SLOVAK GEOLOGICAL MAGAZINE

VOLUME 6 NO 4 ISSN 1335-096X

Geological Survey of Slovak Republic, Bratislava
Dionýz Štúr Publishers

4/2000

ISSN 1335-096X

SLOVAK GEOLOGICAL MAGAZINE

Periodical of Geological Survey of Slovak Republic is a quarterly presenting the results of investigation and researches in a wide range of topics:

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Technical Editor: G. Šipošová

Address of the publishers: Geological Survey of Slovak Republic, Mlynská dolina 1, 817 04 Bratislava, Slovakia

Printed at: Gupress Bratislava

Price of single issue: USD12

Annual subscription rate: USD 48 (4 issues) The price include the postage

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Andrusov D., Bystrický J. & Fusán O., 1973: Outline of the Structure of the West Carpathians. Guide-book for geol. exc. X. Congr. CBGA, Geol. Úst. D. Štúra, Bratislava, 5 - 44.

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SLOVAK GEOLOGICAL MAGAZINE

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60^{th} Anniversary of the Dionýz Štúr State Geological Institute (Geological Survey of Slovak Republic)

PAVOL GRECULA

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The relationship between Earth sciences and the conditions of life on Earth is well known, even if we are not always fully aware of it these days. Our progenitors, significant scientific and political personalities of the past, however, realised the urgent need of scientific institutions for the development of the society and state. Among the first established professional institutions was also the State Geological Institute, founded on 15 May 1940.

First, let us remember now some of the occasions which preceded the establishment of the State Geological Institute. Since the beginning of the organised geological research in Slovakia, our science was always serving society, although its role has changed throughout the time. During the foundation of the Imperial Geological Institute in Vienna in 1849, the main focus was on the compilation of the first geological maps of the whole territory of the Austrian-Hungarian monarchy, including Slovakia. The maps were compiled, for instance, by Zejschner, Hauer, as well as by the first director of the Imperial Institute. In 1869, after the Austrian-Hungarian clearance, the geological exploration of Slovakia was undertaken by the Imperial Hungarian Geological Institute in Budapest (today Magyar Földtani Intézét), with the main task of the investigation of mineral deposits. This investigation is summarised in a very important monograph on the deposits of the whole territory of Hungary, written by K. Papp.

As early as two months after the origin of the Czechoslovakia, the commission of the Czech Academy of Sciences asked in a memorandum for a foundation of the Geological Survey (Geological Institute). This created an impulse for the decision of the Council of Ministers from 7. July 1919 about the foundation of the Geological Survey of Czechoslovakia in Prague, headed by Prof. Cyril Purkyně. Detailed geological mapping and exploration of the republic territory were the main tasks of the institute. Several great Czech geologists worked in Slovakia, and their work resulted in the new synthesis of the Western Carpathians structure, compiled by A. Matějka and D. Andrusov in 1931. Geological backgrounds were created for the localisation of the railway line Červená skala - Margecany, Banská Bystrica - Horná Štubňa, Jelenec dam, Sliač baths, as well as for the Mining factories in the Spišsko Gemerské rudhohorie Mts. and for several other practically oriented geological investigations.

After the third meeting of the Carpatho-Balkan geological association, held in Slovakia, the ambitions for establishing an independent geological organisation increased also in Slovakia. This meeting was an important milestone in the geological research in Slovakia as well as in the whole Carpathians.

60 years ago, new ideas were not being accepted easily. In September 1935, in a memorandum about the necessity of accelerated geological research in Slovakia and Ruthenia, Prof. JUDr. Imrich Karvaš asked the Ministry of Economy of Slovakia to establish a geological institute. The lobbying for the establishment of the Geological Institute went also through the Slovak Mining District, based on the need of mining in Slovakia (exploitation of the new sources of raw materials). Although the leading Czech geologists of that time (J. Šuf, R. Kettner, Zoubek, Matějka, Koutek, Fiala and other) were working in Slovakia at that time, the need to create a large and organised geological institution was felt strongly. After the Mining and Forestry Academy had moved from Banská Štiavnica to Hungary in 1918, neither a university, nor other organisation existed in Slovakia that would educate geologists and concentrate geoscientists for the practical needs of mining. For this reason, a meeting of the Slovak Mining District in 1938 discussed these problems and assigned Prof. Andrusov to prepare a proposal of the guidelines and activity programme of the institute that would be established at the Technical University in Košice as an Institute for the Mineral Research of Slovakia. Prof. Andrusov was appointed as a head of the Institute.

As a consequence of the state - juridical changes, the Technical University was moved to Martin and later to Bratislava, which also meant that the institute could not develop its planned activity. Therefore, the endeavour for the establishment of the independent geological institute strengthened, mainly in the second half of the year 1939. The Commercial and Industrial Chamber, but also the Ministry of Commerce, Industry and Trade of the ČSR (Czechoslovak Republic) as well as Ministry of Transportation got involved. However, numerous organisations had quite different ideas about the institute orientation. Among them, for example, was the idea that this Institute should be focused on engineering geology. The Ministry

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of Education and National Enlightenment, represented by Prof. E. Horniš, proposed an establishment of the institute based on the example of the Geological Institute in Prague (this opinion was also held by Prof. J. Volko-Starohorský). Finally, this proposal was accepted, and the law and government decree about the establishment of the Geological Institute were prepared. Prof. D. Andrusov and Dr. M. Kuthan also joined in the process of preparation. On 15 May 1940, the Slovak Assembly approved the Law No. 119, according to which the State Geological Institute was established by the Ministry of Education and National Enlightenment. The role of the institute was to organise systematic research of geology and mineral riches of the Slovak Republic.

Perhaps somewhat widely, I have described main events leading to the establishment of the Geological Institute. However, sources from the archives point out that many other additional actions lead its birth. I want to stress that need for this Institute was not only expressed by geologists but also by representatives of the economic sphere; thus reflecting a true social necessity for its creation.

For the successful realisation of the task, Prof. D. Andrusov first had to bring up a new generation of Slovak geologists. This generation included M. Mahel', B. Cambel, O. Fusán, J. Kamenický, J. Šalát, V. Zorkovský, B. Leško, J. Bystrický, L. Ivan, J. Jarkovský, J. Ilavský, J. Seneš, J. Kantor, J. Švagrovský, E. Brestenská, V. Kantorová and others. It is admirable what this generation of geologists has done for the evolution of geology of Slovakia. Their names, together with the name of their scientific leader, Prof. D. Andrusov, have been related to all the main geological research and exploration projects in numerous crucial areas of the Slovak economy. Especially after the World War II, when it was necessary to revitalise destroyed Slovak industry. Almost all the geologists took part in the exploration for mineral deposits, in order to provide new resources, mainly of ores and building raw materials. This trend existed about up to the end of the 1950s.

Meanwhile, in 1949, the State Geological Institute was assigned a new controlling government body, the State Planning Office, and it was re-named as the Slovak Central Geological Institution. In 1952, by the government act no. 196, a Geofond (Geological Archive) was also established, and in 1954, a branch of the Geofond was formed at the institute.

The first general Slovak geological conference has represented the first step towards the consolidation of the geology in the new organisation, as well as in its financial and material background. This also included moving of the institute to a new building to Patrónka (9. 11. 1953) and renaming of the institute to the Dionýz Štúr Institute of Geology (GÚDŠ). This was an important occasion, when the great Slovak geologist D. Štúr and his excellent work were memorialised. D. Štúr was a member of the Imperial Geological Institute in Vienna and became its director from 1885 until 1892. On 1 June 1965, the Dionýz Štúr Institute of Geology was detached from the organisation of the Central Geological Institute and it became an independent institute for the basic geological investigation of Slovakia. At the same time the Geofond was detached from the GÚDŠ. It became a branch of the Prague's Geofond until 28 October 1968, when the Slovak Geological Office was established and the Geofond became an independent organisation.

It is also important to remember the origin of the government exploration companies in 1951: East Slovakian and West-Slovakian Ore Exploration, Geological Exploration of Raw Fuels (Coal Exploration) and Civil Engineering Geology. These institutions, together with Geophysics company with its head office in Brno, finally formed a spectrum of independent geological organisation in Slovakia. In the following years, the structure of geological activities changed again due to the establishment of two big exploration institutions, Engineering Geological and Hydrogeological Exploration (IGHP) and Geological Exploration (GP), from which the Slovak Geology was detached (the geological division of the GP) in the process of privatisation. In the next stage of privatisation, also this firm was converted to the share holding company - similarly to IGHP and Geophysics.

The previously mentioned reorganisation of Slovak geology in 1950's started the greatest boom of geological activity in the history of Slovak geological research. GÚDŠ was given a grant for the compilation of geological maps at 1 : 200 000 scale, which were published in a printed edition before 1964. The Central Institute of Geology in Prague and GÚDŠ were awarded the Republic Medal for this map edition. Later, the period of map compilation at 1 : 50 000 scale started, based on detail maps at 1 : 25 000 scale. This period will finish in the near future by the publication of these geological map series.

The geological investigation of the territory of SR was significantly shifted forward by the compilation of the geological maps. The mapping results were successfully presented at the 10th meeting of KBGA and other international events. The development of other geological disciplines was associated with the mapping. Particularly the metallogenetic research connected to the evaluation of mineral resources, development of hydrogeology, engineering geology and related geological research, serving the practical needs of the state economy.

However, after 1989, the role of geology in the society has been gradually, but very significantly changed, along with the changes in the economical system. The position of geology has changed within the framework of economical needs of the state and society. These changes also required changes in the management of the geology in Slovakia.

On 15 January 1996, the Geological Survey of Slovak Republic was established by the decision of the Ministry of the Environment, ceasing the existence of and merging the Dionýz Štúr Institute of Geology, the Geofond and the Slovak Geology. The Geological Survey was established according to the structure of the European geological surveys. The Geological Survey of SR was later again

Fig. 1 Ceremonial speech given by prof. RNDr. László Miklós, DrSc., the Minister of the Ministry of the Environment SR, on the occasion of the 60th Anniversary of the ŠGÚDŠ (GS SR) establishment.

In the front of the photo the directors of geological surveys of neighbouring countries.

Fig. 2 On the ceremonial occasion outstanding workers were appointed the Gold Medal of the Geological Institute. From the left to the right: Academician O. Fusán, Academician B. Cambel, RNDr. O. Samuel, DrSc.

renamed, by the new decision of the Ministry of the Environment of SR from 26 April 2000, becoming valid on 1 May 2000, this time as the State Geological Institute of Dionýz Štúr,

I have touched only some of the significant milestones of the history of the Geological Institute and mentioned only some of the results of the Institute activity. Now a brief vision of the future, in the context of new tasks for the geology, with regard to domestic features and global trends.

The relation of the public to the Earth and to the abiotic nature is remarkably changing. One of the features of this change is the growing awareness that exploitation of natural sources has its limits. The contamination of groundwater, erosion of the Earth surface, global warming, limitation of energy sources and mineral resources are outstanding issues, which the society perceives and pays an increasingly closer attention to.

In the past the significance of geology was seen mainly in its economic contribution. Geologists investigated the Earth so that it could provide raw materials for the industry. Recently, however, the tasks of the geology have diversified and the geology has come under strong economic pressure. The Earth sciences are under budgetary pressure not only in our country, but in all advanced countries. The restriction of the expenses for geology at present is connected with the fact that the traditional function of the Earth sciences - i. e. the localisation of raw

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materials necessary for the industrial growth and national security - disappears. With the end of the cold war, also the call for the discovery of home sources of strategic mineral raw materials weakened.

At the threshold of a new social era, in the period of unavoidable decline of the geological sciences, quite a different approach to their role should be sought, in order to keep its position in the centre of the public life and at the service to the public interest and needs. Geologists not only have to recognise this change, but also have to provide new solutions for the future.

The fundamental task for geoscientists is not only to provide information about the Earth for public discussions, but also to take part in them. In order to have success as public scientists, geoscientists must be able to offer geological knowledge and results of their work to the whole society.

It is obvious that geoscientists alone cannot change the present-day situation in geology. They, however, can stimulate a progress in this direction by the education of students, as well as broad masses, to fulfil the required public role of the Earth scientists.

So, which aspects of geology have to change in order to meet the interests of the society needs ?

- Compilation of the basic geological maps of the territory of the state should remain as the geological priority.
- Water becomes and will be an acute problem. Surface water, the quality of which progressively deteriorates, must be replaced by groundwater from water-bearing rock horizons. This is a key task for geology.
- Securing the food supply for intensely growing human population will require to protect present-day agricultural areas against their degradation as a consequence of salting and soil erosion. The geological institutes have to be in close co-operation with agricultural institutes and/or to adapt to a trend of the connection of both types of institutes and set up independent agro-geological organisations. At the universities, it is essential to devote grater part of the training programs to the properties of the Earth surface (soil, morphology). As the result of present-day practice, students know more about the Earth core than about the contact layer between the Earth and man.
- Air - monitoring of air pollution indicators. Contribution to the development of CO_2 sinks (especially near the industrial agglomerations) by the identification and preparation of suitable geological environment (allumosilicates, for instance, react with CO_2, producing calcite).
- Urban and environmental planning require profound investigation of geological hazards and reduction of their influence on existing urban areas (radon hazard, landslides, soil erosion, geopathogenic zones, seismic zones, etc.).
- The search for the new energy sources of unrecoverable, but mainly recoverable nature, will remain among major requirements also in the 21st century, although an essentially higher proportion of solar energy is expected. Production of solar collectors, consisting of photovoltaic cells, however, requires a great amount of mineral raw materials.
- The problem of the search for new deposits of classical and new types of raw materials is and will remain among the top priorities.
- There will be an increasing interest in mineral raw materials that are suitable not only for economical exploitation and processing, but also that are ecologically friendly. For geology, it will be important not only to find the raw materials, but also to find out if they technologically suit the above mentioned requirements, and how easily they can be exploited and utilised.
- Palaeoclimatology becomes a very important branch of geology (paleontology). Probable changes of climatic conditions on Earth in the future may be predicted on the basis of the climate changes in the past.
- Because of the necessity to dispose of radioactive and other hazardous wastes in deep geological environment, the study of vertical and horizontal movements of crustal blocks becomes an acute task for structural geologists.
- Geomedicine is another new field of geology, which concentrates on the research into the unfavourable influence of geological factors on the human health, and also focuses on the effects of human activity connected to the utilisation of natural resources in the past, which frequently left behind areas acting as ecological bombs.

From this very brief outline of the main issues for the geology in the coming future, it can be summarised that for the permanently sustainable life conditions a balanced utilisation of water, air, soil, energy and mineral raw materials is extremely important. The geological sciences have the key for better understanding of this problem. However, to secure the implementation of the knowledge into the general culture of the nation, it requires better documentation of the geological knowledge, a change of the research orientation, monitoring, modelling of predicted situations and subsequently better understanding of natural phenomena.

I believe that the State Geological Institute is prepared for the challenge of the new geological tasks in the future.

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Origin and exhumation of mylonites in the Lúčanská Malá Fatra Mts., (the Western Carpathians)

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Abstract. The Valča Formation, formerly interpreted as metamorphosed Devonian sediments is herein redefined as mylonite to ultramylonite in the Valča and Trebostovo valleys of the Lúčanská Malá Fatra Mts., (Central Western Carpathians). The mylonites and ultramylonites were derived from the surrounding granitic rocks under pure shear strain ductile deformation. $^{40}Ar/^{39}Ar$ dating of ultramylonite sericite yielded an age of 72 ± 3 Ma and of muscovite from the granitic rocks an age of 345 ± 2 Ma. Two different phases of exhumation and uplift of the ultramylonites are inferred. A Middle Miocene to Pliocene exhumation and a Pliocene to Recent uplift. The calculated rate of exhumation is about 0.5 mm/yr and for the uplift phase is about 1.0 to 1.4 mm/yr.

Key Words: Central Western Carpathians, Lúčanská Malá Fatra, mylonites, $^{40}Ar/^{39}Ar$ dating, exhumation, uplift

Introduction

The Lúčanská Malá Fatra Mts. (sensu Vass et al., 1988) is southwestern part of the Malá Fatra Mts. The basic geological information about the area is summarized in the geological maps by Andrusov and Kuthan (1943, 1946). The modern researches of this area are represented mainly by geological maps and research works by Rakús et al., (1988, 1989) and Gorek (1990).

At the end of the Valčianská dolina valley, at the end of the Trebostovská dolina valley and at its mouth SE ward of the Končiar site (1163 m) in the area of SE part of the crystalline complex of the Lúčanská Malá Fatra Mts. Pulec in Gorek (1990) described group of beds named Valča Formation. We did not succeeded in an identification of the mentioned beds, thus it is not shown in the simplified geological map (Fig. 1). According to the original understanding the beds consist of metasediments, metaclastics with intercalations of black shales that are classified into tufaceous shales metamorphosed in green schist facie. The composition of the rocks of the Valča Formation consists of quartz, orthoclase, feldspar, micas, chlorite, accessory zircon and newly generated rutile. At the end of the Valčianska dolina valley a poor sulfide ore mineralization was fond in black shales (Pulec l.c.). Based on findings of the palynomorphs, the whole formation was classified as Early Paleozoic - Devonian (Planderová et al., 1990). The Valča Formation comes to the surface in the middle of highly metamorphic rocks and granitoid of hybrid character, middle grain-size biotite granodiorite, tonalite with xenoliths of garnet - biotite paragneisses and biotite paragneisses with impurities of graphite. The immediate contact of the Valča Formation with crystalline complex is formed by mylonites (Gorek and Hók, 1992), - metamorphic rocks in which the original rock can by still macroscopically identified. It is difficult to identify the transition from the mylonites to the Valča Formation. The maximal thickness of the Valča Formation including the mylonite zone is about 100 m.

Methods of investigation

The field investigation was focused on study the transition zones among the granodiorite, mylonite and Valča Formation. The sampling was done along profiles in the Valčianská dolina and Trebostovská dolina valleys, as well as in relatively undeformed zones of granodiorite massive of Lúčanská Malá Fatra Mts. The methodology itself included field geological-structural research, documentation of macrostructure elements and textural changes of the rocks. Within the frame of the structural research, we have focused on collection of basic structural elements in the crystalline complex and Mesozoic formations. An interpretation of thin sections followed. The chemical composition and changes of grain size of individual minerals as reaction to deformation processes were verified by use of macroscopic and EDS-microanalyses of electron microanalyzer JEOL JXA 733 SUPERPROBE a KEVEX Delta, with the following parameters: acceleration voltage 15 kV, measuring current 1.2 nA, Taylor synthetic and natural standards were used.

Geological Survey of Slovak Republic, Dionýz Štúr Publishers, Bratislava 2000 ISSN 1335-96X

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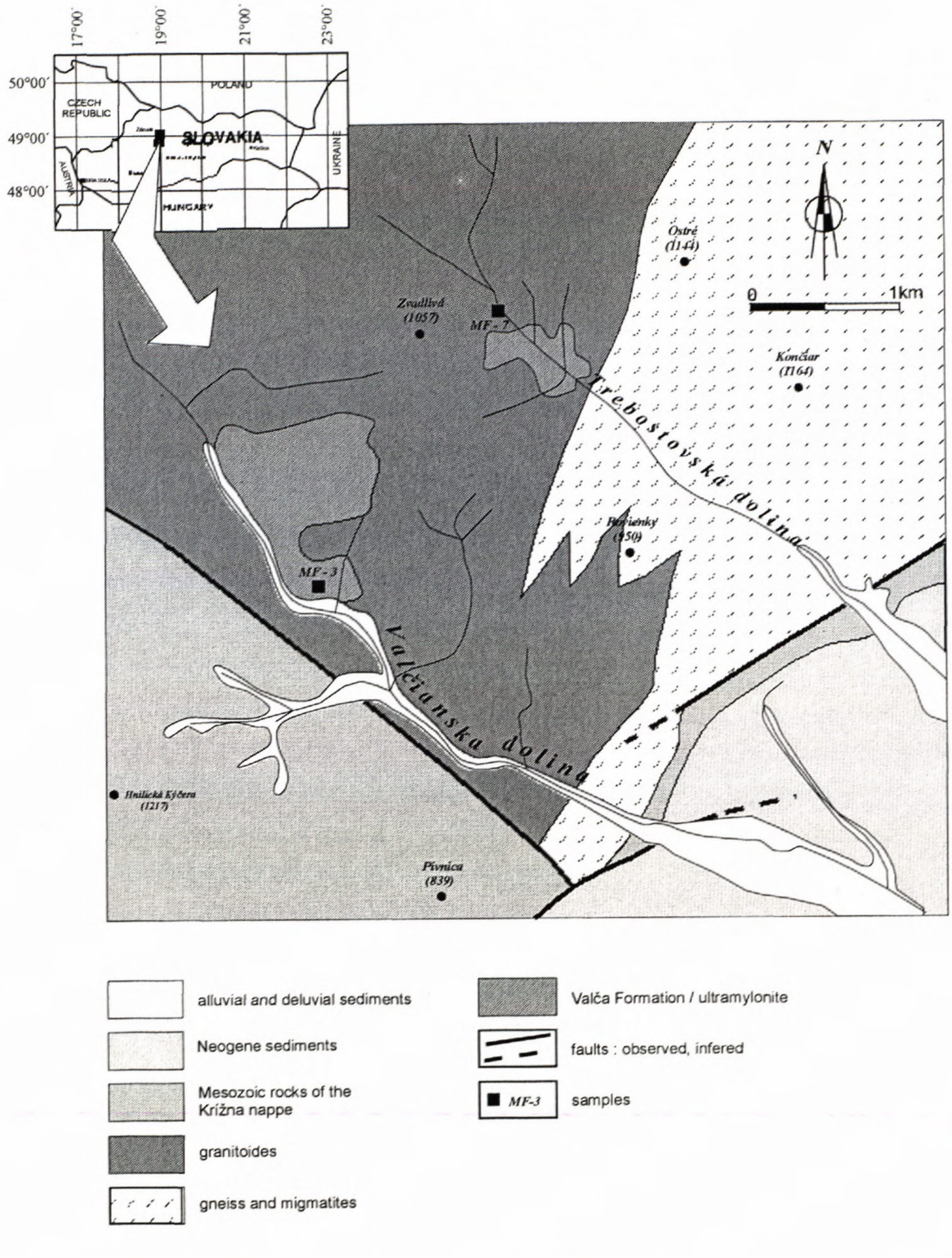

Fig.1 Schematic geological map of the investigated area (modified after Rakús et al.,1989).

After separation of muscovite and sericite from the collected rocks by standard methods (final manual purification under binocular, methanol and bi-distillated water in ultrasonic cleaner), the samples were analyzed in the laboratory GEOZENTRUM, Vienna. Charges with weight 10.4 mg were sealed into quartz tubes and irradiated in ASTRA reactor with dose of fast neutrons (about 10^{17} neutrons.cm^{-2}) together with internal standard WAP that was used to determine J parameter for six samples in the closest position. A degassing of the sample started at temperature 615 - 620 °C, the analysis was over at temperatures 1220 - 1350°C (total fussion), what enabled, with respect to amount of obtained ^{40}Ar and ^{39}Ar, making 6 - 10 analyses of Ar isotope composition. The Ar isotope composition (after its two steps of purification) was measured by mass spectrometer VG 5400 in six cycles. A standard program by fy VG was used for a statistical treatment of the measured isotope composition; the calculations of apparent ages were made with use of interpolated values of relevant isotope ratios, together with accepted age constants (Steiger and Jäger, 1977). More details about this ^{40}Ar/^{39}Ar dating technique can be found, for example, in McDougal and Harrison (1988).

The polished sections were made of the collected rock samples that contained graphite impurities, the light reflectivity of organic matter was determined under standard conditions: microscope Leitz Orthoplan with microphotometer MPV-compact, oil immersion, standard glass prism (R = 1.24 %), photometric field 2.5 x 2.5 microns, lenses 50x, light wave length 545 nm. The maximal (Rmax) and minimal (Rmin) reflectivity of organic matter particles were measured. The obtained values were basis for determination of degree of incarbonization and for calculation of the temperature of the origin.

RESULTS

Macroscopic and Macrostructural Description of the Rocks of the Valča Formation

Clearly developed sub-zonal foliation is typical sign of the Valča Formation. The least effected rocks represent undeformed or slightly deformed granodiorites to tonalites with hybrid character (Fig. 2). The mylonite granodiorite are middle to coarse-grained rocks in which development of metamorphic foliation is possible to observe (Fig. 3). Their shade is grayish, gray-black; the alignment is enhanced by biotite, chlorite, muscovite and deformed eyes of plagioclases, less orthoclase. With increasing deformation (Fig. 4) the rocks get greenish - light green shade caused by dispersed sericite and chlorite. Ultramylonite rocks are very fine-grained, massive, silicified, sometimes with till now preserved plagioclase eyes, else they are detailly folded fine-grained variety with altering 0.5 - 1 mm thick white and greenish strips (Fig. 5). Occasionally we can find layers of dark black fine-grained shales strongly limonitized with content of sulfide minerals: pyrite, chalcopyrite and pyrrhotine.

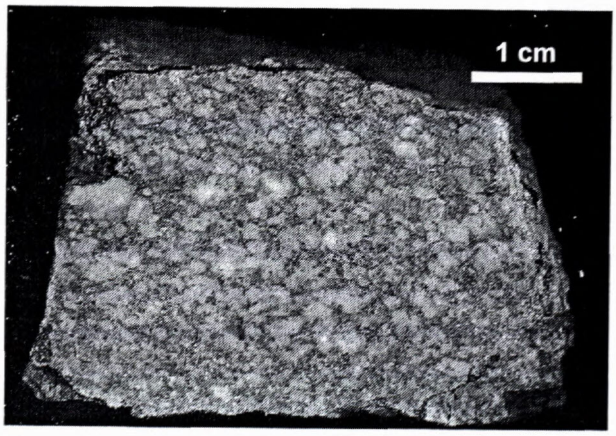

Fig. 2 Photograph of an undeformed granitoid rock. The Valčianska dolina locality.

The Microscopic Research of the Rocks of the Valča Formation

The study of deformation structures in cross sections XZ and YZ planes, development and chemical composition of the minerals confirm macroscopic parameters. Gradual rock mylonitization can be observed - plastic deformation or brittle cataclase. We have divided the samples on the basis of degree of mylonitization into several groups.

The first group includes granodiorite mylonites with preserved mineral composition. They are composed of quartz, plagioclase, phyllosilicates - chlorite, muscovite (phengite, sericite), rarely chloritizated biotite is preserved, and tiny grains of desintegrated accessory minerals - apatite, titanite, rutile, Fe and Ti oxides, zircon, monazite, epidote - allanite. The basic mass is composed of cataclased strongly undulose quartz frequently stripped. The plastic deformation is related to reduction of grain size, local generation of sub-grains and displacement on grain margins. We characterize the microstructure by development of two foliation plains into fine S - C structure (Lister and Snoke, 1984) with inclination about 30°, in zones with higher intensity of deformation stress the degree of surface opening decreases to 15° and less. The detail of the microstructure is shown on the Fig. 6. The foliation planes are defined mainly by chlorite, rarely by chloritized biotite. The dependence of degree of chloritization (decreasing of content of K in the structure) upon degree of deformation is clearly observable. Further they are defined by muscovite of phengite composition, tiny newly developed grains of albite composition from disintegrated plagioclase, rutile, titanite, cataclased apatite and zircon. Some muscovites by their increased content of Ti, Fe and Mg indicate also possible origin by bauertization of biotite.

The second rock group includes mylonites with very fine-grained texture and tiny eyes of deformed minerals (Fig. 7). They have relatively similar composition as the first group, i.e. quartz, plagioclase - albite, phyllosilicates - muscovite prevails (phengite), chlorite, and grains of disintegrated accessory minerals - apatite, epidote, (including content of REE), titanite, rutile, Fe and Ti oxides, zircon. The microstructure is formed by undulosed strongly crushed and partly recrystallized quartz

Fg. 3 Photograph of the ductile deformed granitoid with mylonitic structure. Note the development of mylonitic foliation and stretching lineation. The Valčianska dolina locality.

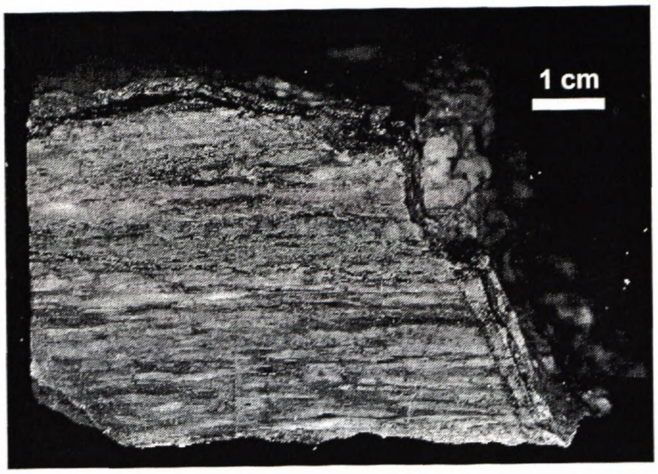

Fig. 4 Photograph of mylonitic structure with light ribbons of quartz and feldspar and dark ribbons of sericite - chlorite mixture. The Valčianska dolina locality.

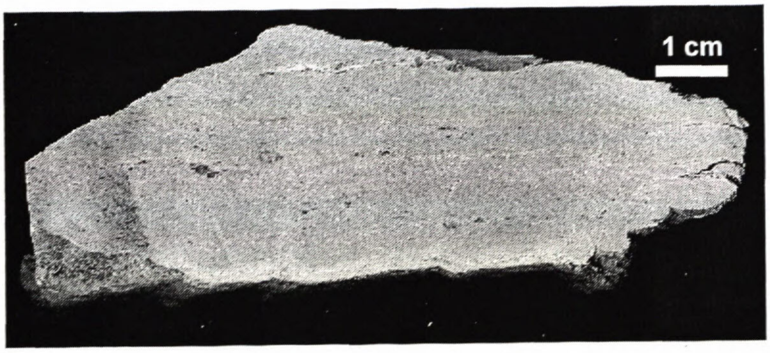

Fig.5 Photograph of a fine - grained ultramylonitic rock with a well developed foliation and kink folds. The Valčianska dolina locality.

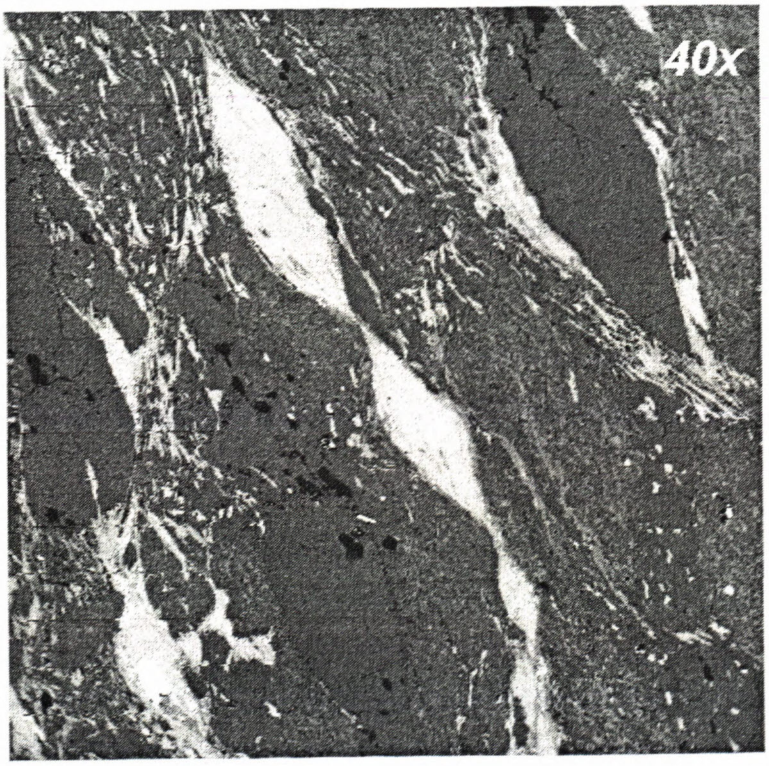

Fig. 6 Photomicrograph of a mylonitic texture. Mylonite, the Valčianska dolina locality. M = 40x

and plagioclase grains up to 1 mm, in fine-grained matrix composed of quartz and phyllosilicates. The foliation plane is formed by altering strips of very fine-grained plastically recrystallized quartz and phyllosilicates, mainly muscovite with phengitic composition. S-C composition is suppressed, S and C foliation plates with growing deformation blend. Similar phenomenon is described by Vernon et al. (1983). In the given deformation field the quartz behaves plastically and plagioclase more brittle, what results in creation of tiny eyes in the quartz matrix. We assume that this phenomenon is caused by a relatively high speed of the deformation. In pressure shadows of the eyes of some of the samples, we observe signs of creation of probably newly developed biotite. Chlorite together with epidote form as if phantom crystals after garnets.

The third investigated rock group is composed of ultra fine-grained, gray to gray-green fine-stripped rocks, ultramylonites (Figs. 5 and 8). The stripping is caused by very strong deformational elongation of the quartz and plagioclase that is changed to epidote, albite and phyllosilicate, although we can observe more ductile behavior of plagioclase than quartz in this rock. Fine-grained titanite and fine-flaked chlorite are abundant, they cause light greenish shade of the rocks. The muscovite is only accessory, however, in fine-grained matrix we can find also disintegrated epidote with a content of REE. The reduction of grain size of the rock basic mass - matrix is extreme and their size reaches range 0.005 - 0.030 mm (Fig. 8). The so-called equilibrium state is reached at this values in the quartz and the grains are not destroyed any more, although the degree of the deformation can be increased yet (c.f. Vernon et al., 1983).

Several conclusions can be draw from our results. In rock microstructures of the so-called Valča Formation we can find clearly provable phenomena of mylonitization: cataclastic structures of tectonic breccia when relatively angular porphyroclasts of plastically deformed quartz or feldspars in fine-grained matrix are preserved; asymmetric structures as S - C structure, fine intraformational folding; rotation of plastically deformed porphyroclasts and metacrystals; plastic deformation of phyllosilicates (originally biotite) along basal planes into shape of so-called mica "fish"; and finally the dynamic recrystallization of quartzite (crystal - plasticity) connected with generation, rotation of sub-grains and border migration in flatten the quartzite strips.

We tried to evaluated thermal - pressure conditions of the deformation with help of structural - petrography criteria. The existence of chlorite as alteration product of biotite

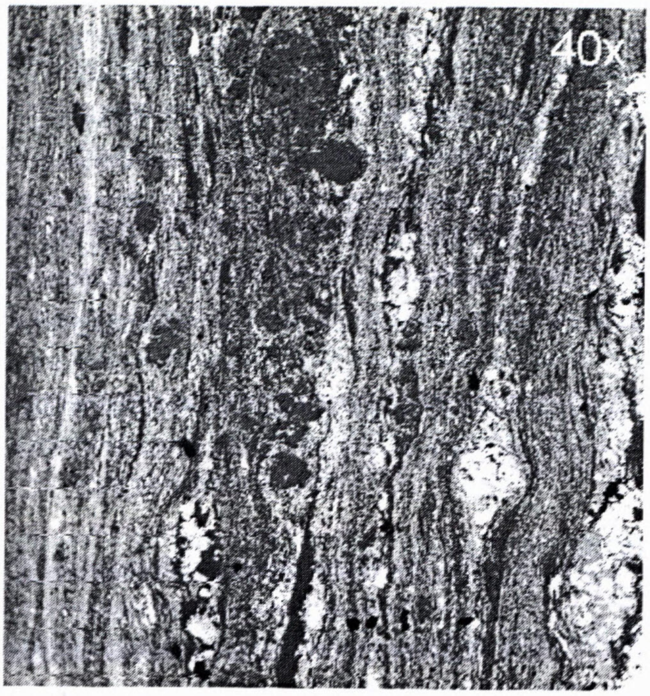

Fig. 7 Photomicrograph of elongate quartz grains and feldspar aggregates. Mylonite, the Valčianska dolina locality. M = 40x.

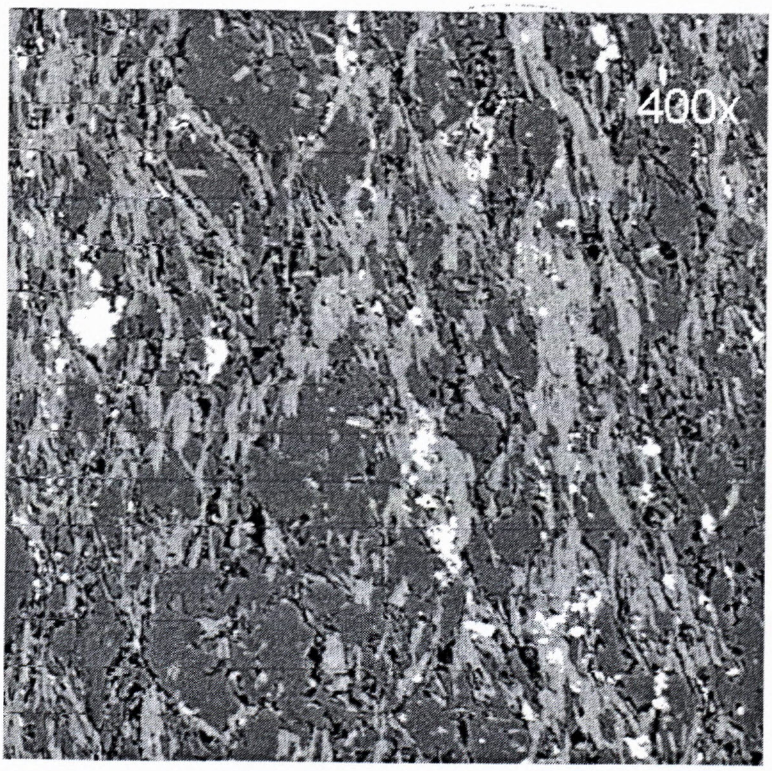

Fig. 8 Microphotograph fine-grained texture. Ultramylonite, the Valčianska dolina locality. M = 400x

biotite and plagioclase and signs of generation of new biotite in pressure shadows indicates reactions under conditions of middle and higher range of green shale facies to lower range of amphibolite facies. Eggleton and Banfield (1985 in Shelley, 1993) suppose the temperature for chloritization reaction 340°C. On the other hand, the crystal - plastic deformation of quartz in some of the samples indicates temperatures with maximum 450°C (Schulmann, oral consultation 1998). This is apparent mainly in generation of long flattened quartz strips with unstable moving margins, what indicates hither temperature solidus deformation, similar as in mylonited ortoschists of the crystalline complex of the Nízke Tatry Mts. (Madarás et al., 1999). According to changes of the rocks, we assume that the first deformation phase was cataclase under brittle conditions, followed by plastic deformation lasting until stage of the ultramylonitization. A release and intrusion of sufficient amount of reaction fluids is needed for such process. These can be obtained under sufficient temperature - pressure conditions by disintegration and transition of minerals containing water in their crystal lattices. The intruding fluids attack surrounding waterless minerals, mainly plagioclases, which change to albite, phyllosilicate and minerals of epidote - zionist group, reactionaly soften, and deform in the oriented pressure. Ultrafine-grained fine-stripped rocks without any macroscopic signs of eye structure can be created by described mechanism from coarse-grained and ocellar structure of granitoid rocks. Only during detail microscopic investigation, we can still observe as rounded relicts of original minerals, quartz and feldspars.

The Results of the Structural Research

The structural research was focused on collection and evaluation of basic structural elements. Foliation, mineral lineation and B - axis of folds in the case of the crystalline complex rocks and so-called Valča Formation. The directions of dips of the bedding and fold axis were measured on the Mesozoic rock successions. The results of the structural research are summarized in diagrams (Fig. 9).

The direction of bedding in the Mesozoic rocks is relatively uniform and it has NE - SW orientation (Fig. 9a). Fold axes with NE - SW direction have similar directional features (Fig. 9d). Rocks of the crystalline complex show considerable dispersion of foliation and fold axes directions, from N - S direction to NE - SW direction (Fig. 9b). This dispersion is caused by Alpine type directional reorientation of originally N - S to NNE - SSW oriented Hercynian structural elements to new NE - SW direction. The reorientation of the structures of crystalline complex is well documented also by lower degree of metamorphic changes of structures oriented in NE - SW direction. The signs of the Hercynian tectonics are most significant on the eastern margin of the area in crystalline schist of the Lúčanský complex (Gorek, 1990). The planes of the original foliation of the gneisses are fixed by granitization, what we considere as result of metamorphism of the

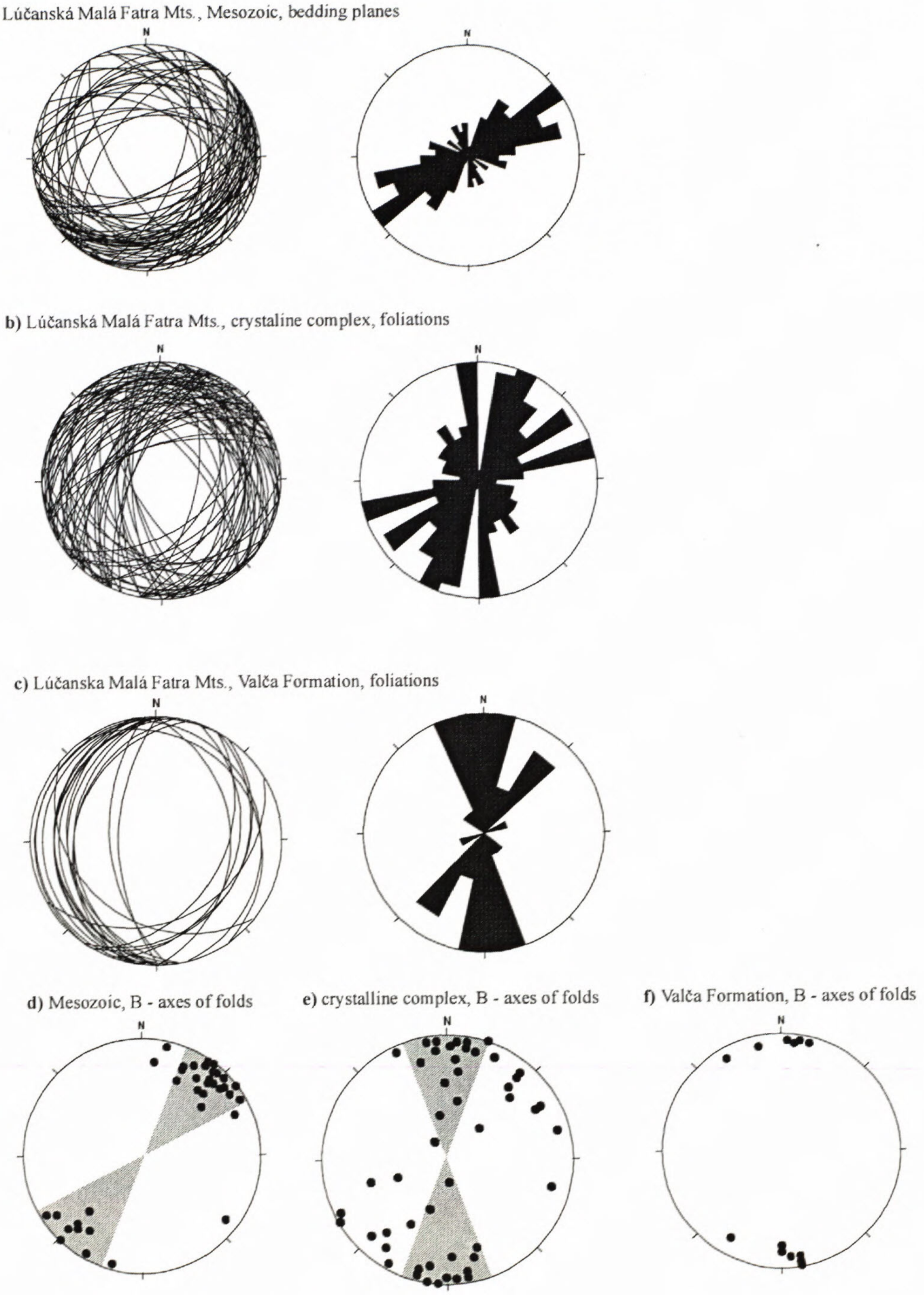

Fig. 9 Tectonic diagrams of the basic structural elements from the investigated area. (Lambert projection, lower hemisphere).

whole complex under conditions of the amphibole facies. The temperature - pressure conditions of metamorphism of the Lúčanský complex reached 700 - 750°C and pressure 8 - 10 Kbar (Janák and Lupták, 1997). The Lúčanský complex was folded to isoclinal and open folds with N - S to NNE - SSW direction. The generation of the folds was joined with compression that was generally oriented in E - W direction, and it could pass to transpresion (Gorek, 1990), what could be indicated also by part of the folds with steeply dipping fold axes (Fig. 9e). In the Alpine orogenetic cycle the planes of the original foliation that represent primary inhomogeneity were again tectonically used, however in lower metamorphic conditions of green schists facies, what resulted in recrystallization of originally higher temperature mineral associations.

Directional characteristics of rocks of the so-called Valča Formation are similar to rocks of the Lúčanský complex, although it is necessary to mention that the statistical data set is not sufficient enough (Fig. 9c, f). Basing on this, we assume that they were created by deformation of the crystalline complex rocks during the Alpine orogenetic cycle. Despite the fact that we have identified numerous asymmetric structures, we have not succeeded in clear determination of the direction of the tectonic transportation. We assume that pure shear without more significant directional component of displacement (simple shear) was significant, mainly during the final stages of deformation of rocks from which the Valča Formation was formed.

$^{40}Ar/^{39}Ar$ Data

The basic analytical data from the measured samples are shown in tables 1 and 2. The apparent age spectra of both samples are significantly different. The muscovite sample MF-7 from a granodiorite (Fig. 10) yielded smooth spectrum of apparent ages in volume of 84% of the total amount of ^{39}Ar released, from 337 Ma to 353 Ma. The plateau age from the four last steps is 345 ± 3 Ma. This age corresponds to rock cooling at temperature about 350°C (Purdy and Jäger, 1976). The characteristic shape of the spectrum in low and middle temperature part can be considered as result of Early Jurassic temperature event that caused partial release of ^{40}Ar of the sample. Although the Early Jurassic age is indicated only by a very low volume of released ^{39}Ar, it can have its real geological importance because spectrum of apparent ages of various minerals starts with these values also in other regions of the crystalline complexes of the Western Carpathians Mts. (Maluski et al., 1993, Kráľ et al., 1997).

Tab. 1. $^{40}Ar/^{39}Ar$ analytical data, sample MF - 7, muscovite, the Trebostovská dolina valley.

Step	T (°C)	% ^{39}Ar	% $^{40}Ar^*$	$^{40}Ar^*/^{39}Ar$	Age (Ma) ± 2 SD
1	615	0.9	91.3	21.25 ± 4.0	173.2 ± 6.7
2	645	1.6	94.4	25.19 ± 2.2	203.7 ± 4.3
3	690	1.2	95.4	29.29 ± 2.5	234.9 ± 5.6
4	735	2.2	95.0	32.09 ± 1.5	255.8 ± 3.6
5	770	3.3	95.6	36.97 ± 0.9	291.9 ± 2.4
6	830	7.1	97.5	40.70 ± 0.7	319.0 ± 1.9
7	865	12.7	98.8	43.21 ± 0.3	336.9 ± 1.0
8	925	21.2	99.4	43.52 ± 0.4	339.2 ± 1.4
9	1065	22.0	99.3	44.28 ± 0.2	344.6 ± 0.7
10	1220	27.7	99.5	45.43 ± 0.2	352.8 ± 0.6

J = 0.004530 ± 0.4 % total gas age: 334.0 ± 2.6 Ma
84 % plateau age: 344.8 ± 2.2 Ma

Tab. 2. $^{40}Ar/^{39}Ar$ analytical data, sample MF - 3, sericite, the Valčianska dolina valley.

Step	T (°C)	% ^{39}Ar	% $^{40}Ar^*$	$^{40}Ar^*/^{39}Ar$	Age (Ma) ± 2 SD
1	620	4.6	60.0	8.48 ± 13.1	71.1 ± 9.1
2	690	6.6	75.1	8.89 ± 10.5	74.5 ± 7.7
3	790	11.8	89.9	8.85 ± 4.2	74.1 ± 3.0
4	920	50.2	96.9	8.44 ± 0.8	70.8 ± 0.5
5	1050	18.7	93.5	9.03 ± 2.2	75.6 ± 1.6
6	1350	8.1	87.3	34.10 ± 4.1	270.8 ± 10.5

J = 0.004530 0.4% total gas age: 88.5 ± 3.5 Ma
92 % plateau age: 72.4 ± 2.7 Ma

The sample of sericite separated from the mylonite (MF-3, the Valčianska dolina valley) has totally different spectrum. The dispersion of the apparent ages is minimal in the first five temperature steps (71 - 76 Ma). The plateau age for volume 92 % of released ^{39}Ar is 72 ± 3 Ma (Fig. 11). We interpret this age as age of generation of the sericite. We reckon the apparent age 271 Ma in the last step (1350 °C) as relative age that indicates either the old cores of the separated sericite or the presence of individual tiny chips of muscovite grains (non-removed by separation) originating from the original granodiorite.

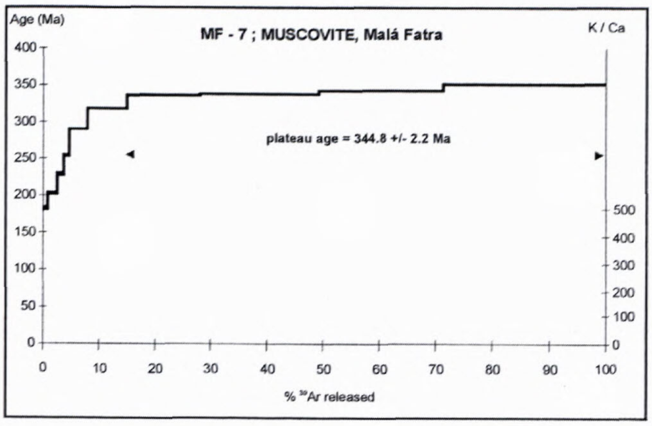

Fig. 10 $^{40}Ar/^{39}Ar$ apparent age spectrum (sample MF - 3; ultramylonite, the Valčianska dolina locality)

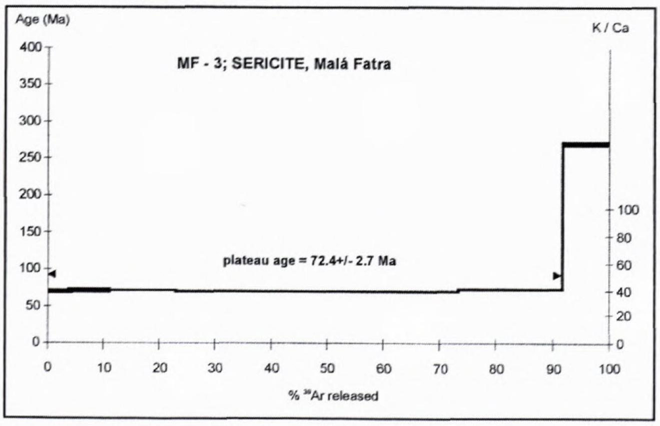

Fig. 11 $^{40}Ar/^{39}Ar$ apparent age spectrum (sample MF - 7, granitic rock, the Trebostovská dolina locality)

Results of Reflectance of Graphite Matter

In the studied samples there is graphite matter consisting mostly of 10 to 20 microns particles ordered in to the direction of rock foliation. The graphite matter has flake like form, it does not have preserved relict structures of the plant origin. The maximal values of reflectance are in range from Rmax = 3.6 - 4.6 % to Rmin = 2.0 - 3.0 %. It is apparent from the results that it is meta-anthracite graphite matter. We attempted with help of software Easy%Ro (Sweeney and Burnham, 1990) to calculate the temperature under which the meta-anthracite was generated. The minimal temperatures were determined to 260°C and the maximal to 360°C. Similar conditions (260 - 290°C) are reported also by Diessel et al. (1978) obtained from Tertiary shales in New Caledonia.

Discussion

The age of the granodiorite intrusion from the Malá Fatra Mts. was determined on the basis of U/Pb dating of zircon to 353 + 11/-5 Ma (quarry Dubná skala, Shcherbak et al., 1990). The attempt for Rb/Sr dating of granodiorite rocks from this area demonstrated very small Rb/Sr dispersion of the analyzed samples, however which can by statistically localized on reference isochrone from granodiorite rocks of the Veľká Fatra Mts. (361 ± 10 Ma, Bagdasaryan et al., 1992). Similar as U/Pb dating of the Malá Fatra granodiorite, this age is within of the analytical errors identical with U/Pb ages of a zircon from the granodiorite rocks of geographically closest crystalline cores (the Veľká Fatra Mts. 359 Ma; Kohút et al., 1999, the Strážovské vrchy Mts. 356 ± 9 Ma; Kráľ et al., 1997). Fission - track dating of apatite from granodiorite rocks of the Malá and Veľká Fatra Mts. are identical within of analytical errors (23 - 25 Ma), what was interpreted as age of Neogene uplifts of the core mountains (Kráľ, 1977).

Based on newly obtained experimental experiences, we assume that the so-called Valča Formation is product of the Alpine mylonitization of granitoid rocks. The mylonitization took place on boundary of Cretaceous and Paleogene (72 ± 3 Ma). At this time the granitoids were cooled down to temperatures lower than 300°C, as we have proved by dating of the sample MF - 7. We assume that the temperature conditions of the mylonitization were in the range 300°C - 350°C (Fig. 12).

The problematic is presence of the Devonian palynomorphs that were described in the rocks of the Valča Formation (Planderová et al., 1990). In the previous studies (Gorek, 1990; Gorek and Hók, 1992) a hydrothermal - metasomatic rock alteration related to sulphidization and probable income of organic, carbonic substance is assumed. The presence of water in the given tectonic environment is common and furthermore it can considerably modify the way of physical and metamorphic reactions (for example Sibson, 1977). The mentioned assumption can explain the presence of black shales with sulphidic mineralization within the Valča Formation. If we assume similar way also the contribution of the palynomorphs than it is questionable weather we can derive them only from surrounding rocks that, however, represent high metamorphic complex with temperatures of Hercynian metamorphism acceding 700°C (Janák and Lupták, 1997). The temperature conditions of the Hercynian metamorphism exclude preservation of the palynomorphs. However, the determined thermal transition of the graphite organic matter would not exclude preservation of the palynomorphs during their transport by hydrotherms (c.f. Planderová, 1991). It is apparent that the palynomorphs can be hydrodynamicaly carried to greater distances. In groundwaters in the environment of crystalline complexes of the Western Carpathians there were found beside Devonian also Early and Middle Miocene association of sporomorphs, which presence is hard to explain from stand point of known geology (Rapant et al., 1986). The palynomorphs could be redeponed into mylonites, however, their source has remained unknown.

On the basis of assumed temperature of the origin of the mylonites, their $^{40}Ar/^{39}Ar$ dating, assumed thickness of sedimentary formations and the average geothermal gradient we have made an attempt to reconstruct exhumation (uplift) of the ultramylonites. The exhumation according to England and Molnar (1990) is defined as transportation of rocks with respect to the Earth's surface (contact of rocks with air or water). The rate of the exhumation is identical with the rate of the erosion. The uplift of rocks is movement of rocks or a part of the Earth's surface (uplift of the Earth surface) with respect to reference level (geoid, sea level). The mean of the rock transportation acts against the Earth's gravitation during the uplift (England and Molnar, I.c.). At given definitions there is a relationship:

Uplift of the Earth surface = rock uplift - exhumation

If we consider geothermal gradient accepted for the Western Carpathians area about 35°C/km (30°C/km - 40°C/km e.g. Kováč et al., 1994. Hurai et al., 1991; Franko et al., 1995) and the temperature of mylonite generation 350°C than the rock transition took place at depth about 10 000 m. On the basis of sedimentation record of rocks from the Rajecká kotlina basin, however mainly of the Turčianska kotlina basin, we have tried to make reconstruction of the rate of exhumation / uplift of rocks of the Valča Formation. We assume that the mylonitization of crystalline complex rocks is a product of the tectonic extension between Cretaceous and Paleogene, which preceded development of the deposition basins and deposition of the Paleogene clastics. In the area of the Rajecká and Turčianska kotlina basins the sediments of the Paleogene were folded in the Early Miocene and the whole area of the Lúčanská Malá Fatra Mts. was in compressional tectonic regime (Hók et al., 1998). The first probative Neogene sediments in the Turčianska kotlina basin derived from Lúčanská part of the Malá Fatra Mts. deposited in the Middle Miocene, their material was derived predominantly from rocks of the Choč nappe and secondary from the Paleogene sediments (Abramová and Slovany Formations, sensu Hók et al., 1998). The thickness of the Paleogene sediments of the mentioned area does not exceed 1 500 m and the thickness of Choč nappe does not exceed 1 000m. The pebbles matter derived from the Krížna nappe and crystalline complex was uncovered in the sediments of the upper most Miocene to Pliocene (Bystrička Formation I.c.). The samples used for FT dating of apatite determined the age when the crystalline complex was cooled down to temperature 120 - 100°C, what with respect to accepted geothermal gradient indicates depth no more than 4 000 m and stratigraphic age Early Miocene (Eger). However, the samples for FT dating were taken from crystalline complexes at places where there is no verification of altitude of the erosive cut off of the crystalline complexes with respect to the Mesozoic cover. In sediments of Pliocene there are material derived from crystalline complexes, which totally domi-

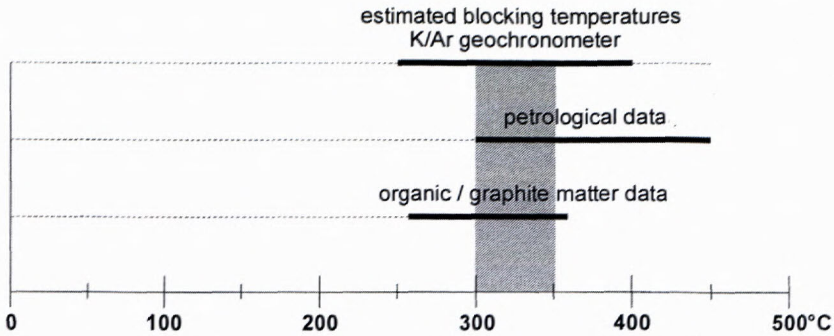

Fig. 12 Temperature of formation of the ultramylonite (shaded area) derived according to different methods (see text for details).

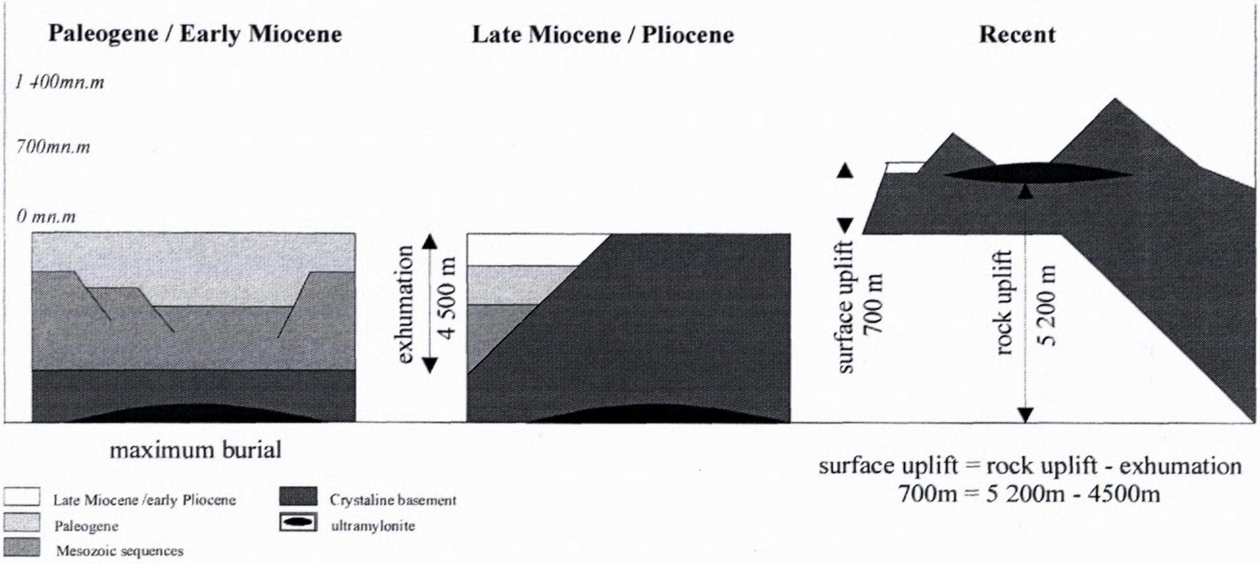

Fig. 13 Model of the exhumation and uplift rates of the Valča Formation (not to scale).

nates in the Quaternary sediments. We assume that crystalline complex of the Lúčanská Malá Fatra Mts. was uncovered in the Upper Miocene - Pliocene time. On the basis of duplex structure of Krížna nappe in the area of the Lúčanská Malá Fatra Mts. we assume its thickness to about 1 500 m. The thickness of the cover sequence does not exceeds about 500 m. However, it is necessary to mention that the cover sequence is not developed in the majority of the area of the Lúčanská Malá Fatra Mts. According to our assumption, the thickness of the sedimentation sequence of the crystalline complexes overburden did not reach 4 500 m. The estimated thickness of the Mesozoic sequence is about 3 000 m. We assume that the maximal burial depth of the mylonites (about 11 500 m) is on the border of Paleogene and Neogene, when in overburden of the Mesozoic sequences there were Paleogene rocks (Fig. 13). The crystalline complex was exhumated at the end of Miocene, what means that since Middle Miocene (The Latest Badenian, about 14 Ma), when the first clastics deposited in the Turčianska kotlina basin, to Pliocene (about 5 Ma) about 4 500 m of deposited overburden was eroded. The average rate of the exhumation is about 0.5 mm/year, which makes 4 500 m per 9 Ma.

The denudation relicts of Pliocene sediments remained in the northeastern part of the Lúčanská Malá Fatra Mts. at altitude 700 - 750 m a.s.l. At the same altitude, there are ultramylonite at the Valčianská dolina and Trebostovská dolina valleys. After the exhumation of the sedimentary sequences the uplift of the rocks of the Valča Formation represented the value about 7 000 m per 5 millions years i.e. about 1.4 mm/year.

If we consider the relationship that was defined by England and Molnar (1990) the value of the exhumation is 4 500 m, the uplift of he Earth's surface is 700 m and the rock uplift is 5 200 m. However, in the given case we do not consider any rate of erosion. From the point of view of above defined relationship the burial depth of the rocks of the Valča Formation would reach 10 000 m. The value of the uplift would reach in the given case about 1.00 mm/year. In both cases, in the calculations there is uncertainty of the main variables. In the first case, it is a value of the geothermal gradient. In the second case, it is the rate of the erosion since Pliocene to Recent. For example, we would consider the value of geothermal gradient 40°C/km, the difference of the uplift intensity estimated by various ways would be minimal. However, in any of the cases of the estimations we can conclude that since the end of the Miocene the rate of the uplift is greater than the rate of exhumation. This assumption could be indirectly confirmed also by positive values of the recent vertical movement tendencies of the Earth's crust in the area (c.f. for example Vanko, 1988; Joó et al., 1992). At the same time, we can accept assumption that the transition of the granitoid rocks to ultramylonites took place approximately at depth 10 000 m.

Conclusions

The rocks at the Valčianska dolina valley and Trebostovska dolina valley, which were defined as the metasediments of the Valča Formation of Devonian age (Pulec in Gorek at al., 1988) are mylonites to ultramylonites derived from crystalline rocks of the Lúčanská Malá Fatra Mts. For this reason, we do not consider as appropriate to use the name Valča Formation. The age of their origin was determined to 72 ± 3 Ma by $^{40}Ar/^{39}Ar$ dating of the ultramylonites. The mylonites and ultramylonites were formed at temperature 300° - 350°C in estimated depth 10 000 m. We assume that they were re-exposed and uplifted in two phases. The exhumation took place since Middle Miocene to Pliocene and rate of exhuation is about 0.5 mm/year. Uplift took place since Pliocene to Recent and its rate is about 1.0 to 1.4 mm/year.

Translated: Dr. Juraj Janočko

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Slovak Geol. Mag., 6, 4(2000), 335–346

Are there tectonic units derived from the Meliata-Hallstatt trough incorporated into the tectonic structure of the Tisovec Karst ? (Muráň karstic plateau, Slovakia)

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Abstract: The Tisovec karst and Kučelach massif area are composed of several tectonic units displaying different degree of metamorphism, character of deformation, age and lithological composition. By means of geological mapping, structural and stratigraphical research, several tectonic units were distinguished. The lowermost one is the crystalline basement and the Permo-Triassic Federata Cover Unit (metasediments) of the southern Vepor Unit. The Vepor Unit is overthrust by (from the bottom to the top) epimetamorphosed Carboniferous sediments of the Gemer Unit and slices of the Meliata Unit(?) and Torna Unit(?) composed of anchimetamorphosed sediments. These include newly described nappe units within this area. The Muráň nappe of the Silica Unit forms the Tisovec karst, which is the uppermost nappe unit. Undeformed postnappe the Paleogene sediments and remnants of the originally volumnous the Neogene volcano-plutonic complex, which cover older structures at the mapped area and are the youngest tectonic units.

Key words: lithology, stratigraphy, nappe tectonics, Silica, Torna and Meliata Units, Western Carpathians, Muráň karstic plateau

Introduction

The Tisovec karst is situated in the central part of the Veporské rudohorie Mts. The Tisovec karst belongs to the subunit of the Muránska planina plateau (Mazúr & Lukniš, 1986). The studied area is cca 70 km^2 large.

Main goal of this paper is to present a study that contributes to the knowledge concerning geology, stratigraphy and tectonics of the Tisovec karst and the Kučelach tectonic outlier northwest of Tisovec town.

This study was focussed on a revision of the geological map of the Tisovec karst area at scale of 1:10,000, with an emphasis on stratigraphy and structural geology. The geological map (Fig. 1) was drawn on the basis of new field work and a review and reinterpretation of archived and published materials (Bystrický, 1959; Bacsó & Valko, 1969; Klinec, 1976; Bezák et al., 1996). This work is an attempt to assemble a comprehesive geological summary of the area, because no throughout study territory has yet been published about this area.

Standard field research methods were used with the geological mapping. These methods include geological observations, structural measurements, documentation of outcrops and sample collecting. Field work was concluded by lithological and stratigraphical analyses.

Regional tectonic structural features were interpreted in this synthesis results gained by geological mapping and laboratory analysis, and applied to regional tectonic situations.

Tectonic unit

Geological structures in the area of Tisovec karst have a sandwich-like character, composed of several superposed tectonic units (Fig. 1, 4), formed compressively during Variscan and Alpine orogenesis. These tectonic units are summarized from the lowest to the highest units.

The southern Vepor Unit

This unit was formed by the Variscan and the Alpine orogene. The Alpine orogeny overprinted on the Variscan Vepor Unit the Alpine orogenic character.

The Vepor Unit is one of the main tectonic units of the Western Carpathians. It comprises a Paleozoic crystalline basement and a Late Paleozoic-Mesozoic metasedimentary cover sequence. The structures although elements of the older Variscan structure were preserved, too.

The Permian-Triassic formations of the Federata Sequence were widely deposited on the Prealpine crystalline basement (Rozlozsnik, 1935; Schönnenberg, 1946; Vozár in Bajaník et al., 1983).

The crystalline basement is represented by granitoids of the Vepor pluton (the Kráľová hoľa complex-sensu Klinec; 1966,1976). The porphyric varieties of the granitoid rocks occur in the northern and northwestern part of this area. The paragneisses, migmatites and granites of the Hybrid complex are located mainly southeast of the Muráň fault zone (Bezák, 1988; Bezák & Hraško, 1992; Lexa & Bezák, 1996). Porphyric granitoids of the Vepor type built mainly higher structural levels of crystalline basement, and they form smaller dikes in the Hybrid complex (Bezák et al., 1999).

The Vepor Unit and the Federata Sequence are epithermally metamorphosed and deformed in a ductile regime.

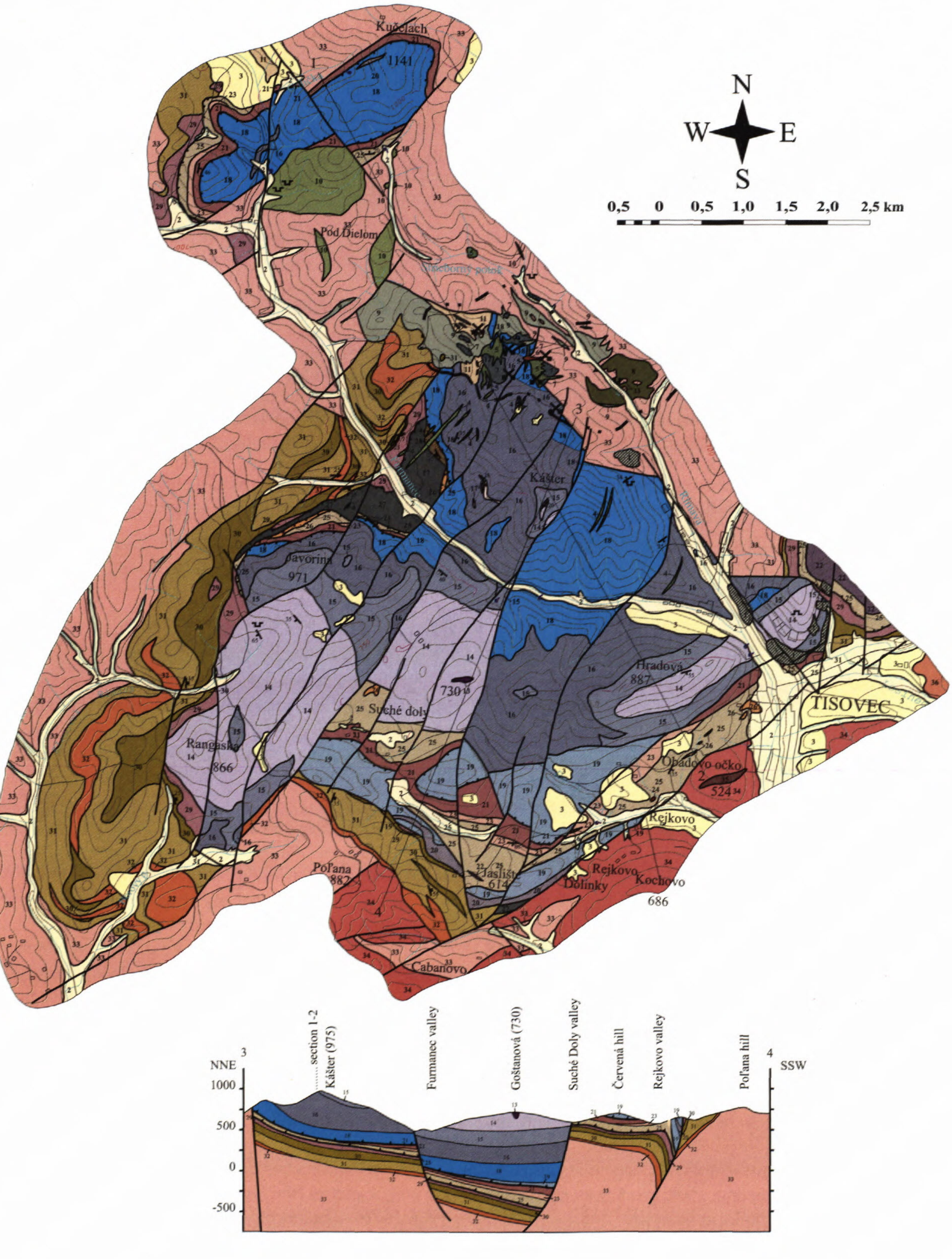


Fig. 1 Geological map of the Tisovec karst and Kučelach massif with geological cross-section and explanations (after Vojtko, 1999).

EXPLANATIONS TO GEOLOGICAL MAP AND PROFILES

(1:25 000)

QUARTERNARY

Holocene - Pleistocene

1 antropogenous deposits; waste dumps and transferred piles

2 fluvial sediments; river and brook alluvium, gravels, sands, loames

3 deluvial sediments; loamy-gravelous

TERTIARY

VEPOR VOLCANO-PLUTONIC COMPLEX

Neogene - Miocene

4 veins of andesites; basalt andesites

5 subvolcanic andesites; amphibole-pyroxene andesites

6 scarns

7 diorites; pyroxene diorites to dioritic porphyrites

8 diorites; biotite-pyroxene diorites

9 subvolcanic andesites; pyroxene andesites

10 subvolcanic andesites; garnet-pyroxene (biotite) andesites

SUBTATRAS GROUP

Paleogene - Eocene

11 Huty Formation?; dark claystones, locally sandstones (Priabonian)

12 Huty Formation?; layers of sandstones and conglomerates (Priabonian?)

MESOZOIC

SILICA UNIT - MURÁŇ NAPPE

JURASSIC - TRIASSIC

13 Goštanová Limestones; dark crinoidal limestones (Rhaetian-Liassic)

14 Dachstein Limestones; light to dark coarse-bedded (Norian - Rhaetian?)

15 Tisovec Limestones; light reef limestones, locally brecciated (Julian-Tuvalian)

16 Wetterstein Dolomites; light-grey, locally dark dolomites (Cordevolian)

17 light-pink crystalline limestones and dolomites (Cordevolian?)

18 Wetterstein Limestones; light-grey, grey with dolomitic lenses (Ladinian)

19 Steinalm Limestones; light massive locally thick-bedded (Pelsonian-Illyrian)

20 Gutenstein Beds; dark-grey limestones and dolomites (Aegenian-Bithynian)

TORNA UNIT

Triassic

21 Reifling, Nádaška, Schreyeralm Limestones and Gutenstein Beds (Pelsonian-Sinemurian)

22 Gutenstein Beds; dark-grey carbonates (Aegenian-Bithynian)

23 Szin Beds; sandstones, shales, marlstones and limestones (Scythian)

24 rhyolitic pyrhoclastics (Scythian)

25 Bódvaszilas Beds; sandstones and shales (Scythian)

26 tectonic breccias and rauhwackes

LATE PALEOZOIC

GEMER UNIT - DOBŠINÁ GROUP

Late Carboniferous

27 Hámor Formation?; shales, sandstones and conglomerates

28 Ochtiná Formations; grey shales and marlstones, dark limestones and conglomerates

PALEOZOIC - MESOZOIC

VEPOR UNIT - FEDERATA SEQUENCE

Permian - Triassic

29 Tuhár Succession?; rauhwackes, dolomites and limestones (Triassic)

30 clayed and sandy shales (Scythian)

31 Lúžna Formation; quartzites and quartzitic sandstones (Scythian)

32 Rimava Formation; arkosed sandstones and conglomerates (Permian)

VEPOR CRYSTALLINE BASEMENT

Paleozoic

33 granitoides; mainly porphyric granitoides (Carboniferous)

34 deformed hybrid granitoides and migmatites (Early Paleozoic?)

