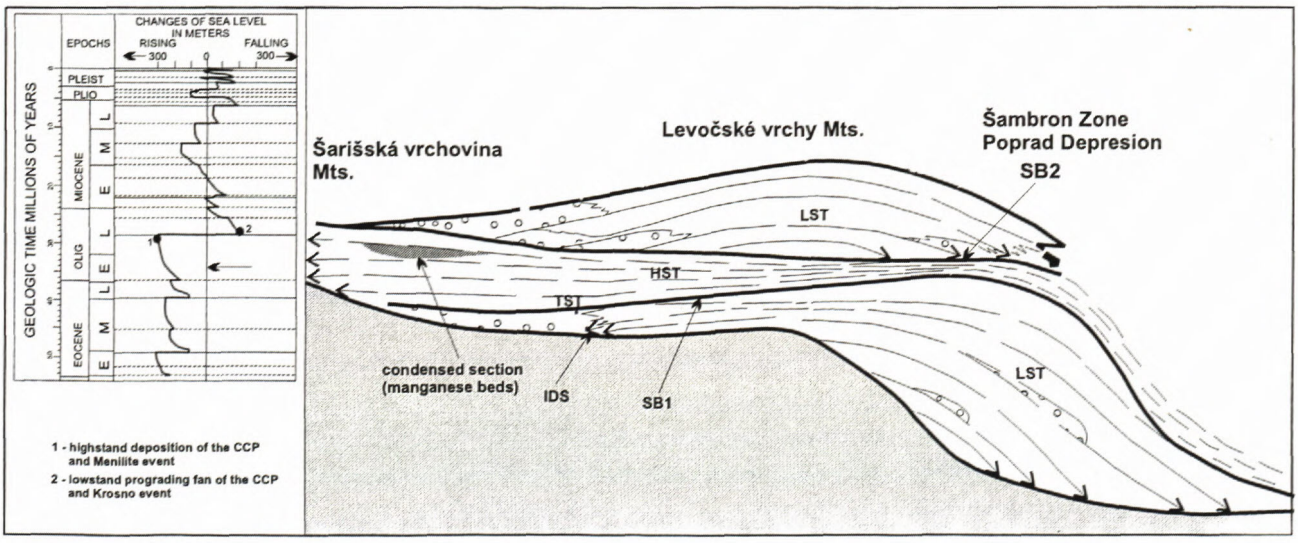



*Fig. 2 Sequence stratigraphy pattern of the Central Carpathian Paleogene showing a lowstand accumulation of the slope fan complex (Šambron Beds), transgressive onlap and highstand deposition of mud-rich fans with maximim condensation in manganese beds (Huty Formation) and lowstand cycle of the upper prograding wedge. Sequence boundary SB2 coincides with the abrupt sea-level fall near 30 Ma (see the Tertiary sea-level curve on the left).*

successions (Bartholdy 1997, Bartholdy & Bellas 1998). In the Priabonian, the sea-level rise grades up to early highstand marked by dominance of hemipelagic deposition and high productivity of Globigerina Marls (Samuel 1973, Blaicher 1973, Nemčok et al. 1990, Leszczynski 1997). With onset of differentiated subsidence, the Globigerina Marls occurred synchronously with turbidite fans deposited in intrabasinal depressions (Šambron Beds). Subsequently, the Central Carpathian Paleogene basin occurred in dominancy of turbidite fan deposition forced by tectonics and/or eustasy.

It is obvious, that the Central Carpathian Paleogene basin was founded as a consequence of the regional tectonic subsidence (e.g. Marschalko 1978, Köhler & Salaj 1997). When the tectonic subsidence rate decreased, the turbidite fan deposition became broadly influenced by eustasy. Thus, the depositional sequences of the Central Carpathian Paleogene could correspond with the global eustatic events. Especially, the sequence boundary of the mud-rich subflysch facies and sandy-rich deep-sea fan deposits closely correspods with the dramatic sea-level fall (for about 300 m) between Lower and Upper Oligocene, which is the most distinct global record on the Tertiary sea-level curve (see Haq et al., 1988) occurred at the time of 30 Ma (near transition of NP23/NP 24-25 biochrons) This indicates, that the Late Oligocene deep-sea fan deposition of the Central Carpathian Paleogene should be accelerated by the global sea-level fall and formed as a lowstand system track. The Upper Oligocene is regarded as a beginning of the Antarctic glaciation under which the Northern Hemisphere became significantly cooler (Robin 1988). It is well known that the frequency of turbidite currents increase in time of the glaciation (ratio of turbidity currents in postglacial time to those in glacial time is at least 1 : 10 - Shanmugam & Moiola 1982, Eberli 1991). Therefore, the acceleration of turbidite deposition in the Late Oligocene deep-sea fans of the Central Carpathian Paleogene could really result from sea-level fall. Eustatic force of the Late Oligocene sedimentation is also apparent when the subsidence curves are compared with rate of accumulation and sea-level history (Fig. 3). These curves show an interference of rapid tectonic subsidence with a high rate of synrift sediment accumulation during the Late Eocene (Šambron Beds). Conversely, an increased supply of the deep-sea fan deposits in the Late Oligocene lacks a tectonic subsidence record, and that it revealed most likely an allocyclic initiation probably in the sea-level fall near 30 Ma.

From the sequence stratigraphy viewpoint, the turbidite fan system of the Central Carpathian Paleogene can be divided into three depositional sequences (DS) comparable to third-order cycles (estimated duration of 1-5 Ma). Lower cycle of the basin-fill sequence is represented by the Priabonian to Early Rupelian sediments of the Šambron Beds. The high accumulation rate of shaly flysch deposits, turbidites and scarp breccias of the Šambron Beds (cca 800 m/Ma, duration 39 - 36 Ma) points to fault-control lowstand deposition of the marginal slope fans (like as Szaflary Fm. - Wieczorek 1989). The lowstand setting of the Šambron Beds is also expressed by the presence of large amount of shallow-marine detrital components (nummulites, coralline algae, etc.) in the deep-water flysch lithologies (shelfs exposed by sea-level fall). The sedimentation of the Šambron Beds occurred probably in the Oxygen minimum zone (anoxic facies, scarcity of microfossils, dark sulphide-rich claystones, etc.) and below upwelled calcite compensation depth (non- or weakly calcareous claystones). The younger third-order cycle of the Central Carpathian Paleogene is represented by the mud-rich subflysch formation. This cycle kept on for about 5 Ma (36 - 31 Ma) with a slow accumulation rate of mudstone deposits (cca 80 - 160 m/Ma). Such dominance of pelagic sedimentation in the basin responds to the highstand system track. Maximum flooding of this highstand formation falls into the horizons of manganese layers that occur mainly in the Poprad Depression (the sequence division below the manganese layers can also be assumed as a transgressive system track). The manganese layers represent a condensed section of the highstand deposition associated with relative abundance of biota (e.g. nannofossils, fish fauna, etc.), glauconite-rich arenites (as a contrary to the lowstand turbiditic sandstones of the Šambron Beds and Upper Oligocene sediments), pelocarbonates and sporadically also tuffaceous intercalations (e.g. loc. Bajerovce, Plavnica). The depositional environment of the claystone lithofacies became well-oxygenated as can be seen from the appearance of bathyal ichnofossils (e.g. Zoophycos - Plička 1987). The Early Oligocene highstand sedimentation of the Central Carpathian Paleogene correspods to so-called Menilite event in the Outer Flysch Carpathians. The sediments of the Menilite Formation and associated nanno-chalk horizons in the Outer Flysch Carpathians (Jaslo Limestones, Dinów Marlstone, Štibořice Member, etc.) were deposited during the coeval sea-level highstand in the Rupelian (cf. Krhovský & Djurasinovič 1993).

The last third-order cycle of the Central Carpathian Paleogene reflects an abrupt sea-level fall near the time of 30 Ma which introduced the Upper Oligocene deep-sea fan deposition. The lowstand setting of the Late Oligocene sedimentation of the Central Carpathian Paleogene is expressed by an offlap break of prior highstand sediments which were eroded and reworked into conglomerate-slope accumulations of the deep-sea fans (e.g. blocks of Mn carbonatic ores - Marschalko 1966). During the Late Oligocene, the fan became progradational, and developed from sandy-poor turbidite system (Zuberec Fm.) to sandy-rich turbidite system (Biely Potok Fm.) sensu Mutti (1985). The sandy-rich deposition of the Central Carpathian Paleogene lasted till to the Early Miocene, as has been already indicated by some nannoplankton and foraminiferal species (e.g. *Helicosphaera scissura*, *H. kamptneri*, *H.* cf. *carteri*, *H.* cf. *ampliaperta*, *Reticulofenestra* cf. *pseudoumbilicus*, *Triquetrorhabdulus* cf. *carinatus*? - Nagymarosi, Hamršíd & Švábenická in Soták et al. 1996, Molnár et al. 1992) and by sandstone content of the neovolcanic clasts related to the Early Miocene volcanic activity (Soták et al. 1996). Hovewer, the Early

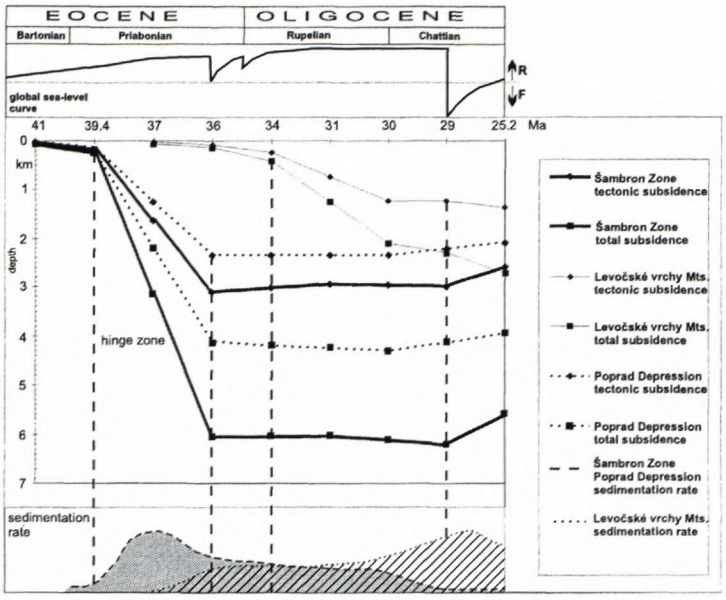


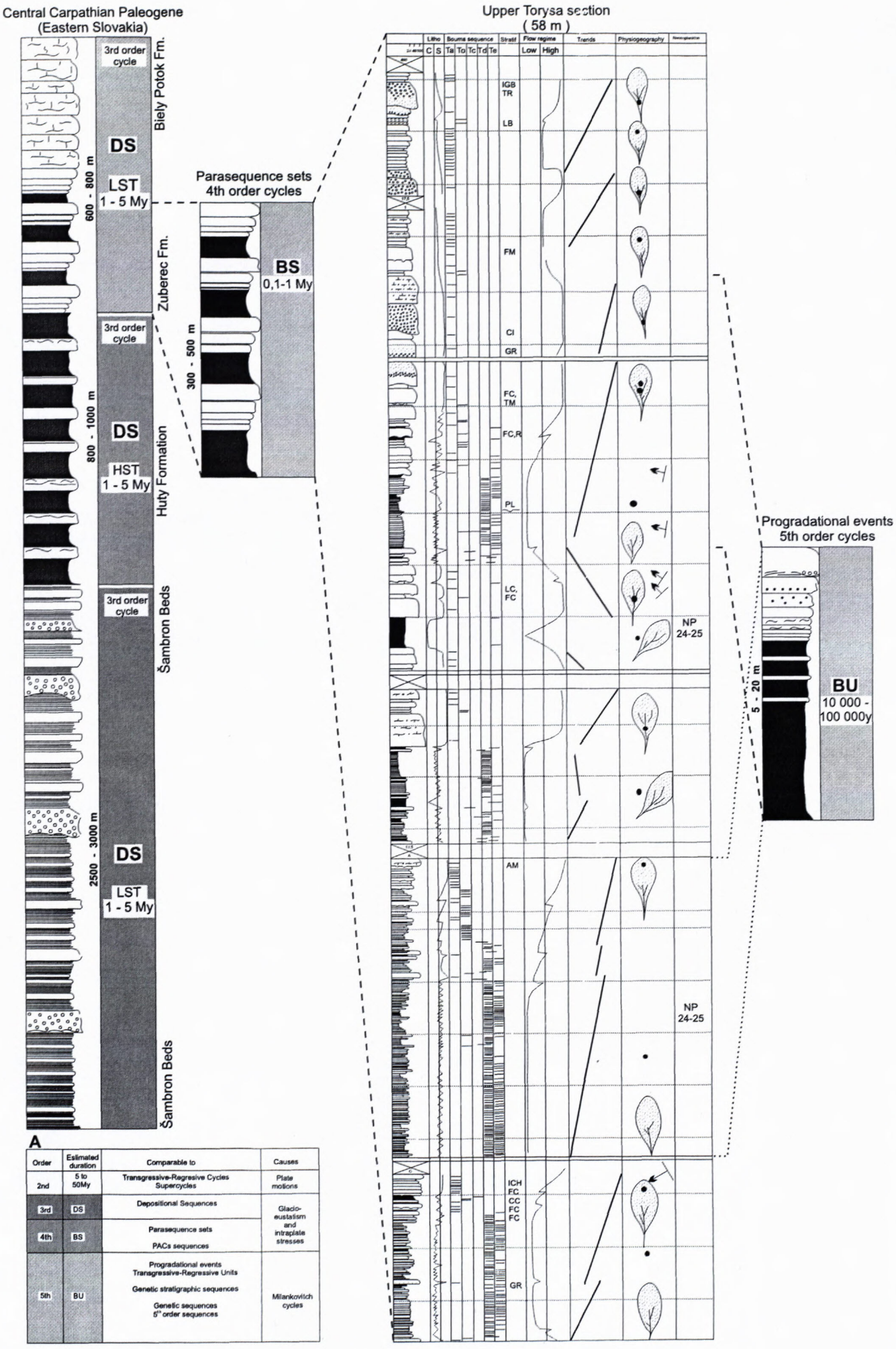

*Fig. 3 Subsidence patterns of the Central Carpathian Paleogene derived from backstripping of the basin-fill sequences, compared with sea-level fluctuations and sedimentation rates*

Miocene age is more apparent from the sequence stratigraphy correlations. During the Late Oligocene - Early Miocene, the global eustasy occurred under a distinctive regression, which led to gradual shallowing and brackishing of Paratethyan basins. In the Central Carpathian Paleogene basin, the Late Oligocene regression is recorded by the Biely Potok Fm. providing an input of sandy-rich deposits and shallow-water brackish species of dinoflagellates (Hudáčková 1998). Regressive trend of the Late Oligocene - Early Miocene sedimentation reached the maximum lowstand on the base of the NN2 zone, when the brackish fauna became to appear (Steininger et al. 1995). Such brackish event, indicating by the fauna of small gastropods, has been recently detected in the sandstone lithosome sediments of the Levočské vrchy (Soták in prep.). By this, the deposition of the Biely Potok Fm. should terminated till to the Early Eggenburgian (Late Egerian sensu Berggren et al. 1995), i.e. to the lowstand phase at the beginning of the NN2 zone, which proceded next transgressive cycle TB 1.5 sensu Haq (1991) occurred on the base of the Prešov Fm. (Hudáčková et al. 1996, Kováč & Zlinská 1998). In fact, the gastropod-bearing sandstones and overlying sandstone lithosomes of the Biely Potok Fm. (cca 300 m in the Levočské vrchy Mts.) could not be assigned to "flysch", but rather to molassa sediments deposited during a retrogressive stage of basin evolution. Nevertheless, the vitrinite reflectance data from the near-surface sediments of the Levoeské vrchy Mts. point out, that up to 2 km of this sequence is missing (Kotulová et al. 1998). So that, the last third-order cycle of deep-sea fan deposition in the Central Carpathian Paleogene kept on for about 7 Ma (29 - 22.5 Ma) giving a high accumulation rate of sandy-rich deposits (320 - 370 m/Ma). The time-equivalent sedimentation in the Outer Flysch Carpathians took also place in lowstand setting, recorded by the Krosno Facies (incl. Ždánice-Hustopeče event - Krhovský & Djurasinovič 1993).