35 biotite gneisses (Early Paleozoic)

36 Muráň Orthogneisses (Early Paleozoic?)

SYMBOLS

lithological boundaries

faults; observed, inferred, covered

thrust fault

35 bedding (dip of direction)

25 schistosity (dip of direction)

springs

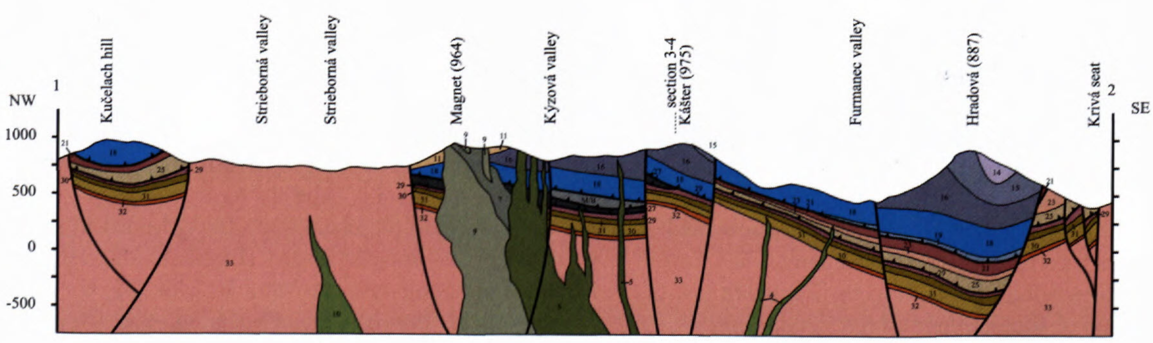
mines out of operation

quaries; in operation, abandoned

TV-10 boreholes

1 2 lines of geological sections

area of the contact metamorphose



The Hybrid complex

Rocks of the Hybrid complex are distributed at the surface in a local area southeast of the Muráň fault zone (Fig. 1). These rocks are probably the oldest elements of the Vepor Unit. The Hybrid complex is made up of remnants of the high-metamorphic gneiss-migmatite mantle, which are incorporated into granitoids of this complex. There are two types of the hybrid granitoids. The first one is distinctively foliated, displaying laminated and augen fabric. The second one is formed of homogeneous granitoids, which developed only weak foliation. Their composition ranges from granodiorite to tonalite (Bezák & Hraško, 1992).

The Kráľová hoľa complex

The basement of all tectonic units is formed by the Kráľová hoľa complex, distributed between the Muráň fault zone and the Zbojská saddle northwestern part of the area.

Overlying the Kráľová hoľa complex is the Federata Sequence, which is situated on granitoid rocks of the complex (Klinec, 1966;1976). It is viewed as the Kráľová hoľa nappe (sensu Putiš, 1989). The Kráľová hoľa complex mainly consists of granitoids of the Vepor pluton. There are middle-coarse grained biotite granodiorites, porphyric granodiorites to granites (Vepor type s.s.), and locally small bodies of coarse-grained diorite (as clasts in debris in the Štrompľovská dolina and Rimavská dolina valleys) and xenolithes of paragneisses (Vojtko, 1999). The pegmatite veins were found locally in this complex. These granitoids are deformed, metamorphosed and to different degrees mylonitised to phyllonitised at the contact with overlying cover sequence (Vrána, 1966; Lexa & Bezák, 1996).

The Federata Cover Sequence (the Permian-Middle Triassic)

The Federata Sequence forms an autochtonous or paraautochtonous sedimentary cover of the Prealpine Southern Vepor basement. Its probable age ranges from the Permian to the Middle-Upper Triassic. This sequence, together with its basement, is epimetamorphosed and intensively deformed in a ductile regime. The Federata Sequence is situated beneath the Gemer Unit.

The Federata Sequence consists of four formations and forms an irregular rim at the western and southern edge of the Tisovec karst and part of the eastern periphery of the mapped area. It occurs also west of the Kučelach outlier of the Silica nappe, where carbonate members are better preserved.

The Rimava Formation (Permian)

The Rimava Formation consists of arkosic quartzitic sandstones and conglomerates. It had strongly a low-grade penetration metamorphic foliation and mineral lineations. The post-deformational thickness of the formation randes from several metres to one hundred metres. The rocks lithologically correspond to the Rimava Formation s.s. of the Revúca Group of southern part of the Vepor Unit.

The Lúžna Formation (Lower Scythian)

The Lúžna Formation consists of a pale grey-green to white imbricated and cleavaged fine-grained quartzitic sandstones, which are deformed and metamorphosed and they have penetration foliations and lineations. Its maximal thickness is 100-200 metres and its composition corresponds to the Lúžna Formation s.s. (Bezák, et al., 1999).

Between the Permian arkoses and the Scythian quartzites are gradual lithological changes. The quartzites and the quartzitic sandstones represent a normal lithological sedimentary succession. The Lúžna Formation pass to the Werfen Formation gradually.

The Werfen Formation (Upper Scythian)

The Werfen Formation consists of dark grinish-grey (originally) clayed, silty and sandy shales with thin intercalations of light-colored quartzitic sandstones in lower part, which indicate gradual development from underlying quartzites in the marine origine.

The Gemer Unit – the Dobšiná Group

The Dobšiná Group is in the tectonic contact both with its underlying (the Federata Sequence) and the overlying Turňa Unit(?), locally the Muráň nappe in Magnet hill domain.

The sequence of sediments belonging to the Ochtiná Formation of the Dobšiná Group is newly described tectonic unit in the structure of this area (Soták & Plašienka, in press).

This sedimentary succesion was formerly described as the Ipoltica Group of the Choč nappe (Vozárová & Vozár, 1988). It forms a tectonic slice between the detrital probably the Upper Carboniferous sediments and the rhauwackized tectonic breccia along the base. The sediments of this formation consist of fine-grained, grey, phyllitic and commonly marly shales with the layers of quartzitic conglomerates. The grey and brown carbonates create both thick beds of the massive or bioclastic limestones and thin imbricated detritic sandy and bioclastic, mainly crinoidal limestones. The limestones are locally ankeritized. The crinoidal internodes are very abundant in the crinoidal limestones. Their thickness is about 20 mm. Locally there are solitary corals with the size about 45 mm. Black carbonates and sandy crinoidal limestones are in the Ochtiná Formation too. The Carboniferous age of the limestones was determined by their microfauna from the Furmanec valley. The microfauna populations (*Stacheoides* and forams *Archaediscus karreri* and *Nanicella sp.*) indicate the Visean age of the limestones (Soták & Plašienka, in press; Bezák et al., 1999).

The sequence of detrital dark sediments is provisionally correlated with the Hámor Formation (the Gemer Unit). The Hámor Formation contrary to the Ochtiná Formation has not carbonate members and clastical mica in phyllites, sandstones and conglomerates (Bezák et al., 1999). The Hámor Formation is comparatible with the Nižná Boca Formation of the Ipoltica Group of the Choč nappe

(Vozárová & Vozár, 1988), but it is tightly spatially and deformationally connected with the Ochtiná Formation of the Gemeric Unit. Rocks observed in the mapped area are lithologically similar to the Hámor Formation (Soták & Plašienka, in press; Bezák et al., 1999). New data mainly from boreholes that are located at the eastern slope of Magnet hill points out also to other solution of this problems, as mentioned below.

The Meliata Unit(?)

Among the borehole drilled in Magnet hill domain, realised while prospecting for of scarn and polymetalic Pb-Zn (Cu) ores, is a very interesting borehole TV-10 (Bacsó & Valko, 1969). This borehole TV-10 cut the Carnian Wetterstein Dolomite, the Ladinian Wetterstein Limestone and the Steinalm Limestone (the Pelsonian-Illyrian) and at the depth span 411-520 m was spotted interesting succesion of rocks including evaporites. This borehole was situated outside of the area affected by the contact metamorphosis, which was considerable because size of grains in recrystallised limestones is often more then 15 mm. Evaporites do not occur in the other boreholes probably due to the effects of contact metamorphosis.

Borehole TV-10 was projected to resolve the lithostratigraphy of this area. Complete lithological profile of the borehole TV-10 is listed below (Bacsó & Valko, 1969):

411,0-430,0 m	black phyllites
430,0-434,1 m	white to grey fine-grained gypsum
434,1-435,5 m	gypsum with layers of graphitic phyllites
435,5-445,5 m	graphitic phyllites
445,5-446,0 m	white-grey fine-grained gypsum
446,0-485,0 m	graphitic phyllites locally with quartzitic veinlets
485,0-487,5 m	quarzite-graphitic phyllites
487,5-489,5 m	chlotite-quartzitic phyllites
489,5-494,0 m	white-grey, light green fine-grained to massive gypsum
494,0-494,5 m	breccia of gypsum and graphitic phyllites
494,5-501,0 m	graphitic phyllites
501,0-507,0 m	chlorite-quartzitic phyllites
507,0-520,0 m	green quartzite-chloritic phyllites disseminated by the pyrite

This succession Bacsó (1973) interpreted as "the Gemer Carboniferous sediments". This idea contradicts the well-known lithology in the area of "the Carboniferous strip" in the Gemer Unit between Dobšiná town and the Podrečany village. Bacsó & Valko (1969) did not excluded this succession represents the Meliata Unit. This for this area surprising succession of rocks has not been noticed for 30 years.

A detailed study of rocks of the borehole TV-10 is necessary for tectonic interpretation of this succession, and other boreholes in Magnet hill domain also could be important for solution of this problem. On the base of above mentioned lithology we do not rule out, that at least part of these rocks represents the Meliata Unit. The duplexes of the Meliata rocks (?) in this area are in a similar position as the evaporite melanges of the Meliata Unit, which were described at many places of southern Slovakia and northern Hungary (Bystrický & Fusán, 1961 in Mello et al., 1997; Bystrický & Oravcová, 1962 in Mello et al., 1997; Réti, 1985; Kozur & Réti, 1986; Horváth, 1997).

The Turňa Unit(?)

The most recent study about the Muráň nappe let to extensive changes of the stratigraphy of its, mainly through the study of cherty limestones (before regarded as the Gutenstein limestone). Havrila (1997) described Conodonta, Holothuria and problematic rests of organisms of the Cordevolian age. On the basis of these data Havrila (l.c.) separated from the Muráň nappe of the Silica Unit "the lower Muráň nappe of the Silica Unit" (consisting of the entire lower part of the succession beneath the Steinalm Limestone). The Steinalm Limestone and higher succession is regarded as the Muráň nappe s.s. The lower Muráň nappe overlies of the Dobšiná Group or the Federata Sequence and is covered by the Muráň nappe s.s. We provisionally include it with the Turňa Unit (sensu Less et al., 1981). This succession was separated from the Muráň nappe s.l. on account of the stratigraphic clasification of dark often cherty and having nodular limestones lying beneath of the Steinalm Limestone belonging to the southern Vepor Unit. The Turňa Unit(?) has features of a duplex structure. It locally taper out, mainly in Magnet hill domain. If we admit that rocks included into to the Hámor Formation (maybe part from them) at the surface can belong to the Meliata Unit then their position is analogous to that in the Slovak Karst area. We include this succession into the Turňa Unit (equivalent of the Slovenská skala nappe) provisionally on the basis of a very similar lithology, stratigraphy, anchimetamorphosis absence of Wetterstein facies and commonly the absence of the light Steinalm or Honce Limestones under the base of dark largely cherty limestones. On the other hand there are rhyolite pyroclastics of the Scythian age in the mapped area, while they are missing in the Silica and the Turňa Units in the area of the Slovak Karst.

The occurences of volcanic rocks are not well substantialed in the saddles north of Kerek and Dlhý hills, in the Turňa Depression and on the southern slope of Horný vrch hill (Melo, 1979). These rocks are included in the Turňa Unit (Mello et al., 1997).

Rocks of rhyolite composition were found in the Muránska planina plateau area at several places (near Telgart village – at Gregová hill where there occurs the largest body of the Scythian rhyolites in the Central Western Carpathians; in the Rácovo valley; on the northern slope of Veľká Stožka hill, Klinec, 1976; and locally debris in the Kačkava stream), in the Tisovec karst area (the Rejkovská dolina valley-in terraces and locality in situ). The rock occurences of the Scythian rhyolite are in the Drienok nappe and also at the base of the Neogene volcanic rocks and the Paleogene rocks of the Štiavnica Stratovolcano (Vozár, 1969; Vozár, 1973). This distribution point to the likeli-

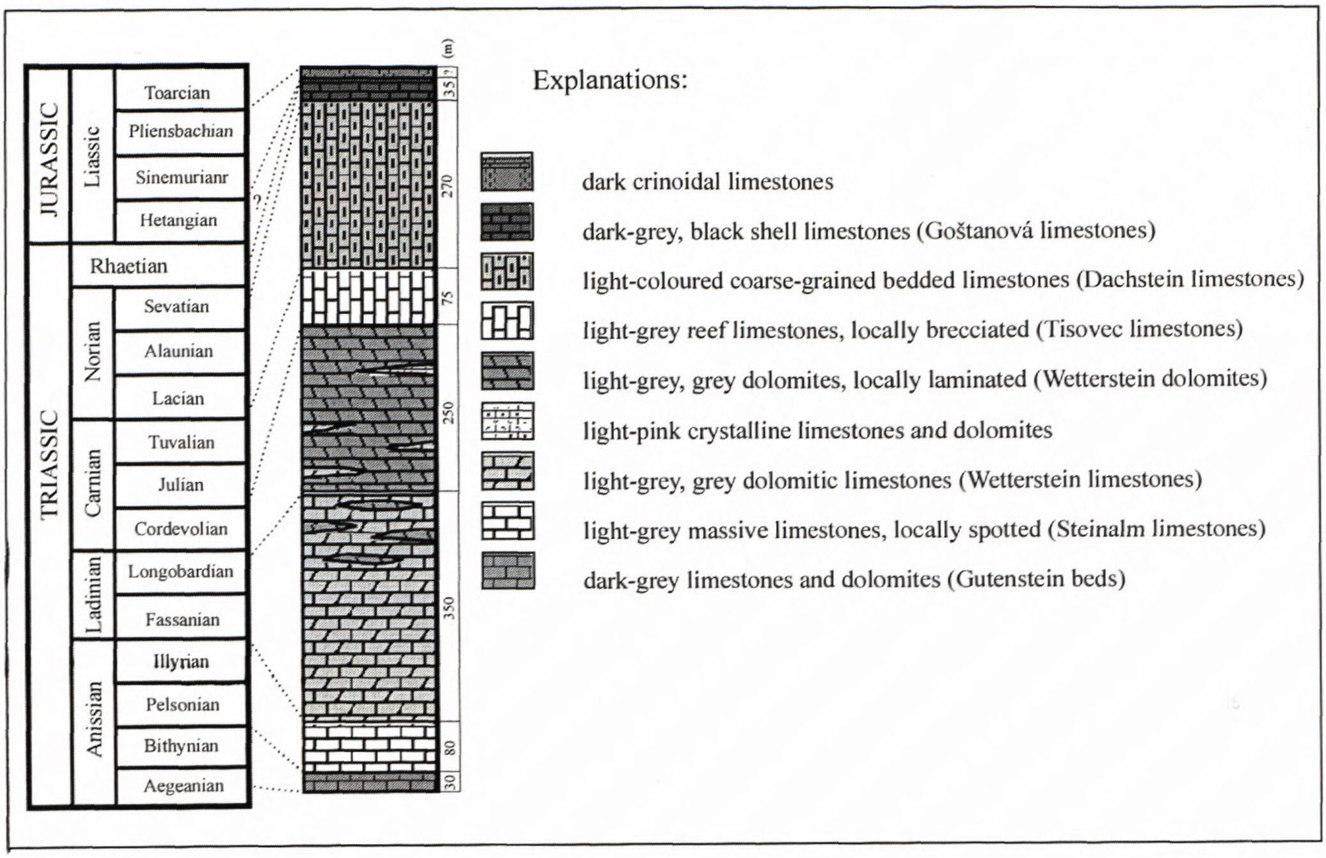


Fig. 2 Lithostratigraphic column of the Silica Unit (the Muráň nappe) in the Tisovec karst and the Kučelach massif area (Vojtko, original figure).

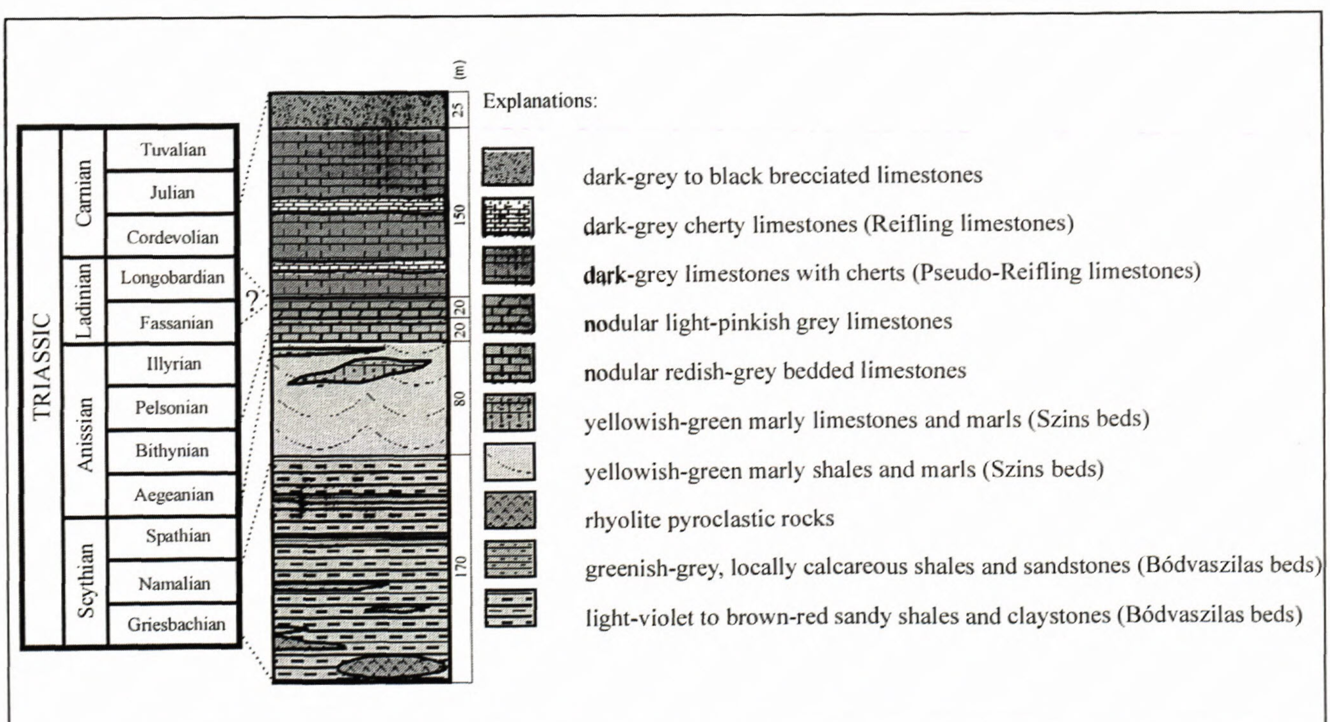


Fig. 3 Lithostratigraphic column of the Torna Unit (?) in the Tisovec karst and the Kučelach massif area (after Vojtko, 1999).

hood that these units in the base of the Muráň nappe, the Drienok nappe and the unit under the base of the Štiavnica Stratovolcano have at least lithological, if not also tectonic similarities. In the Drienok nappe bedded members of the Scythian age, with rhyolite, lying under the Wetterstein Limestone have the same position as in the Tisovec Karst and in the Muránska planina plateau and they could belong to the Turňa Unit too.

In the Turňa Unit in the Slovak Karst there occurs acid volcanic rocks of the Permian age. The rhyolites are even directly in the base of the Bódvaszilas beds and upon the sediments of the Dobšiná Group(?) (maybe the Meliata Unit) in the Rácovo valley.

Bódvaszilas beds (Griensbachian-Namalian)

The Bódvaszilas beds are the lowermost member of the Turňa Unit(?) with an irregular occurence in the mapped area. They occur mainly south of the Hradová fault, where they underlie an outcrop band about 250 m wide, that run from Tisovec town and the Krivá saddle domain to the Rejkovská dolina, Jaslište and Suché Doly valley. Small occurences of this formation are also found on the northwestern slope of Javorina hill, on the southern slope of Magnet hill and Kučelach and Čeremošná hills. They are formed of variegated sandstones and shales, of violet, green and grey colours (Fig. 3).

Szin beds (Namalian-Spathian)

The Szin beds also occur south of the Hradová fault on the southern slope of Hradová hill and the Jaslište area. In the Kučelach outlier they occur only at the western slope of Remetisko hill and the Koryto valley. A local occurence of the Szin beds was also observed on the northwestern slope of Javorina hill.

These rocks are the best exposed on the northern slope of Jaslište hill, where occur marly limestones to lumachela limestones. They are formed by the grey to yellowish-green marly shales and grey to greenish-grey bedded marly limestones with conchoidal fractures (maybe caused by the anchimetamorphosis).

Carbonate complex (Anisian-Carnian?)

This assemblage made us of dark to black limestones, which occurs mainly in the Kučelach outlier and in the Tisovec karst around Červená hill. Small occurences are also founded on the northern slope of Javorina hill.

This complex is formed by the Gutenstein Limestone and Dolomite, Nádaska Limestones and Reifling Limestones (Uppermost Ladinian-Cordevolian) with foraminifera *Turriglominna mesotrasica* ZANINETTI determined by Soták. Reifling Limestones have the character of a basinal facies with cherts.

The Silica Unit – Muráň nappe

The Silica Unit is represented by the Muráň nappe in the study area. It is probably gravitional nappe. Its stratigraphical range of sedimentary rocks is from the Anissian to the Liassic (Fig.2). They are non-metamorphic rocks with a thickness of at much as 800 metres. The Silica Unit forms the uppermost Mesoalpine unit in the Tisovec karst. In the basement of this unit is the Turňa Unit(?) except at the Magnet hill domain. Under the Muráň nappe are the Federata Sequence, the Dobšiná Group of the Gemer Unit or the Meliata Unit in the area of the Magnet hill (cut by borehole TV-10).

Gutenstein Formation (Aegeanian-Bithynian)

Occurences of the Gutenstein Formation are rare, and limited to the south and southeast side of the Hradová fault. The sediments of this formation are preserved only as lenticular bodies.

The Gutenstein Formation is chert free dark thick-bedded limestones with characteristic veins of the white calcite. Dark limestones replaced by dolomite with the characteristic fracturing of the dolomite.

Both their distribution and inclusion into the Muráň nappe are problematic. In the Muráň nappe are included only the Gutenstein Limestone and Dolomite, which are incorporated to the lenticular bodies in the Jaslište domain and to the Muráň fault zone, but we do not rule on that they could belong to the Turňa Unit(?).

Part of dark limestone could belong to the Muráň nappe. It is above the Reifling Limestone with cherts (the Turňa Unit(?)) and below the Steinalm Limestone on the southern slope of Červená hill. Precision of stratigraphy is necessary for outcrop identification and the inclusion of the Gutenstein Formation to the corresponding tectonic unit.

Steinalm Limestone (Pelsonian-Illyrian)

They occur only at the southern part of the study area at Červená and Podhrad hills and they also form tectonic lenticular bodies on the northern slope of Kochovo hill and in the Jaslište domain.

They are light-coloured massive, commonly spotted crinoidal limestones or light-coloured thick-bedded to massive limestones with a finely brecciated fabric. The Steinalm Limestone contain dasycladaceans of the genus *Physsoporella.* The foraminiferas *Meandrospira dinnarica* were described from Červená hill (Salaj et al., 1983).

Wetterstein Limestone (Fassanian-Cordevolian)

The largest occurence of this formation is at the Kášter hill domain and northern slope of Grúniky hill. They occur also in the central part of the Kučelach-Remetisko syncline. Smaller occurences are on the northern slope of Kereška hill and on the southern slopes of Magnet and Pacherka hills. The Wetterstein Limestone are locally dolomitised or changed to dolomites in the Tisovec Karst. Dolomites form irregular bodies, mainly lenses or layers in this limestones. The Wetterstein Limestone are enriched upward by dolomite component. The Wetterstein Limestone gradually change to the Wetterstein Dolomite of the Carnian age. The Wetterstein Limestone are light-grey or grey, massive, and locally thick bedded. They often contain Dasycladaceae. The *Teutloporella herculea* (STOPP.) commonly occurs in the Wetterstein Limestone and big gastropods were found locally.

The upper part of the Wetterstein Limestone is formed by Wetterstein dolomite. This dolomite create an important layer for geological mapping as a distinctive marker because they separate a lower carbonate platform from an upper carbonate platform. This geological layer is well developed in this territory. The thickness of the Wetterstein dolomite is 75 to 375 m; about 250 m is an average thickness.

They occur almost in all of the karst, and mainly on the northern slopes of Grúniky, Hradová, Javorina, Kereške hills and on the central part of Pacherka hill.

Dolomites are light grey to grey, locally white and dark. They have a grained or massive fabric, bedding is visible mainly as the alternating dark and light thin beds. In their lower part occur lenses of light limestones and locally are pink crystalline bedded limestones. This crystalline limestones are describe Bezák et al. (1996), probably also by Bystrický (1959). Limestone lenses are syngenetic and their bedding is congruent with bedding of the neighbouring dolomites. Therefore, we do not expect that these lenses are neptunean dikes of the Jurassic limestones as proposed by Bezák et al., 1996.

Tisovec Limestone (Julian-Tuvalian)

The Tisovec Limestone (sensu Soták 1990) occur mainly on the northern slope of Hradová hill where it forms huge cliffs, as well as at the area of Kášter hill. The Tisovec Limestone do not form a distinctive morphological contrast in the domain of Javorina hill. Their geological age is the Juvalian to Tuvalian. These limestones were not confirmed on the type locality by Krystyn et al. (1990), but Soták (1990) confirmed their occurence. In this paper we use this term for limestones directly overlying the Wetterstein dolomites and the base of the bedded Dachstein limestones. These limestones are light-coloured, locally grey, grained, and bioclastic. Their the most important feature is a brecciaed fabric, which is expressive mainly in the lower part of this layer and second one is the filling of caverns by the aragonite druses between clasts. These druses are recrystallised to calcite. This phenomenon is confirmed by the habitus, which is also still aragonite and locally calcite. These characteristic features were main reasons, to separate them from the bedded Dachstein Limestone.

Dachstein Limestone (Norian-Lower Rhaetian?)

Above the light-coloured the Tisovec Limestone occur light grey to grey limestones, which are distinctively bedded. They occur in all the profile from base to the top in the domain of the Teplica and Suché doly valley, Goštanová hill. Other incomplete occurences are in the Tepličné, Rangaska, Hradová, Čremošná and Kášter hills. They provide records of cyclical sedimentation with features of emersion which were described by Borza (1977).

The Dachstein Limestone are microscopically sparitic and towards the top increases in micrite and only part of limestones are micritic with fenestral structures filled by coarse pellucid sparite on the border between the Sevathian and the Rhaetian(?). Among the fossils were found Amodiscidal Foraminiferas of the species *Agathammina inconstans, Earlandia sp., Opthalmidium triadicum* KRISTAN; and questionable Involutinidal Foraminiferas with manifestations of hard sparitisation and the form of the Nodosarial Foraminiferas. The crinoidal internodes, Ostracoda and oval paramorphosis of the organisms represent a lagoonal facies. The fossils and the Upper Sevathian and the Lower Rhaetian(?) age of this limestones were determined by Soták.

Goštanová Limestone (Rhaetian)

The Goštanová Limestone was mapped by Biely in the Muráň nappe and was named by Michalík (in Bystrický et al., 1973). They were correlated by Kochanová with the limestones of the Bleskový prameň domain on account of their bivalves.

The fauna of the brachiopods (l.c.) is different from the fauna of the Bleskový prameň limestones but similar to the brachiopods of the Hybe Formation (*Rhaetina pyriformis, Zeilleria norica, Zeilleria elliptica, Euxinella subrimosa, Austririnchia cornigeria, Sinucosta emmrichi, Zugmaerella koessenensis*). The Goštanová Limestone are medium grained, grey to pale grey, crinoidal biosparite with brachiopods and bivalves (Michalík; 1977, 1980). They form the lenticular sedimentary bodies and the neptunean dikes, too, in the uppermost part of the Dachstein Limestone succession (Michalík, 1977).

„Crinoidal" limestones (Pliensbachian-Sinemurian)

This formation is represented only by dark crinoidal limestones, which occur very locally in the Tisovec Karst. Goštanová hill is their only occurence, where the limestones form a morphologically conspicuous shallow E-W trending depresion in the marking an erosional remnant or Neptunean dike (Fig. 1). They are above the Dachstein Limestone and are the youngest member of the Muráň nappe in the Tisovec Karst. These limestones are wackestone microscopically, which contain ostracods of the family Ogmoconcha; Amodiscidal Foraminifera – small forms of the Frondicularia, species *Amodiscus incertus* d'ORBIGNY and *Amodiscus multivolutus* REITLINGER; Nodosarial Foraminifera – species *Nodosaria nitidana* BRAND, *Agathamina austroalpina* KRISTAN-TOLLMANN et TOLLMANN; fragments of bivalvia first and foremost the family Periostraca. They contain also crinoidal internodes, sea urchin spines and thecas of punctate brachiopods. These fossils have been described by Soták who interpreted their age as the Pliensbachian-Sinemúrian. Sediments of the Hetangian age have not been found. I think that they do not occur in this territory.

The Subtatras Group

The Subtatras Group was well-known only in the erossional remnants in the Zbojská saddle, the Kučelach massive. The new locality of this group form the sedimentary cover at Magnet hill which is cut into two parts by pyroxenic andesite and pyroxenic diorite.

The northern part is formed by dark Globigerine marly claystones, fine-grained sandy claystones to fine-grainded sandstones. These rocks are contact metamorphosed by Neogene magmatic intrusions and they are converted to shales with cherts and locally by porcelanites (Fig. 1).

Quartzitic conglomerates are distributed in the southern part of the area. These conglomerates create lenses in claystones reported by Bacsó (1964), but he considered all formations from northern part of the area as the Carboniferous sediments of the Gemer Unit (Bacsó; 1964, 1973; Bacsó & Valko, 1969). Underlying these conglomerates are the Wetterstein Limestone and Dolomite of the Muráň nappe, which is contact metamorphosed in the Magnet hill domain. These carbonates are metamorphosed to marbles, as shown in the TV-4 and TV-9 boreholes (Bacsó & Valko, 1969). The Paleogene sediments are at most about 60 m thick and they dip WSW 5°-10° and in the southern part to 20°. Globigerine claystones represent the basinal facies.

The Neogene volcano-plutonic complex

The subvolcanic levels expected in the Tisovec stratovolcano were eroded due to the uplift in the Veporské vrchy Mts. and in this domain were not preserved characteristic morphological features in contrast to the Eastern Slovakian and Central Slovakian volcano-plutonic complexes. Erosion created unique conditions for study of volcanic and geological relations of the deposits and the deeper levels of the volcano-plutonic complex at the surface. Volcano-plutonic complex of the Veporské vrchy Mts. is divided to several lithostratigraphical formations (Burian et al., 1985; Vojtko, 1999).

The Železnícke predhorie Formation is located north of the Rimava basin. This formation build up of the Pokoradz and Blh platforms NNE from Rimavská Sobota town. The peripheral part of the Vepor volcano-plutonic complex is formed by the Železnícke predhorie formation. This formation comprises by explosive products, extrusive sheets, tuffs, tuffites, agglomerates and lahars. These rocks include pyroxene andesites and hornblendpyroxene andesites, which are autometamorphosed (Marková & Vaňová in Burian et al., 1985).

Both types of andesites are related to well-known types in the Tisovec intrusive complex and they are probably products of this intrusive complex. The age of the amphibole-pyroxene andesites was determined by the fission track method as the 16.4 ± 0.6 MA from the epiclastical conglomerates (Višňové village) and from the pyrhoclastic flow of the pyroxene andesites as the 16.2 ± 0.8 MA (the Lower Badenian) near Chvalová village (Repčok, 1981). We expect that subvolcano intrusive rocks Tisovec town have the same age (Vojtko, 1999). The age of the basal formation near Vyšná Pokoradz village was determined by means of macroflora as the Sarmathian (Němejc in Fusán et al., 1962).

The centre of the Vepor volcano-plutonic complex was probably in Tisovec town area according to the volcano paleoflow tracks. These paleoflow tracks are directed from NNW to SSE (Konečný & Lexa in Vass et al., 1982).

The Hájna hora Formation consists of relict of the volcano-sedimentary bedding south of the Brezno depression. Its thickness is 150 m (Konečný & Lexa in Ivanička, 1986). This formation represents a paleovalley fill of NW-SE-trending. NW of the proposed volcanic centre (Magnet hill) fine-grained sediments and sandy sediments predominate. Amphibole, pyroxene, amphibole-pyroxene andesites and andesites with accessory garnet are present in this formation as fragments of volcanic rocks (Konečný & Lexa in Ivanička et al., 1986).

The Pacherka Formation is characterised by penetration of subvolcanic dikes in the southeast zone of the Tisovec intrusive complex. The composition of volcanic rocks is basaltic andesite to basalt. This is the last volcanic activity.

The Magnet hill Formation consists of amphibole-pyroxene andesites, which are very well developed in Tisovec town area. Generally, bodies of amphibole-pyroxene andesites occupy area about 4000 m long and 1500 m width, elongated NW-SE.

The typical phenomenon of andesite intrusions with amphibole-pyroxene composition is that they cut diorite bodies. This phenomenon makes posible to determine their relative succession and age. The Tisovec intrusive complex makes the central zone of the Vepor volcano-plutonic complex, which is situated in the wider domain of Magnet hill in the studied area. The Tisovec intrusive complex is built up of nine diorite bodies with irregular shape, which extent is 2,500 x 500 m. Their occurence begins near top of Magnet hill and ends on the southwestern slope of Huta hill. The biggest and easternmost diorite body is in the middle of the body of porphyric granodiorites. These are relatively strong contact metamorphosed to a distance of 300 m, and rarely even more. This body has the composition of a biotite-pyroxene diorite (Bacsó & Valko, 1969; Bacsó, 1973).

Other diorite bodies are situated in Magnet hill domain and their surrouding is built up of a pyroxene andesites, the Wetterstein Limestone and dolomites of the Muráň nappe and also locally by the Paleogene dark globigerina marly shales.

Hyperstene-diopside diorites dominate especially in the easternmost body and occur as apophysal protrusions, which consist of pyroxene diorite porphyry. The southern part of the largest body consists of pyroxene-quartzitic diorite (Bascó, 1964). The four easternmost diorite bodies at Magnet hill are developed directly in the Tisovec fault zone. Most of the diorite bodies pass to porphyric varieties at their margins.

The Vepor formation in Magnet hill domain is characterised by the penetration of pyroxene andesites, which are relatively older than diorite bodies. The are most abundant

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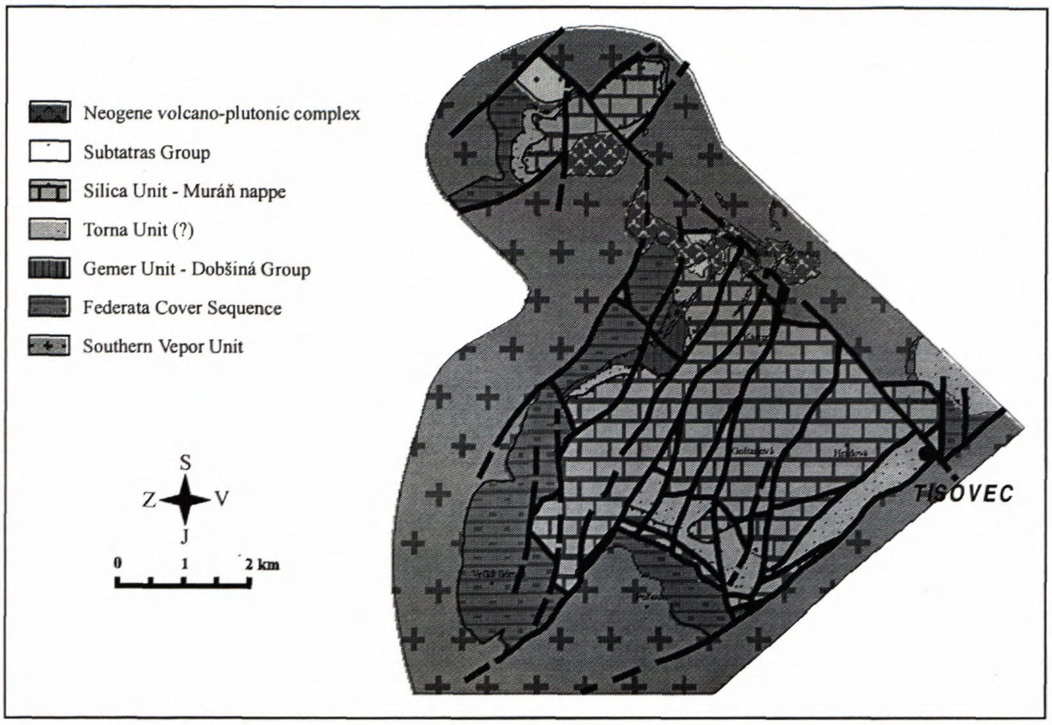

Fig. 4 Schematic tectonic sketch of the Tisovec karst and the Kučelach massif (Vojtko, original figure).

at the western and eastern parts of Magnet hill and southern part of Huta hill.

The Strieborný potok formation occurs northwest of Magnet hill in the study area and consists of garnet-pyroxene±biotite andesites. By analogy with others volcanic Mts. of central Slovakia, they are considered to be the oldest ones. (Konečný, 1998; Šimon, 1997).

Tectonics

The studied area is built up by these superposed tectonic units, which are listed from the bottom to the top:

Paleo-Meso alpine epizonally metamorphosed nappe units (Fig.4):

a) the southern Vepor Unit with the Federata Sequence
b) the Gemer Unit (the Dobšiná Group)
c) the Meliata Unit s.l.(?)

Mesoalpine anchizonally metamorphosed and non-metamorphosed nappe units:

a) the Turňa Unit(?)
b) the Silica Unit (the Muráň nappe)

Post-nappe formations:

a) the Subtatras Group
b) the Neogene volcano-plutonic complex (the Tisovec stratovolcano).

The Paleoalpine epizonally metamorphosed nappe units have features of low-grade metamorphosed recrystallization and ductile deformation, which were overprinted by structures of younger deformation stages (Plašienka, 1981,1993).

The Mesoalpine anchimetamorphosed to non-metamorphosed nappe units are without features of epizonal metamorphosis. Post-nappe formations are deformed only in the brittle regimes. Older deformation stages connected with metamorphosis were not in the focus our interest and so we will not deal with them here.

The pregosau displacement of the Muráň nappe represents the oldest tectonic event in our interest. Both the Turňa(?) and the Silica Units are superficial nappes. Their displacement had only minimal influence on the para-autochtonous Veporic basement (Bezák et al., 1999).

Deformations close to the décolement products of brittle cataclasis, crushing and brecciation of carbonates of the Federata Sequence and the rocks of the Turňa nappe(?). Rauwackised carbonate tectonic breccias held many features of the hydrotectonic phenomena and complicate fluidal regimes approximately in the environment of pore fluids overpressure. These phenomena a enabled smooth displacement of superficial nappes (Plašienka & Soták, 1996; Milovský, 1996; Milovský et al.,1998; Milovský, 2000).

Cataclasis and rauwackization appear mainly along the base of the Turňa Unit(?). On the contact between the Turňa(?) and Silica Units only cataclastic fracturing is developed. The maximum thickness of this zone is 5 m. This phenomenon partly confirms that the Silica nappe was probably carried passively upon the Turňa Unit(?) (Vojtko, 1999), that means the Turňa Unit could be in para-autochtonous position in relation to the Silica Unit.

The tectonics in the area of the Tisovec Karst and Kučelach massif is very complicated, with a evolution of important fault structures. The character of the tectonic structure of the studied area was influenced by the two regional faults (the Muráň and Tisovec faults), which cross on the eastern part of the Tisovec karst.

During the Late Cretaceous time at higher structural levels dextral movement occured along the NW-SE oriented faults (Plašienka, 1993). N-S compression lasted during the Late Cretaceous and the Paleogene ages. The structures due to cooling and uplift became „colder“ and deformation was localised to sinistral SW-NE (the Muráň fault; Marko, 1993a) and dextral SE-NW (the Mýto--Tisovec fault zone; Marko, 1993b) brittle fault zones.

Within the Muráň fault zone operated sinistral transpression. In the Muráň nappe were formed fan like synforms of WSW-ENE strike (Marko, 1993a). The south wings of this synforms are considerably compressed and they have complicated fold-duplexes structure like the south wings of the Tesná skala synform sensu (Bystrický, 1959) the Hradová and Červená synforms in the Tisovec Karst and in Šarkanica hill domain (Marko, 1993a; Plašienka, 1993; Bezák et al., 1999; Vojtko, 1999). The eastern edges of these synforms were truncated by the Muráň fault during the Paleogene (the Pre-Oligocene) period (Marko, 1993a).

The youngest are extensional normal faults, which renew the mainly SW-NE originally transpressional faults. The normal NNE-SSW faults are for geomorphological and the Quarternary development of the Tisovec Karst the most important. They cut the whole karst area and create structure of the irregular grabens with maximum subsidence in the Suché Doly valley and Goštanová hill domain. NNE-SSW faults were supply ways for the Neogene volcanic rocks of Magnet, Pacherka and Kášter hills (Bacsó, 1964, 1973; Bezák et al., 1996; Vojtko, 1999). NNE-SSW normal faults mediate movement of ground waters from the ponores in the Suché Doly domain to the exsurgences in the Furmanecká dolina valley, too. Probably the Badenian-Sarmathian age of volcanic rocks and normal character of the NNE-SSW faults correspond with NNE-SSW orientation of principal compressional stress ascertained for this period in the western part (Marko et al., 1995) and central part of Western Carpathians (Kováč & Hók, 1993).

Conclusions

The geological mapping of the Tisovec Karst confirmed that geological structure and its evolution is more complicated than it was expected in the past.

The southern Vepor Unit forms the basement of the subautochtonous Federata Cover Sequence in the Tisovec Karst area. Metamorphosed formations (the Ochtiná and Hámor Formations) of the Dobšiná Group (the Gemer Unit) are overthrusted. New described tectonic units in this area are the Meliata(?) and Turňa(?) Units. As Turňa Unit(?) we regard members under the Steinalm Limestone of the Muráň nappe (the Silica Unit) with stratigraphical range from the Scythian (the Upper Permian?) to the Cordevolian. The Muráň nappe has a stratigraphical range only from the Anissian to the Pliensbachian, which is stratigraphically confirmed an erossion remnant on Goštanová hill.

Five eruptive phases were earmarked in Magnet hill domain. These phases are distinguished on the basis both their mineralogical-petrological composition and their spatial distribution. In the Magnet hill domain occur the Paleogene sediments. These sediments were regarded formerly as the Carboniferous rocks of the Gemer Unit or the Choč nappe (the Ipoltica Group). The Quarternary sediments do not create more important accumulations and their thickness do not reach more than 3.5 metres.

Several fault structures, which have not been mapped before were identified by the structural research of the Tisovec Karst. The dominant fault structures trend NNE-SSW normal faults, which break the whole territory to asymetric grabens with maximum subsidence in block the Suché Doly valley and Goštanová hill domain.

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The Sparistá dolina Granitic Mylonites – the Products of the Alpine Deformation

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Abstract. Granitic rocks deformed into mylonites and ultramylonites during the Alpine collisional and/or shear zone deformation were studied in the Veporic unit, north of the village of Bacúch (Central Western Carpathians). The brittle-ductile to ductile deformation shows penetrative character throughout the granitoid body. S-C mylonite fabric due to an increase of strain and grain-size reduction is transformed into a single foliation fabric. New metamorphic mineral assemblage replaces the primary magmatic mineral composition. Allochemical processes documenting the mobility of some major and trace element accompanied these mineral changes. Mass-transfer, depending on progressive deformation, was facilitated only moderately by the percolation of external fluids. The composition changes reflect a loss of Fe, Mg, Mn, Ca, Sr, Ba and Ti, and/or a gain of K, Li and Rb. The phengite content up to 7.1 pfu of white mica in the assemblage muscovite-K feldspar-biotite-quartz reveals a peak pressure of ca. 12 kbar, reflecting conditions of the Cretaceous metamorphism in the Veporic basement. However, we infer for the first (progressive stage) of deformation an average pressure ca. 9 kbar, reflecting the creation of ultramylonites, while mylonites in the second period show only 6 kbar of average pressure. These pressure conditions, together with a temperature of 350 – 550°C, are common for orogenic belt metamorphism at convergent plate margins. The well-defined $^{40}Ar/^{39}Ar$ age spectra with 78 ±1.3 Ma (PA and TGA), indicates formations of these white micas during the second deformation period. This age reflects the cooling in the consequence extensional processes leading tectonic unroofing and/or exhumation of basement during the Late Cretaceous, which generally followed the peak time of burial and post-thickening thermal relaxation.

Key words: Western Carpathians, Veporic unit, granitic rocks, mylonites, chemical changes, dynamic metamorphism, $^{40}Ar/^{39}Ar$ dating, tectonic evolution.

Introduction

The polymetamorphic and polyorogenic character of the Veporicum crystalline basement had already been identified by Zoubek (1936). Later this interpretation was supported by Vrána (1966), Kamenický (1977) and Hovorka et al., (1987) among others. Zoubek´s conception (l.c.) of the Veporicum division into regional zones was overcome by a nappe-style classification proposed by Klinec (1966). The distinct sign of the whole Veporicum basement is its deformation. The intensity of the deformation varies from relatively non-deformed domains to ulramylonite zones, commonly indicating the existence of shear and thrust zones. The Alpine tectonic – metamorphic strain deformation of the Veporicum increases not only in the direction from northwest to southeast (Vrána, 1964), due to collision-subduction under the overthrusting Gemericum, but also from the margins to centers of thrusting and shear zones, as for example the Pohorelá lineament and Muráň fault (Hók and Hraško, 1990; Putiš et al., 1997). The Alpine deformation and the recrystallization of the Veporicum crystalline complexes is related to the Middle Cretaceous collision (Andrusov, 1968; Biely, 1989), or to the closing of the Meliata Ocean during the Late Jurassic to Early Cretaceous (Kozur, 1991; Plašienka, 1991).

During the metallogenetic research of the northern Veporicum – Kráľová Hoľa part, several occurrences of granitic rocks were thoroughly studied in the eastern part of the Nízke Tatry Mts. Among others, granitic rocks of the Sparista dolina type, in sense of Miko et al. (1982), were studied. The purpose of this article is to present the existence of the significantly deformed granitoid rocks – mylonites and ultramylonites, in northwestern part of the Veporicum. Based on mineral transformations, the chemical and metamorphic changes in the rocks are discussed. The high-pressure character of the deformation is identical with the degree of deformation of the granites in southeastern part of the Veporicum, in sense of Plašienka et al. (1999). The Late Cretaceous age of the tectonic-metamorphic processes, slightly postdating the movement of the superficial Mesozoic nappes in the Central Western Carpathians (CWC), is a reflection of the exhumation of the basement due to extension movement after collisional crustal thickening.

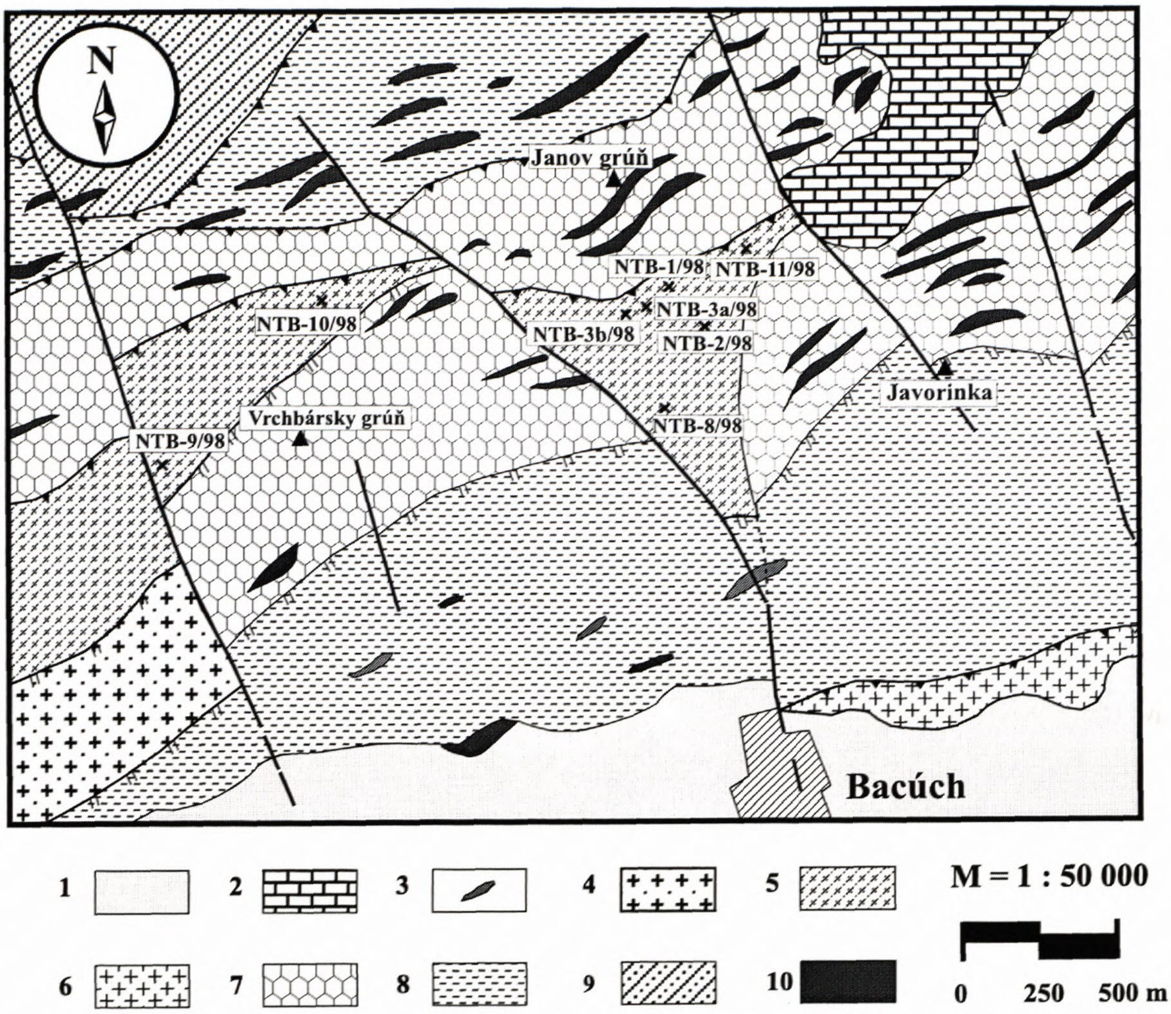

Fig. 1 The geological sketch of the studied area by Biely et al. (1992), slightly modified by authors, with localization of the sample sites. Explanation: 1 – Quaternary, 2 – Mesozoic, 3 – pegmatite and aplite, 4 – muscovite – biotite granodiorite and granite, 5 – SDGM, 6 – biotite granodiorite to tonalite, 7 – chlorite – sericite phyllites of Janov grúň, 8 – mica-schist gneisses – mica schists, 9 – migmatitized gneisses, 10 – metabasalts (green schists ± amphibolites).

Geology Settings

Northerly of the village Bacúch in the valley of Bacúch creek, Leňuša, Kršková, Zamrzlá and Sparistá valleys there are bodies of granitoid rocks in the form of imbricated plate slices. They are tectonically incorporated into the Hron complex in the sense of Klinec (1966) or in the formation of the Janov grúň in the sense of Miko (1981) (Fig. 1).

The first remarks about the sheared granites in the Krakľová zone of the Veporicum at the ridge part of the Nízke Tatry Mts. were done by Zoubek (1935). He considered this rocks as an abyssal analogue of the Permian intrusion, part of these hybrid igneous rocks he identified as „Muráň orthogneiss“. Klinec et al. (1971, 1973) and Klinec (1976) proved the tectonic position of the granitoid blocks overlapping the metamorphites of the Hron nappe, in the area Domárky – Kolesárová – Veľká Vápenica. The Permian age of the granitoids from the Bacúch area was unfixed by the authors (l.c.), based on the conformable fabric. During the late 70-ies and early 80-ies Miko (1981) was working in this area. Within the frame of the Hron complex, he distinguished weakly metamorphosed beds of the Janov grúň formation and tectonically imbricated granite slices. Based on the chemical composition of these granitic rocks the author l.c. assumed their affinity to plagiogranites of the early-orogene gabbro-plagiogranite formation. Miko et al. (1982) gave the first semiquantitative and qualitative geochemical characteristics of these deformed granitoids and for their clear distinction from other Veporicum granitic rock occurrences he named them „granitoids of Sparista dolina type“. During the following periods these rocks were subjected to mineralogical studies; different aspects of chemical composition and typology of accessory minerals were studied mainly by Hraško (1983) and Hraško & Miko (1990). The typology of the zircon expressly excluded the mantle origin of these granitoids.

The age of these problematic rocks has not been reliably resolved yet. The influence of Zoubek´s inferences lasted until the end of 70-ies and it was reflected in the cartographic plots of the general geological maps of CSSR at 1:200,000 and 1:500,000 scales. In these maps were marked these primary igneous rocks as Late Paleozoic – Permian intrusions (Maheľ et al., 1964). Klinec et al. (1971; 1973) classified the deformed granitoids of the

Bacúch area into the Early Paleozoic rock complexes. This opinion was also supported by the U-Pb dating of zircons from the Leňušska valley – 370 Ma (Cambel et al., 1977), which until now is the only relevant date from these deformed granitoids that suggests the time of their primary origin. Although the Early Paleozoic time classification was accepted in the published maps of the Nízke Tatry Mts. at a 1 : 50,000 scale (Biely et al., 1992) and Slovakia at 1 : 500,000 scale (Biely et al., 1996), the primary origin of these deformed igneous rocks is not yet adequately established, due to the intensive Alpine metamorphism. The tectonic-deformation metamorphism of the studied rocks is significant enough. The connection with the Alpine Orogeny is the traditional one, now accepted for a long time, and it was proven of the „granitoids of the Sparistá dolina type“ by Bagdasarjan et al. (1977) by means of the K-Ar method. The age determined was 104 – 97 Ma.

Methods

For the needs of complex research, we took 10 geochemical samples with weights of 10 – 15 kg, from which we have selected 8 samples for the detail geochemical study. The chemical composition of the samples was analyzed in the Geoanalytical Laboratory of GSSR in Spišská Nová Ves with use of AES ICP (Atomic Emission Spectrometry with Inductively Coupled Plasma) and XRF (X-Ray Fluorescence Spectrometry) methods. Quality control was verified on natural, international standards GM (granite) and BM (basalt) from ZGI Berlin. More analytical details are available, for example, in the Geochemical Atlas - rocks of SR (Marsina et al., 1999).

The mineral composition was analyzed in the laboratory of Electron Microanalysis of GSSR in Bratislava with JEOL – 733 Superprobe and KEVEX Delta, energy dispersion analyzes (EDS) under standard conditions 15kV and 12^{-10}A, with the use of natural and synthetic standards Taylor.

The $^{40}Ar/^{39}Ar$ isotopic dating was performed in the joined isotope laboratory of the Geological Survey of Austria (BGA) and the University of Vienna (BVFA ARSENAL). For this purpose we have selected three samples (NTB – 2, 3, and 11) from which mineral phases biotite and white micas – phengite were selected in the separation laboratories of the Department of Isotope Geology of GSSR, Bratislava. The mineral separation was done by mean of standard methods: through the use of a separating Wilfley table, electrostatic separator (made by the Department of Nuclear geology, GÚDŠ), Cook magnetic separator, heavy liquids, manual final cleaning under binocular microscope, and purification by methanol and re-distilled water in ultrasonic scrubber. The procedure for $^{40}Ar/^{39}Ar$ measurements in AVFT AESWNAL is described by Kráľ et al. (1995). The samples (charges of 8 – 10.5 mg) were sealed into siliceous capsules and together with internal standard WAP with the value of radiation parameter J = 0.003274 – 0.00495 (error ±0.4%) irradiated in the ASTRA reactor by a portion of accelerated neutrons about 10^{17} neutrons/cm$_{\llcorner}$. The samples were analyzed with the Mass Spectrometer MS VG 5400. Argon was released through a high-frequency generator by mean of a classical process, i.e., gradual temperature releasing (SBSH) in 6–10 temperature steps to 650–1250°C. The measured isotopic conditions were evaluated by a routine process developed by the company VG, with use of the recommended decay constants for the age calculation according to Steiger and Jäger (1977) and based on McDougall and Harrison (1988).

Petrographic Characteristics

„Granitoids of the Sparistá dolina“ have a pale-gray to greenish-gray color. They are fine- to medium-grained rocks, mostly with equigranular structure. Layers with homogeneous fine-grained fabric (with a maximum grain size up to 1 – 1.5 mm) are altered within relatively homogeneous layers, where quartz-feldspar porphyroclasts with a size of 1 – 2.5 mm are evenly distributed in the fine-grained groundmass. Layers with inhomogeneous mylonitic – porphyroclastic fabric were observed in the lower extent. Larger, unevenly distributed porphyroclasts, as much as 5 mm in size occur only locally. In this rock we can visually identify quartz, feldspars and biotite porphyroclasts. The matrix comprises fine-grained quartzofeldspatic layers and fine-grained muscovite – sericite interlayers, forming an anastomosing (dendroidal) network. The dominating feature of all the studied mylonite samples is their penetrative deformation that is characterized by a significant foliation, in most cases with typical twincleavage S-C mylonite fabric (Berthé et al., 1979; Lister and Snoke, 1984), (Fig. 2a). Locally, we can observe extension fractures (discordant to a C shear- plane and parallel with a first S deformation foliation), filled with chlorite, that were consequently in the second deformation stretched and sigma-like folded (Figs. 2c and 3h). The original fabric of the igneous rock was completely overprinted by the effects of the pressure – deformation and metamorphic recovery to dynamofluidal – plane-parallel, eye-shaped up to mylonitic fabric (Figs. 2a-c). The lineation of sericite can be occasionally observed on the foliation planes.

The effects of the cataclasis, ductile deformation and recrystallization of the original granitic mineral assemblage were also observed in a micro-scale (Figs. 3a-f). The deformation is connected with a reduction of the grain-size (quartz and K-feldspar), locally also with the creation of sub-grains and shifts on the grain rims. Quartz commonly forms fine-grained ribbons with grain-size under 0.01 mm (Fig. 3g). The mylonite deformation is also documented by the formation of mica fish (Fig. 3c). Along with the deformation increase, the angle of two foliation planes (S-C) is changed as well, from values of 45 – 30° to 15 – 10°, and in the ultramylonite it disappears totally (Figs. 3a-e) and we observe only one dominant foliation (cleavage). Of the primary minerals there are present: quartz, plagioclase, K-feldspar, biotite, ± muscovite and the accessories: zircon, apatite and monazite. Of the newly developed minerals, the dominant ones are plagioclase with an albite composition and fine-grained

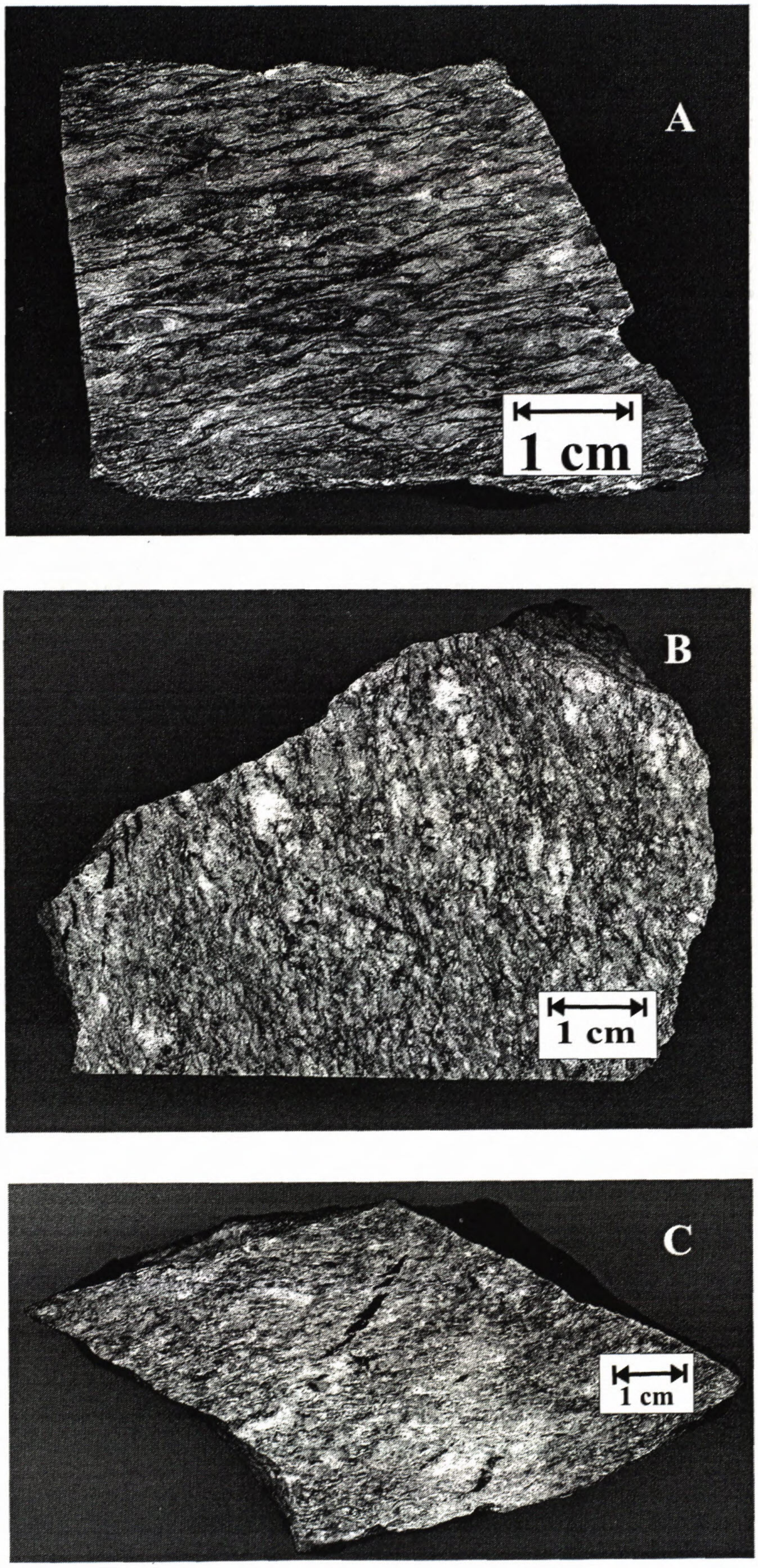

Fig. 2A – eyed fabric of the mylonite, the sample NTB – 3a, S-C fabric indicates crossing shear bands,
2B – indistinctive S-C fabric of mylonite, sample NTB-3b,
2C – extension veins filled with chlorite in ultramylonite of the sample NTB-1.

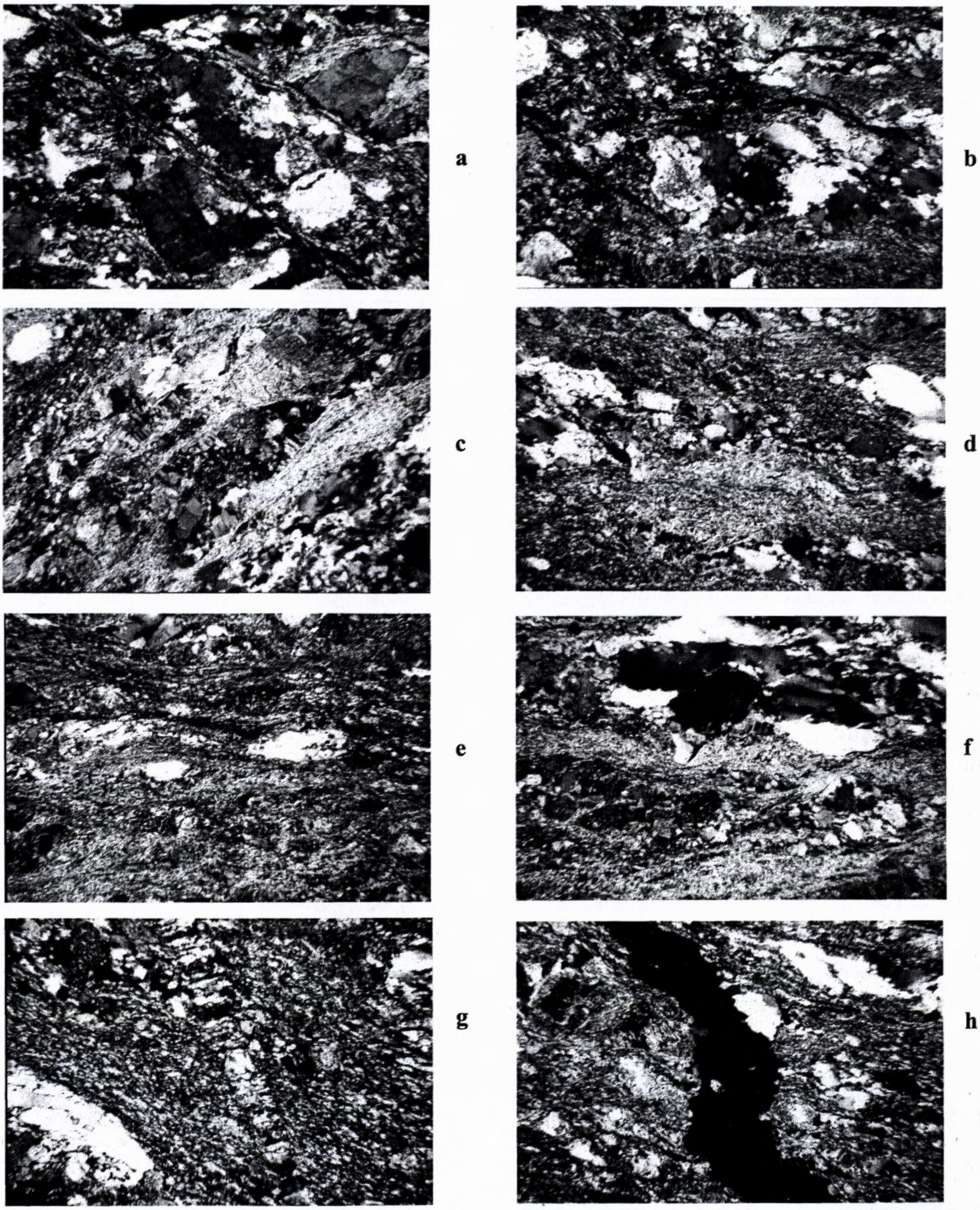

Fig. 3 Microphoto documenting growth of the progressive deformation from porphyroclastic structure – (3a) to ultramylonite structure (3e) in SDGM, 3g – extension vein filled with quartz, 3h – extension vein filled with chlorite. Crossed nicols. Width of the photo = 1.5 mm.

muscovite mica – sericite (phengite). Within the framework of the secondary minerals we observed also epidote, clinozoisite, chlorite (Fig. 3h), carbonates, titanite, ± garnet, rutile, ± bauerite and sagenite. Signs of the original magmatic – hypidiomorphic fabric is observed only in the greater „augen" – porphyroclasts. The dominant fabric of this pressure – deformed rock is porphyroclastic, and/or mylonite fabric. The modal composition is given on table 1.

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Based on the petrographic analysis, we identify this rock as *porphyroclastic (eye-shaped) granitic mylonites to ultramylonites*, or simply as **Sparistá dolina granitic mylonites (SDGM).** Our evaluation of these rocks is partly in agreement with identification according to Miko et al. (1982); however, we prefer the designation „mylonite“ over granitoid, because this rock is more tectonically metamorphosed rock than is a magmatic rock in current form.

Tab. 1 The modal composition of the studied SDGM mylonites.

Mineral	NTB-1	NTB-2	NTB-3a	NTB-3b	NTB-8	NTB-9	NTB-10	NTB-11
Quartz	33,1	32,3	35,7	35,8	32,7	35,4	35,2	33,8
Plagioclase/Albite	24,9	25,4	23,5	25,7	24,1	26,2	24,4	22,7
K-feldspar	5,4	6,3	5,6	5,0	5,8	4,7	4,6	4,2
Biotite	3,6	2,3	1,5	1,2	3,1	1,3	2,2	3,8
Muscovite/Sericite	21,6	22,2	25,4	24,0	22,1	24,5	21,3	22,1
Epidote - Zoisite	4,3	5,2	3,9	3,6	5,5	3,2	5,6	5,2
Chlorite	2,4	1,6	0,8	1,1	1,8	0,9	1,7	1,9
Calcite	2,9	3,1	2,1	2,4	3,0	2,5	3,4	4,4
Accessories	1,8	1,6	1,5	1,2	1,9	1,3	1,6	1,9

Geochemical characteristics

The composition of the mylonites is shown on table 2. The table revealed that the rocks do not have economically interesting mineralization. The analytical values of the samples are in good agreement with the average content of the main and trace elements of the rocks of the upper part of the Earth´s crust and/or the granitic rocks according to Wedepohl (1969), Taylor and McLennan (1985). They are comparable with another nonmetallic (barren) granitic rocks of the Western Carpathians – more data can be found in the Geochemical Atlas - rocks of SR (Kohút in Marsina et al., 1999). Because these rocks are categorized into a group of tectonically deformed rocks, we compared the chemistry of the samples with analogues of the deformed rocks – orthogneisses from the Western Carpathians. The comparison was done with data from the rock catalogue (Marsina et al., l.c.), and also with newer data (Putiš et al., 1997). The particularity of the samples was recognized as a consequence of the high-grade (tectonic + metamorphic) strain deformation.

The rocks (SDGM) represent tectonically deformed analogues of magmatic – granitic rocks. Keeping in the mind the petrographic character of the rocks and the general rules of the changes that are observed during the deformation – metamorphic processes of similar rocks (Vrána, 1964; Kerrich et al., 1977, 1980; Marquer, 1989; and citations therein), and with respect to the chemical composition we do not infer a mantle origin of the magmatic precursor of these rocks. Most probably, the SDGM represent analogues of calk-alkaline magmatic rocks that originated from the hybrid crust/mantle types on the continental margins during the subduction processes (VAG, CAG). The SDGM chemically correspond to greywackes or recycled andesite rocks (basaltic andesite), from which biotite and amphibolite – biotite granodiorite to tonalite were formed. From the granitic rocks occurring in the Veporicum, tonalite of the Sihla type match this characteristic best.

However, for the possibility of tracing the changes in distribution of the individual elements during the deformation processes, we compared our values with an average composition of the relatively unstrained Sihla tonalite that, as we assume, was the precursor of these mylonites (Tab. 3). The average composition was obtained on the basis of published and archived sources (Broska and Petrík, 1993; Marsina et al., 1999). The graphic visualization of the distribution of the main and the most important trace elements of the SDGM compared to the average Sihla tonalite composition is shown in figure 4.

As we indicated from table 3 and figure 4, during the dynamic metamorphism that effected the hypothetical precursor of the SDGM – the Sihla tonalites, allochemical changes due to cataclasis and pressure solution (Kerrich et al., 1977), leaching and recrystallization of a new mineral assemblage have been occurred. A significant depletion (as much as 50%) occurred in the case of iron, strontium, barium and titanium. During the mylonitization, the original tonalite was depleted of as much as 25 % of its magnesium, calcium and zirconium content. In contrast, a 20 - 50 % content increases were registered in the case of potassium, rubidium and partly also of lithium. Also slightly increased are the contents of sodium and silica. A balanced distribution is the case of aluminum, which is in agreement with the generally accepted ideas about its limited mobility. All these chemical changes occurred due to mineral changes during recrystallization. The changes of the content of FeO^T, MgO, CaO and Ba reflects a destruction and the following recrystallization of biotite mica and plagioclase, which is also directly related to the main changes in content of alkalies (K_2O, Na_2O), Li and Rb, due to crystallization of sericite and albite. Other elements (Tab. 3), Ta, Co, Cr, U and V have balanced distributions, Nb, Y, Hf, Ni, Zn, Th and REE have decreasing contents. In contrast, Be, Pb and Cu have increasing contents. These rocks have record of REE identical with that of rocks melted in the active continental arc with a combined crust type – recycled continental with contribution of primitive mantle magma, melted in the lithospheric wedge. Assimilation-fractional processes most probably cause the intermediate character of the tonalite-granodiorite rocks. In the magmatic rocks the REE are preferably bonded with the accessory minerals (monazite, apatite, allanite, garnet, etc.). In the process of dynamic metamorphism these mineral forms were also attacked and broken-up (monazite, allanite), and instead of them mainly REE-rich epidote crystallized, which is in agreement with observations by Petrík et al. (1995) and Broska and Siman (1998). Titanites crystallizing at the expense of biotite micas and plagioclases contributed slightly in increasing of the HREE content (Fig. 5).

Tab. 2 The chemical composition of the studied rocks

Sample	NTB-1	NTB-2	NTB-3a	NTB-3b	NTB-8	NTB-9	NTB-10	NTB-11
SiO_2	68,08	68,33	68,46	67,21	68,22	68,59	69,69	69,09
TiO_2	0,45	0,43	0,32	0,37	0,41	0,35	0,39	0,33
Al_2O_3	15,15	14,47	14,43	16,18	15,08	14,98	13,97	15,07
Fe_2O_3	0,85	0,80	0,78	0,74	0,77	0,95	0,73	0,72
FeO	1,88	1,81	1,41	1,59	1,73	1,34	1,73	1,41
MnO	0,05	0,04	0,06	0,05	0,05	0,04	0,05	0,05
MgO	1,30	1,65	1,26	1,15	1,46	1,00	1,18	1,10
CaO	2,23	2,15	2,68	2,58	2,35	3,03	2,47	2,57
Na_2O	3,67	3,64	4,16	4,73	4,02	4,26	4,89	4,61
K_2O	3,71	3,91	3,30	3,21	3,74	3,31	2,43	2,96
P_2O_5	0,13	0,13	0,14	0,13	0,15	0,13	0,12	0,13
H_2O+	1,75	2,05	2,29	1,31	1,81	1,81	1,96	1,45
H_2O-	0,38	0,39	0,41	0,43	0,20	0,24	0,36	0,37
Total	99,63	99,80	99,70	99,68	99,99	100,03	99,97	99,86
B	7	9	9	11	10	7	9	10
Ba	460	229	236	402	385	315	301	381
Rb	116	136	125	78	124	98	90	96
Sr	249	177	197	457	268	338	354	337
Zr	157	150	123	120	141	154	143	125
Nb	8	7	7	5,3	7	6	0,7	6
Y	10	11	9	7	10	10	10	8
Hf	4	3	3	1	4	4	4	3
Ta	1	1	1	2	2	1	1	1
Be	2,6	2,5	3,7	2,4	2,8	2,6	2,6	2,6
Li	38	35	27	26	34	26	21	29
Co	6	5	4	6	5	5	5	5
Cr	30	27	15	18	25	18	20	19
Ni	10	10	1	8	8	2	3	1
Pb	5	7	5	21	6	20	9	18
Cu	13	8	12	8	11	108	28	9
V	49	43	34	34	44	46	46	42
Zn	45	41	23	58	43	52	52	54
Th	6	6	4	2	6	5	6	5
U	2	2	2	1	2	2	2	2
La	26,00	26,00	24,00	19,00	25,00	23,00	26,00	21,00
Ce	36,00	37,00	35,00	28,00	36,00	35,00	38,00	29,00
Nd	18,00	18,00	15,00	13,00	17,00	14,00	15,00	13,00
Sm	4,50	5,00	4,00	3,00	4,50	3,80	3,00	3,50
Eu	0,90	0,90	0,80	0,60	0,90	1,00	0,65	0,70
Gd	2,60	2,70	2,50	2,10	2,40	2,60	2,60	2,00
Tb	0,40	0,38	0,37	0,30	0,40	0,35	0,40	0,33
Tm	0,24	0,24	0,23	0,19	0,22	0,22	0,23	0,20
Yb	1,10	1,10	1,00	0,80	1,10	0,90	1,20	0,80
Lu	0,15	0,15	0,14	0,12	0,15	0,14	0,14	0,13

Tab. 3 Comparison of the average composition of the precursor – Sihla tonalites (Sihla T) vs. composition of the SDGM, (SDGM – average composition, Min – minimum values of the composition, Max – maximum values of the composition, St.Dev. – standard deviation).

	Sihla T	SDGM	Min	Max	St.Dev.
SiO_2	64,95	68,46	67,21	69,69	0,73
TiO_2	0,81	0,38	0,32	0,45	0,05
Al_2O_3	15,72	14,92	13,97	16,18	0,66
Fe_2O_3	2,05	0,79	0,72	0,95	0,08
FeO	2,53	1,61	1,34	1,88	0,21
MnO	0,07	0,05	0,04	0,06	0,01
MgO	1,76	1,26	1	1,65	0,21
CaO	3,25	2,51	2,15	3,03	0,28
Na_2O	4,01	4,25	3,64	4,89	0,47
K_2O	2,52	3,32	2,43	3,91	0,48
P_2O_5	0,36	0,13	0,12	0,15	0,01
H_2O+	1,45	1,80	1,31	2,29	0,31
H_2O-	0,22	0,35	0,2	0,43	0,08
Total	99,7	99,83	99,63	100,03	0,15
B	4	9,00	7	11	1,41
Ba	640	338,63	229	460	82,13
Rb	71	107,88	78	136	20,22
Sr	541	297,13	177	457	92,33
Zr	174	139,13	120	157	14,65
Nb	11	5,88	0,7	8	2,25
Y	16	9,38	7	11	1,30
Hf	7	3,25	1	4	1,04
Ta	1	1,25	1	2	0,46
Be	2	2,73	2,4	3,7	0,41
Li	26	29,50	21	38	5,68
Co	6	5,13	4	6	0,64
Cr	20	21,50	15	30	5,21
Ni	10	5,38	1	10	4,00
Pb	8	11,38	5	21	7,03
Cu	5	24,63	8	108	34,31
V	48	42,25	34	49	5,52
Zn	70	46,00	23	58	10,98
Th	8	5,00	2	6	1,41
U	2	1,88	1	2	0,35
La	46	23,75	19	26	2,60
Ce	91	34,25	28	38	3,69
Nd	39	15,38	13	18	2,07
Sm	6,5	3,91	3	5	0,73
Eu	1,6	0,81	0,6	1	0,14
Gd	4,9	2,44	2	2,7	0,26
Tb	0,6	0,37	0,3	0,4	0,04
Yb	1,5	1,00	0,8	1,2	0,15
Tm	0,23	0,22	0,19	0,24	0,02
Lu	0,21	0,14	0,12	0,15	0,01

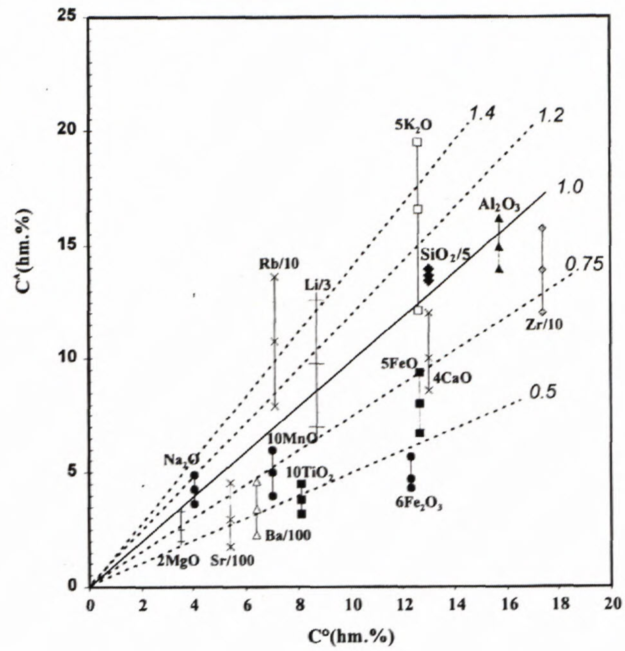

Fig. 4 Relationship between chemical composition of the SDGM (C^A) and composition of the precursor – the Sihla type tonalite (C^o) from Tab. 3 (the composition of the main elements is in wt.%, the trace elements in ppm) in the isocone diagram according to Grant (1986). The composition of the SDGM mylonites is shown by the interval between minimal and maximal content, as well as by their average composition.

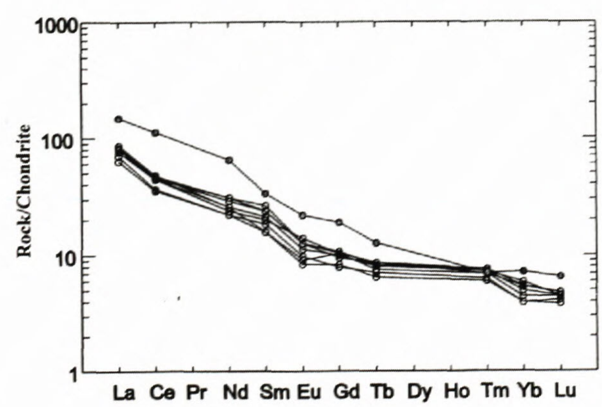

Fig. 5 Diagram of chondrite normalized contents of REE in the studied rocks. Symbols: empty circle - SDGM, filled circle – average composition of the Sihla type tonalite, table 3.

The tectonic-deformation processes, and with them connected allochemical changes, lead to recrystallization of the new mineral assemblage under changed pressure-thermal conditions. During these processes circulation of fluids occurred, not only within the framework of the granitic rocks itself, but also in surrounding metamorphic complexes. The yield of individual components within the granitic rocks and their diffusive transportation in the form of metamorphic fluids had to lead, on the other hand, toward their concentration at another place. Preliminary, within the Sparistá dolina granitic mylonites we have registered only the extension veins filled with silica – albite – chlorite mineral assemblage, which documents the migration of lithophile (Si, Na), as well as siderophile (Fe, Mg) elements. Since this deformation occurred in the extreme depths (>20 km), we assume only limited influence by external fluids, which would be facilitated the shear deformation of the former granitic rocks and its strain for SDGM.

Metamorphic Conditions

Whereas the petrographic studies clearly proved the penetration character of the deformation changes, as well as the metamorphic character of these rocks in the present state, we tried to identify the degree of metamorphism and the overall metamorphic conditions of the origin of these mylonitic rocks. We determined the thermal-pressure conditions of the deformation on the basis of the structural-deformational criteria and mineralogical-petrological conditions. For these purposes it was necessary to know the composition of the mineral phases contributing to the fabric of the mylonites, because during the crystallization they recorded pressure-thermal conditions of closing the crystallization lattice. The chemical composition of the analyzed mineral phases is shown on tables 4 – 6.

Tab. 4 The representative chemical composition of the feldspars (a – plagioclase, o – K-feldspar) from the SDGM. The recalculation is based on 8 oxygens.

	NTB-3/1a	NTB-3/3a	NTB-2/3a	NTB-2/5a	NTB-10/3a	NTB-1/1a	NTB-1/4a	NTB-3/2o	NTB-2/1o	NTB-1/3o	NTB-1/4o
Na_2O	11,35	11,18	11,45	11,08	11,36	11,36	11,70	0,00	0,00	0,21	0,10
Al_2O_3	19,64	20,14	19,99	19,98	20,05	19,89	19,57	18,17	18,38	18,35	18,62
SiO_2	67,92	68,42	67,76	67,68	67,41	68,79	68,54	64,83	65,05	64,55	65,06
K_2O	0,16	0,16	0,11	0,11	0,18	0,00	0,00	16,48	16,77	16,64	16,46
CaO	0,08	0,14	0,18	0,17	0,41	0,16	0,13	0,00	0,00	0,00	0,00
FeO	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,19	0,27	0,00
Total	99,15	100,04	99,49	99,02	99,41	100,20	99,94	99,48	100,39	100,02	100,24
Na	0,97	0,95	0,97	0,95	0,97	0,96	0,99	0,00	0,00	0,02	0,01
Al	1,02	1,04	1,03	1,04	1,04	1,02	1,01	0,99	1,00	1,00	1,01
Si	2,99	2,98	2,97	2,98	2,97	2,99	2,99	3,01	3,00	2,99	3,00
K	0,01	0,01	0,01	0,01	0,01	0,00	0,00	0,98	0,99	0,98	0,97
Ca	0,01	0,01	0,01	0,01	0,02	0,01	0,01	0,00	0,00	0,00	0,00
Fe	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,01	0,01	0,00
Sum	5,00	4,99	4,99	4,99	5,01	4,98	5,00	4,98	5,00	5,00	4,99

As can be seen in table 4, most of the plagioclases have decreasing CaO content, which is a consequence of the decalcification connected with the crystallization of the carbonate within the rock during the deformation processes. Due to their chemical composition, they are classified as albite to acid oligoclase (An 7 – 12). Phengite as a middle member of the muscovite-celadonite mineral group of white micas, reflects very well through its compositions the changes of the P-T conditions during its origin. The mineral assemblage phengite – biotite – K-feldspar – quartz has became the basis for the phengite barometer in sense of Velde (1965, 1967), later improved by Massonne and Schreyer (1987). Just such a mineral assemblage is typical for the mylonitized granitic rocks in the Veporicum. We have observed it in the Sparistá dolina granitic mylonites as well as. The barometer is based on the Si parameter value within a structural formula. The Si parameter values in the samples are within the range of 6.39 –7.10 pfu, which reveals a pressure of 400 - 1200 MPa (4 – 12 kbar) (Tab. 5 and Fig. 6). It is interesting that in the more deformed ultramylonites we have identified phengites with greater Si parameter value (6.65 – 7.10), which corresponds to a pressure of 800 - 1200 MPa and in mylonites only Si = 6.39 – 6.61, which indicates pressures of 400 – 750 MPa. Similarly, Hraško (1998) observed the wide range of phengite Si parameter values in the central and southeastern parts of the Vepor massif. Preliminary we assume that the deformation of the original granitic rocks (granodiorite to tonalite) happened in two phases. The maximum pressure was recorded by the ultramylonites in the first phase during the deepest burial of the complex within the compressional phase of the Alpine Orogeny. During this progressive phase, the pressure may have reach 12 kbar, which is in accordance with Plašienka et al. (1999). Based on the phengite component, we assume the average pressure of 9 kbar (Fig. 6) for this first stage. The mylonites were created during the retrograde phase of the orogeny, after the thermal relaxation due to extension. The identified average pressure in the mylonites – about 6 kbar is more likely a reflection of the second phase, in the consequence rapid unroofing connected with pressure changes (Fig. 6).

The composition of biotite undergoes the greatest changes during the dynamo-metamorphic processes, because this phyllosilicate is composed of several unstable elements (Ti, Fe, Mg, Ca, K), and undertakes various changes (chloritization, baueritization etc.). The secondary minerals as chlorite, ilmenite, rutile, titanite, magnetite,

Tab. 5 The representative chemical composition of the white micas from the SDGM. The recalculation is based on 22 oxygens.

	NTB-3/1	NTB-3/2	NTB-3/4	NTB-3/8	NTB-2/3	NTB-2/6	NTB-10/2	NTB-1/1	NTB-1/4	NTB-1/6	NTB-1/9
SiO_2	48,34	46,43	47,37	48,06	48,48	47,84	52,50	53,40	50,63	50,41	49,57
TiO_2	0,00	0,00	0,57	0,00	0,38	0,56	0,00	0,18	0,15	0,26	0,21
Al_2O_3	29,65	29,94	29,15	31,17	25,88	26,14	30,38	25,27	28,15	27,47	28,04
FeO	3,89	2,74	3,98	2,50	5,92	5,22	1,48	2,70	2,58	2,67	2,58
MnO	0,00	0,00	0,00	0,00	0,10	0,04	0,25	0,42	0,36	0,42	0,00
MgO	2,93	2,77	2,79	2,10	2,50	2,67	1,19	2,65	2,28	2,50	2,43
CaO	0,00	0,00	0,00	0,00	0,16	0,08	0,00	0,00	0,00	0,00	0,00
Na_2O	0,17	0,21	0,17	0,26	0,12	0,12	0,00	0,00	0,00	0,11	0,00
K_2O	11,37	11,72	11,20	11,57	11,14	11,04	9,19	10,44	10,57	10,83	10,64
Cr_2O_3	0,00	0,00	0,00	0,00	0,00	0,00	0,34	0,40	0,22	0,22	0,36
Total	96,35	93,81	95,23	95,66	94,68	93,71	95,33	95,46	94,94	94,89	93,83
Si	6,48	6,39	6,51	6,44	6,66	6,65	6,87	7,10	6,78	6,79	6,73
Ti	0,00	0,00	0,04	0,00	0,04	0,05	0,00	0,02	0,02	0,03	0,02
Al	4,68	4,86	4,66	4,93	4,24	4,22	4,68	3,96	4,45	4,36	4,49
Fe	0,44	0,31	0,43	0,28	0,69	0,64	0,16	0,30	0,29	0,30	0,29
Mn	0,00	0,00	0,00	0,00	0,01	0,01	0,03	0,05	0,04	0,05	0,00
Mg	0,59	0,57	0,56	0,42	0,51	0,55	0,23	0,53	0,46	0,50	0,49
Ca	0,00	0,00	0,00	0,00	0,03	0,01	0,00	0,00	0,00	0,00	0,00
Na	0,04	0,06	0,04	0,07	0,03	0,03	0,00	0,00	0,00	0,03	0,00
K	1,94	2,06	2,00	1,98	1,99	2,01	1,53	1,77	1,81	1,86	1,84
Cr	0,00	0,00	0,00	0,00	0,00	0,00	0,04	0,04	0,02	0,02	0,04
Sum	14,17	14,25	14,24	14,12	14,20	14,17	13,54	13,77	13,87	13,94	13,90

Tab. 6 The representative chemical composition of the biotite micas of the studied rocks. The recalculation is based on 22 oxygens.

	NTB-3/2	NTB-3/3	NTB-3/7	NTB-2/3	NTB-2/6	NTB-8/1	NTB-10/2	NTB-1/3	NTB-1/4	NTB-1/5
SiO_2	34,76	34,47	35,31	35,30	35,17	35,24	35,93	36,12	34,31	35,75
TiO_2	2,68	3,31	2,49	2,04	2,30	2,66	2,96	2,61	3,45	2,08
Al_2O_3	17,41	15,83	14,97	15,49	14,47	16,26	15,70	15,88	15,40	16,76
FeO	22,80	23,34	24,30	24,74	23,87	21,69	20,80	20,45	23,34	20,87
MnO	0,00	0,00	0,00	0,00	0,00	0,57	0,94	0,80	0,79	0,78
MgO	8,38	8,02	8,49	9,07	8,56	8,44	8,71	8,72	7,39	8,36
CaO	0,00	0,00	0,19	0,00	0,00	0,15	0,27	0,51	0,21	0,00
K_2O	8,98	9,83	9,03	8,70	9,78	9,92	9,79	9,85	9,62	10,08
Cl	0,00	0,00	0,00	0,05	0,00	0,00	0,00	0,00	0,00	0,00
Cr_2O_3	0,00	0,00	0,13	0,10	0,00	0,47	0,30	0,34	0,41	0,41
Total	95,01	94,80	94,91	95,49	94,15	95,40	95,40	95,28	94,92	95,09
Si	5,40	5,43	5,55	5,51	5,59	5,48	5,56	5,59	5,43	5,55
Ti	0,31	0,39	0,29	0,24	0,27	0,31	0,34	0,30	0,41	0,24
Al	3,19	2,94	2,77	2,85	2,71	2,98	2,86	2,89	2,87	3,07
Fe	2,96	3,08	3,20	3,23	3,17	2,82	2,69	2,64	3,09	2,71
Mn	0,00	0,00	0,00	0	0,00	0,07	0,12	0,11	0,11	0,10
Mg	1,94	1,88	1,99	2,11	2,03	1,96	2,01	2,01	1,74	1,93
Ca	0,00	0,00	0,03	0,00	0,00	0,02	0,04	0,09	0,04	0,00
K	1,78	1,97	1,81	1,73	1,98	1,97	1,93	1,94	1,94	1,99
Cl	0,00	0,00	0,00	0,01	0,00	0,00	0,00	0,00	0,00	0,00
Cr	0,00	0,00	0,02	0,01	0,00	0,06	0,04	0,04	0,05	0,05
Sum	15,58	15,69	15,66	15,69	15,75	15,67	15,59	15,61	15,68	15,64

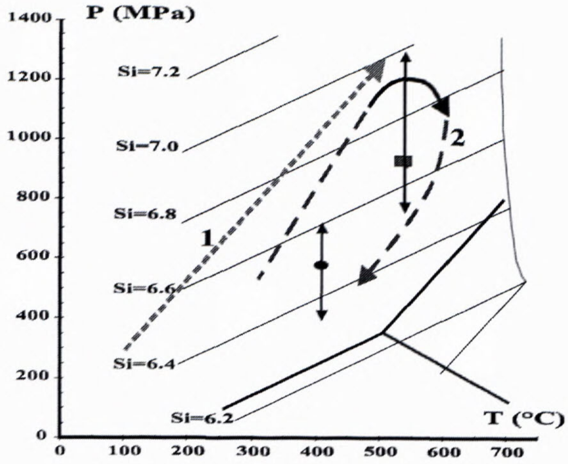

Fig. 6 P – T diagram based on the Si content in the muscovite mica – phengite (Massone and Schreyer, 1987) SDGM. The filled circle represents the average composition of the white micas (Si – component) of mylonites, filled box represents average composition of the white micas in ultramylonites. 1 – the path of the Alpine metamorphism according to Mazzoli et al., 1992, 2 – P-T path of the Alpine metamorphism according to Plašienka et al., 1999.

epidote, eventually also muscovite (bauerite) and phengite can be created. All this is reflected in the composition of the biotite (Table 6). Although we tried to analyze the non-altered phases, the migration of the individual elements within the studied sample is evident.

We also tried to make the temperature estimation on the basis of the mineral changes. The maximum tempeperature ?500°C indicates the origin of myrmekitic textures in K-feldspar (Bell and Johnson, 1989). In contrast to this, the minimum temperature of ?340°C is indicated by the change of biotite to chlorite (Eggleton and Banfield, 1985) in the granitic rocks. This temperature is in accordance with the minimum blocking temperature of the phengitic muscovite (350°C) in the K-Ar or Ar-Ar isotopic system (Purdy and Jäger, 1976), when the mineral lattice is finally closed, the diffusion of argon is stopped and the „isotopic clock“ is set off. The high-pressure conditions in the Alpine deformation is also well documented by the formation of clinozoisite and epidote at the expense of plagioclase at pressures of 8 – 12 kbar (Singh and Johannes, 1996). The deducted thermal (350 – 550 °C) and pressure (4 – 10 kbar) parameters are typical for the continental orogenic zones, as the Alpine Orogeny in the Western Carpathians certainly was.

Geochronology

In order to have an idea about the age of the host rock environment and of the migration of the fluids within the SDGM, as well as about possible relationship with potential ore mineralization, we decided to date these processes by the $^{40}Ar/^{39}Ar$ isotopic method. So far we obtained only one reliable age value from our samples, namely from the sericite sample NTB-3. The results of the measurements are shown on table 7 and their plot is shown in figure 7.

As is possible to state from the table 7 and figure 7, the sericite sample from the mylonite of the Sparistá dolina type – NTB - 3 has an outstandingly smooth spectrum of apparent ages of 76 - 79 Ma. It is equivalent of 96 % of the volume of the total amount of the degassed ^{39}Ar. Only in the first step does it show an apparent age slightly younger that the value of 67 Ma, which is common for similar samples. There were no relict cores of older muscovites in the sample, and it also does not show any signs of disturbances of the Ar/Ar isotope system in the consequent younger periods. Noteworthy is, that none high-pressure phengites were identified in this mylonite, only forms with Si = 6.39 – 6.61 pfu, corresponding to a pressure of about 6 kbar. As can be seen from the data, the sample attracts attention by its perfect match between the „total gas age“ calculated from the total amount of degassed ^{39}Ar and the weighted arithmetic mean of ages from the concordant steps of $^{40}Ar/^{39}Ar$ spectrum (plateau age). Thus, the interpreted age 78 ±1.3 Ma certainly represents the age of the creation and/or closing of the muscovite – sericite lattice during cooling under blocking temperature (Tc = 400 ±25 °C, respectively 350 °C), Purdy and Jäger (1976). More about blocking temperatures can be found in the work by Kohút et al. (1998).

Tab. 7 $^{40}Ar/^{39}Ar$ analytical data from the sericite sample NTB-3a.

Step	T(°C)	$\%^{39}Ar$	40* (mV)	$\%^{40}Ar$	39/37	$\%^{36}Ca$	$^{40}Ar^{*}/^{39}Ar$ (%)	Apparent age (Ma)
1	650	4.3	25.63	90.9	2	4.07	11.05 ±2.1	67.0 ±1.4
2	720	5.3	37.09	99.0	2	31.24	12.97 ±0.7	78.4 ±0.6
3	790	13.2	91.94	96.1	8	2.43	12.86 ±1.0	77.7 ±0.8
4	850	45.6	322.75	97.8	5	6.84	13.04 ±1.2	78.8 ±1.0
5	1000	18.5	126.16	95.5	3	4.99	12.54 ±0.6	75.9 ±0.4
6	1250	13.2	93.37	92.9	5	2.10	13.06 ±2.0	78.9 ±1.6
J = 0.003274 ±0.4%							**Total gas age:**	**77.6 ±1.4**
						96%	**Plateau age**	**78.1 ±1.3**

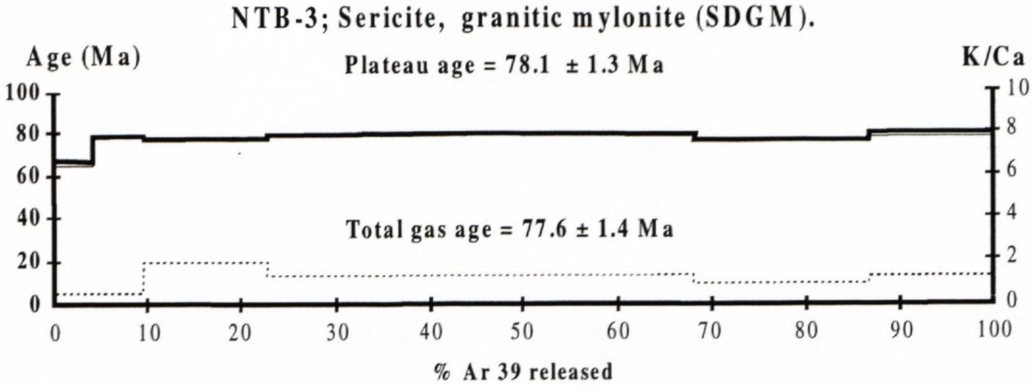

Fig. 7 The diagram of $^{40}Ar/^{39}Ar$ apparent age spectrums of sericite from the sample NTB-3a. The lower spectrum represents the changes of ratio K/Ca in the individual thermal steps.

Discussion and Conclusion

The complex research of the SDGM mylonites did not confirm potential ore mineralization of this rock type; however, at the recent level of knowledge we cannot reject the existence of the hypothetical occurrences of the economic mineral resources.

Based on the petrologic analysis we can confirm that the deformed granitic rocks north of the Bacúch site represents dynamo-metamorphosed, originally magmatic rocks, which is in agreement with previous findings (Miko, 1981; Miko et al., 1982). However, on the base of existing data we cannot confirm the affinity of these original igneous rocks to an early orogenic gabbro – plagiogranite formation in sense of these authors. We assume their origin from the combined crustal type within the active volcanic arch of „Andean type", where they represented medium-potassium, calk-alkaline series. The Middle Paleozoic – Late Devonian age (370 Ma) of these original magmatitic rocks were proven by the U-Pb dating of zircons (Cambel et al., 1977).

The tectonic-deformation processes during the Alpine Orogeny when shearing and recrystallization of these rocks took place, were connected with allochemical changes. The original mineral assemblage: plagioclase, quartz, biotite, K-feldspar, ± muscovite + accessories was replaced by a new assemblage: quartz, plagioclase (albite), sericite (phengite), K-feldspar, carbonate (calcite), minerals of epidote – zoizite group, ± biotite, titanite, ± garnet, chlorite, muscovite (bauerite), rutile, ilmenite, magnetite and accessories.

Based on the study of the element distribution, the mobility of the majority of the elements during the deformation process was confirmed, either in a form of gain or loss into the composition of these mylonites. This is in accordance with observations by Miko et al. (1982) from this area, and by Putiš et al. (1997) from the central part of the Veporicum part of the Slovak Rudohorie Mts., or by Kolaříková et al. (1985; 1994) from the eastern part of the Czech Massif and by Marquer et al. (1985; 1994) from the Alps.

The identified thermal-pressure conditions (350–550 °C and 4 – 10 kbar), which are responsible for the total strain and recrystallization of originally granitic rocks, as well as the character of this dynamic and/or dynamic-thermal metamorphism is typical for the orogenic zones along convergent plate margins (Spear, 1995). It is important to emphasize that the metamorphism (deformation) in these rocks – SDGM has a penetrative character and that it is not bounded only to narrow, several-meter-thick mylonite shear zones. The maximum thickness of the „body" in the Krškova and Bacúšska dolina Valleys reaches 200 m, with the foliation inclined 30 – 50° southwesternward; however, the maximum thickness can be as much as 500 m. Although within the body we observed zones with relatively lower degree of deformation assigned as porphyroclastic eye-shaped granitic mylonites with significant twin-cleavage S-C fabric, in many places there are also zones of fine-grained ultramylonites, where shear bands were gradually transformed into a single foliation fabric. The degree of metamorphism reached the balance level of biotite and garnet zone and/or boundary between greenschist facies and epidote amphibolite facies. A brittle-ductile character of porphyroclastic mylonites and ductile character of ultramylonites also refer to it. Interesting is the fact that deformation-metamorphic character of the SDGM is identical with the deformation of the Veporicum basement in the contact zone of the Veporicum and Gemericum (Plašienka et al., 1999).

The assumed metamorphic condition of the studied area (ca. 400 °C and 6 kbar) most probably represents a retrograde phase of the Alpine type of metamorphism after the climax of the progressive burial of the Veporicum block of the crystalline complex with its envelope under the overthrusting Gemericum from the southeast during the Jurassic – Cretaceous subduction – collision processes. However, there remains the unsolved problem: how and when the SDGM experienced by significant Alpine dynamo-thermal deformation and metamorphism, got into the Variscan low metamorphosed rock complex (Janov grúň formation) (Miko, 1981; Sassi and Vozárová, 1992). This requires detailed, mainly structural, research.

The Alpine metamorphism of the Veporicum is a problem that has long been discussed in the Western Carpathian literature, since the times of Zoubek (1936), who first identified this problem. Vrána (1966; 1980) later discussed this problem. He emphasized that the degree of the Alpine metamorphism of the granitic rocks is directly dependent upon the degree of tectonic deformation and intensity of the metamorphism. The author determined 5 types of D-R granitoids (deformed – recrystallized), from massive metagranites to blastomylonites. Vozárová (1990), Méreš and Hovorka (1991) and Hovorka and Méreš (1997) estimated the temperature and pressure of the Alpine metamorphism of the southwestern part of the Veporicum to 550 °C and 5–8 kbar. Mazzoli et al. (1992) deduced conspicuous pressure character of the Alpine metamorphism (12 kbar) on the basis of b_0 parameters of muscovites changed by the Alpine recrystallization. Korikovsky et al. (1997) determined the conditions of the Alpine metamorphism to about 500 °C and 7–9 kbar. Contrary to this, Kováčik et al. (1997) published temperatures of 350–500 °C and pressures of only 2 – 4 kbar, the obtained higher values of the pressure they explain by fluid influx. Although Kováčik (1998), basing on complex geothermobarometry, admits maximum temperatures of 550–580 °C and pressures of 8–10 kbar, in his model, on the basis of the geological situation, he prefers half-size lower pressures during the overthrusting of the southeastern Veporicum basement with a maximum 12-km-thick hangingwall (4 kbar). Based on structural relationships and geothermobarometric calculations by means of the Alpine mineral assemblage, Plašienka et al. (1999) set the peak of the metamorphism for the deeper parts of the Veporicum crystalline complex at their conception at 550 –600 °C and 8–12 kbar, which indicates burial of the Veporicum basement to a depth of 30–40 km.

The dating of the Alpine orogenic processes in Veporicum has started already by the first K-Ar dating that determined the age of biotite from the southeastern crystalline complex to 106–75 Ma (Kantora, 1960). The age of the Alpine deformation of granitic rocks from the Bacúch area was proven also by Bagdasarjanov et al. (1977) with the K-Ar method to 104–97 Ma. Another dating by Ar-Ar method (Dallmeyer et al., 1996; Maluski et al., 1993 and Kováčik et al., 1996) proved that the main period of the Alpine deformation of the Veporicum crystalline complex took place before 88–84 Ma.

Taking in to account all the above mentioned facts, as well as the present geological situation and position of individual tectonic elements within the Veporicum – Gemericum boundary, based on analogy with another Alpine type continental subduction-collision orogenic mountain ranges (Himalayas, Alps, and Pyrenees), we favor the model of the Veporicum area development in sense of Plašienka (1997) and Plašienka et al. (1999).

During the period between Jurassic and Cretaceous, after the closing of the Meliata ocean, the Meliata accretion complex started to overthrust from southeast over the passive margin of the Central Western Carpathians – Gemericum. The gradual shortening due to compression subsequently caused the overthrusting of the Gemericum with Meliaticum unit upon the Veporicum. The result of the intracratonic crustal shortening was a thickening of the crust, which involved the gravitational instability and rapid exhumation of the Veporicum basement, which was accompanied by east vergent extensive unroofing (Hók et al., 1993). Although partly problematic there remains the question of thickness of the tectonic overburden (roof). Beside the lithostatic pressure during such dynamo-metamorphic processes significant role plays also high strain rate and deviatoric stress, which magnify effect of the lithostatic pressure during collision and tectonic processes.

The SDGM studied by us are indicators of the Alpine tectonic processes. Their deformation reflects probably two tectonic events. The older high-pressure deformation originated as a consequence of maximum burial (P_{MAX} = 8–12 kbar, T = 550–600 °C in the time period before t = 110 Ma?) is overlapped by a younger event. The second episode took place during the retrograde phase of the orogeny and recorded rapid uplift of this crystalline complex block. This was caused by the reduction of thickness of the crust, thickened by the collision, and within the framework of extensive unroofing of overlapping Veporicum complexes at P = 6 kbar, T = 400 °C and in period before t = 78 Ma.

Acknowledgement

We thank Dr. I. Broska and Dr. J. Hók for their peer review of this paper and the suggestions improving its original version. Our gratitude belongs to Dr. J. Kráľ for his help with Ar/Ar dating, as well as for valuable advice improving this work. We owe our gratitude to Mr. D. Zaťovič for his precise mineral separation needed for the isotope dating. Accurate English style brushing by Dr. H. Drewes (US GS) is greatly appreciated.

P.S.: The first author (M.K.) discovered identical mylonitic rocks directly related to granitic rocks of the Nízke Tatry Mts.- Tatricum unit, at the southern slope of mountain range, during field season 2000 work. This finding can significantly changed tectonic interpretation of the SDGM in the future.

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Sample locations

NTB-1/98 – natural outcrop, crest toward Janov grúň, alt.- 825 m,
NTB-2/98 – outcrop in the road cut, Bacúšska valley, alt.- 855 m,
NTB-3a/98 – outcrop by old exploration gallery, Krškova valley, alt.- 770 m,
NTB-3b/98 – rocky outcrop, the slope opposite to 3a, in the Krškova v., alt.- 800 m,
NTB-8/98 – natural outcrop, in the Kriváň valley, alt.-755 m,
NTB-9/98 – rocky outcrop, above road, in the Leňušská valley, alt.- 1080 m,
NTB-10/98 – rocky outcrop, in the Zamrzlá valley, alt.- 1050 m,
NTB-11/98 – rocky outcrop, above a forest road, Sparistá dolina valley, alt.- 800 m.

Slovak Geol. Mag. 6, 4 (2000), 363–366

$^{40}Ar/^{39}Ar$ data from contact aureole of Súľová granite (Gemericum, the Western Carpathians)

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Abstract: $^{40}Ar/^{39}Ar$ method was used for the dating of amphibole and paragonite from contact aureole of Hnilec granitoides in Súľová area. Both samples originated from the Smrečina Formation of the Rakovec Group, from the basal part. The dating of both minerals showed identical ages of 140 Ma. These data are concordant with Rb/Sr ages from this site (145 Ma – whole-rock and mineral ages, Kovach et al., 1986). However, both radiometric ages are in contrast with radiometric ages determined in other occurrences of Gemericum granitoides. The following interpretations are provided: in Gemericum region there were two magmatic events – Late Variscan and Paleo-Alpine; $^{40}Ar/^{39}Ar$ ages can signal a strong thermal Alpine overprint caused by the amalgamation of the northern and southern Gemeric basement.

Key words: Western Carpathians, Gemericum Superunit, granitoides, $^{40}Ar/^{39}Ar$ data.

Introduction

The age of Gemericum granitoides has a long been discussed-for and despite the wide range of radiometric data available, this question has not yet been satisfactorily resolved. On the surface the granitoides form only small apical bodies that predominantly penetrate successions of the Early Paleozoic Gelnica Group. Only in small occurrences in the Hnilec valley are they situated in the basal parts of rock successions of the Rakovec Group. The only granitoid body that penetrates the tectonic contact between rock complexes of the Gelnica and Rakovec Groups occurs in the Súľová area.

The original understanding of the Gemericum as the innermost Alpine tectonic unit consisting mainly of low-metamorphic Variscan basement and its Mesozoic cover unit of "Oberostalpine" type (Andrusov, 1968) has recently been set aside. This resulted through the findings of the differences in deposition-tectonic facies of Late Paleozoic rock complexes occurring in this region (Varga, 1971; Vozárová, 1973; Reichwalder, 1973) and Mesozoic (Kozur and Mock, 1973). Finally, these new data have lead to the partition of the Late Paleozoic cover units and various Alpine structural units (Bajaník et al., 1984) and to definition of a northern and southern Gemericum (Vozárová & Vozár, 1988; Vozárová in Rakús et al., 1998). These units differ in the geodynamic development of their pre-Alpine basement and the development of their Late-Paleozoic-Mesozoic cover units. Their mutual contact is tectonic. Granitoid bodies are predominantly situated in rock complex of pre-Alpine basement of Southern Gemericum that is formed by meta-

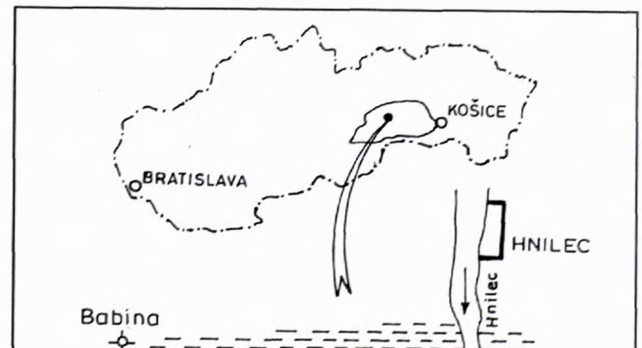

Fig. 1. Geological map of the Súľová-Hnilec area (modified after Bajanik et al., 1984) showing the sample localities.
1 – granitoides: a – Súľová type; b – Hnilec type; Rakovec Group: 2 – metasediments and metavolcanites of the Sykava Formation; 3 – metasediments, metabasalts and metabasaltic tuffites of the Smrečiny Formation; Gelnica Group: 4 – metasediments and metarhyolite tuffites of the Vlachovo Formation; 5 – overthrust plane; 6 – sample localities.

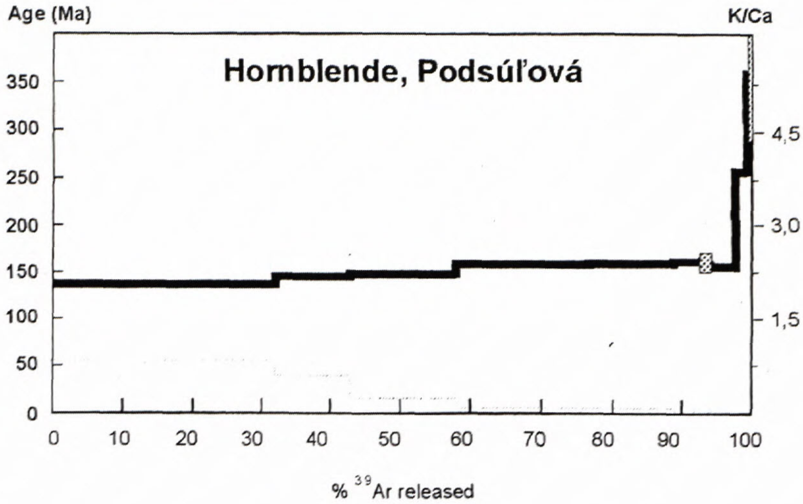

Fig. 2. $^{40}Ar/^{39}Ar$ *apparent age diagram from amphibole, Suľová*

volcanic-sedimentary successions of Gelnica Group. The rare occurrence within pre-Alpine basement of the Northern Gemeric Unit, and the penetration of tectonic contact of both units by a granitoid body, are highly significant. The direct contact is not overlapped by post-Variscan or by post-Alpine cover sediments. For this reason resolving of age problem of the Gemericum granites can resolve the problem of amalgamation of both tectonic units (?Alpine or ?Hercynian). Data originating from structural analyses and analyses of development of Late Paleozoic deposition basin support rather their Alpine age, because the polarity of Hercynian orogenesis derived from Carboniferous-Permian development of deposition basins supports the southward vergency (Vozárová, 1996; Vozárová a Vozár, 1996). This is in contradiction with presently inferred north-vergent structures in both units and in their tectonic contact.

In cover of particular granitoide body there are contact aureoles with characteristic zoned distribution of mineral associations that depends upon distance from the contact (shown on the geological map of Bajaník et al., 1984). It is important to emphasise that minerals formed in contact aureoles of granitoids are younger than the development of the last cleavage, what is valid of both tectonic units. These geologic observations have been the main reasons why these granitoides were previously held to be of Alpine ages.

Brief summary of isotopic dating

The first studies of the stratigraphic classification of the Gemericum granites led to given conflicting opinions about their intrusive ages ranging from Carboniferous to Cretaceous. Since the dating of potassic feldspar from the Betliar granite-porphyry by K/Ar method (98 Ma, Kantor, 1957), a Cretaceous age has been preferred. The following investigation indicated that the problem is much more complicated. Next K/Ar dating confirmed Alpine ages: 70 Ma at the Poproč site (Bojko et al., 1974); 87 Ma on Zlatá Idka site (Bagdasarjan et al., 1977); 141 Ma on the Čučma site (Bagdasarjan et al., 1977). Kantor and Rybár (1979) published many of K/Ar data from various bodies of Gemericum granites and in a case of muscovites they found that values range from 241 to 141 Ma. From this range the authors have concluded that the Gemericum granites were polyphased. Rb/Sr data (Kovach et al., 1986; Cambel et al., 1989) gathered from whole rocks and separated minerals (mainly muscovites and biotites) also documented considerable scatter of WR data and mineral isochrons of particular bodies (for instance Hnilec 290±40 Ma, Betliar 272±40 Ma, Podsúľová 145±6 Ma). Permian ages gained from Rb/Sr whole rock isochrons were confirmed for Hnilec and Betliar granitic bodies with a higher precision by U, Th – total lead determination through the use of a microprobe on monazites (Finger a Broska, 1999).

The dated samples

Samples of newly formed amphibole and white mica (paragonite) that occur in epidote-chlorite schists (metabasaltic tuffites) and fine-grained sandy metapelites of the Smrečina Formation of the Rakovec Group (Fig. 1) were separated from contact aureole of the Súľová granite. Both samples were taken from Northern Gemeric basement.

The sample G-11A was taken from metabasaltic tuffites (site on left slope of Hnilec river, about 700 m northwest of the railway station Delava), in which an association of regionally metamorphosed minerals occurs. These consist of chlorite + epidote ± albite, actinolite, calcite, quartz.

These minerals form fine-grained, markedly alined up aggregate, complexly deformed by transverse cleavage. In this fine-grained structure there are omnidirectionally oriented long-columnar crystals of green amphibole that were separated for radiometric dating. Metabasaltic tuffites alternate with chlorite-muscovite phyllites in which unoriented crystals of light brown biotite occur.

Sample G-13 was taken from the contact of fine-grained metasandstones and sandy metapelites with granitoide body, about 1 km NNW from the Súľová ranger's cottage, from a slope above the forest road. Fine-grained metasediment with a strong lineation contains two mineral associations. The one is a regional metamorphic mineral association consisting of quartz + muscovite ± chlorite. The second and younger lineation consists of thicker crystals of white mica (paragonite) + quartz + tourmaline ± plagioclase. For the radiometric dating the white mica was separated.

Results

The samples were dated by standard method that is recently used in the laboratory of GEOZENTRUM, Vienna. In Table 1 and 2 the basic analytic data from the sample of recrystallized amphibole G-11A and white mice are presented.

Tab. 1: Analytic data from the sample of newly formed amphibole G-11A

Step	T (°C)	%^{39}Ar	^{40}Ar* (mV)	%Ar*	^{40}Ar*/^{39}Ar (±2 SD in%)	Age (in Ma) ±2 SD
1	700	32.0	149.46	94.9	15.92±0.8	137.0±1.1
2	750	11.0	54.55	94.4	16.88±1.0	144.9±1.4
3	800	15.0	76.68	96.0	17.43±0.7	149.5±1.0
4	850	18.7	102.26	96.4	18.62±0.8	159.3±1.2
5	900	11.3	61.81	96.3	18.70±1.4	159.9±2.2
6	950	4.1	22.90	87.9	19.00±2.2	162.4±3.4
7	1000	1.4	8.01	96.1	19.13±5.7	163.4±8.9
8	1050	4.5	24.22	99.0	18.56±1.2	158.8±1.8
9	1100	1.6	14.53	97.1	31.40±5.3	261.3±13.0
10	1250	0.3	4.40	82.3	44.87±21.3	363.2±70.5

J= 0.004735±0.4% total gas age= 151.6 ± 3.1 Ma

Tab.2: Analytic data from the sample of newly formed white mica (paragonite) 171/13

Step	T (°C)	%^{39}Ar	^{40}Ar* (mV)	%Ar*	^{40}Ar*/^{39}Ar (±2 SD in%)	Age (in Ma) ±2 SD
1	650	90.3	27.29	95.4	16.34 ±1.5	140.5±2.0
2	700	9.7	2.3	71.7	12.87±8.2	111.5±8.9

J= 0.004735±0.4% total gas age= 137.7 ± 3.7 Ma

Discussion and conclusion

The ages ^{40}Ar/^{39}Ar obtained from both minerals are partially different, although the ^{40}Ar/^{39}Ar age of paragonite degassed in one step is similar (140.5 Ma) to the apparent ages in low temperature steps of sample G-11A. These data are almost concordant with the youngest Rb/Sr ages. These data emphasise the importance of the remarkable Alpine event that can be interpreted in such different ways:

1. If we understood the transversal minerals in vicinity of the bodies as manifestation of their contact influences, then the granite from Súľová site could be considered as an Alpine one. However, this inference is negated by the fact that the body in Súľová area was until now considered to be a part of the Hnilec body that was dated at Permian by the Rb/Sr whole rock analyses and separated mineral analyses (Kovach et al., 1986) and U/Th, Pb method in monazite (Finger and Broska, 1999). But the Súľová and Hnilec bodies are separated from each other at the surface, although their interconnection under the surface was always assumed. A very important fact supporting the Alpine age interpretation of the granitoid body that occurs in ridge part of Súľová is 145 Ma Rb/Sr age that has been generally interpreted as age of Alpine overprint (Kovach et al., 1986; Cambel and Kráľ, 1989). The congruity of the older dating with the new mineral Ar/Ar dating is significant. This determination shows that the observed radiometric ages confirm an Alpine age of intrusion of the granitoid body in Súľová, with respect to the fact that the granitoids do not bear signs of dynamic-metamorphic reworking, and minerals in contact aureole are randomly oriented i.e., post-cleavage. In that case we should suppose that in the Gemericum zone the granitoides to two separate magmatogene events (Late Variscan and Paleo-Alpine).

2. Newly formed transversal amphibole and paragonite are not genetically connected with granite intrusion, they are products of hydrothermal changes connected with significant Alpine temperature event. It would mean that the total Alpine reworking is the most significant along the contact of two pre-Alpine basements. One possibility of interpretation is that the thermal flow was caused by infracrustal movements during the amalgamation of northern Gemeric and southern Gemeric basements and therefore our data gained from ^{40}Ar/^{39}Ar dating indicate an Alpine age of this event.

The new Ar/Ar dating evoked new questions in the problem of dating the Gemericum granitoids. A reliable solution of this problem calls for the verification of mineral ages of contact aureole of other apical bodies, i.e., those that are situated directly in the south Gemericum basement.

Acknowledgements

Igor Broska is thanked for his helpful reviews. The manuscript was improved by helpful comments as well as careful reviews by H.Drewes.

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Slovak Geol. Mag., 6, 4(2000),367–376

Petrogenesis of Metamorphosed Ironstones Near Kokava nad Rimavicou (Veporicum, Western Carpathians)

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Abstract: Primary ferruginous sediments shared together with their host metapsammites a basic regional tectono-metamorphic event in the condition of the amphibolite facies. This was followed by multistage Hercynian granitization with some periplutonic effects. An Alpine overprint caused only localised cataclasis and/or recrystallisation. At least six basic petrographical types are assigned to these ferruginous rocks. Magnetite is a frequent constituent in a characteristic metamorphic mineral assemblages built with garnet (almandine) - amphibole (grunerite) - apatite ± biotite (annite), quartz, chlorite (brunsvigite), graphite and allanite. Grunerite appears in many petrostructural positions (porphyroblasts, fine-grained aggregates, needles, swarms), which reflect polygenetic metamorphic reactions. Generally, deposition of sandstones, chamosite bearing ferrolites, black quartzites and the occurence of apatite-bearing laminae support an idea about a shallow-level sea sedimentary facies. The uniquity of (meta)ferrolites, among the more common rock association indicate a more specific environment (a bay or a deltaic system) of deposition. The occurrence of magnetite resulted from a complex combination of many factors. A case study of metamorphosed ironstones show an affinity of magnetite to a more pronounced pelitic character, corresponding also to a greater extent of Fe^{3+}. The original Eh of certain beds determine the nature of Fe-phases crystallisation (e.g. ilmenite vs. magnetite or grunerite vs. magnetite) in the process of regional metamorphism.

Key words: regional metamorphism, ironstone, ferrolite, grunerite, magnetite, annite, Hercynian basement, Western Carpathians

Introduction

By the road in the valley 3 km from the Kokava nad Rimavicou site towards to Hriňová (the Hrabina area) there are occurrences of magnetite iron ore (Fig. 1). The site represents an unique accumulation of magnetite mineralization in the rock of the Western Carpathians crystalline complexes. This study includes a brief structural-geological descriptions, a division into lithotypes, petrology of the selected ferruginous schists, and assessing metamorphic development, magnetite generation and genetic aspects of the pre-metamorphic source. The basis for this article stems from research that was focused mainly on problems of graphite in the adjacent metaquartzites (Petro et al., 1998, Kováčik 1998).

Overview of the Geological - Depositional Knowledge

The pilot mining - geological study of the magnetite mineralization was published by Šuf (1938). With respect to the irregularity of the deposit occurrences and of the great hardness of the ore-bearing rocks he considers the ore deposit as non-prospective. Migmatite was recognized as the basic host-rock and the studied area was integrated into the northern migmatite zone (Šuf 1937, 1938). The outlined area was later considered as part of the zone composed mainly of the so-called late-orogenic migmatites and granitoides (Hovorka in Kuthan et al., 1963). This zone is also designated as the hybrid zone, since its magmatic-metamorphic heterogeneousness (Bezák 1988).

During the 50-ies an intensive mining prospecting took place in the wider area of the prior-known occurrences. Zoubek and Nemčok (1951) localized the ore deposit into strongly granitized zone where migmatites and katazoned gneisses evolve from paragneisses. In the ore bearing rocks they distinguished the type composed of magnetite and garnet; garnetstones (without much magnetite) and biotitstones. The authors observed abundant occurrences of apatite and flaky biotite and emphasized a notable lack of hedenbergite. Although the deposit is conventionally classified as the „Kokava scarn", in question of the genesis the preferred idea is that it is a regionally metamorphosed sedimentary ironstone deposit (Zoubek and Nemčok, 1951).

Based on the study of mineral paragenesis, Gubač (1957) proposed a contact-metasomatic origin of the deposit. He understood the original source material as regionally wide-spreaded gneisses of a claystone protolith that were changed by pneumatolitic - metasomatic effect of the pegmatites after the regional and subsequent contact metamorphism. He attributed the development of the

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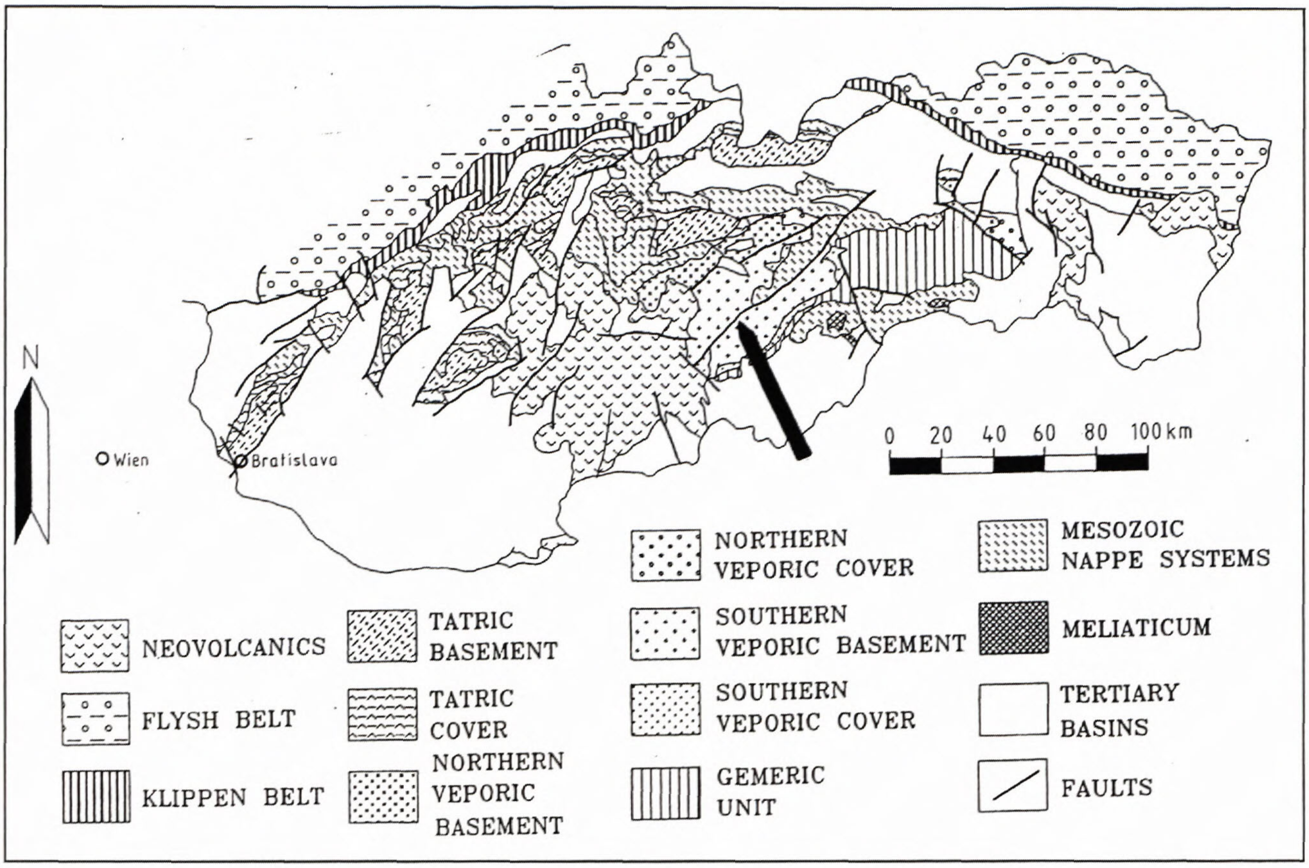


Fig. 1A The geological scheme of the Slovak part of the Western Carpathians; the arrow points out the studied area.

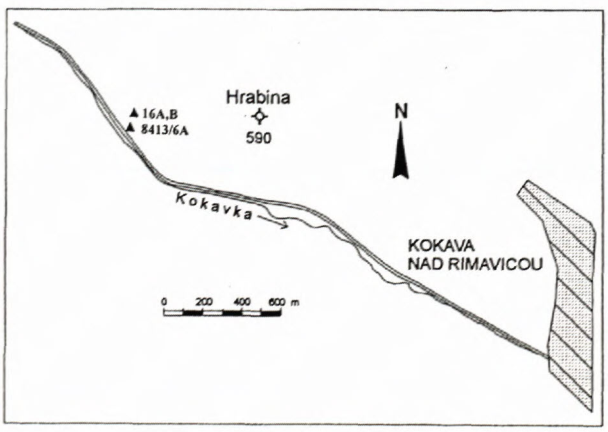


Fig. 1B Site localization of the samples studied on electron microprobe

magnetite ore and accumulation of biotite and garnet to the mineralization process of the pegmatite stage (so-called III. stage). Korikovskij et al., (1989) attributed a sedimentary origin to rock with a mineral assemblage of magnetite-garnet-grunerite. Phanerozoic chamosite ironstones are considered as the most probable source material and the metamorphic temperature is inferred to be about 550°C (Korikovskij et al., 1989). Radvanec (2000) viewed special petrologic questions about the ferruginous assemblages.

Structural - Geological Observations

The construction of a gneiss-migmatite substrate with lenses of graphite quartzites and gneisses, pure quartzites and magnetite-garnet gneisses show a conformable position. Alternations of various types of gneisses (basic gray fine-grained siliceous types, miscellaneous banded types, migmatitized varieties, coarse-grained garnet orthogneisses etc.) can be observed on a mezo- (x-dm) and macro-scale. Lenses of greenish-gray amphibolite rocks occur rarely. The general dip of the rock complex is northward. Folds (many with cylindrical shape) are locally present as well, their shallow to sub-horizontally dipping b-axes are parallel to the strike. The question of the presence of the granitoids seems to be more complicated. Part of the granitoides, including also the dark-gray biotite-plagioclase granitoid rocks of the so called hybrid type, share a mutual deformation history with the gneissic substrate. This is also shown by the rare granite boudins in the gneiss-migmatite base rocks. The migmatitization is probably related to this oldest episode of granitoid formation. In places it can be observed that the biotite gneisses are injected by a light material ("neosome") which created quartz-feldspar eyes or also banded structure. Abundant enclaves of gneisses and oriented relics of graphitic substances enclosed in the later and predominant type of granitoids indicate magmatic assimilation of the metamorphic substrate.

These structures pierce several-decimeters-thick veins of light pegmatitic granite that also indicates formation in a synkinematic regime. In the rigid metaquartzites and in the magnetite-garnet gneisses, there are some nests of light granitic material and feldspathisation connected with silicification. The intrusions of this relatively younger leucocratic melt also indicate a assimilation or partial-anatectic effects on the basic biotite gneisses. In the light pegmatite granite, there are characteristic inclusions of gneiss relics containing as much as 1-2 cm large garnet porphyroblasts, which apparently originated as a result of periplutonic effects of this granite. However, the garnet phenocrysts are in places also present in these granites (residuum from incompletly assimilated gneisses?). Because of the variegated shape and composition each of the granitization-stage, the time and source relation between the predominant Veporic intrusion and the other plutonic events cannot be always expressed explicitly.

The sub-vertical east-west shear-deformation including gneisses and granitoides can be locally observed. Relatively weak sub-horizontal lineation with a similar orientation may indicate a younger Alpine phenomenon. Generally, it seems that the Alpine deformation-recrystallisation processes are not as intensive in the studied area as in the other areas of the southern Veporicum. This is probably caused by more resistant rheology of the rocks in investigated area. The Alpine deformation mostly uses inhomogenities of the pre-Alpine fabrics and it manifests mainly on the lithological boundaries, or follows directions of the Hercynian deformation. For example, the development of the omnidirectional biotitization on the contact with the granite vein penetrating into a fine-grained gneiss is most likely a Hercynian phenomenon. Afterwards, the biotite rims used to be deformed locally, and fine-grained micaceous mixture of presumably Alpine age were generated.

Petrography and Geochemistry of the Rock-forming Minerals

Within the ferruginous rock ensemble there are a number of lithological variations that are often reflected in the banded mesostructure. Generally, the dark to black lithotypes with conspicuous red garnet and shiny faint-green amphibole clusters are dominant. Considerable sedimentary and polymetamorphic variability led to different petrographic division of these rocks (Zoubek and Nemčok 1951, Gubač 1957, Korikovskij 1989). (The rock divisions listed bellow are also only an incomplete overview of the ferruginous lithotypes). The rock composition also shows, that feldspars and white mica are not present here. The rock texture is usually homogeneous, sometime it has a massive, cherty character (for example, straight boundaries of polygonal quartz-grain texture in the fine-grained mineral assemblage in the rock type 2). According to the presence and quantitative abundance of the rock-forming minerals the following petrographic types of ferruginous metamorphic rocks are recognized:

1) garnet > biotite > quartz > magnetite > amphibole;
2) garnet > quartz > magnetite;
3) garnet >> amphibole ≥ magnetite >> biotite;
4) garnet > amphibole ≥ quartz
5) garnet >> quartz > amphibole > magnetite > biotite
6) garnet >> magnetite > amphibole >> biotite

Garnet is the basic component of the rocks; rocks composed up to 90 by vol. % of garnet are assigned to (amphibole)-garnetstones. Garnet is here almost in a "cast" form, individual crystals reach 2 - 3 mm (for example, types 3, 5, 6). As it grows, the garnet encloses about 0.05 - 0.1 mm grains of magnetite, sometimes also tiny biotite, apatite, or grunerite, which probably represent relicts from pre-metamorphic or early-metamorphic stages of the rock development. According to the chemical analyses (Tab. 1, Fig. 2) the garnet is the *almandine type*. In mineral assemblages without magnetite (sample 16A, rock type 4) garnet contains up to 17% of the spessartite components. Garnet associated with magnetite (sample 16B, rock type 3) has up to 82 % of the almandine component and a little bit greater portion of Ca-, but it contains small amount of the Mn-component. A potential chemical zonality is not clear, the garnet can be considered more or less homogeneous.

Tab. 1 Microprobe analyses of garnet in sample without - (16A) and with (16B) magnetite. The chemical composition is recalculated to 12 oxygens; ratio M/MF = Mg/(Mg + Fe_{tot}). (Mineral composition in Tabs. 1 - 3 was made on the device Jeol Superprobe 733 by RNDr. M. Köhlerová a ing. A. Sonáková.)

sample	16A			16 B	
	margin	between	centre	margin	centre
SiO_2	36,73	36,681	36,556	36,651	36,913
Al_2O_3	21,443	21,411	21,4	21,253	21,627
MgO	1,084	1,371	1,158	0,99	1,075
FeO	31,488	31,256	31,101	36,058	35,708
MnO	7,382	7,026	7,398	2,056	1,975
CaO	2,146	1,997	2,175	2,982	2,958
total	100,273	99,742	99,788	99,99	100,256
Si	2,981	2,984	2,978	2,983	2,985
Al IV.	0,019	0,016	0,022	0,017	0,015
Al VI.	2,032	2,037	2,033	2,021	2,046
Mg	0,131	0,166	0,141	0,12	0,13
Fe2+	2,137	2,127	2,119	2,454	2,415
Mn	0,507	0,484	0,511	0,142	0,135
Ca	0,187	0,174	0,19	0,26	0,256
sum	4,994	4,988	4,994	4,997	4,982
Alm	72,147	72,077	71,564	82,46	82,255
Spes	17,117	16,401	17,258	4,77	4,598
Pyr	4,422	5,625	4,762	4,032	4,428
Gros	6,313	5,896	6,417	8,737	8,719
M/MF	0,058	0,072	0,062	0,047	0,051

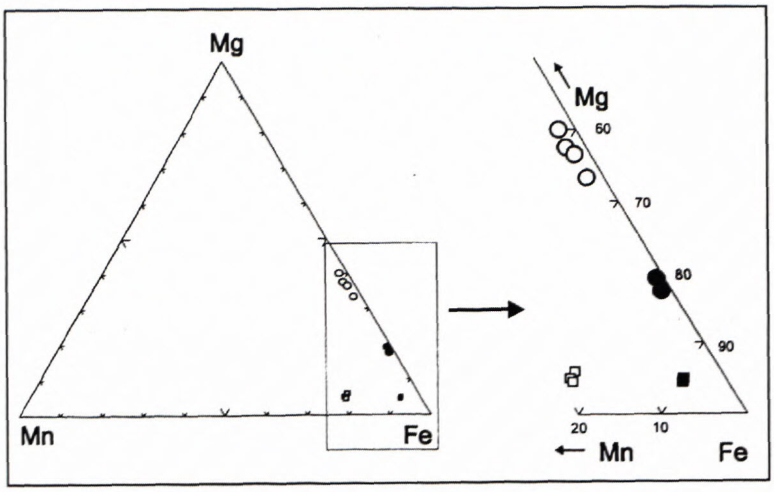


Fig. 2 The projection of the garnet (rectangular) and grunerite (circle) composition in the sample without - (empty symbols) and with magnetite (full symbols). (For more information see the text and Tabs. 1 and 2). The basic mineral assemblage is garnet - grunerite - biotite.

Tab. 2 Microprobe analyses of grunerite recalculated on 23 oxygens. Ratio M/MF expresses also the percentage of cummingtonite component in the amphibole.

sample	16A				16B	
	basic type	light phase	needle-type	needle-type	basic type	tiny type
SiO_2	52,525	52,605	52,688	53,251	49,287	49,631
Al_2O_3	0,23	0,373	0,142	0,189	0,221	0,283
MgO	11,186	9,775	11,864	10,988	5,417	5,001
FeO	31,84	33,81	30,578	34,161	40,129	41,213
MnO	1,28	1,086	1,071	1,04	0,406	0,675
TiO_2	0	0,003	0	0	0	0
CaO	0,13	0,127	0,122	0,026	0,12	0,099
K_2O	0	0,127	0	0	0,035	0,003
Na_2O	0,077	0,105	0	0	0,035	0,002
total	97,268	97,906	96,464	98,453	95,624	96,908
Si	8,05	8,071	8,083	8,072	8,033	8,018
Al IV.						
Al VI.	0,042	0,067	0,026	0,034	0,042	0,054
Mg	2,555	2,236	2,713	2,483	1,316	1,204
Fe	4,081	4,339	3,923	4,161	5,47	5,568
Mn	0,166	0,141	0,139	0,133	0,056	0,092
Ti	0	0	0	0	0	0
Ca	0,021	0,021	0,02	0,026	0,021	0,017
K	0	0,004	0	0	0,002	0,001
Na	0,023	0,031	0	0	0,011	0,001
sum	14,939	14,911	14,903	14,91	14,951	14,955
M/MF	0,385	0,34	0,41	0,374	0,194	0,178

The ***amphibole*** belongs to the group of the monoclinic Fe-Mg-Mn amphiboles and its composition (Tab. 2) fits the field of *grunerite* (Leake, 1997). The grunerite is non-pleochroic, it has mostly variegated interference colors. It commonly forms fine-grained aggregates, found in interstitial spaces among garnets. In a thin section, the grunerite aggregates are intergrown with magnetite (Fig. 3) or do not contain it. This fact may indicate a dependence upon the chemical composition of the primary crystallization domain. An interesting phenomena represents scarce diablastic intergrowths of magnetite and grunerite (Fig. 4), which also can be attributed to a high iron content in a local aluminiumpoor crystallization system. These structural forms, together with the 2-3 mm porphyroblasts, represent relatively the older grunerite generations. Locally also, a skeletal crystals developed - thin long needles or swarms of grunerite grow into quartz or biotite. Figures 5a and 5b illustrate the basic mineral assemblage (rock type 1) garnet - biotite - magnetite and grunerite, which occurs in two morphological forms. Incomplete crystallization of the "skeleton" grunerite in biotite probably indicates an excess of Fe^{2+} in the system, along with a concurrent lack of aluminum and eventually also potassium. Because the metamorphic conditions in these metamorphic minerals are indicated also by the mutual distribution of Fe and Mg (or also Mn), neither the garnet nor the biotite were able to bind greater amount of FeO at the given metamorphic grade. A similar situation is found in the case of grunerite needles grown into quartz, which indicate the predominance of the two-valent iron, otherwise magnetite would be generated (rock type 5).

Within a studied sample, all forms of grunerite (i.e. porphyroblasts, fine-grained aggregates and needles) have practically identical chemical composition (see Tab. 2). This can indicate either equal conditions of metamorphic crystallization, or rather the fact that the metamorphic reactions are to a great extent controlled by the chemical composition of the protolith. The grunerite in the samples without magnetite contains an increased amount of Mg and Mn, analogously as in garnet. Microprobe analysis of the grunerite crystals also did not show a relevant metamorphic zoning.

Biotite - the crystal size considerable depends upon its quantity. If insignificant, then it can be observed as a tiny (about 0.1 mm) inclusion in the garnet, if abundant, then it occurs in 1-3 mm large blasts, similar to garnet or grunerite (Fig. 5). Some of the non-directional biotite flakes were probably generated due to thermal and metasomatic (input of K_2O and H_2O) effects of the surrounding granitoides. On the contrary to the garnet, the fine-grained inclusions of magnetite are not

Tab. 3 Microprobe analysis of biotite and secondary chlorite that documents the ferruginous protolith (decreased K_2O content in the biotite 16B indicates a dealkalization during the initial chloritization stages). Biotite is recalculated to 22 and chlorite to 28 oxygens.

sample	16 B				8413/6A	
	Chl	Chl	Bt (dark ph.)	Bt	Bt	Bt
SiO2	24,213	23,872	34,11	33,936	33,9239	34,132
Al2O3	17,716	17,676	16,004	16,191	14,6753	15,245
MgO	4,367	3,969	3,243	3,735	4,8022	5,258
FeO	39,879	40,257	33,051	32,46	31,259	31,189
MnO	0,18	0,215	0,041	0,035	0,1298	0,147
TiO2	0	0	1,506	1,504	1,882	1,774
CaO	0,04	0,021	0	0,033	0,064	0,275
K2O	0	0	5,662	5,73	8,32	7,893
Na2O	0	0	0,153	0,162	0,189	0,145
total	86,395	86,011	93,769	93,788	95,345	96,309
Si	5,659	5,627	5,556	5,516	5,5008	5,453
Al IV.	2,341	2,373	2,444	2,484	2,5008	2,547
Al VI.	3,318	2,538	0,629	0,618	0,304	0,324
Mg	1,521	1,395	0,787	0,905	1,161	1,252
Fe	7,794	7,936	4,503	4,412	4,238	4,167
Mn	0,036	0,043	0,006	0,005	0,018	0,02
Ti	0	0	0,184	0,184	0,229	0,213
Ca	0,01	0,005	0	0,006	0,011	0,047
K	0	0	1,177	1,188	1,721	1,609
Na	0	0	0,048	0,051	0,059	0,045
M/MF	0,163	0,15	0,149	0,17	0,215	0,231

characteristic for the biotite accumulations, whereas grunerite is there present more frecquently. For this reason it can be assumed that the biotite binds a certain amount of Fe^{3+}. The chemical composition of the biotite (Tab. 3) documents the genesis in the ferruginous - and aluminapoor environment. Considering the classification, the composition of the biotite is close to the end member - *annite* (Guidotti 1984). Comparison of the biotite from two samples (Tab. 3) indicates that a decrease of Al_2O_3 is accompanied by an increase of MgO (and MnO), which represents a certain shift from the siderophyllite to phlogopite component in the classification diagrams.

Quartz occurs in variable abundance, some lithotypes of the ferruginous metamorphic schists do not contain it at all. The synsedimentary origin is indicated by lamina enriched with quartz (mostly to the detriment of garnet) and by quartz lenses with apatite or with the rosette-like grunerite inclusions. Quartz with a genetic relationship to the protolith is usually of finer-grained texture. The younger quartz depends upon the mass influx, connected either with granitization or later hydrothermal processes.

Magnetite is scattered throughout the rock with a frequent grain size of 0.2-0.3 mm. In richer ore accumulation 1 mm or greater crystals are not rare. The coarse-grained magnetite occurs mostly at the contacts with garnet. The samples, in which the garnet encloses a quantum of tiny magnetites, these grains are not observed in grunerite or biotite blasts (Fig. 5a). This points to the development during early stages of metamorphism (or resistance or/and recrystallization in the diagenetic stage?), similarly as do the intergrowths of grunerite and magnetite (Figs. 3 and 4). The grunerite porphyroblasts in places enclose larger magnetite crystals, which can indicate a general increase of grain-size during the progressive metamorphic growths.

The magnetite presence depends to a great extent upon the composition of the primary lithologic domain and probably also upon the redox-potential form of the iron inside a sedimentary bed. The appearence of magnetite is the result of a combination of several factors, from which one example can be mentioned: on the contrary to the rock type 3 (sample 16B), in the rock type 4 (sample 16A) there is no magnetite, but it contains more quartz and grunerite. The absence of magnetite is caused mainly by a lower content of iron in the rock, which also has an effect on the composition of grunerite and garnet (Fig. 2). Except the high iron content in the sample with magnetite, from the mineral composition results, that the rock contains also increased amount of Al_2O_3 (90 vol. % of garnet, presence of biotite and chlorite). A similar situation also occurs in the 1. rock type.

The mineral assemblages are accompanied by ***apatite*** that frequently reach 1 - 3 vol.%, and in some cases even more. Apatite crystallized from the early metamorphic stages (e.g. inclusion in garnet) to the peak metamorphic stage (coarse crystals in textural relations with the basic metamorphic assemblage). It commonly reaches a size of about 0.15 mm, locally it creates up to 1 mm crystals. Apatite frequently forms fine-grained aggregates in lenses (width about 1-2 mm, length 5-10 mm) where it dominates over quartz. The metamorphic assemblages include also ***allanite*** with crystals up to 0.5 mm. ***Ilmenite*** occurs in these ferruginous schists where magnetite is not present (for example sample 16A). Similarly as in the surrounding gneisses and black metaquartzites, the ***graphite*** (this polymorph of carbon was proved by RTG diffract. analysis by Pulec 1989 and Očenáš in Petro 1998) is a not infrequent admixture of ferruginous schists. The petrographic textures show that it is a metamorphic product of synsedimentary organic substances.

From the ***secondary minerals***, rich-green ***chlorite*** is the typical one. It most frequently fills joints in the garnet; however, it also originates at the expence of biotite. Grunerite use to be replaced mainly near contacts with garnet, from which the chlorite probably gains Al_2O_3. The

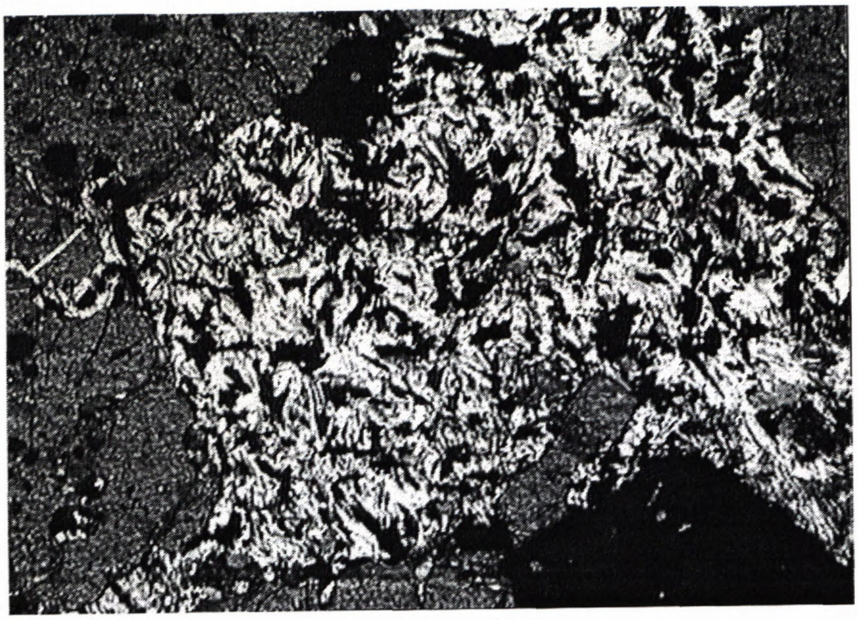


Fig. 3 Fine-grained aggregate of grunerite (bright phases) and magnetite (black) enclosed in dark garnet matter (//N, 34x magnification)

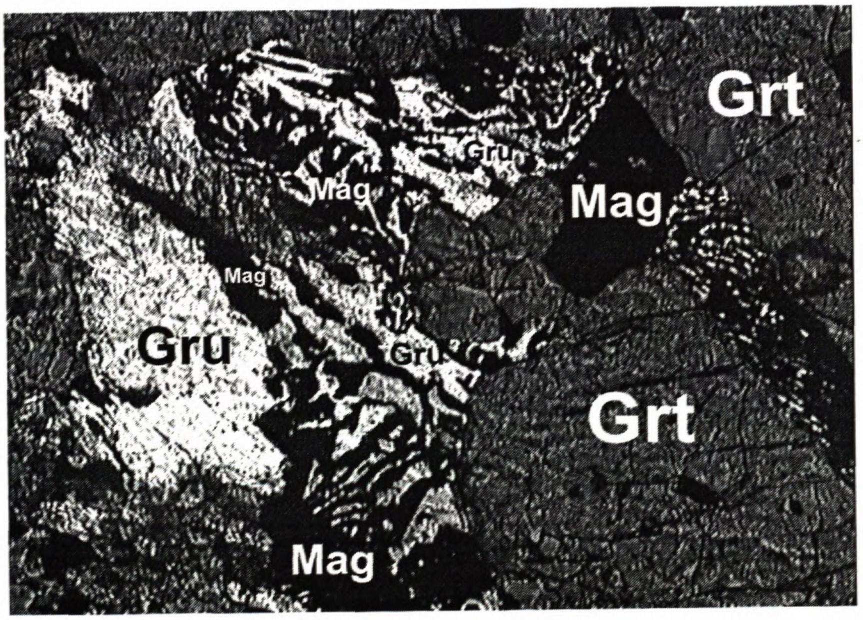


Fig. 4 Diablastic intergrowth of grunerite with magnetite; the surrounding mass forms mainly garnet (//N, 185x magnification).

chloritization, probably of the Alpine age, is locally very intensive and, on the contrary, in some domains it totally absents. The assemblage chlorite - magnetite - apatite indicates that along with magnetite apatite also underwent Alpine remobilization. Locally, the pre-Alpine origin of chlorite cannot be absolutely excluded, mainly if it occurs in the form of inclusions in the garnet. (Under middle pressure conditions Fe-chlorite can resists up to a temperature of about 600 °C, as experimentally proven by James et al., 1976). Also, it is not always clear whether part of the post-kinematic chlorite does not originate in periphery zones of the periplutonic effects of the granites. The chlorite can be classified as *brunsvigite* (Tab. 3), which projective point falls near the border with ferruginous ripidolite (sensu Hey 1954). The geochemical criteria for distinguishing of the potential chlorite generations were not recognized. From other secondary minerals we can mention rare *sulfide dissemination*, *carbonates* and clusters of dark *clay minerals*.