The sequence stratigraphy approach allows to subdivide the basin-fill sequence in more detail - Fig. 4 (according to predictive approach of Ciner et al. 1996). The main depositional sequences (DS) can be divided into particular cyclic units such as progradational sequence of lobes, lithosome sequence of suprafan, channelized sequence of the upper fan zone, etc. (each having a thickness of 300 - 600 m). Although, the BS development can easily be explained by autocyclic factors (compensation cycles levelled deep-sea fan relief - Mutti & Sonnino 1981), we cannot exclude even their origin in the sea-level fluctuations (cf. Mutti 1992, Posamentier et al. 1988 etc.). These BSs are comparable with forth-order cycles, which correspond to parasequence sets spanning the time between 0.1 - 1 Ma. The BS can be further subdivided into the basic units (BU). The BUs are developed as a small-scale cyclic units of lobes, channels, levees, etc. The thickness of these BUs in the Levoča Mts. ranges from 5 to 20 m. For example, in a complete 13 m thick lobe unit there are about 115 individual turbidites. Considering the recurrence time for one turbidite event about 300 - 400 years (Rupke & Stanley 1974) the duration of this unit may be calculated for 35 to 46 Ka. In this case, the BUs might correspond to fifth-order cycles developing over intervals between 10 and 100 Ka (i.e. Milankovitch cycles). Using BU data to calculate the BS (cca 20 BUs), a time span of about 700 Ka to 1 Ma appears to be reasonable for the forth-order cycles.

## Conclusions

The Central Carpathian Paleogene basin accommodates the destructive plate-margin domain. The basin-fill formations began to develop from continental footplain sediments, shoreface sands and carbonate platform deposits (Nummulitic banks) related to Lutetian -Bartonian transgressive cycles 3.5 and 3.6 sensu Haq et al., 1988 (cf. Bartholdy 1997). In the Priabonian, the carbonate platforms were subsequently drowned by sea-level rising, which reached an early highstand with dominance of hemipelagic deposition and high productivity of the Globigerina Marls. Afterwards, the Central Carpathian Paleogene basin occurred under dominancy of turbidite fan deposition. The turbidite sediments of the Central Carpathian Paleogene can be divided into three depositional sequences (DS) comparable to third-order cycles. Lower cycle of turbidite lowstand deposition is represented by the Šambron Beds (Priabonian - Early Rupelian), which are considered to be synrift sediments of the marginal slope fans accumulated along scarps in tilted-fault-block situation (cf. Surlyk 1978). Next depositional sequence reveals the maximum rate of eustatic rise in the Central Carpathian Paleogene basin, which during the Lower Oli-




| Order | Eslimated duration | Comparable to | Causes |
|---|---|---|---|
| 2nd | 5 to 50My | Transgressive-Regresive Cycles<br>Supercycles | Plate motions |
| 3rd | DS | Depositional Sequences | Glacio-eustatism and intraplate stresses |
| 4th | BS | Parasequence sets<br>PACs sequences | |
| 5th | BU | Progradational events<br>Transgressive-Regressive Units<br>Genetic stratigraphic sequences<br>Genetic sequences<br>5th order sequences | Milankovitch cycles |

*Fig. 4 Turbidite systems of the Central Carpathian Paleogene and their possible subdivision to depositional sequences (DS) and cyclic units ( BS - basic sequences, BU - basic units). A - correlation with cyclic sequences proposed in literature, see Ciner et al. 1996.*

gocene starved in highstand deposition of the mud-rich fans (Huty Fm.). Maximum flooding of this highstand formation falls into the horizon of manganese layers (condensed sections of MFS). Early Oligocene highstand sedimentation of the Central Carpathian Paleogene correspods to the so-called Menilite event of the Outer Flysch Carpathians. The last third-order cycle of the Central Carpathian Paleogene reflects an abrupt sea-level fall near time of 30 Ma which introduced the Late Oligocene deep-sea fan deposition. Upper Oligocene sediments were deposited as a prograding lowstand wedge with complex deep-sea zones. Higher up, the sequence grades up to sandy-rich deposits of suprafan, which terminated to the Early Miocene, as is evidenced by the backish event (small gastropods, etc.) indicating the maximum lowstand on the base of the NN2 zone. The time-equivalent sedimentation in the Outer Flysch Carpathians took also place in lowstand conditions recorded by the Krosno Facies. Such correlations indicate a narrower range of the Central and Outer Carpathian basins and their Paleogene (Early Miocene) sedimentation driven by global eustasy.

The turbidite systems of the Central Carpathian Paleogene can be subdivided into more particular up to small-scale units comparable with the forth-order cycles (basic sequences - BS, time duration 0.1 - 1 Ma) and fifth-order cycles (basic units - BU, with time duration of 10 - 100 Ka). The development of these units could have been controlled by autocyclic or allocyclic factors.

The implication of sequence stratigraphy to basin history of the Central Carpathian Paleogene reveals an integrated effects of tectonism and eustasy. The basin began to develop with initial collapse and rapid subsidence, induced probably by tectonics (gravitational collapse due to tectonic erosion - cf. Wagreich 1993). Initial subsidence pattern of the Central Carpathian Paleogene basin reflects a trenchward tilting with a fault-controlled accumulation of marginal slope fans (Šambron Beds). The accommodation space of the Central Carpathian Paleogene basin was enlarged by landward aggradation progressed through alluvial aprons, shoreface banks and shelfal deposits. Landward migration of the basin was driven by a highstand eustasy inferred from overall of pelagic sedimentation. The accumulation of sandy-rich deposits during the Upper Oligocene does not exhibit a tectonic subsidence record (flexural loading). Therefore, the sediment input of deep-sea fans should be initiated by allocyclic (external) factors. Besides of source tectonics, which was a major control of the the deep-sea fan accumulation, the deposition seems also to be accelerated by eustasy (its onset coincides with the abrupt sea-level fall near 30 Ma - Krosno event in the Outer Carpathians). Under the lowstand accumulation of the deep-sea fan clastics the Central Carpathian Paleogene basin became gradually uncompensated with rate of subsidence and overfilled. The sandy-rich submarine wedge grew to form an elevated bulge with a convex-upward relief (suprafan). The sedimentation in the Central Carpathian Paleogene ceased under a tectonic shortening and gradual uplift recorded by the mollase-like sediments of retrogradational cycle (Early Miocene). This paper is a contrib. to VEGA grant no 4077.

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# Gravimetry contribution on investigation of oil shales – alginite in ring structures in Lučenec depression

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*Abstract:* The aim of the contribution is to show possibilities of gravimetry by investigation of structures perspective for the occurrence of oil shales - alginite. Gravity investigation in this problematics was succesfully realized first time in Slovakia. The results from profile and areal measurements are documented from two prospection areas in the Lučenec - Rimava Depression at localities Pinciná and Jelšovec. At this localities volcanic structures were successfully detected and contoured. It was a maar nearby Pinciná and diatreme nearby Jelšovec. At the same time I call attention to differences by the interpretation of areal data gained from detail measurements and data gained from regional gravity mapping at scale 1: 25 000.

*Key words:* gravimetry, map of complete Bouguer anomalies /CBA/, maps of residual gravity anomalies, oil shales - alginite, diatomite, maar

## Introduction

In the period 1992 - 1996 a complex geophysical-geological prospection was realized on the basis of the project of geological-geophysical works (Zbořil et al. 1992, Puchnerová et al. 1993, Puchnerová et al. 1995) in the area of Lučenec - Rimava depression. The prospection was aimed to find suitable structures - maars for potential accumulation of oil shales - alginite. We gathered in considerable extent suggestions for investigation from the prospection results of Hungarian geophysicists and geologists who have long-term experience and positive results from application of geophysical methods for the oil shale - alginite prospection.Considering the available contemporary geological knowledge about the potential occurrence of oil shales, in Slovakia the area of Lučenec - Rimava depression, particularly the area of volcanics assigned to the Podrečany basalt Formation, was chosen for the project. Maar volcanic structures of the Pontian age were the main objective of the work.

The theoretical and practical knowledge of Hungarian geophysicists and geologists from the Geophysical Institute of Lorand Etvös (ELGI, Toth 1975, 1978, 1990) and from the Hungarian Geologic Survey (MAFI) in Budapest (Solti 1987, Ravasz et al. 1992, 1993, 1994) served as a basis for choosing suitable geophysical methods and elaboration of methodologic procedures. Naturally, we considered differences in geology and development of mollase formations in the area of South-Slovakian Basin (Vass et al. 1988, Elečko in Bodnár et al. 1988).

The first etape of the geological-geophysical prospection in the area interested comprised evaluation of the material available, mainly satellite images and areal photographs, aeromagnetic and regional gravity measurements. On the basis of the above mentioned data together with geological knowledge the following localities were chosen for detail geophysical prospection (Fig. 1):

A) Halič - Mašková
B) Podrečany - Točnica - Tomašovce
C) Jelšovec
D) Baňa - Chudač - Čakanovce
E) Pinciná
F) Boľkovce

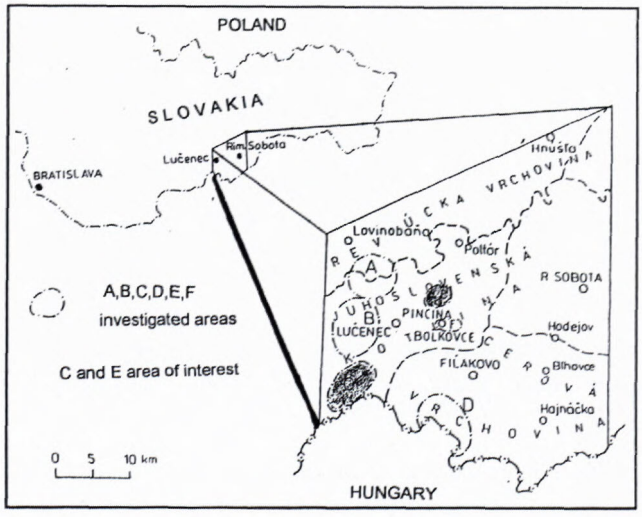


*Fig. 1 Locality map*

In the second etape a detail geophysical prospection was realized at selected localities. It consisted of profile magnetometry, complex geoelectric methods and of profile and areal gravimetry. The profile gravimetry was applied at localities: Jelšovec, Podrečany, Halič and

Pinciná (Puchnerová et al. 1994, 1996a, 1996b, Šantavý 1993). The following localities were investigated by areal gravimetry prospection: Pinciná and Jelšovec with the density up to 20 points per km$^2$ (Šantavy in Puchnerová et al. 1996, Puchnerová et al. 1995). The results of the detail areal gravity mapping were found as very effective because they brought more complex view on structure of volcanic structures and, mainly, they pointed out in more detail on the spatial distribution of the sedimentary - maar fill (Šantavý in Puchnerová et al. 1997).

## Geophysical investigation of the interested area

*a. Regional gravity measurements at scale 1:25 000*

For purpose of the characterisation of regional gravity measurements in a wide surrounding area were mainly used the following works: Šefara et al. 1987 and regional gravity measurements at scale 1:25 000 (Fig. 2) of regional geophysical investigation works (Bárta et al. 1965, Obernauer 1969, Šefara et al. 1970, Bodnár et al. 1973, 1975, 1979). The main objective of gravimetry was observation of morphostructural elements of pre-Tertiary basement beneath mollase deposits and observation of tectonics. A system of high and relatively sunken tectonic blocks has been defined on the basis of areal gravity data interpretation and at the same time more conspicuous density anisotropies in pre-Tertiary substrate were detected (Bodnár et al. 1979).

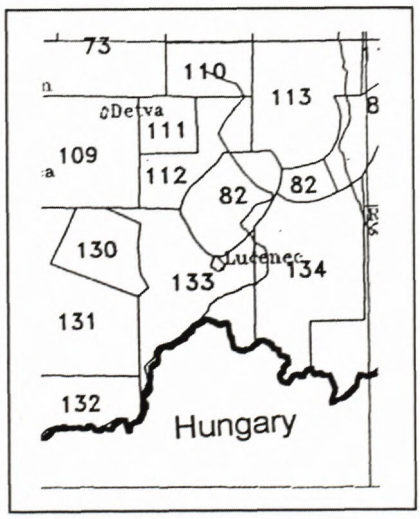


*Fig. 2 Regional gravity measurements in the scale of 1:25 000 (Szalaiová and Šantavý, 1994)*

*b. Detail gravity measurements*

The density anisotropies in the Tertiary fill is expressed less conspicuous in the map of complete Bouguer Bouger anomalies (CBA) because rock complexes differ only little in natural densities. Similarly, also neovolcanic bodies are demonstrated uncospicuously in regional gravity maps because they are represented by areally little extended bodies. From this reason the success of contouring of neovolcanic structure depends on good selection of methodology and on scale of the prospection e.g. the higher

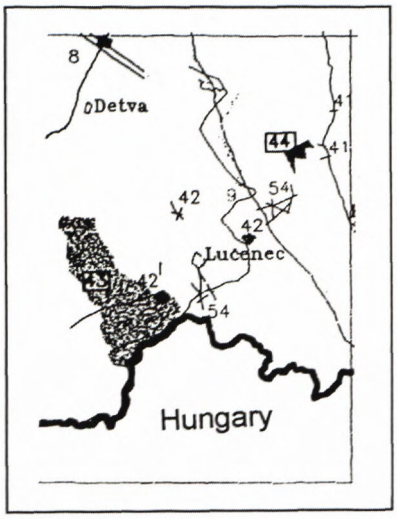


*Fig. 3 Local gravity measurements (Szalaiová, Šantavý and Alfoldy, 1996)*

density of measured gravity points per km$^2$ the higher probability of contouring even relatively smaller volcanic structures. For this purpose more detail gravity measurements were performed (Fig. 3) - areal mapping and also profile measurements. The aim of the work Velich and Mezovský (1988, on the Fig. 3 marked as 54) was to verify possibilities of geophysical methods - detail profile gravimetry and vertical electric sounding measurements by distinguishing the quality of the pre-Tertiary basement of the Lučenec depression. In the area of Luboriečka a geophysical prospection was realized which was aimed at selection of coal-bearing horizons (Džuppa et al. 1987, 1989, 1991). Except areal and profile gravimetry (on Fig. 3 marked as 43) geoelectric measurements of vertical electric sounding were realized. For the purposes of this contribution I mostly used measurement results from the work Puchnerová et al. 1994, 1996 (on Fig. 3 marked 42).