Results and Discussion

A) Pre-metamorphic Genesis

A specialty of the studied locality is, in the first order, the ferruginous protolith. In this relatively small area other unusual elements were also discovered; however, their areal exclusion cannot be confirmed so far. The increased content of apatite, not only in the Fe-rich rocks, is also characteristic. For example, in the fine-grained garnet-muscovite gneiss 1-2 mm thick intercalations containing 30 - 50 % apatite are present, and it is the same in some graphite metaquartzites. The spatial association of graphite metaquartzite and ferruginous gneisses is also noteworthy.

Fig. 5a Two forms of the grunerite on the BSE image: in the middle part there are euhedral grunerite and fibrous clusters in the biotite (bright phase under grunerite crystal and also more to the left). Bright gray porphyroblasts represent garnet, the shining grains are magnetite.

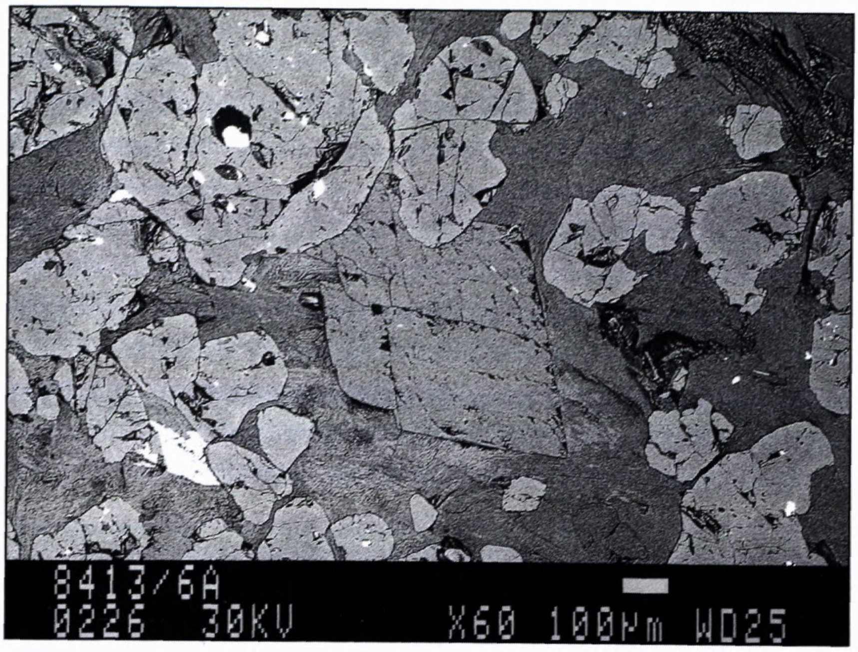
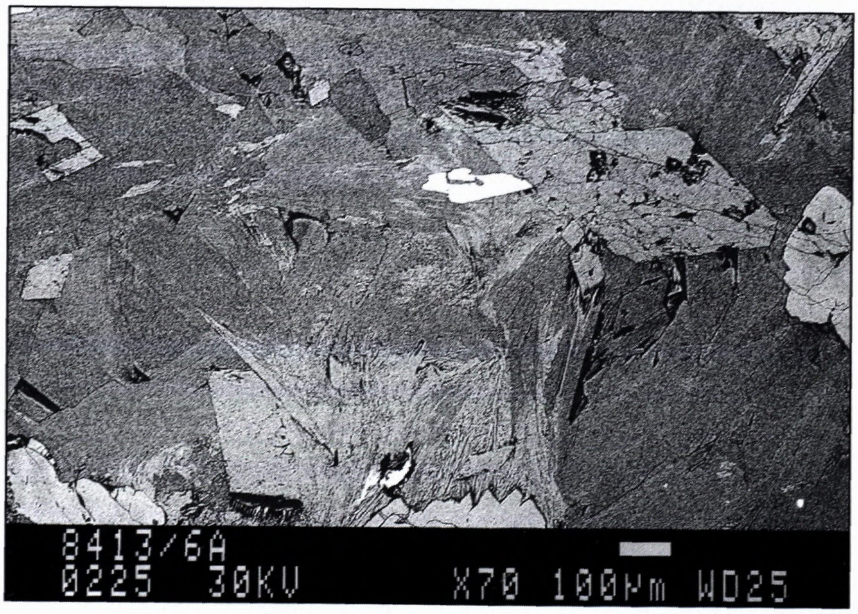


Fig. 5b Detail of the incomplete grunerite development (brighter phases in the middle over the bottom margin) in a dark gray mass of the biotite (biotite composition in Tab. 3). (BSE images made RNDr. F. Caňo on device JSM-840.)



In the question of genesis of iron formation four basic facies are distinguished - oxide, carbonate, sulfide, and silicate - which can mutually overlap to a certain extent (James, 1954). Discussion regarding the origin of the investigate locality is naturally focused on the carbonate and silicate types. In the carbonate Fe-formations the content of the quartz is relatively constant (about 30-35%), whereas the composition of the Fe-silicate rocks vary significantly (James, 1954), which remains the Kokava deposit. The significant aluminum content resulting from the dominat abundance of the garnet designates the chamosite as the main mineral precursor of the studied rocks. The higher aluminum content is understood as a result of a more mature Phanerozoic crust, contrary to the low content of aluminum in Fe-silicate formation of the Precambrian (James, 1954, Bailey and James, 1973). These facts were also pointed out by Korikovskij et al. (1989); thus, we can join the conclusions of their and Zoubek and Nemčok (1951) work to characterize the Fe-protolith as the Phanerozoic chamosite ironstones. Phanerozoic ooidal ironstones were common in Ordovician and Devonian time (e.g. Petránek, Van Houten 1997), which is characterized by a high relative rate of marine sedimentation on the continents (Ronov at al. 1980).

The potential carbonate precursor of the main ferruginous rocks can also be excluded for other reasons: relicts of the synsedimentary layers of carbonates were not found; the increased content of Ca or Mg that are the usual components of Fe-carbonates (Haase, 1982, Tucker 1985), does not manifest itself in the rock composition; it is hard to explain the mechanism of material input (mainly aluminum) into the potential Fe-carbonate layers, as the metamorphics were formed in the pre-granitization stage. The composition and structures of the rocks reflect fairly well the original lithology; infiltration metamorphism of the carbonates (scarnization) generally results in garnet enriched in the andradite-grossularite composition, pyroxene of diopside-hedenbergite type, Ca-amphibole, epidote etc. (e.g. Němec 1991), thus minerals that are not known from the studied site. However, we can assume some siderite admixture in the Fe-protolith, because the carbonates (mostly of diagenetic origin) are common components of chloritic beds in the iron formation (James 1954, French 1973, Korikovskij et al., 1989).

We can assume that the ferruginous metamorphites originally represented fine-grained beds of pelitic to chemogenic character that were deposited in the environment of the sediments with prevailing psammite lithology (i.e. surrounding gneisses). The higher content of Al_2O_3 and eventually K_2O suggests a more significant involvement with the clayey components into the development of ferruginous beds. The chamosite ores are generally considered as near-shore sediments with a diagenetic origin (Tucker 1985); however, there is a lack of the data about a more detailed elaboration about concrete paleoenvironment or sedimentary source. Considering the unique occurrence of the metaferrolites in a broad regional scale, it is probable that a certain specialty of the shallow-water depositional environment played its role, too. It can be attributed, for example, to a bay or a deltaic system prograding to the shelf zone. The accumulation of P_2O_5 reminds, in certain features, of accumulation of shallow-marine phosphorite that commonly is associated with limestones, sandstones, ferruginous sediments and black shales (Kukal 1991). The source of phosphor probably originates from calcareous shells of microorganisms. The minor phases as organic matter (changed to graphite), apatite, ilmenite and part of the magnetite can be considered for relicts from deposition, diagenetic and/or early metamorphic phase of ferruginous rocks development.

Within the rock-complex surrounding the ore deposit there are rarely found amphibole-clinozoisite schists. They are made of as much as 2/3 of clinozoisite, further contain amphibole of actinolite type, quartz, biotite, as well as ubiquitous tiny titanite grains. The clinozoisite belongs to the basic metamorphic assemblage, but it also partly takes place in corrosive reactions. The rock represents an unusual petrographic type that probably belongs to the group of hedenbergite - clinozoisite - quartzite schists described by Korikovskij et al. (1989). The authors recognized the source material as psammite enriched with Ca and eventually with Fe. From our standpoint it seems that the protolith was represented more probably by clayey - calciferous material, which is indicated by the low content of alkalies - K_2O occurs only in biotite and a Na_2O-bearing mineral is not present at all (for example plagioclase). It can be suggested that these rocks are of similar origin to the calc-silicate hornfels with metamorphic assemblage amphibole - clinopyroxene (diopside-hedenbergite type) ± clinozoisite, which are occasionally found in the Kohút massif (Vrána 1964). The rocks with clinopyroxene signalize sporadic pre-metamorphic carbonate admixture in the highly metamorphosed "hybrid" rock-complex of the Kohút zone. The fact that these rocks also contain magnetite, and that the clinopyroxene represents hedenbergite is in accordance with the overall ferruginous character of the Kokava site.

B) Metamorphic Processes and the Questions of the Magnetite Mineralization

The studied area belongs to a polymetamorphosed crystalline complex, where three metamorphic events took place. Regional metamorphism, most probably of Hercynian age, represents the basic metamorphic event (M1). This was followed by several, often unclear, mutually overprinting periplutonic phases of the Carboniferous granitoids, from which the effects of the acid pegmatite granites (M2) are the best distinguishable. The partial reworking of the regionally-metamorphic mineral parageneses caused by younger granitoid phases is a characteristic phenomena of the given area. Although such an importance cannot be attribute to this stage as Gubač (1957) does. In the ferruginous schists the periplutonic stage is manifested by the local development of the biotite, quartz and probably a certain portion of the skeletal grunerite is also developed there.

The Alpine metamorphic processes (M3) are shown in the ferruginous mineral assemblages mainly by chloritization, silicification, partly by sulfide mineralization, and carbonization etc. An intensive garnet cataclasis takes place locally in rheologically weaker domains, mostly in positions where quartz is abundant. In comparison with other areas of the southern Veporicum realm the Alpine thermal reworking in this area of the hybrid zone is not too significant. This is also indicated by the late Hercynian K/Ar and Rb/Sr ages of micas from the migmatitized gneisses near the Chorepa saddle (Cambel et al., 1986).

The study of the ferruginous rocks indicate that the metamorphic reactions of the M1 stage strongly depended on the original geochemical microdomains (there is no pronounced metamorphic segregation as in the common metapelites). Similarly, it can be inferred that the original redox-potential of the individual positions was not considerably changed, and it greatly determines the character of later recrystallization (for example the relationship of the ilmenite and magnetite or grunerite and magnetite). The chamosite is considered to be the most likely primary precursor of the garnet, as assumed by Korikovskij et al. (1989).

The grunerite precursor is less apparent, but it was probably generated in several ways. In the cases where grunerite (mostly the needle-like swarms) intergrow with quartz, in part of the porphyroblasts, eventually in overgrowths with magnetite (Figs. 3 and 4), the generation of grunerite through the reaction of siderite with quartz can be expected. (If this case happened, then the content of the quartz probably significantly exceeded the Fe-carbonate content, which is absent in the primary mineral assemblages). The grunerite, frequently in the skeletal crystals (Fig. 5) grown within the biotite mass, could also represent the minnesotaite-phases in a stilpnomelane matrix. Stilpnomelane is the most probable biotite precursor and since it has a lower content of Al_2O_3, than the biotite, the grunerite could be generated, along with a high content of Fe^{2+} in the system, as a byproduct of this transformation (Miyano and Klein, 1989). A similar situation can also take place in the case where there is a lack of aluminum and an abundance of iron (in the simplified reaction "chlorite + quartz = garnet"), compensated by the production of grunerite. However, in many cases and in accordance with actual observations (Beukes 1973, Klein 1982, Bayley and James 1973) also others, more complex reactions of several ferruginous reactants can be inferred.

The study case of the grunerite - garnet schists with - and without magnetite differs from the example of grunerite - garnet schists given by Korikovskij et al. (1989). Fig. 2 illustrates how the iron content of garnet and grunerite increase at the expense of the Mn and Mg contents in the sample with magnetite, whereas in the work of Korikovskij et al. (1989, see the sample HD-32 and HD-23 in Tabs. 1 and 2) it is to the contrary. The controversy of the both studies suggests that the presence/quantity of the magnetite is not generally related to the indicated geochemical parameters. However, we propone the idea that the presence of the magnetite is stimulated by the higher aluminum character of the protolith, where also the Fe^{3+} form is probably more abundant. In the case of the Kokava locality, the relationship between magnetite content and degree of metamorphosis has not been proven.

Generally, it seems that the main mass of the magnetite was created during the progressive period of M1, less during its peak stage (some of the marginal garnet zones do not contain magnetite) or in the process M2. It can be inferred that under those conditions the crystallization system was already freed of the "surplus" 3-valent iron. An influence of the potential reducing effects of the organic substance was not observed. The presence of magnetite in the rocks with graphite can be explained by the fact that a part of the magnetite content was already created in the diagenetic/early metamorphic stage and the gradual oxidation of the organic substance does not proceed from degradation of the magnetite. (James 1954, published several cases of diagenetic magnetite generation in the presence of decaying organic substance.) In the case of the ilmenite inclusions in garnet (our sample 16A or sample HD-23 taken from Korikovskij et al., 1989) we similarly infer its pre- to early-metamorphic origin, which indirectly points out the primary absence of the magnetite.

Because of the uncertain E_h, dependence upon the p_{H2O}, rock composition, type of geological buffer etc., the more exact *determination of p-T conditions* of the metamorphism in the ferrous formations remains only very informative (for example Fonarev, 1985). The stability field of the Fe-cummingtonite - grunerite has a wide temperature range, from 350 °C to about 760 °C; it is similar also with the determination of the pressure (Lattard and Evans 1992, Fonarev 1981, Miyano and Klein 1986). The bimineral exchange geothermometry cannot be used (Ferry and Spear 1978, Perčuk and Lavrenteva 1983) because these systems are not calibrated for the ferruginous protolith (mainly Fe-biotite causes too high temperature evaluations). There have not been identified ferrohyperstene or olivine in the studied rocks, and the preliminary outlining of the metamorphic conditions can be made only with use of the petrogenetic grid or by looking for an analog rocks of other regions. The reaction "grunerite = fayalite + quartz + H_2O" runs out at middle pressure conditions (about 4 - 8 kb) at 650 °C - 700 °C (Miyano and Klein 1986), or at about 570 °C - 600 °C (Fonarev 1985). The maximum thermal stability of the pure grunerite in the invariant point of the reaction "ferrosilite = fayalite + quartz" is set at 650 °C at the pressure of 9.7 kb (Lattard and Evans, 1992). At a pressure of 2 kb the grunerite is stable up to 550 °C and the temperature, along with the pressure, continuously increases to the mentioned invariant point (Lattard and Evans, 1992). However, from real and experimental systems it results that the grunerite stability increases with an increasing content of MgO. On the basis of several data (Fonarev 1985, Miyano and Klein 1986, Haase 1982, etc.) we assume that the investigated ferruginous rocks were metamorphosed under the conditions of the middle (up to higher) part of the amphibolite facies (i.e. about 600 ± 50 °C).

Similarly, as indicated by Korikovskij et al. (1989), the petrological reconnaissance of the surrounding rocks revealed a higher iron content in the rock forming minerals than is common on a regional scale. The preliminary data obtained from the almost unzoned garnets (the homogenization is probably caused by diffusion processes at temperature over 550 °C, Spear 1991) from the garnet - biotite gneisses showed that the conditions of the basic regional metamorphism (M1), according to the method of Perčuk and Lavrentevova (1983) and Ghent and Stout (1981), were about 675 - 715 °C and a pressure of about 5kb (Kováčik, 1998). This data falls into span of the metamorphic conditions in the southern Veporicum migmatites (Siman et al., 1996). However, the systematic abundance of muscovite and absence of K-feldspar and Al-silicates may indicate that metamorphic temperatures in these gneisses did generaly not exceed 650 °C (Chatterje, Johannes, 1974). Frequent discontinuous incremental zones in the garnet margin of the metapelites are considered the result of the thermal - mass effects of the acid granites (terminal part of M2), which, according to preliminary geothermometric calculations, show similar or a little bit lower thermal condition than the M1 stage.

Acknowledgement: I am thankful to Prof. RNDr. A. Vozárová, DrSc. for her careful review. The impulse for the writing of this work comes from RNDr. M. Petro and ing. M. Radvanec, CSc.

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Palaeogeography of the East-Slovakian Basin

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Abstract. The East-Slovakian Basin is a basin with complex tectonic history determined by oblique subduction of an oceanic slab occurring between the North-European Platform and the ALCAPA plate. The basin represents an autonomous part of the Transcarpathian Basin. It extends mostly on the Slovak territory and is filled by the Neogene clastics, volcanics, caustobioliths and evaporites. Genetic type and spatial distribution of deposits varied during the basin evolution depending on tectonics, volcanic activity, sea level changes as well as sediment input. Spatial distribution of deposits and their sedimentary environments are displayed in maps expressing ten time slices through the Eggenburgian, Karpatian, Badenian, Sarmatian, Late Miocene and Pliocene.

Key words: Neogene palaeogeography, tectonics, sediments, volcanics, pull-apart basin, sea-level changes, sedimentary environment

Introduction

The East-Slovakian Basin is an autonomous part of the Transcarpathian Basin with complex tectonic evolution and high variability in thickness and spatial distribution of sediments complicated by intrabasinal volcanics. Complex geologic evolution is a result of delicate interplay between intra- and extrabasinal processes such as tectonics, sediment input, depositional processes, climate and sea level fluctuation. They determined type of deposition, erosion and denudation, position of depocenters and basinal volcanism. Precise definition of basin fill stratigraphy, spatial distribution and type of deposits as well as type of tectonics comprise basis for unraveling basin history. This is the main role of palaeotectonic reconstructions which helps to understand the basin history.

The first palaeogeographic reconstruction of the East-Slovakian Basin was presented by Rudinec in 1978. The author updated this reconstruction in 1989 and in 1990 using also data from the Transcarpathian Ukraine. Recently, palaeogeography of the East-Slovakian Basin was presented by Kováč et al. (1996), Baráth et al. (1997) and Kováč & Zlínska (1998). The authors used a modern approach to analysis which is a contribution to the knowledge of the basin evolution. However, these reconstructions have often been schematic and lack more several important details.

Extensive and long-term research of the East-Slovakian Basin by the authors of this paper resulted in great amount of knowledge of sediments, tectonics, stratigraphy, palaeoclimate and palaeoecology. All these data provided sufficient database for a new palaeogeographic analysis of the basin. Maps showing ten time slices with thickness and spatial distribution of deposits, distribution of depocentres, erosional areas, volcano locations, character of volcanism and type of depositional environments during the Neogene are the main output of this analysis. The maps should serve not only for basic, but also for applied geological disciplines.

Initial data for the palaeogeographic reconstruction of ten time slices were obtained from deep boreholes. Based on this, sediment isopachs assigned to individual time slices were created. Borehole cores and drilling logs as well as surface outcrops provided information on lithology, facies types and sequences. This was complemented by analyses of reflection seismic profiles. Biostratigraphic subdivision of the basin fill has been based on foraminifera, nanoplankton, molluscs and ostracods.

Interpretation of the main structural elements – faults and overthrusts is based on the surface mapping and seismics. Fault characteristics were deducted from palaeostress fields which interpretation was based on the analysis of the brittle deformation. The sedimentation rate was interpreted on the base of the formation thicknesses. Sedimentological analyses were applied for definition of the geometry of sedimentary bodies and palaeoenvironmental analysis as well as for definition of the sediment input direction. All the data concerning the sediment input direction is related to present day coordinates. Radiometric ages of volcanics, which are barren on fossils, enabled their stratigraphic correlation with the biostratigraphically dated sediments. The palaeomagnetic data yielded basic information on the rotation of the lithospheric fragments underlying the basin fill.

Geological Survey of Slovak Republic, Dionýz Štúr Publishers, Bratislava 2000   ISSN 1335-96X

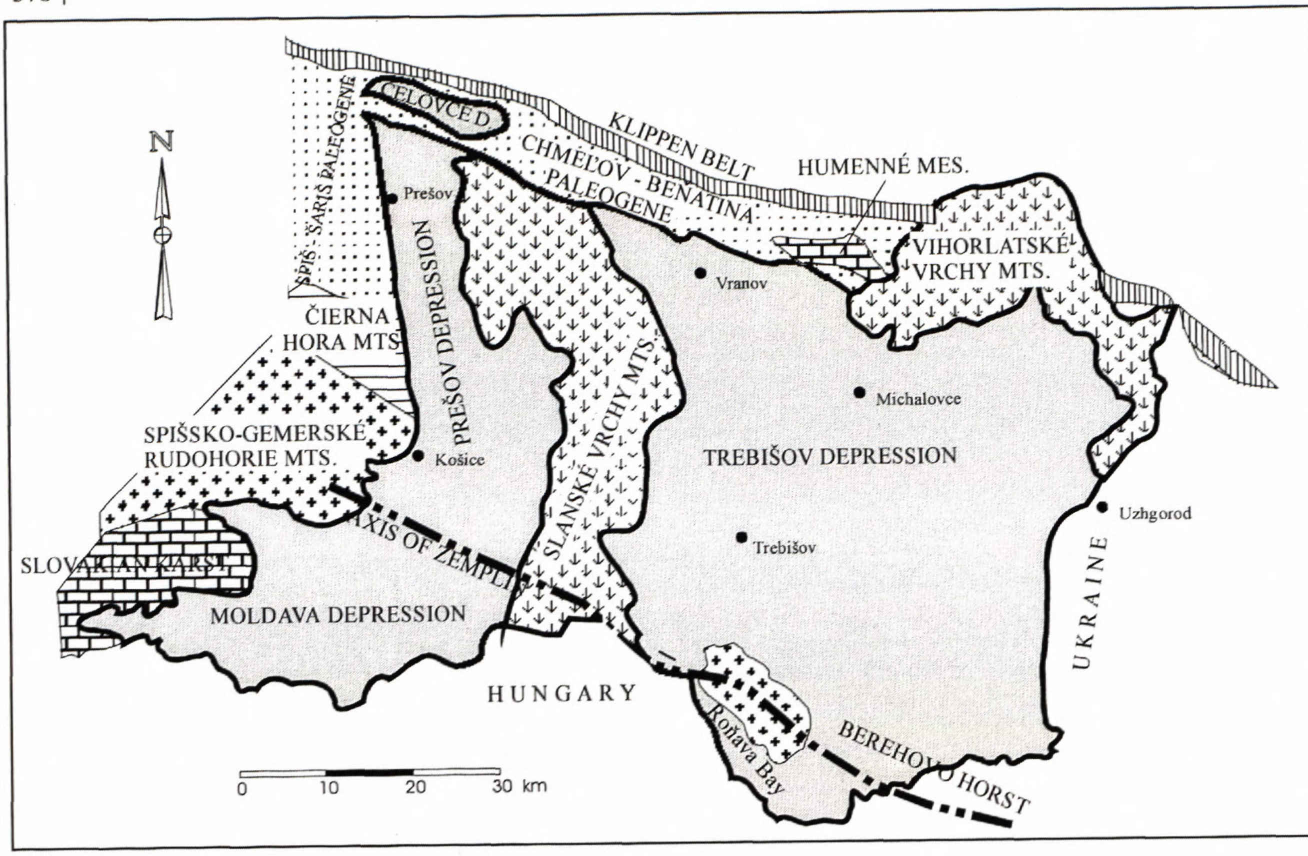

Fig. 1: Regional subdivision of the East Slovakian Basin. The basin includes subunits of Prešov Depression, Slanské vrchy Mts., Trebišov Depression, Roňava "Bay" and Vihorlatské vrchy Mts. Surrounding units are also shown.

Geological setting of the East-Slovakian Basin

The East-Slovakian Neogene Basin comprises western and autonomous part of the Transcarpathian Basin extending from Košice and Prešov in the west to the Uzhgorod in the east (Fig. 1). The basin is fault-bounded to the Chmeľov-Beňatina Paleogene unit in the north and separated by system of faults from the Šariš - Paleogene unit and pre-Tertiary rocks of the Sľubica, Čierna Hora and Spišsko-Gemerské Rudohorie Mts. in the west. To the south it is restricted by the Zemplin-Beregovo Horst separating the basin from the Nyírség Basin of the Great Hungarian Plain. The eastern boundary of the basin is expressed by the buried Seredne transverse horst (Vass et al. 1988, Rudinec 1989).

Striking neovolcanic morphostructure of the Slanské vrchy Mts . divides the basin into two parts – the Prešov Depression in the west continuing into Moldava Depression in the south and the Trebišov Depression in the east with Roňava "Bay" in the southeast (Fig. 1). Both the Moldava Depression and Roňava "Bay" are formally assigned to the East-Slovakian Neogene Basin (Vass et al. 1988) although they genetically belong to the Pannonian back-arc depression to the Nyírség Basin.

The kinematic history of the basin is complex. The basin originated in transpressional regime which later changed to compressional and transtensional. The most important periods of the basin development are connected with a pull-appart regime (Vass et al. 1988). The basin lies on thinned, about 27 km thick crust thickening toward north and north-west where it reaches about 30 km (Šefara et al. 1987). The lithosphere thickness is 80 km (Babuška et al. 1986). Several gravity-magnetic anomalies occur in the basin, from which the most conspicuous are the Sečovce anomaly, anomaly nearby the village Zbudza and anomaly in the Moldava depression. All anomalies are probably induced by bodies of ultrabasic rocks although the opinion on their origin is still ambiguous (e.g. Hovorka et al. 1985, Vass et al. 1988, Gnojek et al. 1991, Šutora et al. 1990, Soták et al. 1993).

Pre-Neogene basement of the basin

Pre-Neogene basement of the basin has a complicated structure composed of several units or superunits (Fig. 3). The complicated structure and its variability are related to the mechanism of the basin opening (pull-apart), when horizontal translations along basin generating faults („dismatch in basement„ of Christie-Blick & Biddle, 1985) resulted in convergence of different geologic units. Today's arrangment of the basement units located in the East-Slovakian Neogene Basin has a NW-SE trend consistent with the trend of the eastern part of the Western Carpathian belt (e.g. Fusán et al. 1972, 1987, Slávik 1974, Rudinec 1978, Ďurica 1982), but perpendicular to the structural trend of the adjacent regions of

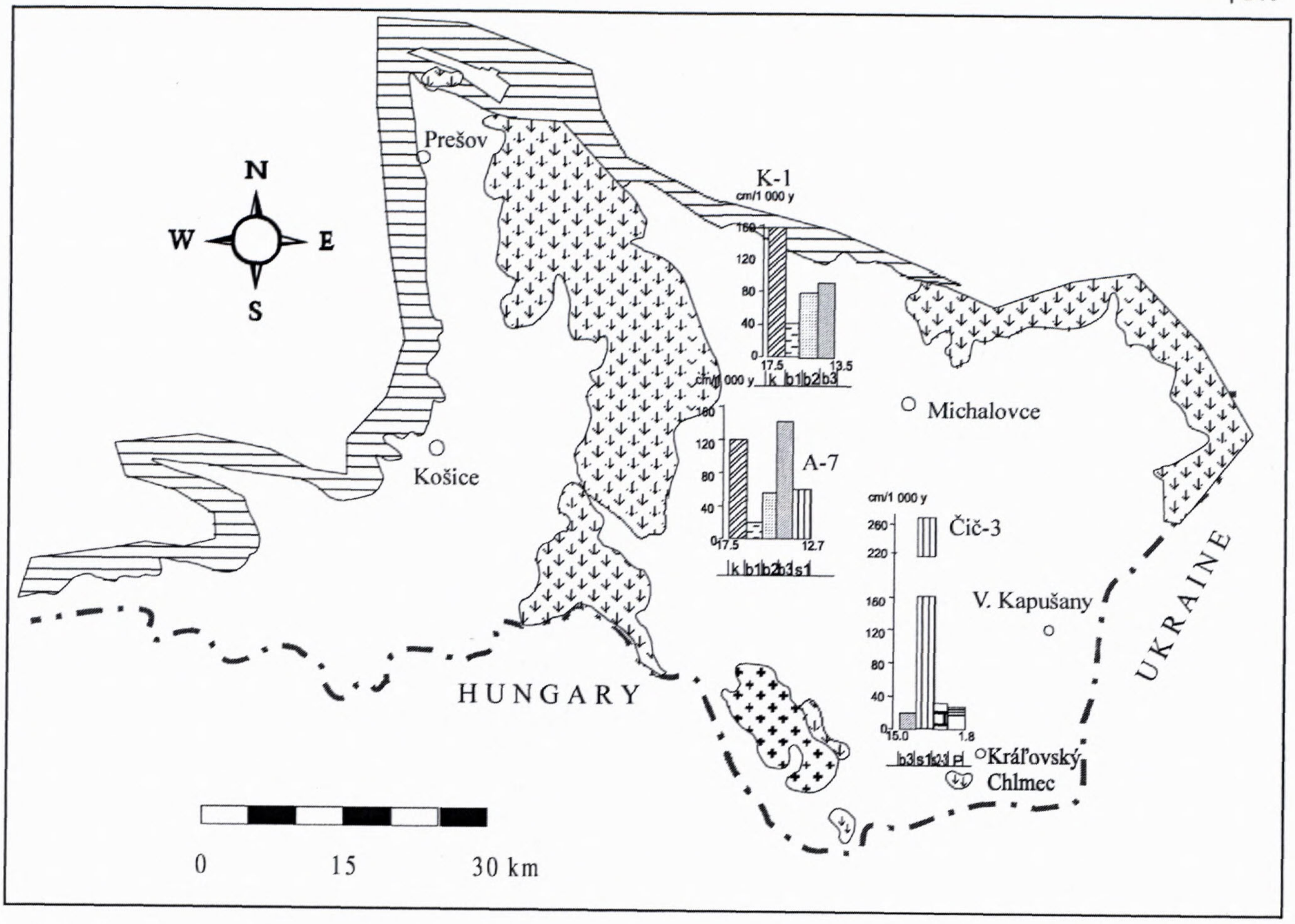

Fig. 2: Migration of subsidence in the East Slovakian Basin. Maximum subsidence of Karpathian stage (k., NW part of the basin), Late Badenian (b_3 central part of the basin), Early Sarmatian (s_1, SE part of the basin). Sedimentary rates calculated according to the data taken from wells KD-1, A-7 and Čič-3. The sediment thickness is considered as decompacted. Abbrevation of chronostratigraphic Central Paratethys Neogene stages: k - Karpatian, $b_{1,2,3}$ - Lower, Middle and Upper Badenian, s_1 - Lower Sarmatian, S_{2-3} - Middle and Upper Sarmatian, white field - Pannonian, Pontian and Pliocene.

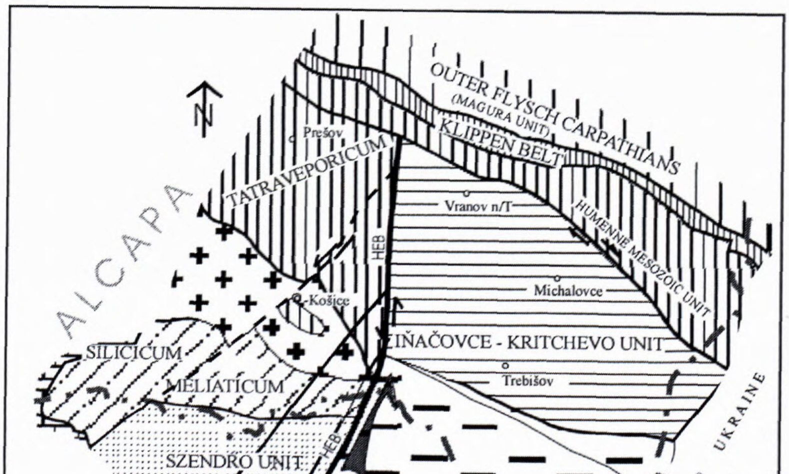

Fig. 3: Pre-Tertiary units of the East-Slovakian basement and its surroundings. The contact between the ALCAPA and Tissia Units is interpreted along the Hornád Fault Belt (HFB)

Hungary (Fülöp & Dank eds., 1987, Dank & Fülöp eds., 1990). This structural arrangement resulted from the Miocene CCW rotation.

The Magura Unit of the Outer Flysch Carpathians and Pieniny Klippen Belt are the northernmost units where the modern NE margin of the basin encroaches. To the south of the Pieniny Klippen Belt the Humenné Mesozoic Unit extends. The basement of the central and southern part of the basin is built by the Pozdišovce-Iňačovce Unit (Iňačovce - Kritchevo Unit), Zemplinicum and Ptrukša Units which belong by our opinion to Tissia. The Prešov Depression and a part of the Slanské vrchy Mts. are underlain by crystalline rocks and by the Mesozoic formations of the Čierna Hora Mts. On Fig. 3 both the Čierna Hora and Humenné Unit are lumped together in the Tatraveporicum. Some authors assume that the Choč Nappe underlies the northern part of the Prešov Depression (e.g.

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Ďurica 1982). The Moldava Depression is underlain by the Paleozoic rocks of Gemericum and by the Mesozoic rocks of Meliaticum, Silicicum (Fig. 3) and eventually Turnaicum. We assigne all these units to the ALCAPA microplate (e.g. Kováč et al. 1999, Fig. 3). The Central-Carpathian Paleogene rocks, do not shown on Fig. 3, represent the youngest unit of the pre-Neogene basement. It underlies the northern part of the basin and it partly covers some of the before mentioned pre-Tertiary units.

Neogene basin fill

The East-Slovakian Basin is a basin with complex tectonic history. Vass (1998) considers the basin as a fore-arc and intra-arc one in relation to volcanic arc generated as the consequence of the North European Platform subduction beneath the ALCAPA Plate. Kováč et al. (1995) characterizes the basin as a back-arc basin. However, the clear evidence about the basin position in relation to volcanic arc is proved only from the Late Badenian when andesite volcanism directly related to the subduction appeared (Lexa & Konečný, 1998). We think that the basin tectonic evolution before the Late Badenian was related mainly to basin position behind the accretionary prism and oblique collision between the North European and ALCAPA plate.

The fill of the basin consists of sediments and volcanics stratigraphicaly ranging from the Eggenburgian to the Pliocene. The maximum thickness of the fill is up to 8 - 9 km. The depocenters migrated in time from NW to SE. In the NW part of the basin the subsidence culminated in the Karpatian stage, in the central part of the basin it culminated during the Late Badenian and finally, during the Early Sarmatian the highest amplitude of subsidence occurred in the SE part of the basin (Fig. 2). Prevailing deposits are composed of siliciclastics, caustobioliths (coal and lignite) and minor evaporite deposits. An important portion of the basin fill is composed of volcanic rocks (mainly volcaniclastics and effusive rocks). They are acid (absolutely prevailing from the Eggenburgian to the Early Badenian) and intermediate (from the Badenian to the Late Sarmatian and/or Pannonian) The volcanic activity culminated in the Badenian and Sarmatian (Tab. 1).

The basin fill is of horst-and-graben style which resulted from mainly syndepositional fault activity. The value of some fault system throws exceeds 1 000 m. Besides the fault structure, the basin fill is also slightly folded. On the N and S margin of the basin the beds dip toward the centre in angle from 30° to 50°. Overthrusts in older deposits are indicated on seismic sections (Keith et al. 1989, Magyar et al. 1997, Mořkovský et al. 1999) and occasionally also on surface (NE margin of the basin). We even found folded Pannonian deposits at some locations. Many listric faults of the basement originally represented thrust planes or reverse faults. Later, during the opening stage of the basin, they became normal faults.

Palaeogeography

Eggenburgian (23.0 - 19.0 Ma, Map 1)

The first Neogene marine transgression into the area of the nowadays East-Slovakian Basin occurred during the Eggenburgian (23.0 – 19.0 Ma). The Eggenburgian deposits crop out in the narrow belt along the northern basin margin (Map 1). The belt continues in Zakarpatie (Transcarpathia) along the northern margin of the Zakarpatie (Transcarpathian) Depression (Burkalovo Formation; Vialov in Muratov & Neveskaja 1986, Rudinec

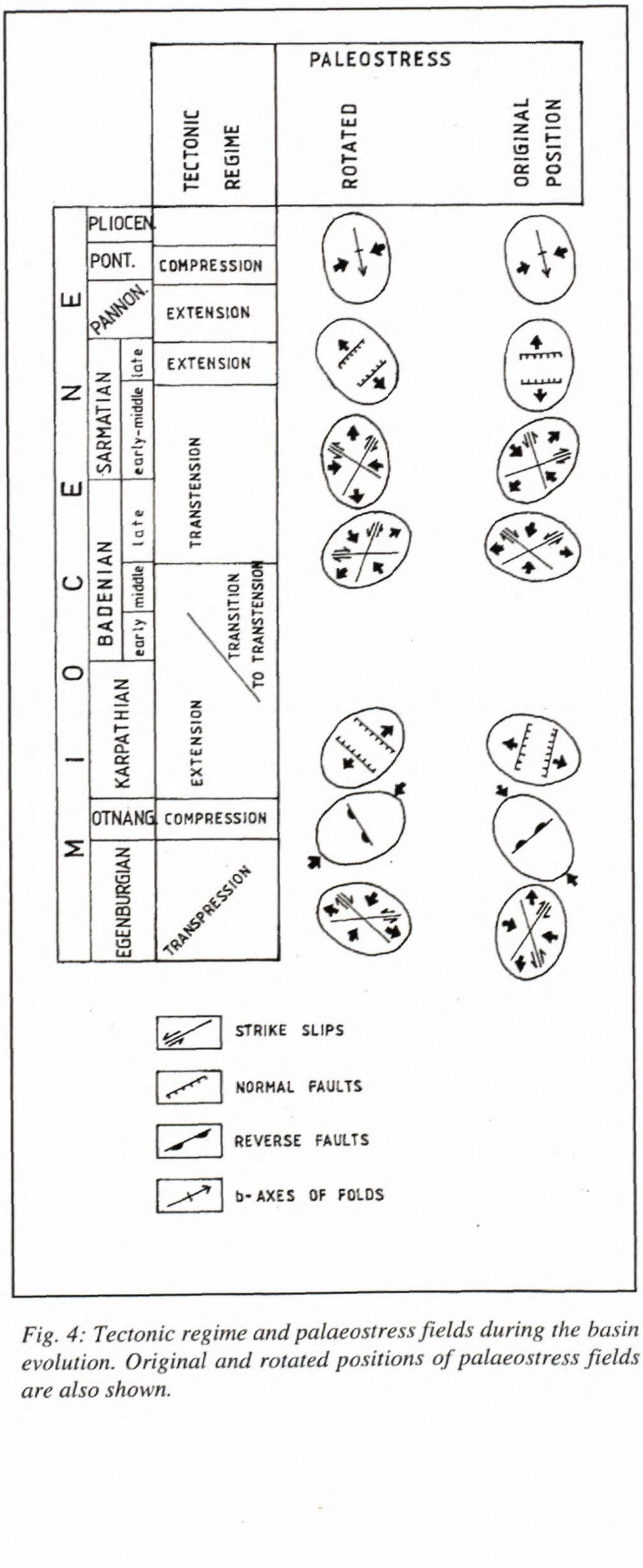

Fig. 4: Tectonic regime and palaeostress fields during the basin evolution. Original and rotated positions of palaeostress fields are also shown.

1989, Petraskiewicz & Lozinjak 1995, Andreyeva-Grigorevich et al. 1997).

The appearance of Eggenburgian deposits in a narrow belt was the principal reason to classify the Eggenburgian basin of East Slovakia as a wrench furrow (Vass, 1998). Recent study of the smectite expandability on the samples coming from the Central Carpathian Paleogene of the Northern Slovakia (Levočské vrchy Mts., region of Orava) enables to estimate the thickness of sediments removed by erosion from the area up to 3–4.6 km and the major part of missing deposits had to be Early Miocene in age (Uhlík, 1999). It seems, the Eggenburgian marine deposits in Northern and Eastern Slovakia were not bounded to a narrow belt, but they originated in a large intensively subsided basin of the fore-arc position.

The subsidence and/or opening of the Eggenburgian basin resulted from an oblique convergence in the subduction zone of the Outer Western Carpathians. Before or during the Eggenburgian culminated lateral escape of the Tissia partial units into the East-Slovakian area (Wein 1969, Grecula & Együd 1977). These units are represented by the Zemplinicum, Ptrukša Unit and Iňačovce-Kritchevo Unit. The escape occurred along sinistral strike-slip faults of the Hornád Fault System (Fig. 3).

Palaeostress interpreted from the brittle deformations would generate dextral and conjugate sinistral strike-slip faults. For the basin the dextral strike-slips played an important role (see Map 1 and Fig. 4). Some of them caused a lateral translation of the Humenné Mesozoic Unit as a partial unit of the Tatraveporicum to the east (see Fig. 3). A distinct seismic anisotropy and steep fault separating the Iňačovce-Kritchevo Unit and the Humenné Mesozoic rocks are well documented by the seismic line 612/88 (Fig. 7). The basin is also disturbed by NE - SW faults. They seem to be slightly younger, perhaps epigenetic normal faults (Map 1).

The opening and subsidence of the East-Slovakian wrench furrow ceased at the end of the Eggenburgian. It was gradually filled by progradational deltas.

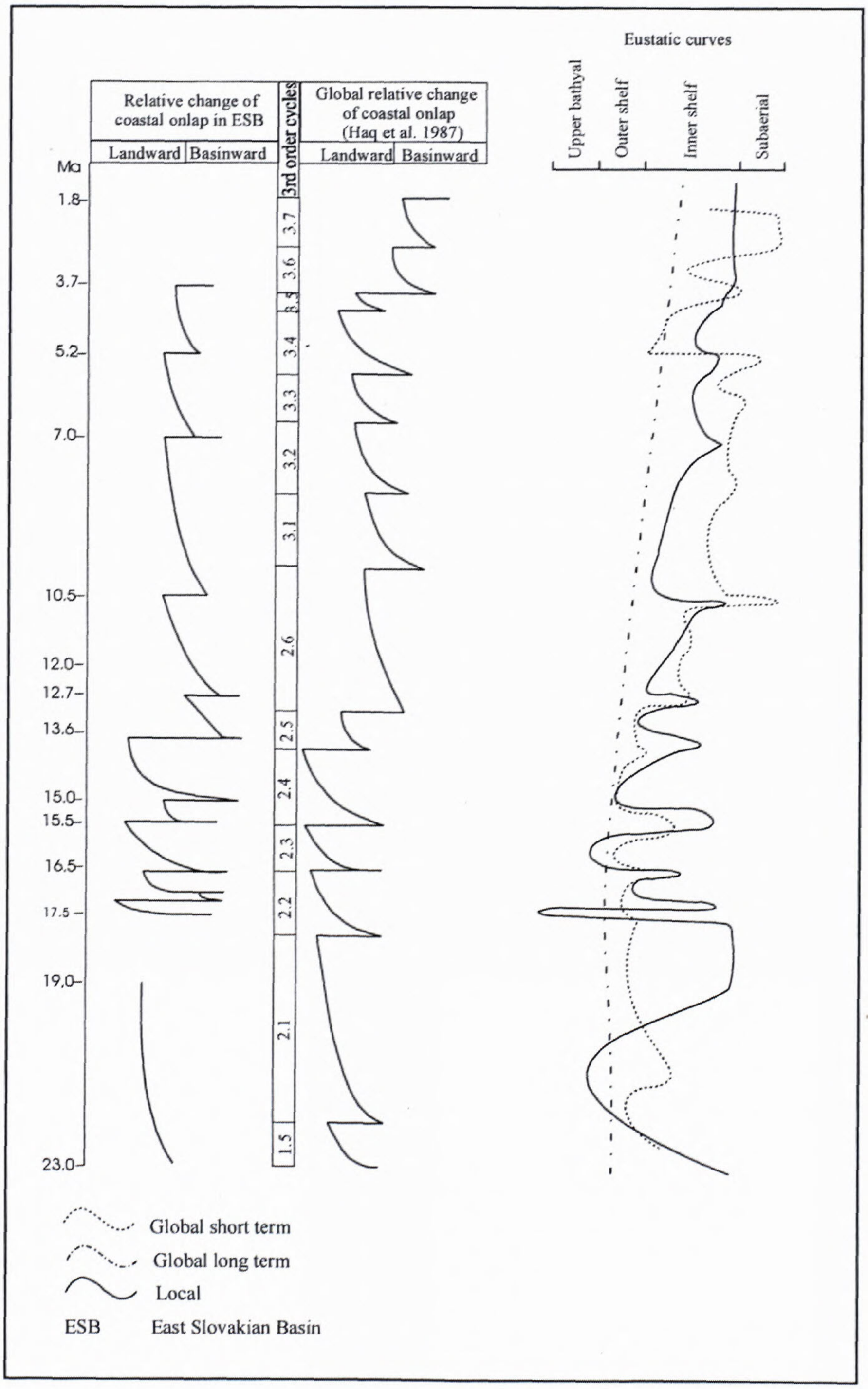

Fig. 5: Correlation between local and global sea level fluctuations

The Eggenburgian deposits of the East-Slovakian Basin have a transgressive character. The transgression is believed to be strengthened by the eustatic sea level rise (Fig. 5). The transgression occurred over the partially emerged Outer Flysch Carpathians e.g. over the Magura, Dukla, Silesian and Subsilesian Units. The Eggenburgian basin communicates over these units with both the Carpathian foredeep and original sedimentary areas of the Skole, Boryslav-Pokuty and Stebnik Units later incorporated into the frontal part of the Flysch Carpathians today.

The area of Eggenburgian sedimentation in East Slovakian Basin itself as it is shown on Map 1, was directly

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EAST SLOVAKIAN BASIN
Legend

ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA

Areas of erosion and denudation

Flat relief

Sharp relief

Areas of sedimentation

Continental

Alluvial

Lacustrine

Swamp

Brackish

Lagoon

Delta

Nearshore (Onshore)

Offshore

Marine

Inner shelf

Outer shelf

Hypersaline

b) Hypersaline sandwiched by marine

Volcanics

Andesite, b) Volcanoclastics

Rhyolite, Rhyodacite, b) Volcanoclastics

Stratovolcano, b) Volcanic extrusion

Lithotypes

Conglomerate, gravel

Sand, sandstone

Silt, siltstone including schlier

Clay, claystone

Evaporite

Coal and lignite

Sand covered by clay

Conglomerate and sandstone covered by
a) claystone, mudstone
b) and evaporite

Other symbols

Synsedimentary fault active during presented time period

Other significant fault

Strike slip

Recent thickness isoline

Presumable direction of sediment input

Presumable direction of sediment transport

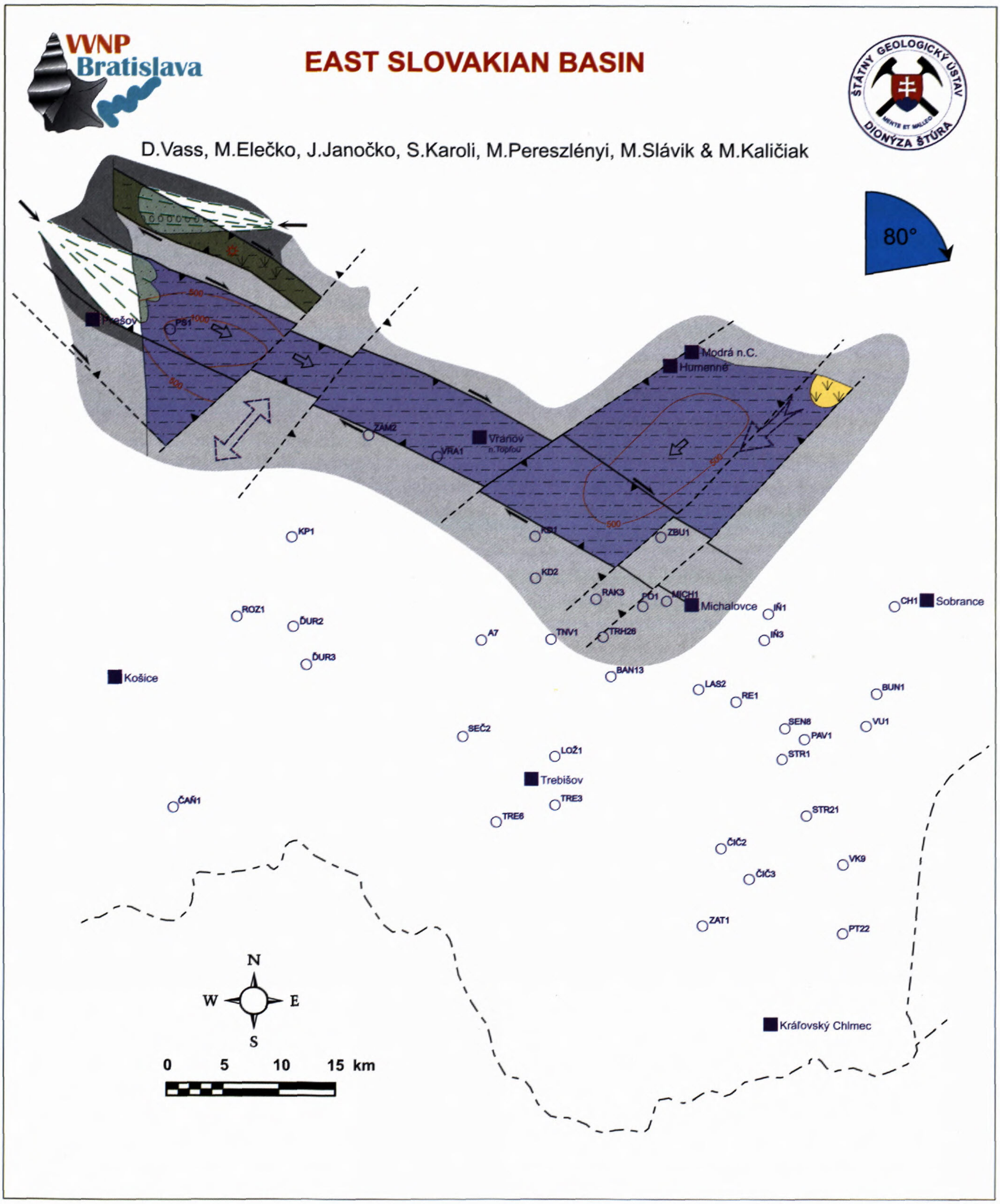
Sea way and/or direction of sea transgression (direction is not object of rotation !)

Inlet and direction of hyposaline sea transgression (direction is not object of rotation !)

Delta contour

Important borehole

Direction and angle of rotation to original position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EAST SLOVAKIAN BASIN
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
D.Vass, M.Elečko, J.Janočko, S.Karoli, M.Pereszlényi, M.Slávik & M.Kaličiak
80°
Prešov
Modrá n.C.
Humenné
Vranov n.Topľou
Michalovce
Sobrance
Košice
Trebišov
Kráľovský Chlmec
N
W
E
S
0 5 10 15 km

connected with sedimentary areas of former Central-Carpathian Paleogene including the pre-Sarmatian Orava Basin (Czieszkowski, 1992).

The connection probably stretched further toward the W over the Inner Depressions of the West Carpathians where the Eggenburgian deposits are preserved (Turiec, Bánovce, Ilava, Trenčín Depressions) to the Vienna Basin and to the Carpathian Foredeep in the SW Moravia. This connection was already considered by Buday et al. (1967).

The basin was also opened to the SW where a sea communication with the Fiľakovo/Pétervására Basin and/or a bay (Sztanó 1994, Halásová et al. 1996) existed.

The sedimentation during the Eggenburgian occurred mainly in the marine environment. The basal transgressive clastics are overlain by pelitic deposits which, in turn, are capped by regressive deltaic and lagoonal deposits.

The Eggenburgian deposits are composed of two formations:

Prešov Formation (Fig. 6) unconformably lies on the Eggerian and older Paleogene and pre-Tertiary rocks and crop out in the NW surroundings of Prešov. It was also penetrated by boreholes beneath younger rocks nearby Vranov. The maximum thickness of the formation is 1 000 m. It consists of basal medium- and coarse-grained polymic, mainly carbonate and quartz conglomerate probably originated in southeastward trending deltaic system (see Map 1). The conglomerate is overlain by gray, calcareous siltstones containing coal detritus and fine-grained calcareous, wacke sandstones (Karoli in Kaličiak et al. 1991) and fine grained conglomerates. The formation also contains tuff and redeposited tuff , which are mostly seladonized and bentonized. They probably originated during the first Early Miocene volcanic explosions in the Eastern Slovakia with center northward of Prešov (Kaličiak et al. 1991).

The Prešov Formation is rich in marine fauna with numerous molluscs (*Pitaria cf. lilacinoides, Cardium cf. moescheanum, Pecten cf. burdigalensis;* e.g. Švagrovský 1952) and relatively poor foraminifera assemblage with prevailing individuals of *Lenticulina* genus (*L. inornata, L. cultrata*, Cicha and Kheil 1962), *L. arcuatostriata*, and *Spiroplectinella carinata* as well as *Pappina bononiensis primiformis* (Zlinská, 1992).

Čelovce Formation (Fig. 6), occurring in the partial Čelovce Depression (Fig. 1), is composed of desintegrated, thick-bedded sandstone containing calcareous, variegated and dark coal clay and thin beds of brown glance coal. Coarse-grained, polymic and fine-grained, quartzite conglomerate intermittently forms small lenses in the formation. The maximum thickness of the formation is about 300 – 500 m. The lower part of the Čelovce depression fill prevailingly contains marine sublittoral and neritic microfauna (*Lenticulina meznericsae, Uvigerina hantkeni, Cibicides budayi, Planulina wuelerstorfi, Ammonia beccarii, Porosononion subgranosum* and others, Cicha and Kheil 1962). The brackish fauna is represented by assemblage of *Ostrea cf. cythula, Polymesoda brogniarti, Congeria basteroti, Pirenella hornensis* (Volfová 1959).

Nearby village of Modrá n/Cirochou E of Humenné a small erosive relic of the Eggenburgian deposits, overlying the Klippen belt and the Outer-Carpathian Flysch, occurs. It consists of gray and dark gray crumbled claystones and siltstones containing lenses and seams of coal (Vass and Elečko 1988). Besides allochthonous foraminifera they also contain autochthonous species *Tenuitellinota angustiumbilicata, Globoturborotalalia rotalita connecta , Globigerina lentiana, Almaea osnabrugensis, Cyclamina praecancelata* suggesting the Eggenburgian age of the deposits (Zlinská 1995). The autochthonous fauna reveals deteriorated conditions, most probably in the coastal lagoon with restricted communication with the open sea. The occurrence of coal seams suggests coastal swamp or marsh environment.

Ottnangian (19.0 - 17.5 Ma)

During the Ottnangian the change of palaeostress field resulted in emerging of the area of the future East-Slovakian Neogene Basin above sea level. The pressure generating shear was changed by a pure compression acting in the SW-NE direction resulting in uplift of the area and depositional hiatus (Janočko & Jacko 2 000). In such a palaeostress condition dextral translation along Periklippen Fault Belt (comp. Baráth et al. 1997) and along the faults separating Iňačovce-Kritchevo Unit and Humenné Mesozoic rocks would be active.

Karpatian (17.5 – 16.5 Ma)

At the beginning of the Karpatian stage a new depositional area opened in the region of the East-Slovakian Basin. During that time a marine transgression, corresponding to the cycle TB2.2 of the global eustatic sea level (Haq. et al. 1987) occurs in the area. The brittle deformations suggest extension in NE - SW direction (Fig. 4) and the basin was opened by normal faults parallel to the recent basin axis. The extension was most probably a result of the upheaval of the Pannonian asthenosphere (e.g. Vass 1995). The faults were often inherited from the older tectonic structure of the upper crust. The original thrust planes along which the Zemplinicum and the Ptrukša Units were overthrusted, were reactivated and changed to listric faults. Similarly, the thrust planes along which the Ptrukša Unit was overthrusted on the Iňačovce-Kričevo Unit, were reactivated (Fig. 8). The subsidence of blocks along these discontinuities resulted in extensional opening of the East-Slovakian Basin.

A marine transgression into the opening basin occurred mainly from the NE, i.e. from the basins occurring in the front of the Flysch Carpathians (from the original sedimentary basins of recent tectonic units of the Carpathian front: Skole, Stebnik and Boryslav-Pokuty units; Oszczypko 1997). Sea might also penetrate from the NW direction where Czieszkowski (1992) has suggested a marine basin in the area of Orava.

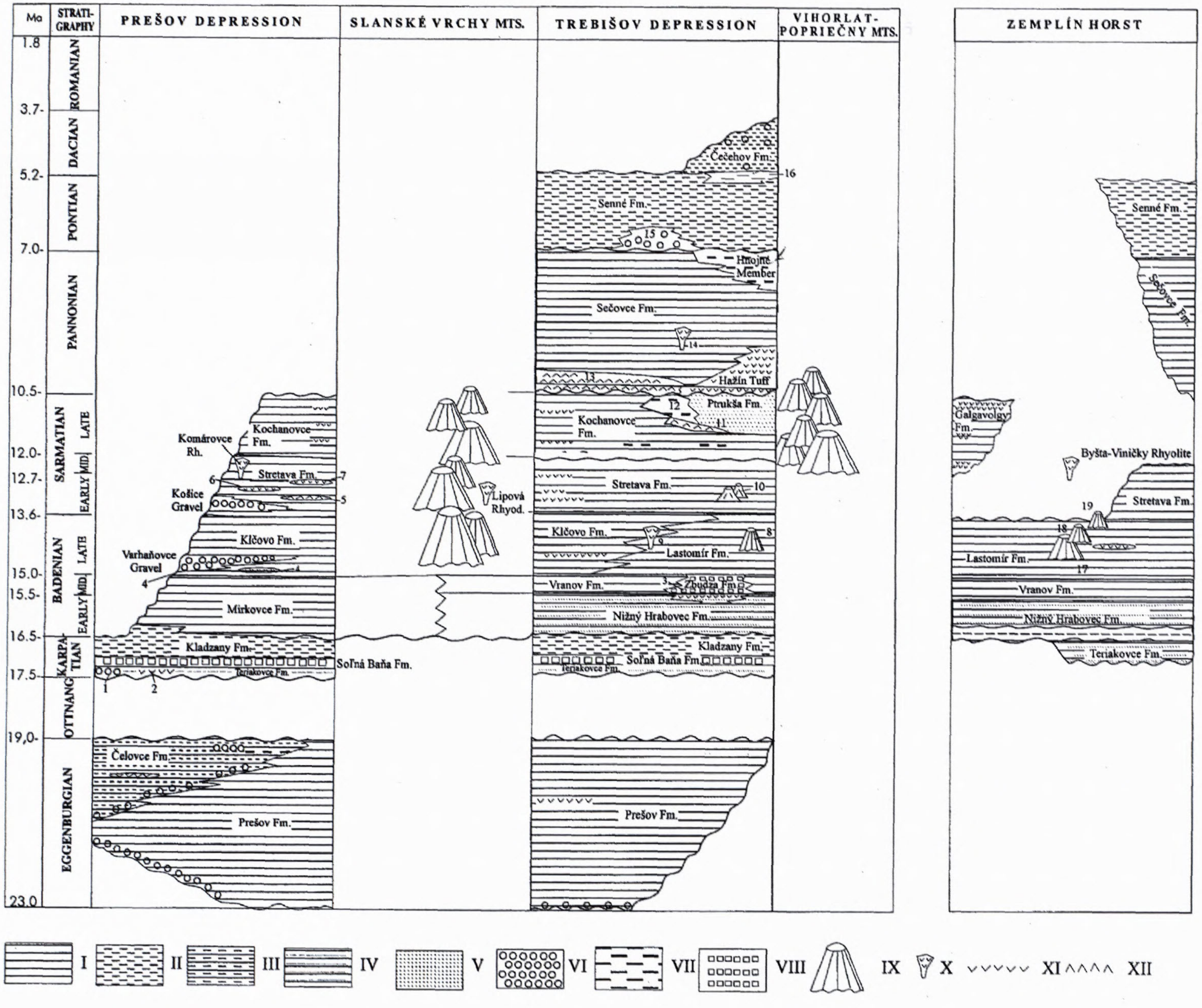


Fig. 6 Lithostratigraphy of the East-Slovakian Basin fill
Explanation:
Legende numbers:
I – calcareous siltstone and claystone; II - multicolored clay/ claystone; III – multicolored variagated and grey siltstone and claystone; IV – sandstone and siltstone; V – sand/ sandstone; VI – conglomerate/ gravel; VII – coal seams; VIII – evaporites; IX – stratovolcano (andesite); X – rhyolite and rhyodacite extrusive body; XI – rhyolite tuff; XII – andesite tuff
Lithostratigraphic units numbers: 1. Lemešany conglomerate, 2. Fintice tuff, 3. Hrabovec tuff, 4. Kráľovce tuff, 5. Olšava Mb., 6. Myšla Mb., 7. Rankovce tuff, 8. Zatin andesite, 9. Lesné rhyolite, 10. Malčice, Beša and Čičarovce stratovolcanos, 11. Lúčky volcaniclastics, 12. Závadka Mb., 13. Albinov tuff, 14. Michalovce rhyolite, 15. Pozdišovce gravel/conglomerate, 16. Iňačovce Mb., 17. Kráľovsky Chlmec stratovolcano, 18. Somotor stratovolcano, 19. Brehov stratovolcano

The Karpatian deposits consist of three formations:

The basal, **Teriakovce Formationn** (Fig. 6), mostly consists of calcareous gray sandstone and claystone (flysch-like sequence) croping out in the NW part of the basin. Conglomer ate, occurring between Prešov and Ďurkov east of Košice (Map 2), comprises the lowermost part of the formation (Lemešany Conglomerate, Karoli in Kaličiak et al. 1991). Genetically the conglomerate represents a wide spectrum of depositional environments varying from fluvial through deltaic and shallow-marine conglomerates. The maximum thickness of the formation is about 250 – 400 m. The formation contains marine fauna *Uvigerina graciliformis, U. parkeri breviformis, U. bononiensis primiformis* (e.g. Kantorová in Kantor & Kantorová 1955, Cicha & Kheil 1962, Zapletalová 1970), *Brissopsis ottnangensis, Ammussium cristatum badense, Macoma elliptica, Lima cf. lebani*; (e.g. Seneš 1955). The calcareous nanoflora of the NN4 zone was also described in the surroundings of Hlinné (Lehotayová 1982). The foraminifera assemblage suggests outer shelf environment (Zlinská 1992, Kováč & Zlinská 1998).

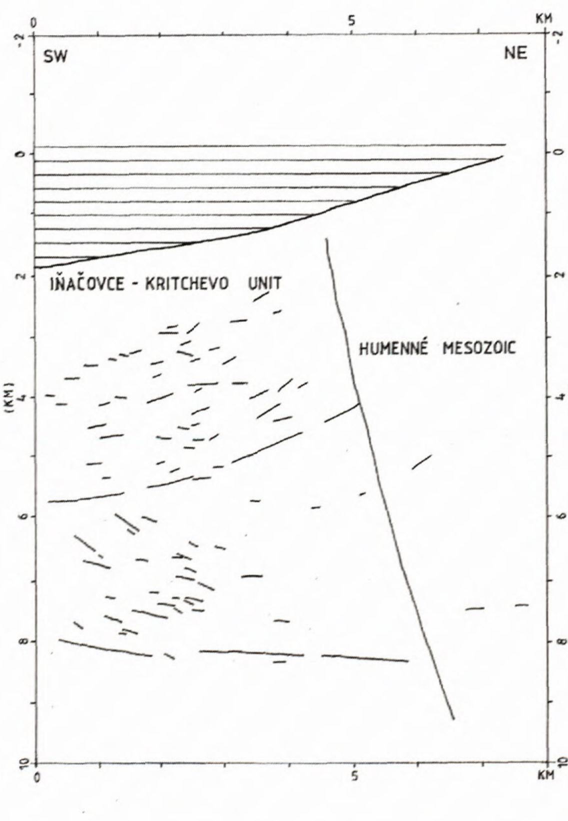


Fig. 7: Deep seismic profile 612/88 showing distinct seismic anisotropy between the Iňačovce - Kritchevo Unit (strong reflections) and the Humenné Mesozoic (no reflection). After Tomek in Vozárová et al. 1993).

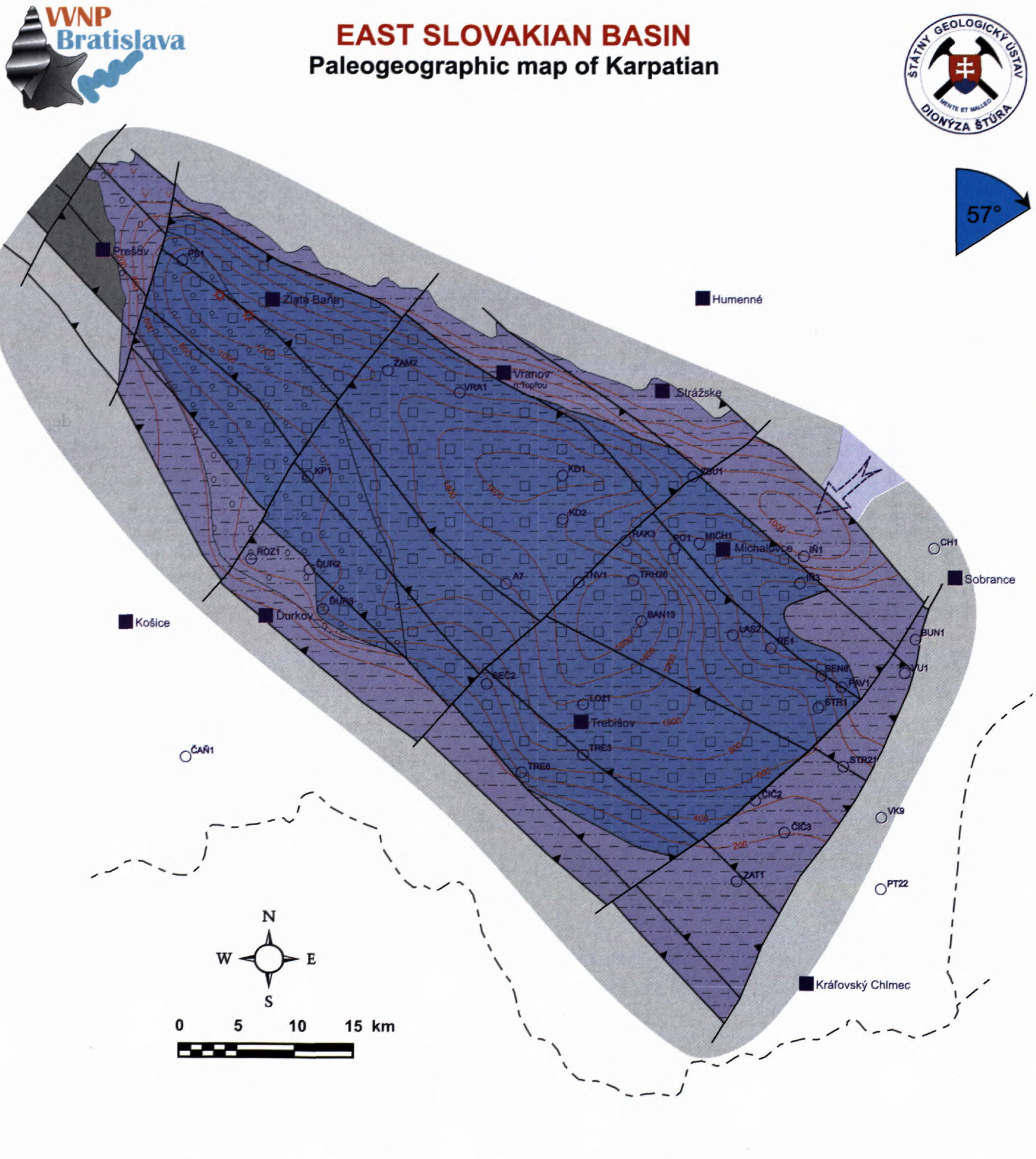
Soľná Baňa Formation overlies Teriakovce Formation (Fig. 6). The decreasing of subsidence during that time resulted in semiclosed environment and prevailing deposition in hypersaline conditions. The formation consists of mudstone with gypsum laminae and sporadic nodular anhydrite at the base, which pass into salt breccia (mudstone and sandstone clasts floating in halite matrix). The uppermost part is again composed of prevailing mudstone deposits containing minor layers of salt breccia (Karoli 1998). The type of deposits and abrupt dissapearance of the foraminifera fauna suggest prevailing evaporite deposition in mud flat and salt pan environments. Maximum thickness of the deposits is 320 m. The formation crop out in the Prešov surroundings but it was were also documented in deep boreholes in the central part of the East Slovakian Basin (Map 2).

The occurrence of fauna in the formation is poor. Foraminifera assemblage composed of species *Florilus, Elphidium, Melonis*, is apparently influenced by hypersalinity of the environment. Typical Early Miocene species, e.g. *Uvigerina parkeri breviformis* and typical Karpathian species *U. graciliformis* (Kantorová 1954, Cicha & Kheil 1962) also occur sporadically.

The deposits of the Late Karpatian **Kladzany Formation** gradually evolved from the Soľná Baňa Formation, or they laterally pass into the Soľná Baňa Formation (Fig. 6). They consist of variegated calcareous claystone and clay with thin beds of sandstone. The structures of deposits together with the occurrence of anhydrite concretions, secondary fibrous anhydrite and fibrous halite filling fissures suggest shallow water deposition. The rocks crop out between Vranov and Strážske and in the Prešov Depression. The sediment almost lacks organic material. Occasionally marine foraminifera, including typical Karpatian species *Uvigerina graciliformis*, and *U. bononiensis compressa, U. acumianta, Bolivina hebes, Cibicides ungarianus*, (Danihelová 1954, Zapletalová 1974, Zlinská in Kaličiak et al. 1991, Holcová in Vass et al. 1996) were found. Foraminifera assemblages are not uniform and reflect local changes of coastal onlaps of the 4th and 5th order (Holcová in Vass et al. 1996, Fig. 9). According to the deep boreholes the formation, which max. thickness is about 1 300 m, occurs almost in the whole basin.

During the Late Karpatian a new tectonic regime appeared which was controlled by stress field formed by renewed activity in the subduction zone of the Carpathians (e.g. Kováč et al. 1994). There is no direct evidence on spatial orientation of the main direction of compression during the Late Karpatian but we asumme that the palaeostress recorded by brittle deformations of Early Badenian rocks (Kováč et al. 1994) was identical with the stress applied in the East Slovakian Basin during the Late Karpatian. The regime is suggested mainly by a very rapid, pull-apart type subsidence resulting in deposition of more than 1 000 m deposits of the Kladzany Formation in the NW corner of the basin (Fig. 2). In the palaeostress field with maximum compression in NW - SE direction the NW normal faults turn to dextral strike-slips and normal faults of NE direction were active as sinistral strike-slips. The dextral translation along Periklippen Fault Belt and faults separating the Humenné Mesozoic rocks and the Iňačovce-Kričevo Unit has been reactivated (see Fig. 4). These faults have been disturbed by sinistral strike-slips of NE - SW direction (Map 3).

EAST SLOVAKIAN BASIN
Paleogeographic map of Karpatian



Slovak Geol. Mag., 6, 4(2000), 377–407

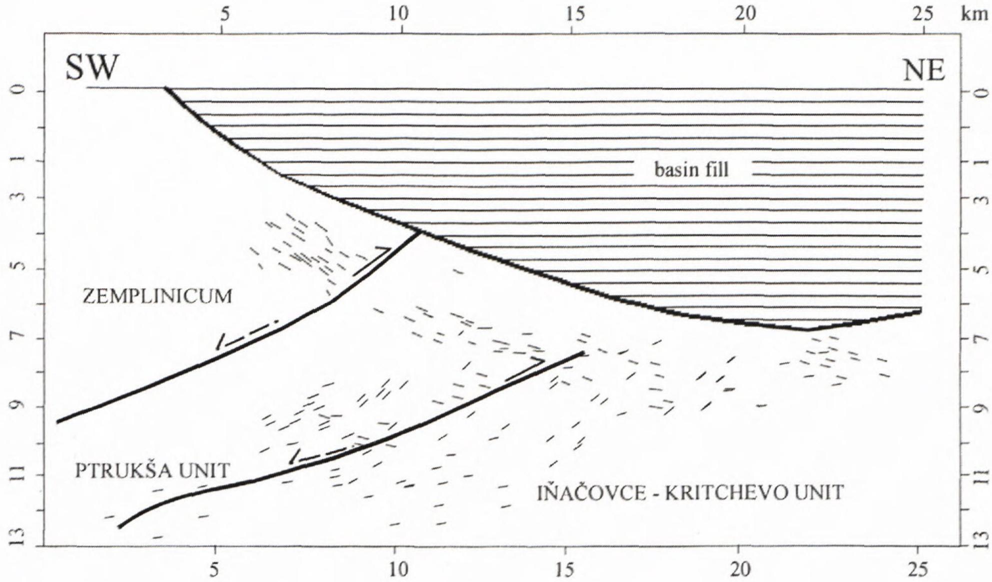


Fig. 8: Deep seismic profiles 597 and 597A/86. As a result of crustal spreading, thrust planes turned into listric normal faults. After Tomek in Vozárová et al. (1993).