Two gravity measurements, differing by purpose and methodology, are compared in the next chapters. We put emphasis on the possibilities of gravimetry for detection of smaller volcanic bodies.

## Brief geological characterization of choosen localities

The choosen localities - Jelšovec and Pinciná, belongs to the Lučenec Depression of the South-Slovakian Basin. The geologic characteristics was mostly made on the basis of the latest data, mainly on the basis of the geologic map of the Lučenec depression and Cerovo hilly land at scale 1:50 000 (Vass et al. 1992).

Pre-Tertiary rocks are represented by two tectonic units - Veporicum and Gemericum which do not crop out in the area under investigation. From the Tertiary molasse deposits Miocene deposits and volcanics of the Eggenburgian age are present at choosen localities. The lower part consists of Fiľakovo Formation of the marine origin overlying Lučenec Beds. The upper part is composed of Bukovina Formation of continental origin. In the northeastern area of the Jelšovec locality Jelšovec

conglomerates belonging to the Fiľakovo Formation occur. Bukovina Formation consists of cyclic alternation of gravel, sand and motley clay. These sedimentary rocks contain beds and layers of rhyodacite tuffs and tuffits. Motley clays and silts are the most common lithotype of the formation. The Otnangian is represented by Salgotarian Formation consisting of two Member successions - from Pôtor and Plachtiná Members. Pôtor Member forms the lower part of the Salgotarian Formation and it is mostly composed of sand containing coal layers, sandy clays and silts. Plachtiná Member lies conformably on Pôtor Member. It is represented by clay, claystone and siltstone. The prevailing part of Plachtiná Member has originated in either lacustrine or brackish environment which is important for genesis of alginite. The Karpatian is absent in investigated areas - it is extended only in the southeastern part of the Lučenec Depression. The Middle Miocene is represented in the investigated area by the rocks of alcaline-calcareous andesite volcanism. The Pontian is characterized by Poltár Formation extending into southwestern part of the depression and into surroundings of Jelšovec village. The formation mostly lies on the Lučenec Formation (Egerian), in the western part of the Depression it lies on the Fiľakovo or Bukovina Formation. The relationship between Poltár and Podrečany basalt Formations is various. The Podrečany Basalt Formation either overlies or underlies Poltár Formation, or eventually it interfingers the Poltár Formation. Two facies were distinguished in Poltár Formation - fluvial and lacustrine (nearby Pinciná) facies, the fluvial one prevails. It consists of gravel, sand and clay. Clay and silt comprise an important part of the Poltár Formation (Vass et al. 1992).

Podrečany Basalt Formation includes relics of lava flows in the northeastern part of the depression (Podrečany, Mašková), maar nearby Pinciná and relics of two maars on the southwestern margin of the depression (west from village Jelšovec). The relationship between the Basalt Formation and Poltár Formation was described above. If Podrečany Formation does not lie within or on the Poltár Formation then it lies on the either Lučenec Formation (nearby Pinciná, partly also nearby Mašková), Plachtiná and Pôtor Members (nearby Jelšovec, partly nearby Mašková) or on the pre-Tertiary basement as it is nearby village Točnica. The assignment of maars located nearby Jelšovec to the Podrečany Basalt Formation is supported only by structural similarity to the other relics of basalt volcanism on the western margin of the depression (Vass et al. 1992).

The Quaternary deposits consist mostly of fluvial sediments of terraces and flood plains of creeks and rivers and alluvial fans of tributaries of axial river system.

For better documentation of the detail geologic structures of Jelšovec and Pinciná localities profiles of selected boreholes drilled on the basis of geophysical investigation are depicted on Figs. 4a, b, c. The results of the borehole investigation were subsequently used for the quantitative interpretation of geophysical profiles.

## Results of gravimetry by the investigation of oil shales

### *A. Locality Jelšovec*

Two profile gravity measurements were realized along profiles PF-A and PF-1 at locality Jelšovec. Besides profile gravimetry geoelectric measurements and profile magnetometry were also realized. In the second etape on the basis of the positive results from locality Pinciná a detail areal gravimetry prospection of volcanic structure was realized aimed at more detail areal location of the structure.

Maps CBA for reduction density 2.00 $g.cm^{-3}$ are calculated from data of regional gravity mapping at scale 1:25 000 with density of 3 – 6 pints per $km^2$ and from detail measurements. Isolines $\Delta$ g in milligals interpolated from gravity data obtained by regional gravity mapping at scale 1:25 000 are depicted on fig. 6. The fig. 5 showes isolines calculated or constructed from detail gravity measurements at scale 1:10 000 with density up to 20 points per $km^2$ which were realized in the framework of the project "Geophysical investigation of oil shales in Slovakia" (Puchnerová et al. 1995, 1996).

A conspicuous gravity gradient of NW-SE direction as well as a gradual increasing of gravity values toward north, resulting from the map of CBA, is obvious. It reflects a gradual pinching out of the pre-Tertiary basement toward surface (Šantavý in Puchnerová et al. 1996). A relatively negative anomaly about 400 m northeasterly from the elevation point 265 m and a relatively positive anomaly about 150 m north of the borehole VJA-3 are considered as the most conspicuous gravity anomalies in the area investigated (Fig. 6). The positive anomaly can be explained by the presence of the Pontian volcanics, but comparison with the positive anomaly obtained by data interpolation from detail gravity investigation at scale 1:10 000 shows its obvious areal inconsistency with these measurements. This is determined by lower density of measured points per $km^2$ causing possible record of the anomalous body of lesser areal extent only by one or in some cases by no measured point. The mentioned positive anomaly obtained from regional measurements is caused by the gravity value gained by the measurement only at one point. Generally one-point measurements are not considered as reliable. Anomaly has to be recorded at least by three points. Similarly, negative anomaly from regional measurements is also detected only by one point. By one-point anomalies we can not surely exclude possible errors which could for example originate by uncorrected record of measured gravity, uncorrectly determined topographic correction or bad located point on map. Just from this reason it is necessary for operator to be careful to interpret individual anomalies. One always should make sure about the degree of probability of generation of anomaly or its source.

The results of detail gravity investigation (Figs. 5 and 7) render comparison of two methodologic and for purpose different measurements from the same area. The greatest difference is the fact that the regional measurements did not record positive anomaly (or they recorded

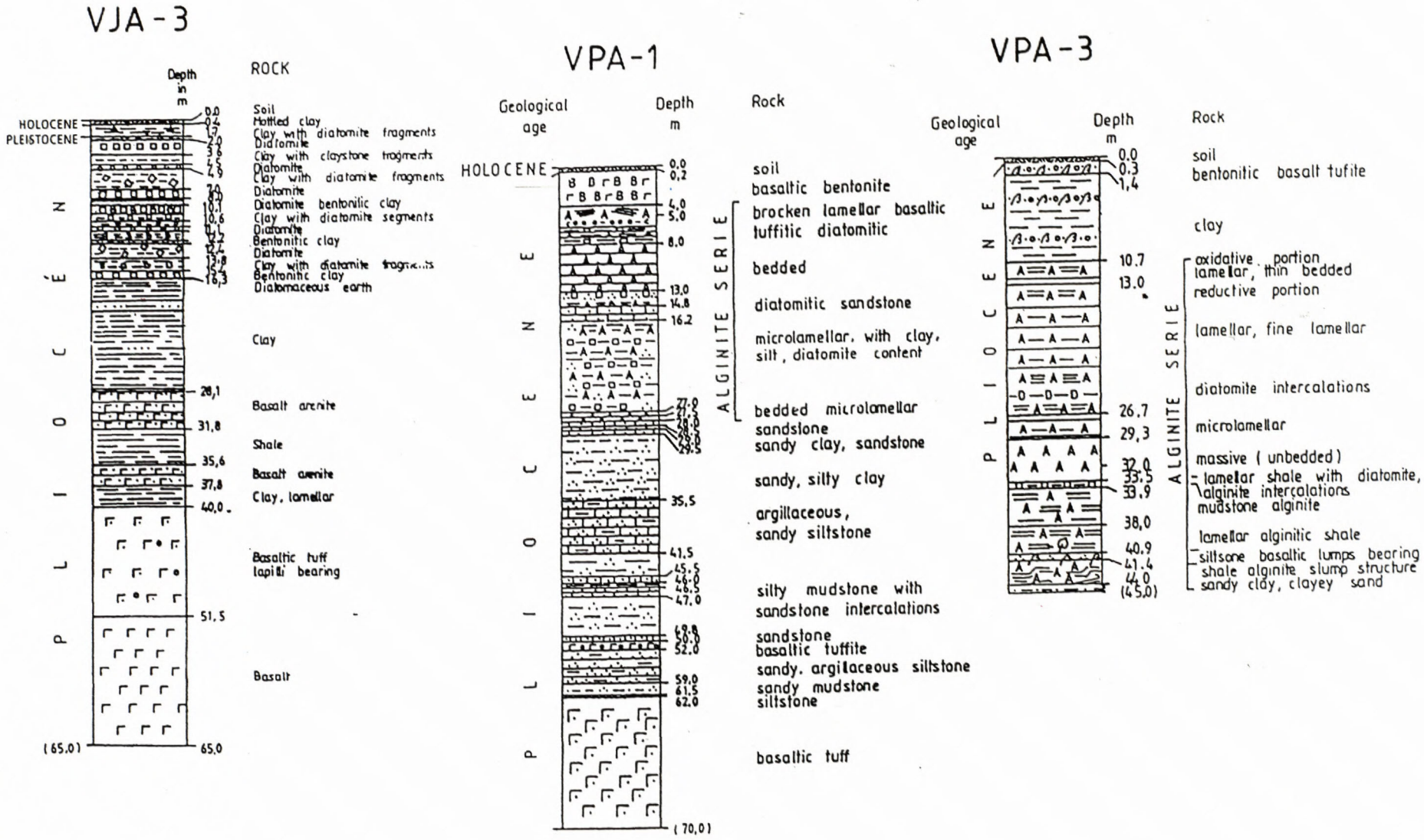


*Fig. 4 (a, b , c) Lithological profile of borehole Jelšovec, Pinciná – (after Ravasz and Solti, 1994)*

only its NNE margin) as a result of low density of measured points. The anomaly with its centrum located about 100 m southwest of the borehole VJA-3 was detected by detail investigation. At the same time the location of the negative anomaly, suggested by results of regional measurements to lie about 400 m northeast of elevation point 265 m, was not confirmed. In this case the anomaly can be considered as "artificial". As a matter of fact, we can not talk about an error of regional gravity investigation because the accuracy of meausurements according to the Technical-organization norm of gravimetry and according to the Instructions for gravity mapping at scale 1:25 000 is 0.5 mGal. The negative anomaly shows intensity -0.3 mGal and, of course, corresponds to the accuracy and purposes of regional gravity measurements. The differences can be also observed on the derived maps. The gravity residuum from regional measurements is generally very unconspicuous as a result of weak representation of high-frequency part while just the high-frequency part represented in detail measurements is a source of frequent local residual anomalies in the given area like is shown on Fig. 5 and more conspicously on Fig. 7.

Results of the detail gravity investigation at locality Jelšovec have contributed to detect the volcanic structure – diatreme, and appoint areal distribution of sedimentary rocks.

**Kvantitative interpretation - modelling on profile PF-A**

On the profile (Fig.8) besides profile gravimetry also geoelectric and magnetometric measurements were done. The profile orientation is NW - SE with the commencement nearby elevation point Hajcov vrch (256 m). Nettleton methods (Šantavý 1993) was used for choosing the reduction density suitable for modelling. The character of particular curves $\Delta g$ calculated for various reduction densities is basically this same even if on the curve CBA for reduction density 2.2 g/cm$^3$ is obvious decreasing of positive anomaly in intervals 550 - 700 m and on the curve CBA for reduction density 2.00 g/cm$^3$ this anomaly practically disappeares. The anomaly source could be seeking in volcanics of the Pontian but the results of magnetometry and geoelectrics are in this part of the profile non-anomalous. According to the change of the gravity field on curves CBA calculated for various reduction densities, we can assume that the above mentioned positive anomaly is probably reflection of the terrain relief and does not connect directly with the geological structure. The second positive anomaly in the interval 1200 - 1600 m is not that conspicous as to intensity, but it includes the influence of both Pontian volcanics and pre-Tertiary basement.

The aim of quantitative interpretation of gravity data was modelling of volcanic structure and its filling. As input data for modelling well data, results of geoelectric measurements (VES - Vertical Electrical Sounding) and petrophysical parameters were used. On Fig.8 we can see a model of volcanic structure and interpreted layer of diatomit, too.

*B. Locality Pinciná*

Also at this locality a complex of geophysical methods similar to measurements realized at locality Jelšovec, was performed. In contrast to the Jelšovec locality, magnetometric investigation could be not realized here because of fenced vineyard.

Similarly to previous case also for this locality maps CBA for reduction density 2.00 g/cm$^3$ were calculated. The calculation was based on data from regional gravity mapping (Fig. 10) at scale 1:25 000 and data obtained from a detail investigation (Fig. 9) made in the framework of the project "Geophysical investigation of oil shales in Slovakia" (Puchnerová et al. 1996). On the processing of detail areal gravity measurements at this locality Dr. Szalaiova, Mgr. Švastová and Dr. Grand took part during the project.

On the first look the gravity isolines obtained from regional measurements are well correlated to results of more detail gravity investigation due to location of the one of the measured point from the regional gravity mapping almost in the centre of the anomalous structure. The anomaly centre obtained from regional measurements is shifted only slightly in about 100 m. The difference is only in the contouring of the volcanic structure. Some disproportions can be observed in the intensities of the gravity, in fact, in their interpolated values between two measurements. This is caused by both different density of the measured points and by interpolation. While on the CBA map computed from detail measurements is a local negative anomaly closed and it is forming a ring structure, on the other hand regional data suggest a certain opening of the structure in its western and southern part. This is also obvious from detail areal measurements and particularly from map of residual gravity anomalies (Fig. 11). The western deformation of the structure forming an elongated local negative anomaly inform us about probable influx (drainage) area which might occurred here during deposition in maar structure. A less intensive northern deformation is suggested from detail measurements. The less intense deformation is probable due to lower thickness of deposits. The southern deformation or opening of the maar has not been confirmed according to these measurements, especially according to the results from residual gravity anomalies map. In the map of residual anomalies we can also divide a NW - SE trending linear deformation crossing the maar centre. The map inform us about the spatial distribution of light deposits belonging to the maar fill. The higher intensity of the negative anomaly in the northern and northeastern part of the maar indicates thicker accumulations of light rocks - deposits, which generate this anomaly. This implies a direct location of spatial distribution of potential alginite deposit by gravimetry determining a suitable location of following prospection works or drilling investigation. The surrounding positive anomalies comprising ring structure are generated by relatively heavier complexes of volcanic rocks of Pontian age, mainly basalt tuffs and basalt s.l.

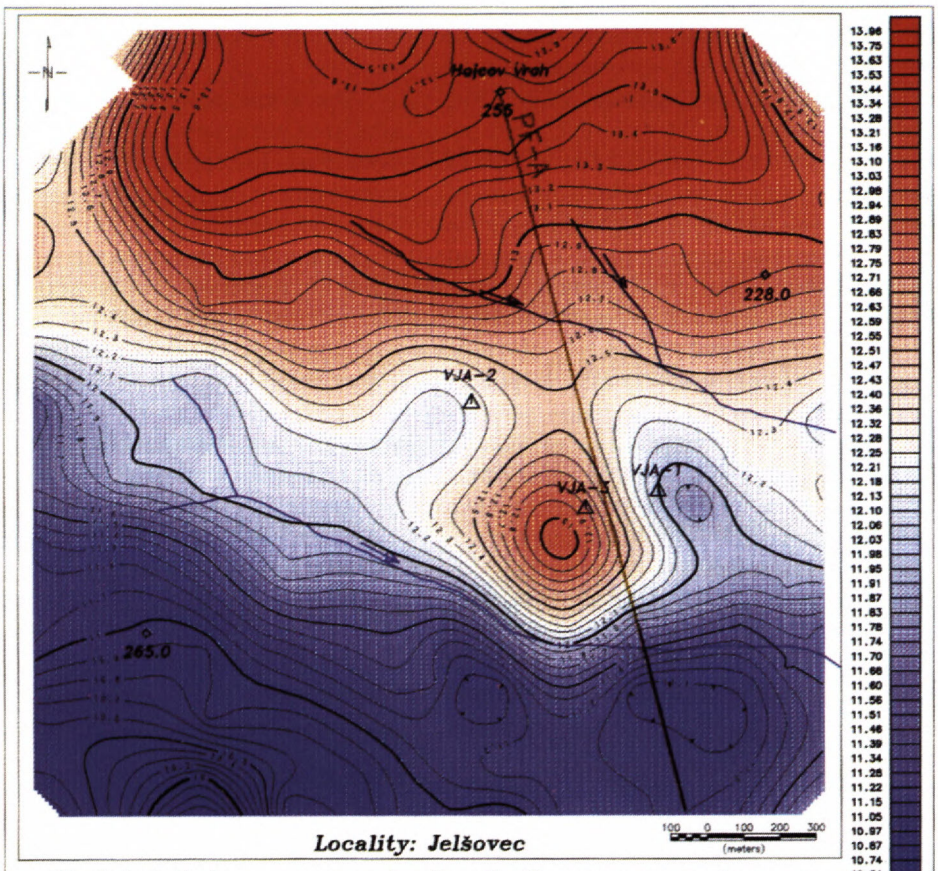


*Fig.5 Detailed measurements: Gravity Bouguer anomaly map density 2.00 g/cm³*
*Compiled by Šantavý J., 1997*

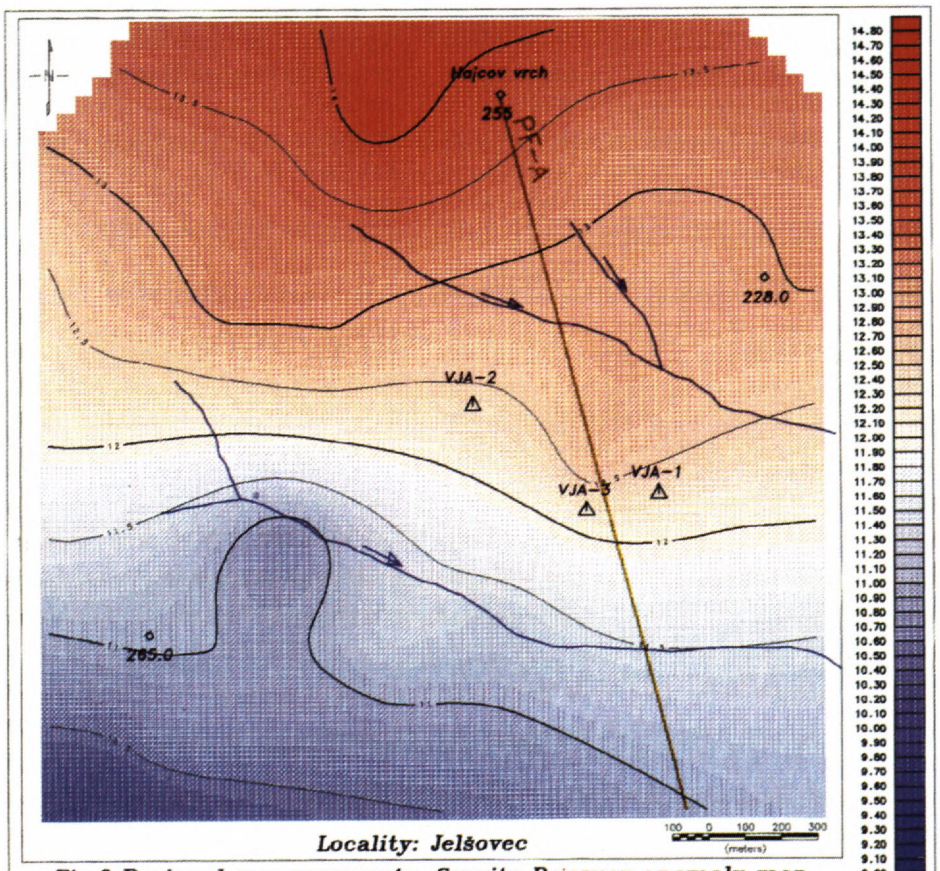


*Fig.6 Regional measurements: Gravity Bouguer anomaly map density 2.00 g/cm³*
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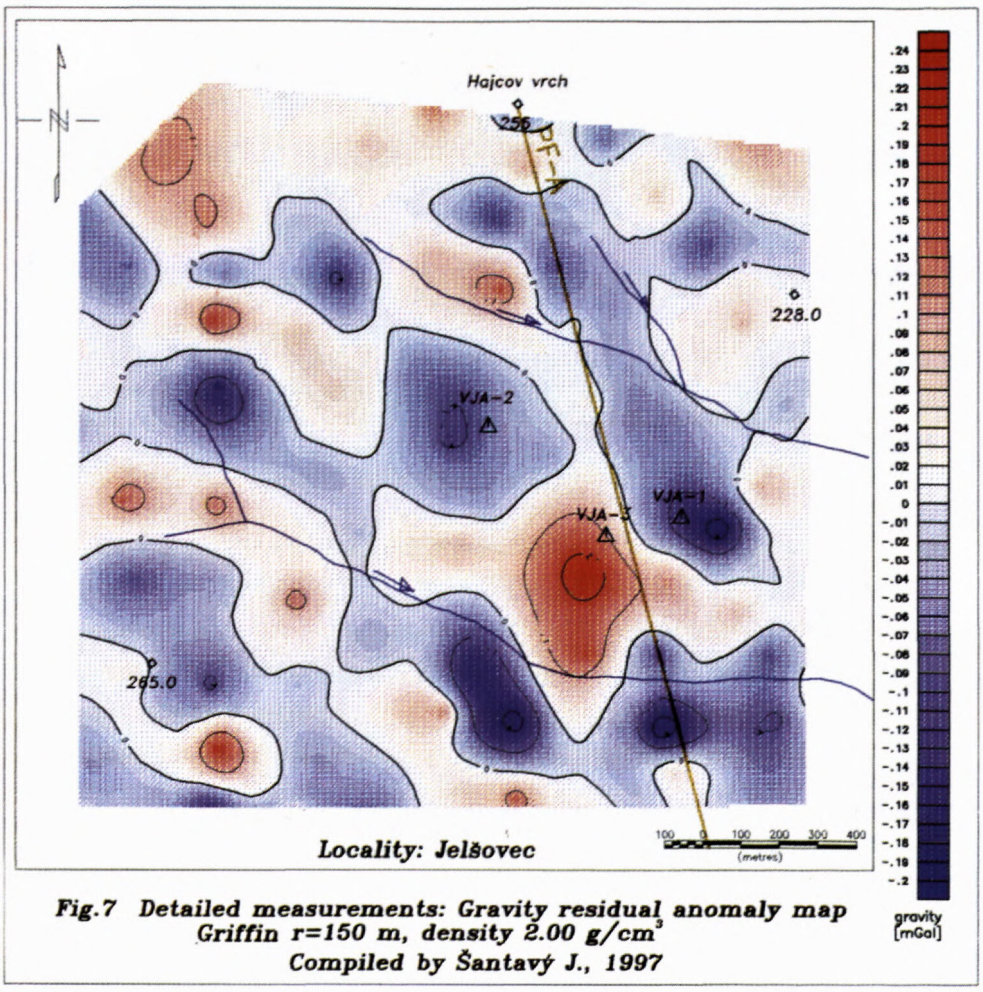


**Fig.7 Detailed measurements: Gravity residual anomaly map Griffin r=150 m, density 2.00 g/cm³ Compiled by Šantavý J., 1997**

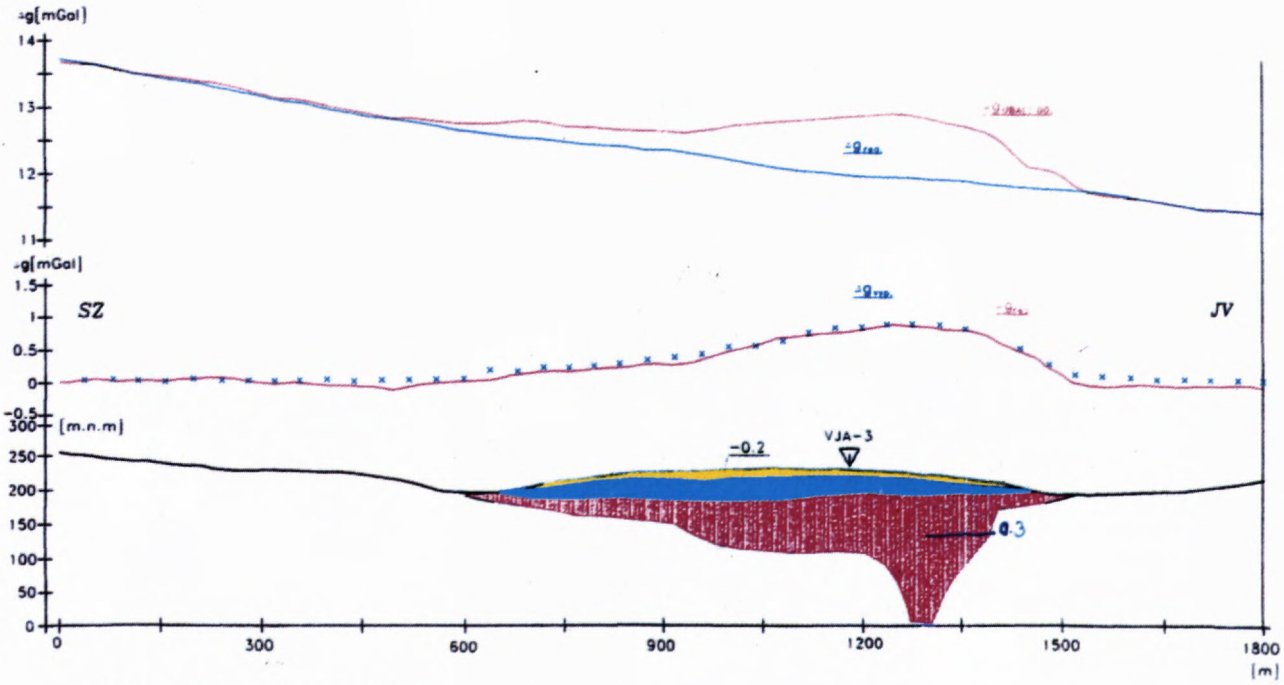


*Fig. 8 Locality Jelšovec – gravity profile PF-A (Šantavý M. in Puchnerová et al., 1996)*

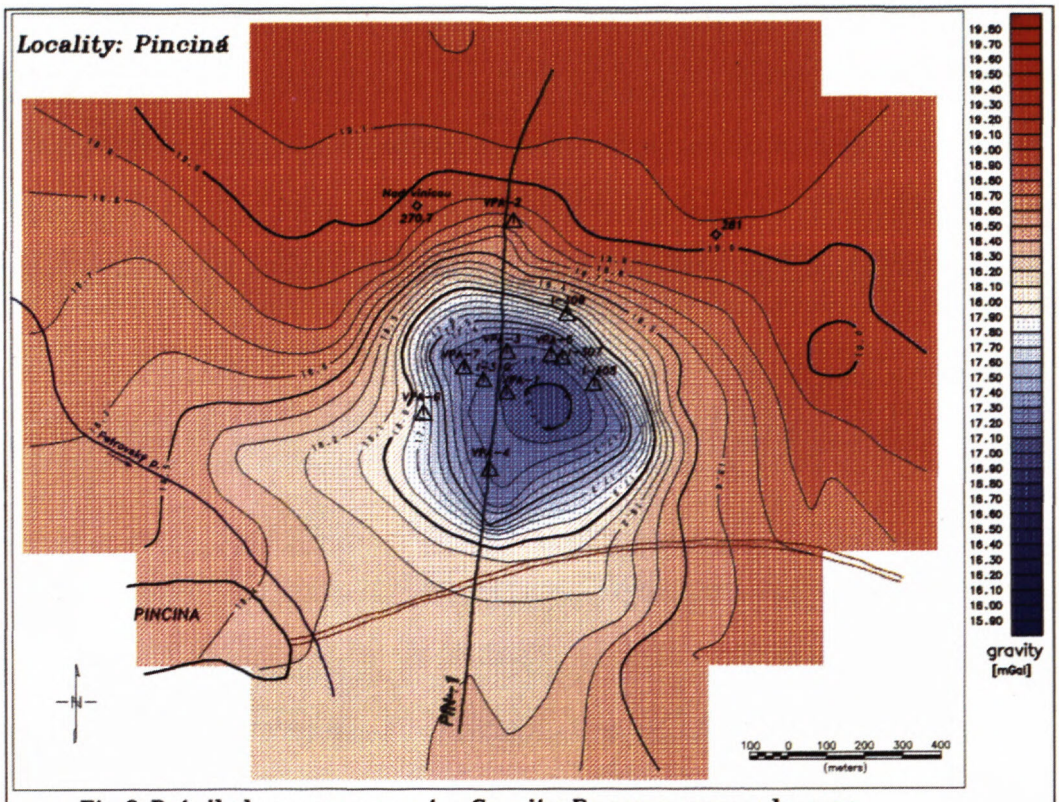


*Fig.9 Detailed measurements: Gravity Bouguer anomaly map density 2.00 g.cm³*
*Compiled by Šantavý J., 1997*