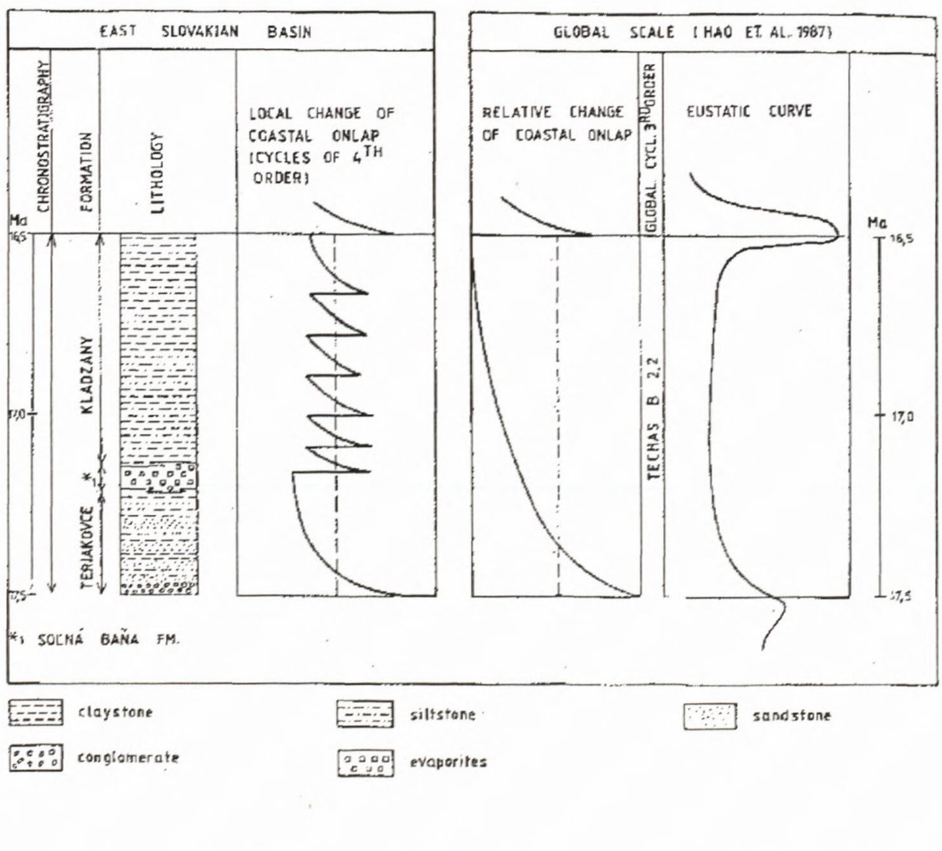


Fig. 9: 4^{th} or 5^{th} order cycles of the sea level fluctuation in the East Slovakian Basin during deposition of the Kladzany Formation, Late Karpatian and comparison with the global scale of sea level fluctuation. After Holcová in Vass et a. 1996).

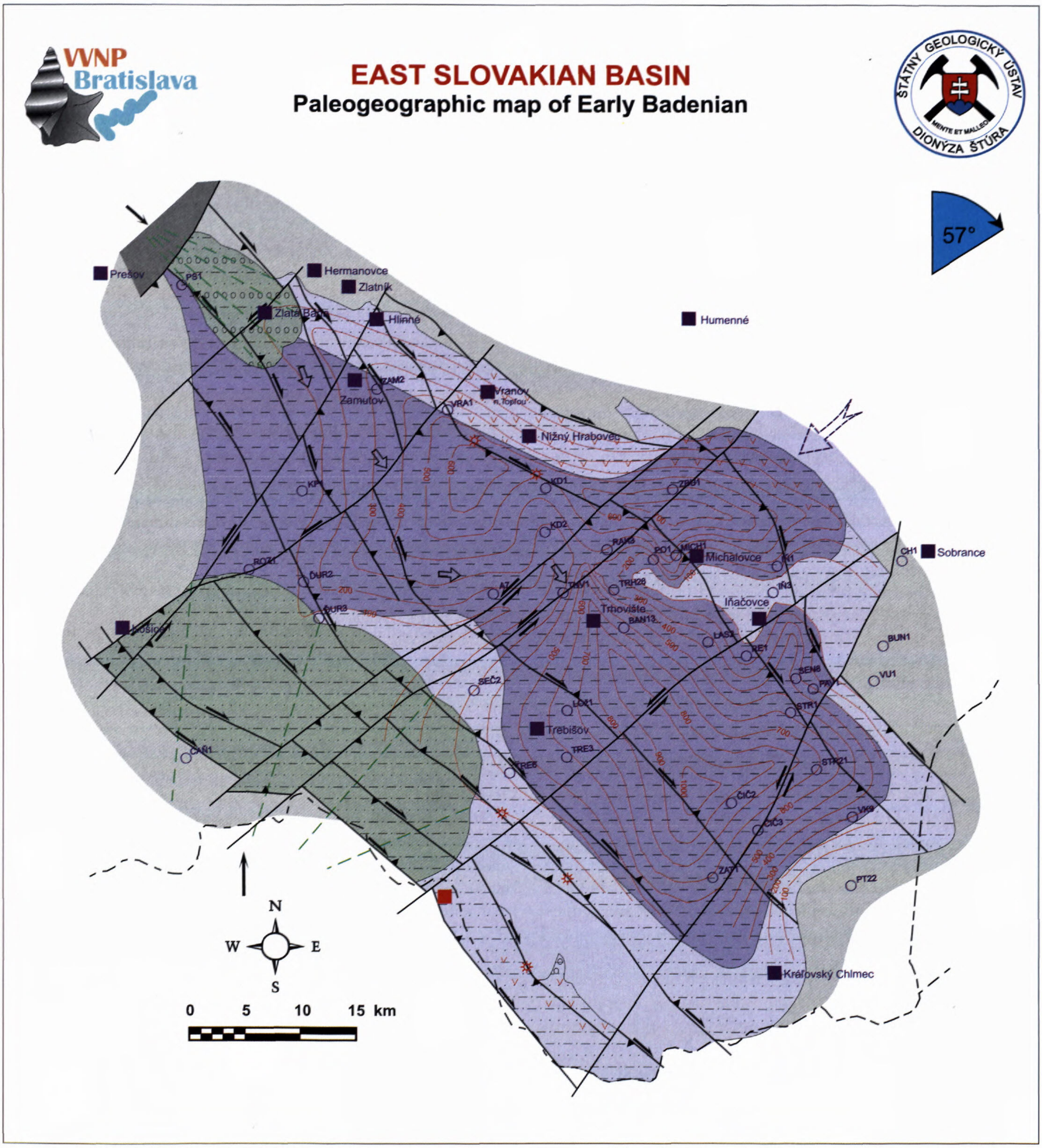
Cumulative thickness of the Karpatian deposits shows that the most intensive subsidence and the thickest deposits occurs S of Vranov, SW of Michalovce and SE of Prešov (Map 2). The thickest part of the deposits belongs to the Kladzany Formation.

Volcanic activity during the Karpatian was represented by acid areal explosive volcanism with crustal origin of magma. The buried products of volcanism occur in the surroundings of Zlatá Baňa (SE of Prešov, Map 2) where also volcanic centers are asummed (Kaličiak et al. 1991) and they outrop NE of Prešov (Fintice Tuff).

At the end of the Karpatian the sea retreated from the larger part of the East Slovakian Basin as it is suggested by reflection-seismic sections from the area.

Early Badenian (16.5 - 15.5 Ma; Map 3)

New widespread marine transgression, corresponding to the TB 2.3 cycle of sea level fluctuation (Haq et al. 1987), occurred at the beginning of the Badenian. The main faults governing the opening of the basin, were dextral strike-slips active in the palaeostress field with maxi-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EAST SLOVAKIAN BASIN
Paleogeographic map of Early Badenian
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
MENTE ET MALLEO
57°
Prešov
Hermanovce
Zlatník
Zlatá Baňa
Hlinné
Humenné
Zámutov
Vranov n. Topľou
Nižný Hrabovec
Michalovce
Sobrance
Iňačovce
Trhovište
Košice
Trebišov
Kráľovský Chlmec
N
W
E
S
0
5
10
15 km

mum compression in NNW - SSE direction. The sea transgression was from NE, i.e. from the basins located in the Outer Flysch zone or from the foredeep and from NW (from Orava marine basin). Occurrence of the Early Badenian deposits on the foothil of the Zemplínske vrchy Hills implies possible transgression also from the area of the Great Hungarian Plain or Nyírség Basin. The foraminifera assemblage analysed from the Early Badenian mudstones occurring close to the Zemplín Hills have a rich planktonic association (Lehotayová in Ivan 1962, Cicha in Čechovič et al. 1963) suggesting sedimentation in deeper, probably outer shelf environment. This, at least, implies seagates along the Zemplínske vrchy Hills suggesting marine communication from the south which, up till now, was first assumed in the Late Badenian. (Rudinec 1989a,b).

The deposits of Early Badenian (Moravian) **Nižný Hrabovec Formation** disconformably overlies the Kladzany Formation (Fig. 6). The only exception occurs in the Prešov Depression where a longitudinal horst trending from Zamutov to Zlatá Baňa determined preservation of shallow sea on the $CaSO_4$ saturation boundary and gradual transition from the Kladzany Formation. In this part the Early Badenian microfauna occurs in the upper part of variegated deposits of the Kladzany Formation (Zlínska & Karoli in Kaličiak et al. 1991). The communication with the sea was improved after a transverse deformation of the horst by faults of NE-SW direction. The Nižný Hrabovec Formation consists of grey calcareous siltstone and claystone interbedded by fine- to medium-grained massive sandstone up to 0.5 cm thick. The sandstone occurs more frequently in the NE margin of the basin where it comprises 0.3 – 3 m thick beds. Siltstone and claystone content increases basinward. In the central part of the basin calcareous clay/claystone prevails. The thickest (about 1 000 m) deposits occur east of Trebišov (Map 3). The deposits crops out on the N margin of the basin in the surroundings of Oblík and in a narrow strip between Hlinné and Nižný Hrabovec in the vicinity of Vranov. The occurrence of the formation in other parts of the basin is proved by deep boreholes. Lenses of fibrous gypsum with dimensions 50x30x7 m occur in dark grey claystone in the surroundings of the elevation point Oblík (Slávik 1967, Karoli in Kaličiak et al. 1991), and suggest lagoonal environment having intermittently oversaturated waters with $CaSO_4$ resulting in evaporitic deposition of sulphates. In the same area beds of polymic conglomerate and gravel associated by sandstone overlie interval consisting of claystone with gypsum. The conglomerate is coarse-grained and blocky. The clasts predominantly consist of flysch rocks and minor limestone and cherts. Similar conglomerate also occurs in the surroundings of Zlatník and Hermanovce SE of Prešov (Karoli in Kaličiak et al. 1991, Vass in Baňacký et al. 1987). The deposits probably originated in deltas entering the basin.

The base of Nižný Hrabovec Formation on the foothill of the Zemplín Hills is composed of polymic conglomerate consisting of local rocks. The conglomerate is overlain by grey sandy benthonitic clay and claystone containing sandstone beds, which is capped by pumice rhyolite tuff with clasts of rhyolite and sandstone composed of Carboniferous rocks fragments (Ivan 1962, Elečko & Vass in Baňacký et al. 1989). The stratigraphic profile of the Nižný Hrabovec Formation was revealed by the borehole Zat-1. The borehole was terminated after 600 m of drilling in the formation (Tereska 1969). The penetrated deposits consist of prevailingly dark grey, bituminous siltstone and claystone interbedded by calcareous sandstone, acid tuff, redeposited tuff and ryodacite which is often propilitized.

A typical member of the Nižný Hrabovec Formation is Hrabovec Tuff. It forms several meters to several tens of meters thick, light green layers. The volcanoclastic material is altered and zeolizated. The original rock was ryodacite or dacite tuff. NE of Sobrance nearby Podhoroď unaltered rhyolite tuffs represent equivalent of the Hrabovec Tuff.

Abundant marine fauna of the Nižný Hrabovec Formation suggests extensive transgression in the basin. The foraminifera assemblage contains for example index species of N8 zone *Orbulina suturalis* accompanied by planktonic forms *O. universa*, representants of species *Praeorbulina*, benthonic species including forms typical for Moravian (lagenids zones): *Lenticulina calcar, L. cultrata, L. auris* etc. (Lehotayová in Ivan 1962, Cicha & Kheil 1962, Cicha in Čechovič et al. 1963, Gašpariková & Slávik 1967 etc.). The calcareous nanoflora assemblage contains individuals typical for the zone NN 5 including index species *Sphenolithus heteromorphus* (Lehotayová 1982).

The Early and Middle Badenian deposits of the Prešov Depression are assigned to **Mirkovce Formation** (Fig. 6). Deposits of the formation crops out on the western foothil of the Slanské vrchy Mts. E and SE of Prešov. They are composed of monotonous, grey calcareous claystone rarely containing layers of fine-grained sandstone. Locally, montmorillonitic clay occurs in the lower part of the formation. In the area of the later stratovolcano Šťavica (SE of Prešov) the claystone was thermally altered by dioritic porphyres to contact chert having dish-like jointing. Maximum thickness of the formation is 630 m. It is underlain by rhyolitic pumice and lapilli tuff and volcaniclastic breccia beneath the Slanské Vrchy Mts. The fragments of breccia are composed of bounded rhyolit.

According to the bioecological and lithologic features several Early Badenian marine depositional environments may be distinguished in the basin (Map 3). In the central part of the basin, where the Nižný Hrabovec Formation has the greatest thickness and prevailingly consists of pelite, we assume relatively deepest depositional environment comparable to outer shelf. It is flanked by shallower environment from both northern and southern sides. This type of environment also occurs on the submarine elevation between Trhovište and Iňačovce having E-W direction. In the NW part of the Prešov Depression a delta, consisting of siltstone, sandstone and conglomerate, entered the basin. A greater delta also entered the basin from the SW. The western margin of the delta extended in to the Moldava or Košice Depression.

During the Early Badenian acid rhyolit-ryodacite crustal areal volcanism was active. The volcanic centers were mainly associated with the elongated fault system near the northern margin of the basin. The main product of this voclanic activity was Hrabovec Tuff representing equivalent of Novoe Selo Tuff in the Transcarpathian Ukraine and Dej Tuff in Romania. The rhyolite pumice, lapilly tuff and volcanic breccias containing fragments of bounded rhyolite, occurring in the centre of the later Zlatá Baňa stratovolcano, are probably also of the Early Badenian age.

Middle Badenian (15.5 – 15.0 Ma; Maps 4 and 5)

The evolution of the basin during the Middle Badenian continued in pull-apart regime similarly to the Late Karpatian and Early Badenian. We suppose any change in stress field and synsedimentary strike-slips occurred along the same faults like in the Late Karpatian and Early Badenian (Map 4). The basin opening continued along dextral strike-slip faults of NE - SW direction. Later, when salt-bearing Zbudza Formation was deposited, the strike-slip faults turned to be normal faults (Map 5).

The Middle Badenian (Wieliczkian), marine **Vranov Formation**, occurring east of the Slanské vrchy Mts. (Fig. 6) and upper part of **Mirkovce Formation,** located west of the Slanské vrchy Mts., suggest a continuation of marine transgression in the basin. The only difference was recorded in the area of the Zemplínske vrchy Hills where alternation of bathyal and littoral fauna (Lehotayová in Ivan, 1962, Kantorová in Baňacký et al. 1989) suggests oscillation of the sea level, most probably due to the tectonic activity of the Zemplín Horst. Main basin communication with open sea in the outer part of the Carpathians was still a seaway to the foredeep in NE. Another communication is assumed to the former marine Orava Basin in NW.

The deposits of Vranov Formation, croping out in Vranov and its surroundings, consist of gray calcareous claystone and siltstone alternating with sandstone. Sandstone and sand, suggesting a shallower environment, occur mainly on the NW margin of the basin in the surroundings of Ruská Nová Ves. Claystone prevails in the basin centre. In the surroundings of the Zemplínske vrchy Hills and nearby Trebišov redeposited vitroclastic tuffs and bentonized tuffaceous clays occur. Maximum subsidence occurred eastward of Trebišov and northward of Michalovce where the thickness of deposits is about 1 200 and 1 000 m, respectivelly. Both subsidence centers were separated by a submarine elevation between Trhovište and Iňačovce. The elevation also determined thickness and facial characteristics of deposits during the Early Badenian. Another depocentrum southward of Vranov was separated from above mentioned centres by a SW - NE trending submarine ridge between Slanec, Albinov and Dlhé Klčovo.

Foraminifera species of the formation correspond by their composition to assemblages of zone with *Spiroplectamina carinata (Valvulineria arcuata, V. marmaroschensis, Bulimina elongata intonsa, Uvigerina pygmoides, U. aculecta, U. hispida, U. aff. rugosa, U. aff. asperula, U. aff. costata* (Lehotayová in Ivan 1962, Zapletalová 1974, Kantorová in Baňacký et al. 1989, Zlinská, 1996).

The absence of coarse clastics suggests a flat relief surrounding the basin. Volcanic activity has not been recorded during the deposition of the Vranov and upper part of Mirkovce Formations.

The tectonic activity ceased during the Middle Badenian. Only normal faults with small throw amplitudes were active. The retreating sea determined development of hypersaline lagoons.

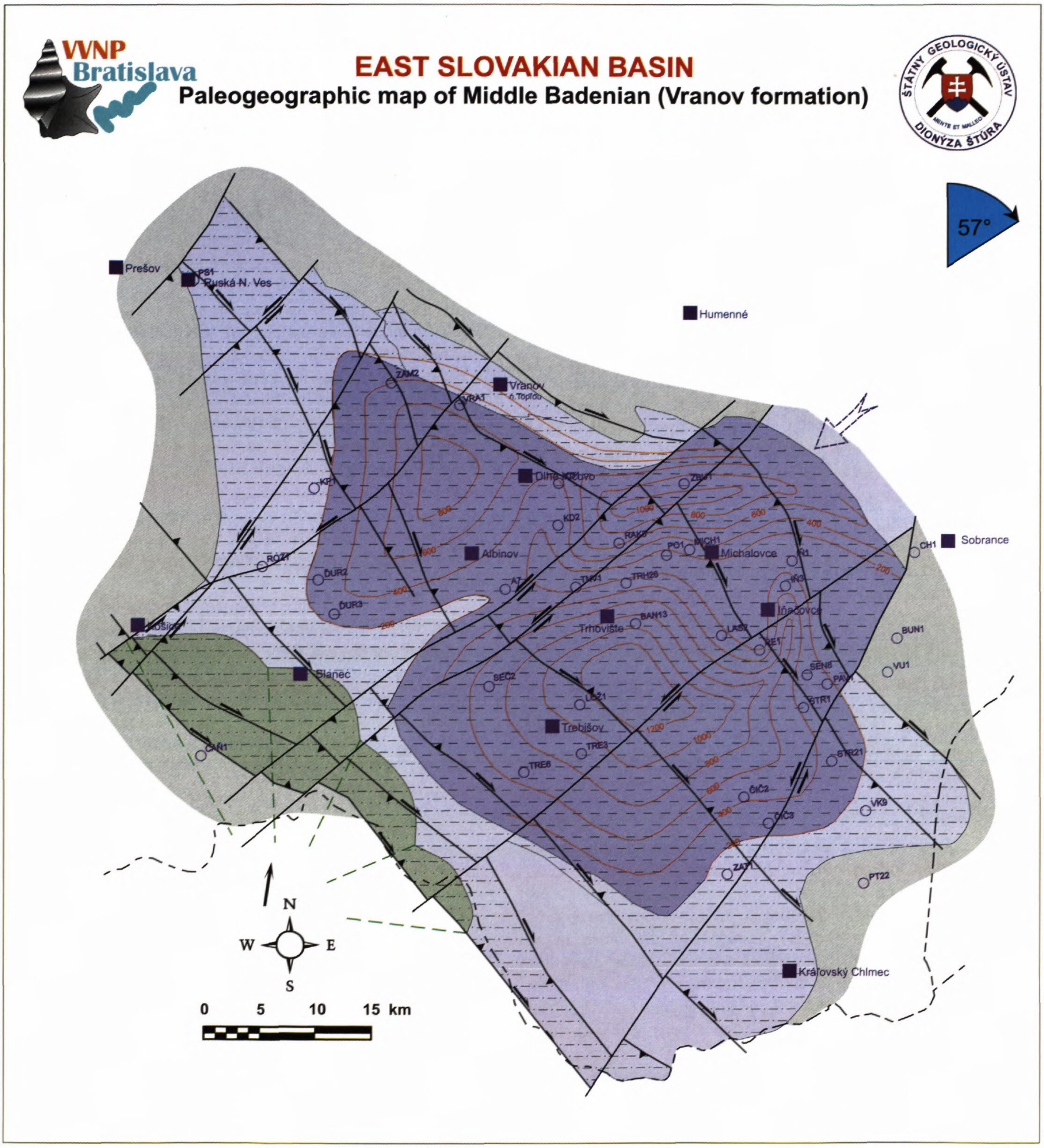
Evaporitic deposition, occurring in salt lagoons and shallow-marine environment, prevailed during deposition of **Zbudza Formation** extended in the central part of the basin (Fig. 6). It does not crop out and its occurrence was only documented by boreholes on the northern margin and in the central part of the basin (map 5). It attains maximum thickness of 300 m. The lower part of the formation consists of fine-grained clastics with chaotically dispersed or aggregated nodules of anhydrite. The clastics are overlain by lenticular, about 60 - 80 m thick and 3 km long, halite bodies. The whole succession is caped by beds consisting of clast-supported halite-rudite (Karoli 1998).

The Zbudza Formation originated during the regional salinity crisis in the Paratethys area. The massive halite originated not only in the East-Slovakian Neogene Basin but also in the Carpathian Foredeep (Wieliczka), in the Transcarpathian Ukraine and in the Romanian Transylvania. The formation contains a poor foraminifera assemblage: *Globigerina aff. bulloides, Globorotalia scitula, Globigerinoides trilobus, Uvigerina aculeata* (Gašpariková 1963).

Late Badenian (15.0 – 13.6 Ma; Map 6)

At the beginning of the Late Badenian (Kosivian), after salinity crisis, the tectonic activity of the basin increased resulting in revitalizing of pull-apart mechanism. The stress field had a maximum compression in E-W direction. The actual palaeostress was not found by measurements of brittle deformations of the Middle Badenian rocks (shortage of suitable outcrops), but on basis of data obtained from the Middle Sarmatian rocks (Kováč et al. 1994). We suppose those data is more reliable for the Late Badenian because palaeostress conditions of the Late Karpatian - Middle Badenian have been interrupted by the tectonic event enabling evaporitic sedimentation in the basin. The main role in a new basin opening played sinistral strike-slips of NW-SE direction and dextral strike slips of NE-SW direction. The faults of N - S direction were normal faults (Map 6 and Fig. 4).

The Late Badenian stage starts with marine transgression having about 0.5 Ma time lag against the global cycle


VVNP
Bratislava
EAST SLOVAKIAN BASIN
Paleogeographic map of Middle Badenian (Vranov formation)
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
MENTE ET MALLEO
57°
Prešov
Ruská N. Ves
Humenné
Vranov n. Topľou
Sobrance
Michalovce
Albinov
Trhovište
Iňačovce
Slanec
Trebišov
Košice
Kráľovský Chlmec
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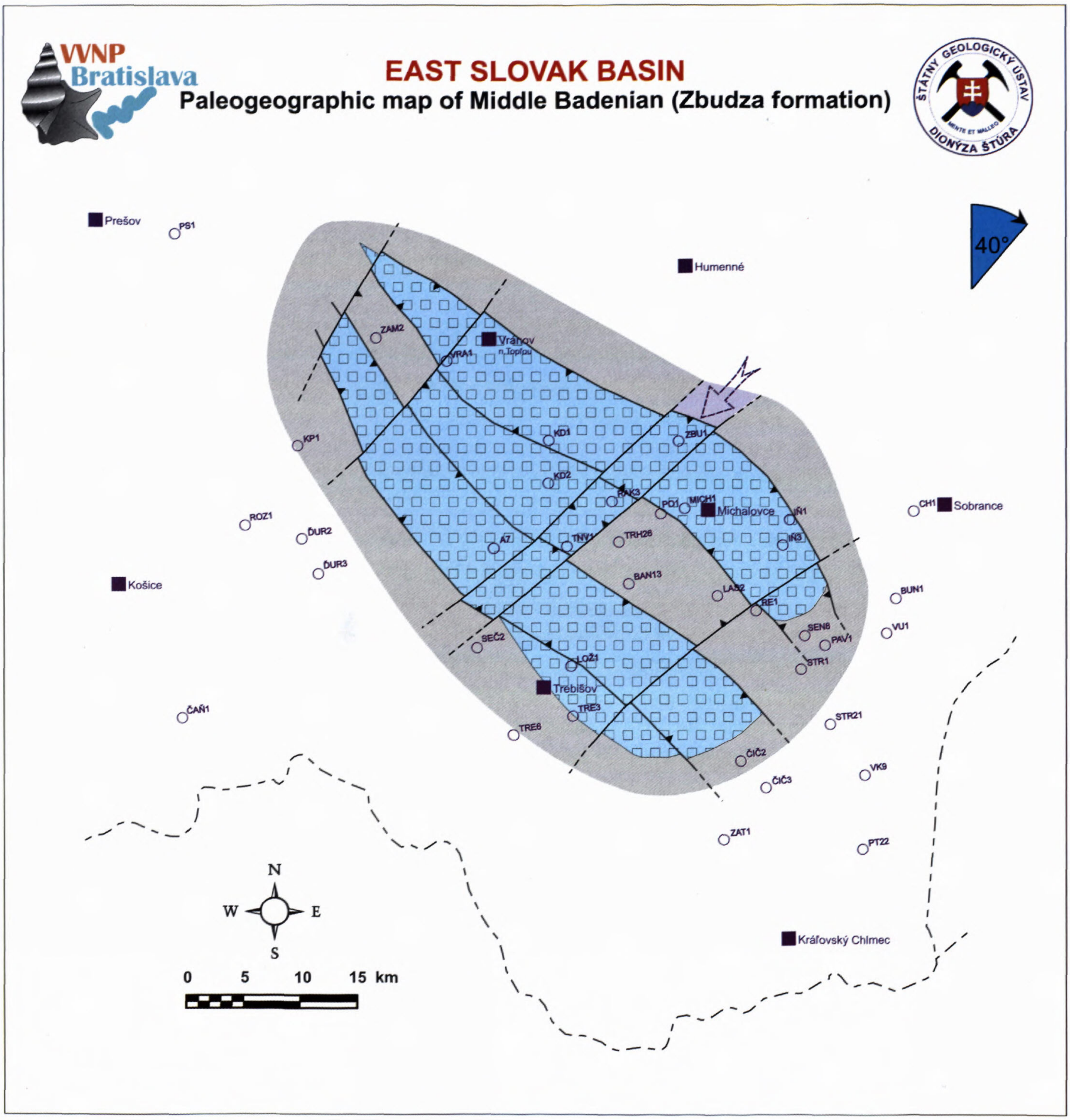

EAST SLOVAK BASIN
Paleogeographic map of Middle Badenian (Zbudza formation)
VVNP Bratislava
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
MENTE ET MALLEO
40°
Prešov
Humenné
Vranov n.Topľou
Michalovce
Sobrance
Košice
Trebišov
Kráľovský Chlmec
N
W
E
S
0 5 10 15 km

TB 2.4 of the eustatic sea level fluctuation (Haq. et al. 1987), which was probably determined by local tectonics. Sea transgression came from NE from the Carpathian Foredeep (Fig. 5) in NE, and also from S and SW (Rudinec 1989).

The Late Badenian transgressive marine deposits are represented by the lower part of **Lastomír Formation**. The upper part of the formation interfingers with deposits of prograding delta assigned to the **Klčovo Formation** (Fig. 6). Deposits of the Lastomír formation are represented by calcareous, gray claystone containing beds of sandstone, acid tuff and redeposited tuff. Acid volcanoclastics occur on the N and NE margin of the Zemplínske vrchy Hills and on the foothills of the Slanské Vrchy Mts. (Map 6). Maximum thickness of the deposits, E of Trebišov, is 2 000 m. Shallow-marine facies of the formation marked as inner shelf occur in the southern and northern part of the basin. Toward the central and south-eastern part of the basin the facies pass into more deep-water facies of outer shelf. According to biofacial indicators the environment could represent a shallow bathyal zone.

Foraminifera assemblages containing species typical for Velapertine biozone (bulimina-bolivina zone) e.g. *Bulimina ex.gr.elongata, B. ex. gr. pupoides, B. ovata, Bolivina dilatata, B. antiqueformis, Uvigerina asperula, U. aff. semiornata, Valvulineria complanata, Cibicides boueanus, C. dutemplei* (Kudláčková et al., in Janáček, 1963, Jiříček 1972) suggest normal marine conditions.

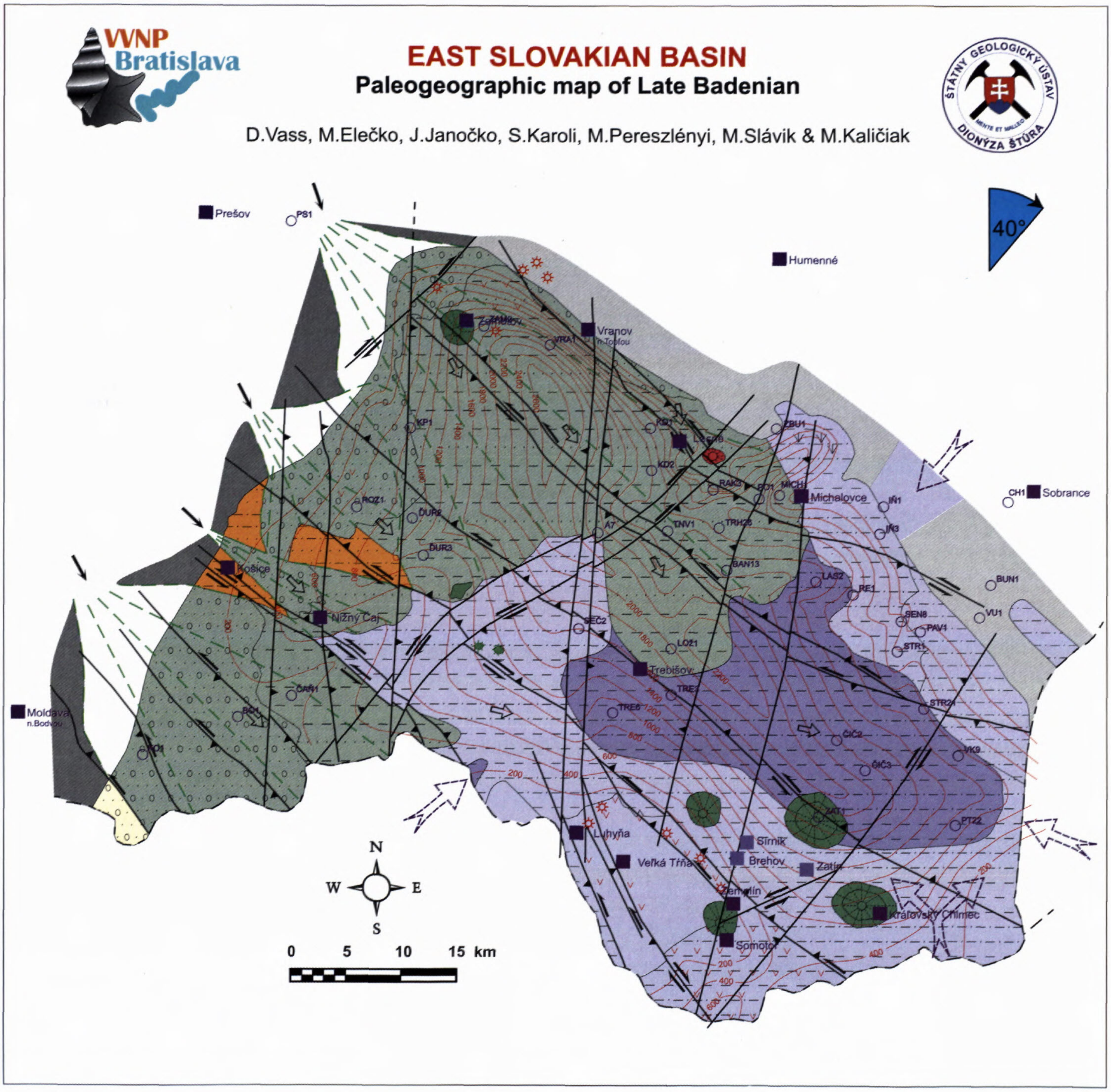
In the surroundings of Zatín the main part of the Late Badenian deposits consists of the Zatín volcanics having max. thickness about 1 400 m (borehole Za-1, Rudinec & Tereska 1972). The volcanics consist of lava flows composed of pyroxenic andesite associated by volcanoclastics and rhyolite. They are interlayered by bituminous clay/claystone and calcareous sandstone. Rarely marine fauna occurs (Jiříček in Tereska 1969). Radiometric ages of andesite and andesite tuff yielded 15.0-+2 Ma age (Slávik et al. 1976).

The Lastomír Formation laterally gradually passes into **Klčovo Formation**. This is proved by change of lithology and by change of some faunistic assemblages suggesting shallowing and hyposalinity of the environment as a result of the Klčovo delta progradations. The assemblages are depleted and they are represented by *Ammonia beccarii, Porosononion comunis, Elphidium sp., Virgulina schreibersi, Roussela spirulosa, Bulimina elongata, Ostrea digitalina, Cardium cff. andrusovi, C. turonicum, C. edule, Corbula gibba, Anomia ephipium , Ervilia dissita podolica, Clithon pictus, Hydrobia stagnalis* (Seneš 1955, Švagrovský 1959, Jiříček 1972). The Klčovo Formation as prograding deltas represents regressive part of the Late Badenian sedimentary cycle. The regression was accompanied by a uplift of the western basin margin which became the main source area delivering clastics into the basin. The deposits of the formation crop out in the Prešov Depression (Janočko 1990). Eastward of the Slanské vrchy Mts. (Fig. 5) it is mostly burried. In this part it was described by Jiříček (1972) and later defined on the base of seismic lines by Beřicha (1992). Its subsurface occurrence is as far as Trebišov and Michalovce. In the lower part of the formation poor brackish fauna assemblage was found (*Ammonia ex gr.beccarii, Porosononion subgranosum, Miliamina fusca, Egerella scabra*). Above this only ostracoda (*Candona strigulosa*) and terrestrial molluscs are autochtonous (Zapletalová unpubl., Jiříček 1972). Maximum thickness 2 800 m of the formation was found nearby Dlhé Klčovo southward of Vranov nad Topľou.

In the Prešov Depression the Klčovo Formation commences with Kráľovce Tuff which represents redeposited ryolite tuff containing pumice (up to 5 cm in diameter), lithoclasts of rhyolitic character and fragments of augit-hyperstenic andesite. The tuff is about 20 – 30 m thick (Kaličiak et al. 1991). The radiometric age of the tuff is about 13.9 +-0.3 Ma (Bagdasarjan et al. 1971). It is overlain by the Varhaňovce Gravel composed of polymic gravel having clasts from the Spišsko-Gemerské Rudohorie Mts. and Čierna Hora Mts. and greenish-gray, locally brownish-spotted calcareous clay. The clay contains layers of sand (up to 10 m of thickness), pebbles, redeposited tuff, coaly clay and rarely also lignite (Čverčko et al. 1968, Jiříček 1972, Janočko 1990, Reed et al. 1992). These sediments were mostly deposited in fan delta environments (Janočko 1990). They contain poor Badenian microfauna (Zlínska in Kaličiak et al. 1991).

Distal deltaic facies interfingering with shoreface deposits were described from the eastern part of the Moldava Depression in the borehole K-8 located nearby Nižný Čaj SE of Košice. They prevailingly consist of sandstone and siltstone, minor claystone and conglomerate. The lithology points to frequent lateral changes of distributary channels and interdistributary environments within the delta plain (Reed et al. 1992).

During the Late Badenian the basin deposition was accompanied by an important volcanic activity. It started by explosive ryodacite volcanism resulting in redeposited and pumice ryodacite tuff underlying andesites of the Kráľovský Chlmec stratovolcano. It was followed by andesite volcanism, products of which redeposited andesite tuff overlain by the lava flows of pyroxenic andesite. The stratovolcanoes occur in the surroundings of Zatín, Brehov, Sírnik and Somotor - see Map 6 (the assignement of these volcanics to the Late Badenian and not to the Sarmatian as it is on the map of Baňacký et al. 1989 is based on the radiometric ages of the Zatín andezite 15.0 ± 0.8Ma, Brehov andesite 14.0 ±1.4 Ma (Bagdassarjan et al. 1971) and mineralization accompanying older pre-Sarmatian volcanism). The extrusive ryodacite volcanism is today represented by individual bodies S and SE of Trebišov (Map 6). Ryodacite (rhyolite) occurring nearby villages Zemplín and Hrčeľ (Cejkov rhyodacite, see Fig. 6) are associated with silicified and adularized breccia. The radiometric age of the rhyolite at Hrčeľ is 13.5 ± 2.5 and 14.9 ± 2.8 Ma (Tsonj & Slávik, 1971). The character of alterations suggests submarine volcanism. The redeposited volcaniclastics are product of explosive-hydrothermal breccia in a submarine environment. The redeposited ryodacite tuff overlying the biostratigraphi-


VVNP Bratislava
EAST SLOVAKIAN BASIN
Paleogeographic map of Late Badenian
D.Vass, M.Elečko, J.Janočko, S.Karoli, M.Pereszlényi, M.Slávik & M.Kaličiak
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
40°
Prešov
Humenné
Vranov n.Topľou
Michalovce
Sobrance
Košice
Nižný Čaj
Trebišov
Moldava n.Bodvou
Luhyňa
Veľká Tŕňa
Brehov
Zatín
Somotor
Kráľovský Chlmec
N
W
E
S
0 5 10 15 km

cally proved Late Badenian deposits, was found by drilling between Veľká Tŕňa and Luhyňa S of Trebišov. The pumice ryodacite tuff occurs S and SE of Trebišov (Kaličiak in Baňacký et al. 1989).

In the northern part of the East-Slovakian Basin a body having fluidal bounded structure was drilled in the area of the central zone of the younger Zlatá Baňa stratovolcano (Dubník located SE of Prešov). South of Soľ (NW of Vranov) a horizon of rhyolite tuff occurs which is associated with epiclastic rhyolite sandstone. The radiometric age of the tuff is 14.2 ± 0.7 Ma (Repčok in Kaličiak et al. 1991). Two bodies of rhyolite crop out at the village Lesné NW of Michalovce. The radiometric age of one of them is 15.2 Ma (Bagdassarjan et al. 1971).

South of Zamutov (W of Vranov) an extrusive ryodacite dome occurs. Its analogues found in the borehole Zam-2 were radiodated to 14.4 Ma (Bagdasarjan et al. 1971).

The evolution of the andesite stratovolcano Ošvárska (W of Vranov) commenced in the Late Badenian. It consists of autochthonous pyroclastics, pyroxenic andesite lava flows and andesite necks. The stratovolcano also was active during the Early Sarmatian.

Early Sarmatian (13.6 – 12.7 Ma; Map 7)

We assume that the main tectonic structure, of Early Sarmatian, is identical to the Late Badenian one. It is represented by NW sinistral, conjugate and NE dextral strike-slips as well as normal faults having N-S direction. The opening of the basin was intensive as it is proved by the highest sedimentation rate in the SE part of the basin (267.8 cm/1 000 years before decompaction and some 390.8 cm/1 000 years after decompaction; Vass & Čech 1989, Král et al. 1990, see Fig. 2).

During the Early Sarmatian the area of the central Paratethys, including the East-Slovakian Basin, began to be separated from the open sea. The Sarmatian epicontinental sea in the Paratethys was brackish with salinity around 20 – 25 ppm. In spite of the isolation, the transgression in the basin coincides with eustatic sea level fluctuation cycle TB2.5 (13.8 Ma , Haq et al. 1987; Fig. 5). The basin was connected with depositional areas in Hungary and the Transcarpathian Ukraine. The connection to the Carpathian Foredeep was terminated and the emerged Outer Flysch Carpathians became one of the main source area.

At the beginning of the Sarmatian the brackish sea incursed the basin from the south. It came round the Zemplin-Beregovo Horst and flooded area as far as Michalovce in the north, the Vranov nad Topľou in the northwest and Košice in the west (Map 7). It also overcame the Uzhgorod Horst and reached the Transcarpathian area. In the western and northern part of the basin sea was shallow and it was influenced by progradig deltas. A shallow sea also occurred in the area of the Zemplín – Beregovo Horst. In the southeastern part of the basin the brackish sea was relatively deeper. At these places the Early Sarmatian deposits attain the greatest thickness, up to 2 400 m.

The Early Sarmatian deposits comprise **Stretava Formation** (Fig. 6), which generally represents transgressive phase of Sarmatian periode. It consists of gray calcareous clay interbedded by layers of pale and variegated clay, sand, gravel and minor acid tuff. Tuff and reworked tuff occur in the Olšava and Myšľa Members (Švagrovský 1956, Rankovce Tuff, Seneš 1955). The intrusive ryodacite bodies or necks (ryodacite of Lipová), associated by swarm of dikes, crop out on the northern margin of the basin NW, of Vranov. Their radiometric age is 13.2 ± 3 and 13.3 ± 1.2 Ma (Repčok 1977, Merlič & Spitkovskaja 1974). Among Malčice – Beša – Čičárovce andesite stratovolcanos are interfingered with Stretava Formation (Ďurica 1965).

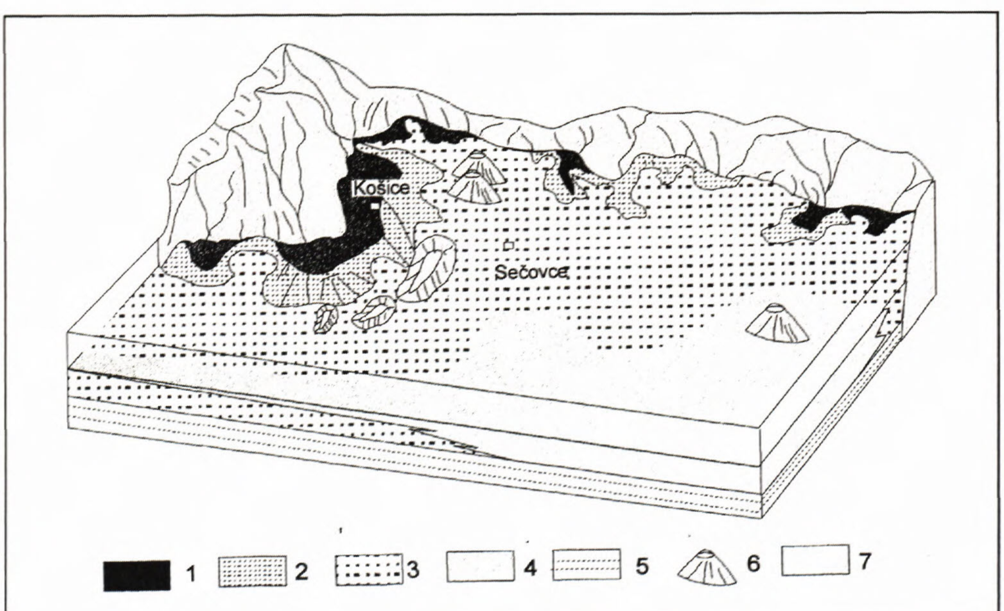
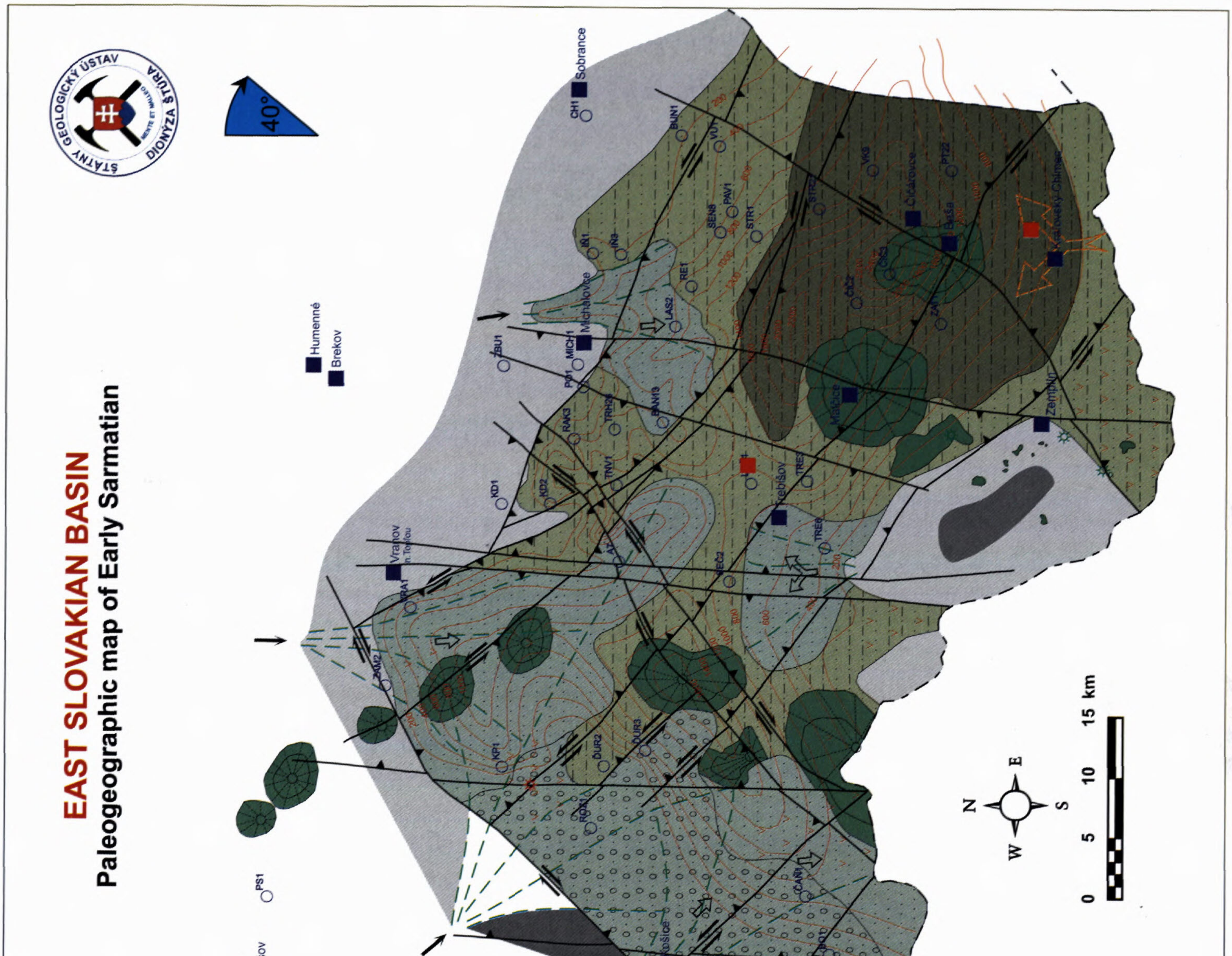


Fig. 10: Blockdiagram showing assumed extension of the Early Sarmatian deltas in the East-Slovakian Basin. The extension of deltas in western part of Basin are limited by the contemporary stratovolcanos of Prešov – Tokaj Mts. Explanation: 1 – fan deltas, 2 – deltaic deposits, 3 – inner shelf, 4 – outer shelf, 5 – pre-Sarmatian deposits, 6 – volcanoes, 7 – pre-Neogene rocks


EAST SLOVAKIAN BASIN
Paleogeographic map of Early Sarmatian
Prešov
PS1
Humenné
Brekov
Vranov n. Topľou
Michalovce
Sobrance
Košice
Trebišov
Zemplín
Kráľovský Chlmec
Moldava n. Bodvou
40°
N
W
E
S
0 5 10 15 km
WNP Bratislava
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA

Slovak Geol. Mag., 6, 4(2000), 377–407

The formation contains brackish fauna assemblage: *Cardium politioanei politioanei, C. vindobonensis, Evilia dissita dissita, Tapes vitalinus, Mohrensternia sarmatica, Acteocina lajonkaireana lajonkaireana, Pirenella picta mitralis* (e.g. Čechovič 1940, Körössy 1940, Švagrovský 1952, 1956, 1960). Based on the foraminifera assemblage it is possible to subdivide the formation into two biozones: (Gašpariková, Cmuntová, Prokšová in Brodňan 1959, Lehotayová in Čechovič & Vass 1960, Brestenská & Priechodská 1959, Zlinská in Kaličiak et al. 1991, Zlinská, 1992).

– a biozone containing *Elphidium reginum* with associated species *E. crispum, E. macellum, E. aculeatum*: the Early Sarmatian sensu Grill (1943) which deposits are showed in Map 7;

– a biozone containing *Elphidium hauerinum* with associated *species E. listeri, e. cf. aculeatum, E. cf. minutum, Porosononion bogdanoviczi* suggesting the Middle Sarmatian sensu Grill l.c. Deposits correlated with the biozone are included in Map 8.

In the western part of the basin the Stretava Formation contains coarse-grained deposits marked as Košice Gravel (Vass 1989, Janočko et al. 1991, 1998). The gravel is polymic and alternates with sand and gray and dark-gray clay containing sporomorphs assigned to the Early Sarmatian. The sedimentary succession originated in deltaic and shallow-marine environment. In the outcrops in the surroundings of Košice small cycles consisting of alternating deltaic, shoreface and offshore deposits occur. The change of deltaic to shoreface deposition was determined by allocyclic (sediment input and tetonics) and autocyclic (delta lobe switching) factors (Janočko 1998). The emerging volcanic range of the Slanské vrchy Mts. partly restricted the extension of the delta toward the east (Fig. 10). We also assume deltaic deposits at the NW foothill of the Zemplín Horst. Another deltas prograded into the basin from the north. The Laborec River palaeodelta entered the basin through the Brekov Gate and continues toward Michalovce. The Topľa River delta spread the deposits in the area S and SW of Vranov nad Topľou. Smaller deltas developed in the Moldava part of the basin and on the foothill of the forming Vihorlat and Popriečny Mts. (Map 7).

Late Sarmatian (12.7 - 10.5 Ma; Map 8)

The palaeogeography expressed in Map 8 comprises the situation during the Middle and Late Sarmatian according to biostratigraphical subdivision (e.g. zone with Elphidium hauerinum and Porosonium subgramosum resp., Grill 1943). Decreasing sedimentation rate and prevailing extensional tectonic structures suggests change of the basin tectonic style from the pull-apart regime to simple extensional one. The direction of extension was NW-SE. Distribution of faults was similar to the Early Sarmatian system but former strike-slips changed to normal faults (both NW and NE trending systems) and the faults of N-S direction changed to dextral strike-slips (Map 8). The Late Sarmatian deposits represent a new sedimentary cycle (Fig. 5) consistent with T.2.6. cycle of Haq et al. (1987). The beginning of the cycle is well-defined in seismic profiles (Janočko in press).

In the western part of the basin the Middle and Late Sarmatian deposits are represented by **Kochanovce Formation** (Fig. 6). The formation is dominantly composed of light gray clay with scattered volcanic material, reworked tuff and pumice and pumice-lapilli andesitic tuff. Volcanic and volcanoclastic rocks often occur in the surroundings of the Slanské vrchy Mts. where andesite lava flows enter the formation (Kaličiak et al. 1991). Volcaniclasics are locally bentonised. Layers of lignite and coal clay occur in the formation too. The formation contains freshwater fauna (*Planorbis sp., Characeae, Candona II, Illiocypris* (Jiříček 1972, Holzknecht 1970 unpubl.). According to Planderová (fide Kaličiak et al. 1991) the sporomorpha assemblage represents the Late Sarmatian-Pannonian association.

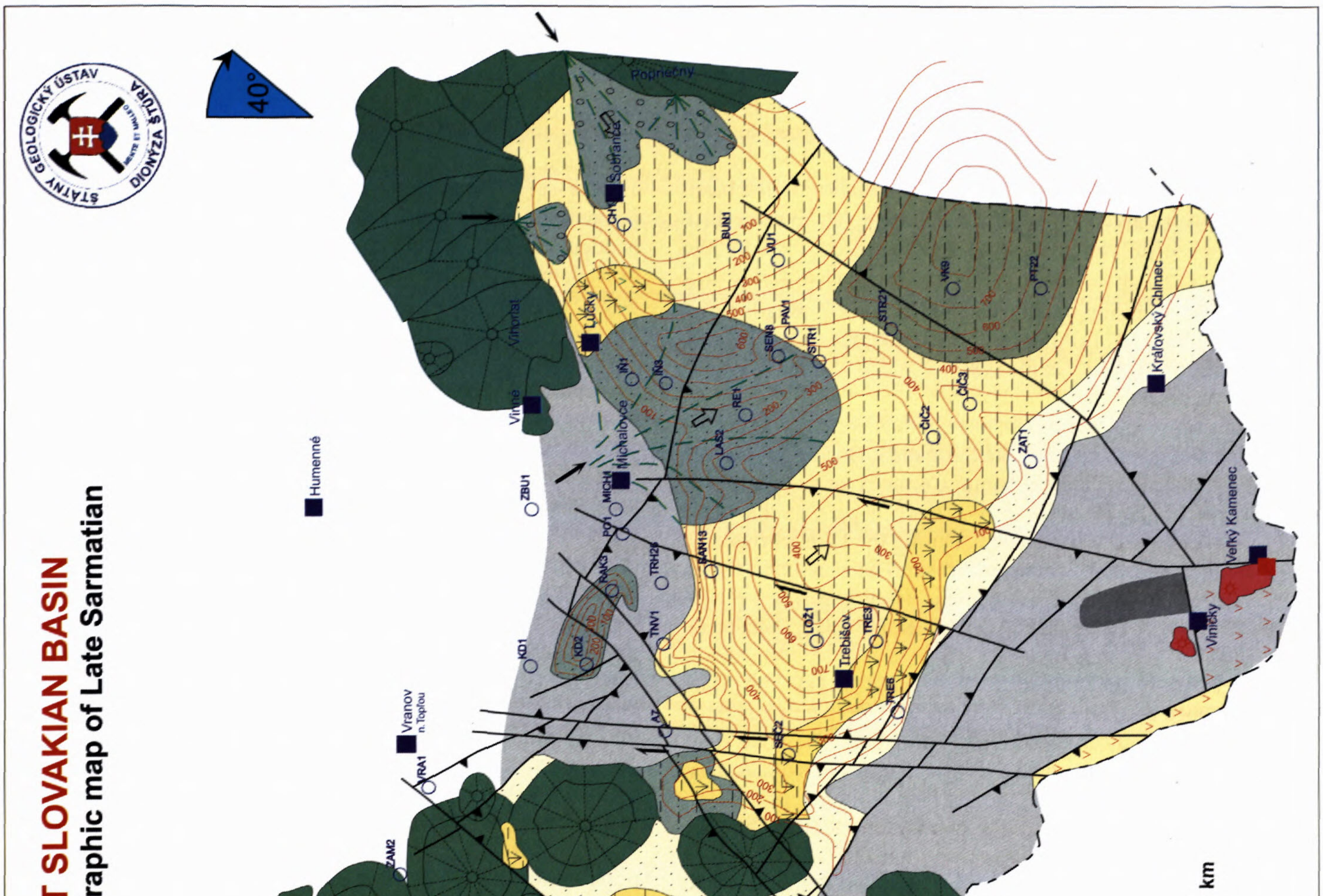
Ptrukša Formation, about 130-200 m thick, represents Late Sarmatian Formation in the eastern part of the basin. It consists of light gray calcareous sand and sandstone with layers of clay and reworked tuff. It contains fauna assigned to the Late Sarmatian biozone with *Porosononion subgranosum* sensu Grill (1943).

Lučky Volcaniclastics are equivalent of the upper part of the Kochanovce Formation and the lower part of the Ptrukša Formation (Fig. 6). They consist of andesite breccia, tuff, reworked tuff and autometamorphic lava of pyroxenic-amphibolic andesite. Tuffitic and silicious limestone occurs too. This member, which probably represents equivalent of the Vinné andesite, is about 50-80 m thick.

Závadka Member overlies the Lučky Volcaniclastics and passes into the Ptrukša Formation basinward. Závadka Member mainly consists of gray calcareous clay and claystone. Beds of sandstone, gravel and reworked tuff also occur. Coal seams grouped into four coal measures containing 13 seams are typical lithologies of the member. The seams are developed in lense forms and they attain thickness from a few cm to 5.9 m . Another typical characteristics of the formation is the occurrence of ironstone balls and layers (concretions and layers of pelosiderite). The Závadka Member contains the Late Sarmatian fauna.

Galgavölgy Rhyolite tuff Formation (formerly Tokaj Formation; Baňacký at al. 1989) is equivalent of the Kochanovce Formation in the Roňava „Bay" (see Fig. 1 and Fig. 6). The formation mainly consists of rhyolite, rhyolite tuff, reworked tuff which are often bentonised. They are associated with lignite seams. Also rhyolite extrusive bodies S of T·ebišov nearby the Slovak-Hungarian state border should be assigned to this formation (Vass, Elečko in Baňacký et al. 1998). The radiometric age of the rhyolite from Viničky is between 11.1 and 12.2 Ma (Vass et al. 1978).

The Sarmatian was a period of an eventful volcanic activity. Except above mentioned volcanics, which comprise a part of the basin fill (andesites and rhyolite tuffs), the main mass of volcanics forming the Slanské vrchy


EAST SLOVAKIAN BASIN
Paleogeographic map of Late Sarmatian
Prešov
Košice
Vranov n. Topľou
Humenné
Michalovce
Trebišov
Vinné
Vihorlat
Lúčky
Sobrance
Popričný
Kráľovský Chlmec
Veľký Kamenec
Viničky
Moldava n.Bodvou
40°
N
E
S
W
0 5 10 15 km

Mts. and Vihorlat Mts. was formed during the Sarmatian. The radiometric ages of the volcanic rocks from the Slanské vrchy Mts. vary from 10.0 to 13.6 Ma (Bagdasarjan et al. 1971, Slávik et al. 1976, Ďurica et al. 1978). The stratovolcano Ošvárska in the northern part of the mountains finished its activity during the Early Sarmatian. At that time also new and main andesitic stratovolcanoes started to form in the Slanské vrchy Mts. (see Fig. 6). Several isolated andesite and rhyolite bodies on the basin margin, rhyolite ignimbrite and perlite (at Viničky 11.1 and 11.4 Ma), including those penetrated by the borehole Ko-1 (12.1 ± 0.7 Ma) in the Moldava Depression also originated during the Sarmatian (Vass 1967, Bagdasarjan et al. 1968, 1971, Pulec and Vass 1969, Konečný, Lexa & Dublan in Baňacký et al. 1989).

According to the radiometric ages the Vihorlat and Popriečny volcanic mountains commenced to form during the Middle and mainly Late Sarmatian. During the Middle Sarmatian the andesite Vinné Complex originated. Ryodacite of the Beňatinská Voda are a part of the older structure of the Vihorlat Mts. During the Late Sarmatian a number of stratovolcanoes arose. Their activity terminated during the Pannonian (Bagdassarjan et al. 1971,Slávik et al. 1976, Ďurica et al. 1978, Žec et al. 1997).

Rhyolite and rhyodacite volcanoes Veľký and Malý Kamenec and Viničky (S of Trebišov) were formed on the Zemplín horst during the Late Sarmatian. They had short lava flows accompanied by various types of volcanoclastics and tuffs which form an important volume of the Galgavölgy Formation on the southern foothill of the Zemplín Hills. The explosions of „nuées ardentes„ type resulting in formation of above mentioned ignimbrite, tuff and perlite, also occurred in the Moldava Depression.

Pannonian and Pontian (10.5 – 5.2 Ma; Map 9)

Termination of the East-Slovakian Basin continued during the Late Miocene e.g. during the Pannonian and Pontian. During this stage of the basin evolution NE-SW and occasionally NW-SE normal faults prevailed (Map 9). During the Pontian the basin was subjected to stress resulting in a weak folding of the basin fill. This is indicated by seismic sections (Keith et al. 1989, Magyar et al. 1997, Mořkovský et al. 1999) and by small fold deformations found in the Teriakovce Formation of the Karpatian age. Another indications are steeply-inclined strata on the basin margin. The Pannonian deposits of Sečovce Formation are also slightly folded (Kováč et al. 1994) giving an important information for timing of the stress shock as post-Pannonian.

Beginning of the Pannonian is emphasized by onlaps on seismic profiles. The Pannonian deposits only extend in the middle and SE part of the Trebišov depression and in the Moldava Depression. They transgressively and unconformably overlie deposits of various biozones and lithostratigraphic units of the Sarmatian. The thickness of the Pannonian and Pontian deposits in the SE part of the basin is about 1 300 m. In the partial depression between Sečovce and Trebišov and E of the Zemplínske vrchy Hills the thickness of deposits is maximum 500 m.

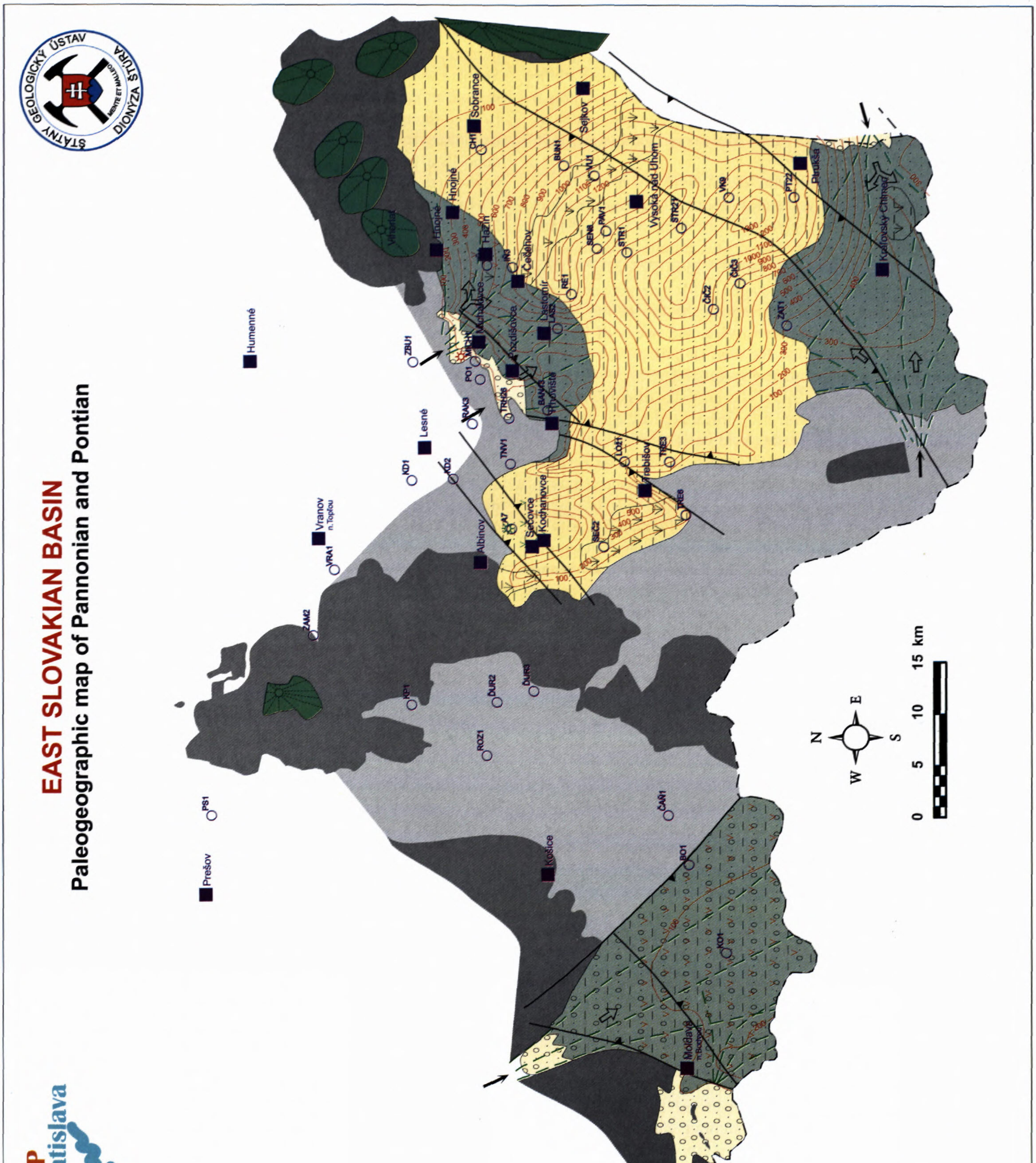
The whole Pannonian sedimentary cycle in the basin is represented by **Sečovce Formation** (Fig. 6). From the underlying deposits of the Kochanovce Formation it differs by variegated colour in the lower part of the sedimentary succession (Janáček 1959). In the upper part the formation consists of gray calcareous clay containing coal clay or seams and lenses of lignite with tuff and redeposited tuff layers. In the surroundings of Sečovce thick-bedded, coarse-grained Albinov Tuff (Albínovská Hôrka Tuff, Janáček 1959, 1963) of amphibolite-pyroxenic andesite occurs in the lower part of the formation. The thickness and grain size of deposits decrease toward the basin centre.

Jiříček (1972) described the „Pannonian type" fauna (*Ciprideis tuberculata, Bithynia aff. tentaculata*) from the Sečovce Formation in the Ptrukša surroundings. Ostracods (*Candona aff. sp.II., Cyprinotus salinus, Limnocytherae sp.*) and gastropods were found in the surroundings of Trebišov (e.g. Janáček 1963). The palynomorpha assemblage has Early Pannonian characteristics (Planderová in Baňacký et al. 1989).

Deposits of the Hažín Tuff and Hnojné Member, both of the Panonian age, occur at the Vihorlat foothill (E of Michalovce). **The Hažín Tuff**, 30 - 40 m thick, discordantly overlies the Závadka Member and Ptrukša Formation (Late Sarmatian, Fig. 6). It consists of pumice granatic grayish-green, palish-gray tuff and reworked tuff. Clasts of palish-gray pumice and rhyolite clasts are frequent. They are accompanied by tuffaceous clay, coal lenses and ironestone balls and layers. Jiříček (1972) reports ostracods and *diatomae Carychium minimum, Cyprinotus sp., Candoinella albicans, Darwinula stewensoni, Melosira arenaria* from deposits, which he considers as equivalent of the Hažín Tuff. He correlated them with the Early Pannonian „A" biozone.

The Hnojné Member overlies the Hažín Tuff. It mainly consists of gray and palish-gray clay containing seams and lenses of lignite. So called main coal seam measure has the largest areal extension. The coal seams in this member are 2.5 – 5m and occasionally up to 10 m thick. The main seam is divided by a layer of tuffitic limestone and spongodiatomic combustible shale into two beds in the eastern part of the Hnojné deposit. Higher up 10 – 15 seams having irregular thickness and development occur within clays. Reworked tuff, ironstone balles and thin layers also occur. The Hnojné Member contains fresh-water fauna from the family *Planorbidae*, diatomae of *Melosira* family and leaf imprints. From the beds considered as the equivalent of the Hnojné Member Jiříček reported ostracode assemblage of fresh-water fauna (*Carychium minimum, Bithynia tentaculata, Candoniella albicans, Ciprideis tuberculata, Pisidium sp.*) and he correlated this member with the Pannonian „B" biozone.

Lignite bearing beds nearby Sejkov are equivalent of the Hnojné Member (Rudinec & Čverčko 1970) and they are bounded to the NE slope of the depression between Lastomír and Vysoká nad Uhom (Map 9).


EAST SLOVAKIAN BASIN
Paleogeographic map of Pannonian and Pontian
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
WNP Bratislava
Prešov
Košice
Vranov n. Topľou
Humenné
Lesné
Albínov
Michalovce
Sobrance
Hnojné
Trebišov
Sečovce
Kochanovce
Vysoká nad Uhom
Pribeník
Moldava n. Bodvou
0 5 10 15 km
N
S
E
W

Deltas entered the Pannonian and Pontian lake in the basin area. Two of them, obviously deltas of Laborec and Ondava palaoerivers, occurred on the northern margin of the lake. Lignite seams originated on the delta plain of the Laborec River delta. The delta, probably of the Latorica River, entered the basin from the east. A separate deltaic system developed in the Moldava Depression (e.g. Janočko & Šoltésová 1993).

During the Pannonian, the activity of the Slanské vrchy Mts. andesite volcanoes (about 10 Ma old, Slávik et al. 1976) as well as the activity of the most Vihorlat Mts. stratovolcanoes were ceasing. The youngest volcanics of the Vihorlat Mts. are about 9 Ma old (Slávik et al. 1976). The Pannonian radiometric ages (9 Ma) are reported from some andesite dikes occurring on the top of the Mount Vihorlat. Similar ages have dykes and necks of basaltic andesites and dacites at locality Dubník in the Slanské vrchy Mts. (Slávik et al. 1976, Kaličiak et al. 1991, 1996, Žec & Ďurkovičová 1993). The products of coeval explosive andesitic volcanism are represented by the Albínov Tuff northward of Sečovce. Acid volcanism was active on the southern periphery of the Vihorlat Mts. The rhyolite body of Hrádok Hill at the town of Michalovce seams to be Pannonian in age (Márton et al., 2000).

Opinions on the occurrence of the Pontian deposits in the basin vary. Janáček (1959) assigned to the Pontian the variegated formation which was later defined as **Senné Formation** by Vass & Čverčko (1985). It occurs in the middle and SE part of the basin, partly it also extends into the N part of the basin (Fig. 6). The maximum thickness of the formation is up to 600 m. It crops out on the Pozdišovce horst between Pozdišovce, Trhovište and Lesné. A member of the Senné Formation, the Pozdišovce Gravel consisting of gravel and interlayered variegated clay extends on the northern margin of the basin. It is interpreted as fluvial-deltaic deposit of palaeorivers Ondava and Laborec. The gravel beds are up to 30 – 40 cm thick and they almost exclusively consist of pebbles derived from the Outer Flysch - mainly sandstone, minor chert, occasionally Neogene rhyolite. The andesite clasts do not occur (Vass & Elečko 1977). Basinward the thickness of the Senné Fm. increases but the gravel pinches out. The main volume of the formation is composed of variagated clay.

Jiříček questioned the Pontian age of the Senné Formation and he assigned it to the Pannonian but later he admitted the correlation of the formation with the Pontian. Based on conception by Jiříček (1972) and proves by Slávik (1974), mainly emphasizing the absence of the andesite clasts in the Pozdišovce Gravel, Vass & Čverčko (1985) assigned the Senné Formation to the Pannonian. However, it is difficult to substantiate the basinwide hiatus during the Pontian. The deposits are also lithologically similar (especially variegated kaolinic clays) to the Poltár Formation extended in the Southern Slovakia and Moldava Depression, which age is well proved (Planderová 1986, Balogh et al. 1981). From this reason we prefer the Pontian age of the Senné Formation though there are not unambiguous biostratigraphic and other chronostratigraphic evidences.

A hiatus and erosion might be expected between the Sečovce and Senné Formations, especially in the marginal parts of their extension. The transgressive character of the Senné Formation is evidenced by its transgressive and disconformable position above the older Miocene formations in the northern part of the Trebišov depression. The tectonic events - slight folding of the Sečovce Formation deposits also suggest possibility of hiatus and discordance between the Sečovce and Senné Formations.

In partial Čečehov Depression (occurring from Michalovce to Vysoká n.Uhom) the upper part of the formation passes into the **Iňačovce Member**. It consists of gray clay with layers of coaly clay and lignite. Sporadically occurring gravel does not contain andesite clasts and it resembles Pozdišovce Gravel.

The Senné Formation contains poor fresh-water fossil assemblage of *Limax crassus, Valvata cf. variabilis, Candoniella albicans, Candoniella sp.III* (Jiříček 1972).

Pliocene (5.2 – 1.8 Ma; Map 10)

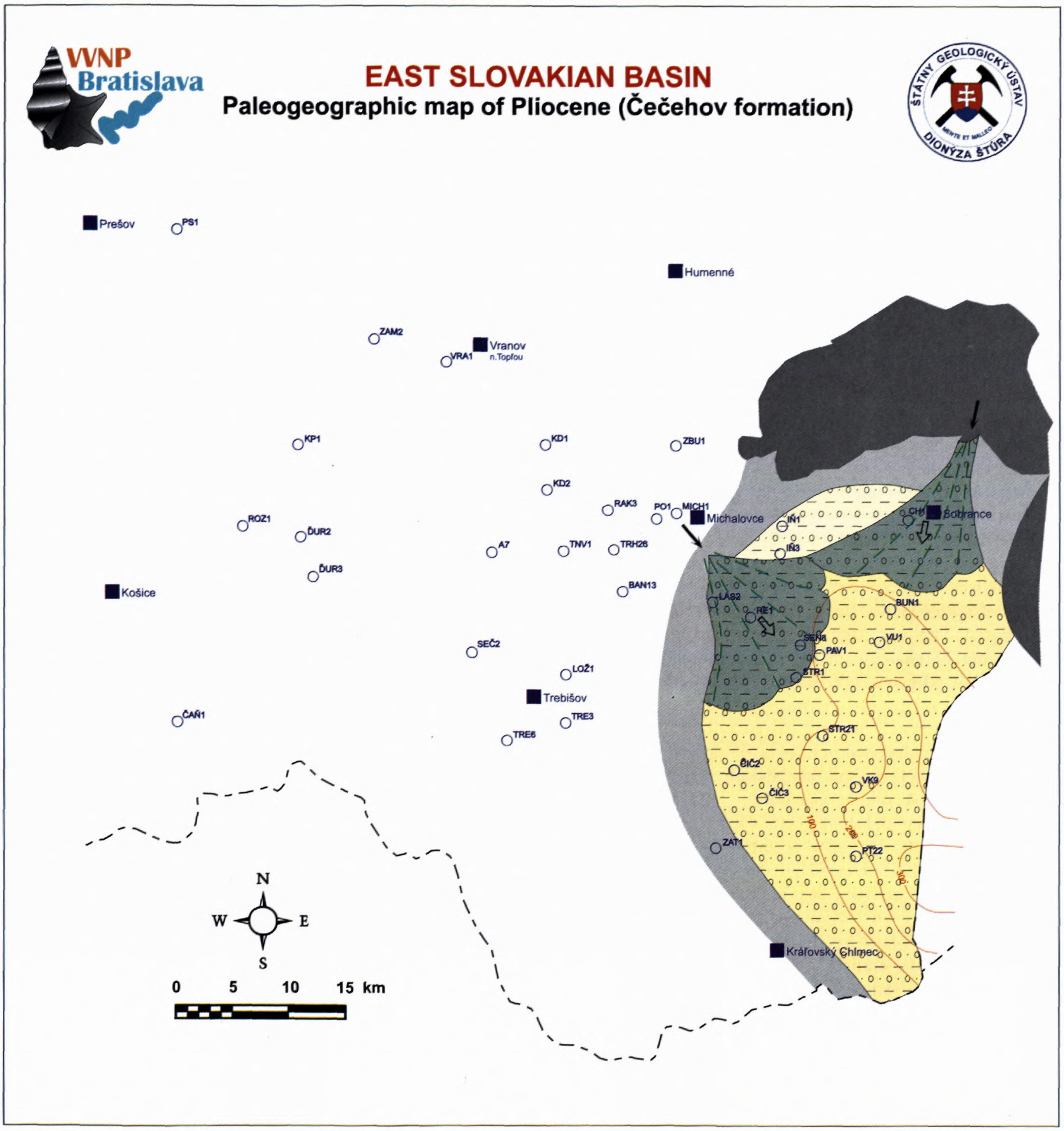
The basin markedly diminished in the Pliocene. Lacustrine deposits from this stage comprise **Čečehov Formation** discordantly overlying the Senné Formation (Fig. 6). The transgressive character is suggested by the position of the Čečehov Formation above the Hnojné Member of the Pannonian age in the Sub-Vihorlat Depression. The formation is about 120 – 200 m thick. It is composed of variageted clay containing layers of andesite gravel, sand and reworked tuff. Clastic andesite material was delivered to fresh-water lake by creeks draining the volcanic Vihorlat Mountains. Small deltas originated in the northern margin of the basin.

Fauna of the Čečehov Formation is very rare, the assemblages consist of fresh-water organisms rests (*Candoniella sp. III, Candona candida, Cyclocypria globosa, Cypria candonaeformis, c. tambovense, Planorbis sp.*) most probably Pliocene in age (Jiříček 1972). Pollen spectrum has a Pliocene in character too (Planderová in Baňacký et al. 1989).

Discussion

Ideas on the structure of the East-Slovakian Basin pre-Neogene basement vary. Soták et al. (e.g. 1993) consider the Iňačovce - Kritchevo Unit as the Penninicum. They mainly reason by lithologic similarity, grade of metamorphosis and age. Vozár et al. (1993) correlates the same unit with the Szolnok - Maramures flysch (Tissia). This is based on its position at rear of the Carpathian units, heterogenity of metamorphic rocks, stratigraphy and lithology. We accept this opinion and we believe that the unit together with the Zemplinicum escaped from the Pannonian area into the Carpathians by a large sinistral strike-slip (Fig. 3) anticipating CCW rotation of blocks underlying the East-Slovakian Basin.

Vass et al. (1988) assume that the East-Slovakian Basin was opened principally by pull-apart mechanism. Based on brittle deformation analyses, Kováč et al.


VVNP Bratislava
EAST SLOVAKIAN BASIN
Paleogeographic map of Pliocene (Čečehov formation)
ŠTÁTNY GEOLOGICKÝ ÚSTAV DIONÝZA ŠTÚRA
MENTE ET MALLEO
Prešov
PS1
Humenné
ZAM2
Vranov n.Topľou
VRA1
KP1
KD1
ZBU1
KD2
RAK3
PO1
MICH1
Michalovce
Sobrance
ROZ1
ĎUR2
ĎUR3
A7
TNV1
TRH26
Košice
BAN13
SEČ2
LOŽ1
Trebišov
ČAŇ1
TRE3
TRE6
LAS2
RE1
BUN1
VU1
PAV1
STR1
STR21
ČIČ2
ČIČ3
VK9
ZAT1
PT22
IŇ1
IŇ3
Kráľovský Chlmec
N
W
E
S
0 5 10 15 km

(1994) reason that at the beginnning of the Karpatian the opening of the basin resulted from heterogenous extension caused by astenosphere upheaval in the Pannonian domain. After deposition of the Soľná Baňa Formation the basin started to open by pull-apart mechanism. Kováč et al. (1995) define the basin as a back-arc basin. Later (Kováč et al. 1996) they assume that during the Karpatian and Early Badenian the basin was opened by a pull-apart mechanism, and later it became extensional back-arc basin although their figure 3 on page 13 suggests the intra-arc position of the basin.

Palaeomagnetic analyses performed on sediments and volcanics suggest their CCW rotation. Rotation occurred in several pulses while remanent magnetism of the oldest Neogene rocks indicates total rotation about 80° (Márton et al. 2000). According to several published data CCW rotation is preceded by sinistral strike slip and CW rotation is by dextral strike slip (Terres & Sylvester 1981, Sengor et al. 1985, fide Allen & Allen 1992). This is not consistent with Kováč & Márton (1998) relating dextral strike slip between the ALCAPA and Tissia - Dacia microplates to the CCW rotation. We presume that this rotation may be related to the sinistral strike slip between the Pelsó Unit and Central Western Carpathians and Eastern Alps along Raaba - Hurbanovo - Plešivec - Rožňava line. We assume that CCW rotation of blocks underlying the East-Slovakian Basin was preceded by sinistral strike slip along the Hornád Fault System enabling lateral escape of units belonging to the Tissia (Zemplinicum, Iňačovce - Kritchevo and Ptrukša Units) into the area of the East-Slovakian Basin (Fig. 3). We did not include rotation into our palaoegeographic maps due to small amount of analyses in the basin and northward of it (Pieniny Klippen Belt, Outer Flysch Zone). However, backward rotation of the basin into its original position may help to elucidate some palaeogeographic relation and therefore we depicted in maps 1 - 8 sphere sectors representing direction and magnitude of basin rotation according to its original position.

The present palaeomagnetic data suggest that the first rotation ca. 20°-25° CCW occurred in the Karpatian during stress relaxation causing extension (deposition of Teriakovce and Soľná Baňa Formations). The following rotation ca. 17°-20° CCW occurred in the Middle Badenian when salt-bearing Zbudza Formation was deposited. The third, Late Sarmatian rotation with magnitude about 40° CCW was terminated before the Pannonian. It is suggested by non-rotated Pannonian rhyolite extrusion the Hrádok Hill nearby Michalovce. The palaeomagnetic measurements do not support one-event, ca. 80° CCW basin rotation after or during the Ottnangian suggesting by some authors (e.g. Baráth et al. 1997).

Conclusion

The East-Slovakian Basin represents an autonomous, eastern part of Transcarpathian Basin. It occurred behind the accretionary wedge rising due to convergency of the European Platform and ALCAPA microplate. Since the Late Badenian, when subduction-related volcanic arc evolved (Lexa & Konečný, 1998), the basin may be classified as an interarc basin. Due to oblique subduction the basin had complex tectonic history represented by extensional (Karpatian, Early Badenian, Late Sarmatian, Pannonian), transtensional (Middle and Late Badenian, Early Sarmatian), transpressional (Eggenburgian) and compressional (Otnangian, Pontian) regimes. The basin is mostly filled by Neogene, shallow-marine deposits containing caustobioliths and evaporites. Volcanics also comprise a significant portion of the basin fill. The spatial distribution of deposits, depositional palaeoenvironments and volcanism, depicted in maps representing ten time slices throughout the Neogene, were strongly determined by subsidence history, sea level fluctuation and sediment input.

For the basin opening, an important role played tectonic escape of the Tissia units by the left lateral motion along the Hornád Fault System into the area of the basin. Beside the basin opening the motion of the basin triggered the later CCW litosphere block rotations attaining a cumulative value 80°. Other tectonic phenomena taking part in the basin opening and evolution were strike-slip and normal, predominantly synsedimentary faults. This tectonic character determined prevailing pull-apart character of the basin. This interpretation is also suggested by thick basin fill (8 000 - 9 000 m) deposited in relatively short time period in depocenters shifting from NW to SE. Compressional shock at the end of the Miocene, recorded by slight folding of the fill, definitely terminated pull-apart character of the basin.

Volcanism significantly influenced the basin palaeogeography by both; large volume of volcanic rocks belonging to the basin fill and building a volcanic mountain chains playing a role of barrier for the transport of clastics in the basin. Eustatic and relative sea level changes as well as tectonic movements were responsible for opening and closing of the sea straits connecting the basin to the open sea.

Progressive decrease of salinity at the end of the Miocene determined hyposaline, and finally fresh-water conditions. In the Pliocene the history of the East Slovakian Basin was definitely achieved.

Acknowledgement

The work is a contribution to the project „Tectogenesis of sedimentary basins of the Western Carpathians„ financially supported by Ministry of Environment of Slovak Republic. Additional support was provided by Grant VEGA Project No. 1/52222/98.

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Slovak Geol. Mag., 6, 4(2000), 409–418

Kinematic Evolution of the Central-Carpathian Paleogene Basin in the Spišská Magura region (Slovakia)

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*Abstract.*The Subtatric – Ružbachy Fault System played a considerable role in the Cenozoic evolution of the northern part of the Central-Carpathian Paleogene Basin. The system, which represents a shear zone with NE – SW direction, restricts the eastern continuation of the Mesozoic rocks of the Tatra Mts., bounds the Mesozoic Ružbachy Island to the Paleogene deposits and it has governed the deposition in the area since the Paleogene. In the area neighbouring to the system we distinguished four deformation stages connected with 1) NNW-SSE compression resulting in NE-SW trending thrusts, 2) E-W compression resulting in strike-slip movements along the system, 3) NNE-SSW compression resulting in WNW-ESE overthrusts and 4) NW-SE extension associated with normal fault formation. Rock sequences record activity of the system from the Eocene and Oligocene up to the Recent.

Key words: Central-Carpathian Paleogene Basin, Subtatric-Ružbachy Fault System, structure, tectonics, sediments, basin evolution

Introduction

The striking Subtatric - Ružbachy Fault System (ST - R in the following) trending from the SW to the NE and bounding the Tatra Mts. to the deposits of the Central-Carpathian Paleogene Basin has been subject of geological investigations and discussions for many years. The system sharply bounds the Paleozoic and Mesozoic units of the Tatras Mts. to the Paleogene deposits in the south and uplifted Ružbachy Mesozoic Island in relation to the Paleogene deposits on its both sides. This system, which we call in this paper Subtatric - Ružbachy Fault System in order to emphasize its geographic position, has played a considerable role in the geologic evolution of the studied area. It governed uplift of the Tatra and Spišská Magura regions, accumulation of thick Quaternary deposits on the foothill of the Tatra Mts. and probably also evolution of the Paleogene turbidite system in the Central-Carpathian Paleogene Basin. Because the ST-R Fault System is still active, it is interesting not only from the viewpoint of basic geology, but also from the viewpoint of applied geological disciplines.

The aim of the presented paper is to analyse tectonic structures in the area of the Spišská Magura region and to interpret these structures in order to get kinematic evolution of the area. This should also provide new knowledge on character and chronological succession of brittle deformation along the ST - R Fault System.

Geological setting

The Spišská Magura Region is a part of the Central-Carpathian Paleogene basin (CCP Basin) and represents the northernmost part of the Central Western Carpathians (Fig. 1). It consists of the Paleogene deposits transgressively overlying Mesozoic rocks of the Krížna Nappe, and tectonically bounded in the north by the Pieniny Klippen Belt and in the east by the East-Slovakian Neogene Basin. The deposits have a wide stratigraphic span ranging from the Middle Eocene to the Late Oligocene (e.g. Gross et al. 1999, Janočko et al. 1998, Janočko & Jacko 1999) and they prevailigly consist of turbidites filling the CCP Basin. Only the lowermost part of the basin fill is composed of continental and shallow-marine deposits, mostly consisting of breccias, conglomerate and nummulitic sandstone and limestone (Borové Formation). The overlying deposits comprise Huty (Middle Eocene - Late Oligocene) and Zuberec (Late Eocene - Late Oligocene) Formations (Gross et al. 1984, Janočko & Jacko 1999). The Huty Formation prevailingly consists of dark shale sandwiching conglomerate representing a fill of a canyon incised into shales. The Zuberec Formation consists of alternating shale and sandstone representing channel and levee and interchannel deposits of turbidite systems (Janočko & Jacko 1999, Fig. 2). The CCP Basin is usually defined as a forearc basin (e.g. Soták et al. 1996), however, the unambiguous evidence proving the basin position is still missing. From this reason we define the CCP Basin in this paper as a basin with a complex, prevailingly extensional and subordinate compressional history.

The structure of sediments, their succession and kinematics imply a complex character of tectonic evolution in the eastern part of the CCP Basin (c.f. Hrušecký et al., 1995; Sperner & Ratschbacher, 1995). The complexity of structural evolution of the basin is most pronounce in the studied region where thrusts, normal faults and strike-slip faults occur.

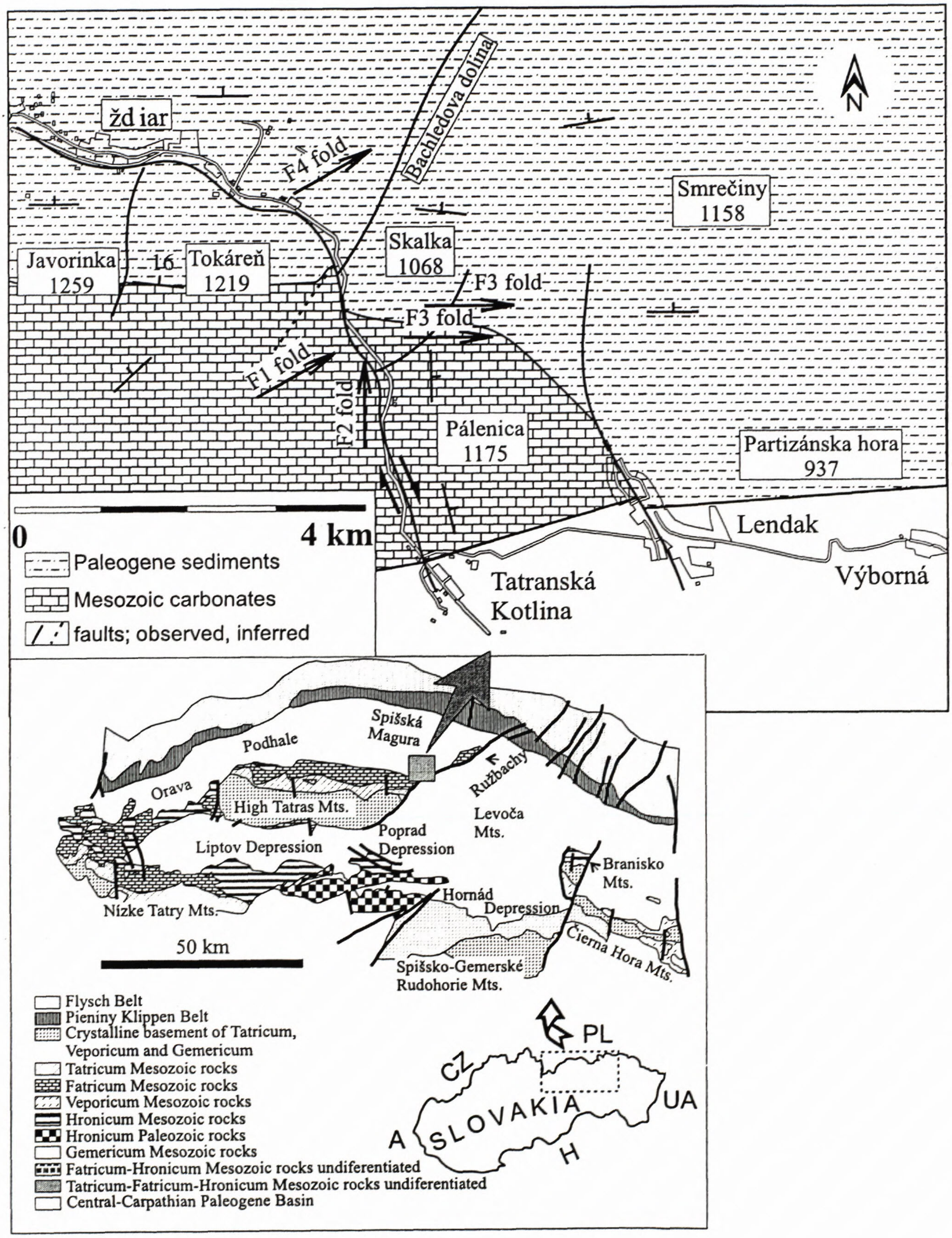


Fig. 1: Location of the studied area and schematic map with geological structures among Ždiar, Tatranská Kotlina and Lendak. Position of the Central-Carpathian Paleogene Basin is also shown.

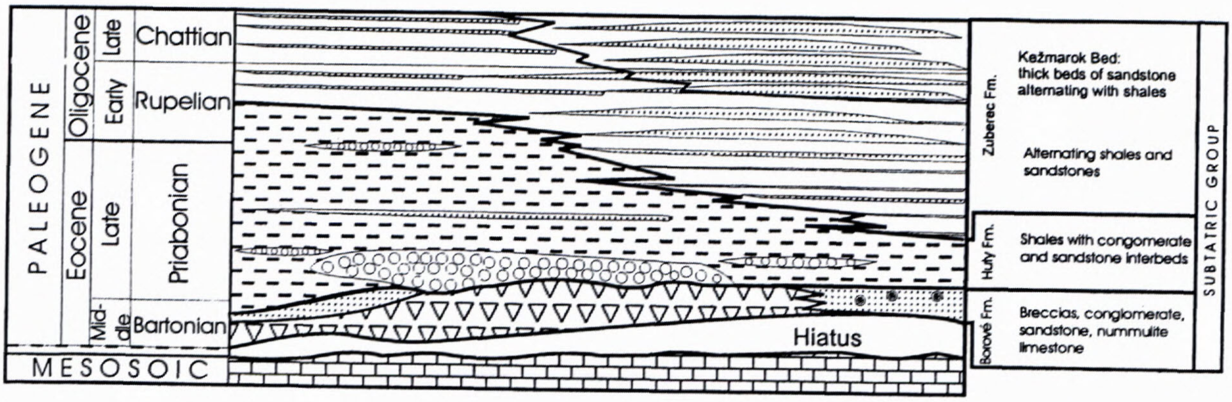


Fig. 2: Lithostratigraphic column of the Paleogene deposits in the studied area

The most expressive structure in the studied area is represented by ST - R Fault System (Fig. 1). It is a brittle shear zone having NE-SW strike and steep dip toward SE. It tectonically restricts the southern and eastern parts of the Tatra Mts. and probably passes to the Krynica Unit of the Outer Carpathian Flysch Zone in NE. In the area of Ružbachy the system restricts a horst of the underlying Mesozoic rocks belonging to the Krížna Nappe. Most of the authors up till now considered the ST - R Fault System as a reverse fault structure (Koutek, 1931, 1936; Matějka, 1935; Gorek, 1954; Andrusov, 1958, 1968; Chmelík et al., 1963; Fusán et al., 1963; Maheľ et al., 1967; Gross, 1973; Nemčok et al., 1993; Hrušecký et al., 1995).). Based on the description of the borehole CH – 1, located SE of the studied area, Gross et al. (1980) documented normal slip kinematics of at least a part of the system. Timing of the activity of the system is constrained by apatite fission track ages of the Tatra granitoids which yielded 15 Ma (Kráľ 1977) and 10 - 19 Ma (Kováč et al. 1994), respectivelly, suggesting Early to Middle Miocene. Assumed amplitude of the Neogene uplift of the Tatra Mts. is about 2 700 m and amplitude of the Quaternary uplift is assumed to be about 400 m (Nemčok et al., 1993). However, the recent results from the sedimentological analyses of the CCP Basin suggests occurrence of a submarine high located in the same direction and area like the ST-R Fault System suggesting possible older activity of the system. The occurrence of the high is proved during the deposition of the Zuberec Formation i.e. in the time span Late Eocene - Late Oligocene (Janočko & Jacko 1999). Strike-slip movements were recorded in several sections of the ST-R Fault System (Nemčok et al. 1993, Hrušecký et al. 1995, Sperner and Ratschbacher 1995, Sperner 1996) suggesting an even more complex kinematics.

Methods

The construction of the geologic map from the Spišská Magura region (Janočko et al. 2000, Janočko & Jacko 1998) required a detail structural analysis of the Paleogene deposits and their basement in order to understand the structural evolution of the area. Based on the analysed structures, we divided the whole region in three, generally NW-SE elongated domains (Fig. 6B). The first domain, consisting almost exclusively of carbonate belonging to the Krížna Napppe, occurs in the SE part of the region between Tatranská Kotlina and Lendak. The second domain, composed of carbonate of the Krížna Nappe and deposits of the Paleogene Borovo and Huty Forma-

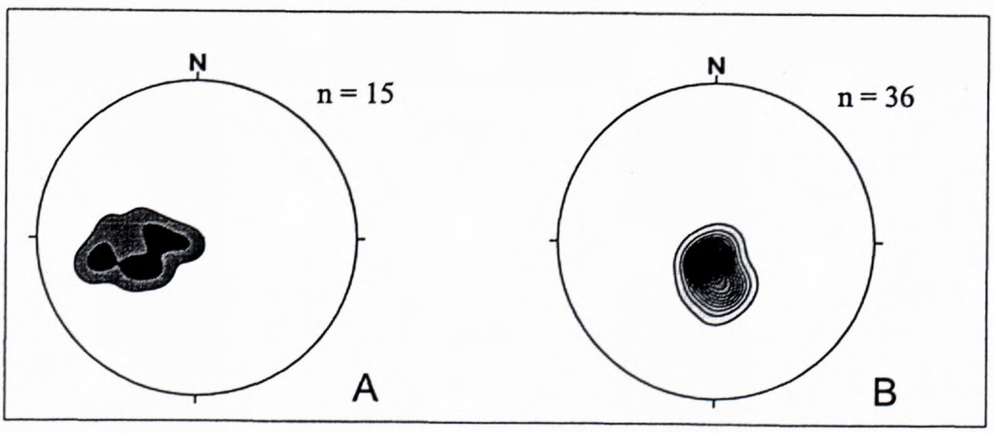


Fig. 3: Distribution of bedding poles: a – Mesozoic carbonates; b – Paleogene deposits
The figure represents non-folded data at lower hemisphere projection

tions, occurs in the central part of the region. The third domain, extending between Ždiar and Bachledová dolina, consists of the Central-Carpathian Paleogene deposits overlying carbonates of the Krížna Nappe. We studied synsedimentary depositional structures indicating the filling of the basin (Janočko & Jacko 1998), geometry of mesoscopic structures yielding primary data for structural analysis, bedding, fold structures, cleavage and brittle shear zones. Geometric classification of folds has been based on the Hudleston (1973) principle of visual analysis of fold structures. Statistic evaluation of the obtained structural data also provided a comprehensive information on succession and spatial - lithological relations of the individual measured sections and relations between the sections. The orientation of palaeostress axes was deduced by application of Angelier's „direct inversion method„ (Angelier 1990).

Description of structural domains

Southeastern, marginal domain

The pre-Paleogene structures, necessary for interpretation of kinematic evolution of the area in the Paleogene, were divided on the base of the structural analysis of the underlying Mesozoic rocks. The area SE of the CCP Basin consists of the Krížna Nappe. It extends among Tatranská Kotlina, Lendak and Toporec, its northern boundary occurs about 500 m south of the elevation point Skalka (1068 m. a. s. l., Fig. 1). The Krížna Nappe prevailingly consists of Ramsau Dolomite (Nemčok et al. 1993) with characteristic bedding parting and occurrence of intensively tectonized dark shales. The bedding of dolomite has N - S direction and dip about 50° toward W (Fig. 3A). The most conspicuous overthrust structures are penetrative, often conjugated and they have NE-SW strike and dip 30°- 50° toward NW and SE. The NE - SW trending dextral strike-slip structures steeply dip (75°-85°) toward NW. The succession of structures ends with extensional structures represented by normal faults of N - S and NE - SW strike dipping toward E and SE. They are often filled by calcite which forms 3 mm thick veins.

Central domain

The central domain occurs in the surroundings of Tokáreň and Skalka (1 068 m.a.s.l., Fig.1). It consists of the Triassic carbonates comprising Carpathian Keuper, Fatra Formation and Kopienec Bed belonging to the Krížna Nappe (Nemčok et al. 1993). These rocks are transgressively overlain by basal Paleogene deposits (Borové Formation, Gross et al. 1984, Janočko et al. 1999) and capped by deposits of Huty Formation.

The Mesozoic deposits have commonly N - S bedding dipping 5° - 30° toward W. The Paleogene deposits have general WNW - ESE bedding diping about 5°-25° toward NNE (Fig. 3B).

We find fault structures common for Mesozoic and Paleogene deposits of the studied area. They are represented by dextral strike-slips having E - W and ESE - WNW orientation and dip 45°- 80° toward S and SSE. The normal faults have E - W direction and they dip with 40°- 65° toward S. Conspicuous extensional fault system suggests NW - SE direction of $\sigma 3$. The extensional structures are filled by calcite which forms up to 3 mm thick veins of NE - SW direction dipping about 60°- 80° toward SE. Another conspicuous structure is sinistral strike - slip system having WNW - ESE orientation with 75°- 86° dip toward NNE and SSW.

F1 and F2 fold structures were identified in the Triassic carbonates of the Krížna Nappe in this domain. The axes of the F1 folds have 5^0 dip toward NE (Fig. 4A). The folds are typically developed in the Carpathian Keuper of the Krížna Nappe. They are larger than 1 m and represent parallel, slightly open folds of 1B class according to Ramsay (1967) or 1D folds according to Hudleston (1973). They only occur in the surroundings of Tokáreň.

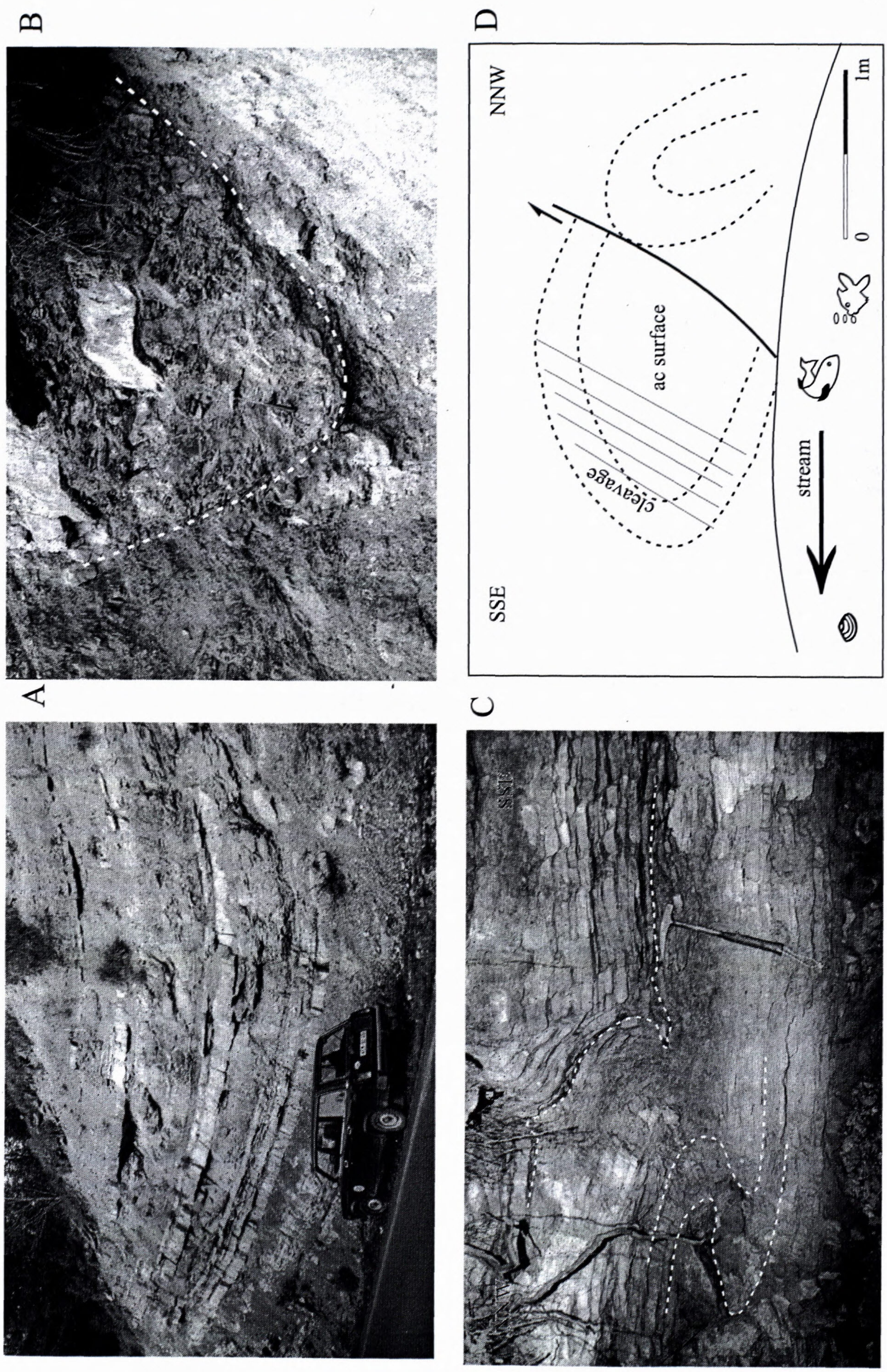
The subhorizontal dip of fold axes of the relatively older mesoscopic folds (F2, Fig. 4B) is15°- 45^0 toward the North. The folds are asymetric, often disharmonic, parallel, uprighted, inclined and they belong to the class 2D sensu Hudleston (1973).

The axes of the folds of F3 type (Figs. 4C and 4D) are oriented from N to S and they dip with 10^0 - 35^0 toward E. The folds are recumbent, parallel, compressed, isoclinal (Fleuty 1964) and according to Hudleston (1973) they represent fault propagation folds with south vergency of the 1C class. They are associated with dark mudstones with sandstone interlayers of the Huty Formation. They have also been observed in marginal parts of the Carpathian Keuper.

Northern marginal domain

The domain occurs between Ždiar and Bachledová dolina. It only consists of Huty Formation deposits dipping with 20°-30° toward N (Fig. 3B). The geologic structure of the area is strikingly influenced by normal fault tectonics. The movement occurred along NW - SE extensional faults dipping 65°- 80° toward NE and SW, occasionally it also originated along WSW - ENE faults dipping 70°- 80° toward NNW. Two systems of inverse fault structures are typical in the area: the first one is represented by NW - SE inverse faults dipping 80°- 85° toward SW and the second one is composed of ENE - SWS inverse faults dipping about 70^0 toward NNW. The last fault structures in this area are conjugated fault pair with NE - SW and E - W strike dipping 65° toward S and SE.

Fig. 4: Representation of fold structures around Ždiar area:
A - Alternation of dolomite and variegated shale of the Carpathian Keuper
B - Disharmonic folding of the Carpathian Keuper and distinct segmentations of dolomites
C - Plastic deformation in the Mesozoic shales belonging to the Carpathian Keuper and in the Paleogene deposits.
D – Sketch: isoclinal folding of Paleogene deposits


A
B
C
D
SSE
NNW
cleavage
ac surface
stream
0
1m

We found F4 folds in the area (Fig. 5A) with fold axis direction dipping 10° toward NE. The folds larger than 1 m range from overturned and recumbent to open and parallel 2D folds according to the classification Hudleston (1973).

Interpretation of structural domains

The results of structural analysis imply more deformation etapes in the studied region. They also confirm discordancy between Mesozoic basement and Paleogene deposits of the CCP Basin.

Generally, Mesozoic carbonates have N - S bedding diping toward W (Fig. 3A) and Paleogene deposits have WNW - SES direction of bedding diping toward N (Fig. 3B). The disjunctive structures have NW - SE and W - E directions corresponding to the orientation of the ST - R Fault System. The different kinematic activity of the structures reflects only regional geologic processes related to the postdepositional evolution of the basin. Different dynamics within the basin is also documented by entirely different disjunctive structures measured in the Paleogene deposits.

Southeastern, marginal domain

The oldest tectonic etape is probably related to the activation of the Subtatric - Ružbachy Fault System which is only developed in the southeastern marginal domain. The results of structural measurements suggest more deformation etapes directly connected to the tectonic activity in this extremely exponed tectonic zone. Based on the analyses we divided three deformation etapes here:

The first, probably oldest deformation etape is connected to the compressive palaeostress field with maximum compression $\sigma1$ of NW - SE direction (Fig. 6/Ia). In this palaeostress field NE - SW thrust faults, corresponding to the regional direction of the Subtatric - Ružbachy Fault System, developed. The maximal E - W extensional field $\sigma3$ resulted in origin of extensional tectonic structures of N - S and NE - SW directon, which have been filled by calcite.

The second palaeostress field (Fig. 6/Ib) is characterized by E - W compression resulting in dextral NW - SE strike-slip movements. The tensional axis $\sigma3$ has NNW - SSE direction.

During the third, youngest tectonic etape (Fig. 6/Ic), mainly the NW - SE extensional stress $\sigma3$ was active. It gave the rise to NE - SW normal faults with joints often filled by calcite.

The palaeostress measurements from a vicinity of the ST - R Fault System confirmed a polyphase tectonic activity of the system which started in the Middle and Late Miocene or even earlier, in the Paleogene. The activity of the system in the Paleogene is suggested by change of palaeoflow directions of the Paleogene turbidites along the fault course. This change was probably caused by existence of submarine high stretching from the Štrba area along the foothill of the Tatras Mts. to the Ružbachy Mesozoic Island (Janočko et al. 1999 in press) which determined evolution of two separated depositional systems on its both sides. The activity of the system since the Miocene is proved by evidence of the High Tatras Mts. uplift during this period (Kráľ 1977, Nemčok et al. 1993).

Based on structural analysis of the domain it is also possible to recognize a conspicuous cyclic succession where rotation of compressional $\sigma1$ field occurs. It is also confirmed by change of angle of maximum compressional stress $\sigma1$ in the marginal part of the studied area, which evolves from SE to E and NNE in anticlockwise direction. The maximal extensional stress $\sigma3$ rotated from NW to SE also in anticlockwise direction. Based on these results we think that in the studied region a gradual rotation took part which may be related to the transtensional movements in the High Tatras Mts. region.

Central and northern domain

The tectonic situation is different north of the ST - R Fault System, in the central and northern marginal domain, where tectonic deformation was not so intense. However, also here we can observe indications of polyphase tectonic deformations most probably related to the tectonic activity on the southern margin of the mountains. The structural analyses in this part of the region suggest four tectonic phases:

The first, probably the oldest phase, is related to the activity of extensional stress $\sigma3$ with NW - SE direction (Fig. 6/IIa). The structures are characteristic for both Mesozoic and Paleogene deposits. They consist of normal faults having NW - SE and E - W directions. Kinematic response of this paleostress field is also conjugated Fault System with hanging walls of NW - SE direction dipping 80^0 toward SE and footwalls of N - S direction with 40° dip toward W.

The second palaeostress field is characteristic by maximal compressional stress σ in ENE - WSW direction and maximal extensional stress $\sigma3$ in NNW - SSE direction. In this palaeostress field associated pair of dextral strike-slips of ENE - WSW direction and sinistral strike-slips of WNW-ESE direction originated. All these strike-slips have character of Riedel shears R and R` (Fig. 6/IIb). A dextral strike-slip in the main shear zone of NNW - SSE direction was formed during an increased deformation (Fig. 1). This zone is followed by creek Belá between Skalka and Tokáreň (Fig. 1). We also relate formation of F2 folds with fold axis diping toward north to the activity of this palaeostress field.

The third palaeostress field in the central part of the area is characterized by maximal compressional axis $\sigma1$ in NNE - SSW direction and maximal extensional component $\sigma3$ in WNW - ESE direction (Fig. 6/IIc). The products of this stress field are WNW - ESE overthrust structures diping toward NNE as well as extensional structures of ENE - WSW direction. On the overthrust structures F3 fold structures of fault propagation type and duplex structures were formed.

The fourth palaeostress field is characteristic for the northern marginal domain of the studied region, the Paleogene deposits crop out (Fig. 6/III). In this palaeostress field normal fault structures of ENE - WSW and WNW - ESE direction were formed which were determined by $\sigma 3$. These structrures are expressed by folded beds (Fig. 5B). Along the normal faults, induced by $\sigma 1$ compression of NNW - SSE direction, the Paleogene deposits were gradually uplifted and folding generating F4 folds (Fig. 5A).

The analysis of discussed palaeostress fields suggest, similarly to the southern part of the region, rotation of the maximal compressional stress $\sigma 1$ counterclockwise from the W toward SSE. The extensional component $\sigma 3$ is not evolved so regularly like in the surroundings of the Subtatric - Ružbachy Fault System because is was not originated at one tectonic line.

Based on the structural analysis, it is possible to suggest correlation of the second deformation phase between the southern and central part of the region (Fig. 7). The deformation phases in both regions are characteristic by strike-slip faults. Also the third deformation phase of the southeastern and central domain shows analogous relations. In this part extensional component prevails resulting in normal faults.

Discussion and conclusion

The Spišská Magura region between Ždiar, Tatranská Kotlina and Lendak, consists of Paleogene deposits (Middle Eocene - Oligocene) unconformably overlying carbonates of the Krížna Nappe. The main geologic evolution was influenced by the ST - R Fault System representing the most striking structure of the region. This shear zone with NE - SW direction and steep dip toward SE restricts the eastern continuation of the Tatras Mts. and continues toward NE. In our study we found that several deformational indicators in the entire studied area are closely related to the kinematics at this structure.

The oldest deformation phase, influencing the evolution of the region, is NNW - SSE compression occurring in the SE part of the studied area. The compression induced activity of the Subtatric - Ružbachy Fault System along which Paleogene deposits were uplifted. This resulted in the subsequent erosion of the Uppermost Oligocene deposits which were described in the Levoča Mts. south of the Subtatric - Ružbachy Fault System (Soták et al. 1996, Janočko et al. 1998). The effect of compression is most striking in the vicinity of the Subtatric - Ružbachy Fault System where the uplift amplitude was highest. In this area also NE - SW overthrust structures originated diping toward NW and SE. In the central part of the studied region smaller uplift amplitude was observed. The compression is manifested by NE verged F1 folds occurring in the Carpathian Keuper deposits. However, it is

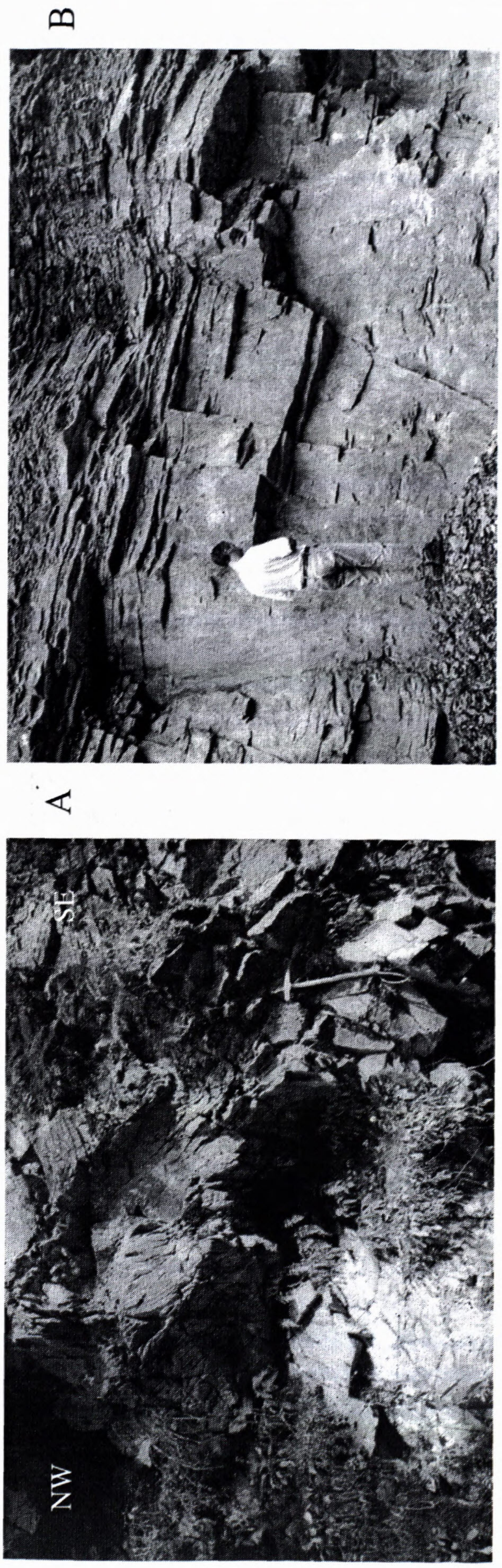


Fig. 5: Photo of A) fold structure in the Paleogene rocks with different competence in the Bachledova dolina, B) Conjugate pair of normal faults with little deflection in layers of thick sandstone beds (marker bed in the study area) in Bachledova dolina.

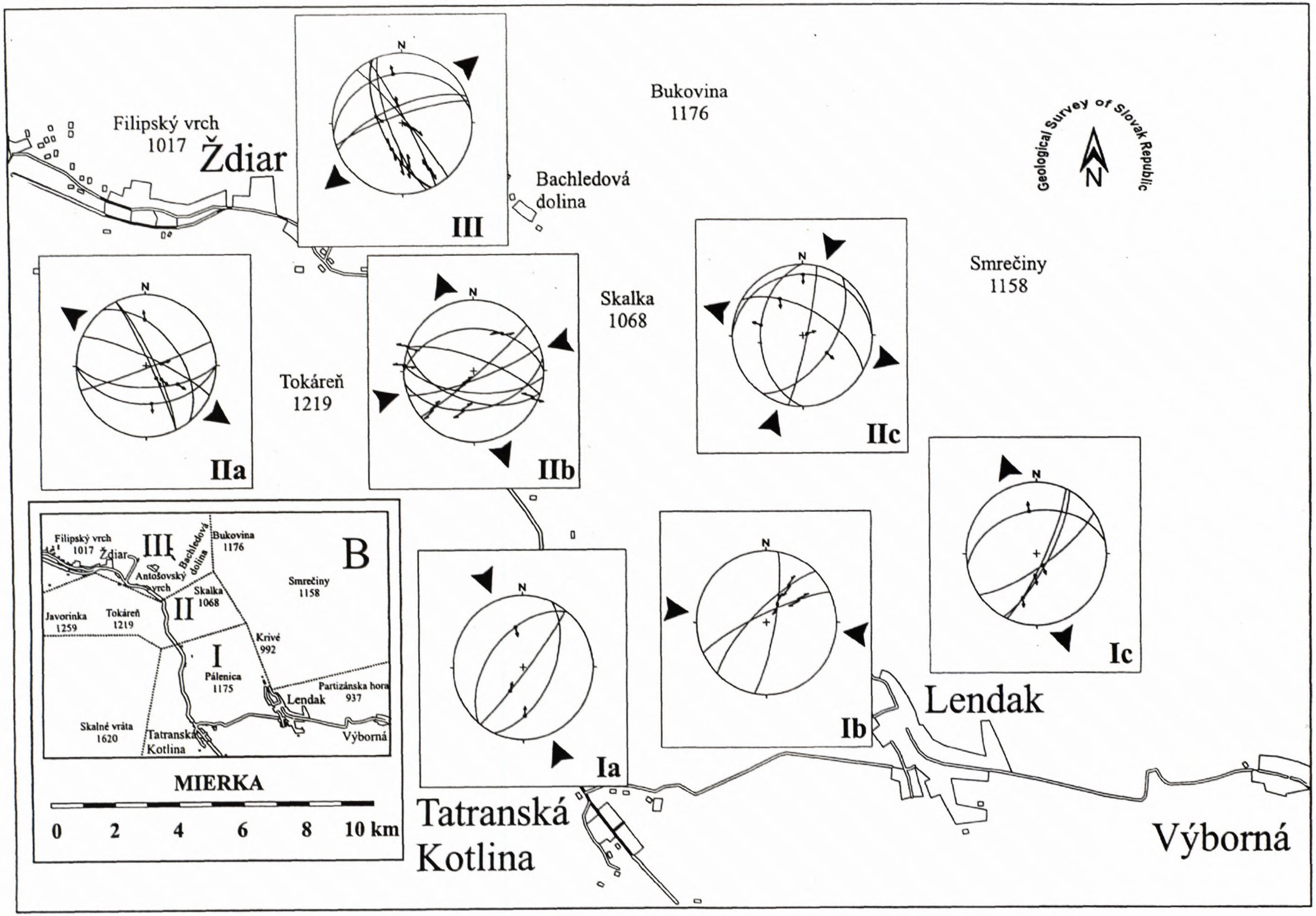
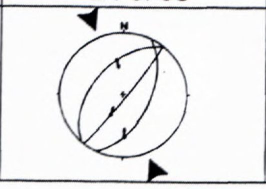
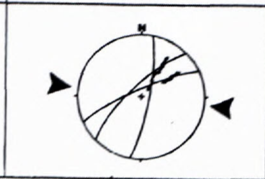
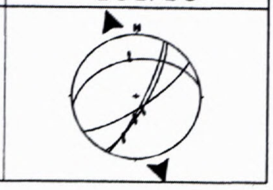


Fig. 6: Stress tensors and localization of structural domains

	Ia	Ib	Ic
σ1	157/12	97/18	13/72
σ2	61/24	201/36	254/9
σ3	270/63	346/48	161/16
Paleostress field			

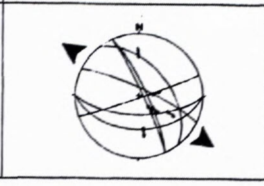
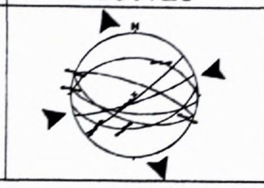
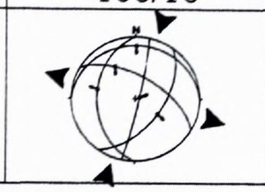
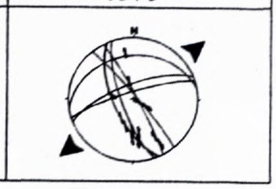
	IIa	IIb	IIc	III
σ1	263/67	253/10	200/25	160/66
σ2	29/14	5/65	356/63	316/23
σ3	124/18	159/23	106/10	49/9
Paleostress field				

Fig. 7: Summary table of calculation of paleostress poles and their values in the study area.

probable that initial compression in the central and northern part was replaced by extension inducing subsidence of the CCP Basin fill. This is proved by NW - SE normal fault structures oriented perpendicularly to the main compression direction and by well-preserved Late Triassic and Paleogene deposits. The age of the deformation may be is suggested by timing of the Tatras Mts. uplift (some 15 Ma, Kráľ 1977) and by existence of submarine high in the Oligocene which had the same course like the ST - R Fault System (Janočko et al. 1999 in press). The proved Quaternary uplift also suggests the recent activity of the Fault System. The Miocene uplift also determined exposure of the oldest Paleogene sequences in the Spišská Magura region due to subsequent erosion of overlying younger deposits and slight tilt of the Paleogene formation toward N.The second deformation stage in the SE part of the region is related to the E - W compressional stress σ1. It resulted in NE - SW strike-slip faults of the Subtatric - Ružbachy system with dextral movement. Their activity induced NW – SE conjugate fault structures. These structures are well observable in the SE marginal part, where they often segment the ST - R Fault System.

The activity of this fault system probably also induced maximal compressional stress σ1 of ENE - WSW direction and maximal extensional stress σ3 of NNW - SSE direction in the central part of the studied region. In this area, associated pair of ENE – WSW dextral strike-slips and WNW – ESE sinistral strike-slips originated, which have character of Riedel shears R and R´. The formation of mesoscopic shear folds of F2 type is also related to the activity of these pair translations. With increasing deformation dextral strike-slip movement originated in the main NNW - SSE shear zone between Tokáreň and Skalka. The origin of the NW - SE and NNW - SSE structures in the whole Spišská Magura region is probably related to the transpressional movements in the area of Tatra Mts. described by Hrušecký et al. (1995). Our geological-structural analysis showed that the dislocations have frequently strike-slip character with local, small normal fault occurrences. They are often associated with subparallel systems of subvertical joints filled by carbonates. The occurrence of travertine along the structures suggests their recent activity continuing from the Neogene.

The third deformation stage in the central part of the studied region is related to the maximal compresional stress component σ1 in NNE - SSW direction and to the maximal extensional stress component σ3 in WNW - SES direction. The existence of these stresses resulted in WNW - ESE overthrusting structures with NNE vergency and ENE - WSW extensional structures. On the overthrusting structures flexures, duplex structures and fold structures of F3 type originated. The fold axis, diping with 10°, is oriented toward E. There are characterized like fault propagation folds. This deformation stage which is probably of local character, is closely related to the formation of NNW - SSE dextral shear zone between Tokáreň and Skalka which we describe in the second deformation stage. Intensive tectonic activity at the shear zone induced local overthrusts consistent with dextral movements in the area. The compressional σ1 and extensional σ3 stress of this deformation etape only partly rotated from the main compressional and extensional stress field found in the second deformation etape of the central part of the studied region. We assume almost synchronous development with the NNW - SSE shear zone.

The youngest deformation stage on the SE margin of the studied area is characteristic by prevailing extensional stress component σ3 with NW - SE direction. This etape of the evolution of the Subtatric - Ružbachy Fault System is associated with normal faults of NE - SW direction. The normal fault was convincingly manifested by the borehole CH – 1 (Gross 1973). Similarly, in the central part of the studied region we recorded activity of maximal extensional stress component σ3 resulting in formation of NW - SE and WSW - ENE normal faults.

The system of NE - SW faults belongs to the youngest dislocation structures strikingly segmenting lithologically different deposits of the CCP Basin and deforming NW - SE faults. This infers the Late Miocene and/or Pliocene age of the faults.

Acknowledgement

We thank the Ministry of Environment of the Slovak Republic for supporting the project „Tectogenesis of the Sedimentary Basins of the Western Carpathians“ that led to the development of this paper. L. Fodor and P. Kováč offered valuable suggestions for improvements in their review of the paper. We also thank M. Potfaj for helpful comments on the manuscript.

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Slovak Geol. Mag., 6, 4(2000), 419–429

Picrite Rocks in the Vicinity of Banská Bystrica (Krížna unit, Western, Carpathians)

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Abstract: In the Mesozoic of the Western Carpathians several characteristic volcanic formations occur: teshenite-picrite formation in Cretaceous of the Flysh Belt, alkali basalts/basanites formation in Cretaceous of the Cover Units and Subtatric Units and tholeiite formation in the Triassic of the ophiolite complex of the Meliaticum. In this study we present the results of the petrological research on the Mesozoic picrite formation in of the broader area of Banská Bystrica. Based on the clinopyroxene composition and on their rare elements composition, we assign the picrites into a formation of alkali basalts/basanites and they occur in the Cretaceous of the Cover Units and Subtatric nappes.

Key words: picrites, mineralogy, geochemistry, Cretaceous, Subtratric nappe, Western Carpathians

Introduction

Already in the past picrites were described in various geological units of the Western Carpathians. The Silesian Unit of the Western Carpathians Flysh Belt is the district with the most common occurrences of these rock bodies (hypabysal and surficial), where picrites are a part of the already classic, teshenite-picrite formation (Pacák 1926, Smulikowski 1929 in Hovorka- Spišiak 1988).

In the broader area of Vienna (Vienna basin) more of the picrite occurrences are known (Prey 1975). According to Tollmann (1985) more than 30 rock bodies are present there.

During oil-well exploration in the northern part of the Vienna basin near Gbely (Slávik 1930), and subsequently also near Kúty (Matějka in Budaj et al. 1963), picrite bodies were drilled. According to Benešová (1957), they metamorphosed the surrounding Albian sediments.

However the picrite described in detail by Ivan (1991) from the Strážny hill near Medzev occurs in the Gelnica Unit (Lower Paleozoic) and was metamorphosed to greenschist facies.

The process of gravitational differentiation is the typically process, that leads to the origin of picrites. Heavy olivine crystals accumulate and, due to the subsequent effusion and cooling, the origin of picrite and picrite basalt type of rocks occurs. As clinopyroxenes in basic effusive rocks, and consequently also in picrites, are highly informative value from all the present silicate phases (type and crystallisation conditions of magma), and in the case of the studied picrites they are and relatively well preserved, we focused on the study of their composition using an electron microprobe.

In the association with a review of Mesozoic volcanism products in the Western Carpathians, we focused on the study of picrites and alkali basalts/lamprophyres (Hovorka – Spišiak 1988, 1993, Spišiak – Hovorka 1997, Hovorka et al. 1999). Despite their rare occurrence, picrites have a high petrologic and geotectonic importance. As a shallow (hypabysal) rock, they offer complementary data not only about the character of the volcanism, but also about the type and composition of the upper mantle in the place of picrite magmas generation.

Generally is the idea still accepted, that picrite rocks in the Mesozoic of the Western Carpathians were either Triassic, or Neogene with the basaltic rocks (Hovorka - Slavkay 1966, Slavkay 1979). From casual view the picrites do not have any equivalent among the Lower Cretaceous alkali volcanic rocks of the Central Western Carpathians, but if the Mesozoic volcanism of the Outer and Inner Western Carpathians are compared in detail, the picrite rocks are an integral part of the whole volcanic province. Hence the resolution of the character and type of the volcanism was the main aim of this study.

The picrites were studied optically and geochemically (5 samples). Minerals were analysed using the electron microprobe CAMECA (Dipartimento di Mineralogia e Petrologia, Universita di Padova), or JEOL Superprobe (GS SR) at standard conditions. Petrogenic features and trace elements were analysed by XRF, or SPA at Geological Institute Sloval Academy os Sciences. Rare earths elements, or selected trace elements were analysed using ICP at the Ibaraki University (Mito, Japan).

Geology

Horný diel near Banská Bystrica

On the southern slopes of the Horný diel (ground elevation 1000 m above sea level) about 4 km north of Banská

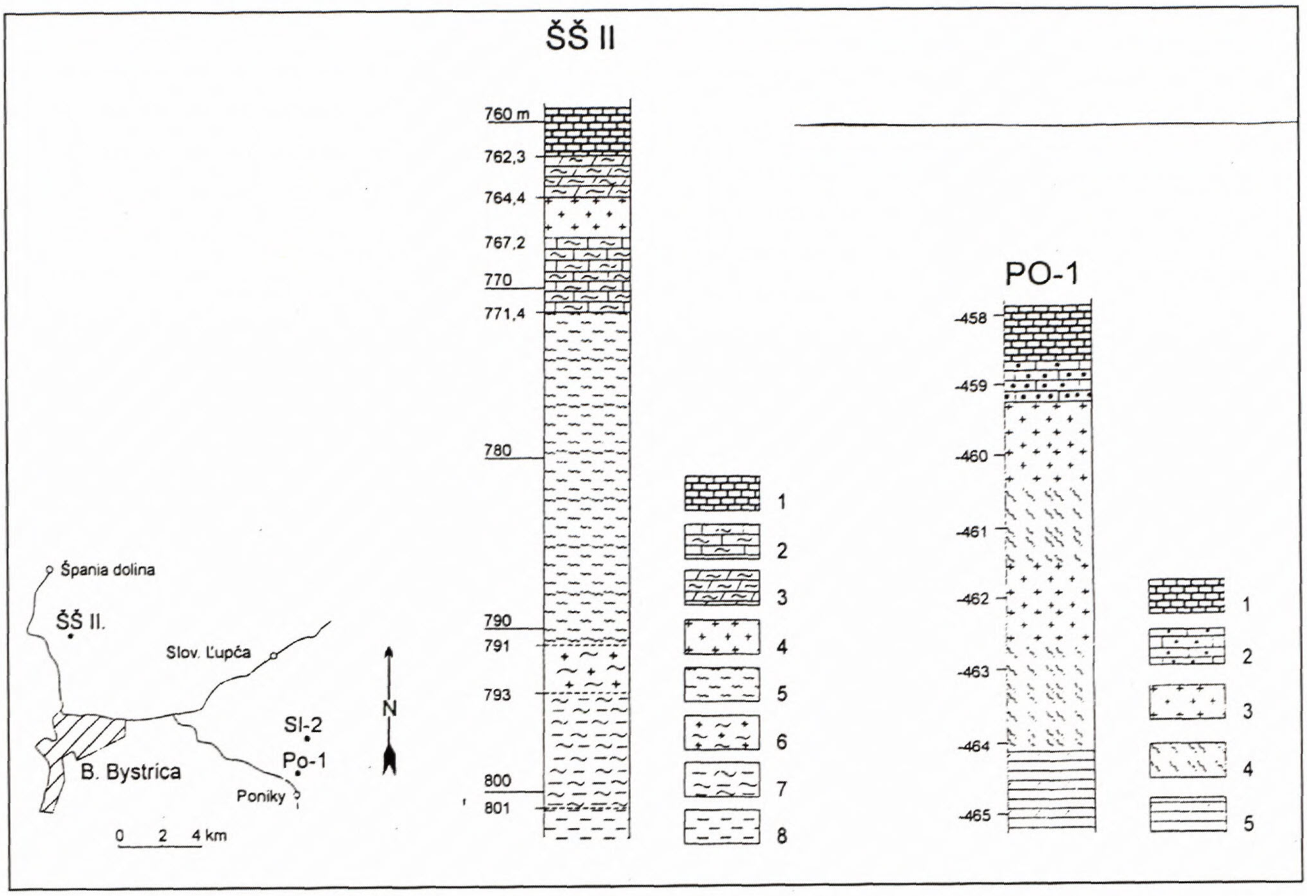


Fig. 1a. Schematic map of picrite occurrences locations (a) a part of the section of the drill hole ŠŠ-II (b) (according to Slavkay, 1979). 1 – grey limestones with interbedded fractured dolomites, 2 – grey, heavily broken limestones, 3 – tectonic breccia composed of dolomite, limestone and clay fragments, 4 – picrite, 5 – tectonic clay, fragments of limestones and claystones, 6 – heavily broken picrite from the tectonic zone, 7 – tectonic clay with fragments of violet–brown shales and sandstones, 8 – violet-brown shales, 1-3 – Middle Triassic, 5-7 – tectonic zone, 8 – Lower Triassic; Po-1, Sl-2 – indication of drill holes.

Fig. 1b. A part of the section of the drill hole PO-1 (according to Hovorka–Slavkay 1966). 1 – broken grey to dark-grey dolomites (fissures filled by calcite), 2 – broken dolomites with clastic fragments of picrite, 3 – biotitic picrite, 4 – broken biotitic picrite, 5 – broken grey dolomites alternating with layers of heavily broken bioititic picrite

Bystrica a picrite body was drilled in the drill hole ŠŠ-II (Fig. 1a) in the interval 764.4-767.2 m (Slavkay 1979). The surrounded rocks consist of Middle Triassic limestones of the lower sub-plate of the Krížna nappe (1.c.). The contact zones with ambient rocks were not obtained using the related core run due to the extensive tectonic fracturing therefore, the mode of deposition of the picrite body and its relationship to the surrounded carbonates is unknown.

Poniky district

North of the Poniky, about 8 km SE from Banská Bystrica in Middle Triassic dolomites of the Choč nappe (Bystrický 1964, in: Slavkay 1965), volcanic rocks of the picrite type were found in two drill holes. Their basic characteristic is described in studies of Hovorka & Slavkay (1966) and Slavkay (1979). In the drill hole Po-1, located about 250 m W of the northern margin of the village Poniky at a depth of about 460 m (fig. 1b), a volcanic rock body was discovered with an apparent thickness nearly of 5 m, occurring in grey dolomite breccia. Along veinlets of carbonate, pyrite and gypsum were present. Carbonates and pyrite represent a product of the overprinting hydrothermal processes. Gypsum probably is a product of migration from the overlying Lower Triassic Drienok sequence, in which layers of gypsum and anhydrite occur. Based on the presence of small glassy fragments of picrite in the superposed dolomites and based on the missing or still nodetermined contact-thermal influence of picrite on the surrounded dolomites. Hovorka and Slavkay (1966) proposed that the body represents a product of the submarine volcanic activity in the surrounding carbonate sediments. However the number of fragments decreases slowly in direction from the contact. It is not possible to determine the relationship of the picrite to the overlying dolomites from the drill core. The picrite from the immediately underlying part of the core is extensively fractured, and the tectonite is cemented by carbonates.

In the drill hole St-2, situated close to the hill Stráž (697 m) about 2.5 km north of from Poniky (approximately 2 km from the drill hole Po-1). It is a body with a consistent geological position and consistent petrographic character as the body in the drill hole Po-1 was discovered (Slavkay 1979).

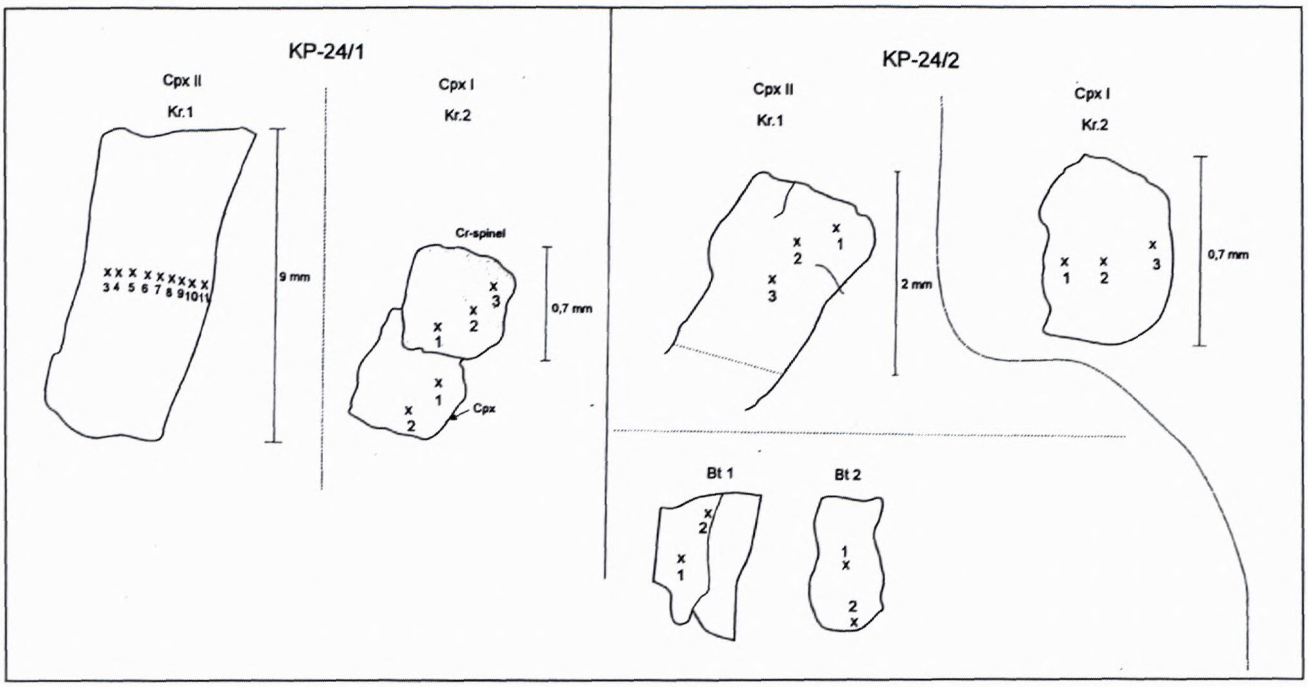


Fig. 2. Diagram of the analysed clinopyroxenes.

Petrology

The basic rocks from the drill hole sections are represented by fragments of centimeter size. The freshest rocks are dark-green, but the tectonized ones are pale green. These rocks contain pale-green porphyric olivines (2-6 mm) of lens-like sections, short (2-5 mm) dark columns of clinopyroxenes, leaves of dark mica (2-3 mm) and products of younger hydrothermal (pyrite, carbonate). The groundmass of the volcanic rocks is deep-green and macroscopically aphanitic. Locally, textures of poikilitic mutual intergrowths of pyroxene, amphibole and dark mica of II. generation are clearly visible.

The rocks contains phenocrysts od more abundant olivine and less abundant clinopyroxene.

Olivine (20-50 % by volume) is present in form of pyrphyritic phenocrysts only. It is extensively serpentinised, and isaccompanied by typical loop-like structures (Photo 1). The cores of the individual loops are formed by serpentinised olivine, as well as talc, calcite and low birefringent to isotropic serpentine of the chrysotile-lizardite group. The middle part of the "channels" is filled by fine-grained ore pigment or by a pale to colourless type of serpentine. Olivine could not be analysed due to the high degree of alteration.

Clinopyroxene is present in the form of columns of three generations or types:

- *Clinopyroxenes I* are made up of tiny columns 1 - 3 mm in size. They occur closely associated with olivine or with Cr-spinel (they are in direct contact with them, Fig. 2, Photo 2). They do not show any optical or chemical zoning.
- *Clinopyroxenes II* - porfyric phenocrysts (up to 10 mm in size) are of idiomorphic appearance. They are pink-violet to brown-violet with a distinct pleochroism (Photo 3, 4). The basic optical properties are the following: Z/c = 54-56°, 2V = 48-50°. Part of the porphyric clinopyroxene phenocrysts are zonally structured and are marked by the gradual ("undulatory") extinction of crystals in the sections perpendicular to the c axis. Oscilatory and sectional zoning is characteristic for the Cpx. Oscillatory zoning is produced by an increase of TiO_2, Al_2O_3, Cr_2O_3, or by reduction of SiO_2, MgO and Na_2O in direction from the centre to the margins of the crystals (Table 1, Figs. 2, 3). The crystal structure of the zoning is a disequilibrium state and is formed during a quick ascent of a melt i.e. change in PT conditions - predominantly pressure.
- *Clinopyroxenes III* - clinopyroxenes of the groundmass. They form tiny 0.1 to 0.4 mm columns or clusters of irregularly confined grains. They have similar optical and chemical properties to the margins of the porphyric phenocrysts.

The composition of clinopyroxenes in picrites was studied by an electron microprobe (Table 1, Fig. 2). In the classification diagram (Fig. 4) these clinopyroxenes project into the diopside field, while the clinopyroxenes I. are shifted in direction of the lower Fe content and higher Mg content. The Cpx II from the picrites have a similar composition to those of the alkali basalts from the Cover Units and of the Krížna nappe (Hovorka – Spišiak 1988). Clinopyroxenes from the picrites are chemically more homogeneous than the clinopyroxenes of the alkali basalts from the Cover Units.

In the clinopyrexenes the Al^{IV} and Ti contents have a close mutual association. Yagi and Onuma (1967) infered, that ions of these elements are present in a hypothetical molecule of Ti-pyroxene $CaTiAl_2O_6$, whose solubility in diopside is markedly decreased at high pressures. Tracy and Robinson (1977) obtained similar results too.

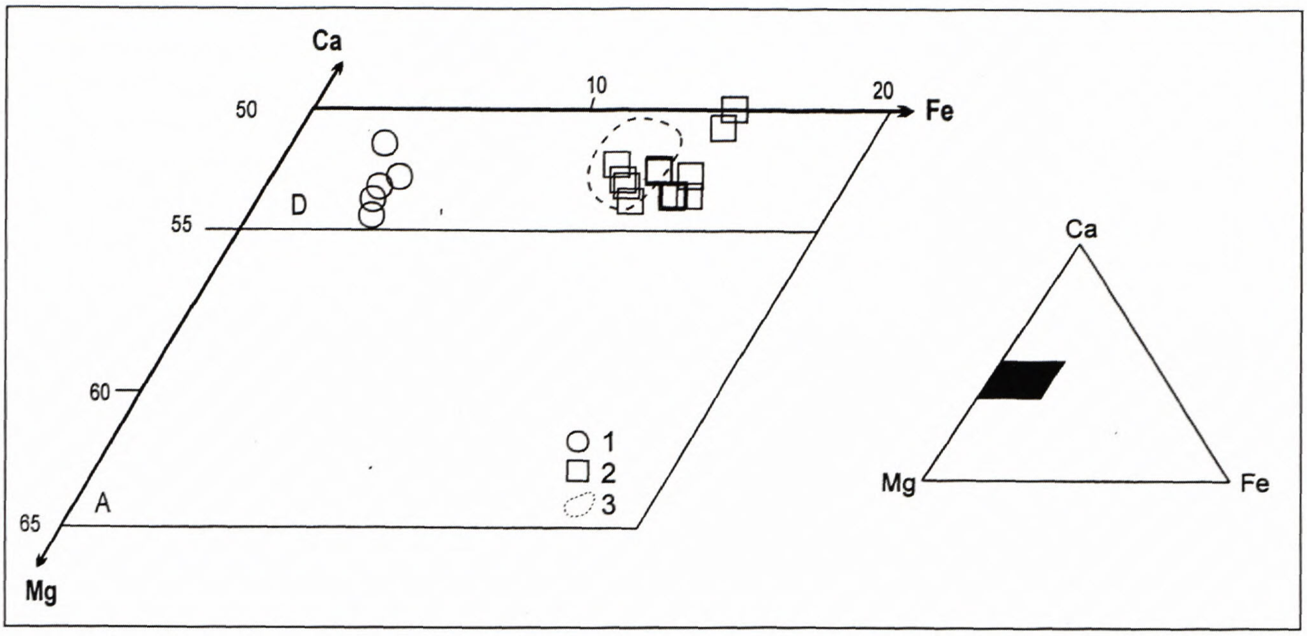


Fig. 3. Linear section with TiO_2, Al_2O_3 and Na_2O contents in the clinopyroxene II grain (kr. 1)

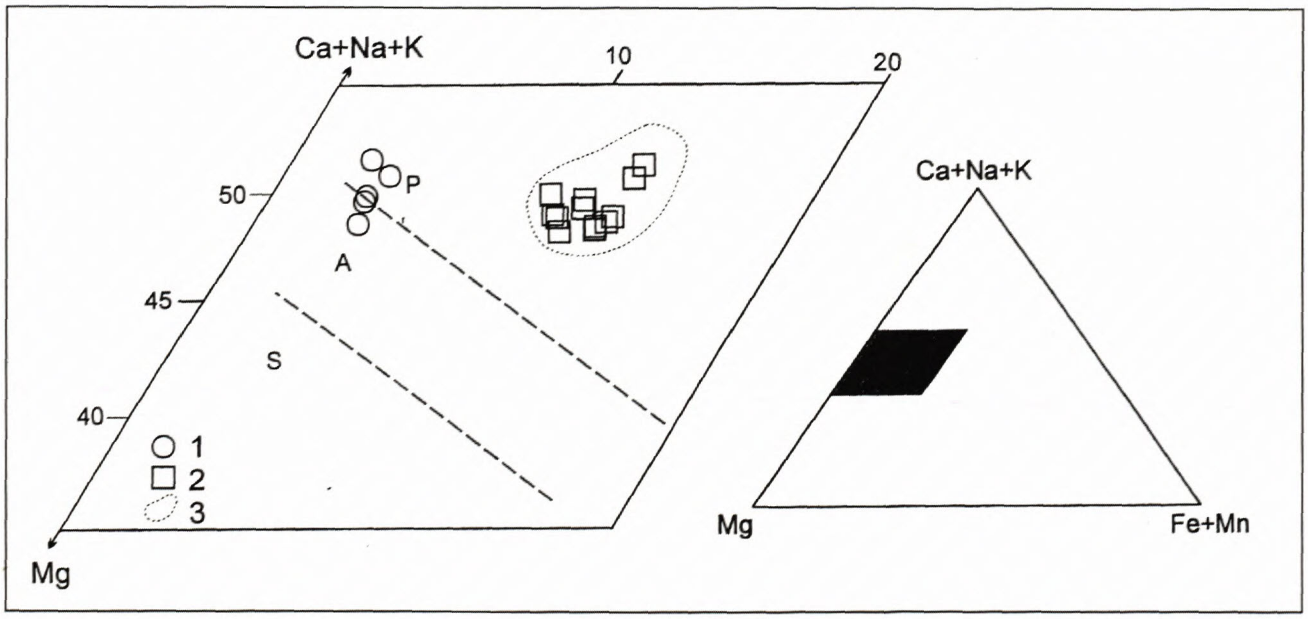


Fig. 4. Classification diagram of clinopyroxenes (according to Morimoto et al. 1988) from picrites from the Poniky area with the following marked fields: A – augite, D – diopside, 1 = clinopyroxenes II, analyses of clinopyroxenes from the figure 2 and table 1, 3 = field of analyses of clinopyroxenes from picrites of the Outer Western Carpathians.

Cpx zoning (rimward enrichment in Al^{IV}, Ti and Fe and Mg and Si and Na depletion) may be explained in three ways (Bédard et al. 1988): (1) it represents a kinetic effect, (2) it reflects a differentiation of the melt as it cools and crystallises, or (3) it reflects changes in pressure during crystallisation. With regard to all possibilities and the presented data (zoning character, Cpx and rock composition, etc.) we infer that polybaric conditions and fractional crystallisation were the dominant process of the Cpx zoning generation. Rimward, Ti-Al^{IV} enrichment is a result of polybaric crystallisation during the ascent of the magma. As pointed out by Wass (1979), data on the solubility of Ti in Cpx in dependence on pressure and temperature are often controversial. There was found a relationship between Ti + Al^{IV}: Si ratio and pressure, and fractionation trends can strongly influence the ratio. The ratio of Al^{IV} : Al^{VI} in Cpx seems to be the most suitable indicator of relative pressure (Thompson 1974, Wass 1979). Through a study of Cpx from the Upper Createaceous lamprophyres in Velence, Buda Mts. (Hungary, Dobosi and Horváth 1988) determined the presence of two types of Cpx: 1. High pressure Al-augites (with low

Photo 1. A spinel grain at the contact with Cpx I (lower left), Sample KP-24, 1N, magnification 54 x, Photo Oswald

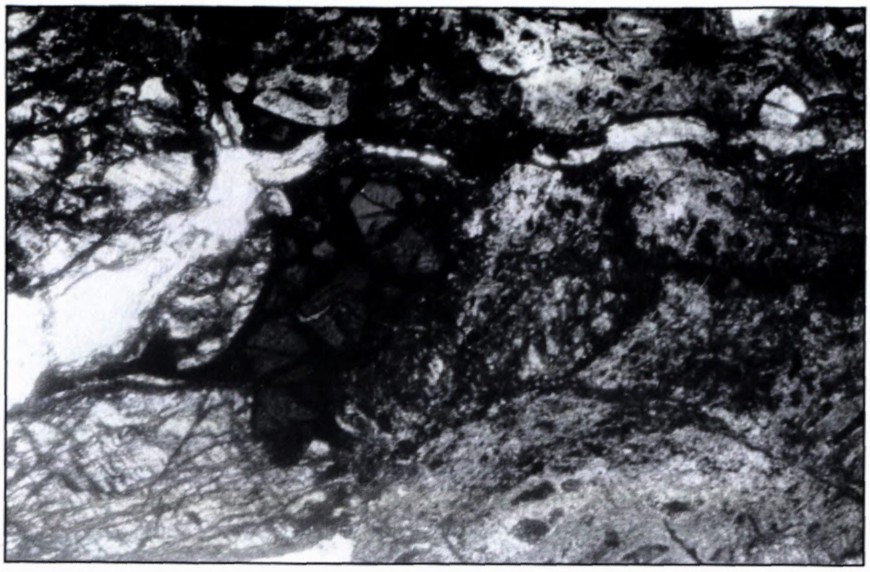
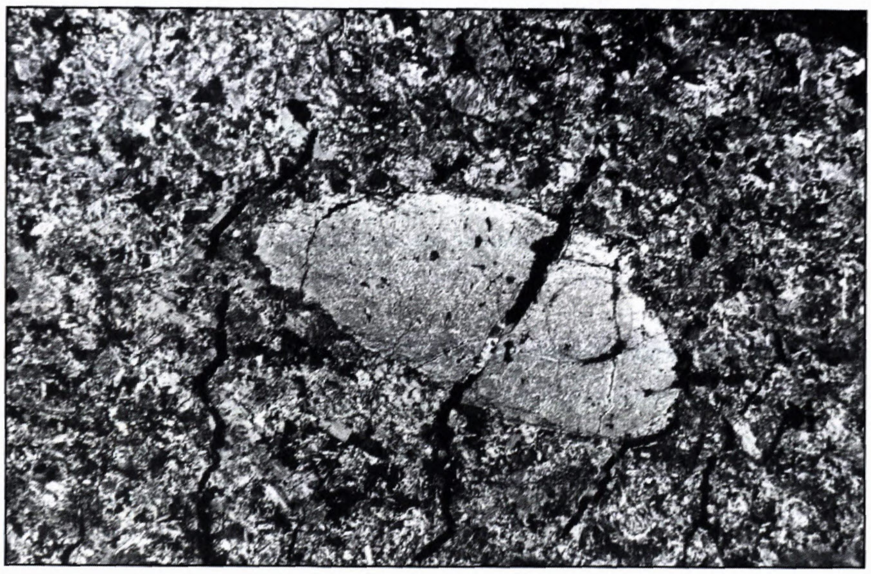


Photo 2. Cpx II phenocryst (in the middle) in the fine grained groundmass, sample KP-24, 1N, magnification 7,7 x, Photo Oswald



TiO_2 concentration and high Al_2O_3 and Na_2O concentrations) and 2. lower pressure zonal phenocrysts with margins enriched in TiO_2.