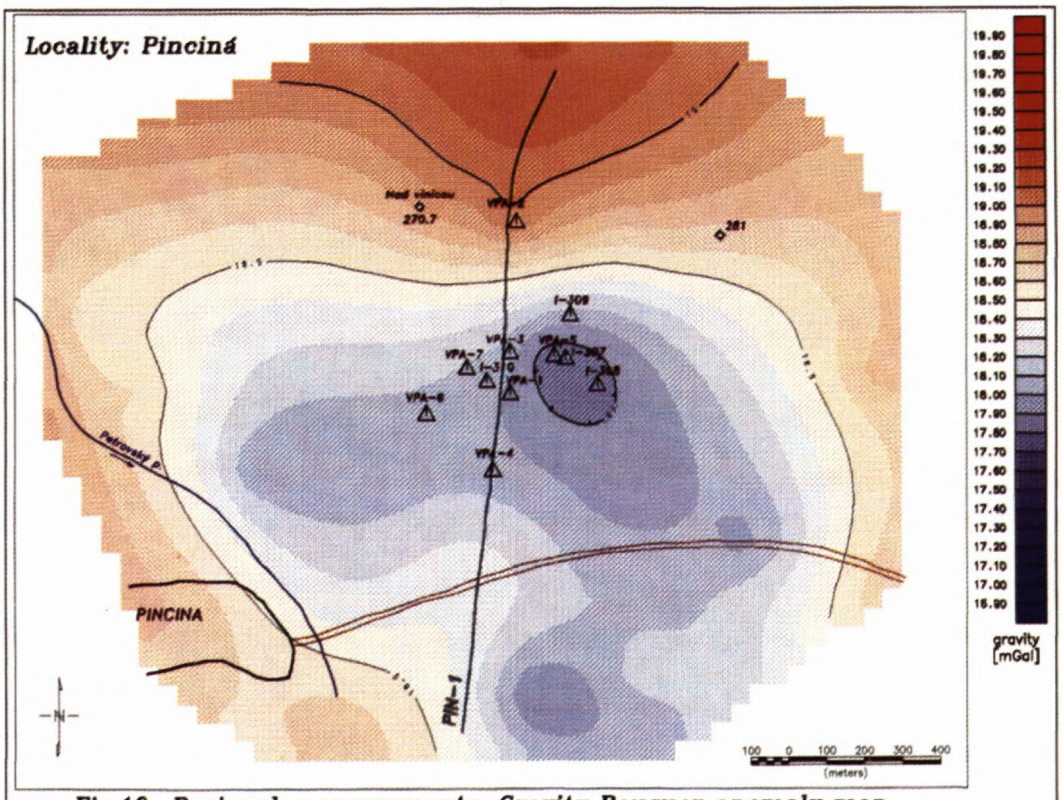


*Fig.10 Regional measurements: Gravity Bouguer anomaly map density 2.00 g/cm³*
*Compiled by Šantavý J.,1997*

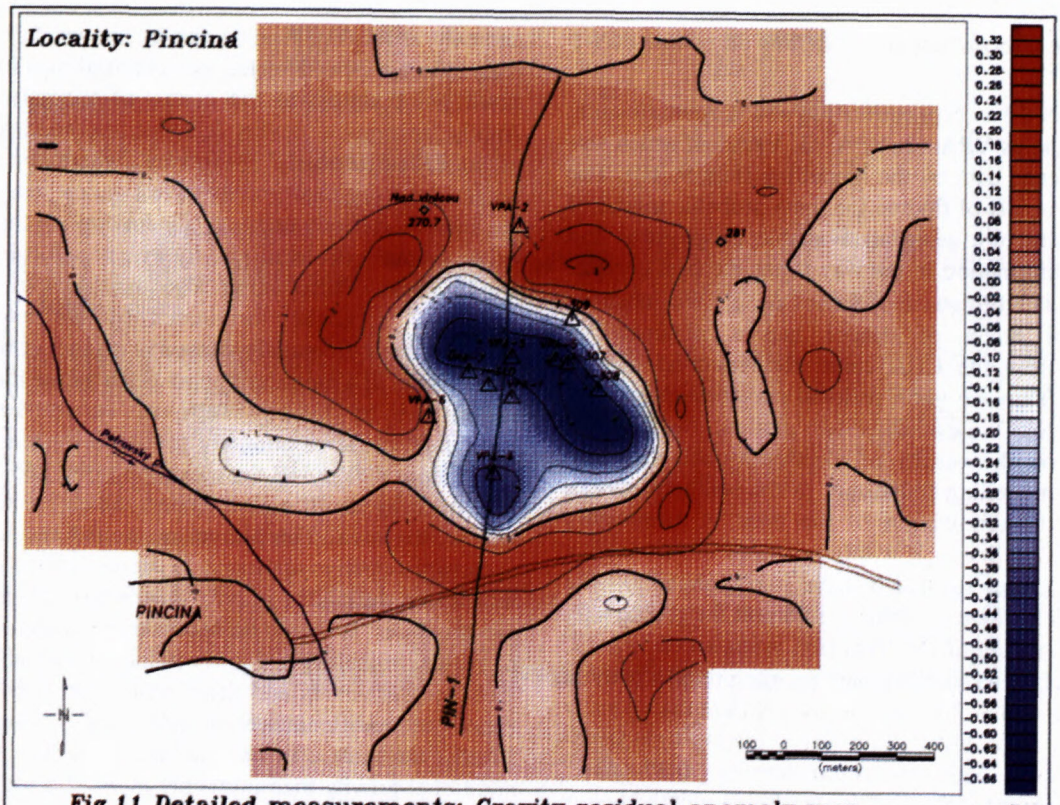


**Fig.11 Detailed measurements: Gravity residual anomaly map Griffin r=150 m, density 2.00 g.cm$^{-3}$ Compiled by Šantavý J., 1997**

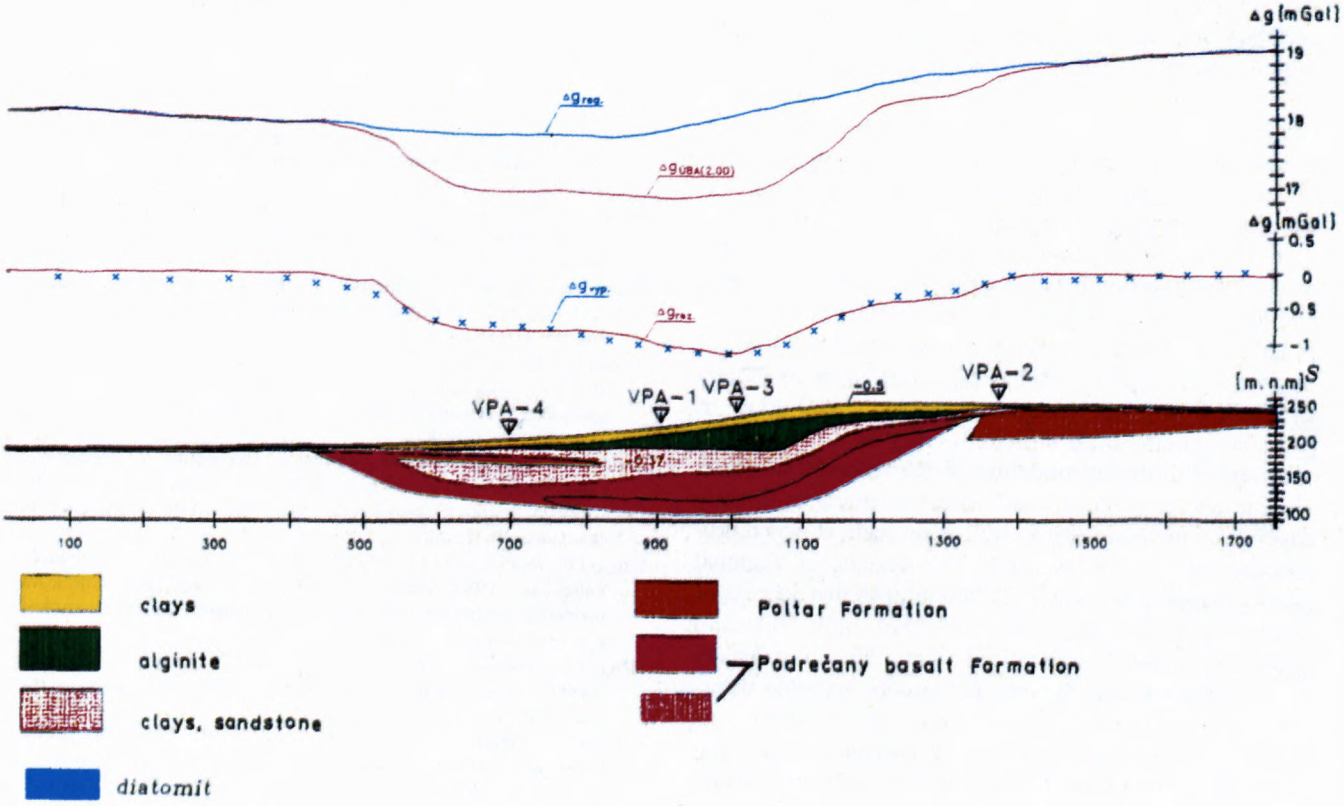


*Fig. 12 Locality Pinciná - gravity profile PIN-1 (Šantavý M. in Puchnerová M. et al., 1994)*

## Quantitative interpretation - modelling on profile PIN-1

Profil PIN-1 has N-S orientation and it approximately crosses boreholes VPA - 2, VPA - 3, VPA - 4, which served as reference data by the modelling (Fig. 9, 10). The modelling consists of finding the most probable model of geological structure using all available geophysical and geological information in the way it has the best correlation to calculated (measured) values $\Delta g_{rez}$. We substracted regional effect of deeper masses obtained by calculation from the CBA curve for reduction density 2.00 g/cm$^3$. Thus we got a residual curve which we further interpreted. We calculated two models. The first one was based on the results of electromagnetic measurements. The modelling found that in the southern part of the profile the termination of the maar is presumably about 150 m earlier as it is inferred from the curve $\Delta$ g UBA. This model seemed to have low probability. The second model, where we tried to interpret the problem by the geologic model of Dr. Konečný, is shown on Fig. 12. In this model, in its southern part we interpreted a body in the interval 600 - 800 m showing differential density +0.05 g/cm$^3$ as an tuffitic interlyer overlying light deposits of maar fill with differential density - 0.37 g/cm$^3$, which were absent in the first model. Subsequently, below this layer of maar deposits we consider volcanics comprising a basis of maar.

## Conclusion

It is remarkable that this ecological uncommon kind of raw material has not been discovered until 1995 in Slovakia. The discovery may be prescribed to application of suitable chosen methods of geophysical investigation as well as to knowledge obtained during a cooperation with Hungarian experts who have a long experience and positive results in solving this problematics. Slovak geophysics has completed Hungarian methodology and has introduced of the profile and areal gravimetry the first time for the prospection of oil shales - alginite. This method was testified for contouring of volcanic structures and considerable contributed to the knowledge of spatial distribution of sedimentary fill in maar structure. A considerable advantage of geophysical investigation is mainly contrasting physical parameters of individual kind of rocks comprising maar structure - ring as well as its fill with regard to the surroundings of the volcanic structure. The success to contour small volcanic structures directly depends on methodology as well as on scale at which prospection works are realized. The results of regional gravity mapping at scale 1: 25 000 implies that success of contouring of smaller structures (not only volcanic) depends on network of localized points. The point density 3 - 6 points per km$^2$ for regional gravity mapping determines average distance between points about 700 m. This is not sufficient for contouring of structures which diameter is smaller than 700 m. In this case the success depends on the accidental location of measured point into the area of anomaly - structure. As an example a negative anomaly at locality Pinciná can be shown. In this case a maar structure was detected as a result of location of measured point almost to the centre of anomaly structure. Another example - locality Jelšovec is a practical example of so called anomaly shadowing, where the measured point of the regional gravity investigation lies outside of the anomalous body causing "shadowing" of the local positive anomaly of detail gravity investigation by regional gravity field. There exists also examples of so called "artificial" anomalies originated by the interpolation. They are most frequently caused or triggered by an erroneous value of a measured point. Such a case it is possible to observe at locality Jelšovec where there is an "artificial" negative anomaly triggered by one measured point about 400 m NE the elevation point 265 m. This is confirmed by detail measurements which are not anomalous in the area. The above mentioned results imply a necessary selective approach to the individual anomalous indications during the gravity prospection of small geologic structures. It should be necessary to keep the principle of critical reliability and to try to increase reliability of the local measurements and their interpretation by an application of additional geophysical and geological data. Finally, the most efficient method is realization of detail additional gravity measurements as we successfully realized at the mentioned localities.

Finally, I would like to state that the application of both detail areal and detail profile gravity investigation brought very positive results for contouring of structures perspective for occurrence of oil shales - alginitu and because of this we recommend application of detail gravity measurements for further prospection works at perspective localities of similar thematic orientation.