If applying these findings to the studied pyroxenes, than the clinopyroxenes I crystallised in the upper mantle at high pressures (together with olivine) and the margins of the clinopyroxene II probably represent a product of crystallisation at the mantle/crust boundary at relatively reduced pressure. At the Ti: Al diagram (Fig. 5) a marked differences can be seen between clinopyroxenes I (the association with olivine and Cr-spinel) and clinopyroxenes II. While for the Cpx I very low Ti concentrations are characteristic, Cpx II have higher Ti contents and the projection points are distributed along the line of Ti: Al ratio = 1:3, whereas there is also a positive corelation of these elements (with increasing Al concentration the increase of Ti content also occurs). The projection points of the clinopyroxene II analyses are located on the right from the T line, which indicates the maximal concentration of the presented elements in clinopyroxenes of tholeiite basalts, i.e. they are located in the field of alkali basalts.

Letterrier et al. (1977) used the dependency of clinopyroxene composition on the composition of melts, from those they originated, for the determination of the type of basalts (Fig. 6). At the diagram fields for Cpx of alkali basalts (A) and tholeiitic with alkali-calcareous basalts (TH + CALC) are selected. The projection points of clinopyroxenes II. from the studied picrites are present in the field A (alkali basalts) and a large part of the analyses overlap with the field of clinopyroxenes of the Mesosoic alkali basalts from the other tectonic units of the central zone of the Western Carpathians (the dashed field). This also results in the similarity of clinopyroxene composition of picrites and alkali basalts from this zone. The flat differences inside the given tectonic zone of the I. order are

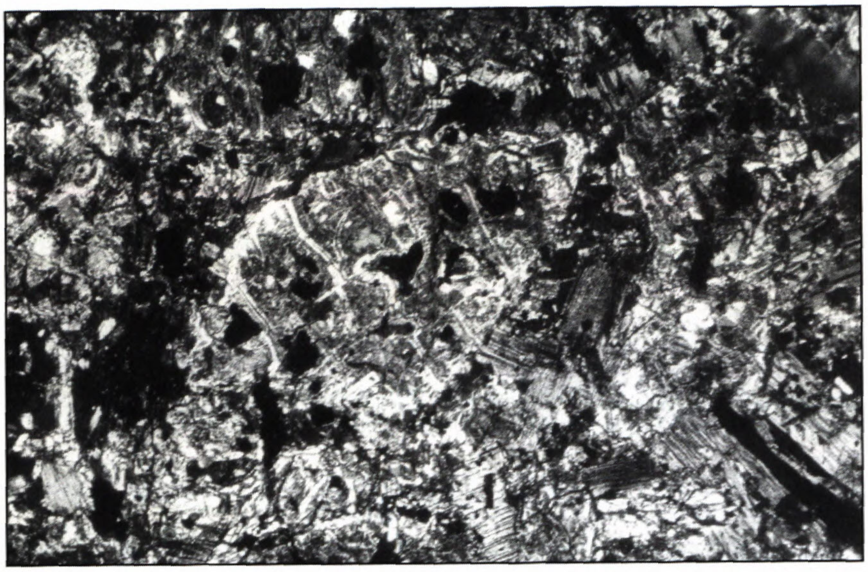


Photo 3. Pseudomorphousis after olivine (in the middle) filled with serpentine group minerals. Around – fine-grained matrix composed of phlogopite, carbonate, serpentine group minerals and ore minerals, sample KP-24, 2N, magnification 27 x, Photo L. Oswald

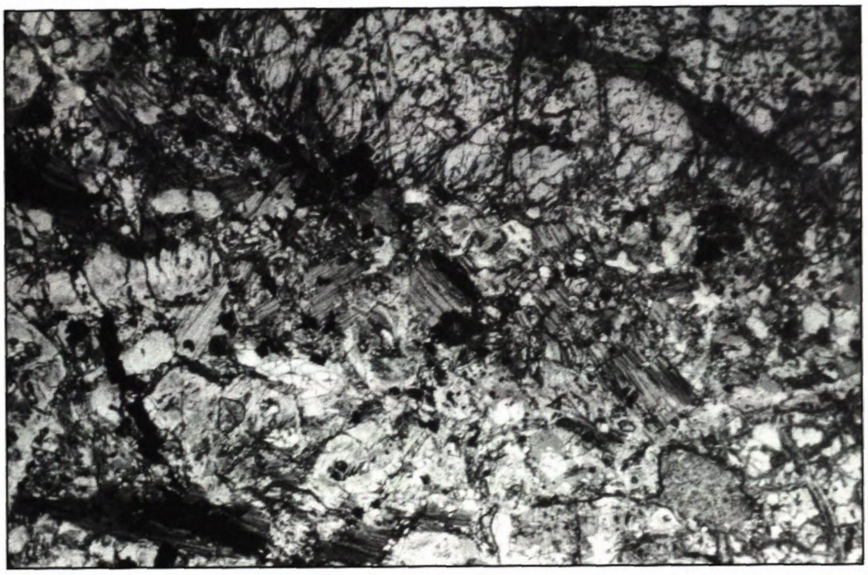


Photo 4. Cpx II phenocryst (at the top) in the fine grained matrix. Under Cpx phenocryst –tiny flakes of phlogopite, sample KP-24, 1N, magnification 27 x, Photo L. Oswald

conditioned by flat differences in the composition of the presented type of eruptives. Considering they have little Ti content, clinopyroxenes I. have markedly different position. Thus, finally we can state that the composition of clinopyroxenes II., as the terminative minerals of picrites from the vicinity of the Banská Bystrica, indicates their classification into the group of alkali basalts.

Dark mica – phlogopite, is represented by deep brown to brownish-red (Z) coloured leaves of the size 0.1 - 3 mm. Its distribution is uneven in detail - locally it forms irregularly confined leaf clusters. The dark mica content in the rock is as much as to 5 volumetric %. Near the hydrothermal veinlets with carbonate filling it gains more bright hue with a pleochroism of bright yellow-brown to green hue. According to the chemical composition (Table 2) the dark micas belong in the field of phlogopites (Fig. 7) in the classification diagram after Deer et al. (1962), and they have an identical position to the phlogopites from picrites of the outer Western Carpathians (Kudělásková 1982).

Apatite is a typical accessory mineral in picrites. It forms long columnar to acicular crystals with an elongation ration 1:25 and it reaches a length of 1.5 mm. Most of the apatite crystals are clear, only in the middle of some crystals is there a dark inclusion. Apatite does not occur among the minerals of the intratelluric phase (a part of the matrix) - it originated among the last phases of the given association.

Fine-grained amphibole (0.01 mm) is present in picrites in the amount of an accessory mineral. Its brown colour and extensive pleochroism are characteristic. It occurs very rarely and it forms idiomorphic columns, but mostly irregular crystals. The pleochroism varies between brown-yellow and brown hue. According to this colour, a

Tab. 1 Selected analyses of clinopyroxenes

Sample	KP-24/1											KP-24/2					
Place	kr. 1 CPx II									kr. 2 CPx I		kr.1 CPx II			kr. 2 CPx I		
Number	3	4	5	6	7	8	9	10	11	1	2	1	2	3	1	2	3
SiO_2	46,99	48,92	48,72	49,78	50,12	49,11	49,52	43,71	44,87	52,67	52,13	49,20	48,26	48,05	52,13	52,43	52,74
TiO_2	2,94	2,28	1,86	1,58	1,24	2,14	1,87	4,76	3,92	0,18	0,00	2,06	2,62	2,79	0,01	0,01	0,21
Al_2O_3	7,06	5,52	5,49	4,50	4,56	5,38	5,34	9,56	8,94	5,53	5,31	5,80	5,89	5,78	6,03	5,75	5,56
Cr_2O_3	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,15	0,14	0,71	0,62	0,01	0,00	0,00	0,85	0,74	0,67
FeO^+	7,46	6,87	7,47	7,53	7,74	6,41	6,84	7,12	7,21	1,89	2,53	6,02	6,81	6,56	2,50	2,30	2,32
MnO	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	0,00
MgO	13,05	13,37	13,51	13,64	13,35	13,84	13,27	11,67	12,03	15,34	16,43	13,56	13,99	13,86	15,44	15,88	15,89
CaO	22,04	22,16	22,03	22,14	21,97	22,17	22,08	22,11	22,09	21,98	21,50	22,04	21,93	22,11	21,50	21,58	21,24
Na_2O	0,33	0,71	0,55	0,52	0,78	0,55	0,50	0,31	0,38	1,15	1,06	0,73	0,55	0,59	1,35	1,09	1,18
K_2O	0,11	0,16	0,16	0,17	0,17	0,14	0,12	0,15	0,12	0,09	0,10	0,11	0,18	0,14	0,16	0,17	0,15
TOTAL	99,98	99,99	99,79	99,86	99,93	99,74	99,54	99,54	99,70	99,54	99,68	99,53	100,23	99,88	99,97	99,95	99,96
Formula based on 6 Oxygens																	
Si^{IV}	1,76	1,82	1,82	1,86	1,87	1,83	1,85	1,65	1,69	1,91	1,89	1,83	1,79	1,79	1,89	1,90	1,91
Al^{IV}	0,24	0,18	0,18	0,14	0,13	0,17	0,15	0,35	0,31	0,09	0,11	0,17	0,21	0,21	0,11	0,10	0,09
Al^{VI}	0,07	0,06	0,06	0,05	0,07	0,06	0,08	0,07	0,08	0,15	0,12	0,08	0,05	0,04	0,15	0,14	0,14
Ti	0,08	0,06	0,05	0,04	0,03	0,06	0,05	0,14	0,11	0,00	0,00	0,06	0,07	0,08	0,00	0,00	0,01
Cr	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,02	0,02	0,00	0,00	0,00	0,02	0,02	0,02
Fe^{2+}	0,23	0,21	0,23	0,23	0,24	0,20	0,21	0,22	0,23	0,06	0,08	0,19	0,21	0,20	0,08	0,07	0,07
Mn	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	0,00
Mg	0,73	0,74	0,75	0,76	0,74	0,77	0,74	0,66	0,67	0,83	0,89	0,75	0,77	0,77	0,83	0,86	0,86
Ca	0,88	0,88	0,88	0,88	0,88	0,88	0,88	0,89	0,89	0,85	0,84	0,88	0,87	0,88	0,84	0,84	0,82
Na	0,02	0,05	0,04	0,04	0,06	0,04	0,04	0,02	0,03	0,08	0,07	0,05	0,04	0,04	0,09	0,08	0,08
K	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,01	0,00	0,00	0,01	0,01	0,01	0,01	0,01	0,01

FeO^+ = total Fe as FeO

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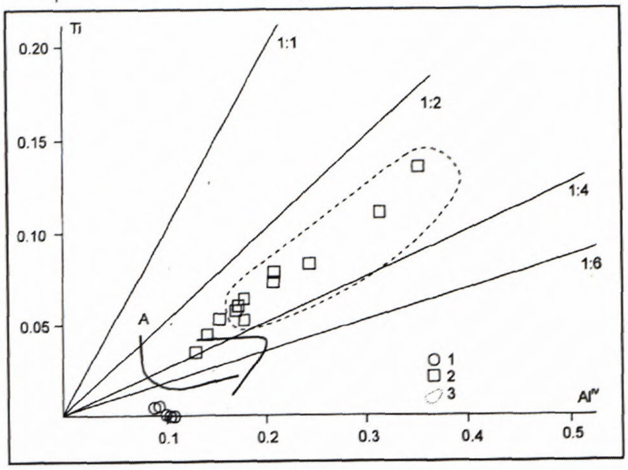

Fig. 5. Diagram Ti vs. Al^{IV} for clinopyroxenes with marked fields for: maximum Ti, Al contents in tholeiite basalts (T) and their minimum contents in alkali basalts (A) (fields according to Maruyama 1976, in: Takeda 1984); explanations as in figure 3.

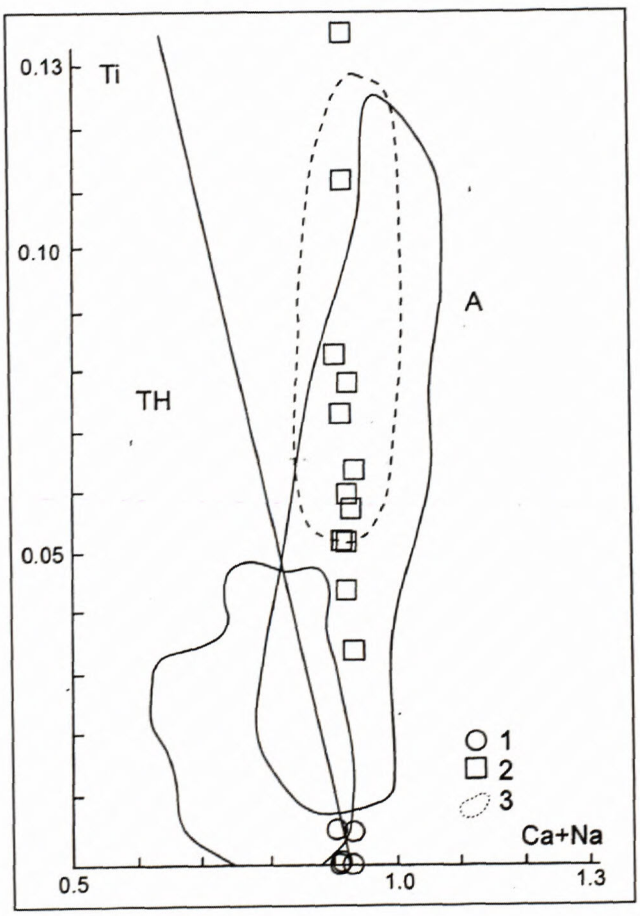

Fig. 6. Diagram Ti vs. Ca + Na for clinopyroxenes (according to Leterrier et al. 1982). A - field of clinopyroxenes from the alkali basalts, Th = field of clinopyroxenes from tholeiites and calc-alkaline basalts; explanations as in figure 3.

general mineral association, as well as according to the analogy with amphiboles from alkali basalts of the Krížna nappe, we are dealing with an amphibole of kaersutite type.

Fe-Al-Cr-spinel is a typical ore mineral of tobacco-brown colour. It has a sinus-like, skeleton and it is overall

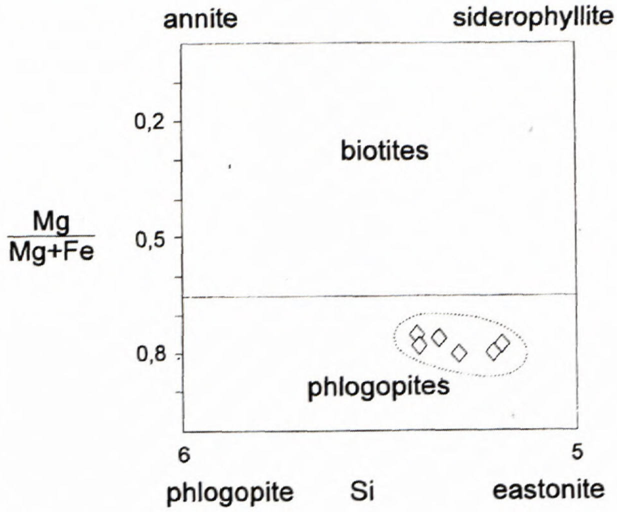

Fig. 7. Classification diagram of dark micas (according to Deer et al., 1962) from picrites from the Poniky area; analyses of micas from the table 2. Dashed field - phlogopite analyses from picrites of the Outer Western Carpathians (Kudělásková 1982)

very irregular confined. Most of the crystals of the spinel mineral group have at the margins tight rims formed by an opaque ore phase. It corresponds to Cr-spinel by its composition (Table. 2, Fig. 8).

The isometric ore grains are present in the form of poikilitic inclusions in clinopyroxenes of the I[st] generation also are, mostly - *magnetite* ± Ti-magnetite (Kevex). Furthermore, these minerals occur also separately closely associated with unaltered parts of olivines. The spaces between the mineral phases of the I. and II. generation are locally filled by brownish coloured, glassy groundmass (20 - 30 vol. %), in which locally the microspherulitic structure, irregular clusters of tiny ore minerals and clusters of *chlorite* leaves are evolved. Locally, in the glassy aggregate fine-foliacoeus, low-birefregent serpentine, probably lizardite, is present. The association *calcite, talk, pyrite, chlorite, limonite* is a product of various types of overprinted alterations.

Geochemistry

Despite the relatively high degree of alteration we studied also the chemical composition of picrites (Table 3). From the main elements, low concentrations of SiO_2, Al_2O_3 and relatively high contents of TiO_2, MgO and P_2O_5 are characteristic for the studied picrites. The high value of LOI is influenced by the strong alteration of rocks - predominantly by hydration. From the trace elements, high concentrations of Cr and Ni, i.e. elements typical of basic and ultrabasic rocks, are characteristic for picrites from the broader area of Banská Bystrica. For the geochemical evaluation we have used predominantly incompatible elements (mainly HFSE) and a group of rare earths elements. In the ternary Zr:Nb:Y diagram (Fig. 9) the picrites are located in the field of intra-plate alkali basalts (WPA). The identical position of the studied rocks

Table 2 Chemical composition of phlogopites and spinels

	phlogopites						spinels		
	1	2	3	4	5c	5r	1	2	3
SiO_2	36,32	37,17	35,30	37,16	37,96	37,60	0,09	0,00	0,00
TiO_2	4,49	3,89	6,38	4,89	3,99	4,88	0,00	0,14	0,00
Al_2O_3	16,17	16,23	16,36	16,02	16,48	16,32	52,52	50,06	52,09
Cr_2O_3	0,00	0,00	0,00	0,00	0,00	0,00	13,36	16,76	14,35
$Fe_2O_3^*$	0,00	0,00	0,00	0,00	0,00	0,00	3,37	3,25	4,10
FeO	10,29	10,98	9,26	9,49	9,71	9,76	11,20	11,51	10,80
MnO	0,00	0,00	0,00	0,00	0,09	0,12	0,00	0,00	0,00
MgO	18,39	18,59	18,24	19,01	21,83	22,13	18,99	18,74	19,37
CaO	0,00	0,15	0,00	0,10	0,02	0,11	0,00	0,00	0,00
Na_2O	0,00	0,00	0,00	0,00	1,11	1,23	0,00	0,00	0,00
K_2O	8,42	8,39	8,76	7,89	7,51	7,51	0,00	0,00	0,00
Suma	94,08	95,40	94,30	94,56	98,70	99,66	99,54	100,46	100,71
Formula based on 22 Oxygens									
Si^{IV}	5,35	5,41	5,19	5,40	5,30	5,21			
Al^{IV}	2,65	2,59	2,81	2,60	2,70	2,70	1,65	1,58	1,62
Al^{VI}	0,16	0,19	0,02	0,14	0,01	0,00			
Ti	0,50	0,43	0,71	0,53	0,42	0,51	0,00	0,00	0,00
Cr	0,00	0,00	0,00	0,00	0,00	0,00	0,28	0,35	0,30
Fe^{3+}	0,00	0,00	0,00	0,00	0,00	0,00	0,07	0,07	0,08
Fe^{2+}	1,27	1,34	1,14	1,15	1,13	1,16	0,25	0,26	0,24
Mn	0,00	0,00	0,00	0,00	0,00	0,00			
Mg	4,04	4,03	4,00	4,12	4,54	4,57	0,75	0,75	0,76
Ca	0,00	0,02	0,00	0,02	0,00	0,00			
Na	0,00	0,00	0,00	0,00	0,30	0,33			
K	1,58	1,56	1,64	1,46	1,33	1,32			

$Fe_2O_3^*$ - in the case of spinels the ferrous iron calculated based on structural formula
c = core of a grain, r = rim of a grain

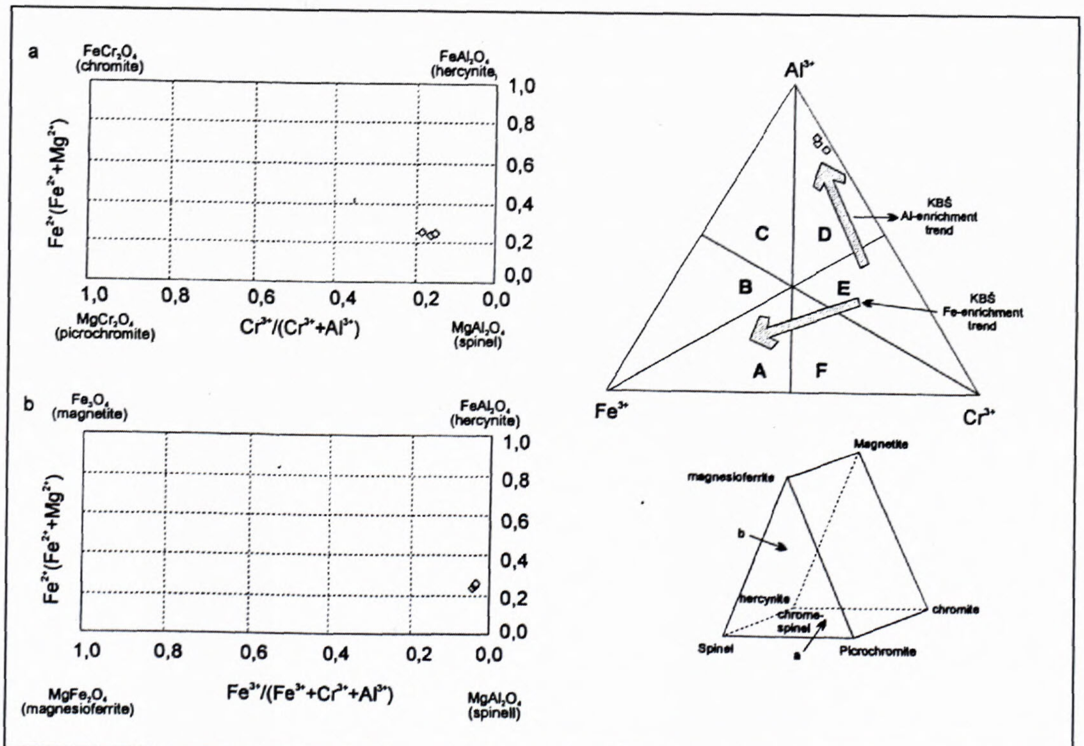

Fig. 8. Classification diagram of spinels (according to Stevens, 1944, Fe and Al - enrichment trend according to Henderson, 1975), analyses of spinels from table 2.

Table 3 Chemical composition of rocks

sample	HD-1	HD-2	KP-24	KP-24/1	KP-27	MV-81	SVP
SiO_2	34,29	29,76	34,06		38,12	37,3	41,01
TiO_2	2,45	0,28	1,34		2,02	1,74	0,48
Al_2O_3	16,22	9,4	5,72		7,57	6,24	8,35
Fe_2O_3	3,63	3,21	6,75		6,52	*9,65	5,93
FeO	6,08	2,31	3,93		4,04		9,5
MnO	0,1	0,03	0,12		0,19	0,14	0,18
MgO	11,1	13,87	22,26		16,94	22,83	23,25
CaO	9,07	10,9	9,05		10,65	5,92	2,61
Na_2O	0,7	0,09	0,41		0,96	0,26	0,01
K_2O	0,79	3,44	0,92		1,02	0,94	0,01
P_2O5	0,67	0,08	0,74		0,85	0,65	0,14
H_2O^-	1,06	0,83	3,07		1,86		0,51
H_2O^+	13,36	25,04	10,87		8,15	*13,5	7,53
Total	99,52	99,24	99,24		98,89	99,17	99,51
La			53,85	39,17		33	1,05
Ce			101,21	72,94		75	
Nd			46,02	32,63		29	
Sm			10,3	7,27			0,52
Eu			2,61	1,75			0,154
Gd			6,82	4,83			
Dy			4,64	3,19			
Er			2,03	1,41			
Yb			1,33	0,96			0,31
Lu			0,18	0,13			
Y			18,85	12,83		14	
Ba			455	303		383	
Co			110	114	117	81	113
Cu			48	39	27		
Nb			71	53		50	
Ni			873	1232	1040	690	1290
Sc			18	15	21		
Sr			562	291		285	
V			173	138		150	
Zn			135	124	117	110	
Zr			209	159	224	163	
B			21,4		24,5		
Cr			680		710		1380
Ga			14		10		
U						1	
Th						8	

Samples HD-1, HD-2 and MV-81 = Predný diel, samples: KP-24, KP24/1, KP-27 = Poniky, SVP - picrite from Strážny vrch (Iván 1991), * = total iron as Fe_2O_3

and Mesozoic alkali basalts/basanites of the Central Western Carpathians suggests their geochemical relationship (dashed field). Alkali rocks have very specific REE contents (high concentrations of light REE and low concentrations of heavy REE). For a better interpretation we compared the composition of the studied picrites to the composition of the alkali basalts of the ocean islands and to an average teshenite (Fig. 10). The studied rocks have very similar character of the normalised curve with characteristic sharp incline in direction to the heavy rare earths and with a nearly invisible Eu anomaly. The composition of picrite from the Strážny vrch (Gemerikum, Ivan 1991) is markedly different.

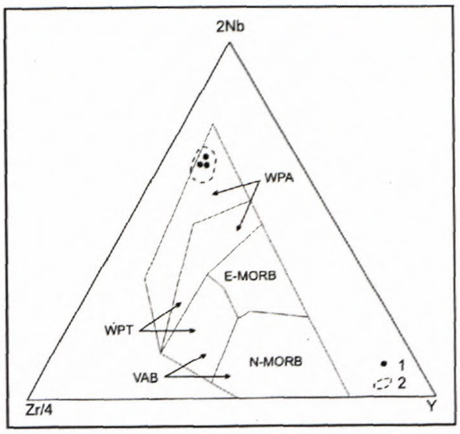

Fig. 9 Ternary diagram Zr:Nb:Y (according to Meschede, 1986) for various geotectonic types of basalts; 1 - analyses of the studied rocks from table 3, 2 - Mesozoic alkali basalts of the Central Western Carpathians (according to Hovorka-Spišiak 1988, Spišiak – Hovorka 1997, Hovorka et al. 1999).

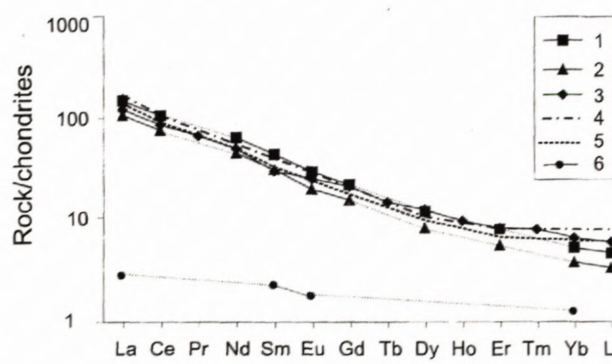

Fig. 10. Normalised curve of rare earths distribution in picrites from the Banská Bystrica area (1 - 3 analyses from the table 3). 4 = average composition of alkali basalts from the oceanic islands (OIB, Sun - McDonough 1989), 5 = average composition of teshenites (Rosi et. al. 1992), 6 = composition of picrites from the Strážny vrch (Iván 1991)

Discussion and Conclusions

– Picrites from the broader area of Banská Bystrica do not have equivalents among the basic volcanic rocks of the Mesozoic of the Western Carpathians. This results from their characteristic mineral composition, structures and chemical composition. Abundant olivine, present in the form of porphyric phenocrysts only, the presence of two Cpx generations, phlogopite and particularly the skeletal and sinus-like brown spinel are typical for these picrites.

– Despite the fact, that Hovorka and Slavkay (1966), based on the clasts of mainly volcanic glass in the overlying dolomites in the drill hole P-1, regarded the relationship of picrites and dolomites for a normal stratigraphic relationship, the discovery of picrites of identical composition also in the Krížna nappe from the study area (Slavkay, 1979) enabled new aspects of the problem to occur that in the time, when just picrite from the drill hole Po-1 was known, were not actual. Note, that picrites probably occur in two tectonic units of markedly allochtonous type. Those areas of sedimentation were separated from each other by tens to hundreds of kilometres. At the same time, there is a low likelihood of intrusions/effusions of nearly identical types of lavas into Mesozoic sedimentary sequences of the two basins. Thus there is a possibility to relate the penetration of picrite melts just after the moving of the Krížna and Choč nappes into their present-day position occurs, while their penetration was focused along the NS-oriented lineament - the Zázrivá-Revúca fault zone. Similar interpretation was already outlined by Hovorka (1976b) and Slavkay (1979) in their studies. Despite also taking into account the results of the study of Biely (1979) about the picrite occurrences area near Poniky (drill hole Po-1, Hovorka - Slavkay 1966), the insertion of dolomites with the associated picrites into the Choč nappe is not entirely clear. If the studied dolomites do not belong to the Choč nappe, but in the Krížna nappe, the above discussion would be not necessary. However, the existing data do not allow a definitive interpretation, therefore we do not regard this problem as entirely solved.

– Based on geochemical criteria the studied rocks from the broader area of Banská Bystrica belong to the alkali intra-plate basalts/basanites. Their composition is similar to the Mesozoic alkali basalts/basanites of the Central Western Carpathians and to the average teshenites.

– Picrites from the broader area of Banská Bystrica can be assigned to the province of the Mesozoic (Lower Cretaceous) alkali basalts/basanites.

This study represents a partial output from the grants No. 2/7091/00 and 1/6000/99 VEGA. Finally, we would like to thank Mrs. N. Halašiová for the computer processing of figures and to referees for valuable remarks, RNDr. P. Konečný for critical review and the English advisor H. Drewes for suggestion helps.

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Slovak Geol. Mag., 6, 4(2000), 431–440

Upper Carboniferous granitoid stage in the Veporic: transition from I - to S-type magmatic events.

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Abstract. The Upper Carboniferous (about 300 Ma) magmatic stage produced relatively large volumes of the magma in the Western Carpathian basement. Recent mapping and petrology study in the Veporic Alpine Unit revealed a polystage character of this magmatic event. Granitoid types formerly designed as Sihla (I-type) tonalite (with mixed mantle-crustal origin) and porphyritic Ipeľ granodiorite with pink K-feldspars belong to this magmatic stage. The third magmatic event, represented by the more leucocratic magmatic bodies with crustal origin cuts the Sihla tonalite. This third stage (S-type) is represented by tonalites to granites (granodiorites prevails) with frequent metasedimentary xenoliths and restitic biotite rich enclaves. Unresorbed quartz xenocrysts are also preserved. Thus, the progressive melting of the crustal material and decreasing of influence of mantle magmatic activity is probably responsible for the different magmatic products during one magmatic period. Intrusions of the youngest magmatic portions is connected with the ductile deformation (NE-SW direction) of older Sihla tonalite.

Key words: Western Carpathians, Hercynian granite magmatism, Veporic Unit, Upper Carboniferous, mineralogy, petrology

Introduction

After the main collisional event (Mesohercynian orogenic stage) and the formation of principal Hercynian lithotectonic units (Bezák et al., 1997), tectonic movements dominated on deep faults systems in a transtensional regime. The role of the Neohercynian orogenic stage (340-260 Ma) is significant in Hercynian granite magmatism of the Western Carpathians. During this stage, deep seated melts generated in the lower crust, in interaction with a mantle material and produced mainly I-type granitoids (the Sihla type).

The main products of the Hercynian magmatism in the Veporic Unit formerly described in literature as Sihla tonalite, or trondhjemite (Zoubek, 1936; Broska & Petrík, 1993a) and Ipeľ granitoid types (Krist, 1979) are on the bases of isotopic dating Upper Carboniferous in age. U/Pb zircon ages show 303 ± 2 Ma for the Sihla type (Bibikova et al., 1990) and 305 ± 5 Ma on zircon and monazite for the Ipeľ type (Michalko et al., 1998). Recent mapping and studies of these granitoids (Hraško et al.,1998; Bezák et al., 1999) in the Veporic unit detected their rather complicated intrusive history.

Mapping works revealed intimate spatial relations among these Upper Carboniferous magmatic pulses (Fig.1). Cross-section documents the spatial relation of Upper Carboniferous magmatic rocks among older and younger intrusive phases in the SW part of Veporic, as well as internal mutual position of this magmatic pulses of the Upper Carboniferous magmatic stage (Fig. 2).

The main goal of the paper is to characterize the Upper Carboniferous magmatic stages in this territory, from point of view of petrography, mineralogy, geochemistry and field works. A possible geotectonic scenario is presented for this magmatic events.

Methods

Mapping works of investigated area were done in the scale 1: 25 000. The representative granite specimen were collected for geochemical and mineralogical analyses in the weight 5-10 kg, Accessory minerals were studied from the heavy fraction of the crushed rocks and the thin-sections. Concentration of heavy minerals was done by classical separation methods comprises crushing, sieving, using Wilfley table, heavy liquids and isolation according to magnetic properties. For determinations of mineral compositions a microprobe JEOL – 733 was used, with application of natural standards.

Chemical composition of rocks was analysed in Geoanalytical Laboratories of the Dionýz Štúr Geol. Institute in Spišská Nová Ves. Elements like Ba, Cu, Ga, Nb, Ni, Pb, Rb, Sr, Y, Th were measured by RFS method; Be, Co, Cr, Gd, Hf, Li, Lu, Nd, Sm, V, Yb by AES-ICP method. La, Ce were measured by RFS, in the case of all analysed REE elements – by AES-ICP. REE in samples – HV-7, HV-33, HV-36 were analysed in Ecological Laboratory in Spišská Nová Ves.

Geological Survey of Slovak Republic, Dionýz Štúr Publishers, Bratislava 2000 ISSN 1335-96X

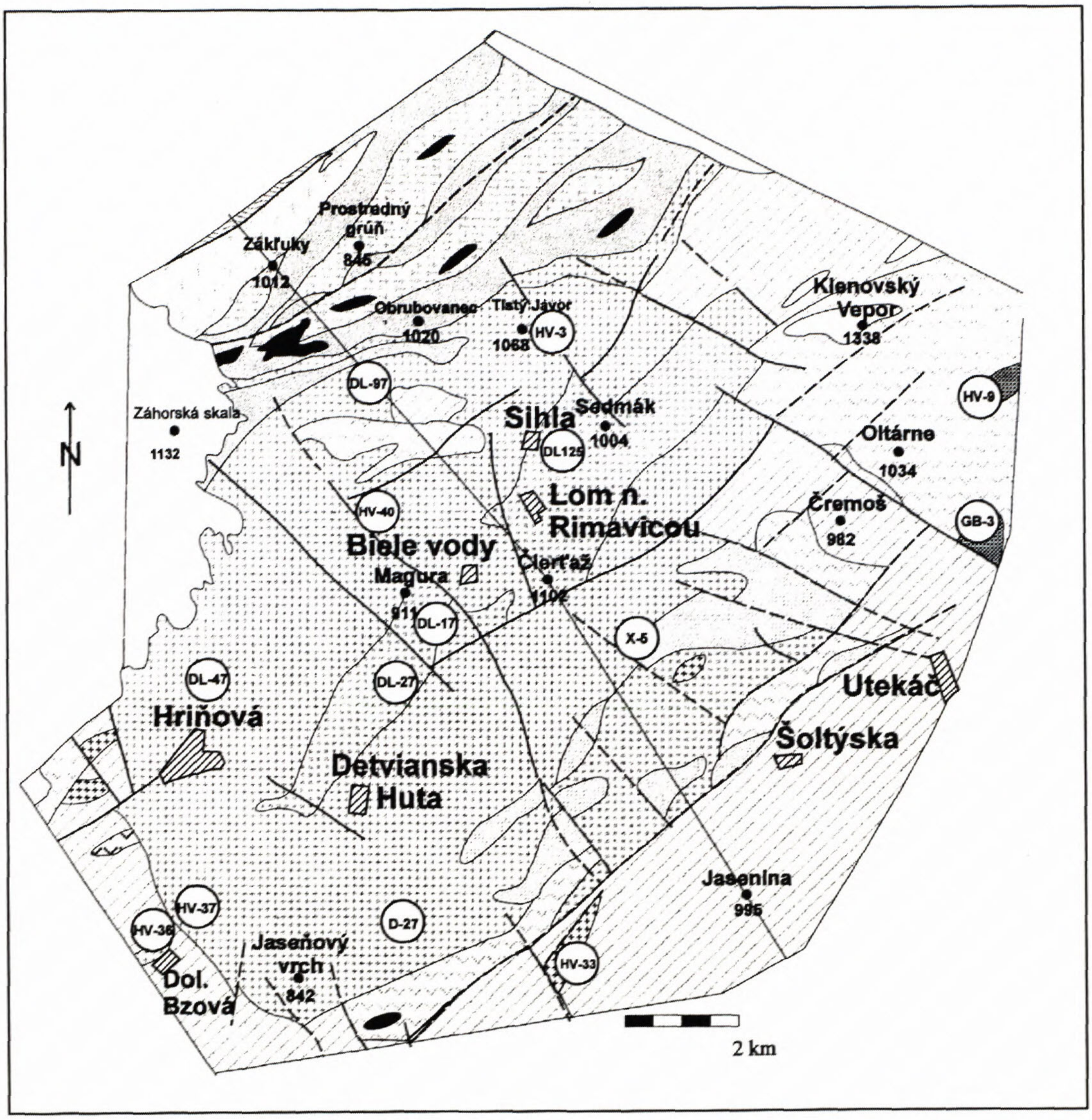


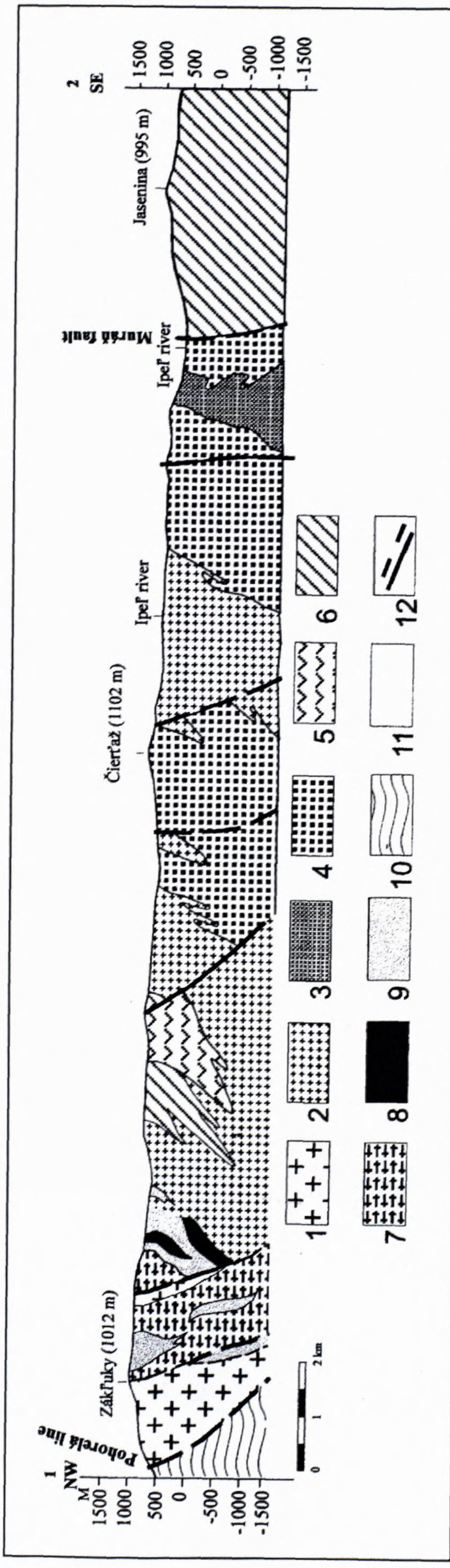
Fig.1: Simplified geological map of investigated area (after Bezák et al.,1999), sample localisation and position of cross-section 1-2.Explanations are in Fig.2.

Main products of the Upper Carboniferous magmatic event in the Veporic unit.

The *Sihla type of granitoid* (tonalite – granodiorite) forms larger (tens km^2 in size) or smaller dykes 0.x - x m in thickness. Mostly high angle bodies prolonged in the direction NE-SW and less E-W. Older granitoid bodies (granitoids of the so called hybridic complex) and older Vepor type porphyric granitoids form a peripheral part of the body, or can be found as a solid blocks drown into the magma of the Sihla tonalite (see the geological section in Fig. 2). The Sihla tonalite is rarely cut by more leucocratic granitoids rich in plagioclase. Their contacts against the older granitoids (hybrid or Vepor porphyritic granite types) or metamorphic rocks in the wall rocks have a clear intrusive character. Mafic microgranular enclaves derived from basic magma are abundant in the Sihla granitoids (Broska & Petrík, 1993b).

A zoning was observed in the Sihla magmatic bodies. Tonalites sometimes form their central part, and are enriched by K-feldspar towards the chamber roof (granodiorites). The Sihla type comprise also further petrographic types: tonalite-granodiorite with tiny idiomorphic K-feldspars and porphyric tonalite to granodiorite with pink K-feldspar (with low modal abundance). A typical loaf shaped mafic microgranular enclaves resulted from co-mingling with more felsic - Sihla melt. The mafic enclaves may represent remnants of a basic magmatic body which triggered melting of the lower crust (Broska & Petrík, l.c.).

Sihla tonalite – granodiorite are plagioclase bearing granitoids containing more than 50 % of plagioclase of the oligoclase – andesine composition, sometimes in porphyritic development (up to 10 mm). Due to strong alteration of plagioclases, pale-green colour is typical for these granitoids. Sometimes, anortite component of unaltered plagioclase cores reaches up to value An_{35}. K-feldspar is not common, usually interstitial and only locally porphyric, occasionally albitised. Albitisation is connected with formation of postmagmatic muscovite



crystals (Fig. 3 – F). This process is partially responsible for the higher peraluminosity of some samples. Biotite is greenish-brown. Modal compositions of selected samples is in table 1. The typical feature is presence of allanite, often rimmed by the epidote crystals (Fig. 3 – E) and presence of titanite. Composition of biotite has rather high Mg / (Fe_t+Mg) ratio – over 0.55 (Fig. 4).

Tab. 1. Modal composition of selected granitoid samples

Sample	DL-17	DL-27	GB-3	X-5	VG-54	DL -47	DL-97
Type	BV	BV	Ip	Si-Ip	Si	Si	Si
Qtz	47,80	33,30	29,80	26,20	19,40	25,00	21,90
Plg	31,65	51,75	26,30	54,20	65,10	61,40	61,25
Kfs	14,25	5,65	18,20	7,75	1,90	1,20	0,70
Bt	3,15	2,90	7,00	11,25	10,80	0,70	1,40
Ms_{1+2}	2,36	5,35	11,80		1,90	0,40	2,35
Ep_{1+2}	0,55	0,80	5,00	0,20	1,00	4,00	2,40
Aln				0,30	0,17		
Ttn					0,90	0,30	0,50
Ap				0,30	+		
OM			0,10	0,05		0,40	0,20
Chl	0,10	0,20	0,80			3,50	9,30
Cal						0,20	
Gar		0,05					
Acc	0,15						
Total	100,00	100,00	99,00	100,0	101,17	97,10	100,00

Abrev.: Qtz – quartz, Plg – plagioclase, Kfs – K-feldspar, Bt – biotite, Ms_{1+2} – primary and metamorphic muscovite, Ep_{1+2} – primary and metamorphic epidote/clinozoisite, Aln – allanite, Ttn – titanite, Ap – apatite, OM – ore minerals, Chl – chlorite, Cal – calcite, Gar – garnet (metamorphic origin), Acc – accessory mineral together

Fig. 2: Cross-section of the granitoid bodies of the central part of the Veporikum (after Bezák et al.,1999) with position of Sihla-Ipeľ magmatic stage products.

Hrončok granite (Permian - Triassic?)

1 - leucogranites a granite-porphyries - Hrončok type, often phyllonitised

Sihla - Ipeľ magmatic stage (Upper Carboniferous)

2 - Sihla type tonalites prevail; 3 - Ipeľ type; massive porphyric granodiorites with pink K-feldspar porphyrocrysts; 4 - Biele vody type; massive, fine to coarse-grained leucogranodiorites to leucogranites, locally with xenoliths of gneisses

Pre-Upper Carboniferous granitoids

5 - Vepor type - porphyric granodiorites to granites, usually in subhorizontal position, foliated; 6 - Hybrid (schliered) tonalites-granodiorites, less migmatites, gneisses

Pre-Carboniferous magmatic and metasedimentary rocks

7 - orthogneisses of granitoid composition, often phyllonitised 8 - diorites, amphibolites, amphibole gneisses, often diaphtorised; 9 - migmatites, migmatitised gneisses and biotite, garnet-biotite paragneisses,, pearl gneisses, locally injected by aplitic veins, often phyllonitised; 10 - phyllonites of gneisses; 11 - faults, tectonic lines - undivided

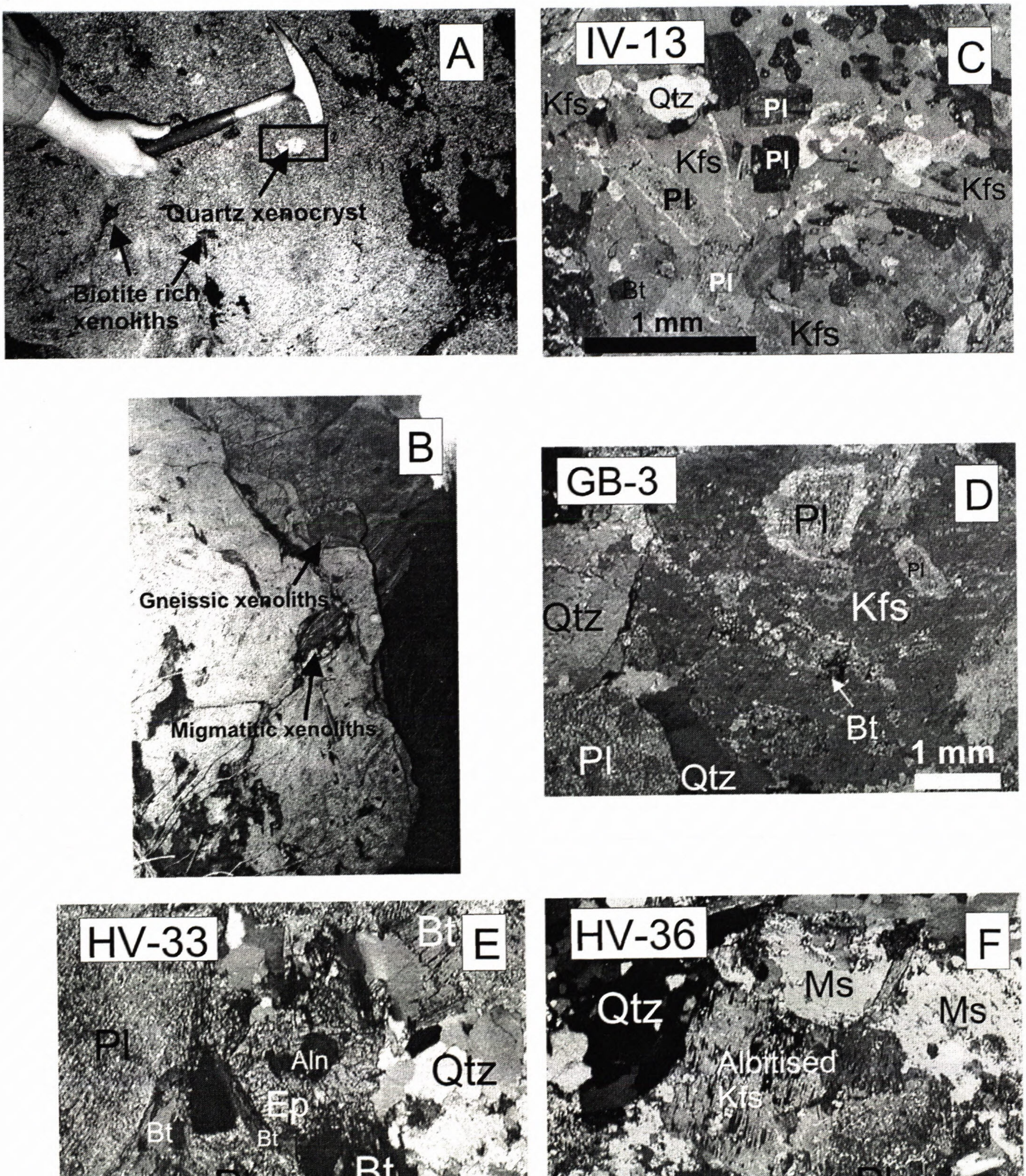


Fig. 3: Photo table of A, B – outcrops of Biele vody granitoids, C-F - microphotos of selected samples of investigated granitoids (explanations in the text)

Common accessory assemblage is represented by magnetite, allanite and titanite along with zircons S_{12}, S_{13}, S_{16} subtype (Fig. 5) after Pupin classification (Pupin, 1980). Macroscopically visible titanite is characteristic for early phases of Sihla granitoids.

Ipeľ type of granitoids are volumetrically less abundant in comparison with Sihla granitoids (Fig. 1, 2). They crop out mainly SE of the main Sihla granitoid body.

Similary as Sihla type, Ipeľ types form high angle bodies and dikes cutting Sihla tonalite and Early Palaeozoic schliered granitoids. Real time relationship with Sihla tonalite is still unclear and field transitions between Sihla and Ipeľ granitoids are often transitional.

Basic varieties are represented by the massive, medium to coarse-grained, usually unfoliated porphyric granitoids with pink K-feldspar usually as phenocrysts (around 1-2 cm). Ipeľ granitic varieties are represented by granodiorites, granites less frequently by tonalites (Tab. 1). Plagioclase is hypidiomorphic, intensive sericitised. K-feldspar is idiomorphic, perthitic, locally with microcline domains, very often with inclusions of plagioclase, rounded quartz and biotite (Fig. 3- C-D). It is relatively late in mineral succession. Biotite is interstitial greenish-brown with Mg/(Fe_t+Mg) ratio around 0.50 - 0.52 (Fig. 4).

Similar accessory mineral associations as in the Sihla type granitoids is typical also for Ipeľ granitic rocks. Slightly different zircon compared to Sihla subtypes (Fig. 5) are present (zircons S_3, S_4, S_{12}, S_{13}), but the assemblage is the same: apatite, allanite, titanite, zircon, ilmenite, magnetite and sulphides (pyrite, arsenopyrite).

Fine- and medium-grained leucocratic granitoids (Biele vody type) (partly also leucocratic porphyritic tonalites and granodiorites - locally granites) intrude the Sihla type of granitoids through steeply inclined structures. The fine- and medium grained leucocratic granitoids form the biggest body (2-4 km in thickness, 18 km in length) between Detvianska Huta and Biele Vody settlements and NE and SE from Lom nad Rimavicou. They are situated mainly in the central part of Sihla magmatic body and follow the same direction as Sihla granitoids.

Peripheral parts of the body are usually porphyric. In the new geological map in the scale 1: 50 000 (Bezák et al., 1999) this type is merged with the typical more mafic Ipeľ type granitoids for simplification.

There are common xenoliths of crustal origin (gneisses and migmatites captured in granitoids forming several dm ellipsoid xenoliths in size – Fig. 3A) and quartz xenocrysts up to 3-5 cm in size (Fig. 3B). In the central part of this body, steeply inclined inclusions of biotite-rich bearing rocks (up to 5 cm) were found (Fig. 3A). These are considered as restite. A chaotic distribution of xenoliths, found in central parts of the body, indicate a high melt mobility. Leucocratic granodiorites are often fractured and cross-cutted by the net of the aplitic and haplogranitic veins.

Plagioclase is hypidiomorhic, up to 3 mm. K-feldspar is interstitial, rarely porphyritic, sometimes with inclusions of biotite flakes, plagioclases and isometric

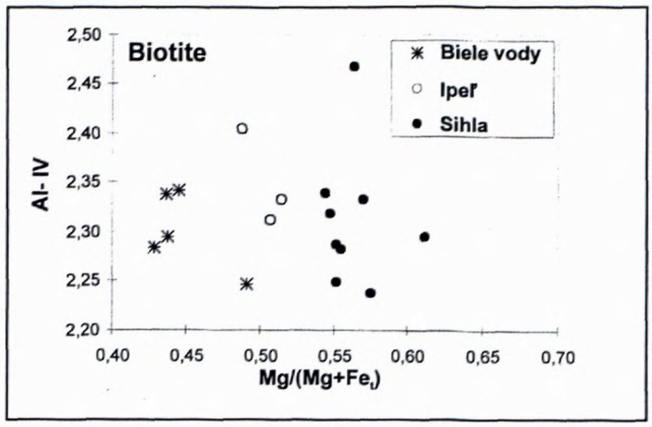


Fig. 4: Composition of biotites of the granitoid types

quartz crystals of the 1st generation (modal composition in tab. 1). Biotite is brown with Mg/(Fe_t+Mg) ratio varies from 0.42 to 0.49 (Fig.4).

More siliceous types of granitoids contain monazite instead of typical allanite and titanite, which is usually missing. The amount of xenotime is very low. Sulfides as pyrite, arsenopyrite and molybdenite (sample DL-17) were identified. Sometimes, accessory amphibole is present. Zircon morphology with subtypes S_1, S_2, S_6, S_7 (Fig. 5) indicates typical crustal origin (S-type granitoids).

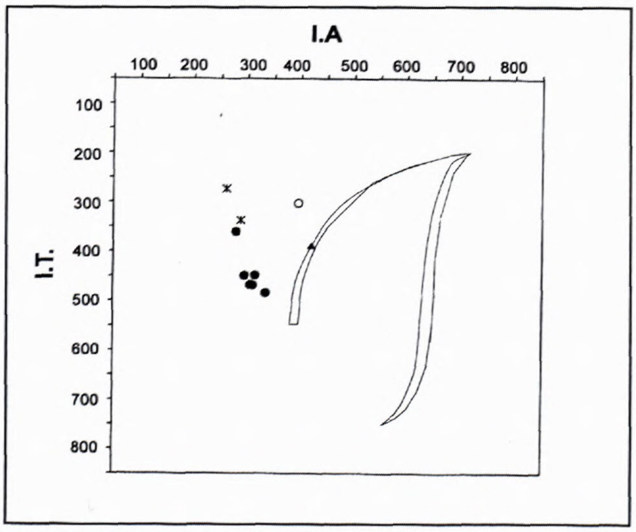


Fig. 5: Zircon typology diagram with the projection of mean points, Explanations: full circles – Sihla granitoids, triangle – Sihla - Ipeľ granitoid, open circles – Ipeľ granitoids, stars – Biele vody granitoids

Geochemical characteristics

Main and trace elements of investigated Upper Carboniferous granitoid rocks are presented in table 3a,b. For more detailed characteristic of Sihla tonalite-granodiorite, which is principal granitoid type of this stage, the chemical statistical parameters are shown in Tab. 2.

Tab. 2: Statistical parameters of the Sihla granitoid chemical compositions

w. %	N° of samples	Arihm. mean	Med. mean	Stand. deviat.	Lower quart.	Upper quart.
SiO_2	22	64,92	65,13	1,61	63,9	66,3
TiO_2	22	0,79	0,81	0,13	0,68	0,87
Al_2O_3	22	15,6	15,63	0,56	15,26	16,06
Fe_2O_3	17	1,97	1,9	0,67	1,55	2,32
FeO	17	2,48	2,44	0,49	2,25	2,85
MgO	22	1,73	1,67	0,42	1,49	1,87
CaO	22	0,073	0,08	0,015	0,06	0,08
MnO	22	3,21	3,22	0,69	2,69	3,76
Na_2O	22	3,95	4,02	0,37	3,72	4,19
K_2O	22	2,51	2,45	0,29	2,29	2,65
H_2O^+	22	0,59	0,35	0,51	0,27	0,98
H_2O^-	17	0,36	0,33	0,1	0,3	0,39
CO_2	17	0,96	0,85	0,35	0,73	1,04
F	22	0,055	0,055	0,016	0,05	0,06
Cl	17	0,015	0,01	0,01	0,01	0,02
Ppm						
B	16	9,7	10,0	4,7	8,1	10,0
Ba	22	1220,0	1312,0	408,0	1000,0	1374,0
Be	22	1,9	1,9	0,4	1,5	2,5
Ce	20	106,0	114,0	28,0	86	118,0
Co	22	7,4	5,0	6,0	4,0	10,0
Cr	20	32,0	28,0	13,0	25,0	35,0
Cu	16	5,4	4,5	5,4	2,0	6,2
Ga	22	19,0	19,0	2,7	18,0	20,0
Li	22	27,1	27,5	7,1	21,0	33,0
Ni	22	10,7	9,4	4,4	8,0	14,0
Pb	16	9,2	8,0	5,3	6,2	10,0
Rb	22	71,0	62,5	23,6	54,0	84,0
Sn	16	2,4	2,0	1,5	1,5	2,5
Sr	22	656,0	712,0	198,0	562,0	794,0
Y	22	18,7	19,0	7,2	13,0	23,0
Zn	16	73,5	74,5	13,0	68,0	81,0
Zr	22	218,0	221,0	40,0	193,0	246,0

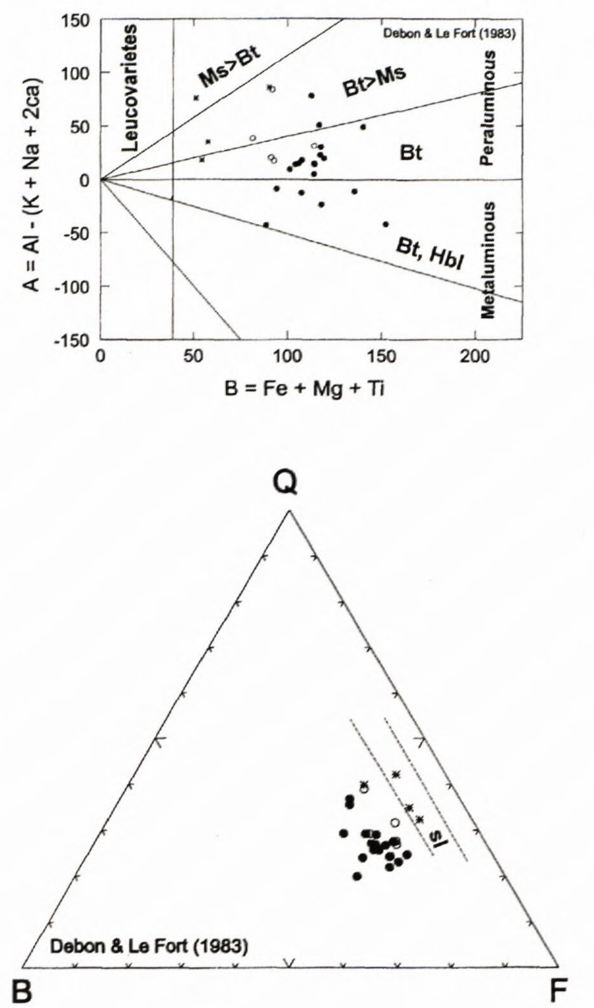


Fig. 6: A - Multicationic diagram A-B of Debon-Le Fort (1983) B - Multicationic diagrams B-Q-F of Debon-Le Fort (1983), Q = quartz, B= mafic minerals, F= feldspars, sl – subleucocratic facies (symbols see Fig.7)

A typical SiO_2 content of Sihla is around 65 wt. %, Ipeľ is around 69 wt.% and Biele vody – more leucocratic granitoids have slightly above 70 wt. % (Tab. 2, 3a-b). The contents of FeO, MgO, CaO, TiO_2 also increasing in the range Sihla-Ipeľ-leucocratic type as well as ratio K_2O/Na_2O.

Composition of Sihla type is metaluminous to peraluminous with a higher content of mafic minerals, Ipeľ and Biele vody types are more peraluminous (Fig. 6 a). Compositional variations in Fig 6b are caused by a different quartz/feldspar ratio, biotite content, and presence or absence of the K-feldspar and high content of accessory minerals. Biele vody type belong to subleucocratic field.

Trace elements are presented in Tab. 2 and 3a,b. For Sihla is typical high Sr (712 ppm) and low Rb (63 ppm) content, Rb/Sr ratio rarely exceeds 0,1 (Fig. 7). In Ipeľ and Biele vody types, have the similar content of these elements (Sr = 326, 428 ppm respectively, Rb = 106, 86 ppm), Rb/Sr ratio is more than 0,2 (Fig. 7). In spite of low contents of K-feldspar in Sihla granitoids, high amount of Ba is typical for this type (1312 ppm in average). Ipeľ and Biele vody type have lower content of Ba although, extremely enriched samples are also presented (Fig. 7). The similar features is also valid for Zr – Sihla granitoids are richer (221 ppm), than associated granitoids (less than 200 ppm). REE patterns are without Eu negative anomaly (Fig. 8A, B) and show strong LREE enrichment (Ce/Yb_N = 10,5 - 14,9 for 3 samples, but also extreme values 34,9 and 55,2 for Sihla type; Ce/Yb_N = 5,4 - 7,1 for Ipeľ type – 2 samples; Ce/Yb_N = 13,3 – 22,9 for Biele vody type - 3 samples).

Tectonic reworking of granitoids

Alpine metamorphic overprint is distinctly observed in all investigated granite groups. Main lineation and fo-

Tab.3a.Chemical composition of selected Sihla type (Si) tonalites to granodiorites

Sample	HV-33	HV-36	DL-47	DL-97	D-27	HV-3	HV-37	HV-39	HV-40
Type	Si	Si	Si	Si	Si	Si	Si	Si	Si
	w. perc.								
SiO_2	61,77	64,77	65,77	66,89	66,70	67,08	64,44	65,00	65,34
TiO_2	0,86	0,90	0,86	0,67	0,87	0,80	0,84	0,76	0,90
Al_2O_3	15,43	15,56	15,87	15,87	14,96	15,26	15,57	15,68	15,23
Fe_2O_3	3,83	2,32	2,51	1,77	1,90	2,10	1,56	2,06	2,41
FeO	1,58	2,35	2,01	2,95	2,38	1,58	2,85	2,25	2,17
MgO	2,75	1,76	1,76	1,56	1,53	1,40	1,68	1,49	1,61
CaO	4,59	2,79	3,76	2,7	2,69	3,66	3,03	2,72	3,25
MnO	0,113	0,078	0,071	0,057	0,076	0,047	0,078	0,068	0,077
Na_2O	3,95	4,18	2,67	2,82	3,91	4,01	3,90	4,33	3,89
K_2O	2,46	2,42	1,95	2,25	2,53	2,26	2,69	2,65	2,45
H_2O+	0,12	0,43	0,16	0,12	0,32	0,23	0,28	0,56	0,27
H_2O-	0,21	0,33	0,26	0,27	0,39	0,16	0,35	0,41	0,32
CO_2	0,91	0,74	0,2	0,38	0,73	0,61	1,52	0,73	0,80
P_2O_5	0,39	0,33			0,30	0,45	0,32	0,29	0,32
F	0,05	0,07	<0,05	<0,05	0,05	0,05	0,06	0,06	0,06
Cl	<0,01	0,02	<0,03	<0,03	0,02	0,03	0,02	0,03	0,02
total	99,01	99,05	97,84	98,31	99,35	99,73	99,19	99,08	99,11
	ppm								
Ba	1374	1337	934	1423	1337	1378	1471	1340	1453
Be	1,3	1,4	1,3	2,4	1,3	1,7	1,5	2,1	1,5
Ce	118	115	69	54	146		120	113	113
Co	6	5	5	5	4	<2	5	3	4
Cr	25	50	20	25	40		35	35	35
Cu	0	2	3	26	4		2	2	2
Eu	2,1	2,6	1,8	1,4					
Ga	16	20	22	23	18	<20	18	20	18
Hf	5,5	5,8							
La	81	70	70	59	78	71	79	70	68
Li	34	34	14	16	28	17	27	18	19
Lu	0,51	0,33	0,18	0,07					
Nb	10,0	17,0	12,0	9,0	15,0	0,0	<5	9,0	11,0
Nd			44	37					
Ni	7	8	10	11	8	8	15	6	15
Pb		5	11	8	9		7	10	6
Rb	67	61	50	59	59	48	66	57	53
Sm	11,0	14,0	7,2	6,0					
Sr	722	699	1027	805	716	720	683	706	827
V	90	85	70	50	70	75	75	65	70
Y	35	30	19	8	27	24	12	23	21
Yb	2,9	2,1	1,2	0,4					
Zn		67	74	71	74		91	69	83
Zr	250	221	282	285	231	205	221	225	230
Rb/Sr	0,09	0,09	0,05	0,07	0,08	0,07	0,10	0,08	0,06

liation is NE-SW direction, following of Hercynian deformation of the same sense and both, massive and foliated granites, exhibit formation of new Alpine mineral assemblage. Final stage of Alpine deformation is represented by sericite-chlorite shists with abundant quartz-carbonate lenses.

Although, most of deformation in the granitoid bodies in the Veporic tectonic Unit are considered to be Alpine in age, ductile deformation of Sihla tonalite in some places, undoubtedly Hercynian in age, is represented by the low angle lineation (NE-SW direction) of biotite crystals, mainly in the contact zone with leucocratic varieties of the latest stage. This indicates emplacement of the intrusion of these younger magmatic products in the subvertical shear zone in relatively older Sihla tonalite. Preliminary Ar/Ar study of new formed muscovites (ca 280 Ma) of metamorphic origin (W. Frank et al.- unpublish. data) from the deformed Sihla type supports an idea of the pre-Alpine deformation of the Sihla tonalite, however, Alpine metamorphism and deformation is also present.

Evolutionary model of the Upper Carboniferous magmatism in the Veporicum.

It is clear from the rock geochemistry, that Sihla tonalites and younger magmatic intrusions of the Ipeľ type

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Tab.3b.Chemical composition of Biele vody (Bv) and Ipeľ type porphyric granitoids (Ip). Sample Si-Ip represents transitive type to Sihla type.

Sample	DL-125	DL-17	DL-27	HV-7	HV-8	HV-9	IV-13	GB-3	X-5
Type	BV	BV	BV	Ip	Ip	Ip	Ip	Ip	Ip-Si
	w. perc.			w. perc.					
SiO_2	71,17	69,26	72,89	66,94	71,56	66,95	69,76	68,77	68,45
TiO_2	0,40	0,57	0,32	0,56	0,32	0,58	0,38	0,51	0,68
Al_2O_3	14,9	15,97	15	15,98	14,47	15,80	14,72	15,72	15,64
Fe_2O_3	0,91	1,82	1,02	1,31	0,60	0,98	1,29	1,31	1,63
FeO	1,58	1,87	1,15	2,31	1,51	2,70	2,02	1,94	2,01
MgO	0,75	1,34	0,72	1,39	0,86	1,39	0,83	1,22	1,33
CaO	2,28	2,65	1,98	3,09	1,95	2,91	1,21	2,62	2,4
MnO	0,047	0,075	0,072	0,058	0,056	0,063	0,051	0,058	0,062
Na_2O	3,19	2,29	2,44	4,08	3,68	4,18	3,79	3,64	2,37
K_2O	3,46	2,83	3,3	2,42	3,66	2,55	3,59	2,79	2,97
H_2O+	0,38	0,26	0,16	0,16	0,16	0,20	0,60	0,22	0,12
H_2O-	0,05	0,28	0,26	0,31	0,21	0,19	0,36	0,15	0,3
CO_2	0,26	0,02	0,04	0,73	0,45	0,48	0,62	0,22	0,52
P_2O_5				0,23	0,14	0,25	0,16		
F	<0,05	<0,05	<0,05	0,02	0,01	0,04	0,04	<0,05	0,05
Cl	<0,02	<0,03	<0,03	<0,01	<0,01	<0,01	0,02	<0,02	0,03
total	99,38	99,24	99,35	99,59	99,64	99,26	99,45	99,18	98,56
					ppm				
Ba	1698	608	891	1014	1104	819	733	839	1063
Be	1,8	1,4	1,8	2,3	1,9	2,0	2,8	2,0	1,9
Ce	73	53	36	71	51	72	52	23	60
Co	3	<1	4	2	<1	2	1	4	4
Cr	23	30	20	30	10	15	35	20	25
Cu	7	2	8				4	4	3
Eu	1,0	1,1	0,7	1,2				1,1	1,2
Ga		23	20	17	14	18	19		24
Gd	2,5							2,7	
Hf	4,8			3,6				5,3	
La	34	53	37	46	31	40	39	11	65
Li	27	15	11	34	24	32	32	20	20
Lu	0,11	0,12	0,10	0,33				0,11	0,12
Nb	6,0	12,0	7,0	11,0			7,0	7,0	11,0
Nd	29	38	22					12	36
Ni	17	13	16	2	4	12	6	12	12
Pb	44	15	22				15	17	20
Rb	84	80	94	97	108	120	124	79	96
Sm	5,0	7,3	3,7	16,0				3,0	5,7
Sr	473	394	419	390	314	321	220	384	440
Th	16							3	
V	34	55	25	50	25	45	35	52	50
Y	11	15	12	12	11	16	17	12	14
Yb	1,2	0,6	0,7	2,6				1,1	0,8
Zn	88	55	72				51	75	77
Zr	171	206	139	180	112	190	135	182	297
Rb/Sr	0,18	0,20	0,22	0,25	0,34	0,37	0,56	0,21	0,22

along with more leucocratic intrusions – Biele vody type, can not simply represent comagmatic intrusions. The genesis of this magmatic stage of the Upper Carboniferous Hercynian orogenesis is proposed according following time succession with decreasing age and decreasing of the melting temperature:

1. The oldest Sihla granitoid type probably represents a lower crustal melting product, triggered by under plated mafic melts with contribution of mantle. This stage produced the hottest I-type magma.

2. Younger or synchronous intrusions of coarse-grained porphyric K-feldspar granitoids (Ipeľ type-2nd stage) are derived from the lower crustal melting of a K-rich immature crustal protolith. This magmatic stage represents a broad melting episode during the final stage of the melting evolution.

3. The youngest fine (-medium) grained more leucocratic granodiorite (3rd stage) with S-type characteristics (Biele vody type) originated probably by lower temperature melting of quartz-plagioclase (± biotite ±muscovite) lower-middle crustal protolith, caused by thermal effect of Sihla and Ipeľ magmas. Low content of biotite and rare primary muscovite indicate low participation of the micas during the melting. The middle crustal

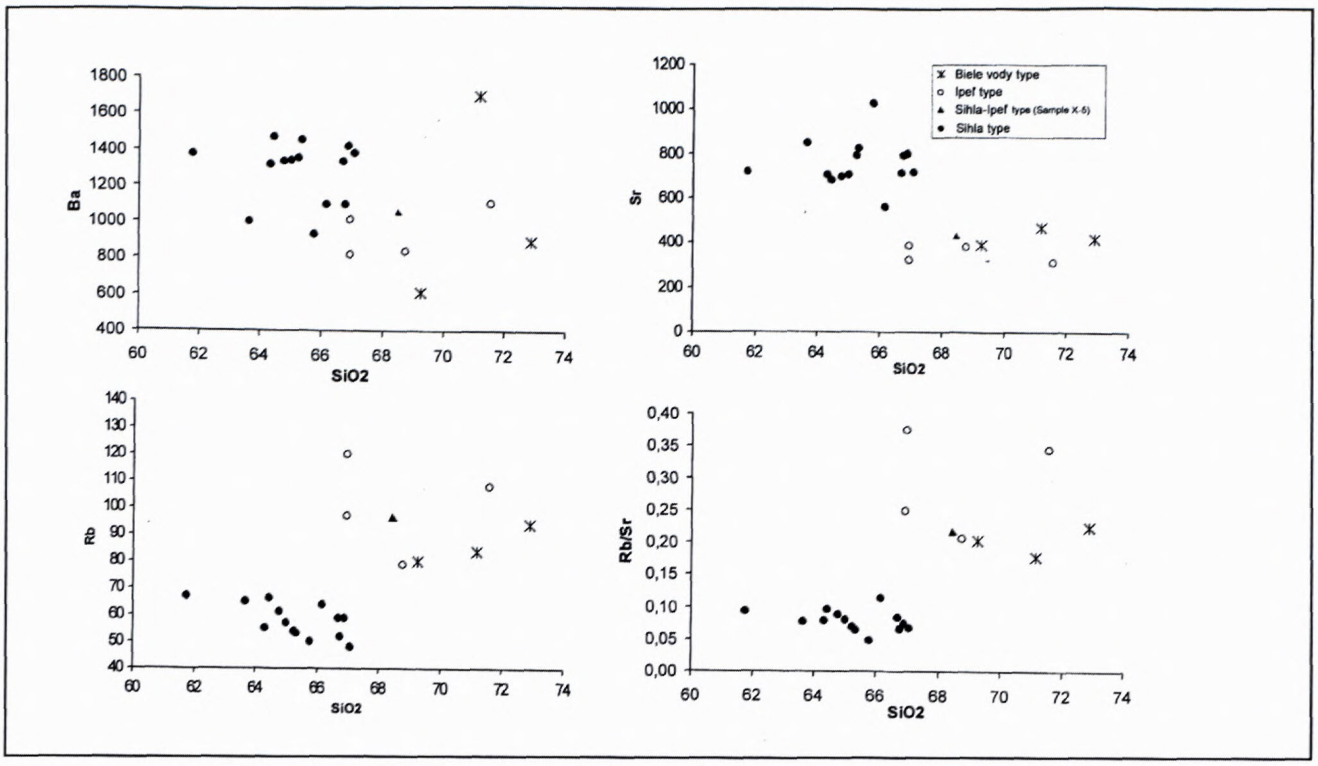

Fig. 7:Harker´s diagrams Ba, Rb,Sr and Rb/Sr versus SiO_2, for the granitoid types.