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## • *Discussion*

# Central Carpathian Paleogene and its constrains: replay to Gross & Filo' and Potfaj's comments

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The reviewed work Soták et al. (1996) is a part of selected publications originating as products of the project "Geodynamic model of the Western Carpathians". The publication represents the only work from the Paleogene sedimentary cycle in the West Carpathians resulting from this project. The paper sources from the themes solved on the mentioned project (not on the project Levoča Mts. - this project is identified by the authors of the review as their credit which is acceptable without doubts). The paper of Soták et al (1996) has been aimed at the search of ophiolite detritus sources in sedimentary formations of the Western Carpathians. The solution of this specific problematics in the eastern segment of the Central Carpathian Paleogene is introduced by a brief outline of lithofacial development and paleogeography of the Levoča Basin, which is analysed by detective approach by the authors of the comments (Gross & Filo and Potfaj 1998, last volume - l.v.). Unfortunatelly, as became a practise, the atitudes of the authors are not restricted for correct professional polemic but also for non-standard reactions. It is obvious that the shortened version of the paper Soták et al. (1996) does not give a room for a complete bibliography of the works dealing with Central Carpathian Paleogene, and it only selects works directly related to the topic of the study - e.g. to the petrofacial composition of flysch arenites and paleogeography of sources. Therefore it is not substantiated to consider selection and effectivity of references as an ignorance and violation of ethics (international periodics require it strictly). On the contrary the real information is often hidden and misrepresented by reference to the manifold (mainly own) group citations (see below). According to the authors who require to refer everything and about everything, hardly any paper would not violate the ethics. Even the omision of the nomenclature of lithostratigraphic units (Gross et al. 1984) is not possible to consider as non-ethical and it is not due to its complex rejection (they are well-founded in maps), but because they express a dominant lithology and not vertical and facial changes in deep-sea fan environments. Yet Zuberec Formation with megaturbidites near Dolný Kubín totally differs from Zuberec Formation in the development of thin to middle rhythmical turbidites (e.g. near Revištné, Oravice, Upper Torysa river in Levoča Mts., etc.) from the sedimentological point of view. Similarly Huty Formation characterized by a predominantly mudstone lithology includes of mudstone subflysch deposits but it also includes mudstone facies from various deep-sea fan zones (e.g. slope mudstone drapes, bypassing muds, interlobe levee deposits, basin-floor facies, etc.). On the contrary deposits of different lithology may occur in the same lithostratigraphic unit. For example, Biely Potok Formation prevailingly consists of conglomerates in the slope part of the deep-sea fan (Šariš Upland) while in the middle deep-sea fan zone it consists of massive sandstones of progradational lobe and suprafan sequences sometimes interfingered by levee mudstones (e.g. in profiles in the area of Upper Torysa river in Levoča Mts., in Orava region - Gross et al. 1993, p. 112). Therefore, it seems an actual need to redefine the formations of the Central Carpathian Paleogene (in the valid nomenclature of Gross et al. 1984) as depositionary systems of the deep-sea fans ("facies tracts" - cf. Mutti 1992). Then they will be fully acceptable not only from the viewpoint of descriptive lithology but also from the viewpoint of genetic classification.

By the description of sequences in the introductory part of the paper terms Tomášovce and Kežmarok Beds were mentioned as a potential lithostratigraphic members, which have been submitted for formal acceptance (by P. Gross and I. Filo). The mentioned terms were identified with marking of certain type sequences. The mention about them in the paper (only in explanatory brackets) should serve for elucidating described facies types from the viewpoint of their prepared nomenclature without any attemps which should evoke an impression of introduction of a new lithostratigraphic units. The reaction to this marginal remark is therefore inadequate and in the slightest it can not question the Gross and Filo's authorship of above mentioned lithostratigraphic units (the paper of Filo & Siráňová 1996, in which they codified Tomášovce Member, could not be referred because it has issued at the same time as the paper Soták et al. 1996). The term Kluknava Beds, was not taken only from Andrusov (1965, Kluknava "development", p. 246 – in Gross & Filo, l.v.) but also from Marschalko (1978, Kluknava Formation, p. 59), who´s proper description appears to be sufficient for lithostratigraphic acceptance (or for supplementary identification). Moreover, the name of Markušovce Beds proposed as a new lithostratigraphic unit by Filo et al. (1995), was identical to the unit defined in the Permian of the Northern Gemericum (Novotný & Mihál 1987). The purity of nomenclature, pressed by authors, is correct and they should try themselves to publish it as required the rules of stratigraphic nomenclature (the reference to manuscripts is not acceptable, e.g. Gross & Filo referred to 27 references and 14 of them are manuscripts). Only in this case they would be entitled to do such critique which otherwise only an evocation of feelings and useless search of conflicts there, where do not exist, It is particularly sorry, that Gross & Filo (l.v.) are representing their opinion as a state of collective opinion and institutional attitude. By this they for example suvereignily miss the problems around so called Šariš "Oligomiocene" and new results of stratigraphic and sedimentologic researches at Geological Survey of Slovak Republic (Janočko 1998, Janočko et al. 1998, etc,).

Much usefull are "matter-of-fact" comments (mainly in review of Potfaj), on which it is possible to response directly. Gross & Filo (l.v.) object against the Lower Oligocene age of transgression in the southern flank of the Levoča Basin, which is partly represented by Tomášovce Member. Their opinion is based on data of biostratigraphic research with numerous references to works on flora, macrofauna, microfauna and nanoplankton. The only data on macroflora from the eastern part of the Central Carpathian Paleogene basal formations originate from Hazslinsky (1852). They are from Radačov Sandstones

(Tomášovce Member sensu Filo & Siráňová 1996) and suggest Oligocene age. According to Němejc (1961) the macroflora from the surroundings of Radačov, but also from the other sites of the Central Carpathian Paleogene (Kluknava, Víťaz, Hrišovce, Smižany and others) contains of the Early Tertiary elements. He even does not exclude an Early Oligocene age. Macrofauna of the basal lithofacies deposits (Volfová 1962) consists of shells containing euryhaline species in the stratigraphic range Eocene - Oligocene, and that younger than in the assemblages of the Central Carpathian Paleogene of the Western Slovakia (p. 26). Similarly, the mollusc fauna of the Odorin limestones (planorbide gastropods) resembles the Lower Oligocene fauna of the Moutnice limestones of Ždánice Unit (Volfová 1963). The Upper Eocene age of transgression is according Gross & Filo (l.v.) also limited by the age of the Huty Formation overlying Tomášovce Member, which is mainly based on foraminifera data. Recently the foraminiferal fauna of Huty Formation has been studied by Samuel (1995) in the report resulting from the project Levoča Mts. at the Geological Survey of Slovak Republic. From the Huty Formation Samuel (l.c.) referred also foraminiferal species which appear from the base of the Oligocene (*Globigerina postcretacea*, *G. cf. tapuriensis* and *Turborotalia cf. densoconvexa*, p. 18); other Oligocene foraminifers, such as *Cibicides lopjanicus*, *Plectofrondicularia hauerina* and *Amphimorphina hauerina*, are common for example in Kiscelian Clays (p. 19). Further O. Samuel writes (p. 19): "Based on this reality we can state that the main part of the Huty Formation was deposited in the Lower Oligocene, while it is not possible to exclude that their lowermost part interfinger with the upper part of the Borové Formation" (e.g. Tomášovce Member). In order to support the Upper Eocene age of the Huty Formation Gross & Filo (l.v.) also introduce data on nanoplankton, mainly from the results of Raková and Korábová-Žecová. In the report of the second author (Žecová 1995, p. 1) it is stated that the analyzed assemblages of nanoplankton from the Huty and Zuberec Formations suggest Eocene up to Early Miocene age. At the same time the deposits of Huty Formation from the southern and northern part of the Levoča Mts. and Šariš Upland (localities Jablonov, Dačov, Šariš castle, Podolinec, Malý Šariš), which were analyzed by the author, yielded Lower to Middle Oligocene age. Results of Žecová (1995) are fully consistent with results of A. Nagymarosi, B. Hamršmíd and L. Švábenická, who studied nanoplankton of the Central Carpathian Paleogene of the Levoča Mts. (in report of Soták et al. 1996). The mentioned authors proved the Lower Oligocene age of the Huty Formation deposits also at localities Torysa, Jakubany, Rožkovany, Nižný Slavkov and others. The deposits of Huty Formation from Hornád Depression are mostly barren of nanoplankton (or they only contain stratigraphically unimportant forms), except the locality Dravce and Jablonov, where the Upper Oligocene age of deposits was proved (similarly as Žecová 1995). A similar consistency can be observed in results of Žecová (1995) and in the report of Soták et al. (1996) concerning determination of the Upper Oligocene formations and even in the registration of nanoplankton elements with affinity to the Early Miocene species (*Helicosphaera scissura*, *H. kamptneri*, *H. cf. carteri*, *H. cf. ampliaperta*, *Reticulofenestra cf. pseudoumbilicus*, *Triquetrorhabdulus cf. carinatus?* and others). It is mostly peculiar that the mentioned information on foraminifers and nanoplankton from the reports of Samuel (1995) and Žecová (1995) does not appear anywhere. On the contrary Gross & Filo (l.v.), who know best these reports, put forward their results as evidence for the Upper Eocene age of the Huty Formation. It is especially contradictory in the context of their suspicion concerning ignorance of results and misleading in rendered informations. To argue by reports of Samuel (1995) and Žecová (1995) in favor of the Upper Eocene age of the Huty Formation is possible only in conviction that these reports are restricted for internal use. But at the same time the authors point to the inaccessibility of reports of Soták et al. (1996), which are deposited in archives (Geofond - GS SR, Nafta Gbely, Nafta-východ Michalovce and GU SAV), where they are frequently look up for informations (e.g. Potfaj l.v.).

The most serious reluctance concerning the Upper Eocene transgression in Levoča Mts. stems from the fact of the absence of large foraminifers in deposits of basal formation (including Tomášovce Member). Gross & Filo (l.v.) offer an explanation that this is result of "decreasing salinity and muddy-rich nearshore water caused by rivers". The explanation is only hardly acceptable because the deposits of Tomášovce Member are not barren of fossils, on the contrary, they contain marine and relatively aboundant foraminiferal fauna but without nummulites. Foraminifera of Tomášovce Member mainly consist of representatives of *Rotallidae*, *Chiloguembelinidae*, and *Heterostegínidae?*. The marine origin of Tomášovce Member is also documented by abundance of crinoidal ossicles and more seldom fragments of coralline algae. Even the frequently referred fresh-watering of Odorín limestones is not substantiated because they even contains an admixture of foraminiferal plankton (neither the basal sediments of Paleogene nearby Veporic mainland, for example nummulite limestones in the Upper Hron area, are not fresh- and muddy-watered). The fresh-watering is not an explanation for the absence of nummulites in the Spiš segment of the Central Carpathian Paleogene because its southern margin is not a real coast line of the marine transgression, which undoubtelly extended more southerly. It is for example documented by findings of bathyal ichnofossils *Zoophycos* directly in the deposits of the basal lithofacies (locality Spišské Vlachy, Plička 1987) and also the development of the overlying deposits, which according to Marschalko (in Soták et al. 1996) shows immediately from the base a distal features (no marks of a river influx from the close mainland). Therefore it is necessary to accept, that the age of the basal deposits of the Levoča Basin in the Hornád Depression is younger as the range of the Eocene nummulites.

Gross & Filo (l.v.) also requires in the case of Kežmarok Beds an evidence of its Upper Oligocene age. Kežmarok Beds represent the uppermost part of the Zuberec Formation (Gross et al. 1996). Samuel (1995) refers foraminiferal assemblage of mostly Oligocene age from the Zuberec Formation and he states: "Because Kežmarok Beds form a sort of „transitional" strata between Zuberec and Biely Potok Formations we can consider their Oligocene age on the basis of superposition ..." (p. 22). Except of Poprad Depression the Kežmarok Beds are also developed as thick-rhythmical flysch deposits in the NE part of Levoča Mts. (e.g. in profile of Upper Torysa river area - Janočko et al. 1998), where their Upper Oligocene age is clearly documented by nanoplankton of zones NP 24 - 25 (Soták et al. 1996, Janočko et al. 1998). In context of stratigraphy of the Levoča Mts. it is possible to skip certain reminiscence. A success of the stratigraphic studies in the project of Levoča Mts. is essentially an agreement in opinion about younger age of formations in this part of the Central Carpathian Paleogene which has been achieved by independant approaches. It is only sorry that there is not a back reflection on periods of strict rejections of any ages younger than Lower Oligocene. So the denouement of these rejections is just here. At this time the new stratigraphic interpretation of the Central Carpathian Paleogene of the Podhale Basin has been published (Olszewska & Wieczorek 1998, Fig. 1). The results are fully in accordance with those obtained from the Levoča Basin (Soták et al. 1996, Soták in this volume),

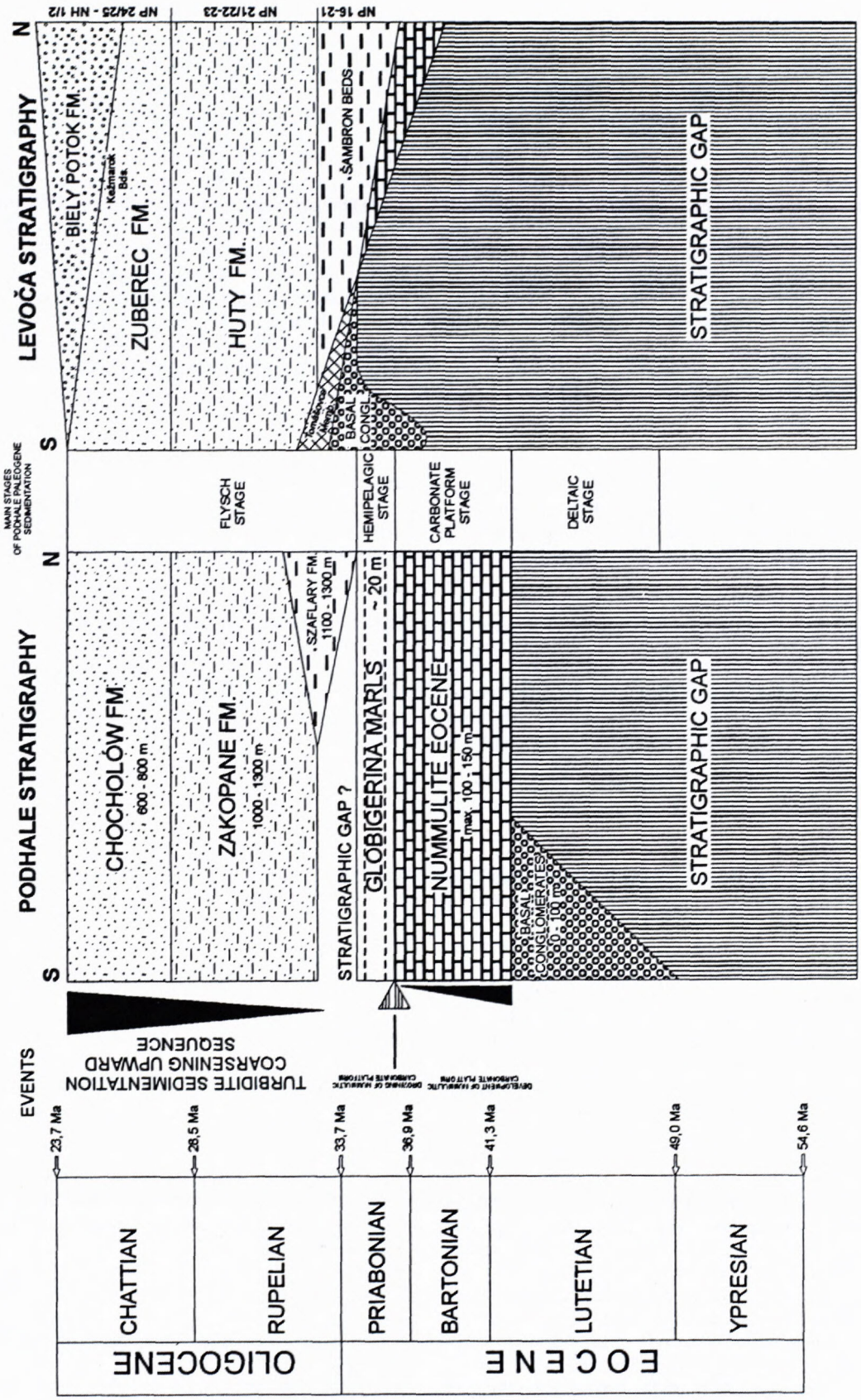


*Fig. 1 Correspondence of Podhale stratigraphy and Levoča stratigraphy in new results of micropaleontological investigations. Podhale stratigraphy according to Olszewsk-a & Wieczorek (1998). Levoča stratigraphy as is interpreted in the works of Soták et al. 1996 a, b, Soták in this volume (Tomášovce Member sensu Filo & Siráňová 1996, Kežmarok Beds sensu Gross in press).*

which have been refused and critized as unacceptable. Refusing these results, the Podhale stratigraphy becomes new and applicable for whole Central Carpathian Paleogene.

In the case of Kežmarok Beds Gross & Filo (l.v.) also question its distal development. But the reasons introduced by them (thick-rhythmical character, Ta stratification of sandstone beds) are not decisive from the viewpoint of distality or proximality. The Kežmarok Beds are developed in the fringe part of the depositional lobe area within the deep-sea fan system of the Šariš Upland and Levoča Mts., e.g. in the lower part of the middle fan. Its distality is already obvious from the paleocurrent orientation of this deep-sea cone, prograding in direction from SE to NW (Marschalko 1981). The thick-rhythmical character of the Kežmarok Beds does also not contradict it because fringe lobes gather clastics winnowed from the distributary channels of the middle fan. Therefore the Kežmarok Beds, representing deposits of concentrated suspensions, are developed in C facies types of Mutti (1979) as graded and massive sandstones, but with the less conspicuous development of progradational cycles (e.g. nearby Holumnica, Toporec and elsewere), rare appearance of channel erosion etc. The distality and deep water deposition of the Kežmarok Beds is at best documented by presence of ichnofossils *Taprhelminthopsis sp.* and *Fucoides graphicus* (directly at locality Kežmarok).

The main part of review of Gross & Filo (l.v.) and Potfaj (l.v.) is related to the problem of serpentinitic sandstones of the Šambron zone. Their findings are a success of the petrography of flysch arenites in the Western Carpathians, highly valued for example on international excursions. Already the primary information from the paper Soták & Bebej (1996) is interpreted erroneously. According to Potfaj (l.v.) the authors described "Šambron Beds with serpentinitic sandstones". This formulation does not occur anywhere in the mentioned paper because it is not precise. Soták & Bebej (1996) described serpentinitic sandstones from the Šambron zone from the Upper Oligocene formations younger than Šambron Beds. All following questions concerning origin of serpentinitic sandstones and modeled reconstruction of the Central Carpathian Paleogene fans stem from this misunderstanding. Thus, the authors of review lose this context and therefore the submitted reasoning seems dubious for them. Accordingly, the interpretation of suprafan seems to be dubious for Potfaj (l.v.). He writes that a suprafan is a morphological terms which can not be used for identification of development stage of deep-sea fan. The term suprafan does not defines only morphology (lobate-shaped bulge - Normark 1970), but also a depositional model for accumulation of sandy-rich fans. Because progradational types of fans are characterized by increasing rate of sandy deposition, their development commonly terminates in the sandstone accumulation of suprafan. Also according model of Mutti (1979) a suprafan progrades over deposits of basin plain, firstly by successions of sandstone lobes passing into non-channeled and later into channeled deposits and finally, into deposits comprising a sedimentary fill of the feeder channels. Such a model of suprafan is fully acceptable for reconstruction of deposition in deep-sea fans of the Central Carpathian Paleogene (similarly like Janočko et al. 1998). The sandy-rich formations of suprafan in the Central Carpathian Paleogene terminate a cycle of the deep-sea fan accumulation, but they probably not terminate the entire deposition (vitrinite reflectance data from the near-surface sediments of the Levoča Mts. respond for about 2 km thick of unknown overlying strata - Soták et al. in press).

Potfaj (l.v.) is also questioning a fact that the reviewed works do not deal with occurrence of conglomerate bodies in the Šambron Beds. In the work Soták et al. (1996), where Potfaj (l.v.) does not see any mention about conglomerates of Šambron Beds, it is devoted almost the whole page 346 to them (and to breccias). The absence of data on conglomerates in the work Soták & Bebej (1996) is logic because they do not occur in the described sequence of serpentinitic sandstones (only one layer of revorked autoclasts, Fig. 4) and their sequence is even not a part of the Šambron Beds. However, in the lithostratigraphic scheme of the Paleogene formations of the Šambron zone (Fig. 1) the Šambron Beds with intraformational conglomerates, are conspicuously marked. Further, Potfaj (l.v.) makes a query about the sense of commonly used term of Šambron zone (Šambron-Kamenica zone - Marschalko 1975 and elsewhere). By this meaning the author also determines position of conglomerates and suggest a possibility, that they does not have to be unconditionally a part of the Šambron Beds. At the same time the conglomerate bodies of the Šambron Beds are well exposed in the whole extent of the Šambron zone (from Šambron to Jakubovany or Pavlovce, respectively) and drilled by numerous boreholes (e.g. Šambron - 1, Plavnica - 1, Lipany - 2, 3, 6). Besides of the intraformational conglomerates of the Šambron Beds there are also different types conglomerate deposits which position is not so clear.

The term Šambron zone marks a belt of tectonically disturbed flysch sediments of the Central Carpathian Paleogene at its boundary with the Pieniny Klippen Belt. The Šambron zone is an antiform structure cored by the Eocene formations (Šambron Beds with intraformational conglomerate bodies, in boreholes Lipany - 6 as much as 4000 m), in its depressed parts the Lower Oligocene deposits of the northern mudstone lithofacies (Huty Formation) occur. In the slice belts of the Šambron zone also Upper Oligocene deposits occur occasionally, which exibit a thin-rhythmical character and have a specific detrital contents (serpentinitic sandstones, sandstones with tuffitic admixture). These formations do not have clear analogues in the Central Carpathian Paleogene (for example according to personal information of T. Ďurkovič they resemble Krosno Beds). The Upper Oligocene age of these deposits is erroneously related to Šambron Beds by Potfaj (l.v.) resulting in contradiction with their definition, correlation with Szaflar Beds and also with older data of biostratigraphical researches.

Potfaj (l.v.) also questions determination of the Upper Oligocene age of formation within the Šambron zone (erroneously understood as Šambron Beds) by nanoplankton. The Upper Oligocene age of these deposits was mainly proved by the occurrence of nanofossils *Cyclicargolithus abisectus*, which appear at the base of the biozone NP 24 (e.g. Krhovský & Djurasinovič 1992). However, Potfaj (l.v.) objects that this species also occurs in older sediments referring numerous supporting citations. Among first references it is a work of Dudziak (1993) here, who should describe an occurrence of *C. abisectus* from the Podhale Paleogene. But Potfaj (l.v.) missed a fact that the described species is not *C. abisectus* but *C. bisectus*. Similarly neither Oszypko (1996) refers species *C. abisectus* from the Upper Eocene and Lower Oligocene. The spesies, referred here, is marked by the author as *Cyclicargolithus aff. C. abisectus* and she writes „it has been described as transitional taxon between *Cyclicargolithes floridanus* and *Cyclicargolithes abisectus* occurred in NP 19-20 which is earlier than normal *C. abisectus* (p. 12). At the same time the author writes that „boundary between NP 23 and NP 24 is defined by first occurrence of *Cyclicargolithes abisectus* or *Helicosphaera recta* (p. 12). The following works referred by Potfaj (l.v.) in order to support the earlier apearance of the species *C. abisectus* are already studies from the Middle/Late Oligocene deposits of the zone NP 24 (for example from Štibořice Member - Bubík 1996, from the deposits of the Banská Bystrica Depression - Bystrická 1979). Neither the following data of Potfaj (l.v.) about syn-

onymity of the species *Cribrocentrum reticulatum* is not correct, because its synonym is not the species *Reticulofenestra lockeri* but the Eocene species *Reticulofenestra reticulata*. Therefore the objections against the Upper Oligocene age of the deposits containing *Cyclicargolithus abisectus* does not correspod to the real knowledge from the biostratigraphy of nanoplankton. At the same time their age determination is not based only on three species from the locality of serpentinitic sandstones, but on the knowledge about distribution of nanofossils in the Central Carpathian Paleogene of the Levoča Mts. (Nagymarosi, Hamršmíd & Švábenická in Soták et al. 1996), where the species *Cyclicargolithus abisectus* is absent in the Šambron Beds (zones NP 16 to NP 21) and also in the deposits of the mudstone subflysch (Huty Formation, NP 21/22 - NP 23). It appears as soon as in the deposits of the Upper Oligocene age (zone NP 24 and NP 25).

Potfaj (l.v.) and Gross & Filo (l.v.) widely comment a figure from the work of Soták et al. 1996 (Fig. 2) reproaching its basic shortage. The figure is a modeling construction of the eastern branch of the Central Carpathian Paleogene and it should illustrate a longitudinal orientation of a deep-sea fan (according to Shanmugan & Moiola 1988 typical for active-margin fans), its marginality to eastern sources, perisuture position of the northern margin of the Central Carpathian Paleogene etc. Thus, it is very complicated reconstruction of the collisional orogen system depicting the subduction of complexes beneath the orogen (underplating), collapse of the plate margin in the zone of synconvergent extension, oblique plate convergence on collisional boundary, buoyancy of the subducted complexes in the strike-parallel wrench zone, accretion in the trench zone etc. It is natural that there is no model reconstruction without errors, the reconstruction can only be more or less successful. The model is actually a proposal of some solutions which may or may be not accepted. In this way it by rule installs more questions as it solves itself. But the authors of these reviews require definite solutions without making more clear all interdependences.

The most conspicuous feature of the critized model is reconstruction of the deep-sea fan consistently with paleocurrent directions proved by Marschalko (1966, 1981) and confirmed by a new paleocurrent research of Soták et al. (1996). The researches from Podhale (Radomski 1958, 1959, Krysiak 1976 and others) does not solve a presented model, because they document opposite paleocurrent systems. Inspite of a clear depicted SE - NW longitudinal system of fans in the scheme, Potfaj (l.v.) criticizes its absence. The further reasoning of Potfaj (l.v.) is erroneous because he changes suggested alternatives for the detrital origin of the Šambron Beds ("from slopes of the active Central Carpathian plate or from its northern collisional edge") with the longitudinal system of the Upper Oligocene deep-sea fans. The works from the Polish Podhale (Mastella 1975, Mastella et al. 1977, 1978) referred by Potfaj (l.v.) do not clearly prove the presence of the klippen material in Szaflar Beds. Mastella et al. (1977) only writes about transport directions with orientation 70 - 80$^{O}$, mostly from the west (p. 495).

In both comments the problems connected with the position of the Klippen belt are shown. Its position within the collisional system is a puzzle always suggesting a lot questions, which can be as follows: Is a material contrast of the Šambron zone and Klippen belt conglomerates an evidence for juxtaposition of originally very distant units? Were they bringing together due to ocean reduction south of the Klippen belt (an alternative of Iňačovce-Krichevo subduction and perisutural position of the Central Carpathian Paleogene Basin is well illustrated by Kováč et al. 1994)? Is it possible that the Klippen belt with some affinities to the North-European platform (e.g. Klippen belt fauna of *Cardioceratidae* in the more southern units not known - Kutek & Wierzbowski 1986, heavy mineral associations with pyrope-type garnets which are similar to those in eclogites of the Bohemian Massif and not in the Central Carpathian units - Aubrecht et al. 1997, termination or shallowing of the Czorsztyn succession during the regional regression on the platform etc.) would comprise a collisional margin of the Central Carpathian plate? Is the Klippen belt a remnant of a narrow and 800 km long continental ribbon like as "seamounts" or it is an annected part of the North European platform formed as soon as in onset of the continental collision and strong transpression on the plate boundary? Is it possible that Magura Unit as a largest trench-type unit of the Western Carpathians would be accreted from the platform? One of the possible solution of these questions is also integration of the Oligocene basins of the central and external Carpathians without more conspicuous morphostructural expression of the Klippen belt, which is illustrated by criticized scheme. Gross & Filo (l.v.) object against the neighbouring of two sea basins not separated by mainland because Central Carpathian and Magura realm represent according them two marine provinces with the entirely different formations and with different depositional regime. During the Oligocene the Magura basin and Klippen belt occurred in the depositional area of the Malcov-Menilite serie. The facial development of the Malcov Beds in the depressed parts of the Klippen belt (e.g. in Ujak-Plaveč area) is already very similar to the Central Carpathian Paleogene formations (Uhlig 1890, Nemčok et al. 1990, p.63). It conspicuously resembles for example the Lower Oligocene sediments of the mudstone lithofacies in the Šambron zone (Huty Formation). In the area of Nowy Targ in Poland the Malcov Beds were even originally mapped as Zakopane Beds of Podhale Paleogene and they also exhibit the same paleotransport patterns (M. Cieszkowski pers. communication). On the other hand the influence of the Malcov - Menilite type deposition of the Outer Carpathians is also manifested in the Central Carpathian Paleogene (Soták this volume) whether they are Menilite Beds (Leško 1960 a), *Globigerina* Marls or Šariš Beds strikingly resembling lithofacies of Krosno Beds (Leško 1960 b). It is also remarkable that the paleocurrent system of the Central Carpathian Paleogene is not reoriented even in the close proximity of the Klippen belt, it tends obliquely to it with orientation to NNW (for example between Sabinov and Ratvaj). Therefore an idea of an undissected basin covering the margins of the Central Carpathian plate, trench zone and areas of Krosno-Menilite sedimentation is acceptable for the Oligocene paleogeography, which is shown on the criticized scheme. The interpretation of the Central Carpathian Paleogene comes out from the same idea in the work of Nemčok et al. (1996). He sees in it a "proximal facies of the West Carpathian Flysch Belt" (p. 321).