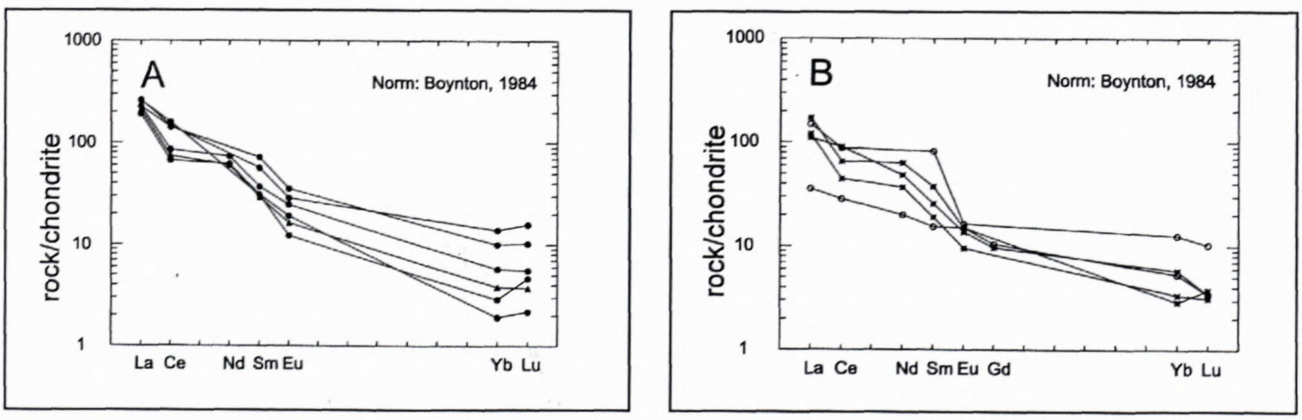

Fig. 8: Rock/ chondrite normalised REE spectra for selected samples (A - Sihla type, B – Ipeľ and Biele vody type). Explanations in Fig. 7

contamination is documented by the presence of migmatites and gneisses as xenoliths in the leucocratic magma. Unresorbed quartz xenocrysts reflect lower magma temperature or short time interval between contamination and magma solidification.

Conclusions

During the Upper Carboniferous magmatic stage, volumes of granitoid magmas intruded the basement of later Alpine Veporic Unit. The oldest stage is represented by intrusions of the hottest, most mafic, metaluminous to slightly peraluminous magmatic bodies (I-type) into the metamorphosed Pre-Carboniferous rocks and older granitoids. This Sihla type tonalite – granodiorite has a mixed lower crustal - mantle origin, similarly as 2nd Ipeľ magmatic stage, but with lower mantle contribution. During the time, progressive melting of a crustal material led to the middle crustal generation of magmas with typical crustal characteristics (S-type granite - 3rd stage) and the intrusions of the youngest magmatic portions (Biele vody type) was connected with the ductile deformation (NE-SW direction) of older Sihla tonalite. Considering the fact that the heat for the production of the 3rd magmatic phase was provided mainly by Sihla and Ipeľ granite melt, it is assumed that the magmatic events were stopped almost simultaneously. The three magmatic stages during the Upper Carboniferous is an example of the increasing of the crustal components during the magmatic evolution.

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Due to relatively narrow time span of intrusive succession in the Upper Carboniferous magmatic stage, precise isotopic datings are not effective, and so, the presented definition of three magmatic events follows mainly from the mapping field works.

Sample location:

Sihla type

HV-33 the big quarry near water dam wall in Málinec, 7 km SE from Detvianska Huta settlement

HV-36 Bzovský potok brook, 1650 m E from the elev. point Žiarina (575m),. 600 m NNE from Dolná Bzová village

DL-47 the big quarry near water dam near Hriňová village

DL-97 Kamenistá dolina valley, 620 m W from gamekeeper´s house, small stone-pit near the road, 3.5 km NW from Sihla village

D-27 900 m NW from the elev.point Vrchdobroč (918m) , 3 km S from Detvianska Huta village, 2,2 km NE from Jaseňový vrch hill, alt. 900 m

HV-3 = VG-54 Tlstý Javor elev. point, stone-pit near the road Čierny Balog-Hriňová

HV-37550 m W from Kováčsky vrch hill (736m), road-cut near gamekeeper´s house in Horná Bzová settlement, alt. 600 m

HV-391.6 km SW from Lom nad Rimavicou village, Studená voda valley, 1.35 km N-NNW from the elev. point 1018 m, alt. 815 m

HV-40Studená voda valley, 1 km N-NNW from the elev. point Magura (911 m), alt. 710 m

Ipeľ type

HV-7 road-cut, 1.4 km SW from Kostolný vrch hill (1058m), 4.3 km ENE form Klenovský Vepor hill, 11 km NNW from Klenovec village, alt. 740 m (out of the schematic map)

HV-8 road-cut, 2.1 km SSW from Kostolný vrch hill (1058m), 4.7 km E from Klenovský Vepor hill, 9.5 km NNW from Klenovec village, alt. 610 m(out of the schematic map)

HV-9 Mazúrka valley, 1.2 km from the elev. poin Rozsypok (1128 m), 2.7 km E from Klenovský Vepor hill, 11 km NW from Klenovec village, alt. 780 m

IV-13 E slope of Hronec valley, 900 m WSW from elev. point 1217 m, 5 km SSW from Závadka nad Hronom village, alt. 1020 m (out of the schematic map)

GB-3 Veporský potok valley, 6 km NW from Klenovec village, 2 km NE from Cisárska hoľa and 3.5 km from Oltárne (1034) elev. point

Transitional Sihla-Ipeľ type

X-5 100 m V form the road junction Málinec-Utekáč-Hriňová, 3 km NW form Šoltýska settlement

Biele vody type

DL-125 road-cut Sihla-Utekáč, 1 km E from Síhla village, alt. 915 m

DL-17 Biele vody valley, 750 m SSE from the elev. point Magura (911 m), 1.5 km SW from Biele vody settlement

DL-27 Sučí potok brook, 2.5 km NNE from Detvianska Huta village, 370 m W from the junction with Jankovský potok brook

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Slovak Geol. Mag., 6, 4(2000), 441–456

Jointing in Eocene flysch strata of the mid-eastern Magura Nappe, Polish outer Carpathians: implications for the timing of deformation

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Abstract. The paper focuses on early joint pattern in the mid-eastern segment of the Magura nappe in the outer West Carpathians of Poland. The studied stations in the Krynica, Bystrica and Rača subunits of the Magura nappe represent northward-younging members of the Eocene flysch sequence. Morphological properties of cross-fold shear or hybrid shear joints point to their formation at the time when the host strata were nor fully lithified, whereas their geometric relation to map-scale folds implies genetic relationship with the early stages of syndepositional folding. The age of this folding was different in different subunits of the Magura nappe, migrating in time from the Early through Late Eocene times. The associated shear joint-related σ_1 showed a 15 ° counterclockwise rotation, from N35 °E in Krynica subunit to N10 °E in Rača subunit, possibly due to a change of the sense of subduction of the European plate under Alcapa. The subsequent Miocene folding and thrusting, as well as minor rotations of fault-bounded blocks disturbed the Eocene pattern of cross-fold jointing. Fold-parallel, extensional joints represent younger episodes of deformation, related to post-orogenic, variably orientated extension.

Key words: jointing, palaeostress reconstruction, Magura nappe, outer West Carpathians, Poland

Introduction

The aim of this paper is to test usefulness of joint pattern analysis in reconstructing structural history of a portion of the outer Carpathian fold-and-thrust belt that shows variable orientation of map-scale folds and thrusts.

The outer Carpathian belt was formed as an accretionary prism during the southward-directed subduction of the European plate under Alcapa (i.a. Pescatore & Ślączka, 1984; Oszczypko & Żytko, 1987; Tomek & Hall, 1993; Zoetemeijer et al., 1999; and references therein), resulting in north-verging folding and thrusting, followed by major rotation of either the regional stress field (Aleksandrowski, 1985 a; Decker & Peresson, 1996; Decker et al., 1997; Zuchiewicz, 1998 a) or the belt itself (Márton et al., 1999), postdated by regional collapse associated with normal faulting (Decker et al., 1997).

Some authors claim that jointing generally precedes faulting (Shepherd & Huntington, 1981; Segall & Pollard, 1983) and folding (i.a. Cook & Johnson, 1970; Tokarski, 1977), whereas others maintain that systematic jointing in fold-thrust belts postdates the main orogenic compressional event (Meere & Rogers, 1999) and that it can rarely be used as a far-field stress indicator (e.g. Pollard & Aydin, 1988). Still another group of geologists concludes that joints can be initiated before, during and after folding (Hancock, 1964, 1985; Rixon et al., 1983). It has recently been suggested that joints in young fold-and-thrust belts are related to far-field stresses and that they can easily be used in palaeostress reconstructions (Mastella et al., 1997; Świerczewska & Tokarski, 1998; Zuchiewicz et al., 1998; Tokarski & Świerczewska, 1999). Testing this hypothesis in a structurally complicated portion of the largest outer Carpathian nappe is the main objective of our study.

We have chosen the mid-eastern segment of the Magura nappe (Fig. 1) because this area has a fairly good coverage by detailed geological maps and that jointing has been studied here extensively during the past decade. The term "joint" is applied in this paper as a field term, following the definitions given by Hancock (1985) and Dunne & Hancock (1994).

State of research

Joints are ubiquitous structures in the Cretaceous through Tertiary flysch strata of the Polish outer Carpathians, and have been studied by numerous authors (i.a. Bober & Oszczypko, 1964; Książkiewicz, 1968; Tokarski, 1975, 1977; Lenk, 1981; Mastella, 1988; Aleksandrowski, 1985 a, b, 1989; Mardal, 1995; Zuchiewicz & Henkiel, 1995; Mastella et al., 1997; Zuchiewicz, 1997 a, b, 1998 a, b; Rubinkiewicz, 1998; Tokarski & Świerczewska, 1998). Relation of jointing to regional fold trends has become easy to establish owing to recently published calculations of map-scale fold axes within homogeneous domains throughout nearly the whole of the Polish outer Carpathians (Mastella et al., 1997; Zuchiewicz et al., 1998; Szczęsny, 1998).

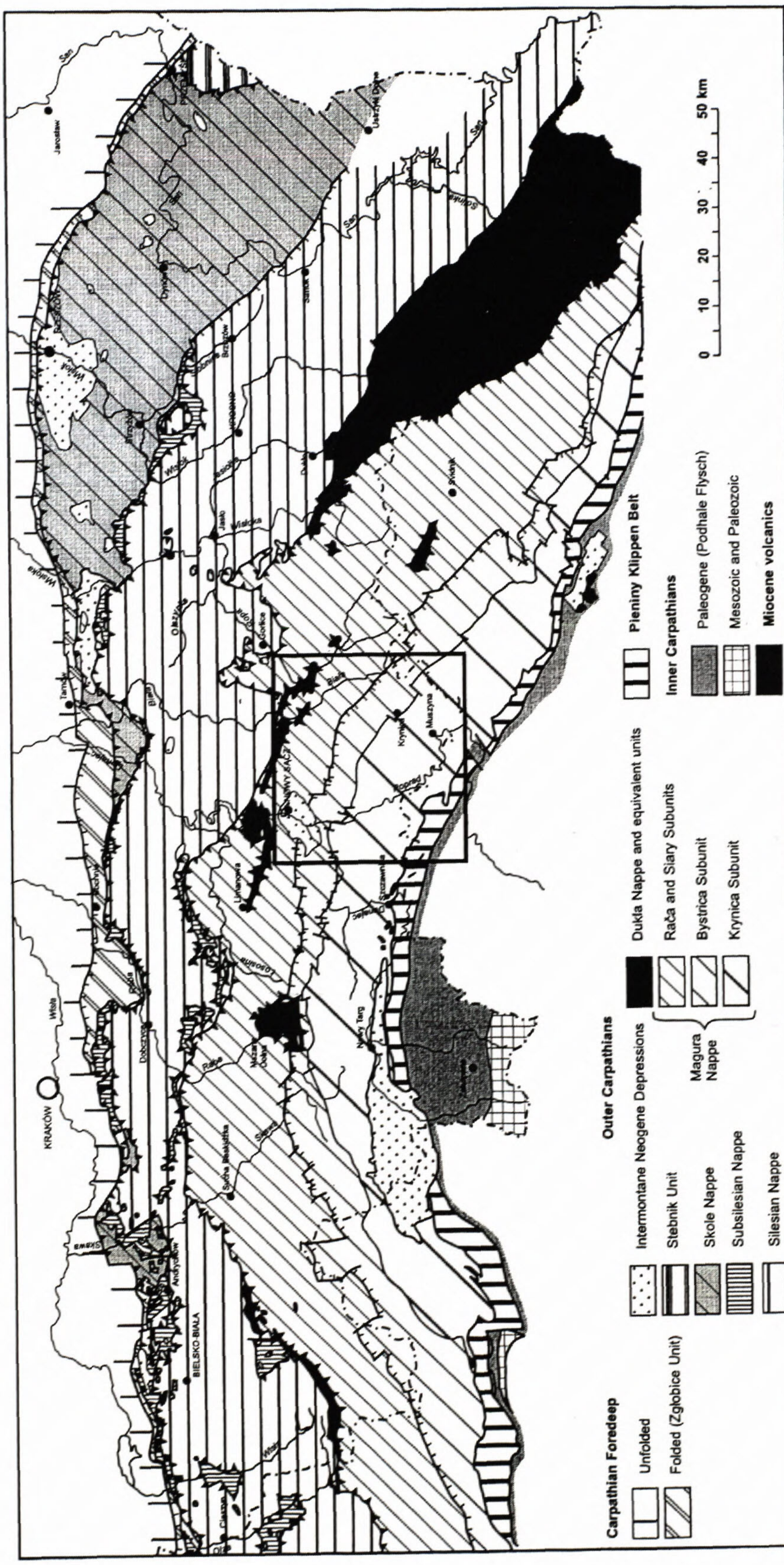


Fig. 1: Tectonic sketch of the Polish Carpathians (based on Żytko et al., 1989). Area shown in Fig. 2 is boxed.

Książkiewicz (1968) was the first who distinguished in the Polish outer Carpathians transversal (cross-fold joints clustering into two conjugate sets of shear or shear-extensional joints, whose acute bisector strikes perpendicular to map-scale folds), and longitudinal (fold-parallel, mostly extensional) joint sets. The former were to be associated with folding (the age of jointing being different depending on depth), the latter were considered to postdate the main episode of folding. Some of cross-fold joints were thought to be extensional fractures. Książkiewicz (*op. cit.*) also noted the presence of diagonal joints, less distinctly marked as compared to the other sets, and thought them to have originated during post-orogenic uplift.

Tokarski (1975, 1977) hypothesized about pre-folding age of two sets of shear-extensional cross-fold joints, and syn-folding age of shear and extensional sets of longitudinal (fold-parallel) joints in the medial segment of the Magura nappe.

Aleksandrowski (1985 a, b, 1989) described from the Western outer Carpathians three sets of cross-fold joints, including a pair of conjugate shears (T_1, T_2) and one set of extensional joins (T'), two sets of fold-parallel joints (L, L'), as well as two sets of diagonal joints (D_T, D_L). Sets L, T_1 and T_2 were to originate during fold-and-thrust event in Oligocene-early Miocene times (F_L folding), L' and T' sets during the early Badenian compression ($F_{L'}$), and D_T and D_L sets in the post-early Badenian (Aleksandrowski, 1989) or post-early Sarmatian (Aleksandrowski, 1985 b) compressional event, associated with folding in the Eastern Carpathians (F_D). Longitudinal folds (F_L) in the Western Carpathians were to be superimposed by diagonal, buckle folds (F_D), the latter being accompanied by transversal folds (F_T) formed at the same time due to cross-fold faulting.

Mastella et al. (1997), summarising a few year study project on jointing in the Silesian nappe, have distinguished cross-fold joints, composed of a single set of extensional (T) and two sets of conjugate shear and/or hybrid (D_1, D_2) joints, formed in poorly lithified, horizontal strata before regional folding, as well as fold-parallel joints without giving any explanation as to origin of the latter. Similar conclusions can be drawn from papers published by Rubinkiewicz (1998), Tokarski & Świerczewska (1998) and Zuchiewicz et al. (1998) with respect to the other nappes in the Polish outer Carpathians.

Recently, Tokarski & Świerczewska (1999) and Tokarski et al. (1999) have concluded about early-folding age of cross-fold joints, followed by opening of fold-parallel joints during synsedimentary folding and thrusting in the Krynica subunit of the Magura nappe. Subsequent stages of structural evolution are thought to include several episodes of faulting, nearly exclusively utilising the pre-existing joint surfaces, and associated first with 90 ° dextral rotation of the outer Western Carpathians in middle Miocene times, and then, since the late Miocene, with structural collapse of the orogen.

Geological setting

The Magura nappe is the largest and innermost nappe of the Polish segment of the Outer Carpathians which during the thrusting movements has been completely uprooted from its substratum along the ductile Upper Cretaceous strata. The oldest rocks of this nappe (Albian/ Cenomanian spotty marls) are exposed only close to the southern margin of the Mszana Dolna tectonic window (Burtan et al., 1978; cf. also Fig. 1). On the basis of facies differentiation of Palaeogene deposits, the Magura nappe has been subdivided into four facies-tectonic subunits, namely the Krynica, Bystrica, Rača and Siary subunits (Koszarski et al., 1974; Książkiewicz, 1977).

Lithostratigraphy

The Upper Cretaceous-Palaeogene sedimentary sequence of the Magura nappe is subdivided into three turbiditic complexes of the Campanian/Maastrichtian-Paleocene, early through late Eocene, and late Eocene through early Miocene ages (Oszczypko, 1992; cf. also Fig. 4). Each complex begins with pelitic basinal deposits (variegated shales) which pass into thin- and medium-bedded turbidites with intercalations of allodapic limestones and/or marls, being replaced higher upwards by thick-bedded and, finally, thin-bedded turbidites.

The Upper Cretaceous sequence begins with variegated hemipelagic mudstones bearing intercalations of thin-bedded turbidites of Cenomanian/Turonian through Santonian/Maastrichtian age (Malinowa Fm.; cf. Fig. 4). The Malinowa Fm. passes upwards into thin- to medium-bedded turbidites of the 50-m-thick Kanina beds which include up to 30-cm-thick intercalations of turbiditic limestones (Cieszkowski et al., 1989). These strata are replaced higher up the section by thick-bedded sandstones and conglomerates of the Szczawina sandstones, 100 m to 350 m thick. The youngest member of this complex, 80-300 m thick, is composed of thin- to medium-bedded turbidites (Ropianka beds) of Paleocene age (Malata et al., 1996).

In the Krynica subunit, the variegated shales of the Malinowa Fm. are overlain by thin- to medium-bedded, calcareous turbidites of the 300-m-thick Szczawnica Fm., dated to Paleocene-early Eocene time (Birkenmajer & Oszczypko, 1989; Oszczypko et al., 1999 b). The Szczawnica Fm. passes upwards into thin-bedded turbidites of the Zarzecze Fm. (Early Eocene), 400-500 m thick, intercalated by thick-bedded sandstones and conglomerates of the Krynica Member, up to 250 m thick (Oszczypko et al., 1999 b).

North of the Krynica subunit, the Ropianka beds are overlain by 20-150-m-thick variegated shales of the Łabowa Fm., Early to Middle Eocene in age (Oszczypko, 1991). These shales pass upwards into thin-bedded turbidites of the Beloveža and Hieroglyphic Formations, a few hundred metres thick (Fig. 4). In the Bystrica subunit, the Beloveža Fm. is overlain by thin- to medium-bedded turbidites with intercalations of the Łącko-type marls (Oszczypko, 1991).

In all the subunits studied, the youngest deposits of Early through Late Eocene age belong to the Magura Fm., some 1,500 m thick (Oszczypko, 1991, 1992; cf. also Fig. 4). This formation is represented by thick-bedded turbidites and fluxoturbidites. In the Krynica and Bystrica subunits, the Magura Fm. includes Middle Eocene variegated shales of the Mniszek Mb. (Birkenmajer & Oszczypko, 1989). In the Krynica and Rača subunits, the Magura Fm. is locally overlain by the Oligocene Malcov Fm. (Birkenmajer & Oszczypko, 1989; Oszczypko-Clowes, 1998). The upper part of the Malcov Fm. in the Nowy Sącz Basin region, represented by marls and glauconitic sandstones (Oszczypko, 1973), has recently been distinguished as the Zawada Fm. of Early Miocene age (Oszczypko et al., 1999 a).

Tectonics

The Magura nappe is flatly overthrust upon the Fore-Magura group of units and partly upon the Silesian nappe (Fig. 2). East of the Raba river, the Fore-Magura group of units is represented by the Dukla nappe and the narrow and discontinuous Grybów unit, which is wedged between the Dukla and Magura nappes.

The total amplitude of the Magura overthrust is at least 50 km, the post-Middle Badenian displacement exceeding 12 km (cf. Oszczypko, 1998). The northern boundary of the nappe is of erosional character, whereas the southern one coincides with a subvertical strike-slip fault that follows the northern margin of the Pieniny Klippen Belt (PKB; Fig. 2). According to Nemčok (1985), the PKB and Magura nappe are separated east of Szczawnica by a subvertical backthrust fault.

The Magura sole thrust is developed mainly in ductile Upper Cretaceous variegated shales. The sub-thrust morphology of the nappe is highly differentiated, controlling the shape of the northern limit of the nappe and the distribution of its tectonic windows. The "embayments" of the frontal thrust are related to transversal bulges in the Magura's basement, whereas the "peninsulas" are located upon basement depressions. The zone of tectonic windows (Mszana Dolna, Szczawa, Klęczany, Ropa, Ujście Gorlickie, Świątkowa) is situated 10 to 15 km south of the

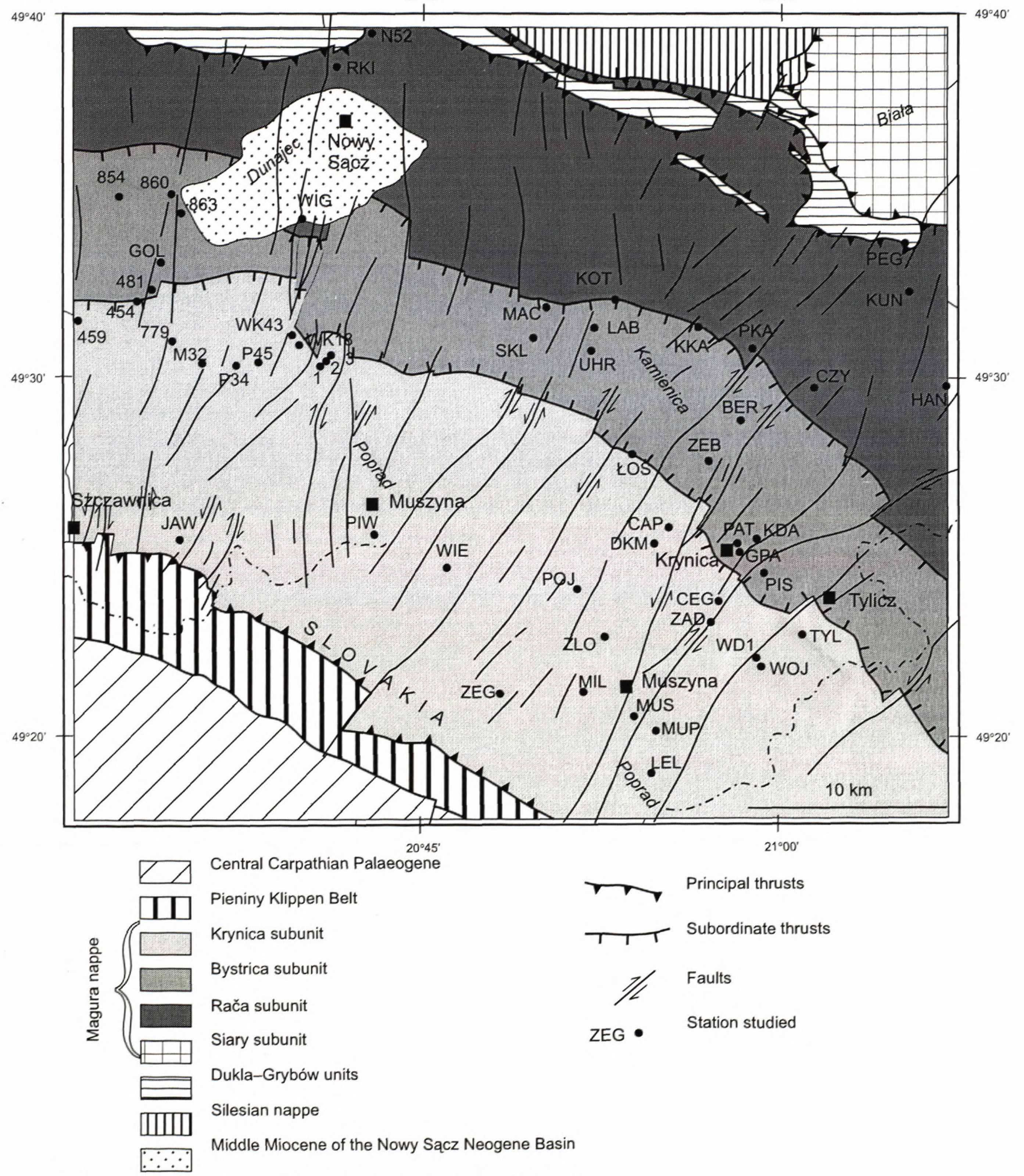


Fig. 2: Simplified structural map of the mid-eastern segment of the Magura Nappe.

Magura frontal thrust, coinciding with the elevated Fore-Magura basement. Farther south of the window zone, the dip of the Magura sole thrust increases, so at the northern boundary of the PKB the thickness of the nappe exceeds 5 km.

Among Polish geologists a controversy still exists as to the mutual relationship between facies zones and tectonic subunits of the Magura nappe. According to Książkiewicz (1977), in the western portion of the nappe, boundaries of individual facies zones do not coincide with major sublongitudinal thrust faults, whereas in the eastern portion tectonic and facies boundaries appear to follow one another (Sikora, 1970; Świdziński, 1972). As far as the eastern segment of the Magura nappe is concerned, only the northern limit of the Bystrica subunit is mapped as an important thrust fault (Żytko et al., 1989). In our opinion, the Krynica, Bystrica and Rača facies zones east of the Mszana Dolna tectonic window are separated by faults (cf. Oszczypko, 1973; Oszczypko & Wójcik, 1992; Oszczypko et al., 1999 b; Malata et al., 1996), the character of which, however, can be observed at few localities only. Due to the lack of necessary pieces of evidence, it is difficult to define these zones as thrust slices or thrust sheets. The Bystrica zone is the only unit that fulfills the definition of a thrust sheet. That is the reason why the above-mentioned zones have been named "facies-tectonic units" (Sikora, 1970; Koszarski et al., 1974), "facies-tectonic zones" (Oszczypko, 1973) or "tectonic subunits" (Birkenmajer & Oszczypko, 1989). The problem can be solved both by new drillings and detailed structural mapping of the contact zones.

West of the Poprad river, the Krynica and Bystrica subunits are probably separated by a W-E-trending thrust fault (Fig. 2). The Krynica subunit is thrust upon the youngest strata (variegated shales of Middle/Late Eocene age) of the Bystrica subunit. The frontal part of the Krynica subunit is represented by a narrow zone of few imbricated folds, composed of the Szczawnica and Zarzecze Formations, as well as by a syncline built up of the Magura Fm. (Oszczypko et al., 1999 b). East of the Poprad river, the Bystrica and Krynica subunits contact along a NW-trending, subvertical thrust fault dipping to NE (cf. Oszczypko et al., 1999 b). This fault, known as the "Krynica dislocation", is well recognised by wells drilled in the Krynica spa (Świdziński, 1972). In the latter area, the marginal part of the Krynica subunit is built up by a 200-500-m-wide anticline, composed of strongly deformed Paleocene-Lower Eocene strata of the Szczawnica Fm. Mesoscopic imbricated folds, accompanied by shear zones lined with 20-cm-thick calcite veins are a common feature in this zone. The overlying, Lower through Middle Eocene strata of the Zarzecze and Magura Formations are moderately deformed. It is important to note that the eastern segment of he Krynica fault and the northern boundary of the PKB are of backthrust character.

The Bystrica and Rača subunits are separated by a NW-trending thrust fault (Figs. 2, 3). The base of the Bystrica subunit is composed of Lower Eocene strata of the Beloveža and Łabowa Formations. In the Nowy Sącz Basin area, the Bystrica subunit is subvertically thrust upon Lower Miocene strata of the Rača subunit (Oszczypko et al., 1999 a). This thrust fault was reactivated during the Late Badenian.

All these units in the discussed portion of the Magura nappe display two different orientations of map-scale folds. -diagrams constructed by Szczęsny (1998) for homogeneous structural domains reveal that in the western part of the study area W-E to WNW-ESE orientated, westward-plunging (2-9°; in the northern Bystrica subunit exceptionally 15°) folds occur, whereas in the eastern segment (i.e., east of 20 °45 'E) folds plunging 3-16° southeastwards dominate.

The study area is cut by numerous, mostly NE-trending, both sinistral and dextral strike-slip faults (Fig. 3). Of particular importance is the Poprad-Dunajec fault system (Fig. 2) which was active at the time of deposition of the Late Badenian-Sarmatian infill of the Nowy Sącz Basin (Oszczypko, 1973; Oszczypko et al., 1992). Detailed studies in the Krynica area imply that the originally N35 °E - orientated sinistral faults were subsequently reactivated as dextral and dextral-reverse ones, whereas the younger generations of predominantly normal faults strike N10 °W and, less frequently, N45 °W and N85 °W (Zuchiewicz, unpublished data).

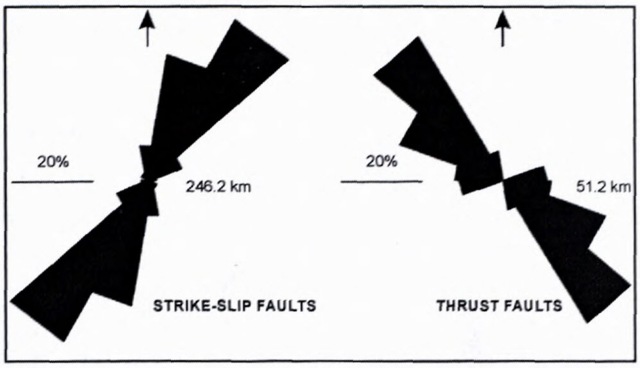


Fig. 3: Dominant orientation of map-scale thrust and strike-slip faults in the eastern segment of the study area (between Poprad and Biała rivers).

The Magura nappe reveals upward-diminishing degree of tectonic deformation. Close to the Mszana Dolna and Szczawa tectonic windows, the basal part of the nappe composed of upper Cretaceous-Paleocene flysch strata is strongly deformed (cf. Oszczypko et al., 1991, 1999 c), whereas in the Rača and Krynica subunits, built up mostly of Eocene flysch strata, broad, E-W trending synclines, separated by narrow anticlines, predominate. The southern limbs of synclines are frequently reduced and overturned. In the Bystrica subunit, in turn, subvertical thrust faults are a common feature. Both the northern limbs of anticlines and southern limbs of synclines are tectonically reduced and usually overturned. The youngest, Late Eocene-Early Oligocene, weakly deformed strata of the Krynica and Bystrica subunits unconformably overlie the older Eocene rocks (cf. also Oszczypko, 1973; Oszczypko & Żytko, 1987).

Material and methods

We have analysed 55 stations located in Eocene flysch strata of variable thickness and age that are exposed in the Krynica, Bystrica and Rača subunits of the mid-eastern segment of the Magura Nappe (Figs. 2, 4). Most of the data from the Krynica subunit represent Lower and Middle Eocene strata, those of Bystrica and Rača subunits coming from the Middle and Upper Eocene rocks. In all the subunits, stations located in thick-bedded sandstones dominate (75 %; Fig. 4).

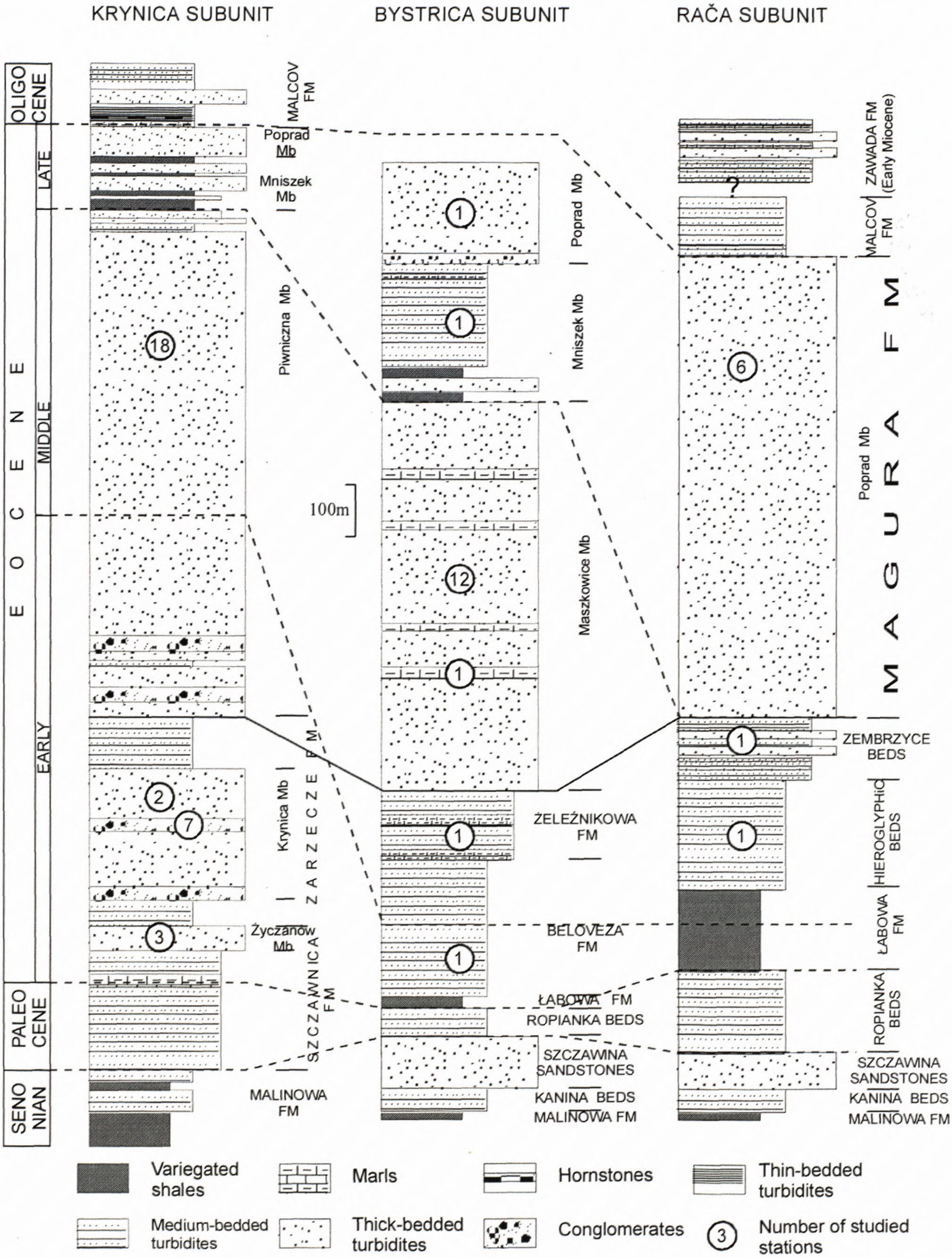


Fig. 4: Lithostratigraphic logs of the Krynica, Bystrica and Rača subunits. Circled numbers denote the number of stations studied.

At each station 50 to 100 measurements have been taken. The data have been plotted on lower hemisphere Schmidt projection and then bedding- and fold plunge-corrected. Interpretation of individual stereoplots is shown in Figs. 6 and 7.

Joint pattern: description

The regional joint network comprises five sets (Figs. 5-7). At individual exposures, however, usually two to four sets can be encountered. The sets maintain a relatively stable orientation in respect to the strike of map-scale folds. Sets showing the same regional orientation display both similar surficial features and the type of intersection with bedding surfaces.

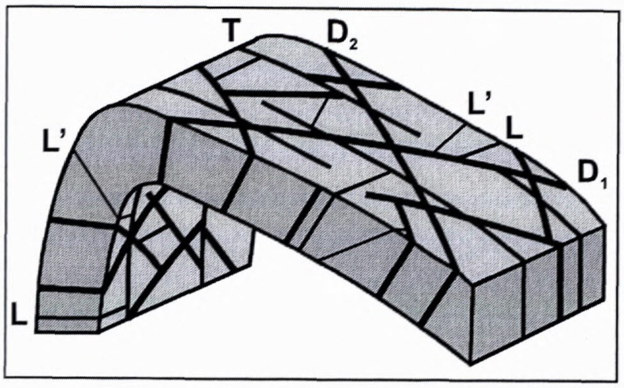


Fig. 5: Joint pattern in a folded sandstone bed (after Tokarski et al., 1999; modified).

Cross-fold joints comprise a single set (T) striking perpendicular or subperpendicular to map-scale fold axes (80-90°) and two sets (D_1, D_2) striking under high angles (60-80 °) to these axes. The acute bisector between these two sets is orientated perpendicular to map-scale folds. *Fold-parallel joints* (L, L') strike parallel or under small angles to map-scale fold axes and are perpendicular or subperpendicular to bedding (70-90°).

Cross-fold joints

The surfaces of *T joints* are non planar, and their traces on bedding surfaces are usually curvilinear. The T joints are not accompanied by feather and *en echelon* fractures, and are lined by mineral (usually calcite) veins, particularly in calcareous thick-bedded sandstones. Joint surfaces are usually devoid of fringe structures. Fissures associated with this set are commonly open, unlike those related to the diagonal sets.

The strike of T joints varies along the studied segments of the Magura nappe from N30 °W to N60 °E in Raca, through N55 °W to N70 °E in Bystrica, to N25 °W - N75 °E in Krynica subunits, clustering at N15 °E and N5 °W in Bystrica and N15 °E in Krynica subunits (Figs. 6-8). These orientations are highly scattered, depending on local attitude of map-scale fold axes which changes from W-E to NW-SE and WNW-ESE in the western, central and eastern portions of the studied segment of the Magura nappe, respectively.

The surfaces of *D_1 and D_2 joints* are planar and their traces on bedding surfaces are rectilinear. Some of D_1 joints terminate on D_2 joints and vice versa. Both sets intersect one another under acute angle, whose bisector is orientated NNW to NE throughout the study area (Figs. 6, 7, 9). Fissures associated with both the sets, a few millimetres wide, are filled at places by calcite.

Numerous joints of the D_1 and D_2 sets are accompanied by millimetre-scale feather fractures of dips not exceeding 30° that represent low-angle Riedel shears (*sensu* Riedel, 1929; Bartlett et al., 1981). Locally, instead of a linear trace, an *en echelon* array composed of low-angle Riedel shears can be encountered, passing sometimes into a continuous joint surface. Some of *en echelon* cracks are filled by material derived from country rocks or by calcite contaminated by this material.

The orientation of D_1 joints changes from N25 °W in Raca and Bystrica subunits to N5 °E in Krynica subunit (Figs. 6-8), that of D_2 joints being N40 °E, N55 °E, and N45 °E in Raca, Bystrica and Krynica subunits, respectively. The scatter is also notable, reflecting differentiated strike of map-scale folds of the region. No regional variability of orientation of both these sets has been observed, although Bystrica subunit displays a wider scatter of dominant joint attitudes. The acute angle between the D_1 and D_2 joints is 50-60° in Raca, 30-80° in Bystrica, and 25-80° in Krynica subunits, averaging at 60-70°. The lowest figures (25-40°) have been encountered at individual stations in Krynica and Bystrica subunits, although average values for all the three units discussed tend to diminish northwards: from 61° and 60° in Krynica and Bystrica subunits, to 55° in Rača subunit (Figs. 6, 7).

Fold-parallel joints

Fold-parallel (longitudinal) joints (L, L') strike subparallel to the map-scale folds and comprise two sets of different orientation (Figs. 6-8). Set L comprises joints striking N65 °W, N30 °W and N55 °E in Rača, N75 °W, N55 °W and N60 °E in Bystrica, as well as N60 °W, W-E and N25 °W in Krynica subunits; those of set L' being orientated, respectively, W-E and N35 °E; N75 °E, N30 °W and N55 °W, as well as N65 °W, N35 °W and N85 °E. In general, the prevailing N65-75 °W strike of L joints in Rača and Bystrica subunits is locally overprinted by N30-35 °W and N55-60 °E orientations, whereas in Krynica subunit it is overprinted by W-E and, rarely, N25 °W orientations. The pattern of L' joints is even more chaotic. In most cases, the L joints are subparallel to the map-scale folds, usually occurring in hinges of large-scale structures and normal to the T joints (Figs. 8, 10). Joints of the L' set, in turn, strike under small angle (20-25°) to the map-scale fold axes and are nearly perpendicular to the acute bisector between the D_1 and D_2 sets.

Joint pattern: interpretation

Cross-fold joints

The morphology of *T joints* indicates that their development proceeded without the initial "shear" stage and

KRYNICA SUBUNIT

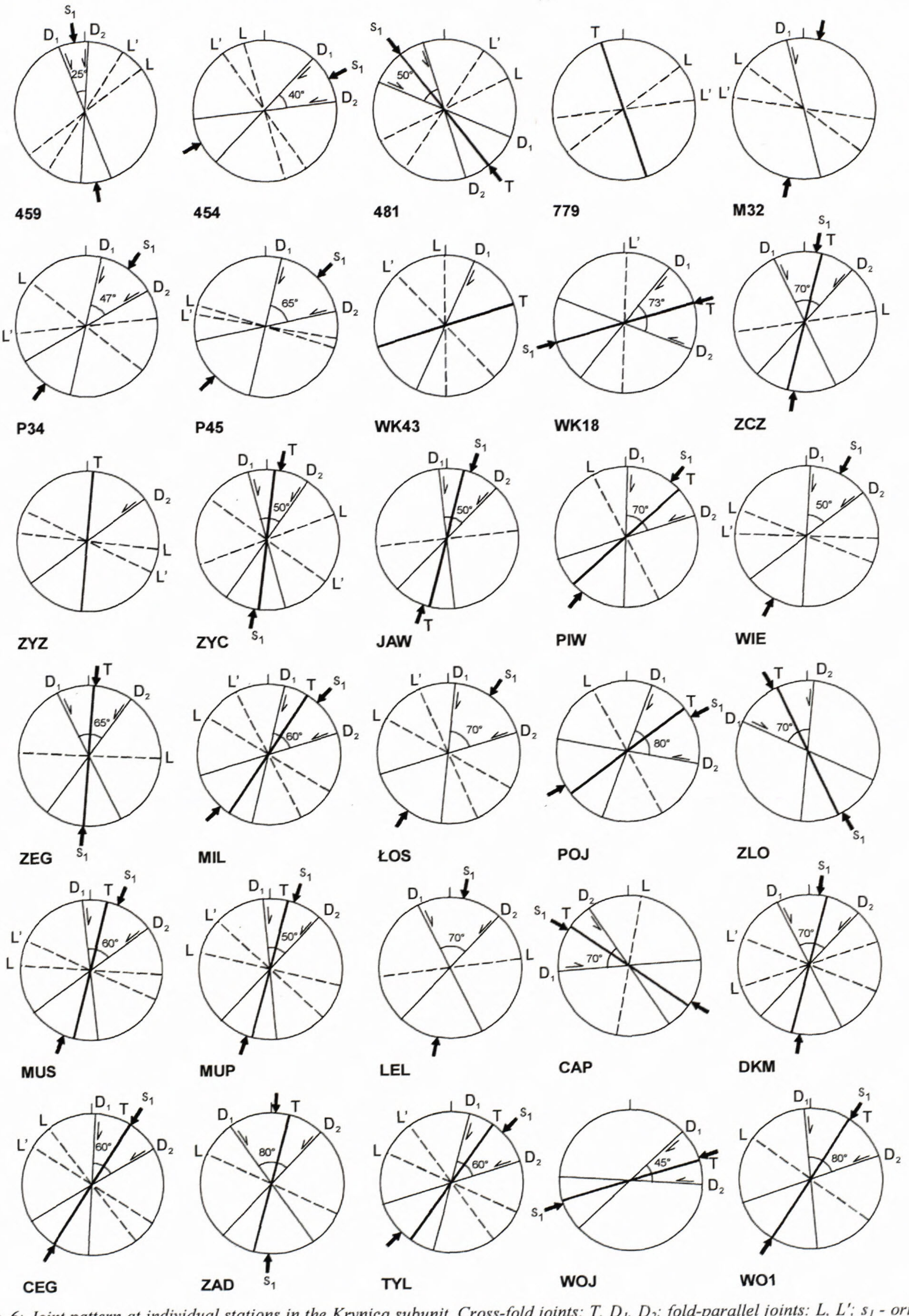
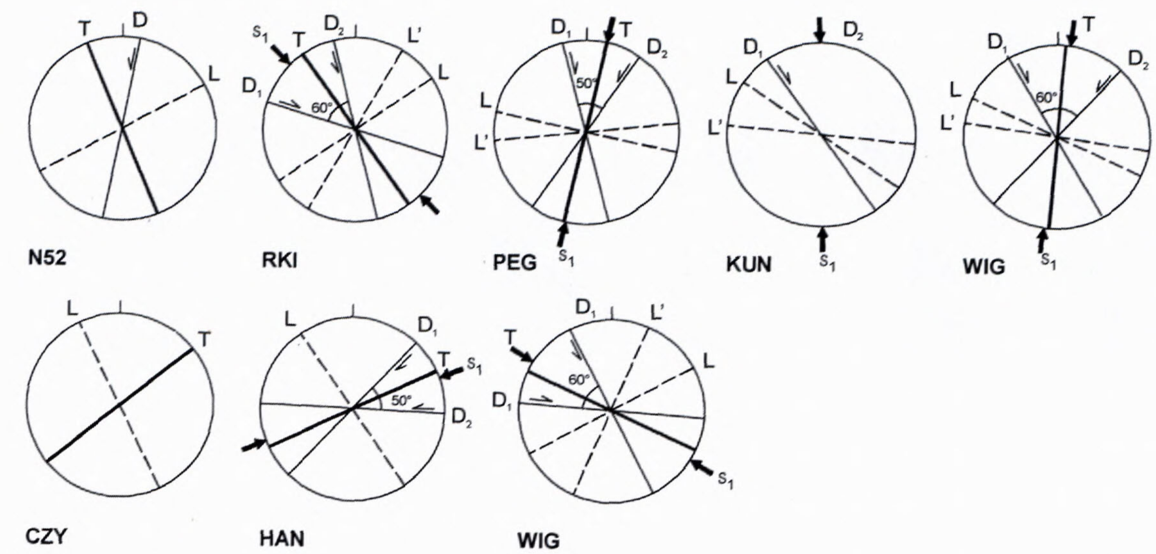


Fig. 6: Joint pattern at individual stations in the Krynica subunit. Cross-fold joints: T, D_1, D_2; fold-parallel joints: L, L'; s_1 - orientation of the maximum horizontal stress related to the cross-fold shear joint formation.

RAČA SUBUNIT



BYSTRICA SUBUNIT

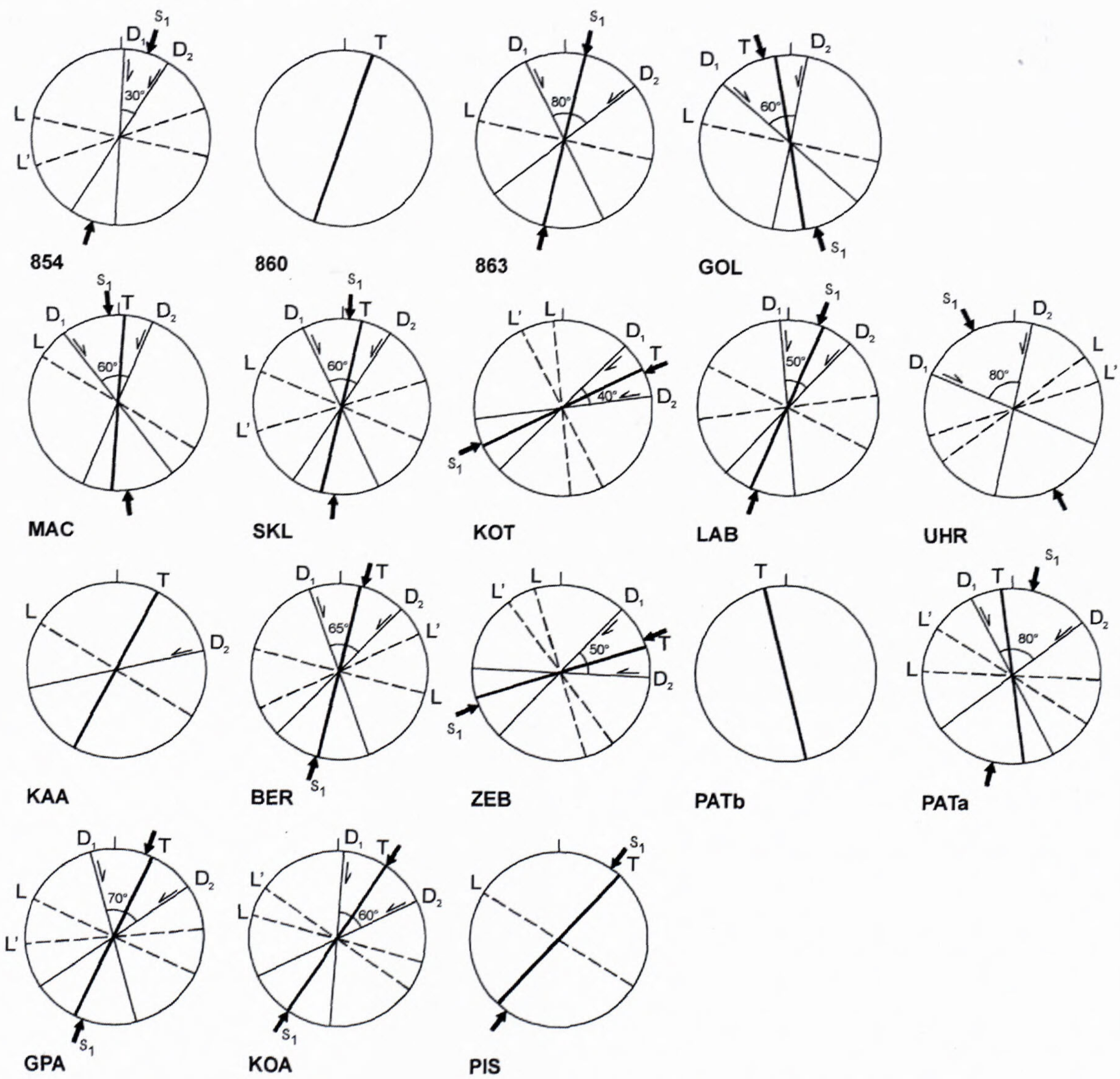


Fig. 7: Joint pattern at individual stations in the Bystrica and Rača subunits. For explanation - see Fig. 6.

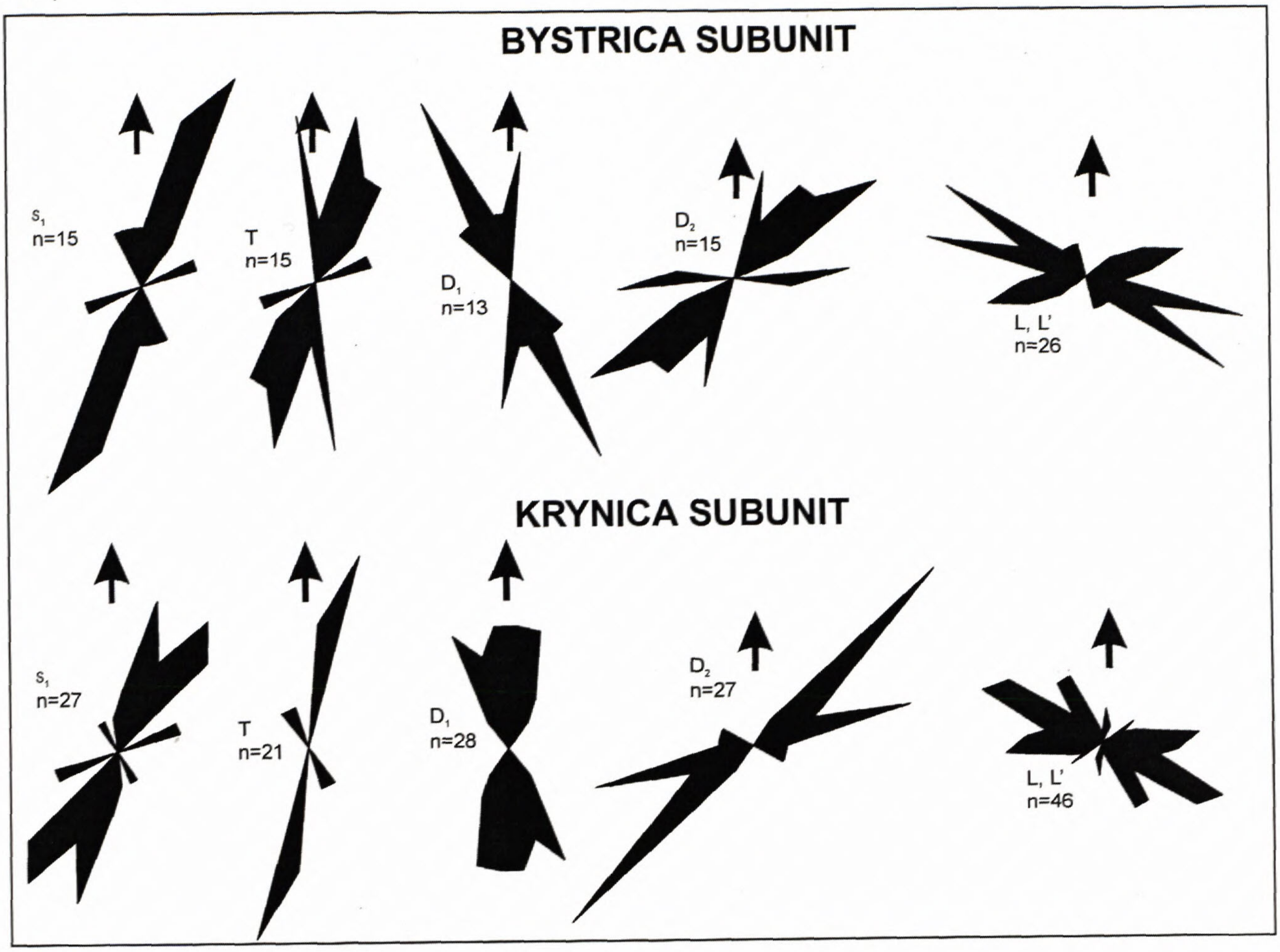


Fig. 8: Dominant orientations of principal joint sets in the Bystrica and Krynica subunits.

that they are extension (mode I) fractures (Price & Cosgrove, 1994).

Morphological properties of the *diagonal (D_1, D_2) joints* indicate that the incipient stage of their development was a shear one, whereas their further opening proceeded in extensional mode. The pattern of *en echelon* arranged gashes and feather fractures shows that the D_1 and D_2 sets represent, respectively, dextral and sinistral shears, whose acute bisector is orientated differently throughout the mid-eastern Magura nappe: from N45 °W - N70 °E in Rača, through N55 °W- N70 °E in Bystrica, to N30 °W - N70 °E in Krynica subunits. The prevailing orientations are, however, N10 °E, N20 °E, and N35 °E, respectively. Deviations from these dominant trends are-particularly noticeable for stations located close to major strike-slip faults, like those associated with the Poprad and Dunajec river courses or situated shortly west of Krynica (Fig. 9).

Furthermore, abutting relationships suggest that the D_1 and D_2 joints are roughly coeval and were formed as "potential shear surfaces" in the triaxial stress field. The local occurrence of plumose structures, in turn, points to subsequent extensional opening of some of these joints.

We suppose that these joints are shear and, at some stations in Krynica and Bystrica subunits, hybrid fractures (Hancock, 1985) that form a conjugate system. Moreover, some of the joints were formed when the host strata were poorly lithified, as indicated by contamination of mineral veins that fill the joints by material derived from the host strata.

The acute angle comprised between the D_1 and D_2 sets (double value of the angle of shear, 2 ?) changes at individual exposures from 25° to 80° (Figs. 6, 7). Nearly 75 % of our data, however, fall into the interval of 60-70°. A regional tendency towards not very significant decrease of these values from the south to the north (61° to 55°) is to be noted.

The extensional T joints, in turn, appear to represent a younger episode of deformation closely associated with folding, particularly that related to extension in fold hinges. This set is probably coeval with fold-parallel L joints.

The above results concerning cross-fold joints in the mid-eastern Magura nappe appear to fulfill criteria listed by Hancock (1985) which lead to a conclusion that extension joints usually form single sets, and that conjugate hybrid joints enclose small dihedral angles (10-50°, commonly 35-45°), whereas conjugate shear joints enclose angles of 60° or greater.

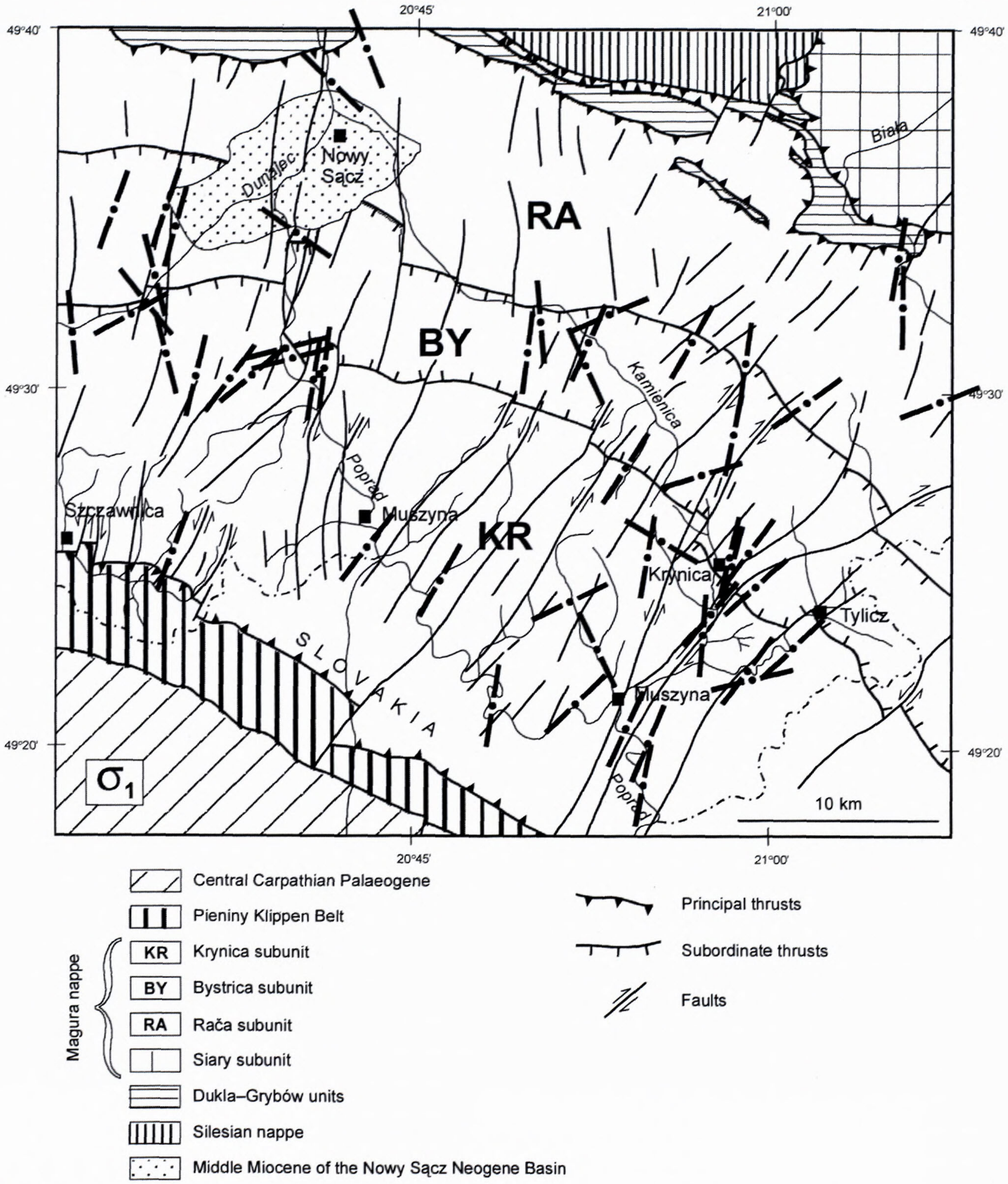


Fig. 9: Spatial distribution of cross-fold joint-related s_1 in the study area.

Fold-parallel joints

Both the L and L' sets are devoid of properties that would point to their shear origin. The morphology of joint surfaces, occurrence of plumose structures (particularly common in the L' set), and characteristic discontinuous or fading, at places nonlinear, traces of intersection with bedding planes appear to indicate extensional origin.

Discussion

Morphological properties of diagonal shear and hybrid joints imply that their formation occurred during early stages of syndepositional folding, when the host strata were not fully lithified. This is confirmed by the presence of calcite contamined by host rock-derived material within *en echelon* arranged veins related to the

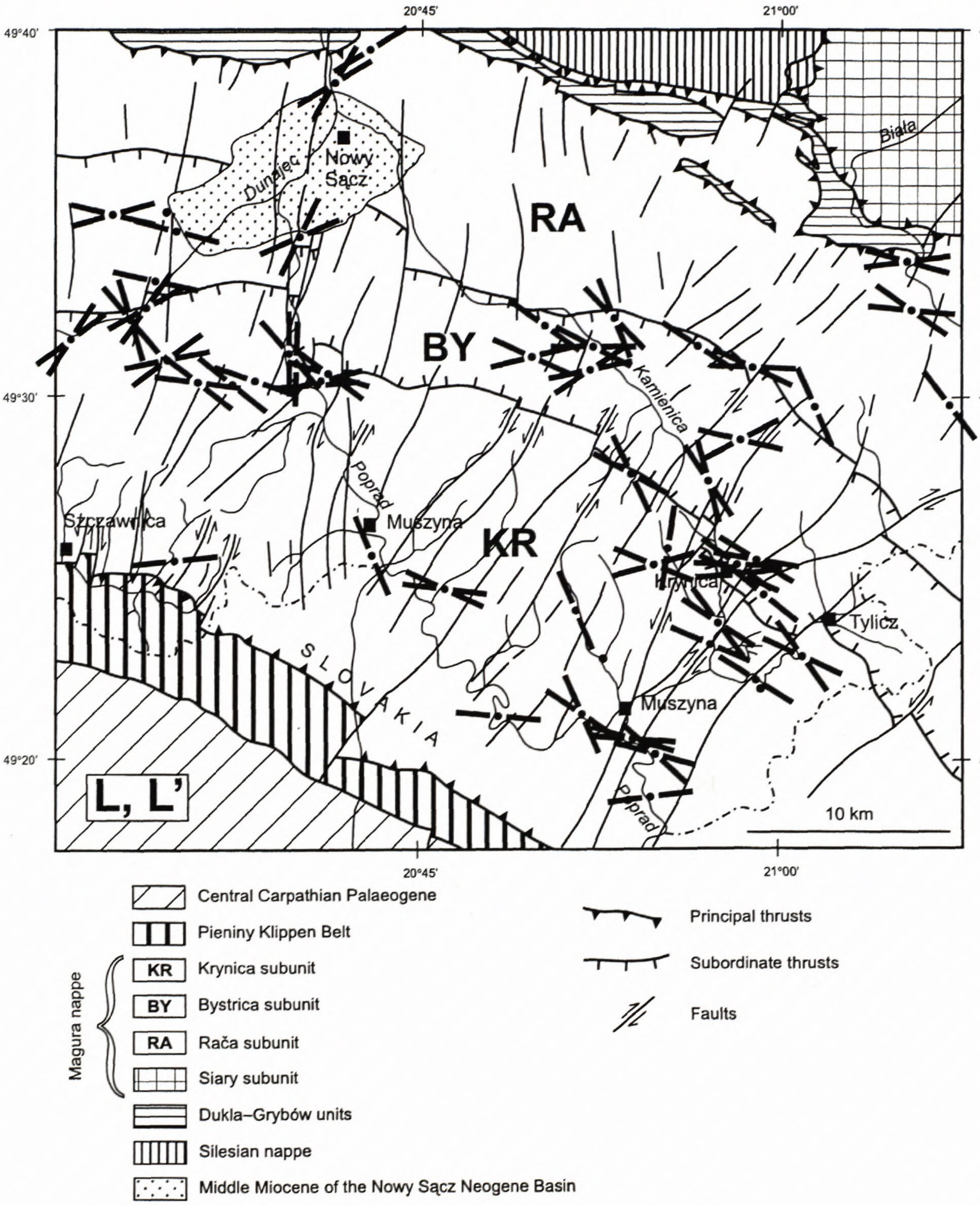


Fig. 10: Spatial distribution of fold-parallel joint sets (L, L') in the study area. Note that these sets are normal to the longitudinal joint-related s_3

cross-fold D_1 and D_2 sets. Fig. 11 shows the most frequently observed configuration of joint sets in the Magura nappe subunits, indicative of a minor anticlockwise rotation of the shear joint-related s_1 when proceeding from the south to the north, during the early through late Eocene folding. The acute angle comprised beween most frequently occurring diagonal joint sets appears to be changing from 40° in Krynica, through 80° in Bystrica to 65° in Rača subunits, whereas average values calculated for individual subunits tend to show another, northward-decreasing, trend. This apparent paradox results from diversified number of data available for each subunit and

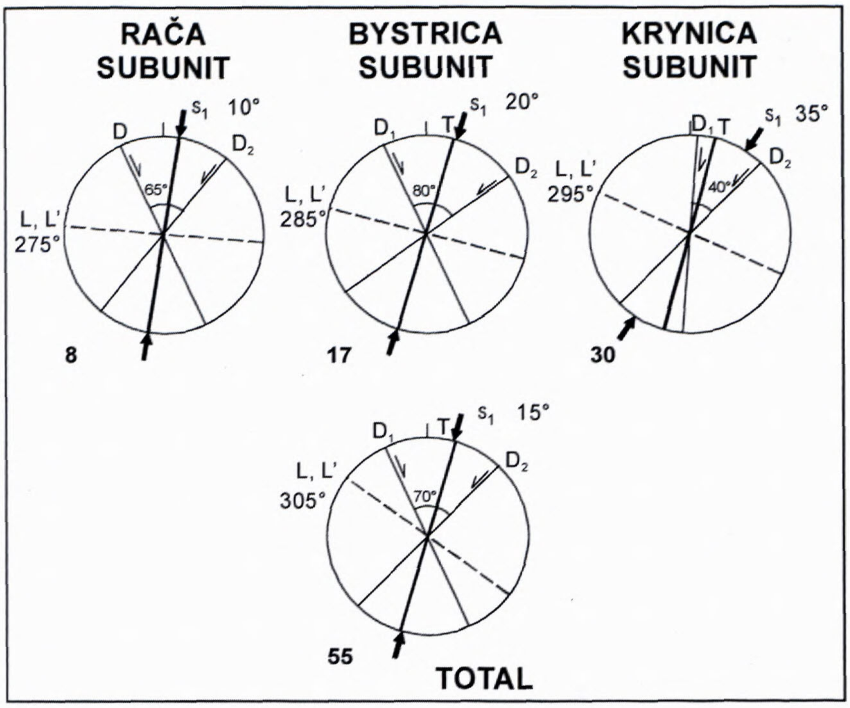


Fig. 11: Joint pattern in the mid-eastern segment of the Magura Nappe. Note anticlockwise rotation of the early joint-related principal compressive stress axis from the south to the north, i.e. from the Krynica, through Bystrica to Rača subunits.

their variable scatter. This contradiction makes it impossible - at this stage of research - to conclude about possible differentiation in overburden at the time of syndepositional folding and, consequently, shear jointing. The most frequently occurring pattern of cross-fold joints, however, points to relatively greater thickness of overlying strata in the Krynica subunit.

According to hitherto-published opinions, folding in the outer Carpathian basins took place several times: at the end of the Early Cretaceous, during the Laramian orogeny (Senonian - Paleocene), during the Eocene (synsedimentary folding in the Krynica subunit of the Magura nappe; cf. Żytko, 1977, 1999; Świerczewska & Tokarski, 1998), following the end of the Eocene and, predominantly, at the turn of the Oligocene/Miocene or during Miocene times (Książkiewicz, 1977; Oszczypko et al., 1991; Roca et al., 1995; Oszczypko, 1998; Ślączka & Kaminski, 1998).

The outer Carpathians are regarded as a remnant oceanic basin which developed between the colliding European continent and intra-oceanic arcs (cf. Oszczypko, 1999). Similarly to other orogenic belts, the outer Carpathians were progressively folded towards the continental margin. This process was initiated during the Paleocene by the formation of the Magura accretionary wedge. The first stage of basin shortening and coeval growing of the Magura accretionary wedge was probably completed before the Priabonian, being connected with submarine folding in the Bystrica and Krynica subbasins and development of the Krynica thrust. At the end of the Eocene, this part of the Magura basin was overlapped by the Globigerina Marls and then by flysch deposits of the Malcov Fm. (Oszczypko-Clowes, 1998). The northern part of the basin (Rača and Siary subbasins) was still occupied at that time by the deposition of the Magura Fm. and Wątkowa glauconitic sandstones, respectively. Deposition in the Magura basin persisted until the Late Oligocene (Oszczypko-Clowes, 1999). After the late Oligocene/early Miocene folding episode, the Magura nappe was thrust northwards onto the terminal Krosno flysch basin (Oszczypko, 1999). In the course of the Burdigalian transgression, the middle portion of the Magura nappe was invaded by the sea (Oszczypko et al., 1999 a). This transgression was followed by the intra-Burdigalian (late Ottnangian) folding, uplift, and thrusting of the marginal part of the Carpathians which overrode the European platform (Oszczypko, 1998). This episode of compression is manifested in the Magura nappe by thrusting of the Bystrica subunit onto the Burdigalian strata of the Nowy Sącz area. Subsequent shortening in the Magura nappe took place during the Late Badenian movements, leading to the steepening of the Bystrica and Krynica subunits. The last compressive event was postdated by formation of the Late Badenian/Early Sarmatian Nowy Sącz and Orava extensional basins.

According to Kovač et al. (1998), the maximum compressive stress in the Western Carpathians progressively rotated from NW-SE (Eggenburgian), through NW-SE (western segment) and NE-SW (eastern segment) during the Karpatian, to NE-SW in late Badenian and Sarmatian times.

The results of hitherto-conducted research into the history of jointing in the Polish outer Carpathians imply that the Palaeogene history of all the nappes was probably dominated by strike-slip stress regime, with more or less constant orientation of the maximum principal stress (Zuchiewicz et al., 1998). This regime has been active throughout the area since the Paleocene and, at least in the eastern segment of the Silesian nappe, since the Cretaceous (cf. Rubinkiewicz, 1998). The post-Palaeogene structural development of nappes located north of the Magura nappe was different from that of the Magura nappe (Zuchiewicz, 1998 a; Tokarski & Świerczewska, 1998; Zuchiewicz et al., 1998), in which the early joint pattern was disturbed by refolding, drag on cross-fold strike-slip faults, and/or by rotations of blocks bounded by gently-dipping shear zones (Aleksandrowski, 1985 b; Oszczypko et al., 1991; Decker et al., 1997; Zuchiewicz, 1998 a; Tokarski et al., 1999). The position of shear joints-related maximum stress axis was horizontal, its orientation changing from NW through NNE to NE in the western, medial and eastern portions of the outer Carpathians, respectively. Extension related to the formation of fold-parallel joints in the corresponding segments of the Carpathians was orientated NW, NW to N-S, and NE (cf. Zuchiewicz & Tokarski, 1999).

Deviations of the reconstructed cross-fold joint - related s_1 from the position normal to map-scale fold axes, observed at some places in the mid-eastern Magura nappe (Fig. 9), can be explained as a result of subsequent, i.e. post-Eocene episodes of folding and thrusting, as well as by rotation of some fault-bounded blocks, like those situated close to the Poprad and Dunajec river courses or west of Krynica.

Abutting relationships indicate that fold-parallel joints are younger features which originated during late stages of map-scale folding (L) or shortly after, during post-orogenic, mostly cross-fold extension (L').

Orientation of s_1 associated with the cross-fold shear joints coincides with that of recent horizontal stresses (NNE to NE) recorded by breakouts in the western and medial segments of the outer Carpathian flysch nappes (Jarosiński, 1998), but differ from those inferred from focal solutions of recent earthquakes (N to NNW) in the Krynica area (Wiejacz, 1994; Dębski et al., 1997).

Conclusions

The studied stations in the Krynica, Bystrica and Rača subunits represent northward-younging members of Eocene strata. Morphological properties of cross-fold shear or hybrid joints point to their formation at the time when host strata were not fully lithified, whereas their geometric relation to map-scale folds implies genetic relationship with early stages of syndepositional folding. The age of this folding was different in different subunits of the mid-eastern Magura nappe, migrating in time from the Early through Late Eocene times (cf. also Żytko, 1977, 1999). The regional stress field associated with these processes was probably characterised by a 15° counterclockwise rotation of shear joint-related s_1 from N35 °E in Krynica subunit to N10 °E in Rača subunit throughout the Eocene, possibly due to a change of the sense of subduction of the European plate under Alcapa (at present coordinates). Fold-parallel, extensional joints, represent younger episodes of deformation, related to post-orogenic, variably orientated extension. Local deviations observed throughout the area, and consisting in the position of shear joint-related s_1 different from normal to map-scale folds, are interpreted as a result of younger episodes of Miocene folding and thrusting that affected the whole studied portion of the Magura nappe, as well as due to still younger rotations of individual, fault-bounded blocks.

Acknowledgements

We would like to express our thanks to A. K. Tokarski for the permission to refer to his data concerning joint stations 454, 459, 482, 779, 854, 860 and 863.

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