In the discussed model also an existence of the oceanic crust in the subducted substrates of the flysch units is questioned by Gross & Filo (l.v.). But based on the today's ideas it is clear that without oceanic crust the space shortening, which is enormous in the Western Carpathians, is not possible, and just on the detriment of the northern flych units. On this basis not only individual approaches but also research teams of international projects work (e.g. PANCARDI). Just findings of serpentinitic sandstones in proximity of the main suture related to the boundary of the Central and Outer Carpathian units (Soták & Bebej 1996) support an existence of oceanic crust. Their importance can not be lowered by their alleged sporadic occurrence (Gross & Filo l.v.) because ophiolite traces are also on big suture zones mainly preserved in form of residual detritus (for example Cr-spinels) and a rare occurrence of serpentinitic sandstones already directly indicates a proximity of serpentinite belts (see Arai & Okada 1991, Okada 1964, Dickinson 1982, Zimmerle

1968, Critelli 1991, 1993, Wagreich 1993 and elsewhere). Moreover, the occurrences of serpentinitic detritus in the Šambron zone by far are not already sporadic. Besides own localities of serpentinitic sandstones nearby Kamenica (Lipany creek, Slané Mláky, Putnov creek), numerous clasts of serpentinites were recorded nearby Šambron (borehole PU-1), Lipany (borehole Li - 6), Hanigovce and Pavlovce. Even more conspicuously the concentration of ultrabasic source detritus in Šambron zone is indicated by heavy minerals consisting of high amount or often even monoassociations of Cr- spinels (Soták et al. 1996). Similarly the Šambron zone is manifested by regional results of heavy mineral concentrate analysis (Križáni 1985) as a conspicuous Cr-spinel anomaly extended as far as Spišská Magura. The distribution of the detritus is not restricted to the Central Carpathian Paleogene as referred by Potfaj (l.c.). The detritus of ultrabasic sources is also recorded in Malcov Beds by Cr-spinels in sandstones (e.g. locality Ľubotín - 16%, Soták et al. 1996) or in heavy mineral concentrates from Malcov-Richvaldy belt (Križáni 1985).

In context of ultrabasic detrital sources Gross & Filo (l.v.) pose a question whether it should be ..."an emerged mainland (trench) in area of Upper Oligocene Klippen belt". The formulation of the mentioned question is contraversial because a source does not to be a mainland and a mainland is not a trech. Soták et al. (1996) assumes a location of ultrabasic sources in front of the collisional margin of the Central Carpathian plate but not as an emerged mainland identical with the Klippen Belt. According to Gross & Filo (l.v.) is not clearly indicate the active plate margin on the figure, which in the illustrated model of synconvergent extension is outlined by marginal blocks tilted toward trench side (the plate margin and trench are also marked by names here). Potfaj (l.v.) is also critical from the viewpoint of mechanisms by which ophiolite detritus might be brought into sedimentary basin. He gives arguments that oceanic crust can not be exposed to subaerial sources (in his case always cordilleras - Potfaj 1997) and he writes "what a morphological and structural body would this source to be: - an intraoceanic peri-subductional cordillera? - such a form is not known to the actual geology, and is hardly interpretable". The expressions of Potfaj (l.v.) are surprising because an existence of forearc serpentinite seamounts forming 10 - 20 km wide and as much as 1 000 m high elevations consisting of ultramafic and mafic rocks (Charvet & Ogawa 1994) is known just from the geology of modern plate boundaries. Such seamounts are interpreted as either serpentinite diapires (Hussong & Freyer 1985) or serpentinite domes (Lagabrielle et al. 1992). The orientation of Potfaj to cordillera-type sources is one-tracked and it was not considered in serpentinitic sandstone sources described in works of Soták & Bebej (1996) and Soták et al. (1996). The topography of plate boundaries is yet even without cordilleras sufficiently contrast to derive subducting oceanic crust, whether by off-scraping, obduction, buoyancy and diapir penetration into sutures etc. Such sources for example might be elevated accretionary wedges giving material of the subducted plate to trench and forearc basin sides (cf. Cloos 1982). Similarly the serpentinitic sandstones of the Šambron zone were interpreted as distal to trench turbidites derived from oceanic crust slices imbricated on plate boundary.

The indications made by Gross & Filo (l.v.), concerning the general knowledge about the decreasing carbonate detritus toward the higher formations of the Central Carpathian Paleogene and increasing amount of siliciclastics are not enough factual. They need to be supported by a detail, in the case of the work of Soták et al. (1996) petrofacial reseach. Generally known facts have then not occur as valid by rule in detail (e.g. evidence of an amount of volcanoclastic detritus in the sandstones of the Biely Potok Formation or evidence of serpentinitic detritus in the youngest formations of the Šambron zone etc.).

In the conclusion of both reviews it is preffered a search of ultrabasic detrital sources of the Šambron zone in the Gemericum. According to Gross & Filo (l.v.) such an approach provides simplier solution and it is consistent with the SE - NW orientation of paleotransport (Marschalko 1978). However, the submitted simple solutions become more complicated by the existence of large displacements on transform faults, block rotations, amputation of sources from marginal facies, younger processes of subduction followed by upwelling and unroofing of underplated crustal segments etc. Gross & Filo (l.v.) argue by presence of serpentinites in clasts of Paleogene conglomerates from the surroundings of Margecany (Šalát 1954) and in the body nearby Sedlice. On the contrary Potfaj (l.v.) excludes the Sedlice body and he does not specify more closely the southern sources. The conglomerates located nearby Margecany probably represent continental deposits (they are red coloured - Šalát 1954, p. 208) deposited from alluvial aprons and local sources (Marschalko 1965, 1970). The clasts composed of Gemeric ultrabasic rocks occur in conglomerates nearby Margecany as well as Kluknava and Markušovce, however, they are metamorphic rocks with Cr-magnetite or Cr-spinels of brown colour. They entirely differ from the red, higher aluminous Cr-spinels of the Šambron zone. The conglomerates of the pre-transgressive formations were accumulated in valley-like depressions filled from SW to NE (Marschalko 1970) it means in the opposite direction than the deep-sea fan orientation of the Central Carpathian Paleogene which distal facies are significantly enriched in serpentinitic detritus. The clasts of ultrabasic rocks does not occur in slope conglomerates of the deep-sea fans in the eastern part of the Šariš Upland. Neither the main volume of deep-sea fan accumulation in the Levoča Mts. consisting of siliciclastic deposits with dominance of garnet-turmaline association in heavy minerals does not contain it (in contrast to Cr-spinel association of the northern zone, Soták et al. 1996).

To solve the provenance of the serpentinitic detritus of the Šambron zone from the Gemeric sources is also problematic from the viewpoint of distance. The findings of serpentinitic sandstones in the Šambron zone are far off about 50 km from these occurrences in Gemericum. In the work Soták & Bebej (1996) a local origin of the ultrabasic detritus of the Šambron zone has been emphasized regarding to the fragility of serpentinitic clasts which would be not preserved in such concentrations during the long-distant transport in the high dispersive pressure flows. The occurrence of Cr-spinels in the Spišská Magura Mts. (Križáni 1985) as well as its easternward tracing up to the Transcarpathian Ukraine (Kruglov 1974) is at all not possible to solve from the position of Gemeric sources. For example, the body of ultrabasic rocks nearby Sedlice, mentioned in the reviews, is not even in its closest surroundings accompanied by the occurrence of serpentinitic sandstones (only scarce spinel grains were recorded here), ultrabasic rocks are even absent in the material of conglomerates in which the body is placed. On the contrary conglomerates cropping out in the close surroundings of the Sedlice body consist of such pebbly material which by no way can be derived from either the Gemericum or Mesozoic series of the Čierna hora Mts. (e.g. Urgonian limestones with *Orbitolina*, Lower Cretaceous shallow-water carbonates of the Brekov Limestone type, Majolika type of *Calpionella* limestones and so on, containing a small amount of Triassic dolomites and limestones, everything lacks of metamorphic recrystalization typical for the Mesozoic rocks of the Hrabkov serie). Also from this reason the position of the Sedlice body is uncertain and it is possible to consider it as a

more important source of Paleogene clastic deposits. As already was pointed out before (Hovorka et al. 1985) the Sedlice body is an exotic peridote klippe occurred within the Central Carpathian Paleogene, which is entirelly different of ultrabasic rocks of Gemericum (minimal degree of serpentinization, different structural character, presence of ophicalcite breccias etc.). It means that even the Sedlice body does not need to be derived from Gemericum. On the contrary they may be an ultrabasic rocks similar to those indicated by a conspicuous magnetic anomaly in the easternmost part of the Šariš Upland (Bzenov anomaly - Gnojek 1987) and to rocks penetrated by a drilling in a shallow depth nearby Prešov (borehole V-1, Slávik 1974). The detritus of ultrabasic rocks, essentially accumulated in Merník conglomerates, probably originated from similar bodies (Soták et al. 1990). Such ultrabasic bodies could be exposed in the Iňačovce-Krichevo Unit in the final phase of their exhumation (Early Miocene).

The attitudes of authors to the Iňačovce-Krichevo Unit lack clarification of problems. It is partly given by a real degree of knowledge on this unit where so far gap exists mainly in the matter of timing of the Iňačovce-Krichevo subduction, tectonic burial, metamorphism, exhumation and uroofing. However, the basic facts are known and in the main features depicted in the model reconstruction of Soták et al. (1996). Iňačovce-Krichevo Unit was formed within the subducting substrates of extratatric domain as a unit with oceanic-type crust and Jurassic-Cretaceous and Paleogene formations of shally-turbiditic lithology. During the subduction the complexes of the Iňačovce-Krichevo Unit were deeply buried and undergone metamorphism in temperature around 350 - 400$^{o}$ C (chloritoid shists) and in intermediate greenshist/blueshist pressure conditions (magnesioriebeckite + actinolite + epidode). The youngest metasedimentary formations of the Iňačovce-Krichevo Unit involved in subduction, are of Eocene age. Their incorporation into subduction-accretionary complex documents neoalpine age of the final phases of Iňačovce-Krichevo metamorphism. The uplift of subducted complexes was being realized in the Late Oligocene and reached a depth of zircon FT blocking temperature (ca 220$^{o}$C) around of 20 Ma (Dunkl in Soták et al. 1997). It means that yet in the Early Miocene the Iňačovce-Krichevo Unit occurred in the depth of some 5 - 7 km underneath complexes of tectonic hangingwall. The final phase of core complex exhumation of the Iňačovce-Krichevo Unit and its displacement to the East Slovakian Basin floor was probably controlled by the Middle Miocene back-arc extension realized through detachement faulting and unroofing. This conception of the Iňačovce-Krichevo Unit development has surely its weak points and unsolved questions but not in such a way as stated Gross & Filo (l.v.) and Potfaj (l.v.).

Gross & Filo (l.v.) and Potfaj (l.v.) are wrong in the basic idea concerning the position of the Iňačovce - Krichevo Unit during the Upper Eocene and Oligocene arguing why its exposed substrates should be not a source of flysch clastics of the Central Carpathian Paleogene. Yet in the work of Soták et al. (1996) and elsewhere (Soták et al. 1993, 1994, 1995, 1996b, 1997) it is emphasized exactly the reverse side view consisting of a deep burial of the Iňačovce-Krichevo Unit in subduction zone evidenced by metamorphose of the Eocene formations (Potfaj l.v. is reasoning very consistently with this idea and it is not clear what is the problem). Therefore the material of the Central Carpathian Paleogene sediments is not derived from the Iňačovce-Krichevo Unit but from its hangingwall substrates probably consisting of Central Carpathian units. The force of these sources resulted probably from exhumation activity of the Iňačovce-Krichevo Unit in a mid-crustal depth during the Late Oligocene. The occurrence of rocks resembling rocks of the Iňačovce-Krichevo Unit in the conglomerates of the Šambron Beds was mentioned in the work of Soták et al. (1996) not in connection with its exhumation (exposing of substrates) but with its subduction (underthrusting of substrates). The occurrence of these type rocks in the Šambron Beds is a reality which can not be disparge by their ubiquitous occurrence (as far as Horná Nitra region - Gross & Filo l.v.). The rocks are specific types of green, dark and cream-brown phyllites, calcphyllites and light-coloured marbles (not gneisses, amphibolites, quartzites as referred Gross & Filo l.v.) which affinity to rocks of Iňačovce-Krichevo Unit is for example manifested by scaly fabrics, small-scale folding, crenulation by cleavage systems, entirely recrystalized primary structures (e.g. on the contrary to the Veporic Mesozoic rocks of Hrabkov Serie) and petrographic composition. A possible explanation of the occurrence of these rocks in the Šambron Beds (perisutural zone) is their retrieval from the subduction scar driven by buoyancy flow similarly like glaucophanites or eclogites appear in blocks of sedimentary melanges.

Gross & Filo (l.v.) question the significance of the Iňačovce-Krichevo Unit from the position of "complete hiding" and only recovering in sporadic boreholes. The approach presented does not appreciate data from the deep structure research. The interpretation of the surface geology without these data is not possible. Yet it is sufficiently known the complexes of the Iňačovce-Krichevo Unit were recovered by a series of 24 deep boreholes in the following sites: Iňačovce-3 (1334 m), Iňačovce-2 (30 m), Iňačovce-1 (10 m) Zbudza-1 (974 m), Lesné-1 (965 m ), Lesné-2 (538 m), Bunkovce-1 (862 m), Michalovce-1 (360 m), Vysoká-1 (100 m), Pavlovce-1 (500 m), Senné-1 (103 m), Senné-2 (608 m), Senné-8 (291 m), Blatná Polianka-1 (302 m), Blatná Polianka-2 (80 m), Blatná Polianka-3 (90 m), Hrušov-1 (155 m), Rakovec-3 (202 m), Trhovište-26 (117 m), Pozdišovce-1 (281 m), Rebrín-1 (106 m), Lekárovce-1 (458 m), Pinkovce-1 (100 m), Pinkovce-2 (3 m). This unit with the entire depth of the recovered profiles about 9 km is one of the best investigated pre-Neogene units on our territory. Only thank of this such peculiar phenomenon as metamorphism of Eocene formations, Penninic-like lithologies, occurrence of ultrabasic bodies even in overthrusted position on the Eocene metasediments (underplate duplexes), subduction-accretionary style of deformation, Early Miocene FT dating of zircons, occurrences of ultramylonites on exhumation detachements etc. (Soták et al. 1993 - 1997) could be identified here. Assignement of these facts into position of unproved facts and wishes (Gross & Filo l.v.) is not substantiated. They can be seen in this position only in the distorted interpretations like that presented in reviews (see above).

The comments to the work Soták et al. (1996) dealing with the petrofacial composition of flysch arenites forced response to questions, which were open prematurely and uselessly in confrontational way (more works are just in press or under preparation). The problems of the Central Carpathian Paleogene gained on dynamics. However, this requires a patient evaluation of new facts and new solution approaches. At the same time the contradictions in the problems of the Central Carpathian Paleogene are from far not so dramatic as they are rasantly formulated by authors (Gross & Filo l.v.). There are mainly some different opinions but from a great part also misunderstandings, which are normally resolved by clarification and not contradiction. The results presented by works of Soták et al. (1996), Soták & Bebej (1996) and others are not commited to be the only one and the most correct, many of them are formulated as preliminary or alternative solutions which may be later supplemented or changed. But they stem out from real facts and arguments and in absolutely responsive and correct approach. It is

only sorry that the authors of reviews are not more open for accepting results from other research activities in the Central Carpathian Paleogene, which they could by their rich experience to inspire and not a priory reject.

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